Extract from
A Dynamic Stratigraphy of the British Isles – A study in crustal evolution

A Dynamic Stratigraphy of the British Isles

A study in crustal evolution

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Preface

This book is in essence a story; an illustrated description of the evolution of the British Isles. Although this in itself is not new the authors believe that there is a need for a change in the approach to the subject. The correlation of rocks is the most fundamental aspect of stratigraphy, but it is necessary to utilise other geological disciplines in order to discover the origin of rocks and unravel the evolution of a terrain. Geochemistry, geophysics, igneous and metamorphic petrology, palaeobiology, sedimentology and structural geology must all play their part together.

This text focuses on the geological events that led to the evolution of the small but complex part of the Earth's crust now called the British Isles. The level of treatment assumes a minimum background knowledge equivalent to a first-year course in geology at university. The evidence is drawn not only from the rocks of the British Isles but also from the adjacent terrains of the present and of the past. The book is divided into four parts. Although they are in chronological order the emphasis is on major events. Thus first, in Part 1, we consider the evolution of the early crust. This is followed in Part 2 by a study of the building of the Caledonides. In Part 3 we trace the development of the Hercynides, and in Part 4 we consider the post-Palaeozoic evolution of the British Isles on the passive eastern margin of the N. Atlantic Ocean.

Since the development of the theory of plate tectonics there have been numerous hypotheses and speculations concerning the evolution of the British Isles. Although there continues to be a substantial outflow of new data it is generally difficult to confirm or totally to refute these hypotheses. This means that our story, especially in the Palaeozoic and Precambrian, must be followed with caution. For this reason we have provided essential data in the form of diagrams and tables. It is hoped that readers will compare the interpretations and views expressed in the text with data located in the figures. Selected references are cited in the text, and key references, those considered most useful, are indicated by an asterisk.

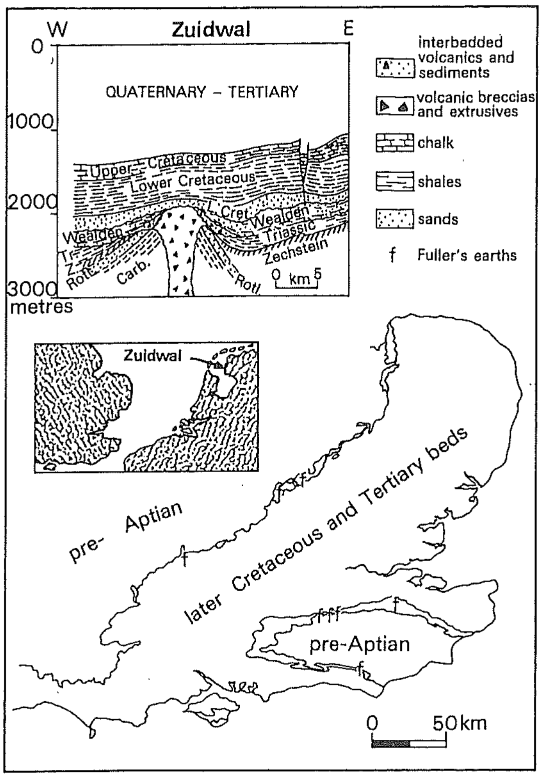


Figure 15.12 Distribution of fuller's earths in the lower Cretaceous of S. England, with insets showing the location of the Zuidwal contemporary volcanic vent and a section through it (after Jeans *et al.* 1977 and Cottençon *et al.* 1975).

recorded by the extensive development of (montmorillonitic) fuller's earths (Fig. 15.12). Hallam and Sellwood (1968) revived earlier speculations concerning a volcanic origin for these deposits, and this view has later been confirmed by Jeans *et al.* (1977) with the recognition of glass shards and well-preserved pyroclastic material of possible trachytic composition. The actual volcanic sites, however, remain a mystery. The Wolf Rock Phonolite of Cornwall (Fig. 15.8) represents the foundered wreck of an early Cretaceous volcano, but it may be too old (*c.* 131 Ma; J. G. Mitchell *et al.* 1975) to have supplied the material. However, other and as yet undiscovered sources might have lain west of Britain, and still others may await discovery on the London Platform itself.

The whole picture deduced above points to an intensification of tectonic, volcanic and transgressive processes during the early Cretaceous. During the Aptian the Rockall Trough, which separates the Rockall Bank from the rest of continental Europe, began to form (Fig. 15.8). This major event marked the initiation of sea floor spreading in the N. Atlantic and the eastward rotation of Iberia (D. G. Roberts 1975, Laughton 1975). Thus a causal link seems to have existed between the more-or-less synchronous events in areas to the east. Spreading in the Rockall Trough was a temporary affair and was succeeded by collapse both of the graben and its margins. One repercussion of the uplift and succeeding collapse phase was the unconformable overstep (Fig. 15.9) of progressively younger Cretaceous strata over the western regions (Hallam & Sellwood 1976*).

15f The Chalk sea

As the late Cretaceous sea encroached on to the craton, the sources for terrigenous clastic material were progressively eliminated. Pelagic deposits became dominant and gradually advanced westwards (Table 15.1). The bulk of this pelagic material is the debris from coccolithophorid planktonic algae, occurring mostly in separate micron-sized plates. However, some plates are still in their original rings, called coccoliths (Hancock 1975*).

Most of the Chalk was probably deposited as magnesium-low calcite, which is relatively stable. The rarity of early lithification may indicate that little magnesium-high calcite or aragonite was originally present (Hancock 1975*). In areas of low heat flow, and where deep burial has not occurred (i.e. most onshore outcrops), initial porosities have been preserved. However, under the thick Tertiary overburden of the North Sea, and beneath Tertiary basalts in N. Ireland, the Chalk has been cemented and porosities have greatly diminished.

Like the Gault Clay and Upper Greensand, the Chalk is a diachronous facies. Its base is normally marked by a condensed marly deposit with a variable quartz and glauconite content (frequently termed the 'basement bed'). It contains reworked phosphatised fossils in addition to an indigenous fauna. Above, the Chalk normally commences in rhythmically alternating chalky limestones and interbedded marls arranged in sequences similar to those described from the Lias (Fig. 14.9; Ch. 14d).

Their rich burrow assemblage (W. J. Kennedy 1967, 1970) proves the depositional origin of the rhythms. There is a progressive loss of detrital clays upwards, and later chalks are normally about 98 per cent $CaCO_3$ (Hancock 1975*).

Normal salinities in the Chalk sea are indicated by the presence of groups such as echinoderms, brachiopods and cephalopods. However, substrate conditions must have been rather soft and therefore inhospitable to many benthic taxa. Certain bivalves (e.g. species of *Inoceramus*, *Pycnodonte* and *Spondylus*) became specially adapted to the soft bottom conditions. Certain *Inoceramus* species had greatly inflated left valves which allowed them to 'float' on the substrate; species of other genera developed spines or buoyant cavernous shells. Encrusting epifaunal organisms that needed firm substrates sometimes took advantage of these 'pioneer' species, and encrustations a few centimetres high can be found upon large *Inoceramus* shells. In view of the intense bioturbation (W. J. Kennedy 1967, 1970; Kennedy & Garrison 1975), it is possible that the actual sea bed was in a thixotropic state, being permanently firm only at depths of a few tens of centimetres.

Over the submerged massifs (Fig. 15.13), chalk facies thin and more stratigraphic gaps are present in sequences. These gaps are frequently associated with signs of contemporary hardening by cementation of the chalky sea floor (hardground development). Encrustation, phosphatisation and glauconitisation are all features seen in these chalk hardground horizons (e.g. Chalk Rock). R. G. Bromley (1967) has shown how cementation a few centimetres below the sea floor caused crustaceans to modify their burrowing activities. Hardground chalks have higher magnesium values than normal white-chalk facies. This probably reflects early cementation by magnesium-high calcite remobilised from the shells of the richer benthos associated with the more stable substrates. There is, however, no evidence for exposure, and the early cementation seems to have taken place under wholly submerged marine conditions.

The European Chalk is characterised by the presence of irregularly shaped flints which mostly follow, but sometimes cut across, bedding planes. Flint consists of micron-sized, randomly arranged, quartz crystals with interstitial water-filled microcavities. Flint nodules frequently envelop fossils and burrows; and, although they did not form penecontemporaneously, since flint pebbles do not occur within the Chalk, they clearly formed before the earliest Tertiary. The silica was probably derived wholly from biogenic sources (e.g. sponge spicules), although some small contribution from distant volcanic sources is possible. Flints probably grew in several stages, with their formation being aided by the carbonate-rich diagenetic environment. The first phase was probably the mobilisation of opaline silica and the subsequent concentration of amorphous opal around sites of organic concentrations. The opal later transformed into low temperature cristobalite during solution and reprecipitation. In its turn cristabolite transformed into chalcedonic quartz (flint), although the precise requirements for this are still a matter of dispute (See Keene 1976 and Wise & Weaver 1974 for recent discussions on silica diagenesis).

In the clear water of the modern tropics, coccolithophorids live at depths of 50–200 m. Their tiny skeletons accumulate from the limits of wave effectiveness to depths in excess of 3 km (calcium-carbonate compensation depth). There is usually little evidence to suggest strong scouring and current action on the floor of the Chalk sea, and accumulation in deepish water is suggested. Fossils provide conflicting evidence for the depth of the Chalk sea. Certain foraminiferans appear to indicate shallow water (a few metres), but the presence of hexactinellid sponges seems to indicate depths of 80–100 m (Reid 1968). The maximum depth is difficult to ascertain, but the well-preserved coccoliths in outcropping Chalk suggest that little etching has occurred. Etching occurs today for coccoliths accumulating below about 1 km. Provided that the carbonate balance of the Cretaceous seas was not vastly different from that of the present, deposition between 100 m and 600 m seems likely (Hancock 1975*). The ancient massif areas were probably more shallowly submerged (Fig. 15.13).

Estimates of Chalk deposition rates vary because of the difficulty in assessing the amounts of both compaction and time. However, Table 15.2 provides a guide to sediment thicknesses for each stage in a number of places. Average figures for the white-chalk facies fall within the range of 20–40 m Ma^{-1}.

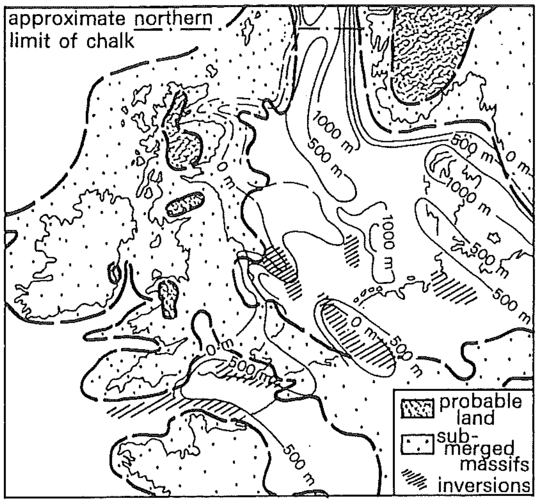


Figure 15.13 Cretaceous isopachs and the distribution of islands and submerged shoal areas in the Chalk sea in NW Europe. The Chalk facies passes into marine shales in the north. Positions of late Cretaceous inversion axes are also indicated (isopachs after Selley 1977, palaeogeography after Hancock 1975*).

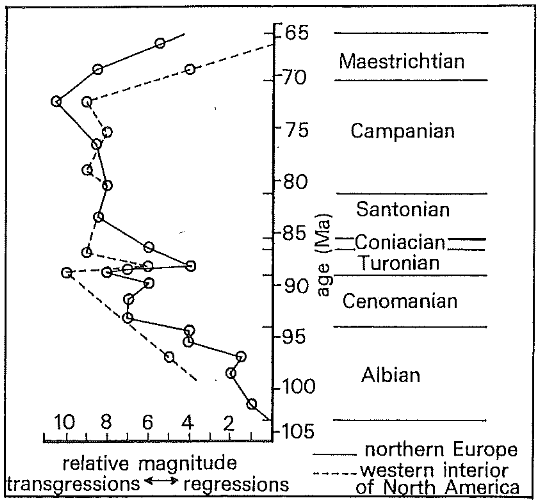
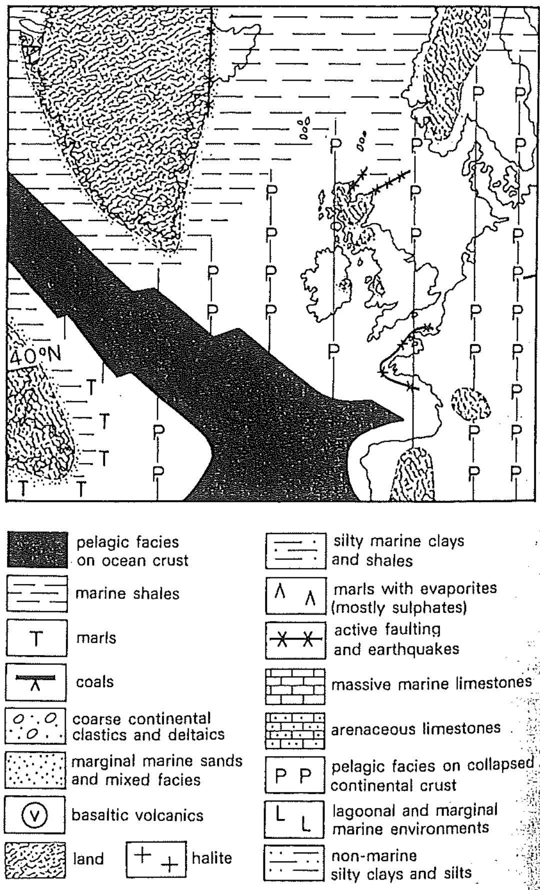


Figure 15.14 Age and relative magnitudes of transgressions and regressions in the late Cretaceous (after Hancock 1975*).

Figure 15.15 Generalised upper Cretaceous (Campanian) facies and palaeogeography for the British area.



The depth of the Chalk sea must also have been affected by eustatic sea-level changes. As in the Jurassic, eustatic fluctuations are interpreted from the evidence of synchronous transgressive and regressive events that correlate over vast areas (Fig. 15.14). There is no evidence for polar ice caps in the Cretaceous, and eustatic fluctuations were probably controlled by the periodic elevation and collapse of ridge systems during successively active and quiescent phases of central Atlantic sea-floor spreading which, on available evidence (Pitman & Talwani 1972), was already well under way in the late Cretaceous (Fig. 15.15).

In Europe, climatic conditions influenced both the rate of organic productivity within the sea and the nature of terrigenous sediment supplied during progressive inundation of the craton. Although sands were generated locally from submerging massifs, clay availability became severely restricted. Terrigenous plant debris is not found onshore in NW Europe after the Cenomanian, and phosphatic nodules in the Chalk are comparable with those from modern marine areas influenced by more arid climates (Hancock 1975*). Northwards, however, into the Viking Graben of the North Sea (Fig. 15.13), the Chalk facies is replaced by silty shales.

Table 15.2 Thickness in metres (and thickness per million years) of the Chalk by stages in Britain (after Hancock 1975).

Stage and Duration		*Isle of Wight*	*Portsdown*	*N. Norfolk*	*Yorkshire Coast*	*Southern North Sea*
Maestrichtian 6 Ma	upper	—	—	—	—	325·6 (54·3) (upper and lower)
	lower	—	—	26·4 (*c.* 14)	—	
Campanian 11·5 Ma	upper	144·8 (*c.* 50)	—	99·1 (17·2)	—	119·0 (20·7)
	lower	104·5 (18·2)	—	79·2 (13·8)	99·7 (*c.* 29)	62·3 (10·8)
Santonian 4 Ma		116·4 (29·1)	100·6 (25·2)	94·5 (23·6)	143·3 (35·8)	154·7 (38·7)
Coniacian 1 Ma		16·2 (16·2)	18·3 (18·3)	22·9 (22·9)	36·9 (36·9)	54·9 (54·9)
Turonian 2·5 Ma		63·7 (25·5)	82·3 (32·4)	56·4 (22·6)	105·5 (42·2)	135·4 (54·2)
Cenomanian 4·5 Ma		55·7 (12·4)	103·6 (23·0)	16·2 (3·6)	29·4 (6·3)	34·6 (7·7)
	Total	501·3	304·8	394·7	414·8	886·5

This may reflect a change to a more humid climate, with the bulk of the clastics being derived from Greenland (Hancock & Scholle 1975).

15g *'Dies irae'?* – requiem for the dinosaur

The lower Cretaceous fault block topographies in the North Sea and over the ancient Hercynian massifs were eliminated as general downwarping and draping took place during the upper Cretaceous. Differential subsidence continued, so that by the end of the period basinal regions had received up to 500 m of upper Cretaceous sediment in contrast to the 100–200 m that had accumulated over the submerged massifs. Some troughs within the North Sea (Fig. 15.13) received exceptional thicknesses (*c.* 1·5 km), which may relate to redistribution of Zechstein salts.

Late in the Cretaceous, some of the basins began to undergo uplift (inversion) in areas peripheral to the main North Sea Basin (e.g. the Weald, English Channel, etc.; Fig. 15.13). Secondary basins formed on the flanks of these newly inverted basins. The major phase of inversion took place from the Maestrichtian (Table 15.1) to the Palaeocene (early Tertiary) and correlates with the last phase of downfaulting in the central and northern parts of the Central North Sea Graben (P. A. Ziegler 1975*). These movements were accompanied by the emplacement of submarine mass flows containing semi-consolidated chalks (Selley 1976).

In S. (Alpine) Europe, the tensional regime that had typified most of the Mesozoic was replaced by a phase of compressive stresses attending the closure of the Tethys. Intense differential movements in the upper Cretaceous were accompanied by the emplacement of Alpine flysch sequences. This change of regime probably provided the driving forces behind the formation of inversion 'axes' on the cratonic foreland. These movements presaged the more general uplift that was to affect the area at the close of the Cretaceous, but these and the more intense Tertiary events will be considered later (Ch. 16).

In the seas, some ammonite families (e.g. the *Hoplitidae*) underwent diversification, and certain genera (e.g. *Turrilites*) adopted a nekto-benthic mode of life, developing asymmetrically coiled shells (Kennedy 1978). The real molluscan success story of the Cretaceous, however, is that of the rudistid bivalves, which constructed colossal build-ups with a morphology similar to that of modern coral reefs. They spread throughout the Tethyan belt, from the Pacific and Gulf of Mexico, through S. Europe, to the Middle East and Asia.

Modern bony fishes evolved during the Cretaceous. However, the large marine reptiles of the Jurassic (e.g. ichthyosaurs and pliosaurs) began to decline, and in the Cenomanian they were replaced by other carnivores such as the mosasaurs (a group of lizards) and long-necked plesiosaurs (Halstead 1975).

On the land, great changes occurred in the composition of the late Cretaceous flora, with angiosperms (modern flowering plants) achieving dominance over the other groups (particularly gymnosperms) by the end of the Cenomanian (Table 15.1). Placental mammals appeared in the early Cretaceous, although dinosaur reptiles dominated the land until the end of the period.

In the air, pterosaurs had greatly advanced and by the late Cretaceous were complex, lightly built and highly efficient flying machines. Some may even have been hair covered. These animals were accompanied by the birds, dinosaur derivatives which first appeared in the upper Jurassic *(Archaeopteryx)*. Their fossilisation potential is low, so that they have left a thin record. However, it is clear that during the Cretaceous many familiar modern forms, such as owls, cormorants and certain waders, had already evolved (Halstead 1975).

At the end of the Cretaceous a major phase of extinction affected many disparate groups. Of the molluscs, ammonites, inoceramids and rudistids died out. The ammonites commenced a steady decline in the Campanian, and there are few records of upper Maestrichtian forms. Belemnites might have just survived into the Tertiary but then soon expired. Other invertebrate groups suffered losses of genera and some of whole families (e.g. planktonic foraminiferans, bryozoans and echinoderms). Major extinctions also affected marine phytoplankton populations. The land flora had already suffered its greatest change earlier in the Cretaceous.

Of course, the greatest popular appeal surrounds the extinction of the dinosaurs, but their demise can only be viewed in the context of the whole wave of end-Cretaceous extinctions. Catastrophic explanations for these extinctions are just as unfounded as they were for the extinctions at the end of the Permian (Ch. 13d). Instead, causes must be sought in the total of Earth disruptions that took place at the end of the Mesozoic era (65 Ma ago). The long phase of global environmental stability was rudely interrupted by a plexus of change. This produced repercussions throughout the world's food chains. The continents continued a phase of rapid migration and dispersal (Smith & Briden 1977*) which must have totally altered the global climatic regimes. Regression on a worldwide scale at the end of the Mesozoic severely restricted the extent of shelf seas and increased the competition for ecological niches on the narrow continent-dominated shelves that remained. The net result was the extinction of the least adaptable groups at virtually every trophic level. Early Tertiary niches were more rigorous and became filled by the descendants of genera 'delivered' by their own adaptability from end-Cretaceous oblivion.

became focused on the newly created Reykjanes–Iceland–Jan Mayen ridge as the N. Atlantic evolved from its rifting into its spreading phase (Bell 1976*).

Events deduced for the Tertiary rifting phase compare in a general way with those of earlier rifting phases (Fig. 14.5). First there was an initial phase of doming, succeeded by a rifting phase accompanied by volcanism. Rifting ended in the early Palaeocene with the emplacement of oceanic crust. Finally, ocean floor spreading was accompanied by a protracted phase of subsidence which affected the new continental margins and continues to the present time.

In E. Greenland, Kangerdlugssuaq Fjord has been interpreted as the northwestern arm of a Y-shaped rift system in which the other rift elements run parallel to the present coast. A major dyke swarm parallels the coast and may represent the onshore remnants of oceanic crustal dykes emplaced during the rifting and early spreading phase. In Britain, however, there is no monoclinal flexure, and the regional dyke swarms run NW–SE. The newly formed continental margin runs roughly at right angles to the NE–SW-trending ridge-and-basin structure in the Hebridean area (Hallam 1972). One explanation for this geometrical discrepancy is that when separation was finally achieved between Greenland and Eurasia, opening was swifter in the south than in the north (Bell 1976*). The result was a skewing of the two plates, which produced a system of tensional fissures that were roughly perpendicular to the rift structure, into which the dykes were injected. However, the NW–SE trend of dyke swarms in Britain is roughly parallel to the axis of the Labrador Sea and may reflect the Tertiary invasion of tensional lines inherited from the Cretaceous opening of this oceanic area. Clearly, further work is needed to integrate the structural and petrological aspects of this story.

16d Tertiary environments in the British area

Much of the British Isles became land during the earliest Tertiary, and the almost ubiquitous blanket of Chalk began to be eroded. Remnant outliers exist in Ireland and W. Britain. At the opening of the Palaeocene (Table 16.1) S. Britain lay at about 40°N and with the rest of cratonic Europe was steadily drifting northwards. In the late Miocene (10 Ma ago) the area lay at nearly 50°N (Smith & Briden 1977*).

Britain probably began to assume its present tectonic style early in the Tertiary, with the uplift of areas in the north and west and the subsidence of areas in the southeast. The western areas provided terrigenous clastic sediments to both the subsiding eastern regions and to fault-controlled basins that developed in the west (e.g. Bovey Tracey, Petrockstow, Celtic Sea, Bristol Channel, Cardigan Bay, Lough Neagh, Sea of the Hebrides). Deposition on the southeastern fringes of early Tertiary Britain records the interplay between marine transgressions from the North Sea and the subsequent progradation of terrigenous coastal-plain sequences into the newly formed marine area. Repeated transgression and regression resulted in the cyclic Tertiary sequences of the London–Hampshire (and Paris) Basins (Fig. 16.7).

In Britain the earliest transgressive deposits are of Palaeocene age and are confined to the area of E. Kent. However, later (Eocene) beds overstep earlier ones, and at its maximum the sea seems to have penetrated beyond W. Berkshire (Fig. 16.8). It is tempting to relate these transgressive–regressive episodes to phases of British tilting, but the possibility of Tertiary eustatic changes cannot be discounted (Vail 1977).

In each cycle the transgressive beds tend to be bioturbated glauconitic sands, containing flint pebbles derived from the Chalk and diverse marine faunas dominated by bivalves and gastropods. More euryhaline environments were inhabited by oysters and, where marine beds directly overlie the Chalk, the contact is sometimes infested with borings. Early Eocene transgressive deposits are mostly thin (Fig. 16.7), but they give way upwards to thicker regressive sands and red-mottled kaolinitic clays. The red-beds probably represent reworked lateritic clays (Montford 1970) deposited in lagoons and marshes. In the west, sandy sequences such as the Reading Beds, with strongly unidirectional cross-bedding, represent alluvial deposits derived from western source areas.

The most widespread marine advance during the Eocene extended marine conditions from the North Sea over the whole of E. Britain from East Anglia to

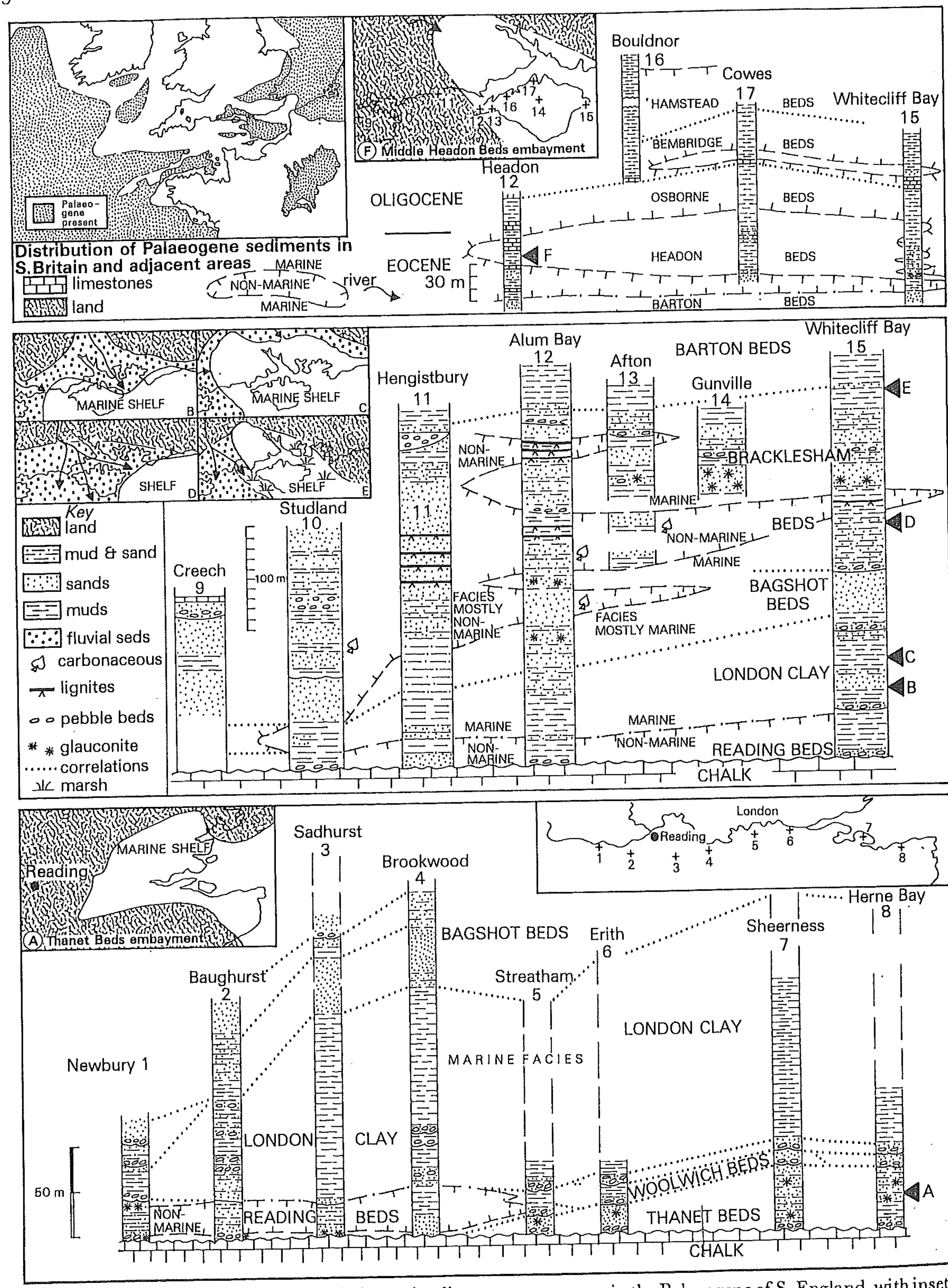


Figure 16.7 Stratigraphic correlation, facies and sedimentary sequences in the Palaeogene of S. England, with insets to show the palaeogeographic situation at selected times (compiled from Curry 1965 and 1966 and Murray & Wright 1974).

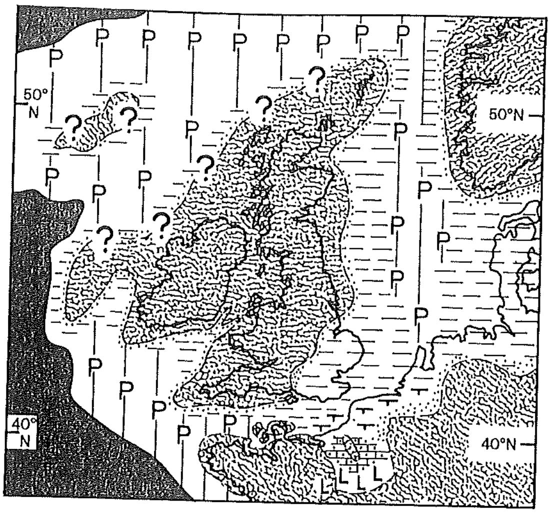


Figure 16.8 Generalised palaeogeography of the British area during the Eocene. For key see Figure 15.15.

Dorset and southwards over the Low Countries, N. France and NW Germany (Fig. 16.8). The western English Channel was also transgressed at this time, but from the direction of the Atlantic. The effect of the transgression was to spread a typical North Sea-type shale facies, the London Clay, over E. Britain. The London Clay is a fairly monotonous and sometimes intensely bioturbated mudstone containing an abundant and diverse fauna of molluscs which, although dominated by bivalves and gastropods, also includes nautiloids. Crustaceans, fish and reptiles (crocodiles and turtles) represent the more exotic macrofauna, the microfauna includes essentially pelagic forms, such as diatoms, radiolaria and globigerine foraminiferans.

Although many of the early Tertiary deposits contain plant debris, the London Clay has a wealth of terrigenous vegetable material (>500 species). This includes pollen, spores, logs and fruits, all of which provide evidence for the early Tertiary climate (Ch. 16e).

The London Clay grades upwards into sandier beds (Fig. 16.7), with flaser bedding and cross-bedding indicating currents of variable direction and magnitude. These transitional sandy beds frequently exhibit complex crustacean burrows *(Ophiomorpha)* and are comparable with sediments from some modern tide-dominated shoreface environments. As regression continued, shoreface sands gave way, diachronously, to non-marine sands showing strongly unidirectional cross-bedding structures that indicate derivation from the west. On the Isle of Wight, marine beds in the east are seen to pass into non-marine facies in the west. Further west, in Dorset, a series of sands and flinty gravels (the Barton Beds of Fig. 16.7) represents the deposits of braided streams which drained the Cornubian and western uplands. In Devon, gravels capping the Haldon Hills (near Exeter) comprise the western outpost of these gravels. Kaolinitic clays were derived from intense weathering and were deposited within the alluvial sequences from which they are now extracted as ball-clays.

In the later Eocene, marine transgression allowed the British debut of the nummulites. These large benthic foraminiferans were more abundant in S. Europe where they contributed to substantial carbonate build-ups (Heckel 1975). In the Bracklesham Beds of the Isle of Wight they occur as monospecific associations within thin current-winnowed laminations. They were probably near the northern limits of their range, and a more normal Tertiary fauna of marine molluscs occurs in adjacent beds.

The competitive interplay between marine and non-marine environments continued until the end of Eocene times. Latest Eocene and early Oligocene sediments are only preserved onshore in Britain in the Isle of Wight–Hampshire area where, at the end of the Eocene, marine beds were replaced by units containing the freshwater molluscs *Viviparus* and *Unio*. However, in the succeeding Oligocene the freshwater ponds inhabited by these molluscs were invaded by the returning eastern sea. Thus the Oligocene sequence of the Isle of Wight illustrates the continuation of transgressive–regressive cycles. Some beds are packed with freshwater snails *(Galba* and *Planorbis)*; others contain terrestrial faunas that include mammals; still others contain barnacles, serpulids and oysters. But throughout there is a persisting theme of decreasing marine influence westwards.

Devon, Cornwall and the western lands had clearly supplied a large amount of sediment to the Hampshire area during the early Tertiary. Denudation of these western areas must have been accompanied by phases of uplift, otherwise the supply of sediment would have progressively diminished. Clearly, some tectonic control would have promoted a periodic rejuvenation of the sediment source areas. Evidence of contemporary tectonism in Devon is provided by the existence of great thicknesses of Oligocene fluvial and lacustrine sediments in the fault-bounded basins of Petrockstow (*c.* 660 m) and Bovey Tracey (*c.* 1200 m). Both of these basins, and further basins in the Bristol Channel, lie along the Sticklepath–Lustleigh Fault zone (Fig. 16.6). This NW–SE-trending dextral wrench fault was initiated, along with a number of others (Fig. 16.6), during Hercynian movements, but it became reactivated in the Oligocene when a cumulative dextral movement of about 30 km took place in SW England. The Bovey Tracey and Petrockstow Basins developed as sagponds that are comparable, in a small way, with those occurring along the line of the modern San Andreas Fault (Fig. 16.6).

In the Bovey Tracey Basin the Tertiary deposits rest upon a preserved outlier of glauconitic Cretaceous sands. Marginal and basal deposits consist of cross-bedded fluvial gravels rich in reworked Cretaceous cherts. Later beds comprise sandy clays, lignites and high-quality kaolinitic ball-clays. The kaolinite results from the deep weathering of the surrounding terrain, which included exposed granite on Dartmoor, Carboniferous Culm, Devonian shales and limestones, and Cretaceous Greensand. The lignites are largely composed of fragments of *Sequoia* accompanied by *Nyssa* (a 'gum') and *Cinnamomum* (cinnamon); some probably represent foundered masses of originally floating vegetation. Animal remains are very scarce, being limited to freshwater molluscs; the absence of a rich fauna is somewhat enigmatic.

In continental Europe, particularly in Belgium and the Paris Basin, early Tertiary beds are represented by terrigenous clastic deposits including sands, clays and thick lignites. Sequences are comparable with those of S. England, and periodic transgressions and regressions occurred. However, a major difference in the Paris Basin is the greater frequency of limestones and of evaporites in the Oligocene (e.g. *Marnes supragypseuses* Formation). These evaporitic beds may reflect both climatic change and the more enclosed nature of the embayment towards the south (Fig. 16.8).

16e The Palaeogene climate of the British area

Early Tertiary deposits of the British area, such as the London Clay, contain a wealth of fossil plants whose descendants survive at the present time. Snags arise, however, when simple floral comparisons are made between the Tertiary and the Recent. About 47 per cent of the flora is typical of modern lowland tropical regions; however, the remainder includes taxa capable of living in both lowland and extratropical conditions, and of these about 11 per cent represent purely extratropical genera (Daley 1972*).

Paralic sequences and nearshore muds of the early Tertiary are reputed to contain the fruits and pollen of mangroves (Montford 1970), and the fruits of the palm *Nipa* have long been recorded. Such plants clearly compare with those of modern tropical mudflats and at least reflect a frost-free climate. In the past it has been popular to interpret the 'extratropical' genera as the remains of an upland flora that had been transported into the offshore basins. These remains include *Metasequoia*, *Magnolia*, *Engelhardtia* (a sort of walnut), *Betula* (birch) and *Cedrus* (cedars). Daley (1972*) has suggested that this floral anomaly can be explained if the climate was variable but mostly warm, being seasonal (but frostless) and very humid in the low-lying regions. The extratropical genera would have lived away from major waterbodies under less humid conditions, but not necessarily at high altitudes.

Support for a warm climate is derived from the presence of the abundant kaolinitic (ball-clay) sediments of SW England, as kaolinite forms most readily under humid tropical conditions. Even in the northern volcanic districts red lateritic soils developed between extrusive episodes, and, as noted above, lotus grew in the caldera lakes of Skye.

Later, in the Oligocene, humidities decreased; gypsiferous evaporites began to form in the Paris Basin to the south, while on the Isle of Wight calcrete soils formed, recording a decrease in rainfall. In the succeeding Miocene (Table 16.1), climates in S. Europe became very arid, and during this period (Messinian Stage) the last remnants of the Tethys (the Mediterranean) underwent repeated phases of desiccation (p. 256).

level led to minor transgressions during warm phases and regressions during cold phases, although these did not significantly alter the position of the coastline of the British Isles (Ch. 16g).

The preservation of sediments deposited during glacial and interglacial cycles is more complex. During each ice advance, tills were deposited in a range of environments from the highlands to what is now the continental shelf. Fluvial and coastal erosion during the intervening periods, together with glacial erosion during the following ice advances, tended to remove these deposits from the present land areas and the shallower parts of the shelf. Clearly, although the most continuous sequence should be found in the deeper water areas beyond the shelf edge, the most complete record on the continental shelf will be found in the areas of maximum subsidence, i.e. in the central parts of the sedimentary basins. Here, tills deposited from grounded or floating ice will tend to be interbedded with marine sediments, as in the case of the Precambrian Port Askaig Tillite (Ch. 4c), and terrestrial deposits will be less common.

Within the sedimentary basins a fairly complete stratigraphic sequence should be preserved from which the alternation of cold and temperate stages, the climatic record, can be read. The uppermost part of this record can be readily calibrated, as radiocarbon dating is practical back to about 50 000 years BP. However, for the older deposits the usual techniques are inapplicable. K–Ar radiometric dating is restricted to volcanic rocks, and palaeontological zoning is far too crude. The most promising method, the one that has proved successful in dating deep sea cores, is palaeomagnetic dating. It should be noted that this can only be applied to continuous vertical sections or borehole cores, not to isolated deposits.

The climatic curve in Figure 17.1b, constructed from pollen analytical data and calibrated by radiocarbon and palaeomagnetic dating, is for the Quaternary of the Netherlands, an area within the North Sea sedimentary basin. If the Dutch Quaternary stratigraphic column could be used as a standard sequence for NW Europe, it should be possible to correlate the isolated deposits around the North Sea Basin by comparing their climatically controlled fossil assemblages with the standard basinal sequence, provided (a) that the standard sequence is a complete sedimentary record (i.e. it includes all the cold and temperate stages), (b) that it is possible to distinguish palaeontologically between each warm stage in the standard sequence, and (c) that the standard sequence and deposits of unknown age are geographically sufficiently close and similar in their depositional environment to have experienced a similar faunal and floral history. However, as the palaeontological differences between climatic stages are fairly subtle, this requires a very detailed palaeontological knowledge of both the standard and unknown sequences, and problems remain in applying such correlations even during the younger interglacials within NW Europe.

Outside the sedimentary basins, only the youngest deposits, i.e. those amenable to radiocarbon dating, can be dated absolutely. In the absence of a standard basinal sequence it is usual to erect a composite local sequence, with stages based on different type localities. This composite sequence is unlikely to be complete; and as it is based on an interpretation of the stratigraphy, which is itself based on assumptions about the climatic history, it may be of doubtful validity.

The climatic stages of the preglacial part of the British Quaternary scheme (Fig. 17.2) are defined from a borehole (Ludhamian to Antian) and coastal sections (Baventian to Cromerian) in East Anglia. There can be little doubt that they are in the right stratigraphic order but, as they are based on deposits from the edge of the North Sea sedimentary basin, there are likely to be major gaps in the record (Ch. 16g). The glacial Quaternary stages are based on type localities in East Anglia and the Midlands. The validity of these stages cannot yet be checked against a standard basinal sequence, and there is controversy regarding the correlation of events with these stages in the British Isles (Ch. 17d). However, there is little doubt that the last glacial stage, the Devensian, correlates with the last glaciation in many other parts of the world, such as the Weichselian in N. Europe, the Würm in the Alps and the Wisconsinian in N. America. The last interglacial, the Ipswichian, can also be correlated with the Eemian of the European scheme.

17d The pre-Devensian record

On land in the British Isles there is evidence for four

	Dates in years B.P.	Stage	Climate	Examples of typical deposits
QUATERNARY: Recent or Holocene		Flandrian	temperate	silts & peats in Lincolnshire, East Anglia & Lancashire; estuarine deposits – Thames etc.
QUATERNARY: PLEISTOCENE	– 10 000	Devensian	cold, glacial at end of stage	head in S & SW England; Skipsea Till, Irish Sea Till; Four Ashes Gravel; Dimlington Silts.
	– 115 000	Ipswichian	temperate	organic deposits near Ipswich; beach deposits at Sewerby & in Devon
	– 128 000	Wolstonian	cold, in part glacial	Welton Till Fremington Till
		Hoxnian	temperate	organic deposits at Hoxne, Suffolk; Marks Tey, Essex
		Anglian	periglacial and glacial	Lowestoft Till, North Sea Drift, Corton Sands Barham Sands and Gravels; Barham Arctic Structure Soil
		Cromerian	temperate	Valley Farm Rubifield Rubified Sol Lessivé; peats & freshwater sediments – Norfolk
		Beestonian	periglacial and glacial	Kesgrave Sands & Gravels
	~ 600 000	Pastonian	temperate	Westleton Beds Icenian Crag including { Weybourne Crag, Chillesford Clay, Norwich Crag }
	hiatus ~ 1 600 000	Baventian	cold	marine silts & clay – Ludham & Easton Bavents
		Antian	temperate	shelly sand } marine sediments in Ludham borehole and Suffolk
		Thurnian	cold	silt } marine sediments in Ludham borehole and Suffolk
	~ 2 050 000	Ludhamian	temperate	shelly sand } marine sediments in Ludham borehole and Suffolk
	hiatus ~ 2 450 000	'pre-Ludhamian'	temperate	Red Crag
PLIOCENE	– ~ 2 500 000		warm temperate	Coralline Crag

Figure 17.2 Quaternary stages, dates, climatic conditions and some typical deposits (see G. F. Mitchell *et al.* 1973 for more details).

glaciations (Fig. 17.2). The ice limits of the last glaciation, which took place towards the end of the Devensian cold stage (between 26000 and 10000 years BP), and of the glaciation of maximum extent are shown in Figure 17.3 (D. Q. Bowen 1977*). Within the Devensian ice limit, Devensian till blankets most of the lowlands and large parts of the highlands too. This till usually lies directly on bedrock, and older Quaternary deposits are only locally preserved beneath the Devensian till, as at Dimlington Cliff on the Yorkshire coast (Fig. 17.5) and Shortalstown, Co Wexford. In the S. Irish Sea pre-Devensian till is preserved beneath Devensian and Ipswichian deposits, where ice partly scoured out areas of soft Tertiary rocks. This is unusual, as within the Devensian ice limit on the continental shelf Devensian deposits unsually directly overlie the Tertiary or lowermost Pleistocene. Clearly, Devensian glacial erosion has been important in removing the record of earlier glaciations in these areas.

Outside the Devensian glacial limit, pre-Devensian glacial and interglacial deposits are common, although they have suffered more erosion than the Devensian deposits in the lowland areas to the north and are most extensive where they have escaped erosion on the interfluve areas of East Anglia and the Midlands.

The earliest evidence for glaciation in the Quaternary sequence in East Anglia is found in the sediments of the Beestonian Stage, which overlie the Pastonian marine, estuarine and beach deposits (Ch. 16g). The Beestonian includes a degraded till

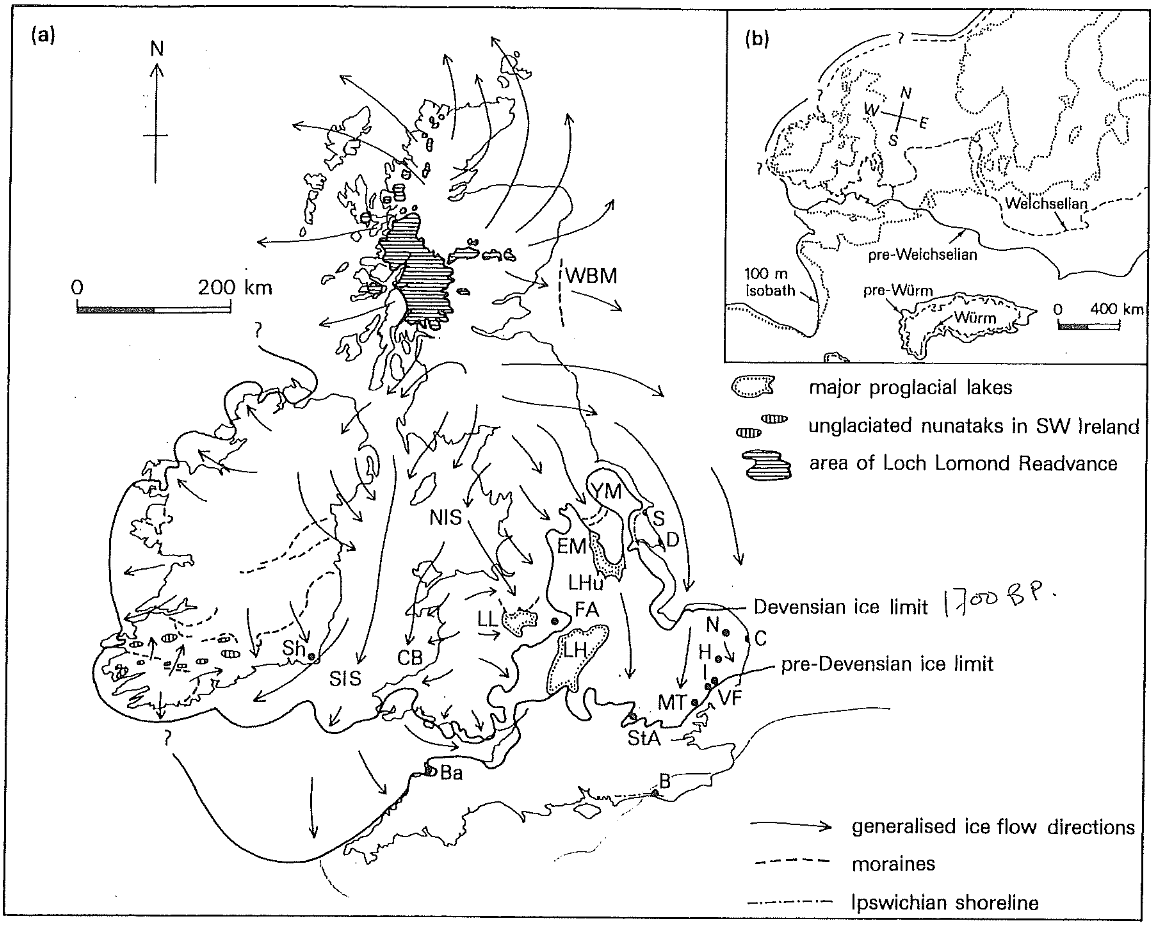


Figure 17.3 (a) Ice limits and ice flow directions in the British Isles (after D. Q. Bowen 1977*). (b) Ice limits for Europe (after Flint 1971).
Localities: B. = Brighton; Ba. = Barnstaple; C. = Corton; CB = Cardigan Bay; D. = Dimlington; FA = Four Ashes; H. = Hoxne; I. = Ipswich; LH = Lake Harrison; LHu = Lake Humber; LL = Lake Lapworth; MT = Marks Tey; N. = Norwich; NIS = N. Irish Sea; S. = Sewerby; Sh. = Shortalstown; SIS = S. Irish Sea; StA = St Albans; VF = Valley Farm.
Moraines: EM = Escrick; WBM = Wee Bankie; YM = York.
For sections N–S and E–W in (b), see Figure 17.6.

remnant in the SW Midlands and extensive fluvial sands and gravels (Kesgrave Sands and Gravels) in East Anglia and Essex (Fig. 17.4b). Abundant ice-wedge casts and involutions indicate that the latter were deposited under periglacial conditions, and palaeocurrent measurements show that, in Essex and S. Suffolk, they were deposited from a major NE-flowing river north of the course of the present River Thames (Rose & Allen 1977*).

During the following Cromerian temperate stage, estuarine silts and freshwater muds and peats accumulated. These are now exposed on the Norfolk coast. Further south, the Cromerian is represented by the Valley Farm Rubified Sol Lessivé, a palaeosol formed under humid warm temperate conditions. Evidence for cold conditions follows with, in Norfolk, the deposition of freshwater and estuarine sediments, cut by ice wedge casts and containing an arctic fauna and flora, and, in Suffolk and Essex, the formation of an arctic palaeosol (the Barham Arctic Structure Soil), loess and blown sand deposits. These features are attributed to a periglacial episode that preceeded the advance of the glaciers that deposited the overlying Lowestoft Till, and they are included with this till as deposits of the Anglian cold stage.

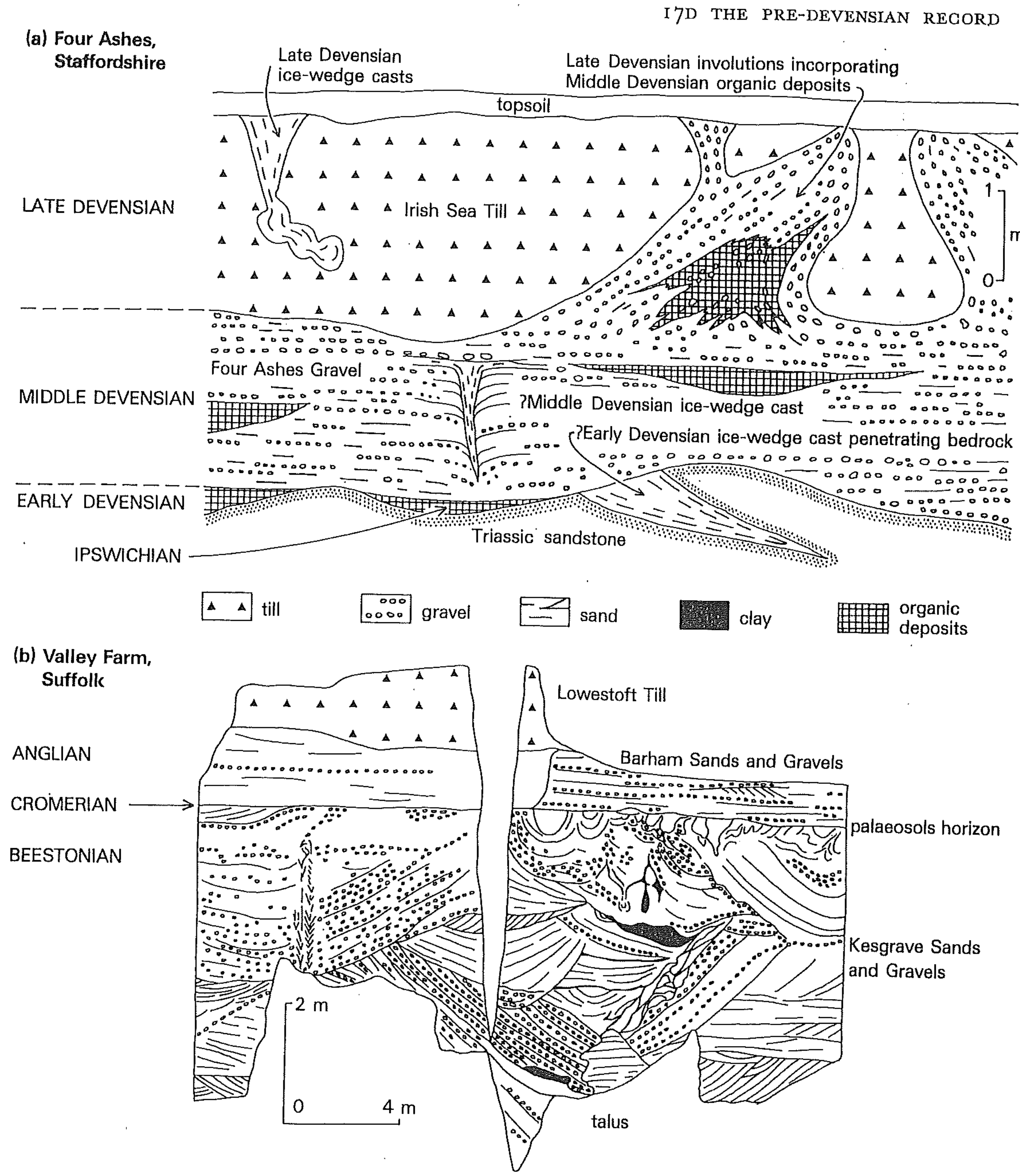


Figure 17.4 Examples of Quaternary stratigraphic sequences. (a) Four Ashes pit, Staffordshire, the type locality for the Devensian: schematic diagram showing the alternation of interglacial and interstadial organic deposits with glacial and periglacial sediments and structures (after Morgan 1973). (b) Valley Farm pit, Suffolk: sketch of north face; the palaeosols horizon includes the Anglian Barham Arctic Structure Soil and the Cromerian Valley Farm Rubified Sol Lessivé which it deforms (after Rose & Allen 1977*).

The Lowestoft Till occupies an extensive area in East Anglia, the E. Midlands and S. Lincolnshire. The bulk of the clasts are chalk, the matrix is dominated by clay, and Perrin *et al.* (1973) have suggested that the North Sea is a likely source area. In NE Norfolk there are up to three tills containing Scandinavian erratics interbedded with sands, which are collectively known as the North Sea Drift. Fabric analysis suggests that these tills were deposited from ice that flowed from the northwest, north and northeast; and although the Lowestoft Till overlies the North Sea Drift south of Corton, all these deposits are thought to be of broadly the same age, i.e. Anglian, but to have been deposited from different ice lobes between which fluvial, lacustrine and estuarine sediments accumulated. Along the N. Norfolk cliffs the North Sea Drift exhibits spectacular deformation structures, including great thrust planes, domes, basins, overfolds and chalk rafts up to 10 m thick. With the exception of the chalk rafts, these are probably load and diapiric structures resulting from the deposition of the overlying outwash sands and gravels (Banham 1975).

A. The major river that flowed northeastwards through Essex and Suffolk during the Beestonian started to migrate south and downcut during this stage. By the Anglian it occupied the position of the present Thames drainage system. Indeed, the major features of the present drainage system of SE England date from the Anglian. Anglian ice produced only a minor southward diversion of the Thames in the Vale of St Albans area. Several subsequent episodes of aggradation and incision have left behind a series of river terraces in the Thames Basin.

All the pre-Devensian tills in East Anglia can be assigned to the Anglian stage. They are overlain by Hoxnian biogenic interglacial deposits. In the Midlands, Hoxnian deposits underlie a pre-Devensian till defined as being of Wolstonian age (Fig. 17.2), although Bristow and Cox (1973) have suggested that it is equivalent to the Lowestoft Till. Thus the deposits of the penultimate glaciation in East Anglia and the Midlands belong to different stages, the Anglian and Wolstonian respectively. It seems either that Wolstonian ice never reached East Anglia, or that the correlation of the Hoxnian deposits in East Anglia with those of the Midlands is invalid, the Anglian and Wolstonian tills both being deposited during the same glacial stage. In view of the fundamental difference between these interpretations it is difficult to produce a synthesis of Anglian–Wolstonian history in Britain, as the interpretation of events in other areas depends on our understanding of the type areas of East Anglia and the Midlands.

In the Midlands, Wolstonian tills with erratics derived from Wales, N. England and S. Scotland overlie Hoxnian interglacial lake deposits and are in turn overlain by Ipswichian fluvial terrace gravels. Tills thought to predate the Wolstonian (i.e. of Anglian age) are of restricted occurrence. As the Wolstonian ice advanced it blocked the Severn Valley, forming a lake (named Lake Harrison) which stretched from Stratford-upon-Avon to Leicester. As the ice advanced further it over-rode the lake, but small lakes might have formed later in the same area when the ice ablated. Outwash from the Wolstonian ice poured southwards into the Thames and other drainage basins, depositing northern erratics. During the following interglacial the valley fills were partly eroded, leaving behind river terraces. Outwash sands and gravels, subsequently terraced, were also deposited in valleys to the north as the ice ablated. B

Further west, Wolstonian ice just impinged on the northern coast of Devon and Cornwall, depositing a shelly till, the Fremington Clay, and glaciofluvial deposits around Barnstable. Giant erratics of granite and gneiss, weighing up to 50 tonnes, are also found overlying wave-cut rock platforms along this coast. These may be the remains of a Wolstonian or older till long since washed away, although, as similar giant erratics are also found along the English Channel and S. Irish coasts, it has been suggested that they were deposited from icebergs.

In NW Europe three glaciations are recognised: the Weichselian, Saalian and Elsterian, in order of increasing age. Saalian ice extended further south that that of the other two glaciations in the Netherlands, although Elsterian ice was the most far reaching in Germany and Poland (Fig. 17.3). The Quaternary succession of the Netherlands can be traced westwards into the southern North Sea. Here, during the Elsterian glaciation, tills were deposited in the north, bordered to the south by fluvial and lacustrine sediments. The damming of the North Sea during each glaciation impounded a

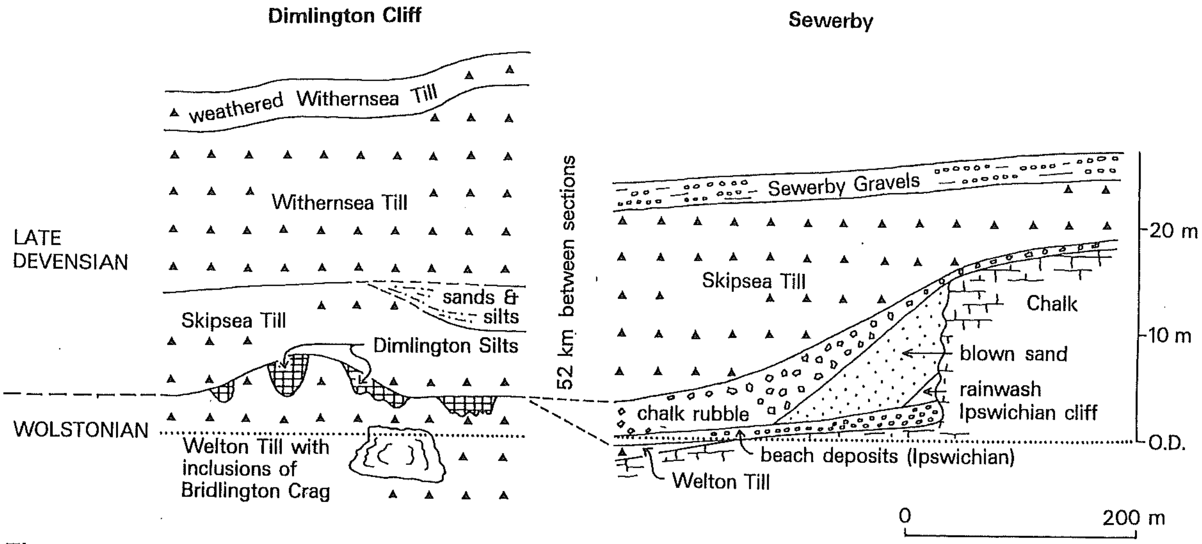


Figure 17.5 Stratigraphic relationships of the Quaternary deposits on the Yorkshire coast (see Figure 17.3 for localities). Heights are in metres above OD (Ordnance Datum). The late Devensian, but preglacial, Dimlington Silts are dated at *c.* 18 400 years BP; peat in kettleholes overlying the Devensian tills is dated at *c.* 13 000 years BP. The rainwash, blown-sand and chalk-rubble solifluction deposits date from the cold early Devensian stage, following the withdrawal of the Ipswichian sea. The Bridlington Crag is a Hoxnian or Wolstonian shelly marine clay eroded from an offshore area by the advancing Wolstonian ice (after Catt & Penny 1966, with terminology after Madgett & Catt, in press).

lake, fed by both outwash streams and glacially diverted preglacial rivers, which eventually drained southwestwards along the bed of the English Channel. During the succeeding Holsteinian interglacial the sea transgressed southwards into the North Sea, and marine sediments were deposited as far south as a coastline which ran E–W at about the latitude of Norfolk. The glaciers of the following maximum glaciation (the penultimate or Saalian of the European scheme) scoured the floor of the southern North Sea, deposited tills, and deformed the underlying interglacial deposits into a series of ice-pushed ridges which are also found in Holland. The marine transgression of the last interglacial was more far reaching than its predecessor. It crossed the present coastline in Yorkshire and Lincolnshire, where a buried Ipswichian cliff line cut in the Chalk along the eastern side of the Wolds is preserved beneath Devensian till (Fig. 17.5), and around the southern coast of England, where wave-cut Ipswichian platforms are found, for example, at Brighton. The continent was clearly cut off from Britain at this time.

Pre-Devensian tills are widespread in Ireland, where all but a few nunataks in the southwest were ice covered during the glacial maximum. The ice was mostly of local derivation, except along the eastern and southern coasts where Scottish ice locally pushed ashore. The paucity of interglacial sites in Ireland makes it difficult to date pre-Devensian events; and although most of the penultimate glacial tills are often assumed to be Wolstonian, there is little firm evidence for this at present.

South of the limit of the maximum glaciation, cold periods were marked by the formation of extensive solifluction deposits known as 'head' or, on the Chalk, as 'coombe rock'. Such periglacial features as involutions, stone stripes and polygons are also found.

17e The story of the last glaciation: the Devensian

In contrast to the confusion surrounding pre-Devensian events, there is a fair consensus of opinion about the detailed history of the last glaciation. This is because Devensian deposits are fresh and extensive, have suffered little erosion and are largely within the compass of radiocarbon dating. A detailed understanding of Devensian events may therefore help in elucidating the more obscure pre-Devensian record.

Climatic curves based on N. Atlantic deep-sea core data show a cooling *c.* 110000 years BP (about the Ipswichian/Devensian boundary) and a further marked cooling 73000 years BP. After this, the climate oscillated, reaching its most severe deterioration during a short cold phase between 30000 and 11000 years BP (Fig. 17.1e). Peat deposits in kettle-holes on top of Devensian tills in Scotland give radiocarbon dates ranging up to 13000 years BP, while organic remains underneath such tills give dates as young as *c.* 27000 years BP. Thus it appears that the advance of the Devensian ice was a short event right at the end of the Devensian, lasting *c.* 10000 years, culminating at *c.* 17000 years BP, and preceded by a long, usually cold but non-glacial phase.

During the Ipswichian interglacial, which preceded the Devensian, pollen and beetle data imply that summer temperatures were several degrees warmer and winter temperatures were milder than those of today, giving rise to a woodland vegetation. Sea level would have been slightly higher than that of today. The Ipswichian cliffline preserved in Yorkshire lies at just about the present sea level (Fig. 17.5), although the higher-level raised wave-cut platforms along the southern coast of England must have suffered subsequent tectonic uplift. As temperatures fell at the beginning of the Devensian, sea level would have fallen slightly as glaciers expanded at higher latitudes (Fig. 17.7), and the woodland would have been replaced by a sparse tundra vegetation. During the early Devensian the British Isles appear to have been very cold but, probably because of a relative lack of precipitation, the lowlands remained ice free. Strong winds would have cut across this inhospitable cold desert landscape. The abrasive action of blown sand formed ventifacts and undercut rocks. The climate was also conducive to the widespread formation of periglacial features, such as cryoturbations, ice-wedge polygons and solifluction deposits.

The study of fossiliferous horizons within fluvial terrace deposits in the Midlands and S. England (Fig. 17.4a) shows that the climate ameliorated briefly twice, resulting in the Chelford interstadial *c.* 60000 years BP and the Upton Warren interstadial *c.* 43000 years BP (Fig. 17.1f). During the former, temperatures warmed to within a few degrees of those of today and a pine and spruce forest vegetation returned. However, the latter warm interval was so short that, although summer temperatures briefly exceeded those of today, the climate deteriorated again before forest had spread northwards into Britain (Coope 1975*). Sea level might have continued to fall throughout this period, with the major rivers flowing along roughly their present courses and cutting across the shallow parts of the present continental shelf. However, the falling sea level did not expose the northern North Sea, where up to 300 m of marine clays with a shallow cold-water fauna, the Aberdeen Ground Beds, were deposited (Fig. 17.6). These overlie an irregular surface, possibly cut by Wolstonian glacial action, which truncates the lower Pleistocene Basal Beds. Dropstones in the Aberdeen Ground Beds indicate the presence of icebergs, which may have calved from the Scandinavian glaciers known to have existed at this time. Lignitised wood samples within the beds may date from the Upton Warren interstadial (Eden *et al.* 1977).

Some time after 30000, and possibly as late as 25000 years BP, the British climate became less continental and more maritime, allowing a sufficient increase in precipitation for glaciers to move out of the mountains and spread southwards. Flow directions deduced from the distribution of erratics and the orientation of glacial landforms are shown in Figure 17.3. Deep-sea core studies show that the glaciation reached its peak globally *c.* 17000 years BP, and it is likely that glaciers extended to the Devensian ice limit by this time. Ice thicknesses probably exceeded 2 km in parts of Scotland during the glacial maximum. As with earlier glaciations, the advancing ice impounded proglacial lakes where it cut off the lower reaches of valley systems. For example, the mouth of the Humber was blocked, forming a large lake (Lake Humber) in the Trent lowlands. Gaunt (1976) has suggested that the leading edge of a glacier extending down the Vale of York may have partly floated in this lake, the maximum extent of the ice being indicated by a belt of glaciolacustrine sands and gravels.

The Devensian ice limit is only rarely indicated by good terminal moraines. Usually the Devensian till has a feather edge and the limit is not easy to find. Within the limit the lithology of the tills varies from gravelly clays to sandy clays, clays and even

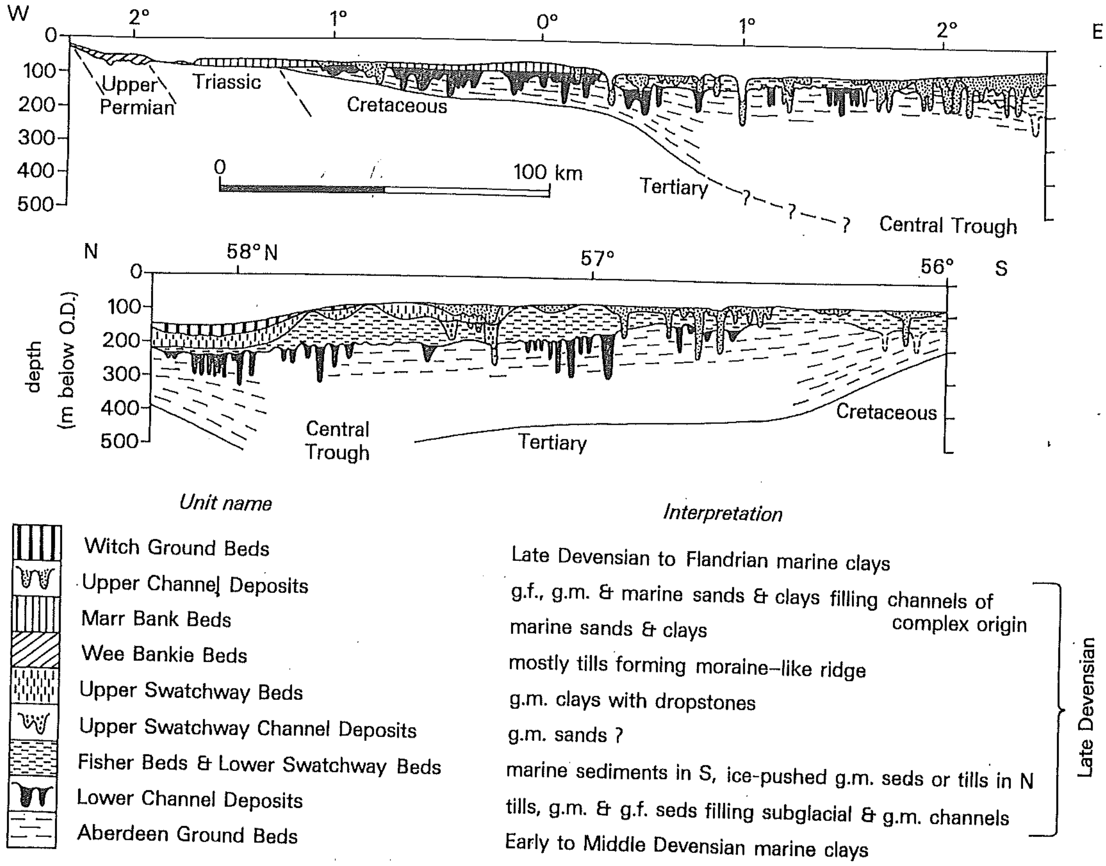


Figure 17.6 Stratigraphic relationships of the Quaternary deposits in the central North Sea, shown along the W–E and N–S sections of Figure 17.3b (after Holmes 1977).
gm = glaciomarine; gf = glaciofluvial; seds = sediments; OD = Ordnance Datum.

sands, depending on the local source material. Several lithologically different tills, all dating from the Devensian, may occur in a vertical section. These may represent deposition from different ice lobes with different source areas, although the topmost till is often just a weathered decalcified variant of the underlying one (Fig. 17.5). Glaciofluvial and glaciolacustrine sands, gravels and clays are often interbedded with, or incorporated into, the tills. These are now accepted as being broadly contemporaneous with the tills, the fluvial and lacustrine environments existing under, within and on top of the ice. However, in the past these deposits led to confusion, as it was thought that sands interbedded with tills indicated interglacial conditions separating two glaciations.

Devensian tills have been widely recognised in offshore areas. In the S. Irish Sea, overlying presumed Ipswichian interglacial deposits, Garrard (1977) has found up to 70 m of till, containing shells and local Palaeozoic and Mesozoic erratics together with far-travelled Scottish erratics. In Cardigan Bay, Irish Sea ice over-rode local Welsh ice flowing westwards out of the mountains. Lateral moraines deposited between the Welsh ice streams in Cardigan Bay are now preserved as offshore ridges known as sarns. Similar tills are found in the N. Irish Sea and on the shelf west of Scotland, except that here the youngest underlying sediments are of Neogene to (?)lower Pleistocene age.

In the central North Sea the situation is more complicated (Fig. 17.6). Here a channelled erosional surface, probably cut by glacial scour and subglacial streamflow modified by marine erosion, truncates the early Devensian Aberdeen Ground Beds in central areas and Mesozoic rocks near the coast. This is overlain by a complex pattern of tills, glaciomarine and glaciofluvial clays, sands and gravels cut by further channels, all attributed to the late Devensian glacial and bracketed by radiocarbon dates at *c.* 32 700 and *c.* 17 700 years BP. Stratigraphic and facies relationships are complex within these deposits, and deposition under grounded ice, floating ice, open marine and terrestrial conditions is inferred, with oscillation of the margins of both the Scottish and Scandinavian ice sheets (Holmes 1977). Possible ice-pushed ridges have also been recognised.

Foraminiferal evidence shows that the N. Atlantic at the latitude of the British Isles started to warm *c.* 13 500 years BP, and oxygen isotope work shows that global ice volumes were decreasing after 17 000 years BP. In the British Isles, ice had probably started to ablate from the lowlands by 14 500 years BP, leaving in places deposits of hummocky till, sand and gravel released from the melting ice. The ice front probably did not retreat in an orderly fashion as it does in mountain valleys. Rather, the ice ablated away in place, leaving dead ice masses impounding small lakes in which laminated clays were deposited. Recessional or readvance moraines of this date are rare. The Escrick and York moraines (Fig. 17.3) record some retreat up the Vale of York. However, numerous other features recorded as Devensian readvance moraines have been largely discredited and are now thought to be areas of topographically controlled hummocky ablation till.

As the ice ablated, the sea advanced across glaciofluvial outwash in the south and till spreads in the north. In the S. Irish Sea the transgression at first turned a major meltwater channel running parallel with the coast into an estuary in which bedded clays and silts were deposited. Further north the till is overlain by laminated muds and fine sands containing dropstones. Clearly the transgression had flooded considerable parts of the shelf while extensive glacier ice was still present. This is confirmed in the Forth Approaches area of the North Sea (Fig. 17.6), where the moraine-like Wee Bankie Beds may mark a recessional or readvance ice front that existed while shallow marine sands with an arctic fauna, the Marr Bank Beds, were being deposited to the east (Thompson & Eden 1977). The latter appear to overlie a plane of marine erosion cut by the transgressing sea. The Marr Bank and Wee Bankie Beds are cut by a series of channels, possibly eroded by the subglacial outflow of water from the Wee Bankie ice front together with tidal scour. Late-glacial marine clays with dropstones, the St Abbs Beds, which pass conformably upwards into Postglacial muds and sands, the Forth Beds, were deposited in these channels and against the moraine.

Although it does not seem to have left a record in the offshore sediments, a short cold period followed the warming and ablation of the main Devensian ice. Between 11 000 and 10 000 years BP ice advanced out of the Scottish mountains, but this time only as far as the Midland Valley, depositing till, outwash and leaving behind a good terminal moraine in many places (Fig. 17.3). This episode is known as the Loch Lomond readvance. Although there is debate about whether ice completely disappeared from the British Isles or just retreated to the high mountain areas during the preceding warm interval, the Late-glacial or Windmere interstadial, palaeontological evidence suggests that the interstadial was marked by temperatures as warm as, or warmer than, those of today (Coope 1975*). The Loch Lomond readvance is equated with a period of periglacial activity, during which Devensian tills and Late-glacial sediments outside the readvance limit suffered cryoturbation and blown sand deposits were formed. The intensity of this periglacial activity is demonstrated in SW Scotland by the main Late-glacial wave-cut rock platform which, although up to tens of metres wide, was cut in a relatively short time *c.* 11 000 years BP.

17f The Flandrian transgression

The melting of the Devensian and equivalent ice sheets throughout the world produced a rise in sea level of about 120 m between 17 000 years and 7500 years BP. The resulting marine transgression was the last major natural geological event to affect the British Isles, although its effects were felt around the world.

If it is assumed that the oxygen isotope curve measures changes in ice volume, a sea level curve can be calculated. This shows an unsteady slow fall during the Devensian, followed by a very rapid rise

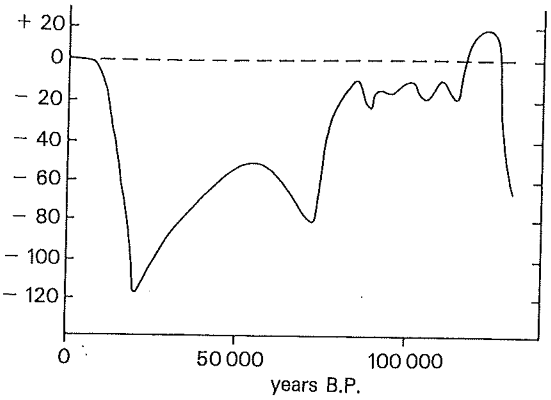


Figure 17.7 Sea level curve calculated from oxygen isotope data. Depths are in metres relative to present sea level (after Shackleton & Opdyke 1973).

(Fig. 17.7) starting *c.* 17 000 years BP. The repeated sawtooth pattern of the oxygen isotope curve (Fig. 17.1c) implies that similar sea-level changes occurred during each glacial–interglacial cycle. The evidence from offshore areas shows that sea level had already risen considerably before the ice disappeared from the continental shelf and that this rise continued, without any significant reversal during the Loch Lomond readvance, into the post-glacial Flandrian Stage. A more detailed curve for Flandrian sea-level rise can be constructed from, for example, the depth of dated peat and intertidal shell samples. This shows the rapid rise continuing until *c.* 7000 years BP, with only a slow subsequent rise of a few metres to the present day.

Although this transgression was much faster than the type of transgression that is familiar from the geological past, some of its geological effects were similar. As first the beach environment and then a strongly tidal shelf environment extended landwards, the top of the till spreads were reworked. The fines were washed out, leaving behind a thin layer of lag gravel. The fluvial sediments of river and meltwater channels were overlain by estuarine and then fully marine sediments, while, upstream, rivers aggraded.

In N. Britain the effect of the eustatic rise in sea level is complicated by the isostatic uplift of the crust following the removal of the ice. This uplift took the form of a doming, with the rate of rebound decreasing radially away from the centre of maximum ice loading in SW Scotland. Around the Forth Estuary a whole series of raised shorelines indicate a progressive emergence of the area during the ablation of the main Devensian ice, culminating in the widely recognised late Devensian shoreline that immediately predates the Loch Lomond readvance. The eustatic rise in sea level eventually gained the upper hand, leading to a transgression which terminated *c.* 6700 years BP with the formation of the main Flandrian shoreline. Subsequent isostatic uplift has led to a regression in N. Britain, where the Flandrian shoreline now stands as high as 15 m above sea level.

17g Theories for glaciation

Having discussed the nature of climatic change and its geological effects during the Quaternary, we may now briefly consider its causes.

Both terrestrial and extraterrestrial causes have been suggested for glaciations. The constancy of the sun has often been assumed, but the suggestion that its radiation output may vary with time is now taken seriously. The reception of energy from the sun may also vary as the solar system moves through parts of the galaxy with different concentrations of dust. The well-known periodic variations in the Earth's orbit around the sun may lead to a varying energy input to the Earth. Among the terrestrial factors are the distribution of continents and oceans. The transfer of heat from the equator to the poles is effected partly by ocean currents. Oceans oriented N–S are therefore more effective at warming the poles than those oriented E–W, and the motion of continental plates could lead to climatic change. The formation of mountain chains affects wind circulation patterns and provides areas for the growth of glaciers. The injection of volcanic dust into the atmosphere could affect the amount of absorbed radiation, as could the composition of the atmosphere itself with changes in carbon dioxide resulting from biological changes on Earth.

All these factors could influence climatic change, acting singly or in any combination. Does the nature of Quaternary record give any clues as to which of the many hypotheses for climatic change are the most likely? A striking feature of climatic curves, especially oxygen isotope curves, is their almost mathematical regularity (Fig. 17.1c). Each glacial

—interglacial cycle seems to follow a similar pattern. Are any of these hypotheses likely to produce such cyclicity? Are any amenable to analysis? The factor that does produce a calculable cyclicity is the astronomical variation in the Earth's orbit. Milankovitch calculated the variation in radiation received at a latitude of 60°N that was due to changes in the Earth's orbit and suggested that it was of the right amplitude and periodicity to produce glaciations. His calculations have been successively refined and have now been compared with oxygen isotope results (Hays *et al.* 1976). The similarity of the curves is evident, and it even stands up to close statistical scrutiny. Hays *et al.* (1976) therefore believe that this shows that the Milankovitch mechanism, as it is known, is the main cause of short-term climatic change, even though the actual links between a change in radiation received and the expansion of glaciers are not known.

Does this mechanism help with the interpretation of older glaciations, such as the late Precambrian or Permo-Carboniferous? Certainly an oscillating climate during such glaciations is suggested by the geological evidence (e.g. Ch. 4c), although it is not yet known whether the long term cyclicity indicated by the recurrence of major glaciations every few hundred million years could result from astronomical or geological causes.

17h Summary

The understanding of Quaternary history requires the integration of geomorphological, sedimentological, palaeontological and isotopic results from the land, the shelf seas and the ocean basins. The Quaternary record on land is very fragmentary, and the chronology of events will only be fully understood by reference to the more complete history preserved in the ocean basins and the more rapidly subsiding shelf sedimentary basins. The most complete Quaternary succession around the British Isles is found in the southern North Sea and the Netherlands, i.e. in the central part of the North Sea Basin. Here at least seven major cold—warm climatic oscillations are recorded, only the last three or four being cold enough to bring glaciers into lowland Europe. Outside this area, both on land and on the continental shelf, it is only Devensian (last-glacial) deposits that are common, and they usually lie erosively on preglacial sediments or rocks. Clearly, the effects of each glacial advance tend to be removed from the land and the shallow shelf seas by erosion during the subsequent interglacials and interstadials and by the glacial erosion of the next ice advance itself. It is only in the centre of rapidly subsiding sedimentary basins that glacial deposits have a good chance of being preserved.