

## Holocene Transgressions and Regressions on the Essex Coast Outer Thames Estuary

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### ABSTRACT

Faunal and lithological changes within the 34 m of Holocene sediments of the chenier plain suggest six episodes (I-VI) of marine transgression interspersed with five episodes of regression. Evidence presented for recognising transgressions includes chenier formation, marsh retreat and vertical changes in fauna, involving brackish water and intertidal molluscs. Regressions are interpreted chiefly from geosols and peat seams. Radiocarbon dates indicate that the transgressive episodes II, III and V first affected the region at about 7500, 4000 and 1400 B.P. respectively. The most recent one (VI) probably commenced about 300 B.P. The dates of episodes I and IV can only be conjectured. Comparison with similar events globally as well as in the Low Countries suggests that eustatic changes in sea level have played a significant role in the Holocene history of the Outer Thames estuary.

### INTRODUCTION

Evidence from deep boreholes and surface exposures in the Essex coastal chenier plain (fig. 1) is now such that it is possible to deduce several transgressive and regressive episodes through Holocene time. There appear to have been 6 transgressions, including that of the present day, interspersed with 5 regressions. In this paper the terms transgression and regression are defined on the basis of retreats and advances of salt marsh fronts respectively. The sea advances and in doing so erodes and diminishes the extent of the salt marshes or the sea retreats giving a seawards extension of the salt marshes.

The causes of the transgressions and regressions are complex and probably include regional subsidence, climatic and sedimentological factors. But, the new data also indicate that eustatic rises and falls in sea level may have been significant in the chronology of events.

The present coastal zone is formed predominantly of marine Holocene deposits now known to reach a thickness of at least 34 m beneath Maplin Sands (cf. Greensmith and Tucker, 1971a, p. 301). To the west the succession wedges out rapidly against Pleistocene gravels and brickearth and to the east, in the Outer Thames area, it also thins (D'Olier, 1972, personal communication).

Marine conditions appear to have been first established in Boreal times (c. 8900 B.P.) and at least 10 m of sediment accumulated. The succeeding marine deposits of Atlantic and early Subboreal age appear to be relatively thin compared with those of later Subboreal and Subatlantic age, a situation in contrast to that of The Netherlands (de Jong, 1971).

The sediments are predominantly calcareous clayey silts and sands and mainly sub-tidal and intertidal in origin. The clayey silts are relatively more dominant towards the bottom of the succession and the sands increase towards the top. This produces a situation in which the Flandrian transgression can be regarded, broadly, as having given rise to a major "coarsening-upwards" succession, though this view must be modified in detail. The Flandrian sea progressively inundated a late-Pleistocene terrain characterised by valleys bounded by gravel-capped terraces. Within these valleys the early Holocene successions are most complex and do not fall easily into the vertical lithological pattern suggested above.

The late-Pleistocene gravels were penetrated by several boreholes. The freshwater mollusc *Corbicula fluminalis* and a slightly worn tooth of the vole *Microtus cf. anglicus* were found in gravels at a depth of -15.5 to -17.5 m Ordnance Datum (Newlyn) beneath Foulness Island. However, some reworking of these gravels occurred during the Flandrian inundation and the top few metres invariably contain fragments of introduced Holocene shells. Other Holocene molluscs found the gravels a very suitable substrate to occupy.

The present day lithofacies distribution is generally straightforward (Greensmith, 1971, fig. 117). Fine sands occupy sub-tidal to mid-tidal areas, only occasionally impinging against the coastline, whereas the mid- to upper tidal areas are characterised mainly by clayey silts. Erosion of salt marshes is commonplace and at the edge of the marshes along the open coast and in estuaries are prominent migratory shell bodies (cheniers) (Greensmith and Tucker, 1965, 1967). Both these effects are witness to a phase of marine transgression at the present day. In the northern part of the area, at the mouth of the Blackwater and Colne estuaries, are gravel pavements of late-Pleistocene age which are being re-exposed and reworked (Greensmith, 1971, fig. 110; Greensmith et al, 1973). At

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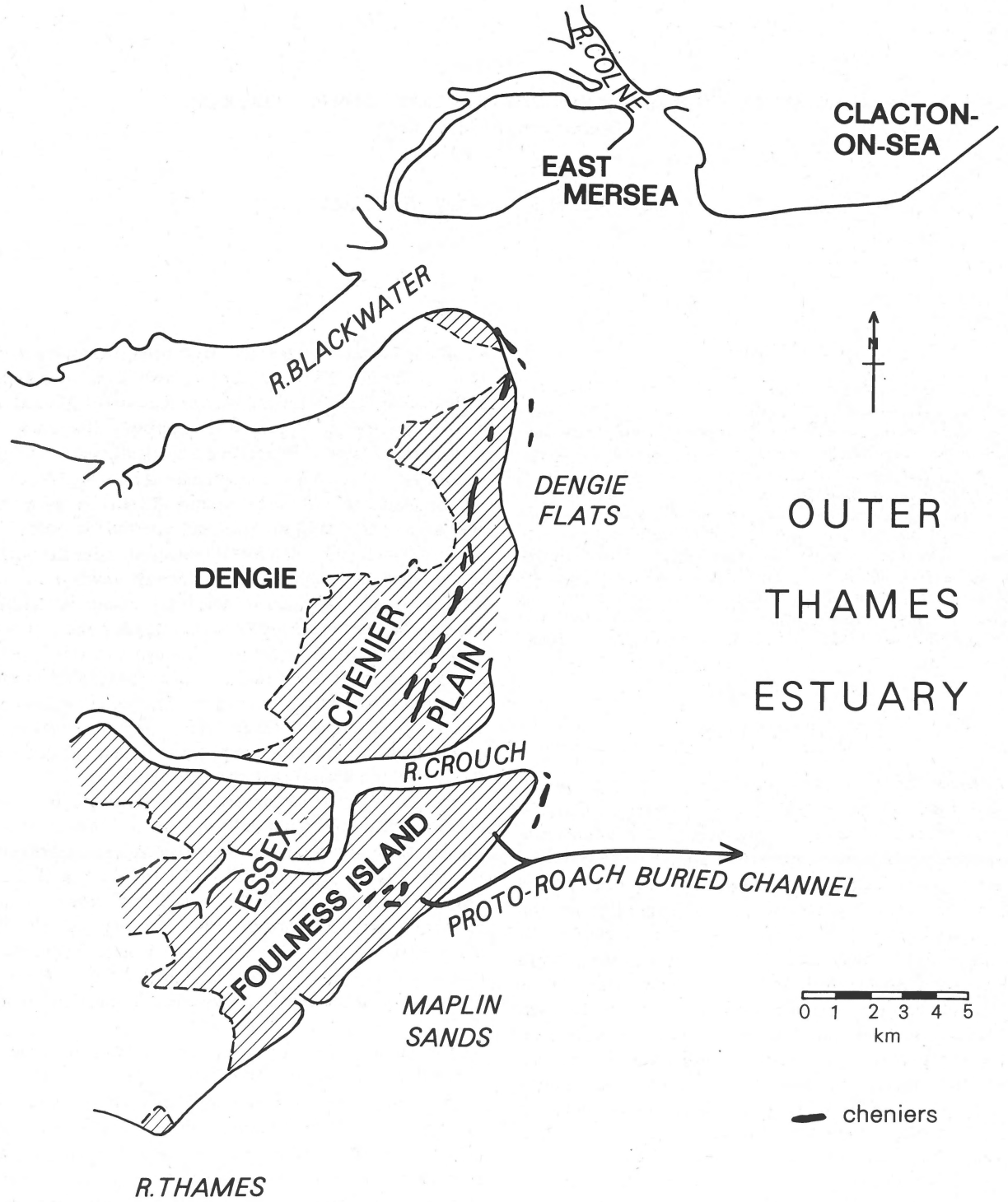


Fig. 1  
The Essex chenier plain (shaded). The inland chenier zone relates to transgression V and the present day cheniers to transgression VI.

mean low water mark the reworking produces elongate gravel ridges which are slowly migrating landwards.

## THE EVIDENCE FOR TRANSGRESSIONS AND REGRESSIONS

For convenience, the information bearing on transgressive and regressive episodes is categorised separately below. But it is necessary to appreciate that the phenomena described are very closely associated in the succession and their mutual relationships are often vital to the interpretation of the episodes.

### A. Transgressions

The key to the recognition of the transgressive episodes is found at the margins of the modern salt marshes where they are undergoing erosion. Because, it is commonly at these points of marsh retreat that sheets and ridges (cheniers) of shell and sand accumulate on the marsh surface. The field relationship recognised comprises a bed of transported intertidal and sub-tidal marine shells resting with slight discordance on marsh top deposits deficient in marine shells, but having an indigenous brackish to freshwater fauna including *Planorbis* sp., *Valvata* spp., *Spirorbis* sp. and *Hydrobia ventrosa* (fig. 2). The transported shells are variably broken and comminuted and include *Littorina* spp., *Hydrobia ulvae*, *Buccinum* spp., *Nassarius* spp., *Mytilus edulis*, *Cardium edule*, *Scrobicularia plana* and *Ostrea edulis* (see table 1 for more complete list). Beneath the cover of marsh vegetation the sediment is mottled and shows evidence of oxidation and leaching; shell fragments are uncommon.

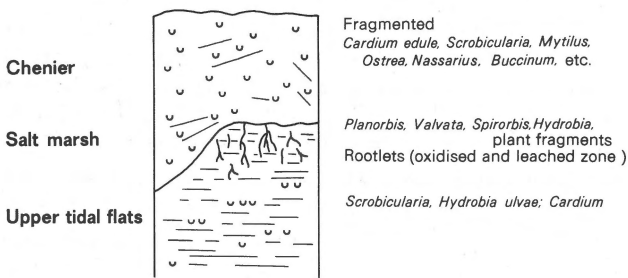


Fig. 2  
Common relationship of beds at the seawards margin of modern salt marshes.

The Essex cheniers are a type of sea-margin barrier, but it is important to realise that they differ in origin from the barriers occurring in the coastal area of The Netherlands (de Jong, 1971, p. 150-152). The Netherlands coastal barriers, dominantly sand in constitution, reflect regressive tendencies and slowing of sea level rise; this is in direct contrast to the conditions appropriate for chenier formation. Generally, barriers pre-date lagoonal - marsh sediments, though with subsequent submergence, synchronous deposition may occur (Hoyt, 1969).

When similar relationships are found in older deposits of

TABLE 1  
Facies and representative fauna in the Essex Holocene succession.

SUPRA-TIDAL (including creeks, salt marsh, lagoons, fresh- water streams and ponds)	INTERTIDAL	SUB-TIDAL
<i>Pisidium</i> spp.	<i>Littorina littorea</i>	<i>Nucula</i> spp.
	<i>L. saxatilis</i>	<i>Parvicardium exiguum</i>
<i>Cerastoderma lamarcki</i> ( <i>C. glaucum</i> )	<i>L. littoralis</i>	<i>Spisula</i> spp.
<i>Assimineia grayana</i>	<i>Hydrobia ulvae</i>	<i>Corbula gibba</i>
<i>Potamopyrgus jenkinsi</i>	<i>Retusa</i> sp.	<i>Ostrea edulis</i>
	<i>Mytilus edulis</i>	<i>Abra alba</i>
	<i>Macoma balthica</i>	<i>Buccinum</i> sp.
<i>Planorbis</i> sp.	<i>Abra tenuis</i>	<i>Gibbula</i> sp.
<i>Valvata</i> spp.	<i>Cardium edule</i>	<i>Mactra</i> sp.
<i>Hydrobia ventrosa</i>	<i>Scrobicularia plana</i>	<i>Venerupis</i> spp.
<i>Spirorbis</i> sp.	<i>Balanus</i> sp.	<i>Barnea</i> sp.
		<i>Nassarius</i> sp.

reclaimed marshland and at depth beneath the surface it is logical to assume that marine transgression is indicated. This assumption appears to be valid for a series of cheniers trending NNE-SSW in a narrow zone across the reclaimed ground of Dengie and Foulness Island (Green Smith and Tucker, 1969). The form, constitution and relationship of these cheniers to adjacent strata is virtually identical with those of the present day, even to the point of having ridge and runnel ("mud-mound") relief preserved on their seaward flank (Green Smith and Tucker, 1967). They are approximately 1400 years old.

Borehole data from a depth of approximately -8 m O.D. on both sides of the River Crouch suggest an even earlier (c. 4000 B.P.) phase of chenier growth and implied transgression. At Ridgemarsh on Foulness Island (fig. 5) drilling penetrated through about 3 m of a shell body now deduced to be a chenier composed of broken valves of *C. edule*, *O. edulis*, *S. plana*, *Macoma balthica*, *Nucula* sp., *Nassarius reticulatus*, *N. cf. incrassatus*, *Littorina saxatilis* and *L. littorea*, all set in a coarse sand matrix. The body rests directly on a plastic clayey silt with rootlet traces, but otherwise devoid of evidence for organic life.

A feature of retreating marsh fronts, whether they be associated with cheniers or not, is the accumulation of degraded marsh-top plant material on the immediately adjacent tidal flats. This plant matter (peatified in older deposits) is occasionally in sufficient quantity to warrant usage of the term plant-rich (peat-rich) clayey silt. At East Mersea this type of sediment can be seen not only forming adjacent to the present marsh cliff but also preserved within a succession of older tidal flat deposits exposed in the cliff face. The old peat-rich deposit in the face is overlain by *Scrobicularia*-bearing deposits of tidal flat aspect. Within these particular deposits there are also preserved peat-balls, up to 0.5 m in size, which clearly imply a phase of marsh erosion prior to that of the present day.

Another example of peat-rich clayey silt, dated as 7516 B.P., occurs beneath Foulness Island at a depth of -18.3 m

O.D. (Greensmith and Tucker, 1971a, p. 316). A significant feature of this level, as with that at Mersea and elsewhere, is the change in the faunal assemblage in the succeeding few centimetres of sediment. The peat-rich layers contain a range of fresh and brackish water forms including *Spirorbis*, *Helix*, *Assimineia*, *Pisidium*, *Valvata macrostoma* and *V. piscinalis*. Many of these shells are in an unworn condition, contrasting with any intermingled intertidal and sub-tidal shells which are worn and broken. This fauna is succeeded by an assemblage more fully intertidal and sub-tidal in composition. The implication is of a progressive, but rapid, change to open marine conditions.

In contrast, the preservation of "in situ" peat layers with similar faunal changes in the sediment above suggests a relatively quiet incursion of the sea across a marsh surface. Evidence from both the modern and older successions exposed at East Mersea and along the Blackwater estuary indicates that this is best accomplished in the more protected marsh tracts at some distance from the marsh front and behind marsh barriers. Excess sediment mobilised by erosion at the marsh front is transported through the creek systems into these protected areas where it is deposited quietly. Ultimately, a transgressive sequence is created in which clayey silts carrying an increasingly marine faunal element rest, with little sign of erosion, on a peat seam.

#### B. Regressions

The formation of "in situ" peat seams requires, in the first instance, high water tables and, in the second, prolonged periods during which vegetation can grow, decay and accumulate in quantity. In the Essex coastal area these conditions appear to have been at their optimum during marine regressive periods. However, the retreat of the sea can be expressed more positively in other ways. One of these is when clayey silt sediments are exposed to prolonged sub-aerial weathering and desiccation, so inducing overconsolidation in the surficial layers (Greensmith and Tucker, 1971b). At present these conditions apply in the innermost supratidal marsh zone and in the low-lying reclaimed ground immediately landward of the marsh zone.

There is abundant evidence for similar geosol-producing conditions existing in the past (cf. Morrison, 1969). Within the Holocene succession there are firm clayey silt layers up to 3 m thick. Fissuring, mottled structure, plant traces and relatively high  $Fe_2O_3 : FeO$  ratios (greater than 4) are typical. The high ratios reflect the oxidation of iron minerals, such as iron sulphides, and this is confirmed by a comparative deficiency in sulphide at these levels. There is also a deficiency in calcareous materials and dolomite is formed occasionally.

Field evidence demonstrating that overconsolidated layers can be readily preserved is found in the northern parts of the area along the flanks of the Blackwater estuary and as far east as Clacton-on-Sea. At East Mersea erosion at high water is exhuming an old, oxidised, overconsolidated surface — the

"Lyonesse Surface" — possibly a surface of exposure during the Last Interglacial but known to have been occupied subsequently by man during the Neolithic and early Bronze Ages, probably as late as 4000 B.P. but no later than about 3500 B.P. (Hazzledine Warren, 1918-19, 1932; cf. Cornwall, 1958, p. 208). The surface is very firm, is overconsolidated to a depth of 2-5 cm, has a reddish-brown colour and overlies firm grey clayey silt. Resting on the reddened surface are 1.5 m of plastic, light grey intertidal clayey silts referred to earlier. Towards the top is a peat seam which has been dated at  $118 \pm 48$  B.P. At the base is a filmy peat-rich clayey silt which is unlikely to be much older. In fact, the field relationship between the peat-rich deposit and a dated peat layer exposed nearby, and which also rests just above the overconsolidated layer, suggests that the age of the peat-rich deposit is  $173 \pm 60$  B.P. Similar plastic, peat-bearing deposits at Clacton-on-Sea were believed to pre-Roman by Hazzledine Warren (ibid).

Probably the most telling point about overconsolidated layers, such as that seen at East Mersea, is the accompanying faunal change from fresh-brackish to marine in the sediments resting immediately on the layers. A marked environmental change is signified.

Perhaps the most puzzling feature of the entire Holocene succession, which may have a relevance to regressive phases, is the presence of green layers within the sediments. These occur at depths between 0 and -30 m O.D. The bulk of the sediments in various shades of grey and brown, but the puzzling levels, up to 2 m thick, have a pale to olive green colour. The lithology is usually clay- and silt-bearing sand, in some instances occurring as pockets and seams in brown sandy gravels. The FeO content of the green sediment is low, between 0.13-0.79 per cent and the  $Fe_2O_3$  percentage is less than 3.

The cause of the colouration, which is more in the nature of a pigment, is obscure though may reflect the presence of ferrous iron within the illite and montmorillonite constituents (Grim, 1951; Keller, 1953). Illite is by far the dominant clay mineral forming some 55-90 per cent of the very low clay fraction. Montmorillonite rarely forms more than 10 per cent. The other clay constituent is kaolinite. It is feasible, but not confirmed, that a small amount of ferrous iron is locked into ferrous-organic complexes adsorbed onto the clay minerals (Friend, 1966).

Chlorite and glauconite have not been detected. Neither has the bluish-green iron phosphate vivianite; the amount of phosphate is very low, generally less than 0.20 per cent. In Essex and Kent known occurrences of vivianite are blue and invariably associated with decomposing Quaternary vertebrate debris set in peat-rich marsh clays.

Excluding deep-sea basins, non-glaucouitic green beds are universally associated with terrestrial deposition. Stagnant ground water, plant-rich, reducing conditions are usually invoked. In Kent, interglacial green silts appear to have accumulated in stagnant pools on alluvial flats; they are rich in *Planorbis*, *Succinea* and *Limnaea* (Carreck, 1972).

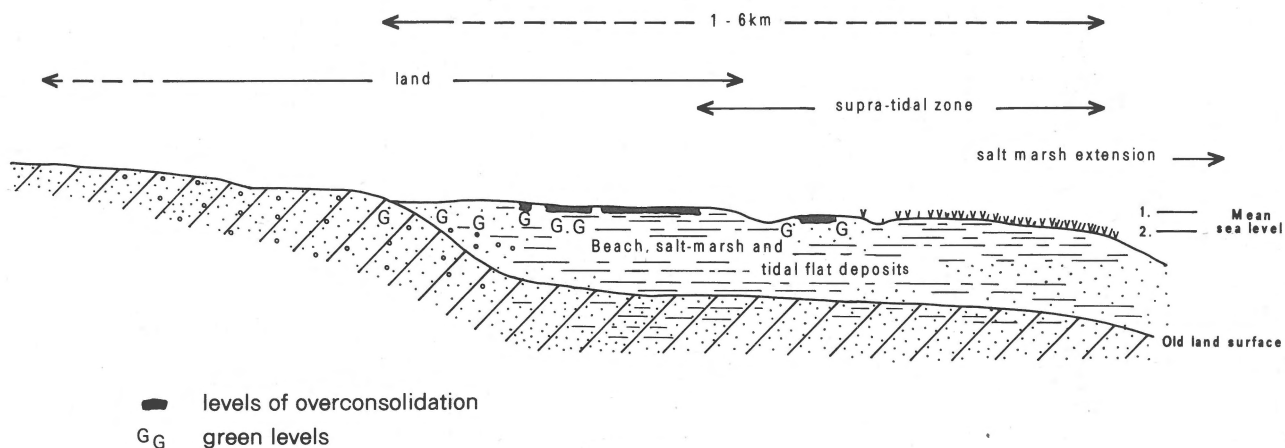


Fig. 3 Schematic illustration of the development of green and overconsolidated levels (geosols) during marine regressive episodes. Salt marsh extension occurred either during falls in mean sea-level (from position 1 to 2) or during stillstand (position 1).

Hazzledine Warren (1932) has recorded at Clacton-on-Sea Quaternary green silts representing a late infill of a channel cut in London Clay. In all these examples the green colour appears to indicate sub-aerial processes probably akin to "gleying" (Perrin and Hodge, 1965, p. 77).

Beneath Foulness Island and Maplin Sands the green levels are also characterised by partial chemical corrosion of the few surviving shell fragments. SrO contents drop to values of 10-155 ppm compared with 100-1200 ppm of the closely associated beds. Some degree of leaching is implied.

The evidence, therefore, allows a tentative interpretation of the green levels as being the residual traces of emergent surfaces once exposed to sub-aerial processes, that is geosols. Relevant to this interpretation is the fact that the faunal content of the immediately overlying sediment again shows a rapid upwards change from fresh-brackish water types to marine types.

In summary, when the total data is assembled the indications are that the margins of Essex were partially emergent above high water mark on several occasions during Holocene times. During these regressive phases it is probable that sub-aerial exposure produced thin soils (geosols) over wide tracts, these merging laterally into salt marshes bounding channels. The schematic relationship is illustrated in fig. 3.

### CHRONOLOGY OF TRANSGRESSIVE AND REGRESSIVE EPISODES

The present evidence suggests that the Flandrian transgression across the late-Pleistocene landscape of the Essex borderland has proceeded intermittently, being interrupted by at least five phases of regression. Six phases of transgression have been recognised including the initial Flandrian (I) and the present day (VI) (fig. 4).

By late-Pleistocene times the basic morphology of the area

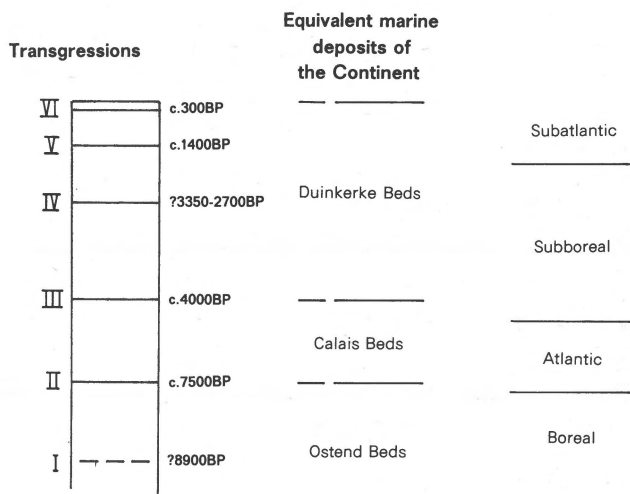


Fig. 4 The Essex transgressions related to the marine Holocene succession of the Continent.

was established. Fluvial channels, alluvial flats and terraces, the lowest at about -15 m O.D., extended from the hinterland eastwards across Essex into an embryonic North Sea.

It has been estimated that sea level at 9600 B.P. was at -45 m O.D. and rising rapidly (Jelgersma, 1961; D'Olier, 1972). Fluvial silts, sands and gravels were being deposited and reworked in the valleys and this condition continued through until about 8900 B.P. when the first impact of the Flandrian transgression was felt at Foulness, Maplin and along the River Crouch depression. Sea level was then at about -34 m O.D. This situation contrasts with that of The Netherlands, where the influence of the sea was not felt until its level had risen to about 20 m below present level (de Jong, 1971, p. 147) From then on in Essex, through Boreal into early Atlantic times, estuarial deposition prevailed, leading to the accumulation of at least 10 m of sediment; a thickness proved along the axis of the proto-

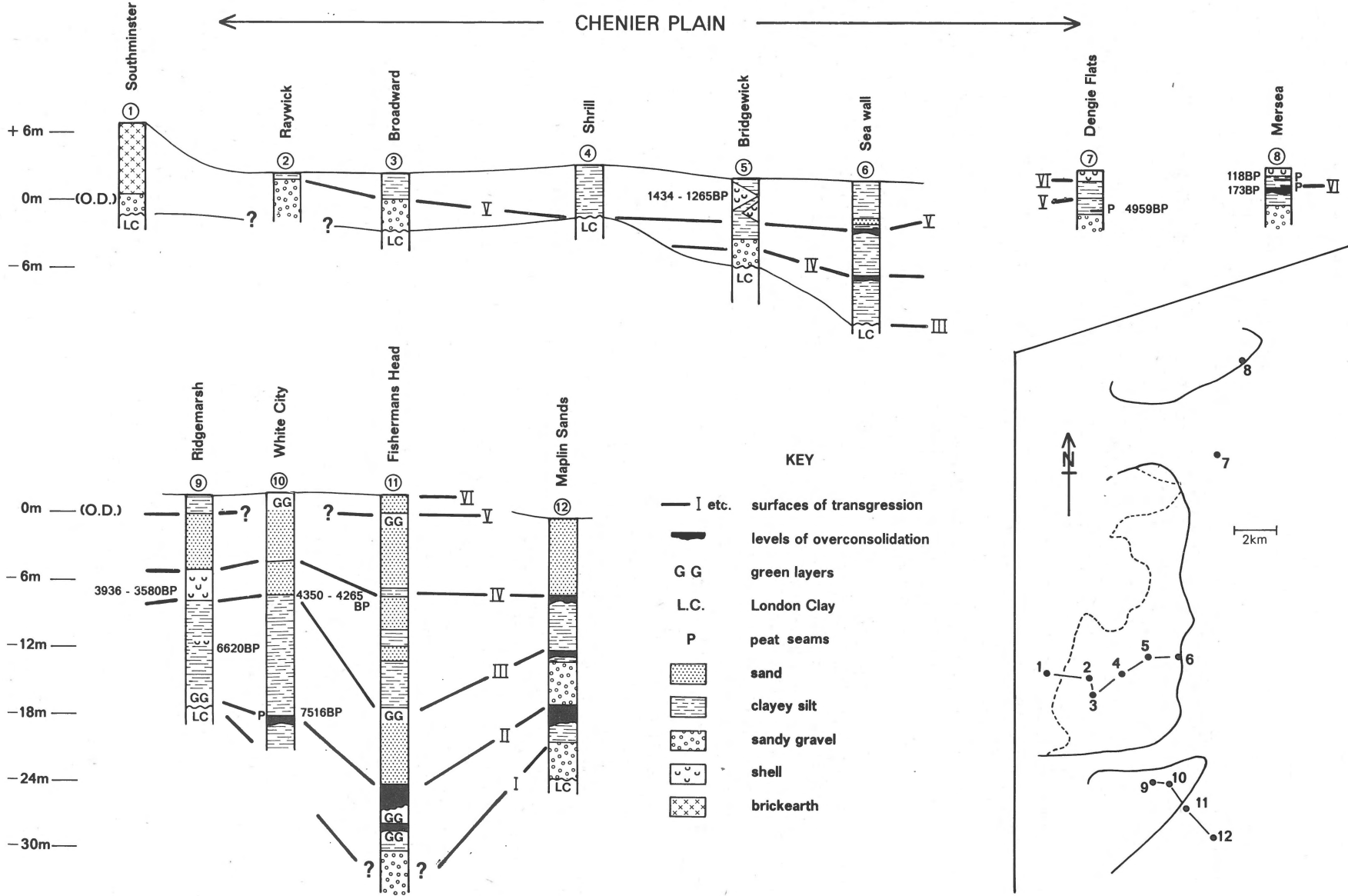


Fig. 5  
Cross-sections through the chenier plain indicating levels of transgression.

Roach channel at depths greater than  $-21$  m O.D. These deposits are younger than originally believed and are probably equivalent to the "old fluvial deposits" of the Dutch coast (de Jong, 1960) and Ostend Beds of Belgium and northern France.

An important feature of these old Essex deposits is their degree of overconsolidation, almost to the full extent of their thickness. The implication is of at least one, possibly two, relatively prolonged regressive episodes during which the sediments became exposed (Greensmith and Tucker, 1971a, fig. 3). The dating of the episodes is conjectural except for some evidence beneath Ridgemarsh, Foulness Island (figs. 1 and 5). At this locality, at a depth of  $-18.3$  m O.D., occurs a plastic, peat-rich clayey silt which has been dated as  $7516 \pm 250$  B.P. and which rests on strongly overconsolidated beds carrying fragmented shells, including *Cardium*, *Spisula* and marine foraminifera. In quality of induration these old beds closely resemble those in the deep channels and it is likely that they were laid down in early Atlantic times. Hence, the regressions pre-date 7516 B.P. In terms of dating, the eustatic sea level curves of Fairbridge (1961) and Mörner (1971a) are pertinent, as both indicate a marked fall in sea level extending over about 1000 years from c. 8700-7750 B.P. The changes in sea level are estimated as being between 2 m and 9 m, figures quite compatible with the Essex evidence.

The peat-rich deposit dated at 7516 B.P. fixes in time a major Atlantic transgressive episode (II in figs. 4 and 5), referred to as the "Hydrobia" transgression by Fairbridge (1958). The deposit marks in Essex the base of the deposits known as Calais beds on the Continent. At this time the local, regional and global evidence points to a pronounced rise in sea level, probably co-inciding with global climatic amelioration – the "post-glacial climatic optimum" – initiated, according to Mörner (1971b), at about 7750 B.P. (also see van Straaten, 1954, fig. 1; de Jong, 1960). Submergence of the irregular Essex landscape was comparatively rapid as dates from unworn and articulated *Ostrea* valves beneath Foulness Island show that sea level must have been at least 4.5 m higher in  $6620 \pm 100$  B.P. By  $5650 \pm 240$  B.P., a date obtained at  $-12$  m O.D., it is probable that most of the early Atlantic deposits and certain marginal parts of the older land surface had been further inundated. This land surface partly comprised green surface layers and pockets in Pleistocene sands and sandy gravels now found at depths between  $-12$  m and  $-15$  m beneath Foulness Island.

On Dengie Flats, a peat seam at about  $-1.5$  m O.D. is dated at  $4959 \pm 65$  B.P. This suggests that the second Flandrian transgression had ceased, being replaced by a regressive phase. The peat is interbedded with intertidal deposits, so probably formed close to high water mark. If this assumption is valid then a comparatively high mean sea level is implied for about 5000 B.P., possibly only a few metres below present mean sea level. Fairbridge (1961) advocates high sea levels at this time.

The regressive episode appears to have spanned a sufficient length of time to produce green beds and overconsolidated layers, which are developed extensively below Maplin Sands and Dengie Flats at depths between  $-11$  m and  $-7$  m. Also, it may be relevant that occupation of the "Lyonesse Surface" commenced at about 4700 B.P. The termination of the regression is conjectural though the faunal, lithological and radiocarbon evidence suggests a date around 4350-4260 B.P. This date is derived from an assemblage of partly broken shells of *Cardium edule*, *Ostrea* and *Nassarius* at a depth of  $-7.5$  m O.D. below Foulness Island. The shells occur at the base of a sand unit which rests with sharp contact on clayey silts.

The marked change in lithology marks the onset of an early Subboreal major transgression (III in figs. 4 and 5) and is probably the Essex equivalent of The Netherlands "Cardium Transgression" (Hageman, 1962). The change also possibly defines the junction between beds equivalent to the Calais Beds and overlying Duinkerke Beds of the Low Countries (fig. 4). At depths between  $-5.5$  m and  $-10.5$  m O.D. the transgression is defined by thick sandy shell bodies. The one beneath Ridgemarsh, for which detailed information is available, resembles a chenier and it can be surmised that other shell bodies recorded in wells at about the same level are also cheniers or degraded shell banks. *Cardium edule* valves from the Ridgemarsh chenier, at a depth of  $-5.5$  m to  $-8.3$  m O.D., provide dates of  $3936 \pm 110$ ,  $3912 \pm 114$  and  $3580 \pm 75$  B.P. According to Fairbridge and Mörner eustatic sea level at this period rose above that of the present day. Such a rise could account for the abandonment of the Neolithic-early Bronze Age coastal occupation sites at Mersea and Clacton-on-Sea. However, the marine deposits resting on the occupation surfaces are much younger than believed by Hazledine Warren (1932), so reasons other than inundation may have caused Bronze Age man to move away (cf. Akeroyd, 1972, p. 161).

From about 3500 B.P. onwards to Romano-British times (0-200 A.D.) evidence for regressions and transgressions is present, but dating is circumstantial because of an absence of radiocarbon dates. Beneath Dengie Flats at  $-7$  m O.D. and in the southern and central parts of Foulness Island and Maplin Sands at depths between  $-4$  m and  $-7$  m there are overconsolidated clayey silts and clayey sands ranging in thickness between 0.5-2.5 m. At the latter localities clean sand carrying an open marine fauna rests with sharp contact on the overconsolidated beds; these are the first deposits of the transgressive episode IV (figs. 4 and 5). Elsewhere under Maplin and Dengie the break is essentially physical and faunal, not markedly lithological.

The data suggest prolonged emergence of certain parts of the coastal plain and it is possible that this relates to a major Subboreal fall in eustatic sea level, deduced by Fairbridge and Mörner to have started at about 3350 B.P. and ended at about 2400 B.P.

From archaeological data it is certain that a regression was in being during Romano-British times, though the time of

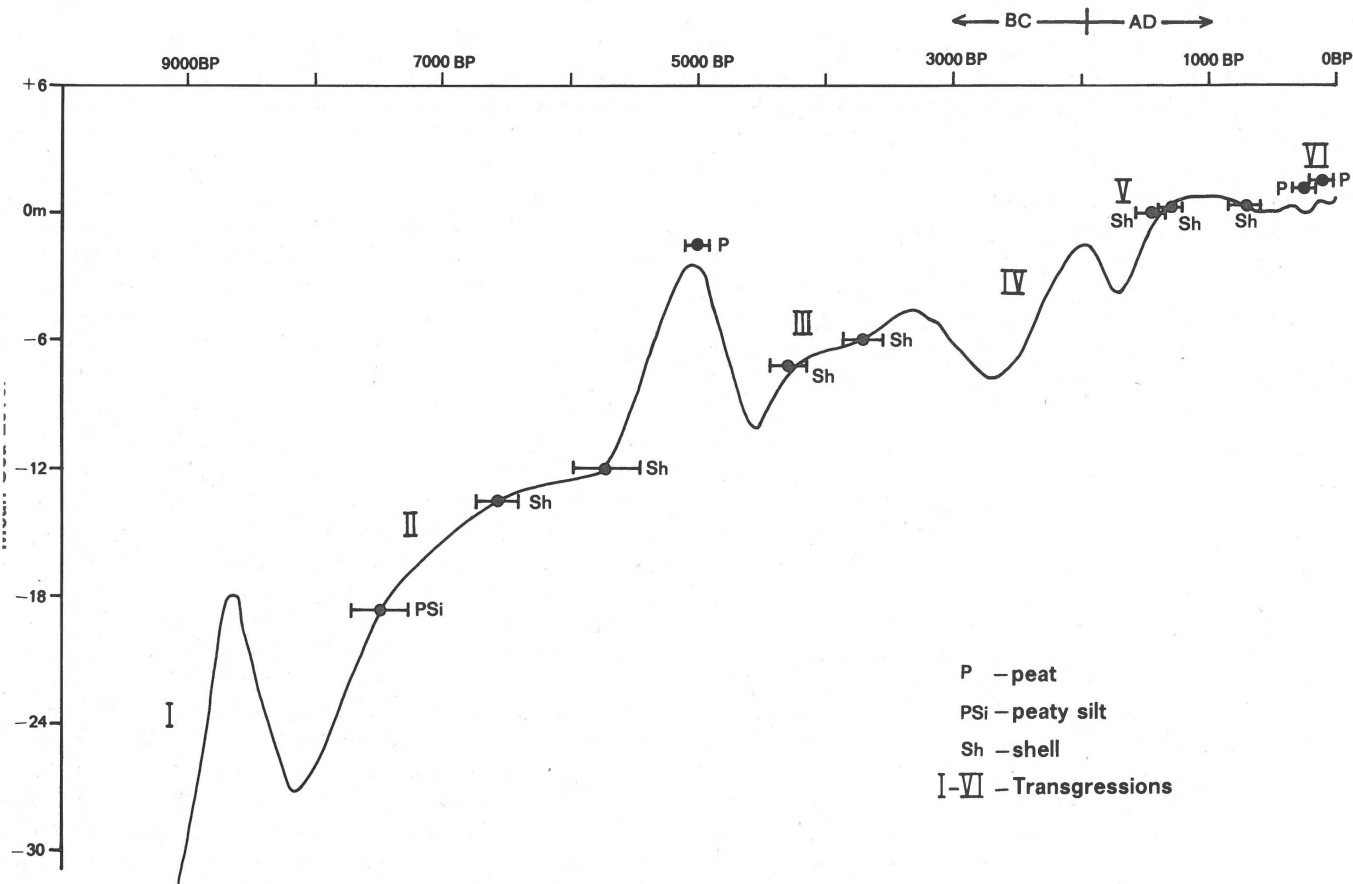


Fig. 6  
Changes in mean sea-level based on radiocarbon dates, lithological, faunal and geotechnical data.

initiation is again in doubt. It is possible that the regression equates with the Florida (Roman) Emergence advocated by Fairbridge (1962); an emergence which reached its maximum in 150 A.D. During the Roman occupation of Essex, river mouth forts and outstations were erected on and immediately adjacent to the upper salt marshes of Dengie and Mersea Island; their foundations now partly lie below high water mark under a younger sediment cover. Coastal saltern ("red hill") sites were also established on the upper salt marshes over the same period; some of these have been subsequently transgressed (Stopes, 1887; Cole, 1905-6; Smith, 1917-18; Linder, 1940). Also relevant to the same phase of regression are overconsolidated and green beds occurring at depths between -2 m and -4 m O.D. below Dengie Flats and Foulness Island.

Above this once emergent Romano-British surface marked faunal changes occur in the succession, the most pronounced being represented by the Dengie and Foulness inland cheniers (Greensmith and Tucker, 1969). The main and ultimate period of chenier growth gives the following dates,  $1434 \pm 110$ ,  $1410 \pm 100$ ,  $1340 \pm 100$  and  $1265 \pm 100$  B.P., but there is evidence for a slightly older minor period of

chenier development in the south of Dengie. The main cheniers are located at an altitude which varies between +1.5 m and -1.0 m O.D.; the base of the minor older series is at -2.5 m O.D. They rest on 2-15 m of clayey silts. On their immediate seaward flanks are shell sheets and lenticles incorporated in clayey silts strictly analogous with those on the present day flats. This information clearly indicates an important Subatlantic transgressive episode (V in figs. 4 and 5). It appears to be equivalent to the regionally important late Flandrian inundation of the Flemish coastal plain recorded as occurring between the 4th and 8th centuries by Tavernier and Moormann (1954). Pons *et al* (1963) record, for the same period, peat erosion and large-scale inundation in the Rhine-Meuse and Northholland areas. In some localities clay and silt were laid down directly on the top of peat.

A feature of the main Dengie cheniers is their present altitude in reclaimed, drained ground known to have subsided. The altitude co-incides closely with present normal high water mark. If one accepts that compaction and settlement of the underlying plastic clayey silts have occurred, then the implication is that, at the time of formation of the cheniers, sea levels slightly higher than those of the present

day existed (cf. Fairbridge, 1958, p. 479, 1961).

The time of termination of the post Romano-British submergence in Essex is not known. If the events in the Flemish plain are at all comparable then it ended in the 8th century. There is then a gap in the record until the marsh extension was coming to a close (Fairbridge, 1958; Davis, 1964). Enclosure and reclamation of salt marshes and sea wall building were widespread at that time (Cole, 1905-6; Grieve, 1959). The subsequent major marine transgression (VI in figs. 4 and 5), with erosion of exposed salt marshes, has continued through to the present day, not without temporary pauses. At Mersea there is evidence indicating regressions at 205, 173 and 118 B.P., but these appear to have been minor events in the generally erosive and transgressive episode.

### SUMMARY

The information available at present suggests that the Flandrian transgression in the east Essex and Outer Thames region has not been a uniform extension of the southern North Sea across a regionally subsiding, dissected hinterland. Instead, there were several pauses of varying degrees of magnitude during which sea level either remained static or fell. At these times, parts of the Essex coastal zone became emergent by accretion, relative uplift or both, and susceptible to sub-aerial weathering processes. At present there is no evidence that coastal sand barriers analogous to those of The Netherlands played a significant role in sedimentation during these emergent phases, or indeed were ever formed.

After the initial Flandrian submergence (I), commencing at about 8900 B.P., further transgressive episodes (II to VI) occurred at about 7500, 4000, 2400, 1400 and 300 B.P. The episodes since 4000 B.P. seem to have been of most importance in determining the ultimate modern morphology of the coastal plain.

The causes of the transgressive and regressive episodes are complex in a region known to have subsided throughout Holocene times. Variations in rates of subsidence and sediment supply, alternations of wet and dry phases and changes in wind direction frequencies in the North Sea all may have played some role. But the evidence, especially for transgressions II, III, V and VI, indicates that eustatic changes in sea level have been the controlling factor. Otherwise it is difficult to account for the fact that the episodes correlate so closely with similar events in the Low Countries and globally. The total faunal, lithological, environmental and radiometric data from Essex appears, therefore, to support the deductions of Fairbridge (1958, 1961, 1962) and Mö rner (1971a, 1971b) on Holocene eustatic changes. Figure 6 summarizes this data in terms of changes in mean sea level, though the inferred curve represents a tentative assessment only.

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