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Holocene Sea Level Rise in the Western Baltic and the Question of Isostatic Subsidence

KYAW WINN, FRITZ-RUDOLF AVERDIECK, HELMUT ERLENKEUSER, FRIEDRICH WERNER

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Sea level related radiocarbon, palynological and stratigraphical data from sediment cores in the Western Baltic have been tested against the existing sea level curves for the region. The relative sea level rise curves for the beginning of the Holocene show no significant deviations between the Kiel, Mecklenburg und Lübeck Bays and hence do not support the previously reported differences in the averaged regional subsidence rates for this time interval. Local subsidence and upheaval due to salt tectonics probably played a greater role than previously suspected in the region.

The sea level possibly stagnated around -28 m during the early Holocene before rising very rapidly to -14 m. The submarine terraces at -30 m and perhaps also at -27 m were formed during the lacustrine phase of the Western Baltic when the water levels were controlled by the main thresholds in the Great Belt.

Dr. F. WERNER and Dr. K. WINN, Geolog.-Paläont. Institut und Museum der Universität, Olshausenstraße 40, D-2300 Kiel. - Dr. F.-R. AVERDIECK, Inst. f. Ur- und Frühgeschichte der Universität, Olshausenstraße 40, D-2300 Kiel. - Dr. H. ERLENKEUSER, Institut f. Reine und Angewandte Kernphysik der Universität, C14-Labor, Olshausenstraße 40, D-2300 Kiel.

Kurzfassung

Eine größere Anzahl von älteren und jüngeren Altersdatierungen und von pollenanalytischen Daten aus holozänen Transgressionshorizonten der Kieler und der Lübecker Bucht wurde zusammengestellt, um den wahrscheinlichen Verlauf der holozänen Meerespiegelanstiegs einzuengen. Die Daten stammen sowohl von den über der Transgressionsbasis liegenden marinen Schlicksedimenten wie von den darunterliegenden Süßwassersedimenten, u. a. Torfen, Torfmudden und humosen bis kalkigen Gyttjen. Die „Transgressionskurve" wird weniger als Kennlinie des genauen Meerespiegels verstanden denn als Einengung eines Hiatus, der oft in der Transgressionsabfolge direkt nachgewiesen werden kann.

Die Daten werden im Hinblick auf die Frage der isostatischen Senkung südlich der eisisostatischen Nullinie geprüft. Hier zeigt sich

- durch das Fehlen eines signifikanten Gradienten des Transgressionsniveaus senkrecht zu den „Senkungsisolinien" zwischen dem mittleren Großen Belt und der mittleren Kieler Bucht,
- durch ebenfalls niveaugleichen Verlauf frühholozäner submariner Terrassen,

daß eine relative, regionale Senkung des schleswig-holsteinichen Küstengebietes im Gegensatz zu früheren Auffassungen nicht wahrscheinlich ist.

Lokale Senkungs- oder Hebungstendenzen aufgrund von Salzstockbewegungen oder anderer Ursachen sind dagegen möglich. Der Vergleich der Kurve mit bisherigen relativen Transgressionskurven unseres Gebietes ergibt folgendes:

- Zwischen einer Stagnation (oder sogar leichter Regression) bei –28 m und dem Niveau –14 m liegt der sehr rasche Anstieg des Meeresspiegels;
- Ein treppenähnlicher oder sogar ein durch kurzfristige Regressionen unterbrochener Verlauf der Meeresspiegelkurve ist im steilen Anstieg der Littorinatransgression nicht nachzuweisen.

Die submarinen Terrassen bei –30 m und vielleicht auch diejenigen bei –27 m werden während der lakustrinen Phase der westlichen Ostsee gebildet.

Introduction

The Western Baltic Sea, comprising the Danish Belt Sea, and the Kiel, Mecklenburg, Lübeck and Neustadt Bays, has generally been depicted as an isostatically sinking area with the tectonic zero line passing through the Belt Seas (GUTENBERG 1941). Although averaged differential subsidence rates which gradually increase southwards have been established for the Holocene in the Mecklenburg, Lübeck and Neustadt Bays (KÖSTER 1961; KOLP 1979), extrapolation of these rates westwards is questionable, mainly due to the presence of active salt bodies which particularly occur in the western part of the Kiel Bay (VOSS 1968; HINZ et al. 1971; PRANGE 1985).

Local studies have already been made for most of the Western Baltic (KÖSTER 1961; KOLP 1965; EXON 1972; WINN 1974; WINN et al. 1982, 1983, 1984; NIEDERMEIER-LANGE 1985; SIMANOWSKY 1985) including a comprehensive paper by SEIBOLD et al. (1971). In this study, we present new core data from the outer Eckernförde fjord, a synthesis of the available radiocarbon datings, palynological analyses and sedimentological-stratigraphical data from sediment cores, and evaluate their significance for the estimation of the subsidence rates and eustatic rise of sea level in the Western Baltic Sea during the early to middle Holocene.

We are grateful to the captains and crews of the research vessels "Wattenberg", "Alkor", "Littorina" and "Poseidon" of the Christian Albrecht Universität for their unstinting help during the investigations at sea. Thanks are also due to Prof. H. D. SCHULZ for granting access to his organic carbon, carbonate and ^{14}C measuring instruments, Miss M. NEVE for the pollen analyses, to Mr. W. BREUER for drafting the pollen diagram, the staff of the University of Kiel Radiocarbon Dating Laboratory for the ^{14}C datings, Miss W. REHDER for the radiographs, and also for the ^{14}C analyses on core 15 496, and to Messrs G. TÜNSCHEL, H. HENSEL and W. REIMERS for help during coring. We also wish to express our gratitude to Prof. R. KÖSTER and K. DUPHORN for critically reviewing the manuscript.

Physical and hydrographical settings

The Western Baltic Sea is a relatively young water body which was formed after the retreat of the ice sheets towards the end of the Weichselian glaciation. An interconnected deep channel system with thresholds or sills between its different sections is typical of this sea. During the Late Glacial and the early Holocene, this physiography gave rise to an intricate network of freshwater lakes with the thresholds at –26 m to –27 m in the Great Belt (WINN 1974) acting as the most important barriers against the Kattegat Sea in the north. After the sea level rose above this level, the marine incursions, generally termed the Littorina Transgression, gradually penetrated into the Western Baltic (KOLP 1965; WINN and AVERDIECK 1984) and subsequently into the main Baltic Sea. The present-day Western Baltic with the Belt Sea and the Copenhagen Sound forms the only major natural connection between the Baltic proper and the Kattegat/Skagerrak Seas, incorporating a complicated system of outflow and inflow (DIETRICH 1950).

Sea level rise: Indices

Palaeogeomorphology

An accurate knowledge of the palaeogeomorphological configurations is necessary in deciphering the complex pattern of sea level rise as depicted by the sediment record. A key position is occupied by the thresholds which have acted as barriers against the incursions of the sea between the different sectors during the early Holocene. The most important threshold for the Baltic is situated north of Sprogö Island in the Great Belt. Since it exists in an tectonically almost stable area (GUTENBERG 1941; MÖRNER 1969), its present depth of -27 m (WINN 1974) in first approximation should also hold in the early Holocene, when the salt water inflow into the Baltic (Yoldia Sea and freshwater Ancylus Lake phases) through middle Sweden was completely cut off due to isostatic upheaval. This depth of -27 m is the maximum assuming that erosion of this threshold is negligible. In the southern entrance of the Great Belt and also between the Vejsnaes area and the southwestern part of the Kiel Bay, further thresholds are found with present-day depths of about -28 m and about -30 m, respectively. In the interpretation of the eustatic rise of sea level and local subsidence rates, these boundary conditions have to be taken into consideration. Thus all depths of the marine transgression recorded in the sediment succession below the major threshold in the Great Belt have to be adjusted at least to this level, after allowance is made for probable isostatic movements. Fore example, a transgressive sequence found at -29 m NN cannot represent a sea level at this depth.

Stratigraphy

A typical sediment succession for the Western Baltic is illustrated by the following sequence found in core 15 496 (fig. 1, plate 1) from the southwestern part of the Kiel Bay. The pollen zonation is after OVERBECK (1975). The related absolute chronology after ^{14}C dating (WILLKOMM, 1976) is shown in parentheses. Zonal transitions may encompass a few hundred years and are not fixed chronologically here.

Core no. 15 496 — Water depth 25.6 m

Pollen-zone	Approx. Chronology	Depth in sediment	Lithology
IX	3 000 B. P.-	0–170 cm	Clay, greenish grey, soft, increasingly compact with depth.
IX	5 100 B. P.	170–220 cm	Clay, greenish grey, compact, with plant remains and small peat fragments showing a peppery texture. Faint laminations discernable. ^{14}C age (organic carbon, see also tab. 3) from 213–217 cm is 6 040 ± 310 years B. P.
IX/VIII		220–223 cm	Transition to dark grey clay gyttja, strongly bioturbated.
VIIIa	7 000? B. P.-	223–275 cm	Clay gyttja, dark to blackish grey, with lighter layer between 240–250 cm having less detrital organic content. ^{14}C-age (organic carbon) from 268–272 cm is 8 370 ± 580 years B. P.

VIIIa		275–282 cm	Transition to light grey lake marl.
VIIIa		282–299 cm	Lake marl, light grey, with thin (less than 1 cm) organic rich layers.
VIIIa/ VIIb	8 200 B. P.	299–312 cm	Lake marl as above, with numerous organic rich layers.
VIIIa/ VIIb	8 200 B. P.	312–342 cm	Lake marl as above, with a few organic rich layers.
VIIb		342–409 cm	Lake marl as above, with numerous organic rich layers.
VIIa	–8 900 B. P.	409–432 cm	Clay gyttja, dark to blackish grey, alternating with lighter layers, units 6–8 cm.
VI	8 900–	432–458 cm	Clay gyttja, calcareous with organic remains, (mainly *Cladium*), increasing with depth.
VI		458–466 cm	Peat, mainly of *Cladium*, wet, reddish to blackish brown with thin gyttja lenses. Hazel nuts present at the base. ^{14}C-age (organic carbon) is 9 550 ± 660 years B. P.
VI	–9 250 B. P.	466–497 cm	Lake marls, grey with abundant shells, especially *Lymnaea baltica, Valvata piscinalis piscinalis,* and *Sphaerium corneum.* Organic rich stringers and vertical roots of *Cladium.*
V	9 250 B. P.– 10 150 B. P.	497–557 cm	Lake marls above, sandier, alternations of lighter and darker layers (6–12 cm units). ^{14}C-age (organic carbon) from 548–552 cm is 9 330 ± 440 years B. P.
IV		557–573 cm	Lake marls as above.
		...	

Analytical results of the water, organic carbon and carbonate contents as well as the dry and wet bulk densities (dry weight/ wet volume and wet weight/wet volume) are given in tab. 2. As expected, the carbonate content in the marine clays is practically zero. Foraminiferal masks and dinoflagellate cysts were observed signifying that carbonate dissolution had taken place. The relatively high $CaCO_3$ content (over 30 %) coupled with the total absence of foraminifera and dinoflagellates in the underlying clay gyttja (organic lake marl?) between 223 – 275 cm further supports the interpretation that these sediments have been deposited in a predominantly freshwater environment. These gyttjas or organic rich lake marls represent a deeper water facies of the subaerial to limnic sands studied by WEFER et al. (1978) in the adjacent Boknis Eck area. The freshwater to brackish (?) clay gyttjas overlying the peats have largely been overlooked in the earlier core studies.

The density and water content data indicate that most of the sediments are not fully saturated with water. Sample extraction had been carried out using a 10 cc steel cylinder immediately after unsealing the core sections in the laboratory. Therefore, losses due to evaporation and to systematic errors are estimated to be minimized and < 5 %. According to our results, the undersaturation ranges from 10 % to 20 %.

Palynology

Palynological analyses have been carried out in the "Institut für Ur- und Frühgeschichte" Kiel, on sediment cores from the Baltic since 1962 (tab. 4). During the early stage of the investigations, only sediment cores with peat or peatlike coarse detrital gyttja which underlie the marine sediments were analysed. Later, the work was

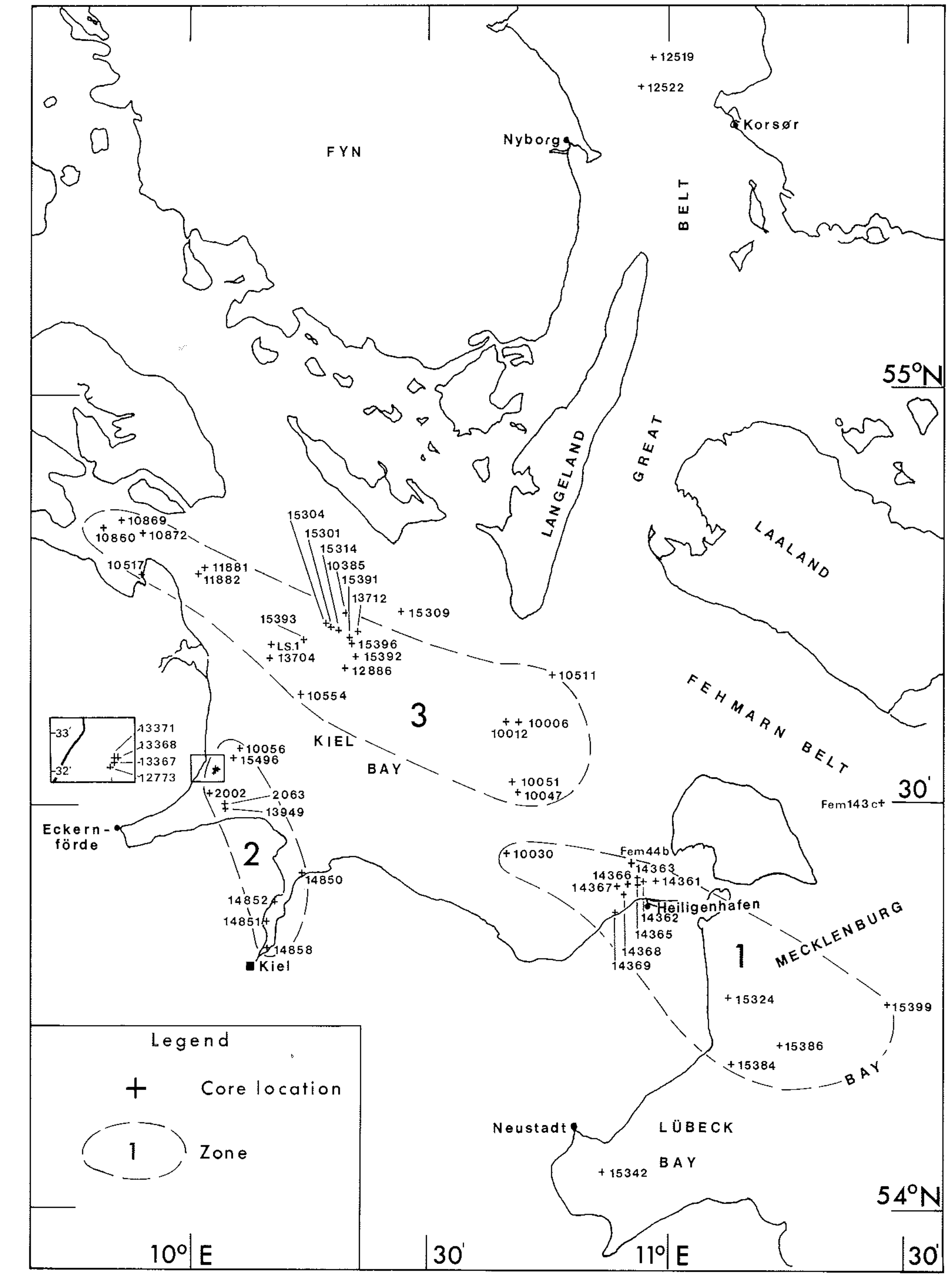

Figure 1 Map of the Western Baltic showing core locations.
Cores within zones are evaluated together.
Zone 1: Heiligenhafen and western Mecklenburg Bay.
Zone 2: Southern part of Kiel Bay.
Zone 3: Middle part of Kiel Bay.

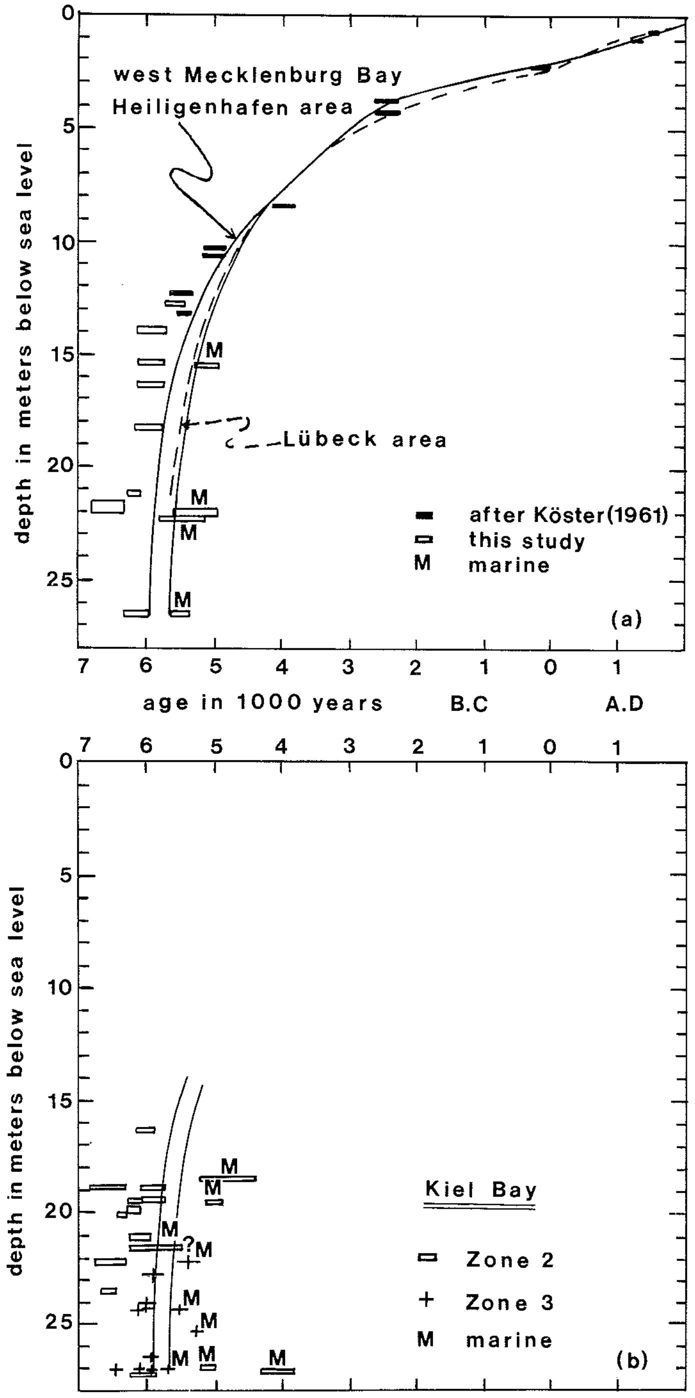

Figure 4 Relative sea level rise curves from the Western Baltic
(a) Heiligenhafen area and western Mecklenburg Bay.
continuous lines: this study,
dashed lines: Lübeck area after KÖSTER (1961).
(b) Middle and southern Kiel Bay.

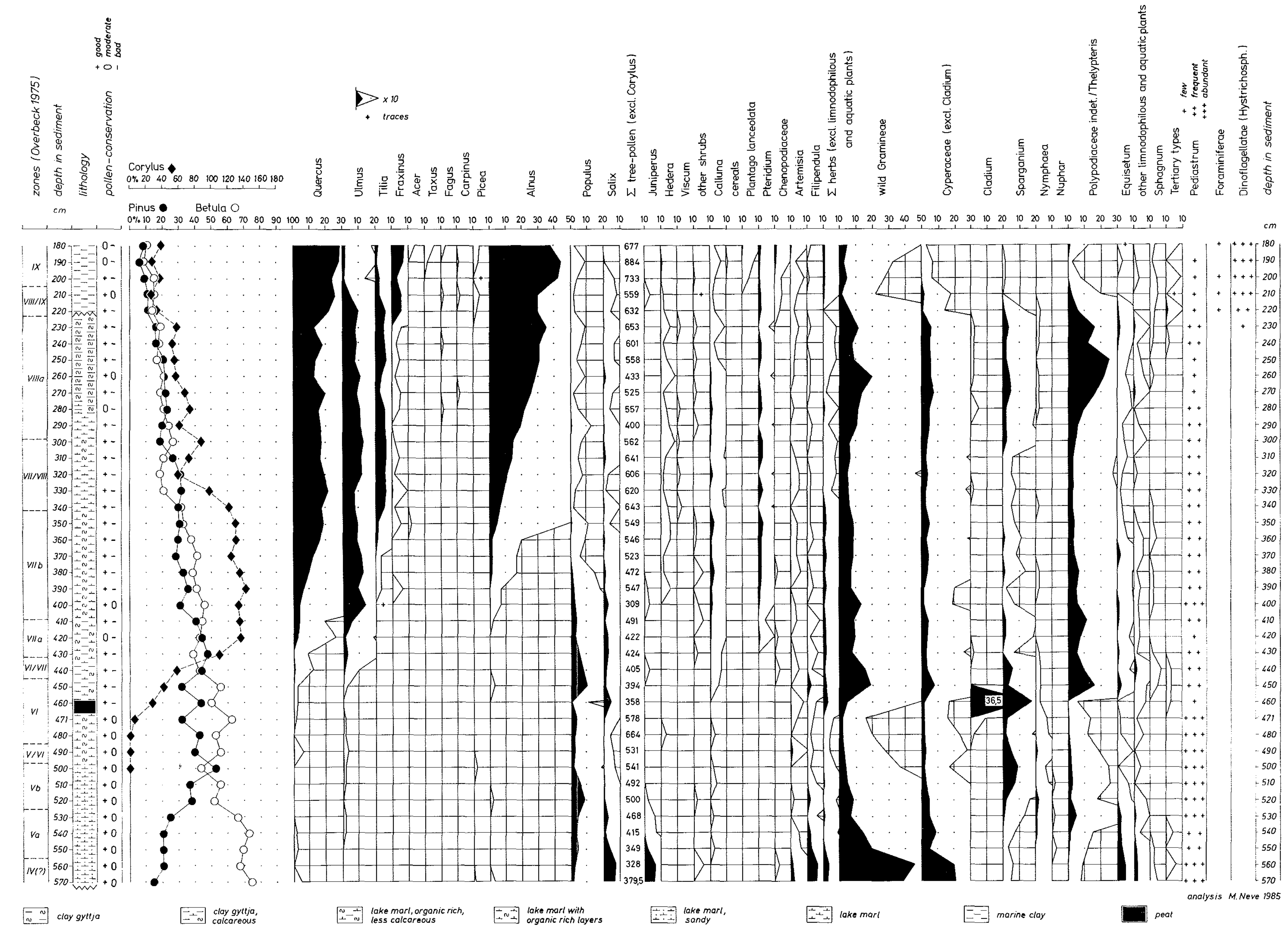

Figure 2 Pollen diagram, core no. 15 496.

expanded to include freshwater to marine sediment cores having a typical Western Baltic sequence.

The pollen assemblages are controlled by the prevailing wind systems, the different vegetational suites of the neighbouring land areas and their distance from the core sites, by the variable durability (preservation grades) of the different pollen types and by the local facies. When pollen analyses were conducted parallel with the ^{14}C datings, both studies served either in their mutual verification or in minimizing inconsistencies. For instance, replicate analyses on core 15 399 showed that the correct pollen zone for the marine clays is Zone VIII and not Zone IX as reported in WINN et al. (1983) (see also tab. 4).

The Littorina Transgression could be dated in 15 out of a total of 41 analysed cores (tab. 4). For the absolute chronology of the pollen stratigraphy, stratigraphical correlations have been made by KÖSTER (1961), EXON (1972), and by WILLKOMM (1976). The analysed transgressive horizons fall in the Atlantic Age (Zone VIII of OVERBECK, 1975), with the underlying freshwater-lacustrine deposits reaching well into the lower part of this zone, mainly as a result of the partially threshold controlled, step-wise incursions of the sea into the Western Baltic (WINN and AVERDIECK 1984). Twenty-three of the analysed cores reach down into the Late Boreal, with a few of them touching the Younger Dryas.

The core descriptions given in tab. 4 were taken from the references cited in the remarks column. The sediment sequence of core no. 13 712 was contributed by W. Richter, Kiel (personal communication).

The pollen assemblages in core 15 496 (fig. 2) are representative for the area. The shallow water lake marl of the upper part of the Younger Dryas have high *Juniper* values, signifying a warmer climate. The *Cladium* peat with hazel nuts suggests subaerial exposure during the early Boreal. *Cladium* has rarely been found forming a peat in this area. Paludification may however occur at different horizons in different areas and depends solely upon the local palaeogeography and vegetation. The lake marls and clay gyttjas were probably deposited very rapidly during the early Atlantic (Zone VIIIa) as evidenced by the slow increase in the *Alnus* values coupled with low *Fraxinus* occurrences (fig. 2). The Littorina Transgression is only recorded in the late Atlantic and is marked by the appearance of foraminiferal masks, dinoflagellate cysts and *Chenopodiaceae*. A hiatus is evidenced by the mixed sediment layer at 220 to 223 cm with marine and freshwater indicators. In some of the other Kiel Bay cores, e.g., 10 872 (AVERDIECK 1972), 15 304 (WINN et al. 1983) and 10 056 (tab. 4), a similar hiatus is also evident palynologically, and has lasted at least a few hundred years if not more. These cores are located in the channels or on their flanks, where reduced sedimentation due to bottom currents has been observed (WERNER et al., in press). A decrease in the occurrence of *Ulmus* and *Tilia* along with the first appearance of *Fagus* and cultivation indicators is typical of the Subboreal in the upper part of the core.

10 150 B. P.–
Radiocarbon Dating...

Detailed radiocarbon datings have been carried out on a large number of cores from the Western Baltic (WILLKOMM and ERLENKEUSER 1969, ERLENKEUSER and WILLKOMM 1971; 1973; 1975; WEFER et al. 1978; WINN et al. 1983; NIEDERMEIER-LANGE 1985; SIMANOWSKY 1985). Of these cores, the ^{14}C data which bear the greatest significance as to the age of the transgression horizons are compiled in tab. 3.

The samples consisted of freshwater lake marls, clay gyttjas, peats, bivalve shells and marine clays. With one exception, the samples were dated on the (total) organic fraction.

^{14}C-age calculations are based on the Libby halflife of 5 568 years for ^{14}C, and refer to 95 % of the NBS oxalic acid ^{14}C-standard activity (STUIVER and POLACH 1977). It was not considered necessary to correct the radiocarbon dates of the organic fraction for isotope fractionation effects, as the δ^{13}C-figures of these samples closely meet – within a few $^0/_{00}$ – the δ^{13}C-reference value of – 25 $^0/_{00}$ for this normalisation procedure (STUIVER and POLACH 1977), so that the corrections are of minor importance. This does not hold for the carbonate date, which has therefore been δ^{13}C-corrected in order to attain comparability with the other dates in tab. 3.

Corrections for secular variations of the past ^{14}C-level in the atmosphere (KLEIN et al. 1982) were not applied, in order to maintain consistency with published data. According to tree ring studies (KLEIN et al. 1982), these secular variations have led to radiocarbon dates which for terrestrial samples are too young by about 800 years (as compared to the true age) in the time range considered here. This might also hold for the early Holocene (STUIVER 1978). The effect on sediment samples from the Western Baltic environments probably is of similar magnitude, but cannot be estimated with confidence due to the complexity of the ^{14}C-pathway between the atmosphere and the sediments.

The ^{14}C-dates of samples from aquatic sedimentary environments are likely to be more or less older than the ^{14}C-ages of contemporaneous terrestrial samples. This is due to reworking and transport processes which intermix particulate matter of different ages (ERLENKEUSER 1979), and to the ^{14}C-deficiency ("reservoir effect") which is generally observed with various magnitudes in lake and ocean water as compared to the atmosphere (WILLKOMM and ERLENKEUSER 1972). In detail, the magnitude of ^{14}C-ages shift depends upon the particular environmental and depositional regime which controls the availability and deposition of the particulate matter. Hence, it is also related to the type of sediment fraction dated (e. g., WEFER et al. 1978). For freshwater deposits, it further depends upon the hydrological conditions in the lake, particularly on the effectiveness of CO_2 exchange between the water and atmosphere (WEFER et al. 1978; ERLENKEUSER and WILLKOMM 1979). The majority of the present samples are peats or stem from very shallow water deposits close to land, and accordingly are likely to contain appreciable amounts of terrestrial plant debris. At least the reservoir effect appears less critical for such samples (WINN et al. 1983). We therefore estimate that the ^{14}C-dates are not higher by more than a few hundred years as compared to the ^{14}C-age of contemporaneous land plants.

The ^{14}C-age offset may be more severe in the marine sediments of Kiel Bay. For instance, studies on the organic carbon of recent sediments yielded (natural) sediment surface ^{14}C-ages of 650 to 850 years (ERLENKEUSER et al. 1974). In some cases higher deviations may occur, particularly if the sediments have been deposited close to subsea exposures of much older peat layers. Consequently, the marine ^{14}C-dates should be applied with caution.

Discussion

The global eustatic rise of sea level in the Late Pleistocene and Holocene has already been studied in detail in the North European region, and a number of eustatic as well as relative sea level rise curves (fig. 3) are available (KÖSTER 1961; MÖRNER

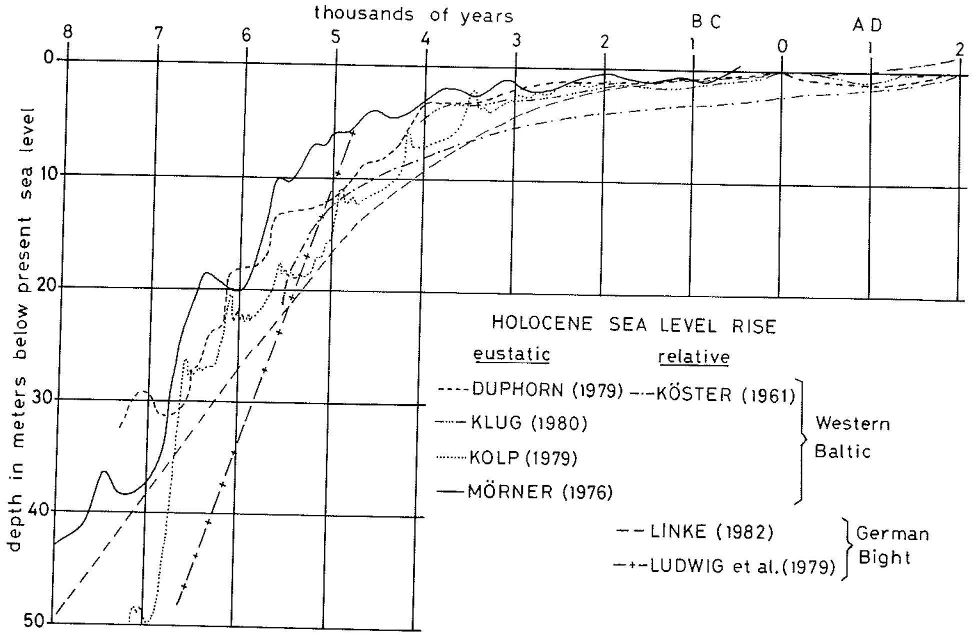

Figure 3 Selected Holocene sea level rise curves from the north European region.

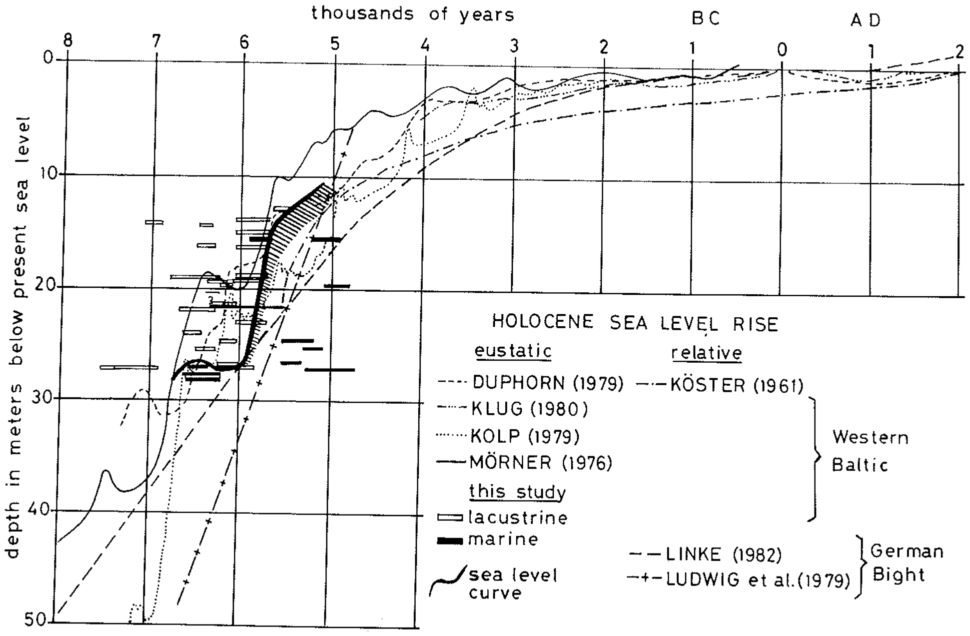

Figure 5 Sea level rise curve from the Western Baltic (this study) superimposed on the published curves.

1976; KOLP 1976; LUDWIG et al. 1979; DUPHORN 1979; KLUG 1980; LINKE 1982 among others). JELGERSMA (1966) has aptly discussed the problems of selecting models assuming either an oscillating sea level, a steadily rising level followed by a stagnant phase or a continuously rising sea level. BLOOM (1967) pointed out the difficulties of separating the isostatic and eustatic components and accounting for the mechanisms of the crustal adjustments involved.

Most of our ^{14}C dated horizons are restricted to the sediments either overlying or underlying the Littorina Transgression. Absolute datings on the crucial horizon itself are rare (tab. 3 and 4). Moreover, the exact age of the transgression is often obscured in a hiatus, with the missing interval either due to non-deposition or later erosion of the sediments. The significance of erosional processes during the early Holocene transgressive phases is evident in the Western Baltic sequences from the shallow seismic (boomer) profiles and sometimes also from the great variety of sediment types ranging from lake marls to clay gyttja and peat immediately underlying the marine succession.

Similar limitations also apply to the reconstruction of the former threshold depths. Although we have defined the major barrier for the region, the original paleobathymetric levels could have been higher. In addition, the height of a threshold cannot simply be equated with the sea level, because the seawater requires a certain surplus of height before the hydrodynamic processes could efficiently develop to control the sedimentary facies. This is evidenced in our study area by the lacustrine marls which, while indicating a rising lake level, show progressively less saline effects the further away they were deposited from the main threshold in the Belt Sea (KOLP 1965, WINN et al. 1983).

We have not initially applied subsea terraces as indicators of sea level stagnations, although a number of studies, notably by KOLP (1976) and HEALY (1981), have attempted tentative correlation of the Western Baltics submarine terraces with sea-level stillstands. The significance of terraces in sea level interpretations clearly depends upon their mode of origin, their relationship to the water level, wind direction and strength, wave amplitudes, sediment types in the original cliffs and shorelines, and the protective effect of relict or lag sediment cover which controls the time required for terrace formation. The geomorphological development of submarine terraces thus seems to be not yet fully understood (SUNAMURA 1983). Moreover, the dating of such terraces is mainly through the overlying sediments under the assumption that terrace formation is directly followed by sediment deposition without a major hiatus. KOLP (1976) also partially inferred a short regression after terrace formation, implying that terrace formation and the lacustrine-marine transgression sequence found on the terrace belong to the same sea-level phase. This combination does not appear in the section of sea-level rise history considered here.

Peat is perhaps the only lake level indicator which does not require correction for formation depth. Some of our cores record the effects of a rising water level in the sedimentary succession – from peats to lake marls to clay gyttjas – before the marine conditions prevailed. Since lake marls and clay gyttjas are not deposited at zero water level, but rather require a certain water depth for their formation, their present depth do not exactly represent the palaeolevel of the lake or sea, but, depending upon their facies, rather represent much shallower levels.

A source of error could lie in the determination of the absolute water depth of the cores. This was measured with an echosounder where the velocity of sound in water played a major role in the calculations. Postdepositional compaction of sediments could also lead to errors in estimating the palaeolevels of the marine transgression horizons. However, since we have shallow water depths (less than 30 m) and the horizons generally lie within 1 m above a hard subsurface (mainly glacial tills), these errors are small and are estimated to be ±30 cm.

Since the Western Baltic is a large heterogeneous region with different palaeogeomorphological configurations, we have subdivided our cores into three large groups (zones 1 to 3 in fig. 1) so as to attain a semblance of homogeneity and to overcome as far as possible the effects of local tidal inequalities, large scale earth movements and variations in the geoid configuration (TOOLEY 1978). Our core data (tab. 3 and 4) have first been plotted without any correction for isostatic movements to obtain the relative rise in sea level. Data compiled by KÖSTER (1961) were incorporated for the shallower levels. The relative sea level curves are then laid between the freshwater deposits and the marine sediments (fig. 4a and 4b). The upper curve is drawn through the top of the freshwater deposits and represents the oldest possible position of the sea level curve. The lower curve represents the older limit given by the marine sediments. Due to the marine transgression occurring much later than the normally dated peats and freshwater gyttjas, the actual relative sea level curve is expected to lie closer to the lower curve in the Mecklenburg-Heiligenhafen area (fig. 4a). For the Kiel Bay, only one set of curves has been drawn for zones 2 and 3, with a better fit to zone 3 (fig. 4b). In the cores of zone 2, the hiatus between the freshwater and the marine deposits encompasses a time span of at least 800 years (tab. 3), and the erosion or nondeposition episodes could not be differentiated.

According to KÖSTER (1961, fig. 24) and KOLP (1979, fig. 10), the difference in the average rates of subsidence during the Holocene between the Lübeck and the Heiligenhafen areas is around 5 cm/100 years. Therefore, with this average subsidence rate, the relative sea level rise curves for the above two areas should show a difference of about 5 m in level at the beginning of the Holocene (10 000 years B. P.). However, our relative sea levels rise curves for the Heiligenhafen area (fig. 4a) and for the Kiel Bay, (zone 3, fig. 4b) do not essentially differ from the curve for the Lübeck Bay. Thus, the high rates of subsidence previously postulated for this area are not substantiated by our data, at least not as time-averaged values.

The identification of a series of early Holocene wave-cut terraces at practically the same levels in various parts of Kiel Bay (HEALY 1981) also supports this finding. If Kiel Bay had been a purely subsiding area with its northern part sinking much slower than its southern sector, then the submarine terraces in the north should be situated in our acoustic records at a significantly higher absolute level than those in the south. This was not found. Moreover, there is little difference today between the levels of the early Holocene marine transgression horizons in core nos. 12 519 and 12 522 from the Great Belt outside the thresholds (WINN 1974) and those in the cores from the northern and middle parts of Kiel Bay (tab. 3 and 4). Even allowing for possible corrections for the original levels of these thresholds, the depths of the transgression in these cores do not confirm the high differential rates of subsidence previously assumed for this region (KÖSTER 1961, KOLP 1979). On the contrary, local differences up to 7 m in the levels of the transgression horizons observed on the profiles between core nos. 10 030, 10 047 and 10 051, and between 13 949, 2 063 and 15 496 (fig. 1, tab. 3 and 4) may, in addition to palaeomorphological influences, also reflect local upheaval and subsidence. VOSS (1972) and PRANGE (1985) also observed such local adjustments during the late Holocene along the west coast of Kiel Bay, which could be attributed to salt tectonics.

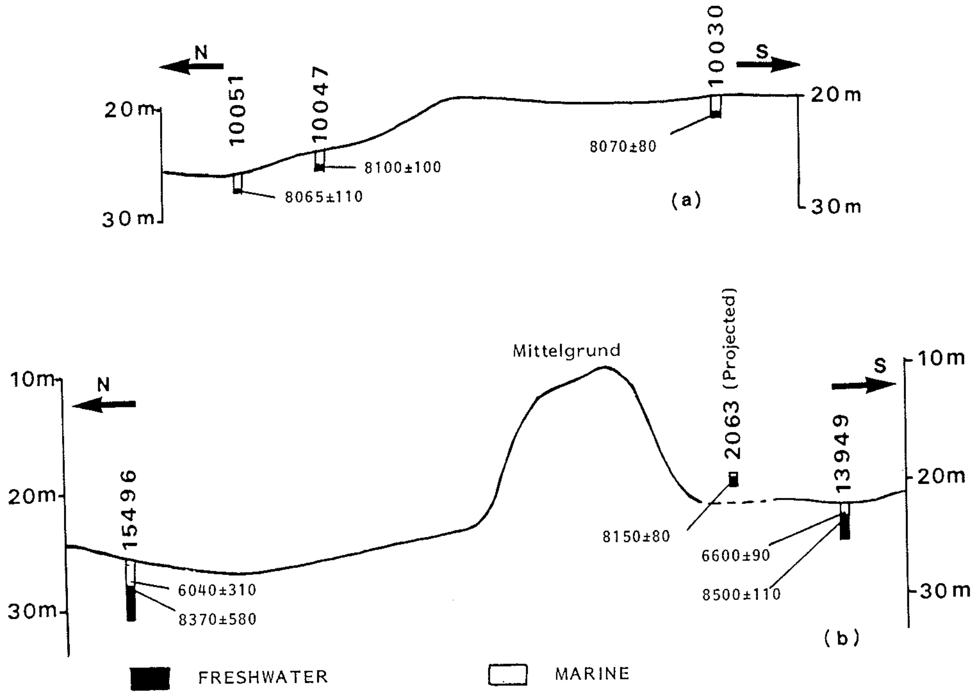

Figure 6 Profiles illustrating local differences in the levels of the Littorina Transgression.
(a) cores 10 030, 10 047 and 10 051, horizontal scale 1 : 50 000
(b) cores 13 949, 2 063 and 15 496, horizontal scale 1 : 25 000

We have therefore corrected the dephts of the marine transgression horizons in our cores only for the threshold effect. The test of these data against the various previously published sea level curves is depicted in figure 5. It shows that the rise in sea level was very rapid between - 26 m and - 14 m and slowed down somewhat thereafter. Between - 29 m and - 26 m, if the ^{14}C ages on the marine sediments reflect true ages, a stillstand or even a slight regression is possible. This phenomenon is partially obscured in our area due to the effect of the thresholds, which may induce a unique offset between the water levels on different sides of this barrier. A second effect which could be important are the older than „true" ^{14}C ages generally observed in marine sediments. If the Great Belt cores are also 650 to 850 years too old, then the stagnation will no longer be supported by our data. It appears noteworthy that LUDWIG et al.(1979) and LINKE (1982) have also presented continuously rising relative sea level curves without any stagnations from data in the German Bight and adjacent coastal areas.

It is clearly evident, that the submarine terraces at - 30 m and perhaps also those at - 27 m in the Kiel Bay are not caused by the eustatic rise in sea level, but are formed during the lacustrine phase corresponding to the Ancylus Lake stage of the main Baltic Sea preceding the marine incursions. Our data also do not support the many stagnations in the sea level curve af either KOLP (1965) or DUPHORN (1979) which obviously were inferred from submarine terraces. We consider that the resolution of the available data is not precise enough to permit these contructions in our study area.

Conclusions

New core data have confirmed that the eustatic rise in sea level above the thresholds in the Great Belt did not at once lead to marine conditions throughout the Western Baltic. The sediments close to this event indicate an initial rise in water level due to the back-up of dominantly fresh water. Therefore, the peats in the early Holocene sedimentary succession are not direct indicators of sea level, but mark the oldest possible age of the marine transgression. The main marine Littorina Transgression occurred only after the sea level rose considerably above the crucial thresholds in the north. All depths of the marine transgression recorded in the sediment succession below this level therefore have to be adjusted to it.

Regional differential subsidence rates previously postulated for the Western Baltic (KÖSTER 1961; KOLP 1979) could not be substantiated by our data. This may partially be due to the effects of salt tectonism. Local supsidence and upheaval are possible, especially in the southern and western parts of the Kiel Bay. Assuming that the marine ^{14}C dates do not deviate much from the true age, a sea-level stillstand around - 28 m is apparent from our core data. Thereafter, the sea level rose very rapidly before slowing down around - 14 m.

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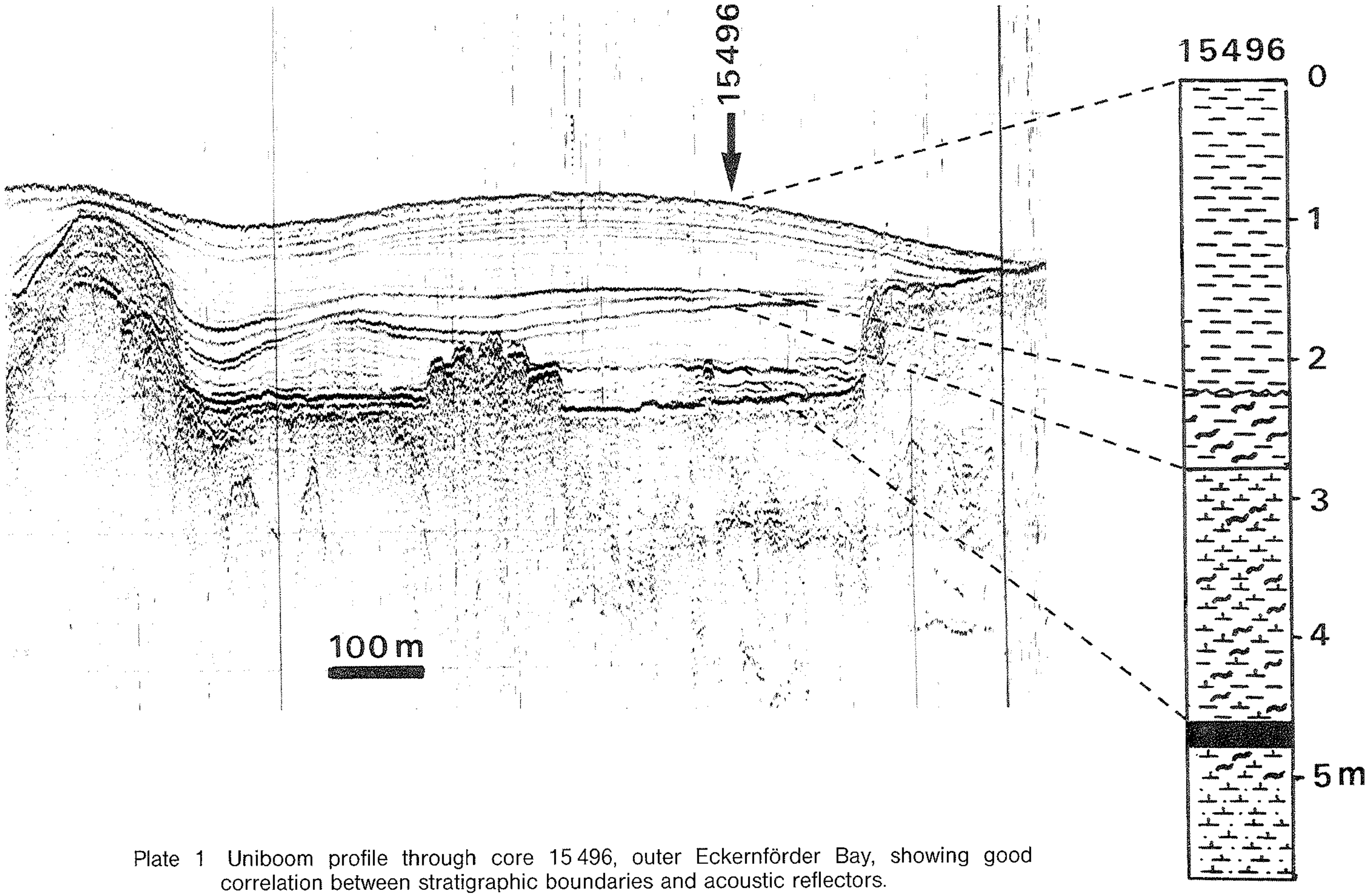

Plate 1 Uniboom profile through core 15 496, outer Eckernförder Bay, showing good correlation between stratigraphic boundaries and acoustic reflectors.
For sediment symbols see legend in fig. 2.

Table 1

Location of cores

Core no.	Latitude	Longitude	Water depth (m)
LS 1	54°41.8′N	10° 9.5′E	31.0
Fem 44b	54°25.4′E	10°55.5′E	12.5
Fem E143c	54°30.0′N	11°27.2′E	29.0
334	54° 5.4′N	14° 0.5′E	48.0
2002	54°31.0′N	10° 2.2′E	24.5
2063	54°30.2′N	10° 4.2′E	19.0
10006	54°36.1′N	10°41.5′E	28.0
10012	54°36.1′N	10°40.0′E	24.0
10030	54°26.2′N	10°39.9′E	20.0
10047	54°30.9′N	10°41.1′E	24.0
10051	54°31.7′N	10°40.7′E	26.0
10056	54°34.2′N	10° 6.2′E	30.0
10385	54°44.1′N	10°19.7′E	23.5
10511	54°39.7′N	10°45.6′E	26.5
10517	54°46.9′N	9°54.1′E	0.3
10554	54°38.0′N	10°13.9′E	22.0
10860	54°50.4′N	9°48.9′E	27.5
10869	54°51.0′N	9°51.2′E	21.0
10872	54°50.1′N	9°53.9′E	26.5
11881	54°47.4′N	10° 1.7′E	32.5
11882	54°47.0′N	10° 0.9′E	29.0
12519	55°24.7′N	10°58.4′E	25.5
12522	55°22.6′N	10°56.8′E	27.0
12773	54°32.3′N	10° 2.9′E	18.0
12886	54°40.2′N	10°19.5′E	22.7
13367	54°32.4′N	10° 3.1′E	18.0
13368	54°32.5′N	10° 3.2′E	21.0
13371	54°32.5′N	10° 3.1′E	19.0
13704	54°40.8′N	10°10.0′E	28.0
13712	54°42.9′N	10°21.1′E	28.5
13949	54°29.8′N	10° 4.3′E	22.0
14361	54°24.2′N	10°58.7′E	13.0
14362	54°24.2′N	10°57.3′E	11.0
14363	54°24.5′N	10°56.6′E	13.0
14365	54°24.1′N	10°56.4′E	13.0
14366	54°24.1′N	10°55.2′E	15.0
14367	54°23.8′N	10°53.8′E	15.5
14368	54°23.1′N	10°54.8′E	12.0
14369	54°22.0′N	10°53.7′E	15.0
14849	54°47.0′N	15° 6.5′E	61.0
14850	54°27.3′N	10°13.8′E	18.0
14851	54°21.3′N	10° 9.3′E	12.0
14852	54°22.7′N	10°10.2′E	16.0
14858	54°19.3′N	10° 8.9′E	16.0
15301	54°43.0′N	10°17.8′E	22.0
15304	54°43.3′N	10°17.5′E	21.0
15309	54°44.3′N	10°26.6′E	12.0
15314	54°43.0′N	10°18.7′E	26.8
15324	54°15.6′N	11° 7.6′E	16.5
15342	54° 2.7′N	10°52.0′E	23.0
15384	54°10.9′N	11° 8.1′E	20.0
15386	54°12.2′N	11°14.3′E	21.0
15391	54°41.9′N	10°20.0′E	32.0
15392	54°40.9′N	10°20.9′E	24.0
15393	54°42.1′N	10°14.5′E	25.0
15396	54°41.9′N	10°20.4′E	32.2
15399	54°15.1′N	11°16.4′E	24.0
15496	54°33.4′N	10° 5.4′E	25.6

Table 2

Core No.15496 25.2m water depth

Depth (cm)	bulk density wet (grams/ccm)	bulk density dry (grams/ccm)	Water content (%dry weight)	C_{org} (weight%)	$CaCO_3$ (weight%)
1				6.32	0.16
10	0.261	1.112	326	6.77	nm
20				4.81	nm
30	0.323	1.179	262	4.94	nm
50	0.455	1.263	178	3.58	nm
60	0.500	1.299	160		
70	0.442	1.230	178	4.58	nm
80	0.428	1.241	190		
90	0.459	1.232	168	3.58	nm
100	0.491	1.281	161		
110	0.490	1.288	163	3.84	nm
120	0.472	1.276	170		
130	0.493	1.280	160	1.50	nm
140	0.495	1.288	160		
150	0.474	1.257	165	3.96	nm
160	0.476	1.279	169		
170	0.517	1.268	145	3.41	nm
180	0.475	1.251	163		
200	0.538	1.312	143	3.76	nm
210				3.81	nm
220	0.638	1.328	116	3.12	nm
230				9.01	nm
235				7.09	13.63
240	0.442	1.204	181	5.44	29.50
245				7.48	30.75
250				8.85	30.79
255				10.60	31.21
260	0.397	1.206	204	12.06	36.50
270	0.371	1.170	215	10.61	51.13
275				9.00	57.17
280	0.467	1.236	165	6.40	71.42
285				4.35	79.21
290				4.84	76.63
295				5.03	75.83
300	0.485	1.253	158	4.65	78.88
310				5.06	78.25
320	0.536	1.297	142	4.38	77.79
330				3.93	76.58
340	0.553	1.309	137a		
350				4.21	74.54
360	0.517	1.265	145	4.17	75.50
370	0.509	1.262	148	4.23	70.38
380	0.496	1.269	156	4.66	69.96
390	0.533	1.326	152	4.69	72.00
400	0.539	1.347	150	4.52	70.17
410				6.47	53.79
420	0.422	1.224	190	9.55	28.17
430				9.92	32.54
435				10.64	11.21
440	0.463	1.297	180	11.18	0.75
445				10.15	10.05
450				14.13	16.67
455				16.68	16.96
460	0.242	1.148	374	49.80	nm
465				45.67	3.25
470	0.371	1.219	229	6.15	82.96
475				8.13	76.58
480	0.381	1.203	216	4.95	84.00
490	0.360	1.194	232	5.49	80.29
500	0.377	1.200	218	3.31	84.46
510				4.41	78.66
520	0.360	1.194	232	3.47	83.21
530				3.77	82.79
540	0.494	1.275	158	3.90	78.25
560	0.633	1.354	114		

nm = not measureable

Table 3

Selected ^{14}C ages of sediment cores in the Western Baltic.

Core no.	Sediment	^{14}C date[2]	Depth(m)	Reference no.[1]
LS 1	sedge peat	9430± 85	31.3	KI-207
LS 1	sedge peat	9780±100	31.5	-208
2063	sedge peat	8150± 80	19.8	-209
2063	sedge peat	8250±130	20.0	-210
Fem44b	sedge peat	7550±140	12.7	-214
10030	clayey peat	8070± 80	21.1	-375
10030	peat	8340±140	21.2	-370
10047	peat	8350±120	25.3	-369
10047	peat	8100±100	25.4	-374
10051	peat pocket in marine clay	8065±110	27.3	-380[3]
10056	sedge peat	11880±100	31.1	-381
10511	marine gyttja	8410±110	28.8	-848.03
11881	marine clay	7060±90	34.8	-619.16
11881	lake marls	10120±120	35.1	-619.18
11882	clay gyttja	8110±100	30.8	-621.094
12519	marine clay	8340±190	28.3	-738.33
12522	marine clay	8370±220	27.4	-736.08
12522	lake marls	8340±200	27.7	-736.14
12773	marine sand	6720±450	18.6	-846.04
12773	lake sand	7870±165	19.3	-846.10
12886	sandy clay	7170±100	25.3	-1106.113
13367	lake sand	8490±300	18.8	-1194.01
13368	marine sand	8540±440	21.4	-1195.01
13368	lake sand	10470±470	21.8	-1195.02
13371	lake sand	8180±140	19.4	-1196.01
13704	lake marls	7880±105	30.3	-1315.04
13949	marine clay	6600± 90	23.1	-1330.03
13949	lake marls	8500±110	23.6	-1330.04
14361	clayey peat	7440±125	13.8	-1706.01
14363	wood peat	8370± 90	14.0	-1707.01
14365	sedge peat	8950±105	14.0	-1708.01
14366	coquina with Cerastoderma edule	7010±160	15.5	-1709.011
14366	marine clay	7770±140	15.5	-1709.01
14851	clay gyttja	7960±120	16.3	-1779.01
14852-1	clay gyttja	8290± 75	20.1	-2311.01
14852-2	marine clay	6910± 85	19.5	-2162.10
15301	clay gyttja	10810±160	24.1	-1837
15399	marine clay	7390±115	26.6	-2002.01
15399	clay gyttja	8030±175	26.7	-2002.02
15496	marine clay	6040±310	27.6	GPI-84/57
15496	clay gyttja	8370±580	28.3	-84/60
15496	Cladium peat	9550±660	30.2	-84/61
15496	lake marl	9330±440	31.1	-84/59

1) University of Kiel Radiocarbon Laboratory, Kiel. except for core 15496 (Geol.-Pal.Institute).

2) except for KI-1709.011, all other ^{14}C-dates were obtained on the organic fraction.

3) revised date.

^{14}C ages on marine sediments are probably 650-850 years too old. See also WEFER et al. 1978.

All depths given are below present mean sea level.

Table 4

Palynological results of Baltic Sea cores

Core no.	Sediment	depth below sea level(m)	Pollen Zone	Remarks
LS 1	sedge peat	31.30-31.50	VI	this study
E143c	lake marl	30.00	III or V	this study
334	marine clay	48.00-48.50	VII or VIII	LUTZE 1965
	marine clay	48.50-48.60	VI	
	clay	48.60-49.50	V	
2002	reworked peat	24.98-25.00	XI	this study
	sandy gyttja	25.37-25.87	XI	
10006	lake marl	28.79-28.95	VII/VIII	this study
	lake marl	28.95-29.14	VII	
	clay gyttja	29.14-29.40	VII	
	clay gyttja	29.40-29.58	VI	
10012	marine clay	24.00-24.82	VIII-XI	this study
	peaty gyttja	24.82-24.87	VIIIa	
	sedge peat	24.87-24.97	VIIIa	
	clay gyttja	24.97-25.13	VIIIa	
	lake marl	25.13-25.78	VII	
10056	marine clay	30.94	IX	this study
	sedge peat	30.94-31.05	VII	
	sedge peat	31.05-31.15	VI	
	clayey peat	31.15-31.35	IV	
	sedge peat	31.35-31.50	III	
10385	lake marl	23.50	VII	this study
10517	peat	0.50- 0.60	XII	EXON 1972
10554	marine clay	24.20-24.30	VIIIa	this study
	clay gyttja	24.41-24.60	VIIIa	
	clayey peat	24.58-24.87	VII	
	lake marl	24.87-24.94	VII	
10860	peat	30.30-30.35	VI	EXON 1972
10869	peaty gyttja	22.00-22.05	VIIa	EXON 1972
10872	marine clay	26.50-28.95	VIII-XI	AVERDIECK 1972
	lake clay	28.95-29.13	VIII	
	peat	29.13-29.20	IV	
13712	clay gyttja	29.20	VIIIa	this study
	sedge peat	29.65	VI	
14361	clay gyttja	13.80-13.96	VIIIa	NIEDERMEIER 1985
	peaty gyttja	13.96-14.22	VIIIa	
14362	Cladium peat	12.58-12.64	III or V	NIEDERMEIER 1985
	clay gyttja	12.64-12.71		
	lake marl	12.71-12.96	III or V	
14363	fern-reed peat	13.94-14.03	VII/VIII	NIEDERMEIER 1985
	sedge peat	14.03-14.38	VII	
	reed peat	14.38-14.72	V	
	peaty gyttja	14.72-14.83	IV/V	
14365	fern-reed peat	13.91-14.10	VII	NIEDERMEIER 1985
	fern-reed peat	14.10-14.20	VI/VII	
	reed peat	14.20-14.40	V	
	Cladium peat	14.40-14.46	V	
	detrital peat	14.46-14.58	V	
	lake marl	14.58-14.68	V	
	lake marl	14.68-14.73	IV	
	clay gyttja	14.73-14.93	IV	
	lake marl	14.93-15.27	III	
	lake marl	15.27-15.87	II or III	
14366	detrital gyttja	15.56-15.69	VIIIa	NIEDERMEIER 1985
	clay gyttja	15.69-15.82	IV	
14367	detrital gyttja	16.33-16.42	VIII	NIEDERMEIER 1985
	clay gyttja	16.42-16.50	VII or VIII	
14368	silty peat	13.70-13.91	VIIIa	NIEDERMEIER 1985
	clay gyttja	13.91-14.12	VIIIa	
	sedge peat	14.12-14.15	VIII	
	sedge peat	14.35	VI	
14369	palaeosol	15.34-15.38	VIII	NIEDERMEIER 1985
14849	humus clay	63.12-63.24	XI ?	this study
14850	reed peat	21.30-21.45	VII	SIMANOWSKY 1985
	lake clay	21.45-21.60	VII	
	swamp peat	21.60-21.80	IV	
	sedge peat	21.80-22.20	IV	
14852	clay gyttja	19.81-21.00	V	SIMANOWSKY 1985
	clay gyttja	21.00-21.06	IV	
14858	sedge peat	18.83-18.87	VIII	SIMANOWSKY 1985
	reed peat	18.87-19.00	VI	
	fern-reed peat	21.00-21.16	VI	
	lake clay	21.16-22.96	IV	
15301	lake marl	22.69-23.34	VII	WINN et al. 1982
	clay gyttja	23.84-24.12	IV or III ?	
15304	marine sand	21.78-22.42	VIIIb	WINN et al. 1982
	clay gyttja	22.42-22.63	VIIIa	
	peat	22.63-22.82	VIIIa	
	lake marl	22.82-23.84	VII	
15309	silty sand	12.70-14.80	IV ?	Late Glacial ?
15314	clay gyttja	28.00-28.38	VI	WINN et al. 1982
	reed-fern peat	28.38-28.42	V	disturbed
	peaty gyttja	28.42-28.80	VI	
15324	reed-fern peat	18.10-18.25	VIIIa	WINN et al. 1983
	reed-fern peat	18.25-18.50	VI-VII	
15342	marine clay	23.00-26.40	VIIIa-XII	WINN et al. 1983
	clay gyttja	26.40-26.85	VIIIa	
	lake marl	26.85-27.05	VII-VIIIa	
	lake marl	27.05-27.15	IV	
	clay gyttja	27.15-27.30	IV	
15384	marine sand	20.00-21.15	VIII/IX	WINN et al. 1983
	marine sand	21.15-22.20	VIIIa	
	sand & clay	22.20-22.60	VIIIa	
15386	lake marl	21.50-22.17	VII	this study
15391	lake marl	32.49-32.55	VII	WINN et al. 1982
	lake marl	32.70-32.97	V	
	clay & sand	32.97-33.85	IV	
15392	marine clay	24.00-26.46	VIIIa-XII	WINN & AVERDIECK 1984
	clay gyttja	26.46-26.65	VIIIa	
15393	laminated clay	27.21-27.51	VIII	WINN et al. 1982
	laminated clay	27.51-27.61	V	
15396	detrital clay	33.58-35.16	IV ?	reworked forms
15399	marine clay	26.55	VIII	WINN et al. 1983
	lake marl	26.67-26.85	VIIIa	
	lake marl	26.85-27.10	VII/VIII	
	clay gyttja	27.10-27.24	VI	
	swamp peat	27.24-27.29	VI	
15496	marine clay	27.40-27.83	IX	this study
	clay gyttja	27.83-28.40	VIIIa	
	lake marl	28.40-28.59	VIIIa	
	lake marl	28.59-28.72	VIIIa/VIIb	
	lake marl	28.72-29.69	VIIb	
	clay gyttja	29.69-30.18	VI	
	Cladium peat	30.18-30.26	VI	
	lake marl	30.26-30.57	VI	
	lake marl	30.57-31.17	V	
	lake marl	31.17-31.33	IV	

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