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RELATIVE SEA LEVEL CHANGES, GLACIO-ISOSTATIC REBOUND AND SHORELINE DISPLACEMENT IN THE SOUTHERN BALTIC

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Abstract. The relative sea level curve was developed for the southern Baltic area, based on a set of 314 radiocarbon datings of different terrestrial and marine sediments, collected at 163 sites located in the Polish part of the Southern Baltic and in the adjacent coastal land area. When developing the curve, relicts of various formations related to the shoreline evolution as well as extents of erosional surfaces, determined from seismoacoustic profiles, were taken into account.

During Late Pleistocene and Early Holocene, i.e. between 13.0 and 8.5 ka BP, the southern Baltic sea level rose and fell three times, the amplitude of changes extending over 25–27 m. In some extreme cases, the sea level was falling at a rate of up to about 100–300 mm/a, the rate of rise accelerating to about 35–45 mm/a. In the Late Boreal, *c*. 8.5 ka BP, the Baltic — its water level by about 28 m lower than the present one — became permanently connected with the ocean. Until the onset of the Atlantic, the sea level had risen to about 21 m below the present sea level (b.s.l.). During 8.0–7.0 ka BP, the sea level was rising, at a rate of about 11 mm/a, to reach 10 m b.s.l. Subsequently during the Atlantic, until its end, the sea level rose to 2.5 m b.s.l., the rate of rise slowing down to about 2.5 mm/a. During the first millenium of the Subboreal, the sea level rose to about 1.3–1.1 m b.s.l., to become — on termination of the Subboreal — about 0.6–0.7 m lower than present. During the Subatlantic, the sea level changes were slight only.

The glacio-isostatic rebound began *c*. 17.5 ka BP, to terminate *c*. 9.2–9.0 ka BP. The total uplift during that time amounted to about 120 m. The maximum uplift rate of about 45 mm/a occurred *c*. 12.4–12.2 ka BP. Within the period of *c*. 9.0 to *c*. 7.0 ka BP, the southern Baltic experienced forebulge migration, a subsequent subsidence ensuing from *c*. 7.0 to *c*. 4.0 ka BP. As from *c*. 4.0 ka BP, the Earth crust in the area regained its equilibrium.

In Late Pleistocene and Early Holocene, the southern Baltic shoreline displaced rapidly and substantially several times, the displacement rate ranging from several tens of metres to a few kilometres per year. The displacement processes involved the seafloor surfaces located at present at 25 to 55 m b.s.l., the shoreline migrating over distances of 30–60 km away from the present coastline. In Middle Holocene, the shoreline moved southwards over a distance ranging from about 60 km in the Pomeranian Bay to about 5 km in the Gulf of Gdańsk. The shoreline location approached the present one at the final phase of the Atlantic. Late Holocene was the period when coast levelling processes were prevailing, the shoreline becoming gradually closer and closer to its present setting.

Key words: sea level changes, shoreline evolution, glacio-isostasy, Late Pleistocene, Holocene, Southern Baltic.

Abstrakt. Krzywą względnych zmian poziomu morza skonstruowano na podstawie 314 dat radiowęglowych osadów pochodzących z różnych środowisk lądowych i morskich. Próbki do datowań pobrano z 163 stanowisk zlokalizowanych na obszarze polskiej części południowego Bałtyku i przyległej strefy brzegowej. Przy konstruowaniu krzywej wykorzystano również relikty różnych form związanych z rozwojem strefy brzegowej oraz zasięgi powierzchni erozyjnych, zlokalizowane na profilach sejsmoakustycznych.

W późnym plejstocenie i wczesnym holocenie, między 13,0 i 8,5 tys. lat BP, poziom wody trzykrotnie wzrastał i opadał, a zakres wahań dochodził do 25–27 m. Poziom wody obniżał się w skrajnych przypadkach w tempie do ok. 100–300 mm/rok, a tempo wzrostu dochodziło do ok. 35–45 mm/rok. W późnym boreale, ok. 8,5 tys. lat BP, Bałtyk uzyskał stałe połączenie z oceanem na poziomie niższym od obecnego o ok. 28 m. Do początku okresu atlantyckiego poziom morza wzrósł do ok. 21 m poniżej współczesnego poziomu morza (p.p.m.). W okresie 8,0–7,0 tys. lat BP poziom morza wzrósł do 10 m p.p.m., w średnim tempie

ok. 10 mm/rok. Do końca okresu atlantyckiego poziom morza wzrósł do 2,5 m p.p.m., a tempo wzrostu zmalało do ok. 2,5 mm/rok. W pierwszym tysiącleciu okresu subborealnego poziom wody wzrósł do ok. 1,1–1,3 m, a do końca tego okresu do ok. 0,6–0,7 m niższego niż współczesny. W okresie subatlantyckim średni poziom morza zmienił się już nieznacznie.
Przebudowa glaciizostatyczna rozpoczęła się ok. 17,5 tys. lat BP i zakończyła ok. 9,2–9,0 tys. lat BP. Całkowity zakres podniesienia (*total uplift*) w tym okresie wyniósł ok. 120 m. Maksimum prędkości ruchów wznoszących, dochodzące od ok. 45 mm/rok, wystąpiło w okresie ok. 12,4–12,2 tys. lat BP. W okresie od ok. 9,0 do ok. 7,0 tys. lat BP przez obszar południowego Bałtyku migrowało nabrzmienie brzeżne, a w okresie od ok. 7,0 do ok. 4,0 tys. lat BP wystąpiły ruchy obniżajace. Od ok. 4,0 tys. lat BP położenie skorupy ziemskiej wróciło do stanu równowagi.
Linia brzegowa południowego Bałtyku w późnym plejstocenie i wczesnym holocenie kilkukrotnie uległa szybkim i znacznym przemieszczeniom. Zmieniła położenie w tempie od kilkudziesięciu metrów do kilku kilometrów rocznie. Procesy te rozgrywały się na powierzchni dna morskiego położonej obecnie na głębokości od ok. 55 do 25 m p.p.m. i w odległości 30–60 km od dzisiejszego wybrzeża. W środkowym holocenie linia brzegowa przemieściła się ku południowi od ok. 60 km w Zatoce Pomorskiej do ok. 5 km w Zatoce Gdańskiej. Położenie linii brzegowej zbliżyło się do współczesnego w końcu okresu atlantyckiego. W późnym holocenie dominowały procesy wyrównywania wybrzeży, a linia brzegowa stopniowo zbliżała się do obecnego położenia.

Słowa kluczowe: zmiany poziomu morza, rozwój wybrzeży, glaciizostazja, późny plejstocen, holocen, południowy Bałtyk.

INTRODUCTION

How did the Baltic Sea emerge? This seemingly straightforward, although ambiguous, question has been haunting scientists for a long time. The question touches an array of many different issues, from the problem of origins of a crustal depression in the area covered by the Baltic at present, to the history of Scandinavian ice sheet expansion and final retreat, to reconstruction of palaeohydrologic changes and identification of evolutional phases, to the history of glacio-isostatic movements and sea level changes in Late Pleistocene and Holocene.

The problem tackled at the earliest, in the late 17th and early 18th centuries by Urban Hjärne, Emanuel Swedenbourg, Anders Celsius and Carl von Linne, concerned the occurrence of readily observed traces of former Baltic shores found at considerable altitudes and far away from the modern coast (Hyvärinen, 2000). Studies on the history of the Baltic as understood at present began, too, in Sweden in the late 19th century. De Geer (1882) was the first to attempt to describe the history of the Baltic. As early as at the end of the 19th century, the basic stages in the evolution of the Baltic Sea were identified, based on changes in the hydrological setting inferred from malacological research. Lindström (1886) introduced the concept of the Littorina Sea. Munthe (1887) published the first coherent account of the Baltic history, arrived at by identifying three developmental stages preceding the present-day Baltic Sea: the Yoldia Sea, the Ancylus Lake and the Littorina Sea. The Baltic Ice Lake, a stage commencing the Baltic evolution, was the last major stage to be identified (Munthe, 1902, 1910). The division of the Baltic history into stages was subsequently elaborated on and refined. Munthe (1910) identified an additional stage, the Mastogloia Sea, separating the Ancylus Lake and the Littorina Sea and another one — the Lymnaea Sea, succeeding the Littorina Sea. Thomasson (1927) set off the Echineis stage as a period between the Yoldia Sea and the Ancylus Lake. The final, youngest stage, i.e., the Mya Sea was eventually defined by Hessland (1945) as the stage that not so much reflected salinity changes as it did the early effects of human activities. The Mya Sea stage had been earlier mentioned by Munthe (1910). *Mya arenaria* Linnaeus is a bivalve species brought by man from America to Europe. Until recently, the introduction had been assumed to transpire subsequent to Columbus's voyages. Recently, *Mya arenaria* Linnaeus has been demonstrated to have appeared in the Baltic earlier, during voyages of the Vikings (Petersen *et al*., 1992). The two youngest stages following the Littorina Sea, i.e., the Lymnaea Sea and the Mya Sea are occasionally referred to jointly as the Post-Littorina Sea (e.g. Mölder, 1946; Andrén *et al*., 2000). As a result, a complex and non-uniform set of Baltic history divisions has been created. Various authors identified different numbers of stages and substages in different parts of the Baltic; in addition, those stages and substages were assumed to persist for differing periods of time. Of the periods identified, those commonly accepted and referred to are: the Baltic Ice Lake, the Yoldia Sea, the Ancylus Lake and the Littorina Sea. The historical stages and substages have been usually identified based on succession of the malacofauna and/or diatom flora; occasionally, the location of former coastlines or sediment lithology were used as markers; not infrequently, a combined approach involving various markers has been used. The stages of the Baltic history, very frequently referred to in stratigraphy, are not always comparable between authors and between countries. A serious problem stems from the fact that the major fossil molluscs used to track down the major stages in the Baltic evolution in Sweden are not found in sediments of the Baltic seafloor (Duphorn, 1979). Generally, the stages identified in the Baltic history define time intervals when the connection between the sea and the ocean opened or closed. Consequently, the stages reflect primarily changes in hydrological (environmental) conditions and less so the changes in sedimentation processes. Towards the end of the 20th century, attempts at introducing some order into the system of divisions of the Baltic history were made (Hyvärinen, 1988; Svensson, 1989, 1991; Berglund, Björck, 1994; Björck, 1995). In addition, formal lithostratigraphic (Ignatius *et al*., 1981; Kotliński, 1989, 1991; Winterhalter, 1992) as well as biostratigraphic, diatom- (Alhonen, 1971) or mollusc-based (Alexandrowicz, 1999) divisions were attempted. Despite all those efforts, stages in the history of the Baltic have not been, so far, properly defined as geochronological units with generally accepted time-frames. Stratotypic pollen and chronostratigraphic profiles have not been worked out so far, either (Mangerud *et al*., 1974; Duphorn, 1979; Hyvärinen, 1988, 2000).

As already mentioned, the problem of the Baltic shoreline displacement in Late Glacial and Holocene as well as that of relative sea level changes and glacio-isostatic movements has been in the focus of science since studies on the history of the Baltic began. In Sweden, Finland and Estonia, i.e. in the areas dominated, in the Post-Glacial, by the Earth crust uplift, the problem of coastline displacement was adequately and satisfactorily solved (e.g. Sauramo, 1958; Kessel, Raukas, 1979; Eronen, 1983, 1987; Cato, 1985, 1992; Påsse, 1990; Svensson, 1991; Björck, Svensson, 1994; Hadenstrom, Risberg, 1999). Diatom research, pollen dating, and particularly the radiocarbon dating of sediment cores made it possible to accurately define the moment when the former marine embayments became isolated and when marine conditions switched over to a freshwater setting. Those data, together with the spatial analysis of the location of former shore formations, rarely found far away from the present coast and at considerable altitudes above the sea level, allowed to date the former shorelines and to follow the relationships between eustatic changes and glacio-isostatic land uplift (e.g. de Geer, 1940; Sauramo, 1958; Berglund, 1964; Mörner, 1976, 1980a, b; Eronen, 1983; Svensson, 1989, 1991; Harff *et al*., 2001).

The extent of knowledge on the location of former coastlines in the southern part of the Baltic Sea, where they are at present a part of the seafloor, is much narrower. Attempts at solving the problem of sea level changes were made at first and were accompanied by research on glacio-isostasy in the Southern Baltic involving seafloor morphology, a particular attention being paid to analysis of the location of erosional submarine terraces (e.g. Rosa, 1967, 1968; Kolp, 1979b, 1981, 1983, 1990). Relative sea level change curves, developed from radiocarbon dating of terrestrial sediments that are at present found on the seafloor or in the coastal zone on land, were published somewhat later (e.g. Duphorn, 1979; Klug, 1980; Kliewe, Janke, 1982; Winn *et al*., 1986; Bennike, Jensen, 1998). The curves were based on radiocarbon datings in the western part of the Baltic and primarily illustrated sea level changes within the last 8000 years. A comprehensive synthetic research on sea level changes in the Baltic Sea has been presented by, i.a. Mörner (1980a), Björck (1995), Harff *et al*. (2001) and Pirazzoli (1991, 1996), the latter dealing also with other seas and oceans. A review of the literature shows, however, the absence of satisfactory and acceptable solutions regarding both the relative sea level changes and glacio-isostatic rebound in the Southern Baltic.

AREA, OBJECTIVE AND SCOPE OF STUDY

AREA OF STUDY

The Baltic Sea, occupying an area of 415,266 km^2 and having a mean depth of 52 m, is divided into the following seven regions: the Bothnian Bay, Bothnian Sea, Gulf of Finland, Gulf of Riga, Baltic Proper, Belt Sea and Kattegat (Fig. 1), the division stemming from the shoreline complexity and seafloor morphology. This division is internationally recognised (Mikulski, 1987). The terms "Southern Baltic" or "southern Baltic area", although not precisely defined, are in commonly use, particularly in the Polish literature, both in monographs and cartographic publications (e.g. Łomniewski *et al*., 1975; Augustowski ed., 1987; Mojski *et al*., eds., 1995).

In this work, the two terms mentioned will be used interchangeably to refer to a part of the Baltic Proper which is bordered by the 14 and 20° meridians from the west and from the east, respectively, by the Polish coast from the south and by the 56° parallel from the north (Fig. 2). This work concerns primarily the Polish coast and the Polish Republic's maritime areas consisting of the internal waters (a part of the Gulf of

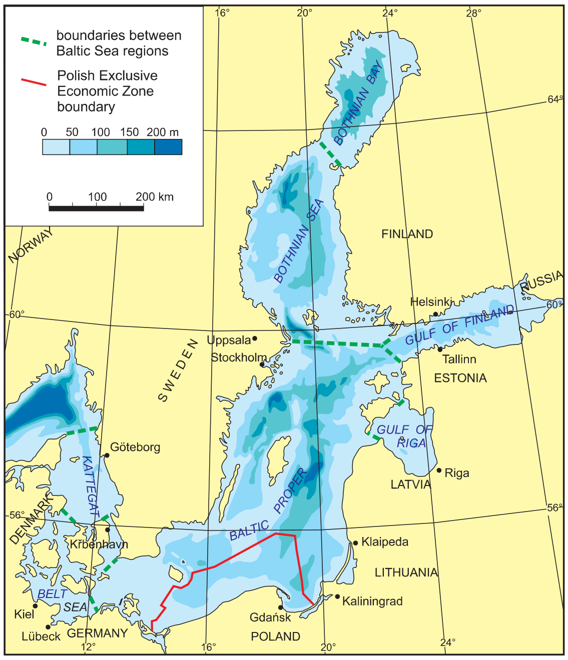


Fig. 1. Location of the study area

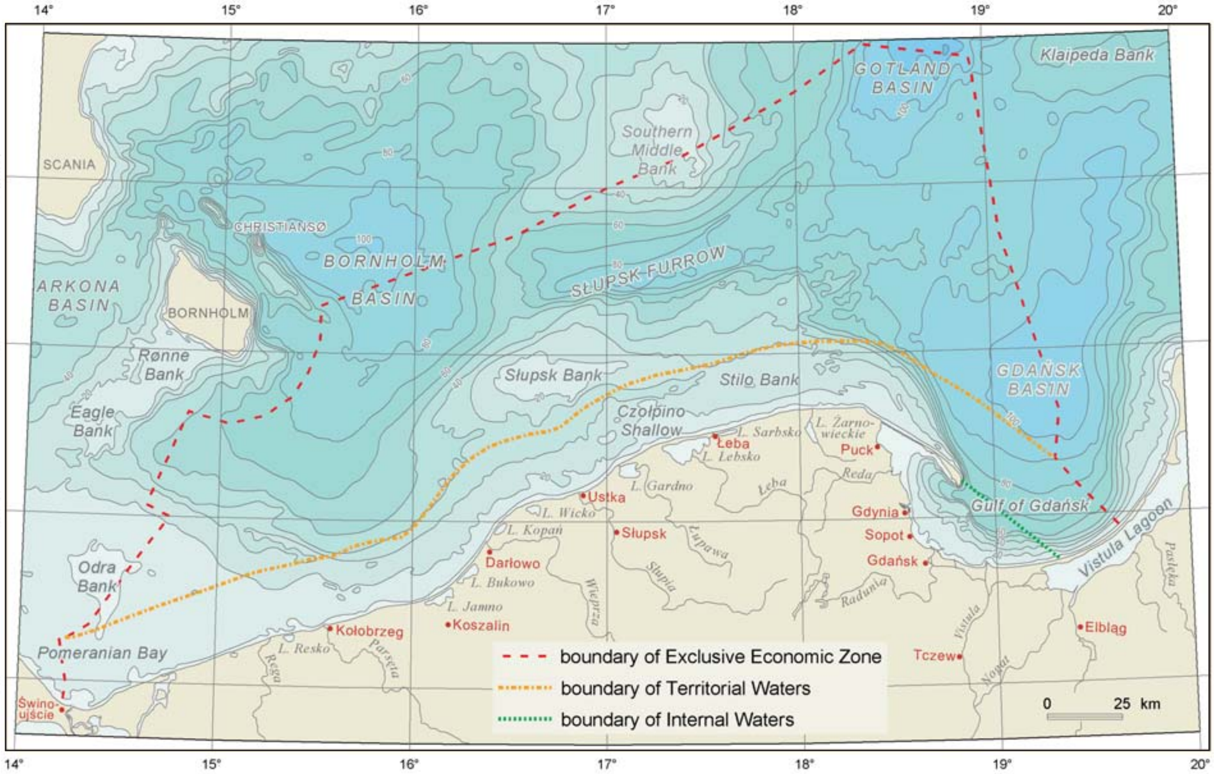


Fig. 2. The Southern Baltic Sea and Polish maritime areas

Gdańsk with the Puck Bay, Puck Lagoon and the Vistula Lagoon as well as the Szczecin Lagoon), the 12 nautical miles-wide territorial sea and the Exclusive Economic Zone; it is in those areas that most of the data used in this work were collected. The Polish Republic's maritime areas (excluding the Szczecin Lagoon) cover 30,533 km^2 and include parts of the major deep Baltic basins (the Bornholm Basin, Gotland Basin and Gdańsk Basin) as well as the Słupsk Furrow (Fig. 2). The basins are separated by sills the minimum depth of which, recorded on the sill that separates the Bornholm Basin and Słupsk Furrow, is about 60 m; the depth of a sill separating the Gdańsk and the Gotland basins is about 85 m. South of the basins there are coastal shallows, characteristic by the raised seafloor of the Odra Bank, Słupsk Bank, Czołpino Shallow and Stilo Bank. The only bank located north of the deep basins, away from the coast, is the Southern Middle Bank, found north of the Słupsk Furrow (Fig. 2). The southern Baltic area, defined as above, encompasses also — west of the Polish Exclusive Economic Zone — the Rønne Bank, Eagle Bank (Adlergrund) and the Arkona Basin; in the north-east, the area includes also a part of the Klaipeda Bank (Fig. 2).

OBJECTIVE AND SCOPE OF THE WORK

This work is a result of an attempt made at solving important and the least known problems related to the emergence and evolution of the Baltic Sea: the history of sea level changes and crust movements in the southern part of the Baltic. The southern Baltic area was the first to be freed from under the Scandinavian ice sheet and it was here that the history of the entire Baltic Sea began. At the same time, the Southern Baltic is the area in which, as a result of the domination of transgression processes, numerous formations and sediments that could have provided potential evidence of the southern Baltic shoreline location in different periods had perished; those forms and sediments that remained are at present located on the seafloor, frequently masked by a younger sedimentary cover.

This work is aimed primarily at reconstructing the history of relative sea level changes, at tracing vertical crust movements in time and following the demise of the last vestiges of the ice sheet, and at reconstructing changes in the shoreline location within the Southern Baltic in Late Pleistocene and Holocene.

Relative sea level changes were inferred chiefly from the analysis of the present position of radiocarbon dated marine and terrestrial sediments. A comparison of the Baltic sea level curve with curves depicting eustatic changes in the ocean level allowed insights into the history of glacio-isostatic rebound. Identification, analysis, and dating of relicts of forms that were associated with development of former coastal zones, sediment distribution, and extents of erosional surfaces related to various developmental stages of the Baltic provided grounds from which to reconstruct the palaeogeographic evolution of the southern Baltic area.

Questions of palaeohydrology, palaeoecology, and sediment stratigraphy of the Baltic are not discussed. A brief review of problems related to those areas was already given in the introduction. To eliminate ambiguity of the descriptions, chronostratigraphic units of Late Pleistocene and Holocene proposed Mangerud *et al*. (1974) are used. The Baltic's history stages and their respective duration follow those given by Eronen (1988) and Svensson (1991), while the time-frame of open connection between the Baltic and the North Sea is that proposed by Björck (1995). The events are timed with conventional radiocarbon age. However, in view of the fact that numerous authors use, more and more frequently, datings calibrated to the calendar time, some figures contain also a time axis marked with calendar years according to calibration curves published by Stuiver *et al*. (1998).

ANALYSED MATERIALS

To reach the objectives listed above, this work is based on materials and information on selected seismoacoustic profiles, sediment cores, and data related to their lithology and stratigraphy as well as on results of numerous radiocarbon datings of marine and terrestrial sediments, collected and stored in the database of the Polish Geological Institute's Marine Geology Branch. Information found in numerous Polish and foreign publications has been made use of as well.

The analysis involved a total of about 5,000 km of seismoacoustic profiling, the seismoacoustic surveys being conducted primarily with equipment (a boomer and occasionally a sparker) manufactured by EG&G. The surveys were conducted within 1982–1990, mainly as a basis on which to develop a 1:200,000 Geological Map of the Baltic Sea Bottom (Mojski ed., 1989–1995) and to document placer deposits (Masłowska, 1993). A 10×10 km network of seismoacoustic transects covered a substantial part of the Polish Exclusive Economic Zone. In gravel prospecting areas (the Słupsk Bank, Southern Middle Bank, vicinity of Koszalin and of Darłowo), the network is much denser, the nodes having an 0.5×0.5 km arrangement. It was only in the Gulf of Gdańsk and the Pomeranian Bay that a lower number of transects, irregularly distributed, was surveyed. Seismic profiling with high resolution reflection seismic equipment was carried out in 1996–1998, along a total length of 4,050 km, to develop a Geological Map of the Baltic Sea without Quaternary deposits (Kramarska *et al*., 1999). The profiles were spaced about 15 to about 25 km apart. Within 1995–2000, when developing a 1:10,000 Geodynamic Map of the Coastal Zone, seismoacoustic surveys with the 3010-B Seabed Oretech sub-bottom profiler were carried out along about 1,100 km in the area between the coastline and the depth of about 10–12 m.

A total of 30 boreholes, 7–30 m deep, were drilled in the open sea area of the Polish Exclusive Economic Zone; 824 cores, 3–6 m long, were collected with vibrocorers and gravity corers. As few as 26 boreholes and corers reached sub-Quaternary deposits. Bore-drilling and sediment coring in the land areas of the coastal zone, carried out since 1995, are a very important source of information as well. A total of about 300 wells, 5–30 m deep, were drilled on beaches and in their hinterland.

The materials collected from sediment cores were subjected to various lithological, macro- and microfaunal, diatom- and pollen-based analyses. Pollen analyses (performed by J. Zachowicz) involved about 200 samples from about 30 cores. Diatom analyses (performed by K. Zaborowska, A. Witkowski and M. Witak) were carried out on about 380 samples from about 25 cores. The composition of macrofauna was studied in about 700 samples from about 340 cores, while the microfauna (mainly ostracods) were identified (by J. Krzymińska) in about 580 samples from about 220 cores. Organic sediments and shells were radiocarbon dated. Documentation involving about 600 radiocarbon datings (e.g. in Tomczak *et al*., 1998; Tomczak *et al*., 1999; Uścinowicz *et al*., 2000) was analysed.

GEOLOGICAL SETTING OF THE SOUTHERN BALTIC: AN OUTLINE

PRE-QUATERNARY

The Baltic is the largest European intra-continental sea, located in its entirety on the regular continental crust. The geological setting of the Baltic is strongly related to that of the surrounding land mass. In terms of its structure and tectonics, the southern part of the Baltic is divided into two clearly different parts. The much larger eastern and central part lies within the Precambrian Craton (East-European Platform), while the western part is situated within the Palaeozoic platform of the central and western Europe. The two platforms are separated by the Koszalin Fracture Zone, a part of the transcontinental Teisseyre–Tornquist Zone (Dadlez in: Mojski *et al*., eds., 1995; Kramarska *et al*., 1999).

The southern Baltic part of the East-European Platform is located between the Baltic shield in the north and north-west with crystalline rocks outcrops and the Baltic syneclise in the east and south-east. The sedimentary cover thickens from north-west to south-east. Quaternary deposits overlie a sequence of Sylurian and Devonian sediments. In the south, those sediments are capped by formations belonging to the Zechstein–Mesosoic structural complex. The widest extent is shown by the Upper Cretaceous sediments overlying older formations. Locally, Permian and Triassic forms crop out from beneath the Cretaceous deposits. Tertiary (from Eocene to Miocene) sediments overlie older deposits across the area from Koszalin in the west to Jastarnia on the Hel Peninsula, and reach to the slopes of the Southern Middle Bank in the north. Tertiary (Lower Palaeocene) sediments occur locally near the western border of the East-European Platform, ingressing occasionally into the Koszalin Fault Zone and extending into the Palaeozoic platform. Tertiary sediments occur locally, too, near the coast of the Gulf of Gdańsk (Kramarska in: Mojski *et al*., eds., 1995; Kramarska, 2000; Kramarska *et al*., 1999; Kramarska *et al*., 2002). Sediments beneath the Cenozoic cover in that part of the Precambrian platform located within the southern Baltic evidence the presence of a sub-latitudinal system of fractures. The fracture zones of (from east to west) Kuźnica, Karwia and Smołdzino delimit the Curonian, Rozewie, Łeba and Słupsk blocks. A broad band of tectonic deformations occurs in the western part of the Precambrian platform, between Bornholm and Darłowo.

The Palaeozoic platform of the central and western Europe, located south-west of the Teyserre–Tornquist Zone is more diverse in its geological and tectonic structure. Lower Palaeozoic sediments, folded and thick, are covered by a platform of a Devonian–Carboniferous complex. The complex is capped, like in the Precambrian platform, by a Permian–Mesosoic formation. The three major fracture zones of (from east to west) Koszalin, Trzebiatów and Kamień Pomorski–Adler delimit the Kołobrzeg, Gryfice and Wolin blocks. The raised western margins of the Kołobrzeg and Gryfice blocks are accompanied by anticlines with Jurassic and Triassic sediments in their core parts. The eastern, subsided, parts of the blocks contain synclines filled with Cretaceous sediments (Kramarska *et al*., 1999; Kramarska, 2000).

The sub-Quaternary surface relief is primarily glacial erosion-related. The deepest location of the top of pre-Quaternary deposits is found in the Bornholm Basin. The Cretaceous and Sylurian sediments underlying the Quaternary deposits occur there from 100 to almost 200 m below the sea level (b.s.l.). In the Gdańsk Basin, the sub-Quaternary surface consisting of Cretaceous formations lies down to 140 m b.s.l. The Tertiary sediment capping in the Słupsk Furrow is found at depths of 90–110 m b.s.l. In the southern part of the Gotland Basin, the pre-Quaternary deposit surface, consisting of Sylurian and Devonian sediments, lies at depths of 110–120 m b.s.l.

The sub-Quaternary surface south of the Bornholm and Gdańsk basins and the Słupsk Furrow lies at much shallower depths, from 40 to 100 m b.s.l. The shallowest location of pre-Quaternary sediments is found at the feet of cliffs near Gdynia (Orłowo and Oksywie) and north-west of Władysławowo (Chłapowo), where Tertiary sediments crop out. The recent bathymetry of the Baltic Sea is largely related to the sub--Quaternary surface morphology. The sub-Quaternary relief is rendered more diverse by subglacial incisions the relative depths of which reaches 100 m. The troughs are frequent within the pre-Quaternary deposit fractures and occur both in deep basins and in shallow-water areas, often continuing inland (Kramarska *et al*., 1999; Kramarska, 2000; Kramarska *et al*., 2002; Mojski, Tomczak, 1994).

QUATERNARY

The southern Baltic seafloor Quaternary cover consists of Pleistocene sediments, deposited mainly during cold periods, but also during interglacials; it contains also Holocene deposits, typical of both terrestrial and marine environments. The Quaternary sediments are 1.3 to about 300 m thick. The thinnest sediment layers occur in the recent deep basins, affected by glacial erosion during Pleistocene. Sediment layers of a particularly low thickness (<10 m) were found in areas affected, also during Holocene, by erosional processes. Such areas include the central part and the southern slopes of the Słupsk Furrow as well as the southern outskirts of the Gotland Basin. Thicker layers are found in shallow areas, and particularly in the inshore zones where the Quaternary sediment cover is locally more than 200 m thick (Kramarska *et al*., Uścinowicz in: Mojski *et al*., eds., 1995).

The Pleistocene is represented primarily by sediments of glacial, fluvioglacial, and limnoglacial accumulations and, to a lesser degree, also by the interglacial and interstadial sediments of terrestrial and marine environments. Most common are tills of various facies and age. A single layer of subaqueous diamicton, less than 10 m thick, covers the southern slopes of the Słupsk Furrow and occurs within the southernmost part of the Gotland Basin. The Bornholm and Gdańsk basins as well as the northern slopes of the Słupsk Furrow usually have two till layers. In the basins, tills are covered by clayey, varved and microvarved limnoglacial (originating in the Baltic Ice Lake) deposits constituting the upper surface, Pleistocene in age (Uścinowicz *et al*., 1988; Uścinowicz, 1995; Uścinowicz in: Mojski *et al*., eds., 1995). The Pleistocene sediment thickness within the basins ranges from 1.3–10 m on the southern slopes of the Słupsk Furrow and in the southwestern part of the Gotland Basin to 10–20 m in the Bornholm Basin and on the northern slopes of the Słupsk Furrow, to 30 m in the central part of the Gdańsk Basin. Thicker (up to 60 m) Pleistocene sediments are found in the basins only locally, within deep incisions in the Quaternary base (subglacial troughs) (Pikies, 2000). They are occasionally associated with fluvioglacial accumulation formations, such as the Bornholm Basin's esker system (Kramarska *et al*., Uścinowicz in: Mojski *et al*., eds., 1995).

Much more complex is the Pleistocene deposit structure on basin slopes, particularly in shallow areas. In addition to several till layers, there are fluvioglacial and limnoglacial accumulation sediments as well as interglacial and Late Glacial barrier deposits of the Baltic Ice Lake, deltaic and limnic-marshy sediments. Sediments of those genetic types make up the cover of sub-Holocene shallow zone. The Pleistocene sediment thickness there varies greatly, from 5–10 m west of Jarosławiec to about 300 m near Czołpino. Like in basins, the thickest (more than 100 m thick) Pleistocene sediments occur in subglacial troughs cutting through the Quaternary base (Mojski, Tomczak, 1994; Kramarska *et al*., Uścinowicz in: Mojski *et al*., eds., 1995; Uścinowicz, 1995, 1996).

The southern Baltic Holocene consists of two parts. In deep basins, the lower member is made up by Preboreal and Boreal silty-clayey sediments of the Yoldia Sea and Ancylus Lake, the upper member being formed by Atlantic, Subboreal and Subatlantic silty-clayey sediments of the Mastogloia, Littorina and Post-Littorina seas. The total thickness of the Holocene sedimentary cover ranges from 7–8 m in the Bornholm Basin to 5–6 m on the northern slopes of the Słupsk Furrow to as little as 3–4 m in the southern part of the Gotland Basin. The thickest (about 15 m) Holocene sediment layers occur in the Gdańsk Basin. Sediments on the outskirts of the basins and in the sills dividing them, and locally also in seafloor elevations within the basins, belong to the upper Holocene part only and consist of sandy-silty and occasionally gravelly-sandy-silty (mixite) facies, their thickness usually not exceeding 20 cm. Whenever the pycnoline converges with the seafloor, the mixite facies is accompanied by ferromanganese nodules.

In shallow areas, the lower Holocene member is made up by Preboreal, Boreal and Atlantic limnic, deltaic, and lagoonal sediments. They are present in the Pomeranian Bay, Gulf of Gdańsk and the Słupsk Bank as well as in some parts of the coastal zone. The shallow water Holocene sediments are thickest in the Gulf of Gdańsk and are associated with the Vistula delta development. Thick (locally more than 10 m) Holocene sediments occur also within the Vistula Lagoon and Vistula Spit as well as in the Pomeranian Bay and the Odra Bank. The lower member sediments (terrestrial Preboreal, Boreal and Atlantic deposits) were, to a large extent, eroded away during transgressions, so that slopes of the basins as well as extensive shallow water areas are covered only by marine sediments of the upper Holocene member. They comprise Atlantic, Subboreal and Subatlantic sands and gravels of the Littorina and Post-Littorina seas. In most of the shallow water areas, the marine sand and gravel layer is usually less than 2 m thick, its thickness over extensive areas hardly reaching 1 m. It is only locally that the Holocene marine sands are more than 3 m thick. The Holocene sands in the southeastern part of the Southern Middle Bank, in the southern part of the Słupsk Bank, and in the Czołpino Shallow are up to 6 m thick. Sand accumulations in those areas, and also in the Odra Bank, are most probably associated with barrier structures developing during the initial phase of the Atlantic transgression. The Holocene formations of the Hel Peninsula and the remaining recent barrier areas are a separate problem. The Hel Peninsula is notable in having the thickest (up to 100 m) Holocene sediments among those recorded in Poland (e.g. Tomczak, 1990; Kramarska *et al*., Tomczak, Uścinowicz in: Mojski *et al*., eds., 1995).

RELATIVE SEA LEVEL CHANGES

THE PROBLEM AND METHODS FOR ITS SOLUTION

Research on the southern Baltic shoreline displacement was conducted primarily by Rosa (1963, 1967, 1968, 1987, 1994, 1997). His publications (1963, 1968, 1987) present a series of the sea level curves and their modifications. Due to the lack of data on the geological setting and seafloor structure of the area of interest, the curves were developed by extrapolating data pertaining to shorelines of the central and northern Baltic onto the southern Baltic area and by interpreting bathymetric evidence (Rosa, 1987, 1991). Rosa's more recent curves (1994, 1997) are a modification of the previous ones, the modification resulting from analysis and interpretation of a single seismoacoustic profile in the Gulf of Gdańsk. Unfortunately, Rosa's interpretations and information on the age and structure of the layers and formations he analysed were not backed up by sediment datings or results of lithological research.

Curves representing relative sea level changes, developed from both a preliminary analysis of radiocarbon datings of marine and terrestrial areas from the southern Baltic bottom and adjacent coastal zone, and from seismoacoustic profiles were also published by the present author (Uścinowicz, 1999b, 2000a, b, 2001).

In addition to curves covering the entire post-glacial history of the Southern Baltic, the Polish 1982–1999 literature contains also curves of changes in the southern Baltic sea level for a shorter periods of time, and for selected parts of the Polish coast (Mojski, 1982; Tobolski, 1987, 1989, 1997; Wojciechowski, 1990; Kaszubowski, 1992, 1995; Tomczak, 1995a; Rotnicki, 1999).

In 1982, Mojski correlated several radiocarbon datings and lithostratigraphic data of the Vistula Delta with stages of the Baltic history and Rosa's 1963 curve of sea-level changes in the Gulf of Gdańsk. In his 1988 paper, Mojski continued his analysis of correlation between the Vistula delta development and changes in the Gulf of Gdańsk sea level by dating the termination of peat accretion in the northern part of the delta and, indirectly, the onset of barrier sediment accumulation, to about 6300 years BP.

Based on his palaeoecological research on and radiocarbon dating of the Gardno–Łeba Lowland sediments, correlated with results of detailed studies on and radiocarbon datings of a reference site at Kluki, Tobolski (1987, 1989, 1997) outlined changes in the southern Baltic sea level over the last 7500 years. He distinguished between 6 transgression phases and dated the maximum Baltic sea level in the Holocene to *c*. 4600 years BP; he assumed the sea level to be then by *c*. 1 m higher than present.

Detailed studies on the Lake Gardno sediment lithofacies, particularly his 20 radiocarbon datings of the peat underlying lacustrine sediments, allowed Wojciechowski (1990) to develop a curve of water level changes in the lake since *c*. 7.0 ka BP. He distinguished between four transgression phases related, in his opinion, to the Littorina Sea transgression history, and pinpointed the maximum lake level, higher by about 0.5 m than at present, to occur *c*. 2500 years BP. Wojciechowski (1990) had neither radiocarbon datings nor results of pollen and diatom stratigraphy of the lake's sediments in his disposal. In the light of earlier pollen- and diatom-based analyses (Bogaczewicz-Adamczak, Miotk, 1985a; Zachowicz, Zaborowska, 1985), it may be contended that marine transgression affected Lake Gardno area not earlier than in the Subboreal; the lake, similar to its present shape, emerged on the onset of the Subatlantic.

Kaszubowski (1992) published a curve of the Baltic sea level changes in the central part of the Polish coast over the last 8000 years, while his 1995 paper contains a curve of Late Holocene sea level changes in the same area. Both curves illustrate a number of transgression-regression phases of amplitudes ranging within 2–4 m. Unfortunately, in neither case did the author backed up his ideas with any detailed data.

Tomczak (1995a) put together radiocarbon datings from lithological profiles obtained from the southern Baltic coast. The profiles supplied a number of indicators providing information on palaeogeographic changes in the southern Baltic area and on the adjacent coast. Information on, i.a., onset of accumulation of organogenic sediments and their origin and on timing of environmental shift from terrestrial to marine conditions made inferences on the sea level changes possible. The curve, developed for the Southern Baltic, reflected a fast (*c*. 10–15 mm/a) rise, occurring over 8.0–6.3 ka BP, from –30 m to a level lower by a few metres than at present. According to that curve, the sea approached its present level 1000 years ago, following a fast rise by about 1.5 m.

Rotnicki (1999) published a preliminary interpretation of the history of sea level changes within the central part of the Polish coast (the Gardno–Łeba Lowland). The ideas he presented were based on results of facial and stratigraphic analyses, including radiocarbon datings of the Gardno–Łeba Lowland sediments (Rotnicki *et al*., 1999; Rotnicki, 1999). According to Rotnicki (1999), the Baltic sea level in the area of the Gardno–Łeba Lowland approached that at present as early as about 8000 years BP and exceeded it for the first time about 6500 years BP. The current level of the Baltic was for the second time exceeded about 3000 years BP. About 7500 and 4000 years BP, the sea level fell by about 3 and 5 m, respectively.

Although based on detailed regional studies and concerning mostly the central part of the Polish coast the curves of the sea level changes described above (Tobolski, 1987, 1989, 1997; Wojciechowski, 1990; Rotnicki, 1999) differ, in many cases and in numerous details, both from earlier ideas of Rosa and from one another as well. The causes of the interpretative differences being so wide could be, perhaps, unravelled by reverting to the original data and re-analysing them.

For the purpose of this work, radiocarbon-dated marine and terrestrial sediments are the basic indicators of sea level changes. When radiocarbon datings were too few or altogether lacking for a period of time or a depth range, additional proxies were sought in the extent of erosional surfaces and progradational structures associated with evolution of the coastal zone, such as deltas and/or barriers. The relative age of those formations was determined from a sequentional seismostratigraphic analysis of seismoacoustic profiles. At some sites, the age of the formations identified was validated by referring to radiocarbon and pollen datings. It should be pointed out that the age determined for those formations, and particularly their relation to the sea level at which they were formed, may be biased, the magnitude of bias being occasionally difficult to assess. The bias resulted from the fact that progradational structures are frequently transformed in part by erosional processes, operating later on. Erosional surfaces could have emerged both when the sea level was rising and when it stabilised. The so-called submerged terraces were not taken into account as indicators of sea level changes in this work. They were made use of in the geomorphological approach to the analysis of sea level changes (e.g. Kolp, 1965, 1976, 1979a, b; Rosa, 1967, 1968; Healy, 1981). The basic problem presented by the terraces, however, involves determination of their origin and structure, their location relative to the sea level during formation, and — most importantly — determination of their age. Similar reservations with respect to the utility of the terraces were earlier expressed by, i.a., Rudowski (1979), Winn *et al*. (1986) and Lemke (1998).

To develop a curve illustrating relative sea level changes one has to plot, in a two-dimensional coordinate system, a set of sea level indicators, each expressed by two numerical values: depth (ordinate) and age. Both values should be accompanied by the range of possible errors introduced when they were determined. The set of data, along with their precision estimates, plotted this way sets a foundation for developing of a sea level curve (Pirazzoli, 1996). A sea level curve not accompanied by the set of data used to plot it is inappropriate because it may then reflect not only the sea level data, but also subjective ideas as well as preconceived theories and interpretations. The user may find then the source data on sea level changes more valuable than a curve that should, in principle, synthesise the analysis (Shennan, Tooley, 1987).

An important problem encountered when developing curves of relative sea level changes involves a possibility of various errors, both those related to the location of dated samples and those associated with dating results themselves (van de Plassche, 1980). One of the important issues is the accuracy of dating, particularly potential errors related to timing of deposition of the sediments dated. A particular caution should be exercised when viewing results obtained for various fine-grained sediments containing scattered organic matter that was dated. That organic matter might have originated not only from primary production in the sedimentary basin, but also from redeposition of older sediments. The presence of redeposited older organic matter in samples of fine-grained sediments results in the radiocarbon dates being "older" than the actual timing of deposition of those sediments. The magnitude of this "ageing" effect may be estimated only by using other sediment dating techniques, e.g. pollen analysis. That this effect occasionally occurs is confirmed by age inversions (lower core parts showing a younger radiocarbon age, compared to the overlaying parts) found in lacustrine and lagoonal sediment cores. Discrepancies, of up to few thousand years, between radiocarbon and pollen datings are not uncommon. For example, sediments in core ZW6 from the Vistula Lagoon, radiocarbon dated to 2500 ±130 (Gd-9450) and 2010 ±140 (Gd-9451) (Appendix 1; Uścinowicz, Zachowicz, 1996), showed both age inversion and incompatibility relative to the pollen spectrum. Radiocarbon analyses dated those sediments to the turn of the Subboreal and the Subatlantic, while pollen spectra, involving pollen of grains and other plants associated with human activities, pointed to deposition taking place in the early Middle Ages, i.e. 1500 years later (Zachowicz, 2001, personal comm.). The "ageing" effect of radiocarbon dating, resulting from an admixture of redeposited, older organic matter, may occur also in marine sediments deposited away from the shore. The effect there is likely to be much weaker and to have a much narrower range, compared to limnic and/or lagoonal sediments. The effect in Middle and Late Holocene marine sediments of the Bornholm Basin was estimated, in a preliminary analysis, at about 400 years (Andrén *et al*., 2000).

Although still possible, the probability that a dated sediment is contaminated by admixture of younger organic matter, thus producing a radiocarbon age "younger" than the actual one, is much lower than the probability of ageing a sediment by the redeposited organic matter.

The problem of apparent *vs*. true age persists also due to the presence of a reservoir effect when carbonate sediments and shells are radiocarbon dated. The effect, for shells collected from the southern part of the Baltic Sea, was estimated at about 400 years (Krog, Tauber, 1974) or at about 320–370 years (Pazdur *et al*., 1999). Based on their dating of the Ancylus Lake mollusc shells from Gotland, Königsson and Possnert (1988) estimated the reservoir effect at about 400 years. In carbonate and lacustrine sediments, the reservoir effect may vary over a still wider range (Pazdur *et al*., 1999). Andrén *et al*. (2000) estimated the marine reservoir effect in the Middle and Late Holocene sediments of the Bornholm Basin, deposited under brackish conditions, at 300 years.

Another important problem arising during development of a relative sea level curves is the accuracy with which the ordinate of the layer dated is determined, relative to the present mean sea level. The area or depth ordinate at a sampling site may carry the largest error. During research carried out in the terrestrial part of the coastal zone, the area ordinate was most frequently read from 1:25,000 or 1:10,000 topographic maps. Even the 1:10,000 maps, available since 1990, do not make it possible to read the area ordinate to less than about 0.5 m. When working on the sea shore or on shores of lakes and lagoons, the area ordinate was most often determined relative to the instantaneous sea level (or the ground water level), rather than relative to the mean level. Short-term (diurnal) sea level variations on the southern coast of the Baltic, except for extreme cases, may amount to about ±0.2–0.3 m. When working on board a research vessel and echo-sounding the depth, one should bear in mind the errors resulting from regional and seasonal variability of sound velocity in water as well as from changes in the shape of the free sea surface. It may be assumed that those factors, as well as — to a smaller degree — equipment errors allow to measure depth with a similar accuracy (about ±0.25 m), relative to the mean sea level. Perhaps a better accuracy was reached in many cases, but — because publications and archived data frequently lack information on positioning techniques used in field work, and in view of the large data set dealt with in this work — sample positioning accuracy of 0.5 m, assumed in this work, does not seem an overestimation.

Still another problem that should be factored in when developing and analysing sea level curves. The problem is encountered when dating transgressive contacts between terrestrial sediments (most frequently peat) and marine sediments overlying them and involves a possibility that the original location of the terrestrial sediment top has changed. One of the causes is erosion of the peat capping during marine water incursion. Thus the actual marine transgression into an area might have, in fact, occurred later than the timing of transgression as determined from the age of peat top (Jelgersma, 1961; van de Plassche, 1980). Errors may stem also from a change in the original location of the layer dated, the change being provoked either by compaction of the sediments (most frequently peat) being dated or compaction of the underlying deposits (van de Plassche, 1980). The original location may change as a result of, i.a., vertical displacement of peat layers under influence of differing weight of migrating dunes. Peats and lagoonal sediments may not only be compacted, but also squeezed and pushed upwards (Gudelis, 1979).

To develop the southern Baltic relative sea level curve, samples collected from sites located below 1 m above sea level (a.s.l.) and younger than 15,000 conventional radiocarbon years were selected. The collection was supplemented by datings published by other authors (e.g. Rotnicki *et al*., 1999) past the year 1997. Altogether, a total of 314 radiocarbon datings from 163 sites located in the Polish part of the Southern Baltic and the adjacent coastal zone on land (Fig. 3; Appendix 1) were used. The used datings had not been calibrated, nor had the reservoir effect been taken into account in shell- and carbonate-containing sediment datings. Most datings were obtained with the classic technique at the Radiocarbon Laboratory of the Silesian Technical University. Five shell datings and one peat dating were performed with AMS ^{14}C at the Leibniz-Labor für Altersbestimmung in Kiel (Germany). Of the samples dated, 77 were additionally subjected to pollen analyses. Sediment deposition environments were identified with diatom-based as well as micro- and macrofaunal analyses performed on 103 dated samples, the analyses concerning primarily limnic, lagoonal and marine sediments (Appendix 1). The information on site locations and the type of sediments dated can be summarised as follows:

Location of sites yielding sediment to be dated: 81 sites (158 datings) were located on the sea bottom and on the bottom of lagoons and coastal lakes; 82 sites were located on spits, in the Vistula delta and on coastal lowlands; 136 sites were lo-

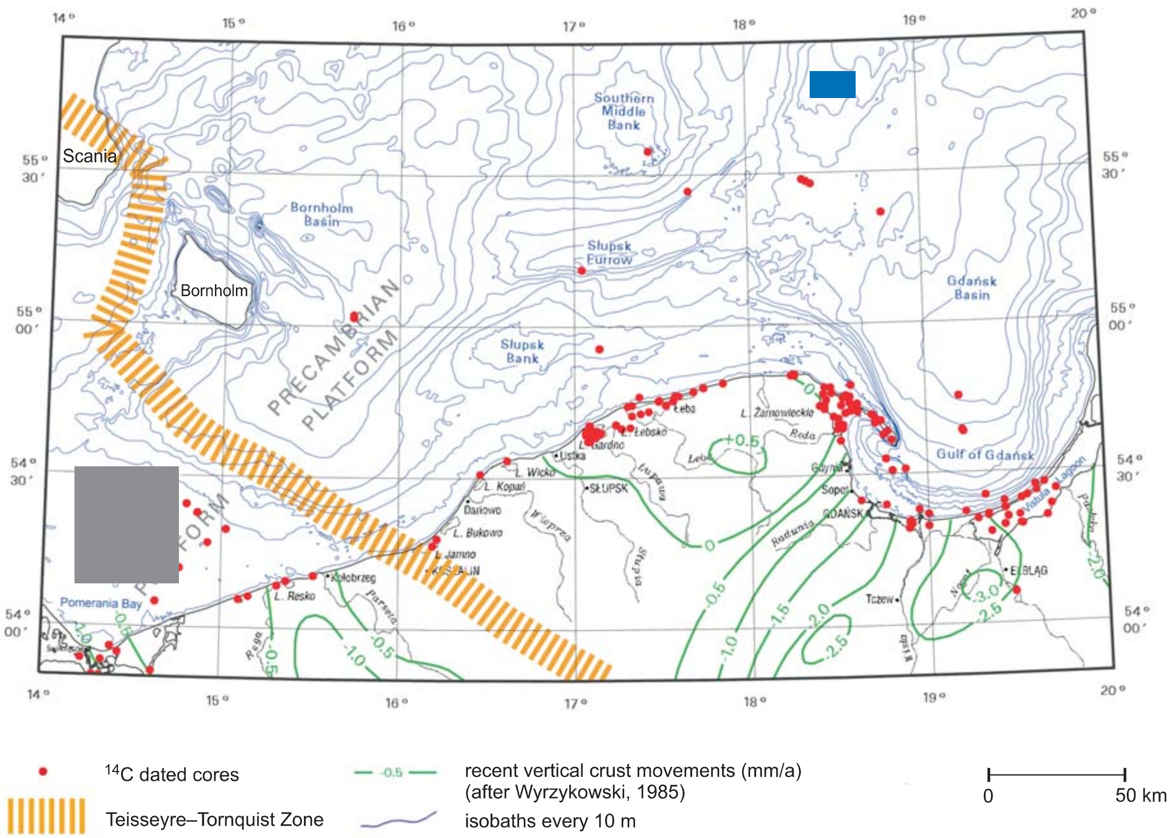


Fig. 3. Location of ^{14}C dated cores

cated on the Precambrian platform, 27 sites being located on the Palaeozoic platform (Fig. 3).

Types of dated sediments: peat, muddy peat, and peaty muds yielded 168 datings; gyttjas, limnic muds and sands with plant detritus — 20 datings; muds, sandy muds, sands with plant remains, lagoonal and deltaic (brackish) sands — 41 datings; marine muds — 52 datings and marine shells (primarily *Cerastoderma* sp., occasionally *Macoma* sp. and *Hydrobia* sp.) — 33 datings.

One of the problems that had to be overcome when developing the relative sea level curve was inhomogeneity of the set of datings used. Most of the datings (53.2%) were performed on samples collected from 0 to 10 m b.s.l.; the 10–20 m depth range produced 18.9% of samples datings, the 20–30 depth range yielding 8.3% of samples dated (Fig. 4). Temporal distribution of datings is far from homogenous as well, as 44.2% of the datings concern the Atlantic (8–5 ka BP) deposits, 26% being produced by the Subboreal and the Subatlantic (5–0 ka BP);

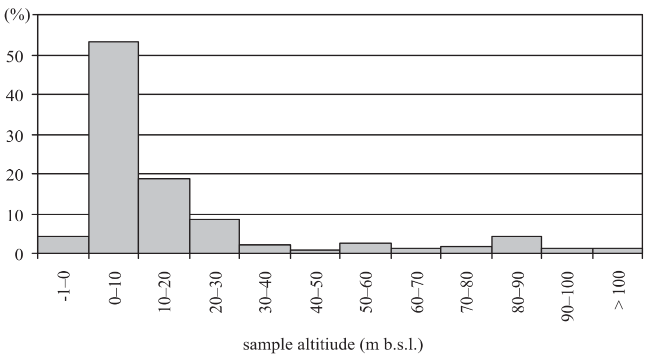


Fig. 4. The frequency of radiocarbon datings in relation to sample altitude

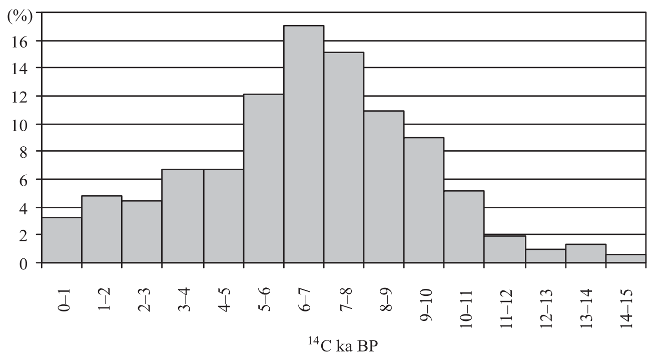


Fig. 5. The frequency of radiocarbon datings in relation to sample age

19.9% were assigned to the Boreal and Preboreal and 9.9% to Late Pleistocene (15–10 ka BP) (Fig. 5). In view of such a distribution of datings, the relative sea level curve development was assisted by evidence provided by seismoacoustic and echo-sounding profiles. Location of various formations, associated with the shoreline development, on the seafloor and extents of erosional surfaces were analysed; the relative age of those formations was determined from sequential seismostratigraphic analyses of seismoacoustic profiles. Age of the sequences identified was determined at certain sites by radiocarbon and pollen dating.

In view of the low number of radiocarbon datings from Late Pleistocene (15–10 ka BP), the time frame of events taking place during that period were determined from more recent literature data, primarily those concerning southern Sweden (e.g. Svensson, 1991; Björck, Digerfeldt, 1991; Berglund, Björck, 1994; Björck, Svensson, 1994; Björck, 1995). Data pertaining to the southern Swedish coast and the western Baltic (i.e. Björck, Svensson, 1994; Winn *et al*., 1986; Bennike, Jensen, 1998; Jensen *et al*., 2000) proved helpful for timing the Early Holocene events and when determining relative changes in the Baltic sea level.

The relative sea level change curve for the Southern Baltic, developed from the materials and data described above, is presented against the backdrop of both all the available radiocarbon datings of marine and terrestrial sediments (Fig. 6) and in relation to the key reference sites, described more extensively in the text and shown in figures (Figs. 7, 8).

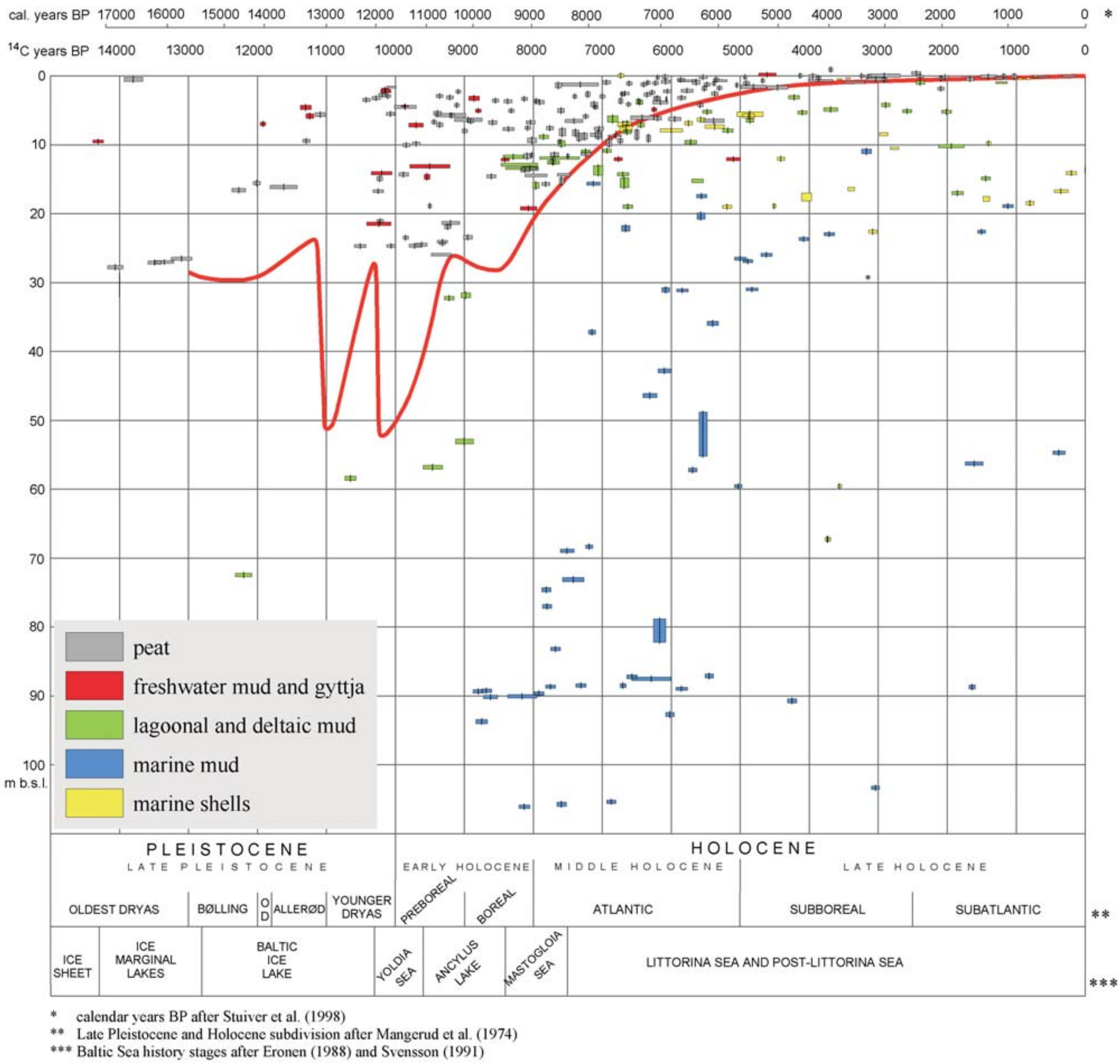


Fig. 6. The relative sea-level changes of the Southern Baltic as inferred from radiocarbon dating of terrestrial and marine deposits

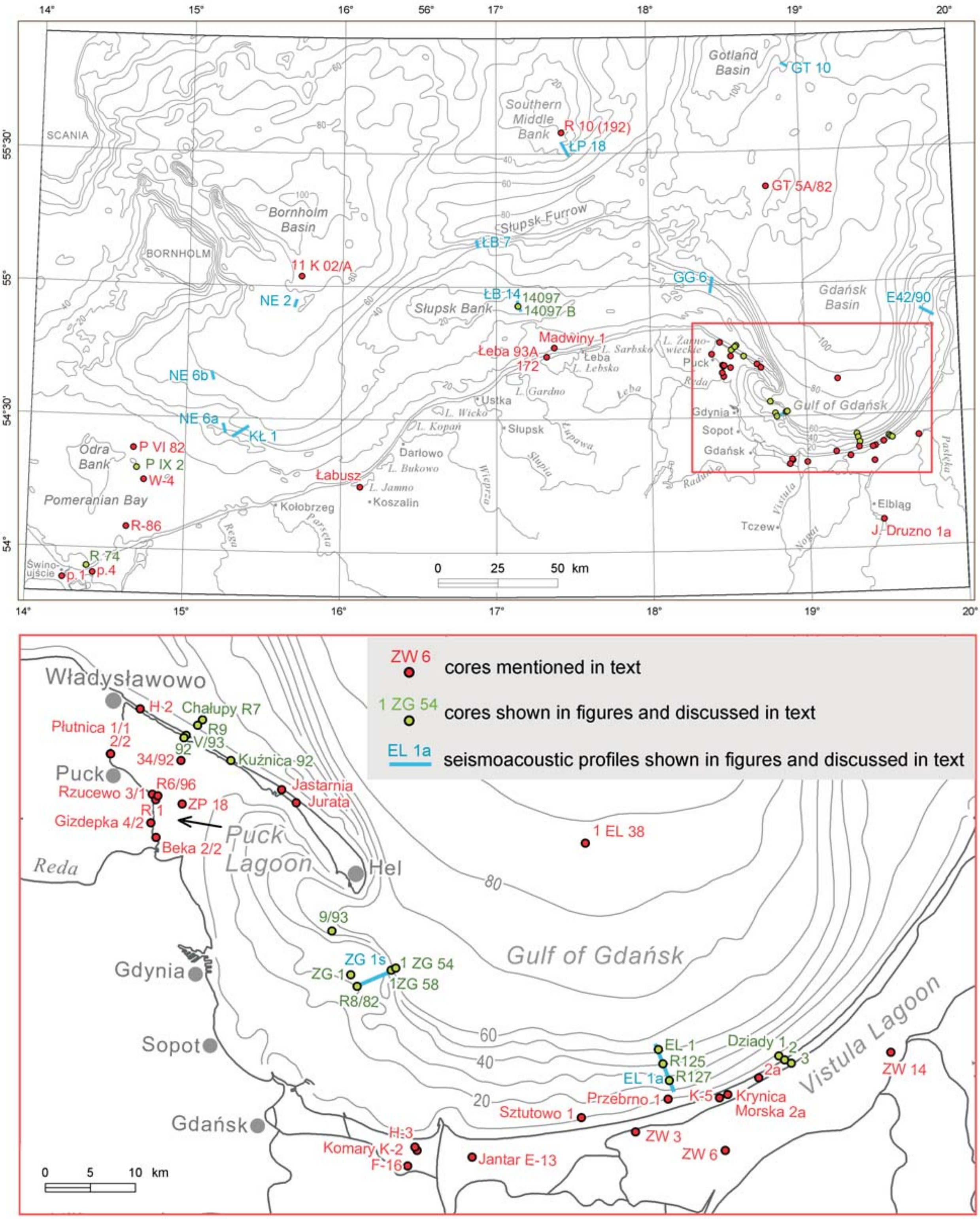


Fig. 7. Location of reference sites in the southern Baltic area

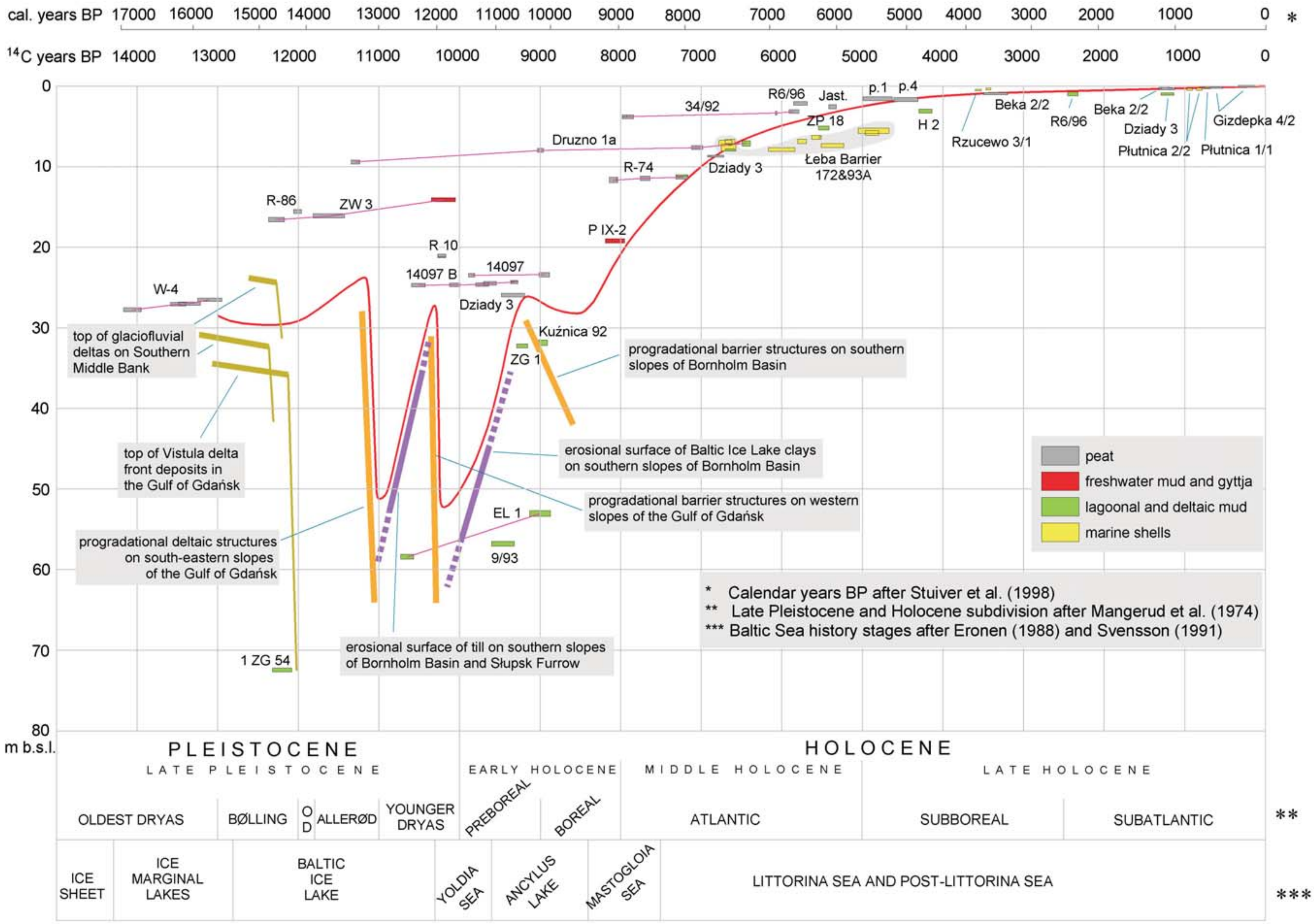


Fig. 8. The relative sea level changes of the Southern Baltic as inferred from reference sites

LATE PLEISTOCENE AND EARLY HOLOCENE

Late Pleistocene, from the Gardno Phase ice sheet retreat from northern Poland and southwestern Baltic (*c*. 14.0 ka BP) until the Southern Middle Bank Phase (*c*. 13.0 ka BP) (Uścinowicz, 1999a), is the least known part of the history of the Baltic Sea level changes. As the ice sheet retreated from the Southern Baltic and a northwards-inclined surface uncovered, more and more numerous and increasingly larger ice-dam lakes were remaining. Their extent and level changed rapidly as new pathways of meltwater drainage emerged and progressed northwards. About 14.0 ka BP, the Danish Straits became ice-free (Lagerlund, Houmark-Nielsen, 1993) and thus the meltwater could be at that time finally drained out from the southern Baltic basin through the Straits. The outflow level and changes in the southern Baltic water level were controlled by an interplay of deglaciation dynamics, glacio-isostatic movements, and erosion of the Danish Strait sills as well as by the eustatic ocean level changes (Björck, 1995).

Emergence of the Late Pleistocene Baltic at the first stage of the Baltic history, i.e. the Baltic Ice Lake, is related to the connection opening between ice-dammed lakes of the Gdańsk and Bornholm basins via the Słupsk Furrow. This happened *c*. 13.5–13.0 ka BP when the ice sheet margin was retreating from the Słupsk Bank moraines (Uścinowicz, 1996, 1999a) (Fig. 9).

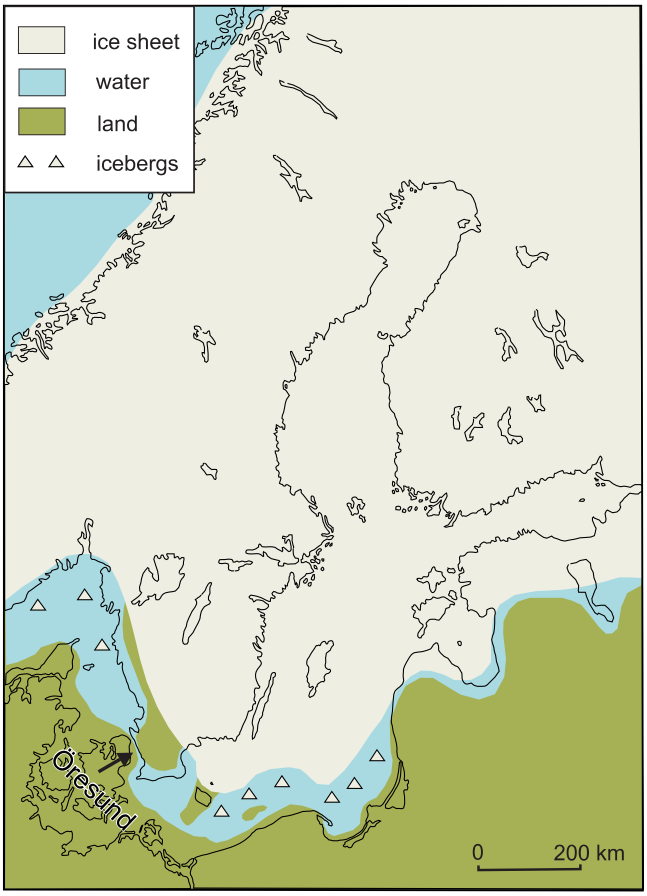


Fig. 9. The Baltic Sea area during early phase of the Baltic Ice Lake (*c*. 13.0–12.8 ka BP)

As inferred from data collected in southern Sweden and in the Danish Straits (e.g. Berglund, Björck, 1994; Björk, Svensson, 1994; Björck, 1995), the outflow at the initial Baltic Sea stage proceeded through Öresund. This is evidenced also by the contemporaneous sedimentary sequences, identified in the Kattegat (Bergsten, Nordberg, 1992). Large amounts of outflowing meltwater coupled with a simultaneous glacio--isostatic uplift resulted in rapid erosion of Öresund, so that the outflowing water could remain at a more or less stable level. The uplifting areas of southern Sweden (Scania and Blekinge) were at that time subjected to regression. Water level changes on the southern shores of the Baltic Ice Lake, most probably uplifting at a lower rate, were in all likelihood slight only. The situation changed about 12.0 ka BP when the Öresund erosion, having reached the pre-Quaternary bedrock, slowed down and when the outflow from the Baltic Ice Lake stopped due to continuing uplifting crustal compensation movements in the area (Berglund, Björck, 1994; Björk, Svensson, 1994; Björck, 1995). The regression in southern Sweden slowed down considerably a slight transgression occurring in the southern part of the Baltic Ice Lake.

Water level of the Baltic Ice Lake, which at that time covered the Gdańsk Basin, Słupsk Furrow and the southern part of the Bornholm Basin, was initially (i.e. 13–12.5 ka BP) lower than the present one by about 30 m. The level could be, with some approximation, determined from the location of peat in the Pomeranian Bay, the location of sediment top of the then Vistula delta front, and from the glaciofluvial deltas of the Southern Middle Bank.

Borehole W-4 in the Pomeranian Bay (Figs. 7, 8; Appendix 1) showed, at 26.5–27.8 m b.s.l., the presence of muddy sands with peat layers. The peat base, located at 27.8 m b.s.l., was dated to 14.06 ±0.22 ka BP (Gd-2928), the peat capping located at 26.5 m b.s.l. being dated to 13.10 ±0.3 ka BP (Gd-4336) (Jurowska, Kramarska, 1990; Zachowicz *et al*., 1992; Kramarska in: Mojski *et al*., eds., 1995; Kramarska, 1998).

About 13 ka BP, the Vistula was draining into the Gulf of Gdańsk (Niewiarowski, 1987; Mojski, 1990; Starkel, Wiśniewski, 1990). Seismoacoustic profiling in the southern part of the Gulf of Gdańsk allowed to detect progradational structures resembling the Gilbert-type delta (Fig. 10). The sandy sediment top of the Vistula delta front, in its proximal part, is at present positioned at about 35–40 m b.s.l. and shows numerous traces of later erosional incisions. The prodelta sandy-silty sediments revealed in core 1 ZG 54 (Fig. 11), with organic remains, were found to contain freshwater mollusc fauna represented by *Pisidium amnicum* (Müller), *P. milium* (Held), *P. nitidum* (Jenyns) and *P. moitessierianum* (Paladilhe) (Krzymińska, 2001). The sediments were radiocarbon dated to 12.2 ±0.24 ka BP (Gd-4634) (Figs. 7, 8).

At the same time, two consecutive glaciofluvial deltas were formed on the Southern Middle Bank, as inferred from two glaciofluvial delta systems identified there (Uścinowicz in: Mojski *et al*., eds., 1995; Uścinowicz, 1996, 1999a). The top of the proximal part of the older, lower system is positioned at

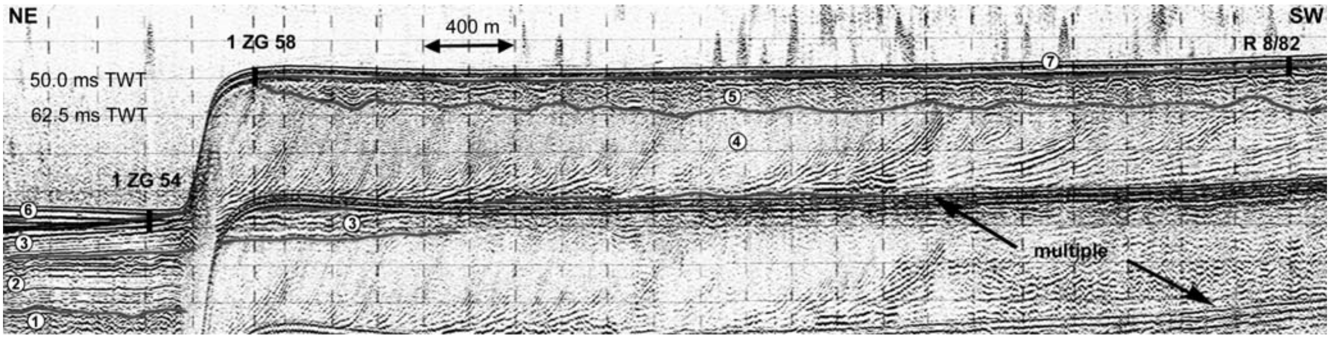


Fig. 10. The Late Pleistocene delta-front of the Vistula River on the southwestern slope of the Gulf of Gdańsk; ZG 1s seismoacoustic (sparker) profile

Pleistocene: *1* — till; **Late Pleistocene** (Bølling): *2* — clay from early phase of the Baltic Ice Lake, *3* — pro-delta silty-sandy sediments, *4* — delta front sandy sediments; **Middle and Late Holocene**: *5* — deltaic sandy sediment, locally mud and peat, *6* — marine mud (Littorina and Post-Littorina), *7* — marine sand (Littorina and Post-Littorina)

about 30 m b.s.l. The upper surface of the second, higher and shifted north by about 4–5 km, is located at about 22–25 m b.s.l. (Fig. 12). Emergence of the second, overlaying delta system is indicative of a slight sea level rise. Differences in the present position of tops of the Vistula delta front sediments in the Gulf of Gdańsk and glaciofluvial deltas of the Southern Middle Bank could most probably have been brought about by differences in the amplitude of glacio-isostatic movements between the Gulf of Gdańsk and the Southern Middle Bank. The glaciofluvial deltas of the latter are located about 100 km north from the Gulf of Gdańsk part of the Vistula palaeodelta discussed here. The Southern Middle Bank area had most probably experienced a stronger glacio-isostatic uplift than the Gulf of Gdańsk. The recent neotectonic movements in the two areas differ as well: the southern coast of the Gulf of Gdańsk subsides slightly (Wyrzykowski, 1985), while the Southern Middle Bank is situated at the borderline of the Fennoscandian uplift zone (Winterhalter *et al*., 1981). It is possible that the intensity and extent of erosional processes in the two areas differed, too, later on.

The extent of transgression that began in the southern part of the Baltic Ice Lake about 12.1–12.0 ka BP is difficult to reconstruct. It may be inferred from the location and radiocarbon datings of peat samples from the Vistula Lagoon (core ZW 3) and from the Pomeranian Bay (core R 86) that, within 12.0–11.2 ka BP, the Baltic Ice Lake level did not reach the present depth of 18–20 m (Fig. 8). At that time, tills of the last glaciation in some areas of the Southern Baltic could be eroded. The till capping, thermoluminescence-dated to 167–97.3 ka BP, found within the Słupsk Bank at the depth of about 24.0–24.5 m b.s.l., suggests that the tills might have originated during the Wartanian or early Vistulian Glaciation. The tills are directly overlain by Late Glacial clayey sediments and peats the base of which was dated to 10.51 ±0.17 ka BP (core 14 079B) (Figs. 13–15). This could be an indication of the extent of transgression, although the youngest tills in the part of the Słupsk Bank in question could have also been eroded earlier, during the meltwater drainage (Uścinowicz, 1996, 1999a).

As deglaciation progressed, a strait located north of Mt. Billingen in central Sweden was opened for the first time; this happened *c.* 11,200 years BP and made it possible for the Baltic Ice Lake to reconnect with the ocean (Svensson, 1989, 1991; Björck, 1995). As a result, the Baltic Ice Lake drained rapidly within *c.* 11.2–11.0 ka BP and its level fell by about 25 m, to become lower by about 50 m than at present (Fig. 8). The magnitude of the first regression of the Baltic Ice Lake can be estimated from the extent of progradational deltaic structures found in the southeastern part of the Gulf of Gdańsk (Fig. 16) and from barrier structures on the western slopes of the Gdańsk Basin (Fig. 17). Those structures are found down to the depth of about 65 m b.s.l., their upper surface at the depth of about 57–60 m b.s.l. being swept away by later erosional processes. Within the depth range of 57–65 m b.s.l., the distal part of the delta front is capped by silty-sandy sediments of the late phase of the Baltic Ice Lake (Fig. 16). The sea level of the first Baltic Ice Lake regression is also approximated by the lower range of erosional cut-off of the till top in the southern part of the Bornholm Basin and Słupsk Furrow, visible on seismoacoustic and geological profiles (Przezdziecki, Uścinowicz, 1989; Uścinowicz, 1989; Uścinowicz, Zachowicz, 1991a, b; Kramarska, 1991a, b) (Fig. 18). The approximations may carry a certain error, because both the barrier structures and erosional surfaces associated with the subsequent transgression could have emerged below the contemporaneous sea level.

The till top in the southern part of the Bornholm Basin and the Słupsk Furrow and the top of deltaic sediments in the southeastern part of the Gulf of Gdańsk were eroded during the subsequent sea level rise. Ice sheet readvance in Younger Dryas (*c.* 10.9 ka BP) resulted in renewed isolation of the Baltic Ice Lake (Svensson, 1989, 1991; Berglund, Björck, 1994; Björck, Svensson, 1994; Björck, 1995). This led to a considerable slow-down of regression in southern Sweden as the uplift was compensated for, in a water body isolated from the ocean, by the water table rise.

The southern coast of the Baltic Ice Lake experienced a rapid transgression at that time, its maximum occurring *c.* 10.3 ka BP (Fig. 19). The highest sea level stand did not, to be

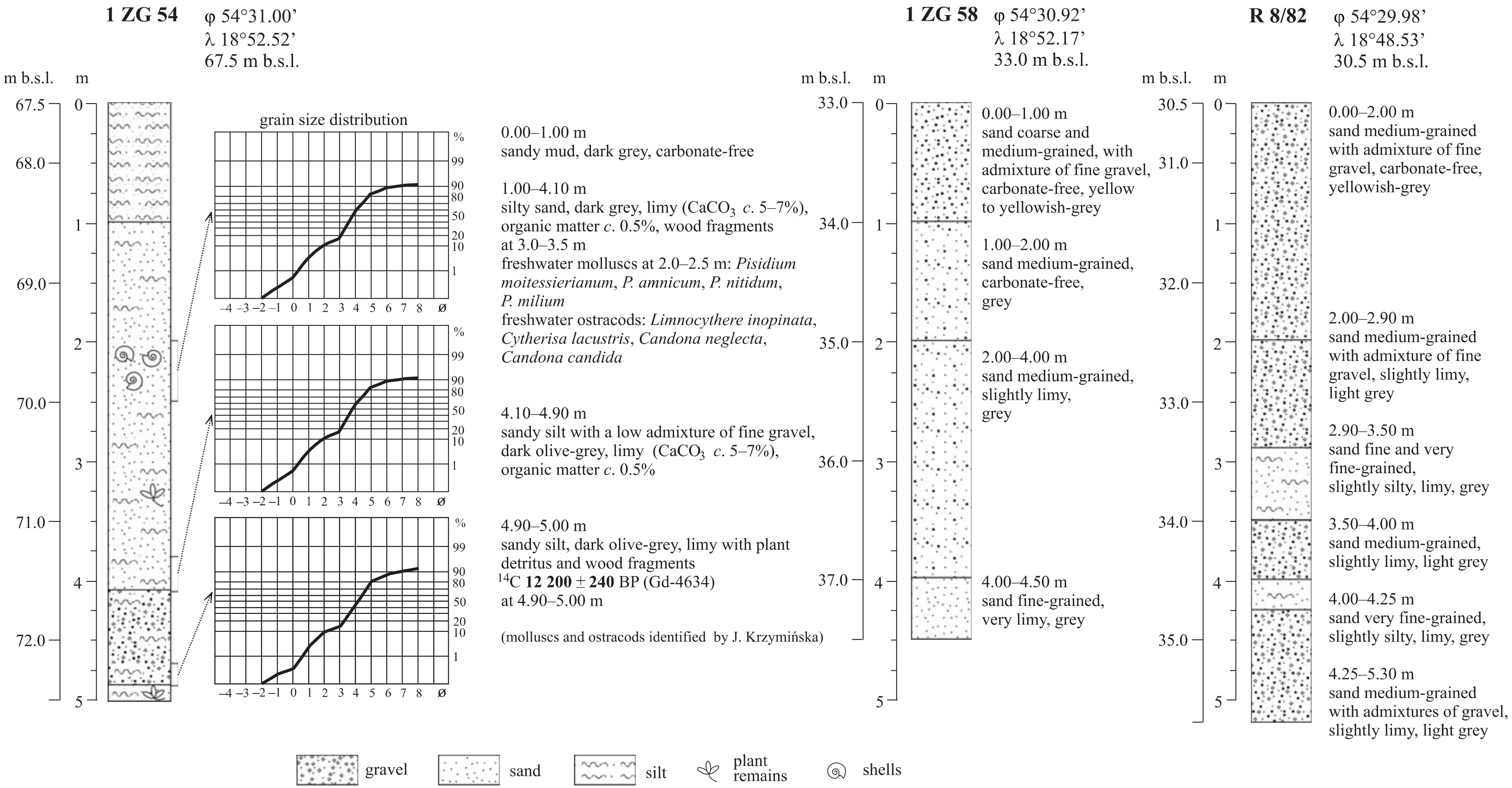


Fig. 11. The Late Pleistocene delta of the Vistula River in core 1 ZG 54 profile and other cores located on seismoacoustic profile ZG 1

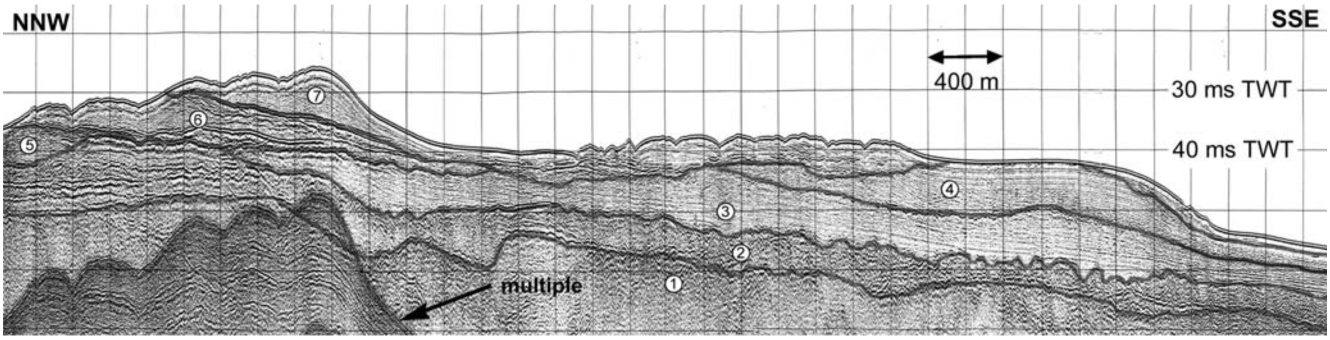


Fig. 12. The Late Pleistocene glacifoluvial deltaic system on the Southern Middle Bank; seismoacoustic (boomer) profile ŁP 18

1 — Pre-Quaternary (?); **Late Pleistocene**: *2* — till, *3* — fluvioglacial sand and silt (?), *4* — glaciofluvial delta sands and gravels (lower system), *5* — limnic sands and silts; *6* — glaciofluvial delta sands and gravels (upper system); **Middle and Late Holocene**: *7* — marine sand (Littorina and Post-Littorina); sediment origin identified from lithological (Masłowska, Michałowska, 1995) and faunistic (Krzymińska, 2001) analyses

sure, exceed 25–26 m below the present sea level, as evidenced by the position and radiocarbon dating of peat in the Słupsk Bank and in the Southern Middle Bank. The Słupsk Bank peat (core 14 097B) occurs at the depth of 24.05–24.72 m b.s.l. their age being dated to from 10.51 ±0.17 ka BP (Gd-4187) to 9.32 ±0.15 ka BP (Gd-4190) (Figs. 8, 15) (Uścinowicz, Zachowicz, 1991a, b). Core R 10 (192) from the Southern Middle Bank, at the depth of 21.05–21.10 m b.s.l. shows the presence of peat dated to 10.22 ±0.1 ka BP (Gd-10304). The site in question supplied less data than the Słupsk Bank peat did. The radiocarbon dating was obtained from a peat sample collected from the core from between sand layers of unknown age. However, even if the sample in question had been a peat ball, it could not have been transported into the area over a wide distance, from beyond the Southern Middle Bank. Thus it may be assumed that the water level rose, within the 600 years (from *c.* 10.9 to 10.3 ka BP), by about 25 m at a mean rate of about 40–45 mm/a (Figs. 6, 8).

Climate warming at the end of the Younger Dryas and ice sheet melting led, again, to opening of a connection between the Baltic Ice Lake and the ocean, via lowlands of central Sweden (Fig. 20). The most recent data (e.g. Björck, Digerfeldt, 1989; Wohlfarth *et al*., 1993) indicate this to have happened *c.* 10.3 ka BP. The drainage was very rapid, almost — as argued by some workers — catastrophic. According to Strömberg (1992), the water level fell by about 25 m during about 90 years, while Svensson (1989, 1991), Berglund and Björck (1994) and Björck (1995) contended that the drainage took as little as a few years.

The magnitude of the Baltic Ice Lake regression within the southern Baltic area can be inferred from certain evidence. For example, thick sediment layers are found at the depth range of about 30–65 m b.s.l. on the northwestern slope of the Gulf of Gdańsk; their origin can be linked to the regression–transgression–regression cycle of the Baltic Ice Lake (Fig. 17), although the type of those sediments and their purported age result from interpretation of the seismoacoustic record structure and sequence of layers. Best developed are progradational barrier structures that can be associated with the first Baltic Ice Lake regression. The surface of sand barrier sediments, tilted northwards and showing large-scale oblique layering, is overlain by parallel silty-sandy layers. Accumulation of the silty-sandy layers can be related to transgression and the high Baltic Ice Lake water stand during 11.0–10.3 ka BP. The sand layers capping the silty-sandy sediments were most likely deposited during the second (terminal) regression of the Baltic Ice Lake. In those layers, structures associated with barrier progradation are poorly visible; they are most probably masked by acoustic signal reverberation. From the depth of about 65 m, barrier sands grade into the silty-sandy facies wedging out at the depth of 70–75 m. Still deeper, at the feet of the slope, clays of the Baltic Ice Lake and till are exposed on the seafloor. The top of the barrier structure, both those originating from the terminal and from the first Baltic Ice Lake regressions as well as the upper part of the silty-sandy sediment separating them are erosionally

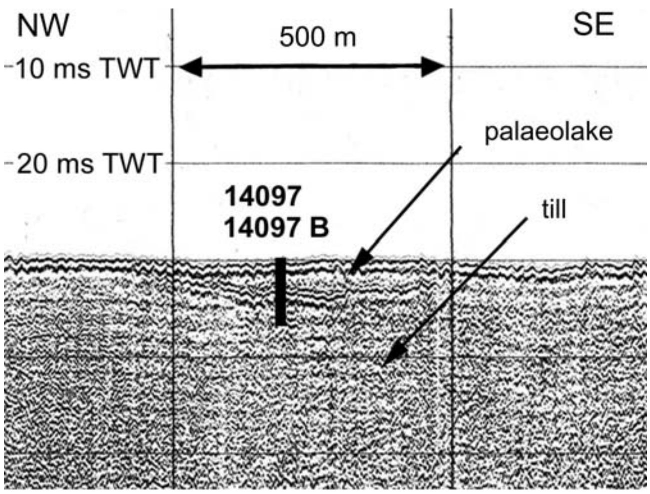


Fig. 13. The Słupsk Bank reference site: a palaeolake on seismoacoustic (boomer) profile ŁB 14

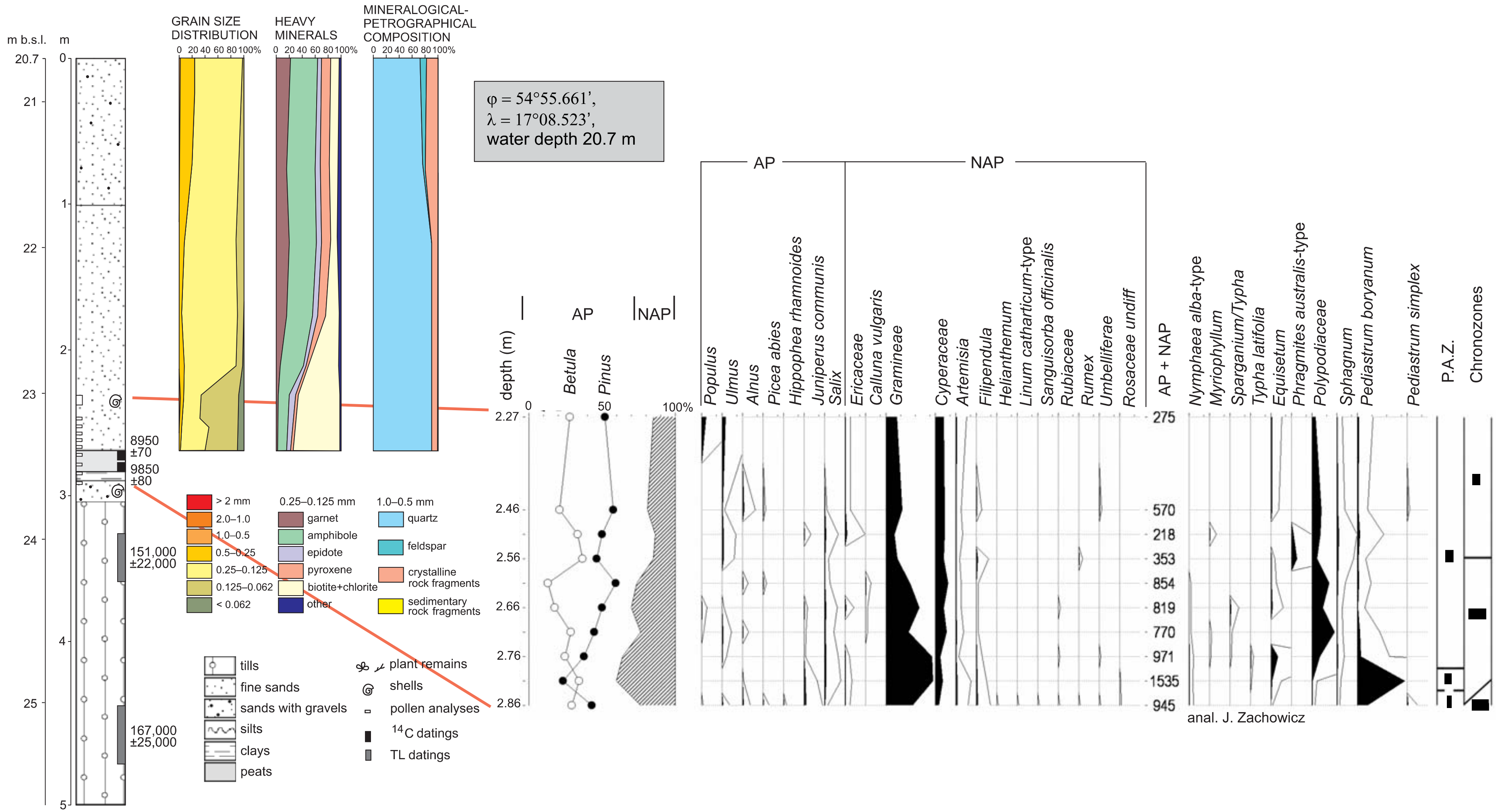


Fig. 14. The Słupsk Bank reference site: core 14 097

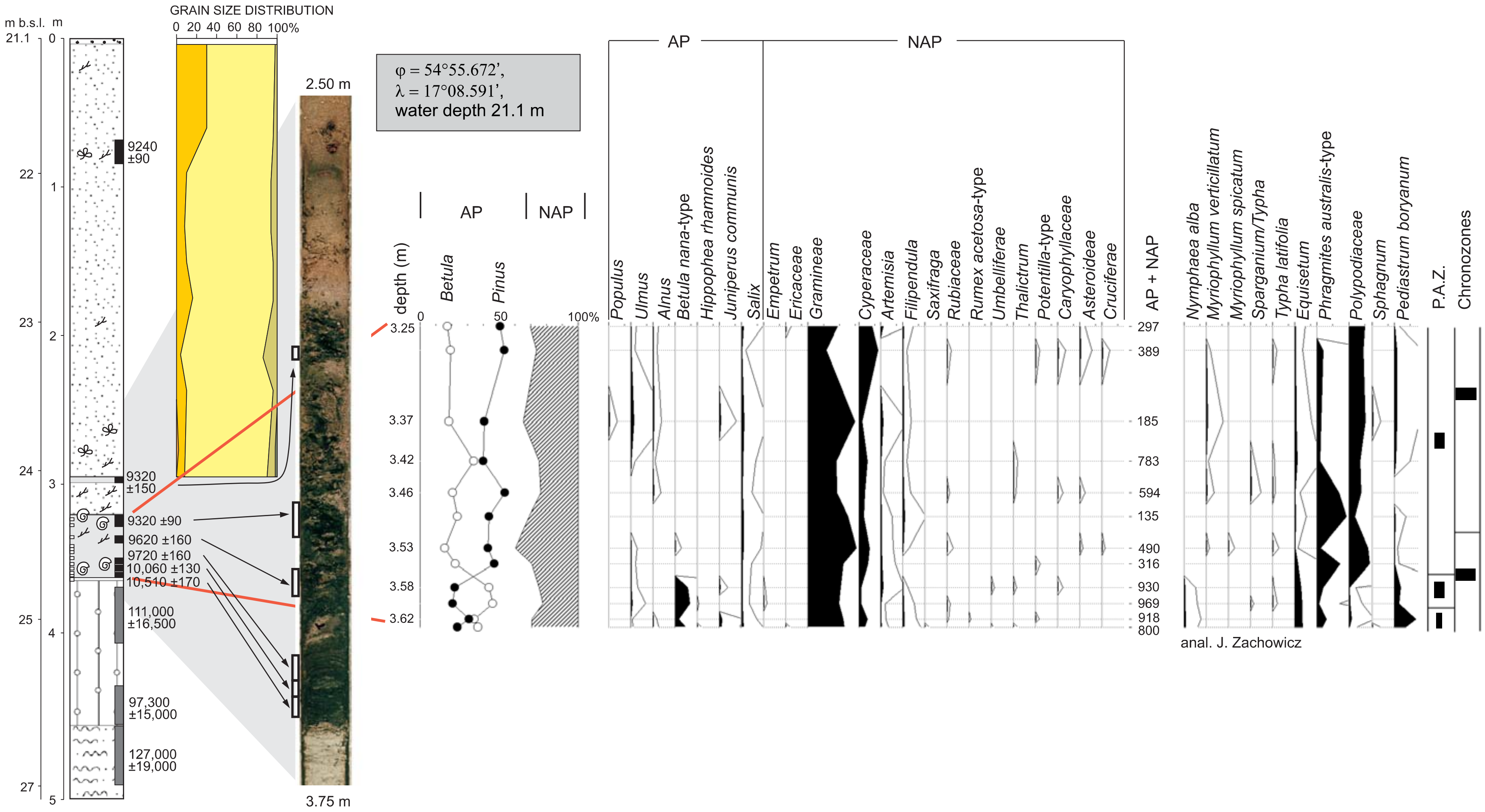


Fig. 15. The Słupsk Bank reference site: core 14 097B

For explanations see Figure 14

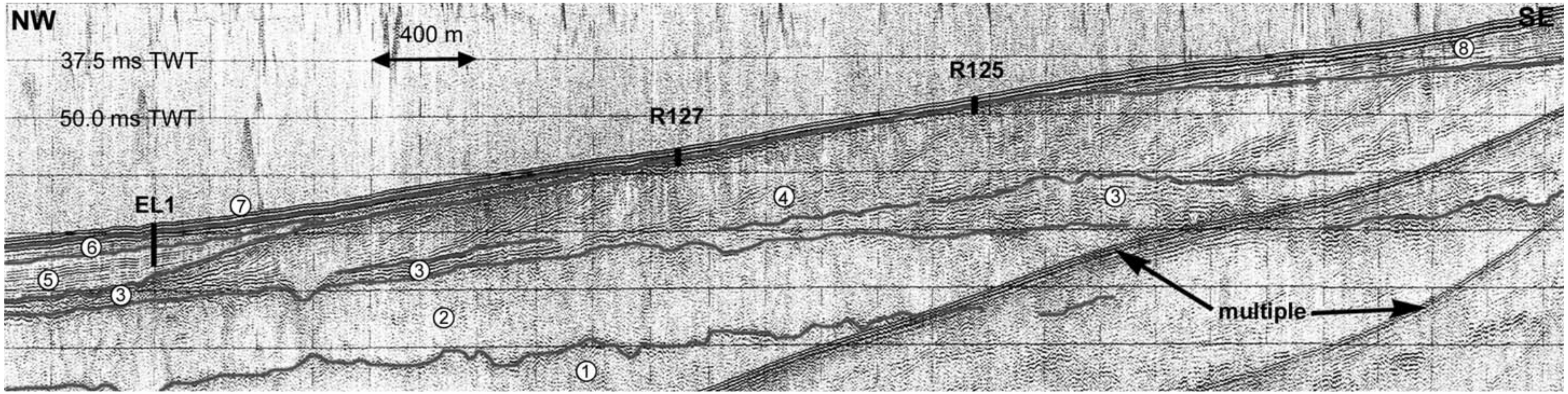


Fig. 16. The Late Pleistocene deltaic progradational structures of the Vistula (first regression of the Baltic Ice Lake) and the Late Pleistocene and Holocene sediment sequence on the southeastern slope of the Gulf of Gdańsk; seismoacoustic (boomer) profile EL 1a

Pleistocene: *1* — till; **Late Pleistocene**: *2* — clay of early phase of the Baltic Ice Lake (?), *3* — pro-delta silty (?) sediments, *4* — delta-front sandy sediments, *5* — silty-sandy sediments (late phase of the Baltic Ice Lake); **Early Holocene**: *6* — silty-sandy sediments (Yoldia Sea and Ancylus Lake); **Middle and Late Holocene**: *7* — silts and sands (Post-Littorina), *8* — barrier sands of the Vistula Spit

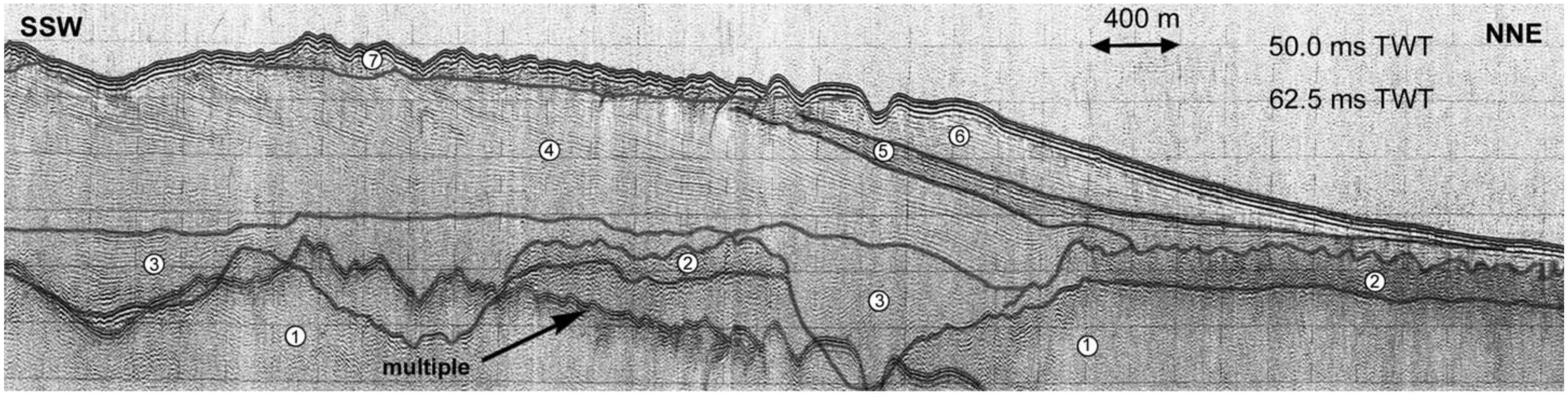


Fig. 17. The Late Pleistocene barrier system on the western slope of the Gdańsk Basin; seismoacoustic (boomer) profile GG 6

1 — Pre-Quaternary (?); **Pleistocene**: *2* — till, *3* — marginal ice lake silt and clay (?); **Late Pleistocene** (Baltic Ice Lake): *4* — progradational barrier sand (first regression), *5* — clay (transgression phase), *6* — barrier sand (final regression); **Middle and Late Holocene**: *7* — marine sand (Littorina and Post-Littorina); the erosional valley (left side of seismogram) could have been formed during the lowest Baltic stand after the first or final Baltic Ice Lake regression; erosional surface at the top of units 6 and 4 is related to an Early Holocene transgression of the Yoldia Sea

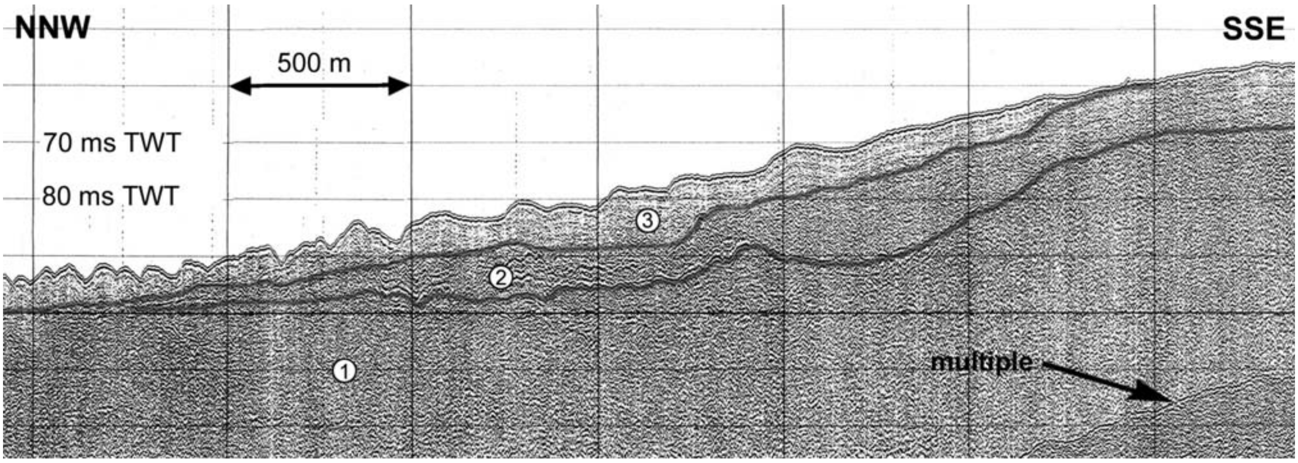


Fig. 18. The erosional surface of the subaqueous till (diamicton) and till on the southern slope of the Słupsk Furrow; seismoacoustic (boomer) profile ŁB 7

1 — Tertiary silty-sandy deposits, *2* — Pleistocene till, *3* — Late Pleistocene subaqueous till (Southern Baltic diamicton)

levelled off. This is a trace of an Early Holocene erosion that occurred during the Yoldia Sea transgression. The erosional valley visible in the left-hand part of Figure 17 could have been formed during the low Baltic Ice Lake level subsequent to both the first and the terminal regression.

In the southeastern part of the Gulf of Gdańsk, on the forefront of the Vistula Spit, an extensive, several metres thick bed of laminated silty-sandy sediments was revealed (Figs. 16, 21). The bed, recorded in a few seismoacoustic profiles, encroaches the Late Pleistocene delta front of the Vistula River (Fig. 16).

Fig. 19. The Baltic Sea area during the final phase (maximum extent) of the Baltic Ice Lake (*c.* 10.4–10.3 ka BP) (after: Eronen, 1988; Björck, Svensson, 1994; Björck, 1995; Lemke, Kuijpers, 1995; southern part changed by the present author)

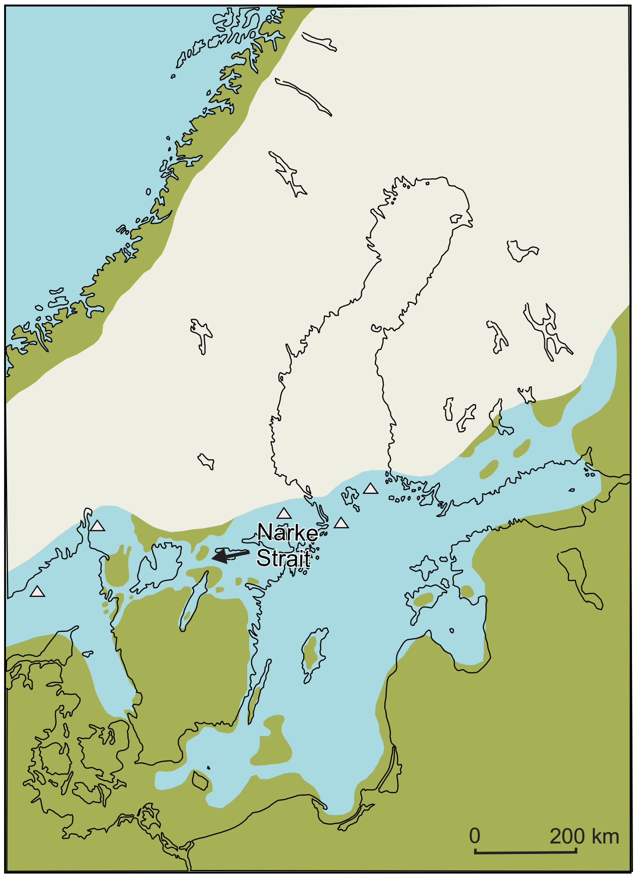


Fig. 20. The Baltic Sea area during early phase of the Yoldia Sea (*c.* 10.0–9.9 ka BP) (after: Eronen, 1988; Björck, Svensson, 1994; Björck, 1995; southern part changed by the present author)

For explanations see Figure 19

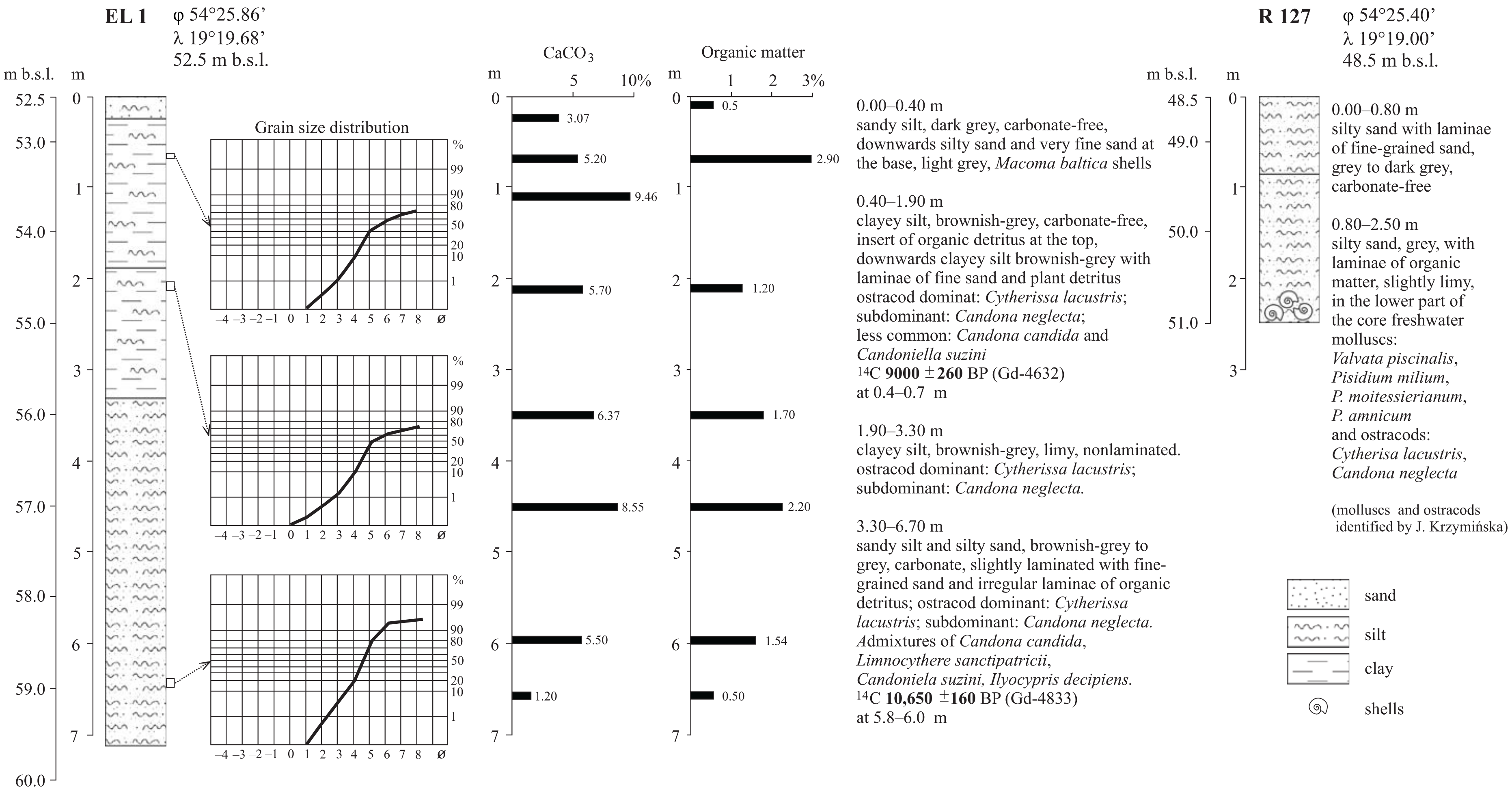


Fig. 21. The Late Pleistocene and Early Holocene sediments in core EL 1 and R 127 profiles located on seismoacoustic profile EL1a

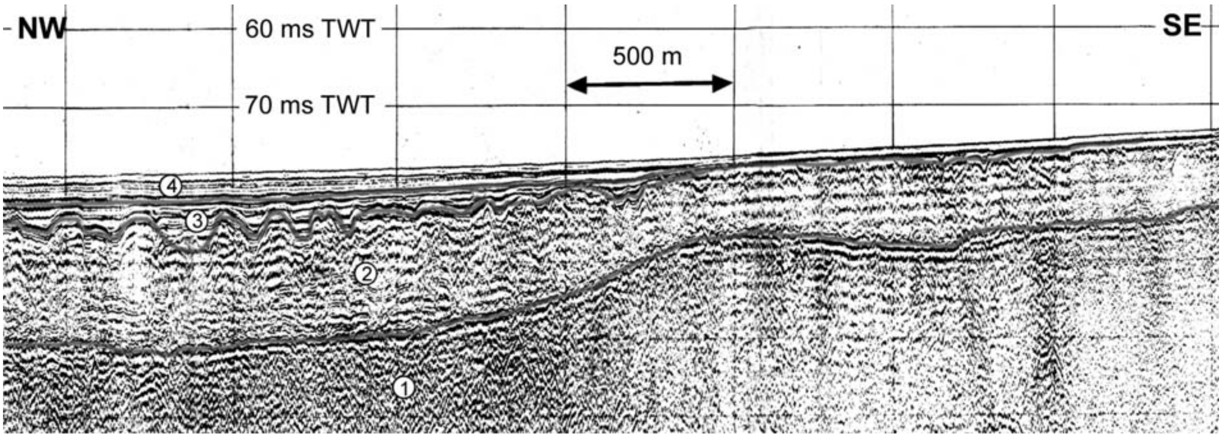


Fig. 22. The erosional surface of the Baltic Ice Lake clay on the southern slope of the Bornholm Basin; seismoacoustic (boomer) profile NE 6a

Pleistocene: *1* — till; **Late Pleistocene**: *2* — clay (Baltic Ice Lake); **Early Holocene**: *3* — clay (Yoldia Sea and Ancylus Lake); **Middle and Late Holocene**: *4* — marine mud (Littorina and Post-Littorina)

The 7 m long core EL 1 (Fig. 21), obtained from the depth of 52.5 m b.s.l., shows the presence of laminated silty-sandy sediments containing abundant freshwater ostracods, dominated by *Cytherissa lacustris* (Sars) (identified by J. Krzymińska).

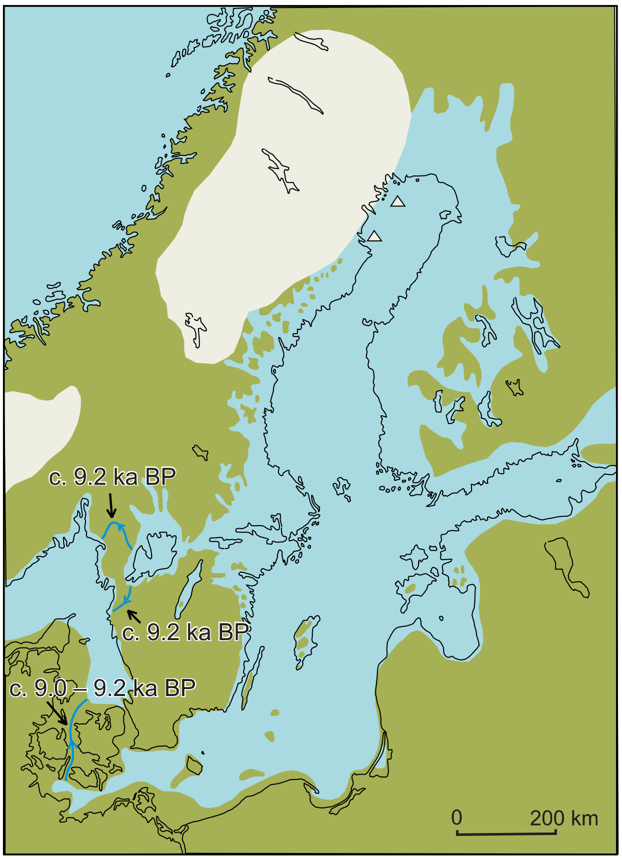


Fig. 23. The Baltic Sea area during the final phase (maximum extent) of Ancylus Lake (*c.* 9.2–9.0 ka BP) (after: Eronen, 1988; Björck, Svensson, 1994; Björck, 1995; Lemke, 1998; southern part changed by the present author)

For explanations see Figure 19

Samples collected from 5.8–6.0 and 0.4–0.7 m below the seafloor were radiocarbon dated to 10.65 ±0.16 ka BP (Gd-4632) and 9.0 ±0.26 ka BP (Gd-4833) (Fig. 21). The layers in question represent most probably Late Pleistocene and Preboreal sediments of the Baltic, deposited in the vicinity of the Vistula mouth.

The Vistula Delta Plain revealed the present of fossil Vistula channels, incised down to the level of about 30–40 m b.s.l. (Mojski, 1988, 1995a). The late-glacial deltaic Vistula sediments in the Gulf of Gdańsk contain, too, erosional incisions reaching down to about 40–45 m b.s.l. (Uścinowicz, Zachowicz, 1993b, 1994).

Similarly, erosional surfaces cutting off the capping of the Baltic Ice Lake clayey sediment, formed during a later transgression, do not extend deeper than about 60 m b.s.l. (Fig. 22).

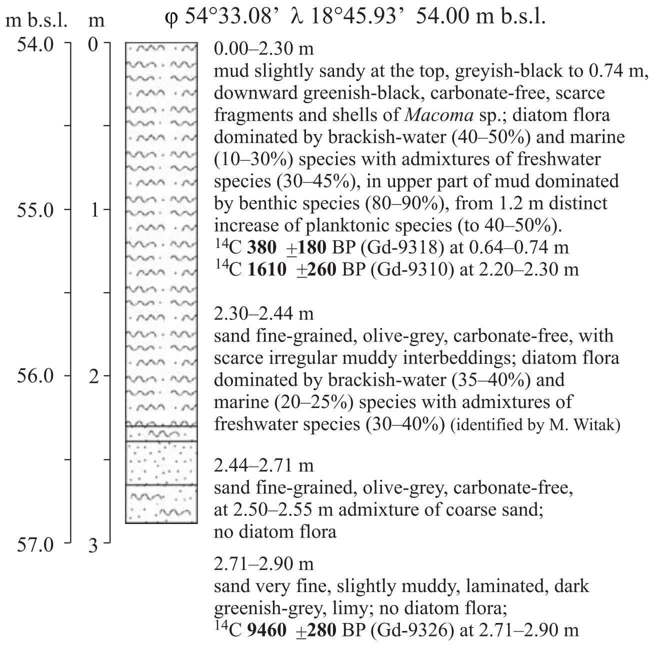


Fig. 24. The core 9/93 profile

The evidence discussed above indicates that the water level during the final drainage of the Baltic Ice Lake did not fall, in the southern Baltic area, below about 50–52 m below the present sea level (Figs. 6, 8). The evidence shows also that the magnitude of the terminal regression of the Baltic Ice Lake, as inferred for the southern Baltic area, did not exceed 25–26 m and is in good agreement with the southern Sweden data, referred to above. In the light of those facts, Rosa's (1987, 1994, 1997) contention that the Gulf of Gdańsk water table dropped to a level lower by about 100–105 than at present should be rejected. The formations visible on seismoacoustic profiles, interpreted by Rosa as cryogenic or thawing structures formed under subareal conditions, could have been formed under water. Such structures emerge in the subsea permafrost and when blocks of the buried ice remain after deglaciation (Elverhoi, Solheim, 1987; Reimnitz *et al*., 1987). Thawing structures associated with melting of the buried ice are known also from the Southern Baltic (Rudowski *et al*., 1999).

The drainage of the Baltic Ice Lake resulted in the water level fall to the ocean level, a consequence of which was a connection between the Yoldia Sea with the ocean, persisting for about 700–800 years, within 10.3–9.6 (9.5) ka BP (Svensson, 1989, 1991; Björck, 1995) (Fig. 20) and facilitating water exchange. The fast glacio-isostatic uplift of Scandinavia, faster than the eustatic ocean level rise, brought about — *c.* 9.6–9.5 ka BP — the final closure of the central Sweden straits, whereby

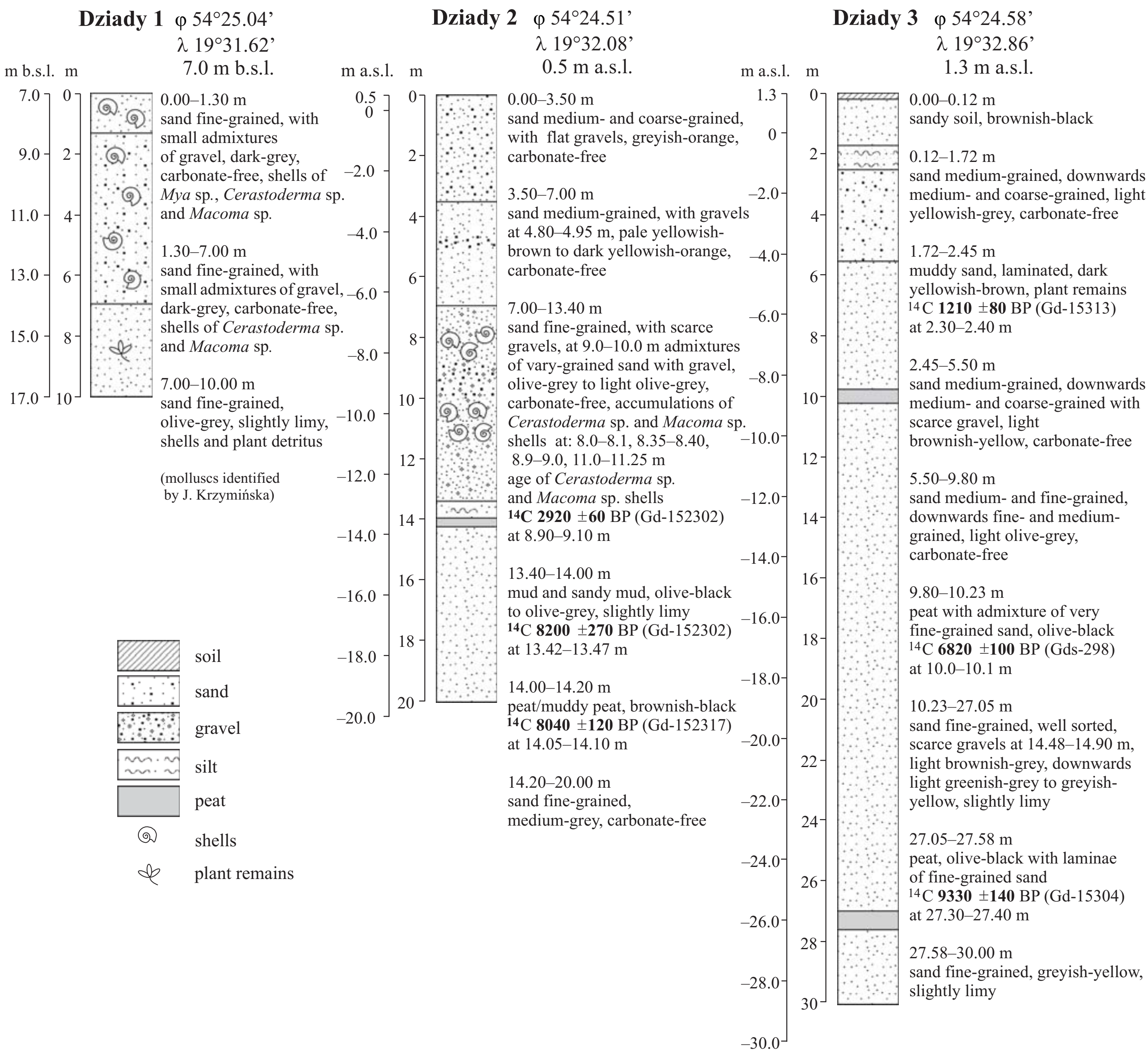
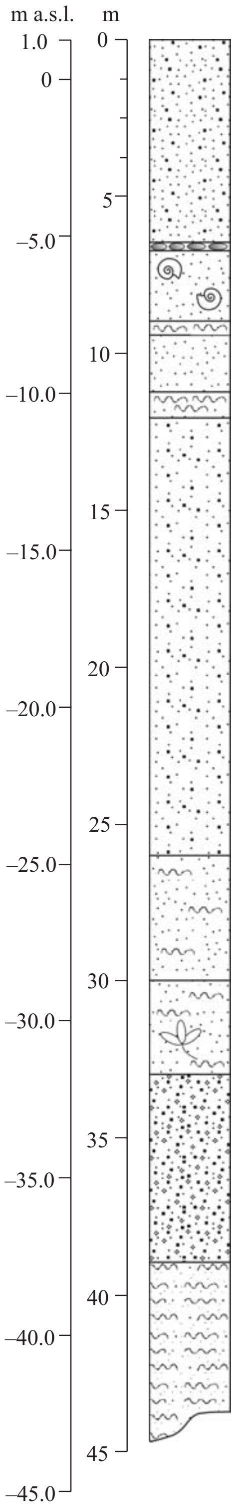


Fig. 25. The core Dziady 1, 2, 3 profiles on the Vistula Spit

φ 54°43.80’ λ 18°36.30’ 1.0 m a.s.l.



0.00–6.50 m
sand medium-grained, yellowish-grey to light grey, carbonate-free

6.50–6.60 m
cobbles (max. ϕ 8 cm).

6.60–9.00 m
sand fine- and medium-grained, grey, carbonate-free, *Cerastoderma* sp. shells, at 7.9–8.0 m fine, flat gravel and shell fragments

9.00–9.25 m
mud, organic matter and plant remains;
diatom flora dominated by marine and brackishwater species:
Campylodiscus clypeus, *C. echeneis*, *Rhabdonema arcuatum*, *Coconeis scutellum*, *Nitzschia punctata*, *Synedra tabulata*, *Rhisosolenia hebetata*, *Hialodiscus scotius*, *Cosconodiscus oculus-iridis*, *Actinocylusehrenbergii*, *Gramatophora oceanica*, *Diploneis didima*, *D. interrupta*, *D. smithii*, *Achnanthes brevipes*, *Navicula digitoradiata*, *Nitzchia trybionella*, *Rhapalodia gibberula*
with admixtures of freshwater species (identified by Zaborowska)
^{14}C **6640** ±**120 BP** (Gd-6874) at 9.00–9.20 m

9.25–11.20 m
sand fine-grained, grey, carbonate-free; diatom flora dominated by marine and brackish-water species as above

11.20–12.00 m
sandy mud, dark grey to black, carbonate-free; diatom flora dominated by marine and brackish-water species as above
^{14}C **6930** ±**120** BP (Gd-9009) at 11.8–10.9 m
^{14}C **7240** ±**130** BP (Gd-6875) at 10.9–11.0 m

12.00–26.00 m
sand medium- and fine-grained, downwards fine- and very fine-grained, grey, carbonate-free; no diatom flora

26.00–30.00 m
sand fine-grained, slightly silty and limy, grey;
no diatom flora

30.00–33.00 m
sand fine-grained, slightly silty, grey, slightly limy, plant detritus and plant remains; no diatom flora and microfauna
^{14}C **8980** ±**140** BP (Gd-9015) at 32.7–33.0 m

33.00–39.00 m
sand medium- and coarse-grained, scarce gravel

39.00–55.00 m
sandy silt, fresh water diatom flora dominated by: *Fragilaria construens*, *Amphora ovalis*, *Opephora martyi*, *Melosira islandica*;
subdominants: *Cymatopelura solea*, *C. elliptica*, *Pinularia viridis*, *Rhopalodia gibba*, *Diploneis domblittensis*, *Navicula placentula*, *N. platystoma*, *N. bacillum*
(identified by Zaborowska)

Fig. 26. The core Kuźnica 92 profile

For explanations see Figure 25

the Yoldia Sea was transformed into the Ancylus Lake (Eronen, 1983, 1988; Svensson, 1989, 1991; Björck, 1995) (Fig. 23). Once the connection with the ocean was closed, the shoreline displacement relied on vertical crustal movements. According to data from southern Sweden (e.g. Svensson, 1989, 1991; Berglund, Björck, 1994; Björck, 1995) and from western Baltic (e.g. Bennike, Jensen, 1998), the transgression peaked *c.* 9.2 ka BP. Until then, the Ancylus Lake drainage had been effected through narrow, and shallowing due to uplift, straits in central Sweden. The progressing transgression in the southern part led to the formation of a new drainage channel (the so-called Dana River) in the area of the present Belt Sea (Fig. 23). This resulted in a regression the duration of which is estimated at the period of 9.2–9.0 ka BP (Svensson, 1989, 1991; Berglund, Björck, 1994; Björck, 1995) or 9.2–8.5 ka BP (Bennike, Jensen, 1998). Incidentally, the more recent data from southwestern Baltic (e.g. Bennike, Jensen, 1998; Lemke, 1998; Jensen *et al*., 1999) show the extent of the regression to be smaller than assumed earlier (e.g. Kolp, 1979a, 1990; Svensson, 1991; Björck, 1995) and not to exceed 5 m. The regression terminated in a renewed equilibration of the water levels in the Baltic and the ocean.

Changes in the sea level during the Preboreal and Boreal (i.e. during the Yoldia Sea, Ancylus Lake and at the beginning of the Mastogloia Sea) are, in the southern Baltic area, rather poorly documented by few radiocarbon datings of formations and sediments. The sea level rise subsequent to the Baltic Ice Lake regression at the end of the Younger Dryas (the onset of the Yoldia Sea transgression) began, as already mentioned, from the depth of about 50–52 m and terminated towards the end of the Proboreal (the maximum Ancylus Lake water stand) at a depth not higher than about 25–26 m. The Gdańsk and Bornholm basins provided evidence in the form of Preboreal (pollen dated) silty-clayey sediments assigned to the Yoldia Sea deposits, containing — among the dominant freshwater species — small admixtures of brackish-water diatoms (Witkowski, 1994; Zachowicz, 1995). The Preboreal silts and silty-sandy sediments from the southern part of the Gulf of Gdańsk contain freshwater diatoms and ostracods only (Uścinowicz, Zachowicz, 1993a, b, 1994). Core 9/93 (Fig. 24) from the western part of the Gulf of Gdańsk, collected from the depth of 54 m, contains — in its lowermost part — fine-grained sands, slightly silty and carbonaceous, the radiocarbon age of which was dated to 9.46 ±0.28 ka BP (Gd-9326). This is an indirect indication that, in the Preboreal, the sediments were deposited in a shallow environment, which supports conclusions drawn from the analysis of core EL 1 (Fig. 21) regarding the Gdańsk Basin's Early Holocene water level.

The maximum water level at the turn of the Preboreal and Boreal (the maximum Anclus Lake level) is indicated by the Słupsk Bank peat. The peat capping in core 14 097, occurring 23.38 m b.s.l., was dated to 8.95 ±0.07 ka BP (Gd-3229) and the peat capping in core 14 097 B occuring 24.05 m b.s.l. — to 9.32 ±0.15 ka BP (Gd-4190) (Figs. 8, 14, 15) (Uścinowicz, Zachowicz, 1991a, b). A similar radiocarbon age, 9.33 ±0.14 ka BP (Gd-15304), was ascribed to the peat sampled from the Vistula Spit, from the Dziady 3 borehole, 26.0–26.1 m b.s.l. (Figs. 8, 25). The Ancylus Lake water level can be inferred from radiocarbon datings of two samples representing a shallow sedimentary environment in the Gulf of Gdańsk. In Kuźnica on the Hel Peninsula (Fig. 7, 8) silty sands with organic matter, dated to 8.98 ±0.14 ka BP (Gd-9015) occur at 31.7–32.0 m b.s.l. (Fig. 26). In the southern part of the Gulf of Gdańsk (Figs. 7, 8 — core ZG 1) sandy silts, most probably deltaic, containing freshwater molluscs *Valvata piscinalis* (Müller), *Pisidium nitidum* (Jenyns) and *P. moitessierianum* (Paladilhe) (Krzymińska, 2001) occur at 32.2–32.3 m b.s.l. Those sediments were radiocarbon dated to 9.22 ±0.14 ka BP (Gd-4777) (Fig. 27).

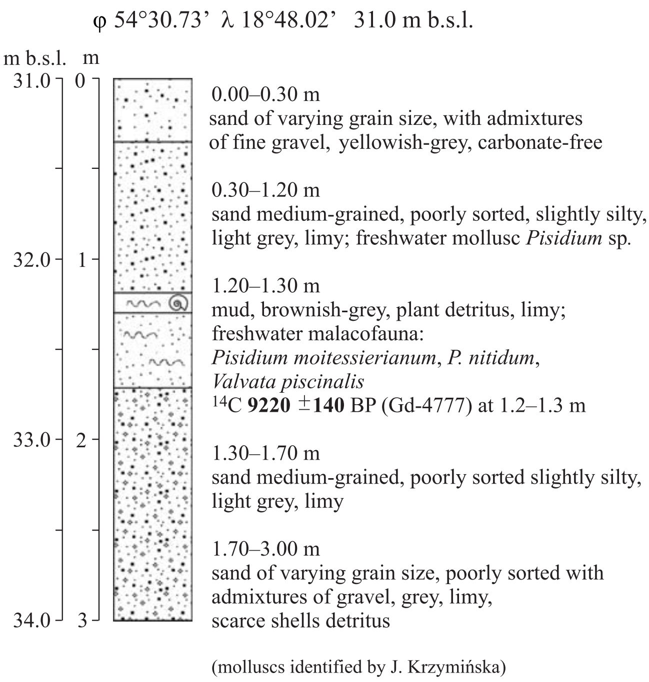


Fig. 27. The core ZG 1 profile

For explanations see Figure 25

The total water level rise in the southern Baltic area within about 1000 years, during the Yoldia Sea and Ancylus Lake transgressions, spanned about 25–27 m, the mean rate of the water level rise amount to about 25–27 mm/a. The water level rise at the time when the Baltic was connected with the ocean (the Yoldia Sea transgression), i.e. from 10.2 to 9.6 ka BP, was generated chiefly by the eustatic ocean level rise. Within 10.0–9.5 ka BP, the ocean level rose at a rate of about 20 mm/a. as estimated from the sea level curves published by Fairbanks (1989) for Barbados and by Bard *et al.* (1996) for Tahiti. A somewhat slower rate of the ocean level rise, i.e. about 16–18 mm/a, is demonstrated by curves of Edwards *et al*. (1993) and Chappell and Polach (1991) for New Guinea. According to the curve developed by Blanchon and Shaw (1995) for the Caribbean, the ocean level rose at that time at a rate of about 40 mm/a. Such a fast rise, according to the authors, reflects the pattern of catastrophic rise events during periods of intensified meltwater supply. The curves of Fairbanks (1989) and Bard *et al*. (1996), too, reveal an accelerated ocean rise rate during that time, the rate of increase being, however, lower.

Within *c*. 9.6–9.5 to *c*. 9.2 ka BP, i.e. between the closure of the Yoldia Sea's connection with the Ocean and the period of the Ancylus Lake's maximum water level, changes in the southern Baltic water level were dependent on differences in the vertical crust movements between the southern and northern Baltic. The rate of water level changes was at that time controlled to some degree by the rise of sills in the central Sweden straits (Berglund, Björck, 1994; Björck, 1995) and by the decreasing outflow.

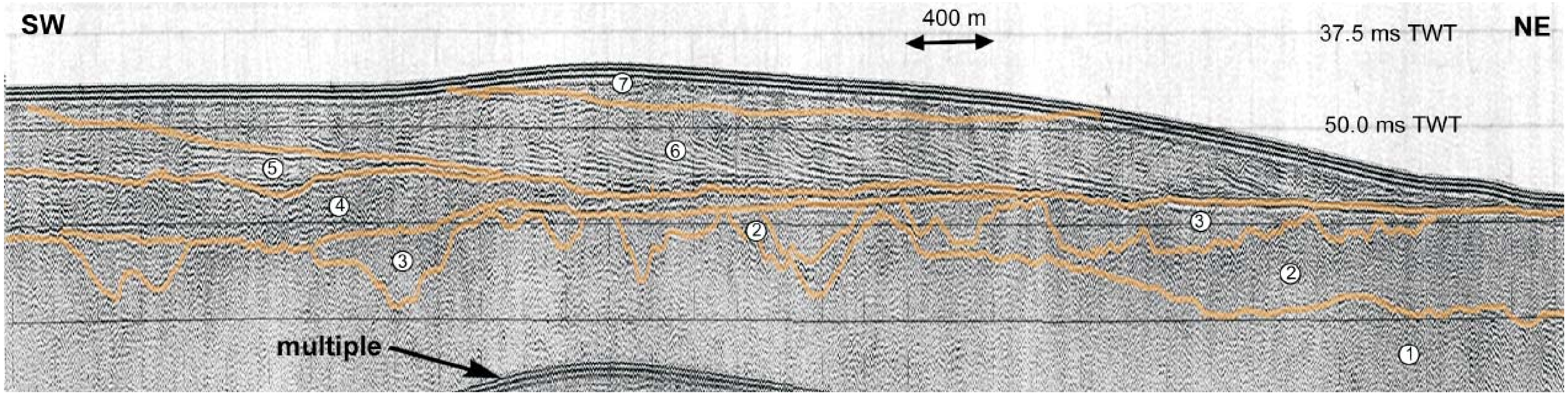


Fig. 28. Late Preboreal/Early Boreal barrier progradational structures (final regression of the Ancylus Lake); seismoacoustic (boomer) profiles KŁ1 — southern slope of the Bornholm Basin

Cretaceous: *1* — sandy silt; **Pleistocene**: *2* — till, *3* — till (?), *4* — marginal ice-lake clay (?), *5* — fluvioglacial sand (?); **Early Holocene**: *6* — barrier sand (Ancylus Lake); **Middle and Late Holocene**: *7* — marine sand (Littorina and Post-Littorina)

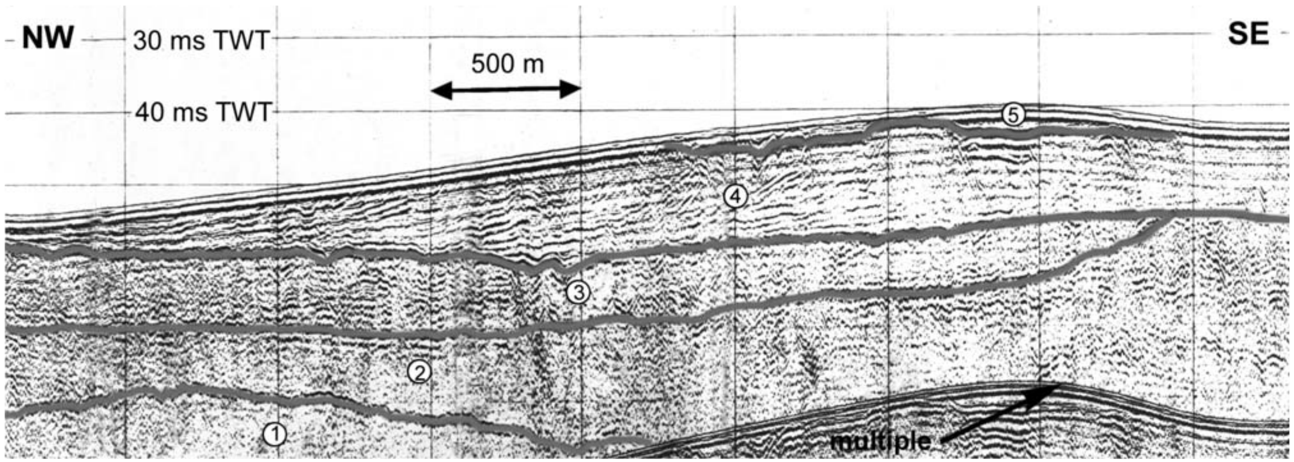


Fig. 29. Late Preboreal/Early Boreal barrier progradational structures (final regression of the Ancylus Lake); seismoacoustic (boomer) profiles NE 6 — southern slope of the Bornholm Basin

Cretaceous: *1* — sandy silt; **Pleistocene**: *2* — till, *3* — clay (Baltic Ice Lake); **Early Holocene**: *4* — barrier sand (Ancylus Lake); **Middle and Late Holocene**: *5* — marine sand (Littorina and Post-Littorina)

Bearing in mind the general pattern, described above, of the sea level rise within *c*. 10.2–9.2 ka BP in the southern coast of the Baltic, and in the light of a low number of datings pertinent to that stage in the Baltic history (Fig. 6), it is, at most, possible to estimate the sea level rise rate during the Preboreal at the Yoldia Sea stage and in the first half of the Ancylus Lake. The rate of sea level rise within 10.2–9.6 ka BP, when the connection with the ocean was open, could not have been faster than that of the eustatic ocean level rise. Most probably, the Yoldia Sea water level rose from about 40 to about 50–52 m b.s.l. at a rate of about 15–20 mm/a, i.e., slightly less than or close the rate of the eustatic ocean level rise at that time (Fairbanks, 1989; Bard *et al*., 1996). Hence, it might be concluded that the major uplift on the southern coast of the Baltic had ceased by that time, the residual uplift remaining at most. Within 9.6–9.2 ka BP (the Ancylus Lake transgression), the water level was most probably rising at a rate of about 35–45 mm/a. The inhibited drainage from the Baltic Basin as well as the glacio-isostatic crustal uplift, much faster in the northern part of the Baltic than in its southern part, resulted in the sea level rise on the southern Baltic shores being faster than the eustatic ocean level rise.

The regression occurring within 9.2–8.6 ka BP, which terminated the Ancylus Lake stage in the southern Baltic, is associated with progradational barrier structures, preserved on the southern slope of the Bornholm Basin. They were revealed at the depth of about 42–44 to about 28 m (Figs. 28, 29). Most frequently, distal parts of the barrier overlie the erosional surface of tills (Fig. 28). Locally, the barrier sands cap the clayey sediments of the Baltic Ice Lake (Fig. 29), which is an indirect indication of the progradtional structure timing. Within the depth range of about 28 to about 35–37 m, the erosional surface cutting off the progradational barrier structures is overlain by a few metres thick of younger marine sands. The location of barrier relicts and the extent of the erosional surface do not allow to draw direct conclusions as to the contemporaneous changes in water level. The thickness of eroded barrier sediments is unknown; the erosional surfaces associated with the subsequent transgression could have emerged below the water level prevalent at that time.

THE TURN OF EARLY AND MIDDLE HOLOCENE, MIDDLE HOLOCENE

No formations or sediments (radiocarbon- or pollen-dated) have been so far found in the southern Baltic that could yield direct information on the depth at which the sea level began to rise once the Baltic had finally connected with the ocean about 8.5 ka BP, at the turn of Early and Late Boreal. In view of absence of radiocarbon-dated sediments from 8.95–8.0 ka BP from depths of 25–30 m (Fig. 6), the time and scope of those events can be inferred from data concerning the western Baltic. Based on his studies on sediments from the area of Blekinge (southern Sweden), Berglund (1964) set the beginning of saline water penetration into the Baltic via the Danish Straits at Late Boreal, about 8.5 ka BP. Subsequent research in the southwestern Baltic, carried out with the use of, i.a., numerous radiocarbon datings, including those of macrophyte remains and shells dated with AMS ^{14}C (Winn *et al*., 1986, 1998; Bennike, Jensen, 1998), indicated the transgression to begin about 8.5 ka BP as well, the depth level from which the transgression began being set at about 28 m.

The oldest radiocarbon datings from brackish diatom-containing basement of marine mud in the southern Baltic deeps range from 8.8 ±0.15 ka BP (Gd-6317) in core 11 K 02/A from the Bornholm Basin to 8.75 ±0.17 ka BP (Gd-6313) in core 1 EL 38 from the Gdańsk Basin to 8.62 ±0.2 ka BP (Gd-9174) from core GT 5A/82 from the southern part of the Gotland Basin (Zachowicz *et al*., 1992; Witkowski, 1994; Kramarska *et al*., Zachowicz in: Mojski *et al*., eds., 1995). More recent data (Andrén *et al*., 2000), too, demonstrate that incursion of the saline North Sea water into the Baltic via the Danish Straits could have begun as early as about 8.8 ka BP.

The ages shown above are most probably somewhat older than the true timing of saline water penetration into the western part of the Baltic (Winn *et al*., 1986, 1998; Bennike, Jensen, 1998). The discrepancy is most likely a result of the presence of redeposited, older organic matter in the samples dated. The discrepancy between radiocarbon datings of organic matter-enriched muds and the true timing of sediment deposition has been recognised and dealt with in the southwestern Baltic (e.g. Winn *et al*., 1986; Bennike, Jensen, 1998; Andrén *et al*., 2000).

The water level at which transgression began once the Baltic had connected with the ocean can be only inferred from seismoacoustic profiles. The profiles conducted in the southeastern part of the Gulf of Gdańsk (Fig. 16) show, at the depth of about 30 m, distal parts of the Vistula Spit. It can be assumed that the depth of 30 m marks the beginning of spit accretion and migration, accompanying the rising sea level, the rise starting at about 8.5 ka BP.

The sea level rise in the southern Baltic is best documented by numerous radiocarbon datings of peats, lagoonal sediments and marine bivalve shells from the period of *c.* 8.0 to *c.* 5.0 ka BP and for depths of about 20 m b.s.l. Fewer data are available for the last 5000 years (Fig. 6). The sea level curve for the last 8000 years, plotted against the entire data set (Fig. 6) does not separate all samples from terrestrial areas from the lagoonal and marine samples. The set consisting primarily of dated samples of peat, a terrestrial formation, contains also 12 datings of lagoonal and deltaic silts as well as 5 datings of marine bivalve shells. In the case of lagoonal sediments, the reason may be sought in the presence in the samples of admixture of redeposited, older organic matter biasing the radiocarbon dating towards a period earlier that the true timing of deposition. Some lagoonal sediments contain also carbonate admixtures, which may also bias the dating due to the reservoir effect. The probability of peat samples being contaminated by younger sediments, in which case the radiocarbon would indicate a younger peat age than the true one, is much lower. In the case of marine shells, once the reservoir effect is factored in, 2 datings are no longer in conflict with the curve presented. The position of the shell datings somewhat (about 0.5 m) above the sea level curve, i.e., among terrestrial sediment samples, can also be an effect of marine sediment deposition during storm surges exceeding the average sea level of the time. Only one dating of *Cerastoderma* sp. shells deviates considerably from the curve presented. The shells, collected from the southwestern shore of Lake Jamno near Łabusz (Fig. 7), show on the contemporary sea level ordinate, the age of *c.* 6.73 ±0.08 ka BP. Not only the position of those shells is different from that of other shells of that age, relative to the present sea level, but it is in conflict with data on Lake Jamno evolution and age. The peat surface in the Jamno bottom, containing marine and brackish diatoms and pollen-dated to the end of the Atlantic, is located on the ordinate of about 5 m b.s.l. The base of muds carrying *Cardium* sp. (*Cerastoderma* sp.) shells and eu- and mesohalobous diatoms, situated at the depth of about 4.0–4.5 m b.s.l., has been dated to the turn of the Atlantic and Subboreal (Przybyłowska-Lange, 1979; Dąbrowski *et al*., 1985). In the light of the evidence presented above, it has to be concluded that the shells collected at Łabusz were not positioned *in situ*.

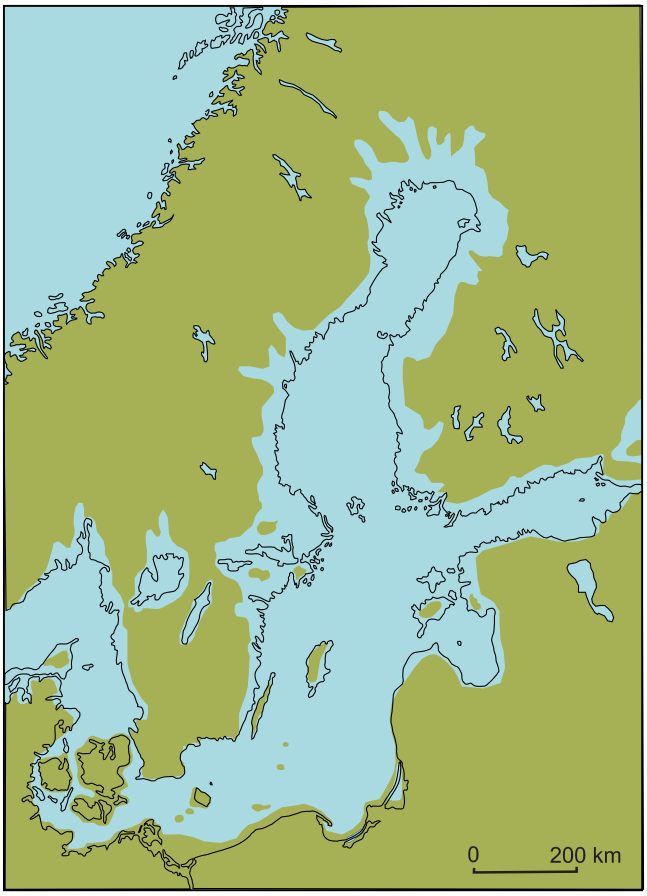


Fig. 30. The Baltic Sea area during the early phase of the Littorina Sea (*c.* 7.5–7.4 ka BP) (after: Eronen, 1988; southern part changed by the present author)

For explanations see Figure 19

Below the sea level curve, among samples of marine and lagoonal sediments and those of sea shells, there are 9 peat samples. At least 5 samples are positioned not more than about 0.5 m below the curve. The shown discrepancies could have, in part, resulted from inaccuracy in determination of the terrain ordinate or water body depth at the actual sampling site. In the case of peat samples collected from the Vistula Spit (boreholes: Jantar E-13, Komary K-2, Komary H-3 and Komary F-16 — Fig. 8; Appendix 1), positioned slightly beneath the curve, the reason may be sought in compaction and subsidence of deltaic sediments. The dated peat layers lie beneath a 9–12 m thick stratum of deltaic sediment and barrier sand. In deltaic areas, peat may form in slightly subsided areas located below the average sea level. In few cases, the discrepancies could have resulted from contamination of the samples with younger organic matter or errors introduced during the analysis.

In addition to the peat samples mentioned above, a single peat sample from the Szczecin Lagon and a single lacustrine sediment site in the Odra Bank are also positioned beneath the curve. The Szczecin Lagoon peat sample age being determined at 5386 ±0.3 ka BP is situated at the ordinate of 6.45–6.53 m below the present sea level. This is one of the first radiocarbon-dated samples in Poland and, perhaps, it may carry an error higher than the officially stated ±300 years. Perhaps the water body depth determination was not correct, either (core Z 13 was collected in 1961). It seems justified, when plotting the relative sea level curve, to ignore that sample as the remaining Szczecin Lagoon samples show a good agreement with data from other areas of the southern Baltic.

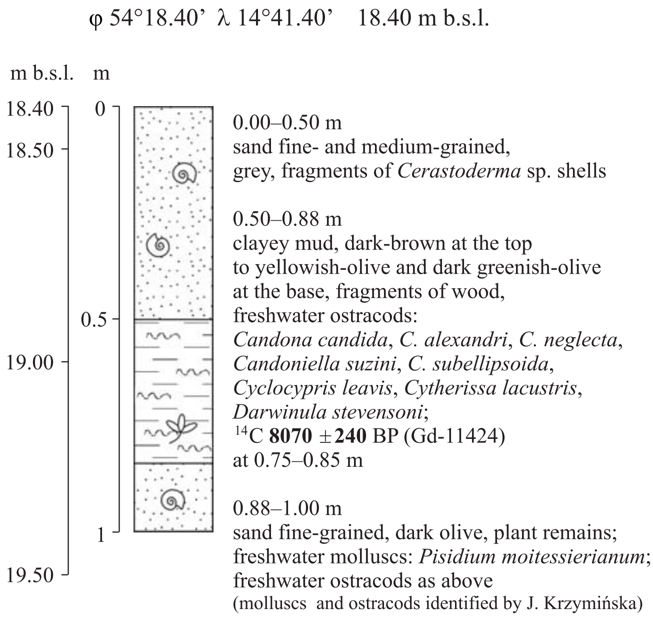


Fig. 31. The core P IX 2 profile

For explanations see Figure 25

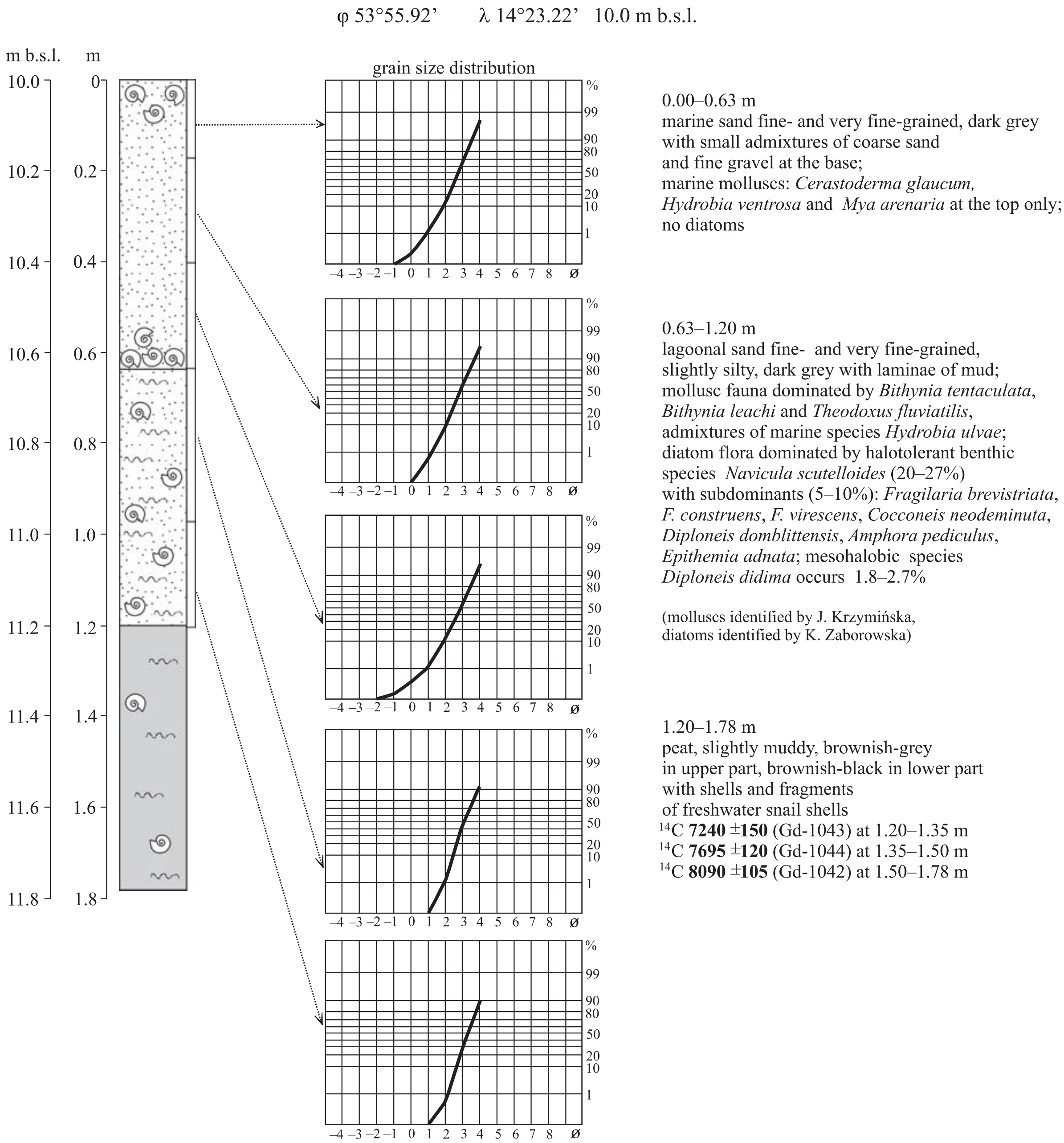


Fig. 32. The core R 74 profile

For explanations see Figure 25

In core P VI 82 collected from the Odra Bank (Fig. 7), a 78 cm thick layer of marine sand, at the depth of 12 m b.s.l., is underlain by a layer of lacustrine gyttja containing fragments of wood. Still lower, a 13 cm thick layer of silty-clayey sands was revealed, but was not cored through to the base; it was found to contain plant detritus and shells of the freshwater snails *Armiger cristata* f. *cristatus* Draparnaud and *Bithynia tentaculata* (Linnaeus) (Kramarska, 1998; Krzymińska, 2001). Both the plant remains and the carbonate fraction from the same level of lacustrine gyttja were ^{14}C-dated to 5.1 ±0.2 ka BP (Gd-9313) and 6.77 ±0.12 ka BP (Gd-10135), respectively. The two datings situate the lacustrine sediments in question below the sea level as determined for the periods by the curve discussed. The absence of seismoacoustic data from the area makes it impossible to find out if the site in question represents the *in situ* lacustrine sediments. If the datings are technically correct and the sediments are indeed the *in situ* ones, it should be assumed that the Odra Bank was, until the end of the Atlantic, an island with a lake. The site discussed would thus be a lake bottom relict, preserved after erosion of the originally elevated Odra Bank areas.

At the turn of the Boreal and the Atlantic as well as in the Early Atlantic, extensive areas of the southern Baltic were in fact a land mass (Fig. 30). The sea level 8.0 ka BP was located not higher than about 20 m b.s.l. The sea level position at that time is evidenced by lacustrine sediments from the Pomer-

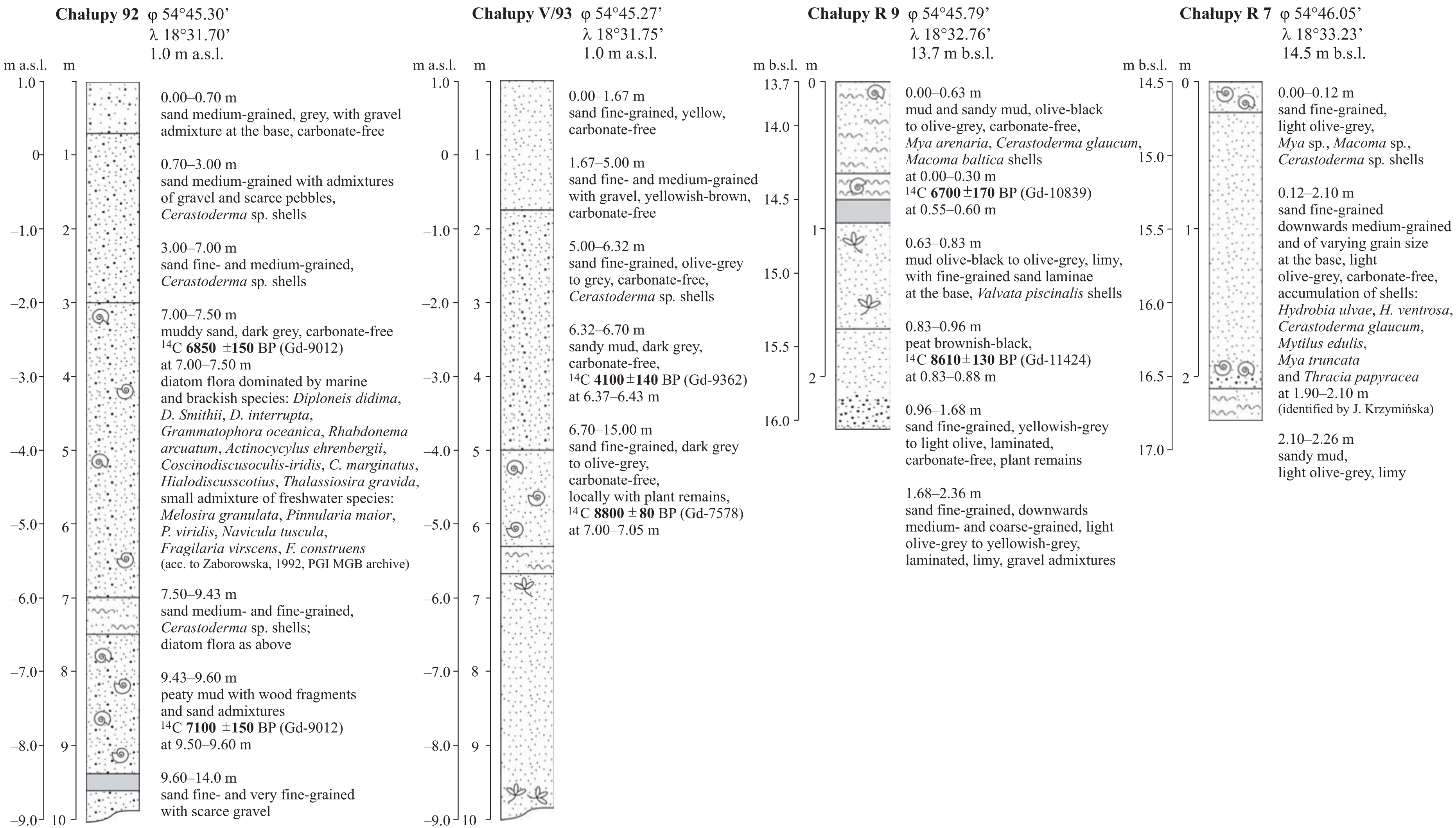


Fig. 33. The core profiles on the Hel Peninsula (Chałupy) transect

For explanations see Figure 25

anian Bay. The lacustrine mud in core P IX 2 (Figs. 7, 8, 31) is capped, at 18.9 m b.s.l., by a layer containing plant remains and freshwater ostracods. A sample of those sediments from the depth of 19.15–19.25 m b.s.l. was dated to 8.07 ±0.24 ka BP (Gd-9309) (Kramarska, 1998).

The few important references used to plot the sea level curve section for the Atlantic include, i.a., core R 74 from the Pomeranian Bay (Fig. 32) and core 1a from Lake Druzno. The first R 74 contains peat overlain by sandy-silty lagoonal sediment, this in turn being overlain by marine sand (Fig. 32). The peat capping, situated at 11.2 m b.s.l., was dated to 7.24 ±0.15 ka BP (Gd-1043) (Kramarska, Jurowska, 1991; Kramarska, 1998). The lagoonal sediments overlying the peat contain both freshwater molluscs, such as *Bithynia tentaculata* (Linnaeus), *B. leachi* (Sheppard), *Theodoxus fluviatilis* (Linnaeus) and the marine snail *Hydrobia ulvae* (Pennant). The lagoonal sediments in question contain also freshwater diatoms with an admixture of brackish-water forms. The overlying 0.63 m thick sand layer contains exclusively marine molluscs (*Hydrobia ventrosa* and *Cerastoderma glaucum*) (Krzymińska, 2001).

Lake Druzno is located south of Elbląg (Fig. 7) and, as demonstrated by biostratigraphic analyses and radiocarbon datings (Zachowicz *et al*., 1982; Zachowicz, Kępińska, 1987), has been a part of the Vistula Lagoon since the Middle Atlantic. The Lake Druzno core 1a radiocarbon dating to 6.44 ±0.05 ka BP (Gd-1131) is confirmed by the pollen spectrum indicating that the onset of brackish diatom-containing lagoonal mud sedimentation occurred in the Middle Atlantic (Zachowicz *et al*., 1982; Zachowicz, Kępińska, 1987). The sample dated was collected from the depth of 7.05–7.15 m b.s.l.

The reference sites marking the sea level curve section pertaining to 7–6 ka BP include also cores Chałupy 92 from the Hel Peninsula (Fig. 33), Dziady 3 from the Vistula Spit (Fig. 25) and ZW 14 from the Vistula Lagoon; the age of peat layers and the ordinate of peat contact with barrier sands in those cores are indicative of the contemporaneous sea level.

An important evidence for the range and level of the sea from *c*. 6.8 to *c*. 4.86 ka BP is a series of radiocarbon datings of *Cardium* sp. (*Cerastoderma* sp.) shells from drilling wells Łeba 93A and Łeba 172, located on the Łeba Barrier (Spit) (Rotnicki *et al*., 1999). Both the datings and sample location in relation to the present sea level indicate that the sea level *c*. 6.8 and *c*. 4.8 ka BP was not lower than 9 and 5.5 m, respectively (Fig. 8).

The Puck Lagoon is an area important for tracking changes in the sea level during the last 6000–5000 years and the sea level rise to close to present. Deposits from the central part of the Lagoon, as found in core 34/92 (Fig. 7, 8), contain — at the depth of 3.12 m b.s.l. — the capping of peat dated to 5.85 ±0.12 ka BP (Gd-6804) (Kramarska *et al*., 1995).

The onset of lagoonal mud deposition in the Puck Lagoon was dated to 5.48 ±0.13 ka BP (Gd-4831) (Witkowski, Witak, 1993). Those sediments (core ZP 18) (Fig. 7, 8), present at 5.15 m b.s.l., contain marine diatoms *Dimerogramma minor*, *Grammatophora oceanica*, *G. marina*, *Rhabdonema arcuatum*, *R. minutum* and *Synedea crystallina*, regarded as oceanic species. It is probable that the dating mentioned above, similarly to many other radiocarbon datings of lagoonal and lacustrine sediments, is slightly biased towards older age by the redeposited, older organic matter. In a drilling well in Jastarnia on the Hel Peninsula (Fig. 7), peat found at the depth of 2.5 m is dated to 5.37 ±0.095 ka BP (Gd-1027) (Bogaczewicz-Adamczak, 1982). The above-mentioned datings of peats and lagoonal silts in the Puck Lagoon determine the position of the sea level curve towards the end of the Atlantic (Fig. 8).

The presented curve (Figs. 6, 8) allows only to tentatively estimate the rate of the Southern Baltic level rise in the Late Boreal. Due to both an abundance of data pertaining to certain phases of the Atlantic and the range of sea level changes occurring from 8 to 5 ka BP, the relative sea level and a mean rate of its rise can be estimated with a higher accuracy for those phases (Table 1). In the Late Boreal, after the Baltic had eventually become connected with the ocean until the onset of the Atlantic, the sea level rose from about 28 to about 21 m b.s.l., the mean rate of rise being about 14 mm/a. During the Atlantic, after an initially fast rise, the sea level rise rate slowed gradually down (Table 1).

Table 1

Sea levels and rate of seal-level rise during the Atlantic

Years (ka BP)	Sea level (m b.s.l.)	Period of time (ka BP)	Sea level rise rate (mm/a)
8.0	21.0		
7.5	15.0	8.0–7.5	12.0
7.0	10.0	7.5–7.0	10.0
6.5	6.5	7.0–6.5	7.0
6.0	5.0	6.5–6.0	3.0
5.5	3.7	6.0–5.5	2.6
5.0	2.5	5.5–5.0	2.4

LATE HOLOCENE

The last 5000 years are more difficult to interpret, both due to a somewhat smaller amount of data and a narrow range of sea level changes during the Subboreal and Subatlantic. Many sites important for estimation of the subsequent sea level changes at the turn of the Atlantic and Subboreal as well as during the Subboreal and Subatlantic are located in the Puck Lagoon and on its shores (Figs. 7, 8).

In the northwestern part of the Hel Peninsula (core H 2), barrier sands are underlain, at the depth of 3.05 m b.s.l., by lagoonal muds dated to 4.22 ±0.16 ka BP (Gd-10405).

The Rzucewo Headland on the western coast of the Puck Lagoon is one of the best described sites. In several sediment cores from that site, the peat capping and the shells of marine bivalves found directly on the peat were radiocarbon- and AMS ^{14}C-dated, respectively. Additionally, pollen analyses were performed in one core (Uścinowicz, Miotk-Szpiganowicz, 2001, 2003). Core R 6/96 shows the presence, at 2.10 m b.s.l., of the peat capping dated to 5.77 ±0.17 ka BP (Gd-14025), overlain by lagoonal sands with scattered organic matter, macrophyte remains and *Cerastoderma* sp. shells. A sample of those sediments from the depth of 0.9–1.0 m b.s.l. was dated to 2.39 ±0.13 ka BP (Gd-10836). A pollen spectrum of core R1 sandy peat sample from the depth of 0.92–0.94 m b.s.l. (the uppermost organic sediment sample in core R1) showed the peat formation to have been stopped in the area during the Early Subboreal. Core Rzucewo 3/1, at the level of 0.24 to 0.34 m b.s.l., shows the presence of sands mixed with gravel and shells of *Cerastoderma* sp. and *Hydrobia* sp. on the erosional surface of peats dated to 5.78 ±0.12 – 5.83 ±0.045 ka BP (Gd-15209, GdA-159). The *Cerastoderma* sp. shells pressed into the peat top were dated to 3.56 ±0.035 ka BP (GdA-171), while those lying directly on the peat were dated to 3.435 ±0.03 ka BP (GdA-169). In addition, the Rzucewo Promontory houses a Neolithic archaeological site with a seal hunters' settlement dated to 4.4–3.7 ka BP (Król, 1997). Geological, palaeobotanical and archaeological evidence allows to conclude that *c.* 4 ka BP the sea level was by about 1.1–1.3 m lower that at present.

The Puck Lagoon data are in agreement with information extracted from the Świna Barrier (Spit). At core Mierzeja Świny p. 4, peats dated to 4.46 ±0.3 ka BP (Gd-47) occur at 1.6 m b.s.l., while at site p. 1, peats dated to 4.81 ±0.365 ka BP (Gd-45) lie at the depth of 1.5 m b.s.l. (Prusinkiewicz, Noryśkiewicz, 1966) (Fig. 8).

The sea level location during the last 1500–1000 years is evidenced at the reference sites Beka, Gizdepka and Płutnica on the western shores of the Puck Lagoon (Fig. 7, 8). Ordinates of the cores collected from Rzucewo, Beka, Gizdepka and Płutnica sites were geodesically determined to ±1 cm relative to the mean sea level. Core 2/2 collected at Beka site showed the peat capping dated to 1.225 ±0.1 ka BP (Gd-15250) to be positioned in the Lagoon's bottom at the depth of 0.4 m b.s.l. Beneath a layer of beach sand in core 1/1 collected at Płutnica site, at the ordinate range from 0.07 m a.s.l. to 0.09 m b.s.l. there are *Cerastoderma* sp. shells dated to 0.72 ±0.025 ka BP (GdA-166). In core Płutnica 2/2 at the depth of 0.37–0.45 m b.s.l. there are *Cerastoderma* sp. and *Hydrobia* sp. shells dated to 0.835 ±0.25 ka BP (GdA-167) and 0.945 ±0.025 ka BP (GdA-168), respectively. The youngest peats occurring beneath the present sea level are found at Gizdepka site. The peats at the ordinate 0.21–0.26 m b.s.l. in core 4/2 were dated to 0.655 ±0.115 ka BP (Gd-17073), their top at ordinate 0.06 m b.s.l. being dated to 0.235 ±0.1 ka BP (Gd-15255).

The presented curve (Figs. 6, 8) allows to assume that the rate of sea level rise began to decrease rapidly in the Late Atlantic. At the beginning of the Subboreal, the sea level was by about 2.5 m lower than at present, while 1000 years later it rose to about 1.1–1.3 below the present level. The rate of sea level rise from 5.0 ka BP to 4.0 ka BP continued to slow down, to an average of about 1.2–1.4 mm/a. The mean sea level 3 ka BP was not lower than about 0.70–0.90 m that at present, the mean rate of rise for the period of 4.0–3.0 ka BP being reduced to less than 0.5 mm/a.

The data presented above allow to infer that the sea level was rising very slowly during the last 3000 years or so, at a mean rate of about 0.25–0.30 mm/a, and approached the present state about 200–100 years ago. Shells of marine bivalves, occurring slightly below the present sea level or at it, dated to 3.435 ka BP (0.24–0.34 m b.s.l.) and 0.72 ka BP (from 0.07 m a.s.l. to 0.09 m b.s.l.) (Figs. 8), would then be an effect of storm surge deposition. Another possibility is that during the last 3000–4000 years, the southern Baltic was experienced alternating transgressions and regressions superimposed on the general backdrop of the rising mean sea level. The discrepancies between peat and shell datings could have then been a result of the mean sea level oscillations during the Subboreal and Subatlantic.

Sea level changes in the southern Baltic can be divided into three periods. The first period, covering Late Pleistocene and Early Holocene, was characterised by rapid and extensive fluctuations of the sea level. The second period, Middle Holocene, involved a constant rise in the sea level, the initially fast rise rapidly slowing down *c.* 6–5 ka BP. The third period, Late Holocene, showed a very narrow range of the mean sea level variations. The sea level rise was either very slow and monotonic or proceeded in stages: oscillations, the so-called transgression and regression phases, were superimposed on the overall rise.

Changes in the sea level during the first period were related primarily to deglaciation dynamics, to differences in glacio-isostatic movements between the northern and the southern parts of the Baltic, and to the connection with the ocean being alternately opened and closed. This period in the southern Baltic sea level changes cannot be compared with situation found in any other sea. Moreover, a potential for comparisons with other parts of the Baltic Sea itself is very limited at best. Each of those areas experienced their own regimes of glacio-isostatic rebound and relative sea level changes.

In the second period, beginning *c.* 8.5 ka BP when the Baltic became permanently connected with the ocean, the sea level changes in the southern Baltic depended primarily on glacio-eustatic changes in the oceanic water level. This is confirmed by a good agreement between the history, extent and rate of transgression in the southern Baltic and the history of sea level changes in other areas located close to the maximum extent of the last glaciation, beyond the zone of dominant uplift. In northwestern Europe, the period from *c.* 8.5 to *c.* 6.5 ka BP was notable by a rapid sea level rise. The deceleration of the sea level rise, beginning *c.* 6.5 ka BP, was global in its range and associated with the Laurentide ice sheet decay. The glacio-eustatic ocean level rise terminated *c.* 5.0 ka BP (e.g. Peltier *et al.*, 1978; Clark *et al.*, 1978; Jelgersma, 1979; Lambeck *et al.*, 1990; Pirazzoli, 1991, 1996; Spek, Beets, 1992; Plag *et al.*, 1996). The southern Baltic relative sea level curve for the last 8000 years shows a similarity to eustatic oceanic curves (e.g. Mörner, 1976, 1980a; Fairbanks, 1989; Bard *et al.*, 1996). It may be then concluded that glacio-isostatic movements were not particularly important for the sea level changes in the southern Baltic at that time.

During the third period, i.e. within the last 5000 years, the sea level rise was small, the rise pattern being very sensitive to all possible climatic fluctuations and neotectonic crustal movements. The relative sea level curve, plotted for the southern Baltic (Figs. 6, 8) is smoothed to represent averaged course of changes. Doubtless, the sea level changes during Holocene, particularly during the Subboreal and Subatlantic, were more complex. Most likely, sea level oscillations resulting from regional eustasy (*sensu* Mörner, 1976) and from neotectionic crustal movements not directly associated with glacio-isostasy were superimposed on the general rising trend. A more accurate reconstruction of sea level changes within the last 5000 years, including the possible eustatic and/or neotectonic fluctuations, based on formations and sediments of the southern Baltic coastal zone is problematic and prone to questioning. The sediments and formations from the period discussed, found in the field and described in the literature, provide evidence of possible small variations in the sea level or may have resulted from short-term extreme events such as, e.g. catastrophic storms breaking the barriers, salt water intrusions into lagoons and formation of appropriate sediments and shore formations.

VERTICAL CRUST MOVEMENTS

GLACIO-ISOSTATIC REBOUND: AN OUTLINE OF THE PROBLEM

The ice sheet weight imposed on crustal plates induces gravitational disturbances in the crust and in the astenosphere mass transfer, manifested as the crustal subsidence. The maximum subsidence occurs in the ice sheet centre. In front of the ice sheet, the crust is uplifted to form the so-called forebulge. Once the weight has been removed, i.e., once the ice sheet has decayed, the surface deformations are levelled off through crustal uplifts proceeding at a progressively decreasing rate. The crustal uplift begins as early as during the ice sheet meltdown. As the ice cover becomes thinner, the uplifting movement, initially slow, picks up during and immediately after ice sheet retreat, to slow down again later on. The subsidence relaxes at the earliest in the marginal part of the ice sheet-covered area, while the disequilibrium affects the central part of the ice sheet-covered area for a long time to come. Simultaneously, the proglacial forebulge vanishes slowly; not only is the forebulge lowered down, but it also migrates towards the ice sheet centre. Following the vanishing and migration of the forebulge from the ice sheet forefront, the area begins to subside. The subsidence is interpreted to be a result of submergence due to collapse of the proglacial forebulge in response to glacio-isostatic uplift of glaciation centres. The causes, history, and physical mechanisms of crustal movements associated with evolution and disappearance of ice sheets have been described and discussed in numerous publications (e.g. Daly, 1934; Peltier, 1974; Starkel, 1977; Peltier *et al*., 1978; Mörner, 1979, 1980b, 2001b; Liszkowski, 1975, 1993; Dawson, 1992).

Authors of a number of studies (e.g. Sauramo, 1958; Kolp, 1979a, b; Ignatius *et al*., 1981; Eronen, 1983, 1988) assumed that areas located more to the south were not subjected to glacio-isostatic uplift and that the zero isobase at different evolutionary stages of the Baltic was present in the southern Baltic. Reconstructions of the total Fennoscandian uplift are, too, based on the assumption that areas located north of that line were uplifted (e.g. Mörner, 1979; 1980b, 2001b; Balling, 1980; Eriksson, Henkel, 1994).

Problems of neotectonic crustal movements, including postglacial glacio-isostatic rebound in the Polish Lowland and in northeastern Poland, have been dealt with more or less directly in the Polish literature (e.g. Rühle, 1968, 1973; Baraniecka, 1975, 1979; Liszkowski, 1975, 1980; Niewiarowski, 1983; Bażyński *et al*., 1984; Rotnicki, 1987; Marks, 1988; Graniczny, 1994; Ber, 2000).

An attempt to reconstruct the history of glacio-isostatic rebound on the southern Baltic coast, by comparing the relative sea level curve with that of ocean eustasy, was made by Rosa (1968). At that time, there were no detailed data concerning the absolute chronology of events within the Baltic area, both during periods of isolation from and connection with the ocean and concerning relative sea level changes in the southern Baltic. In addition, knowledge on eustatic changes of the ocean level during the last glacial cycle was only beginning to be advanced (Fairbridge, 1961; Shepard, 1963). Rosa (1968) found that the results he obtained provided a very tentative information on glacio-isostatic rebound of the southern Baltic land mass; he argued that the data showed it to be uplifting (by about 40 m) in the Late Glacial and to subside, initially rapidly and slowly later only, in Holocene.

The problem of glacio-isostatic rebound of the southern Baltic area was also tackled by Liszkowski (1975). Basing his analysis on Rosa's (1968) results, he contended that intensive uplift occurred on the southern Baltic coast from 15 to 12 ka BP. Subsequently, the uplift slowed down somewhat and *c.* 10 ka BP turned into subsidence that stopped *c.* 6–3 ka BP. Liszkowski (1975) estimated the total uplift extent at about 140 m, the maximum uplift rate *c.* 13 ka BP being estimated at 50 mm/a. Zuchiewicz (1995a) regarded a quantitative analysis of glacio-isostasy during Pleistocene glaciations as one of the most important problems that should be solved in Poland.

Using geophysical modelling, Fjeldskaar (1994) demonstrated the area of northern Poland and southern Baltic to have been subjected to uplift during and following deglaciation. He presented quantitative data with respect to timing and extent of the uplift. He located the zero uplift line in northern Poland, the location of the line being then relatively stable temporally. He also demonstrated that the proglacial forebulge collapsed gently and did not migrate towards the glaciation centre.

GLACIO-ISOSTATIC MOVEMENTS IN THE SOUTHERN BALTIC

History of glacio-isostatic movements in an area can be reconstructed by comparing the relative sea level course with that of eustatic changes in the ocean. General principles of separating eustatic and glacio-isostatic components from sea level curves were proposed by, i.a., Mörner (1976, 2001a). This method of glacio-isostatic movement reconstruction makes it possible to obtain an averaged (smoothed) model for the entire southern Baltic area, comprising both the Pre-Cambrian platform and the Palaeozoic one.

In 1989–1996, a number of eustatic sea level curves, developed from radiocarbon and uranium-thorium datings of submerged coral reefs were published. The best evidence supports the curves for Barbados (Fairbanks, 1989; Bard *et al*., 1990; Blanchon, Shaw, 1995), Tahiti (Bard *et al*., 1996) and New Guinea (Chappell, Polach, 1991; Edwards *et al*., 1993). Sea level curves of the tectonically stable area of the southeastern Asian shelf, based on radiocarbon dating of peats and terrestrial plant remains were published as well (Hanebuth *et al*., 2000). All those curves show a generally similar pattern of sea level changes after the last glacial maximum. They, however, cover different time periods. Moreover, the resolution of sea level changes differs between various periods of time. In addition, the curves concern areas differing in their tectonic activity, hence — among other reasons — the differences between various curves.

The New Guinea curves (Chappell, Polach, 1991; Edwards *et al*., 1993) refer to a tectonically active area; consequently, in spite of factoring in the average crustal movement rate, the sea level change record may be distorted by complex tectonic movements. The curves in question encompass the shortest time period, from *c*. 11.0 to *c*. 6.1 ka BP. The Tahiti curve (Bard *et al*., 1996), although plotted for an area that is rather tectonically stable, covers only the period from *c.* 11.8 to *c.* 2.9 ka BP, the ocean level changes within 6.0–2.9 ka BP being rather poorly resolved. The curve of Bard *et al*. (1990) for Barbados, based on uranium-thorium datings, stretches to 8 ka BP only; similar is the temporal extent of the curve published by Hanebuth *et al*. (2000).

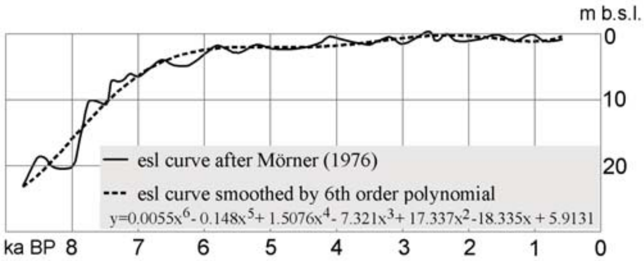


Fig. 34. The eustatic sea level (esl) curve for the Kattegat area (after Mörner, 1976) smoothed by 6th order polynomial

In the present work, the glacio-isostatic movements in the southern Baltic area were reconstructed based on the relative sea level curve developed for the southern Baltic (Fig. 8) and the eustatic ocean level curves (Fairbanks, 1989; Blanchon, Show, 1995). The latter curves were plotted from radiocarbon (Fairbanks, 1989) and radiocarbon and uranium-thorium (Blanchon, Show, 1995) datings of *Acropora palmata* (Lamarck) corals from the Caribbean. *A. palmata* (Lamarck) occurs down to 5 m depth at the deepest. The maximum growth rate, up to 14 mm/a, is one of the fastest among corals. The location and age of the submerged *A. palmata* -dominated reefs allow to infer the sea level with the accuracy of ±2.5 m. The Blanchon and Show's (1995) curve is located somewhat higher, above the reference points, because the corals grow below the water surface. The authors quoted found the sea level rise to have been higher that the maximum growth rate of *A. palmata* during three periods of time, which was reflected on the curve as sharp inflexions. The two eustatic curves were used in comparisons with the Baltic relative sea level curve because they cover the longest period of time and are well referenced, the differences between them stemming primarily from interpretative assumptions adopted by the respective authors. In addition, for the period of the last 8000 years, the eustatic curve of Mörner (1976) for Kattegat, an area closest to the southern Baltic, was used as well. In addition to the long-term increasing trend in the ocean level rise, the eustatic curve of Mörner (1976) shows also short-term regional eustatic oscillation. To render the curve of Mörner (1976) comparable with the ocean eustatic curves, it was smoothed out using a 6th order polynomial (Fig. 34).

For Late Pleistocene and Early Holocene, the curves can be compared for short periods of time only, when the Baltic was connected with the ocean and the water levels were even (Fig. 35), i.e. for the periods of *c.* 12.5–12.4, 11.0–10.9, 10.2–10.1 and 9.7–9.6 ka BP. For the period from 8.5 ka BP, when the Baltic became permanently connected with the ocean, the difference between the southern Baltic relative sea level curve and the eu-

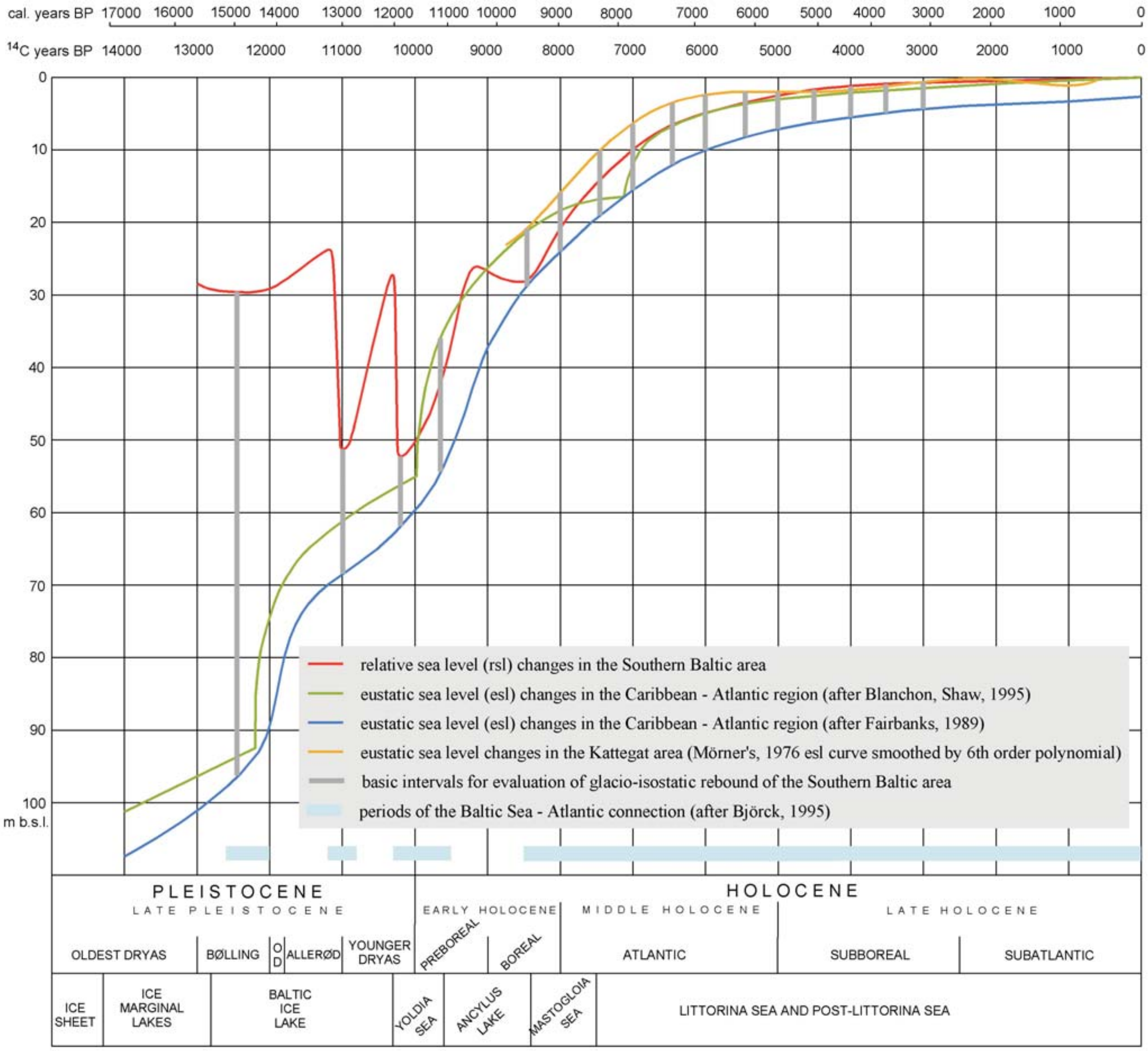


Fig. 35. Comparison of the southern Baltic relative sea level curve and ocean eustatic changes curves

static ocean level curve was determined for 500-year intervals. The curves were compared until 3 ka BP, i.e., until the southern Baltic relative sea level curve crossed the eustatic curves of Blanchon and Show (1995) and Mörner (1976), the eustatic curve of Fairbanks (1989) differing by –2.5 m from the remaining ones (Fig. 35).

Results of the comparison between the curves mentioned above (i.e. the differences between the relative sea level curve for the Southern Baltic and the eustatic ocean level curves) are shown in Figure 36. Because three different eustatic curves were used in the comparison with the southern Baltic relative sea level curve, the glacio-isostatic rebound curve was plotted using mathematical approximation. The best fit was produced by a curve described by the 5th order polynomial (Fig. 36).

The total glacio-isostatic rebound curve for each area, covering the period of glaciation and post-glaciation, consists of three parts (Dawson, 1992): the initial section relating to the onset of deglaciation the so-called restrained rebound; post-glacial uplift; and the final phase of residual uplift. The curve presented (Fig. 36) reflects only a fragment of the history of the post-glacial glacio-isostatic rebound of the southern Baltic area: a part of the essential uplift and the residual uplift phase. Reconstruction of the entire post-glacial rebound history in the southern Baltic from comparisons between the eustatic

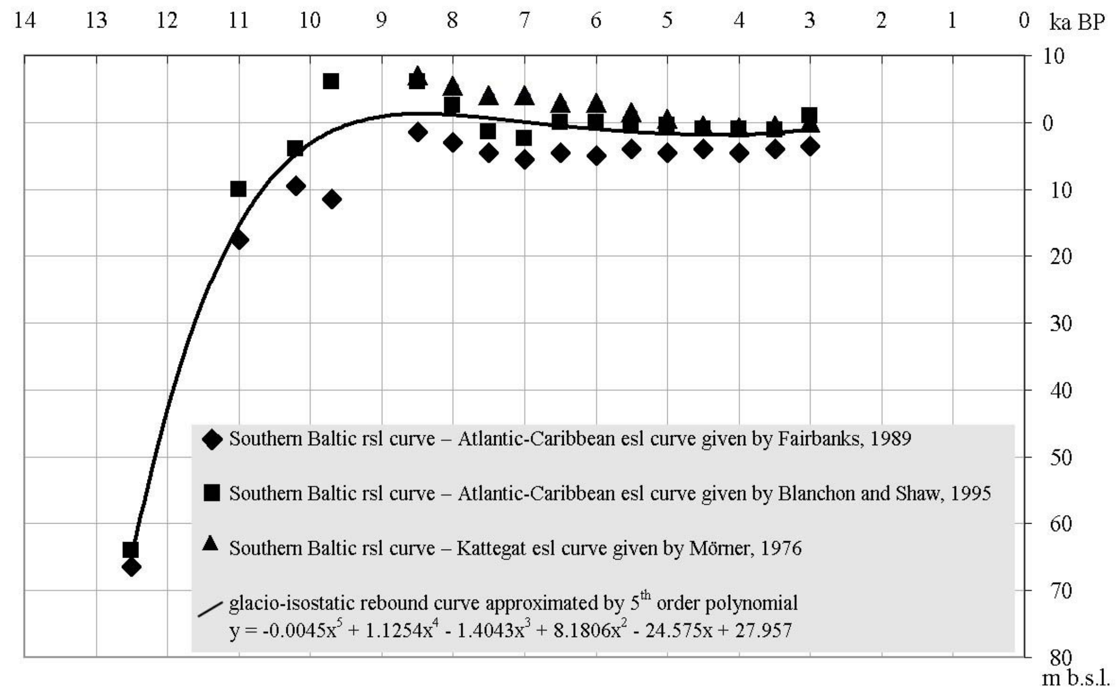


Fig. 36. The result of subtracting the southern Baltic rsl curve and ocean eustatic curves

curves and the relative sea level curve is impossible simply because the Baltic had not existed before a certain point in time. Reconstruction of the entire glacio-isostatic rebound curve for the southern Baltic is possible only through extrapolation of a defined section of the curve, from knowledge on general patterns of glacio-isostatic movements (e.g. Walcott, 1970, 1972; Peltier *et al*., 1978; Mörner, 1980b, 2001b; Eronen, 1987; Lambeck *et al*., 1990; Dawson, 1992) and knowledge of deglaciation history of in northern Poland and southern Baltic (e.g. Kozarski, 1986; Mojski, 1993, 1995; Rotnicki, Borówka, 1995a, b; Uścinowicz, 1996, 1999a).

When developing the glacio-isostatic rebound curve for the southern Baltic it was assumed that:

1. The uplift began about 17.5–17.0 ka BP during retreat of the ice sheet at the Poznań Phase. Not only did the ice sheet front receded, but the thickness of the ice sheet far away from its forefront decreased.

2. The major uplift of the southern Baltic area began earlier than 12.5 ka BP, as early as during the onset of southern Baltic deglaciation *c.* 14 ka BP.

3. When the ice sheet began to retreat *c.* 14 ka BP, the southern Baltic coast must have been higher than the contemporaneous ocean level; otherwise the drainage of meltwater would not have been possible. The ocean level at that time must have been about 107–108 m below the present level (Chappell, 1987; Fairbanks, 1989).

The curve depicting the total glacio-isostatic rebound of the southern Baltic area, combining the part calculated for the period of 12.5–3.0 ka BP and the part extrapolated, under the assumptions listed above, for the period of 17.5–12.5 ka BP is shown in Figure 37. The curve served to calculate the glacio-isostatic movement rate (Fig. 38). The rates for the period 12.5–3.0 ka BP are described by the first derivative of the equation describing vertical movements of the southern Baltic area at that time (the 4th order polynomial).

For the period of 17.5–12.5 ka BP, the rates were derived from appropriate sections of the curve depicting vertical dislocation of the southern Baltic area as a function of time (Fig. 37).

Figures 37 and 38 illustrate glacio-isostatic movements of the southern Baltic. Three major phases in the glacio-isostatic rebound can be isolated. The restrained rebound phase began about 17.5 ka BP and lasted until about 14.0 ka BP. The southern Baltic area became uplifted at that time by about 20 m. Initially, from 17 to 16 ka BP, the uplift proceeded at low rates of about 1.0 to 3.0 mm/a. As the ice sheet receded and its thickness decreased, the uplift picked up to continue, from 15.0 to 14.0 ka BP, at rates of about 7.0 to about 14.0 mm/a. The postglacial uplift phase lasted from *c.* 14.0 to *c.* 11.0 ka BP. At that time, the southern Baltic area rose by a further 85 m or so. During deglaciation of the southern Baltic area, the uplift rate increased to a maximum of about 45 mm/a within *c.* 12.4–12.2 ka BP, i.e., soon after the ice sheet retreat of the Southern Middle Bank Phase. As deglaciation proceeded, *c.* 11.0 ka BP the uplift rate slowed down to about 16.0 mm/a. The residual uplift phase began *c.* 11.0 ka BP to terminate *c.* 9.2–9.0 ka BP. The area was rising at that time at the slowest rate, the uplift extent being relatively small (by about 15 m). The uplift rate decreased to about 5.5 mm/a 10.0 ka BP. The uplifting movements stopped *c.* 9.2–9.0 ka BP. The rapid termination of post-glacial uplift within 11.0–9.0 ka BP resulted most probably from the re-

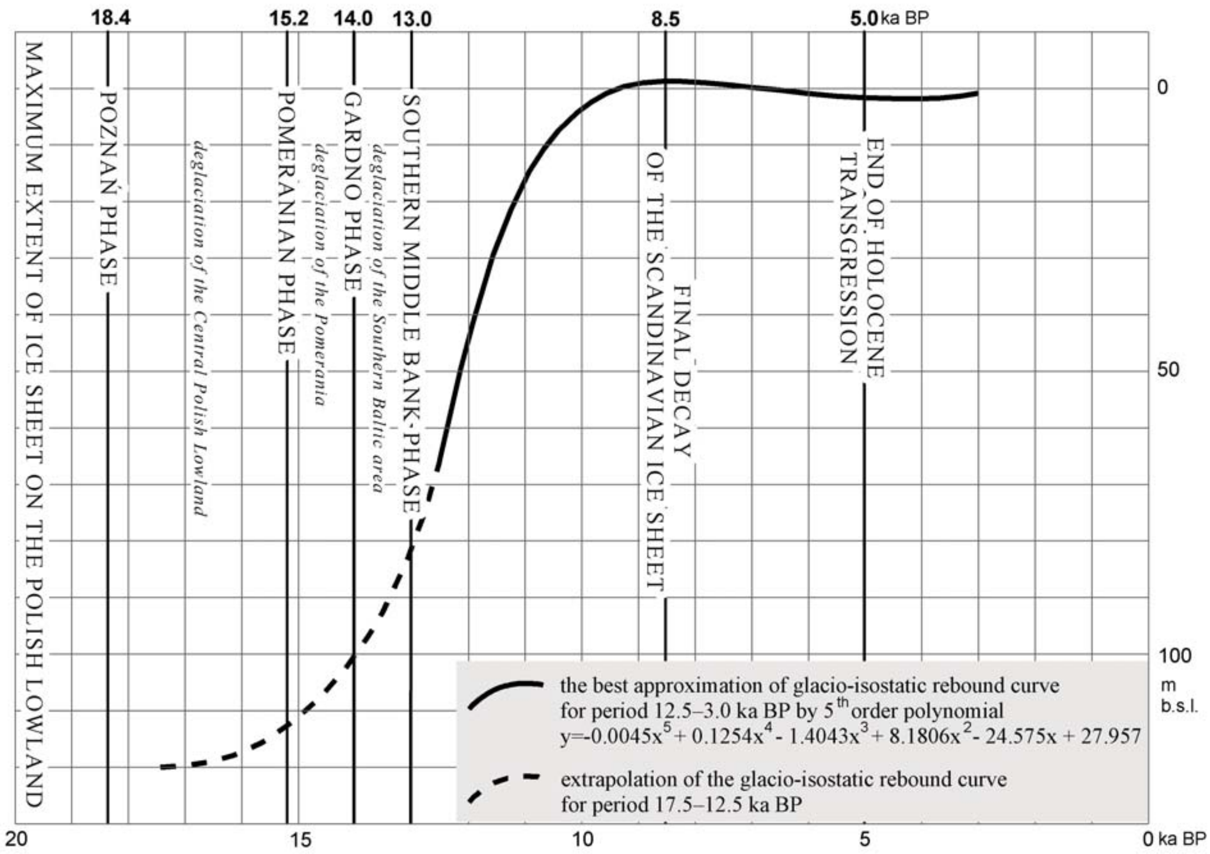


Fig. 37. The glacio-isostatic rebound curve for the southern Baltic area

straining effect of hydro-isostasy and sedimento-isostasy. Since the onset of deglaciation, the southern Baltic area was covered by water bodies the depth of which ranged, within 13.0–9.0 ka BP, from about 40–50 to about 70–80 m. Those water bodies (the Baltic Ice Lake, Yoldia Sea and the early Ancylus Lake) rapidly accumulated thick sediment layers. The Late Pleistocene (Baltic Ice Lake) sediment thickness averages about 10 m in the southern part of the Bornholm Basin and about 15 m in the Gdańsk Basin. The Early Holocene (Yoldia Sea and Ancylus Lake) sediment thickness is 3–8 m. The weight imposed on the southern Baltic area by the water and sediment masses had to contribute to the relatively fast termination of the glacio-isostatic uplift of the area.

The shape of the southern Baltic glacio-isostatic rebound curve (Fig. 37) seems to indicate that from *c.* 9.0 to *c.* 8.0 ka BP the area experienced the forebulge migration, the subsidence occurring from *c.* 8.0 to *c.* 6.5 ka BP. As of *c.* 5.0 ka BP, the crust seemed to regain its state of equilibrium. The narrow range of vertical crustal movements (about 3–4 m at most) within 9.0 to 3.0 ka BP remains within the curve estimation error range for that period of time. For this reason, the interpretation presented is loaded with more uncertainty than that pertaining to the phase of major glacio-isostatic rebound of the southern Baltic area.

The results of an attempt to reconstruct glacio-isostatic movements should be treated as an averaged (smoothed) model for the entire southern Baltic area. The true picture of the glacio-isostatic rebound was doubtless more complex, particularly in the western part, within the Palaeozoic platform and in the marginal, western part of the Pre-Cambrian platform (Bornholm Basin) which are characteristic in their strong block

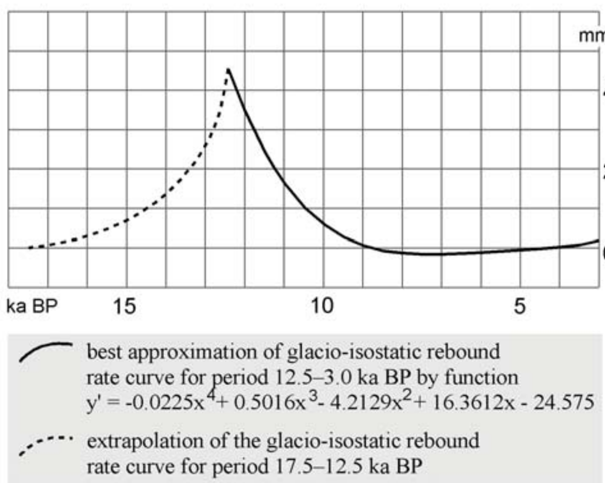


Fig. 38. The glacio-isostatic rebound rate curve for the southern Baltic area

fracturing, individual tectonic blocks having a potential to move separately and the relaxation time being extended (Liszkowski, 1975, 1980, 1993; Dadlez in: Mojski *et al*., eds., 1995). Compensatory movements could also differ to some extent between various tectonic blocks with the Pre-Cambrian platform east of the Bornholm Basin. In that area, the coast is intersected by a few photolineaments which are most likely neotectonic in nature (Bażyński *et al*., 1984; Graniczny, 1994). However, within the marine area located on the Precambrian platform, no translocations of the Pleistocene sediment top that would be indicative of differentiated block movements in Late Pleistocene or Holocene, were observed within the fracture zones. The only example from the southern part of the Gotland Basin (Fig. 39) may be interpreted both as an effect of lithosphere block translocation along the fracture zone and as an effect of that zone on differentiation of glacial erosion of the bedrock. A much more complex is the tectonic pattern within the Palaeozoic platform (e.g. Dadlez in: Mojski *et al*., eds., 1995; Kramarska *et al*., 1999). However, few photolineaments, presumably neotectonic in origin, cut through the shore line (Bażyński *et al*., 1984). That glacio-isostatic movements in the marine area of the Palaeozoic platform were still in progress at the beginning of Holocene is indicated by translocations of the Late Pleistocene sediment top (Baltic Ice Lake sediments) reaching up to 6–7 m and visible on seismoacoustic profiles from the Bornholm Basin. Displacements of the Early Holocene (Ancylus Lake) sediment top are less pronounced and extend across about 2–3 m (Fig. 40).

To sum up, the reconstruction of glacio-isostatic movements of the southern Baltic area modify and refine the results published by Rosa (1968) and Liszkowski (1975, 1980). Particularly important was to demonstrate that the southern Baltic area and northern Poland was subjected to uplift in Late Pleistocene, which is in agreement with predictions of the geophysical model of Fjeldskaar (1994). Discrepancies with respect to that model involve the extent of the southern Baltic uplift. For the period of the last 15.0 ka BP, Fjeldskaar (1994) estimated the southern Baltic to have been uplifted by as much as 300–400 m, i.e. more than twice the estimate resulting from this work. As pointed out by Fjeldskaar (1994), the ice sheet thickness is the least reliable parameter of the model and it is most likely the maximum value, greatly affecting the results of modelling. The ice sheet covering northern Poland and the southern Baltic could have been in reality less thick. Similarly to Fjeldskaar (1994), Ehlers (1990) and Mojski (in: Mojski *et al*., eds., 1995) assumed the ice sheet during the maximum range of glaciation to be about 3.2–3.5 km thick in the centre. However, while Fjeldskaar (1994) assumed the ice sheet on the southern Baltic coast and over the southern Baltic itself to be 2,000 to 2,400 m thick, Ehlers (1990) and Mojski (in: Mojski *et al*., eds., 1995) estimated the corresponding thickness at about 500–1,000 m. The lower ice sheet thickness resulted in a smaller crust deformation and an earlier cessation of compensatory movements. Fjeldskaar (1994) demonstrated the proglacial forebulge during the maximum extent of the last glaciation to subside gradually, without any discernible signs of lateral migration after the ice sheet retreat. In this respect, the reconstruction of glacio-isostatic movements in the southern Baltic, carried out by comparing relative sea level curves with the eustatic ocean curves coincides with results of the geophysical modelling.

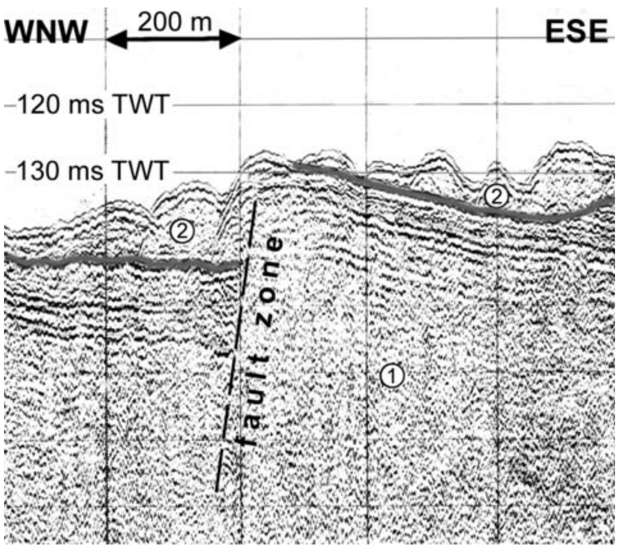


Fig. 39. Displacement of the top of Late Pleistocene deposits in the southern part of the Gotland Basin; seismoacoustic (boomer) profile GT 10

Silurian: *1* — clayey deposits; **Late Pleistocene**: *2* — subaqueous till (Southern Baltic diamicton)

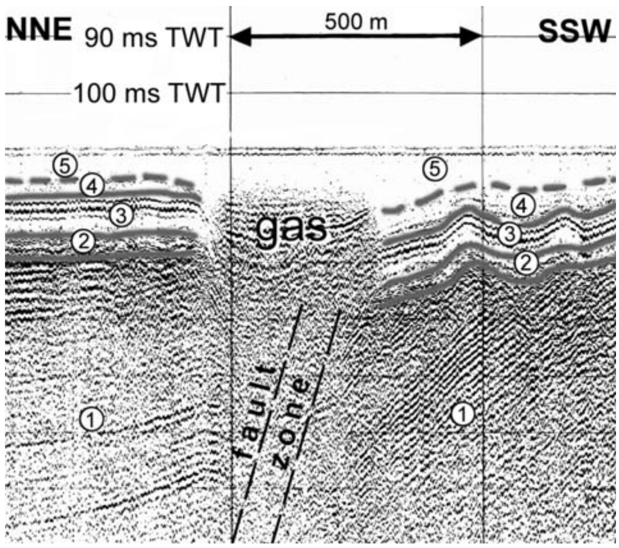


Fig. 40. Displacement of the top of Late Pleistocene and Early Holocene deposits in the southern part of the Bornholm Basin; seismoacoustic (boomer) profile NE 2

Cretaceous: *1* — sandy silt; **Pleistocene**: *2* — till; **Late Pleistocene**: *3* — clay (Baltic Ice Lake); **Early Holocene**: *4* — clay (Yoldia Sea and Ancylus Lake); **Middle and Late Holocene**: *5* — marine mud (Mastogloia, Littorina and Post-Littorina)

RECENT CRUSTAL MOVEMENTS

Recent crustal movements in the southern Baltic area and northern Poland are a net result of interaction of a number of geotectonic processes. Important in this regard are such processes as:

— the on-going glacio-isostatic uplift of Scandinavia;

— subsidence of the area extending from the Valdaj Upland through Masuria and central Poland as a result of collapse of the forebulge that originally surrounded the Pleistocene ice sheet;

— northward drift of the African Plate and processes initiated along the North-Atlantic Mid-Ocean Ridge (Harff *et al*., 2001).

Thus, recent vertical movements in the southern Baltic and northern Poland can be only indirectly and partially associated with glacio-isostasy, as already pointed out by Liszkowski (1975, 1993). The pattern of neotectonic movements generated by large-(planetary-) scale geotectonic processes may be rendered more complex by regional and/or local factors and processes. The pattern may be regionally modified and diversified by the block structure of the Quaternary basement in the southern Baltic area and by local processes such as, e.g. compaction of thick Quaternary sediment layers.

Extrapolation of data collected from adjacent land masses onto the Baltic reveals its southern part as a stable area the recent vertical crust movements of which proceeded at a rate ranging from –1.0 to +1.0 mm/a (e.g. Balling, 1980; Winterhalter *et al*., 1981). This is also the pattern emerging from a map of recent vertical movements (Harff *et al*., 2001) according to which the southern Baltic forms a divide between the rising Fennoscandian area and an extensive north-east to south-west subsidence zone from the Valdaj Highland through the Masurian Lake District to central Poland.

Recent vertical crustal movements within Poland were traced by repetitive precision nivelations and by collecting mareograph data (Fig. 3) (Wyrzykowski, 1975, 1985). The movement rates, related to the mean sea level, and their mean errors for the Polish coast are as follows: –2.9 ±0.6 mm/a at Braniewo, –1.1 ±0.2 mm/a in Gdańsk, –0.8 ±0.5 mm/a at Hel, –0.1 ±0.4 mm/a in Władysławowo, +0.3 ±0.4 mm/a at Łeba, +0.1 ±0.2 mm/a at Ustka, –1.0 ±0.1 mm/a in Kołobrzeg and –0.7 ±0.1 mm/a in Świnoujście (Wyrzykowski, 1985). The values indicate a higher variability of the vertical movements along the Polish coast than that shown on an earlier map (Wyrzykowski, 1975). However, differences in movement rate estimates notwithstanding, the two maps identify identical areas subjected to uplift or subsidence. A slight uplift is observed only on the coast stretch between Władysławowo and Ustka, while subsidence, slight as well, affects the Gulf of Gdańsk coast east of Gdynia and the coast of Western Pomerania. It may be presumed that the neotectonic movements associated with geotectonic factors proceed along the Polish coast at a rate of –1 to +0.5 mm/a. The fastest subsidence (up to 3 mm/a) is typical of the Vistula Delta near Elbląg. This is most likely associated with compaction of the Holocene deltaic sediments, accelerated by exploitation of the aquifer in the vicinity of Gdańsk and Elbląg.

It is difficult to assess how far northwards (seawards) one may extrapolate observations from the southern coast. Still more problem-ridden is the question how far back in time may the data on recent vertical movements of the coast be extrapolated in a reliable manner. A comprehensive review of the scope covered and problems tackled by Polish neotectonic research (Zuchiewicz, 1995b) reveals the absence of studies directly concerning the southern Baltic area.

It seems that, at this stage, data on recent vertical crustal movements on the southern Baltic coast cannot be directly related to the history of relative sea level changes during the Subboreal and Subatlantic, and even less to those happening earlier. A comparison of dated sediment profiles from northwestern and southwestern inshore areas of the Gulf of Gdańsk, i.e., areas where differences in recent vertical crust movement rate may exceed 2 mm/a, shows no significant differences in the vertical location of sediments of identical origin and age. Examples are furnished by the profiles Dziady 3 on the Vistula Spit (Fig. 25) and Chałupy 92 on the Hel Peninsula (Fig. 33) which evidence the formation of peats there during the Middle Atlantic at an identical ordinate. The onset of lagoon mud deposition, too, is dated similarly in the Puck and Vistula Lagoons, to the end of the Atlantic and the turn of the Atlantic and Subboreal (Bogaczewicz-Adamczak, Miotk, 1985b; Witkowski, Witak, 1993; Kramarska *et al*., 1995; Uścinowicz, Zachowicz, 1996; Zachowicz, Uścinowicz, 1997).

Evolutionary trends on the Polish coast over 1875–1983 (Zawadzka, 1999) showed a higher erosion rate to be typical of the eastern rather that of the western coast. West of the Teysseyre–Tornquist (TT) zone, with the prevalence of crustal subsidence (Fig. 3), the mean rate of cliff recession over 1875–1979 was 0.23 m/a, the dune coast receding at a mean rate of 0.09 m/a. East of the TT zone, in the area affected predominantly by uplift, the rate of cliff and dune recession was 0.55 and 0.06 m/a, respectively. That would indicate that the coast evolution trends, at a low variability of recent crust movements, are more dependent on hydrodynamic than on neotectonic factors (Zawadzka, 1999). Nevertheless, there have been so far no studies in which interrelationships between the recent climatic changes and accelerated sea level rise on the one hand and the vertical crust movements on the Polish coast on the other would be analysed. Even more importantly, interactions between the area's neotectonics and relative sea level changes in Late Holocene are still unclear. A further, detailed research is called for as the problem has not received sufficient attention and remains unresolved.

While inferences on past events from the knowledge on recent vertical crust movements on the Polish coast involve a substantial uncertainty, this knowledge may be important for forecasting evolution of the coastal zone, particularly in view of the observed sea level rise and the increasing storm frequency.

COASTAL ZONE EVOLUTION

POTENTIAL FOR IDENTIFICATION AND DATING OF SHORELINE FORM RELICTS

Discussion of the southern Baltic coast evolution throughout its entire history could begin from still another attempt at defining such terms as the shoreline or the coastal zone. Such definitions have been already given many times in numerous publications. Regardless of the scope and precision of the definitions, they are of a limited use for the palaeogeographic analysis of an area subjected to marine transgression, when the evolution of the shore may be only inferred from the seabed relief and geological structure. The seafloor could have, at most, retained, in a better or worse state of preservation, relicts of various forms and sediments associated with evolution of a broadly defined coastal zone. For the changes occurring in Late Pleistocene and Holocene, i.e. within about 15 thousand calendar years (about 13 thousand conventional radiocarbon years), on the southern Baltic coast, precise definitions are not overly important. It is unquestionable that the term "shoreline" can be used only conventionally, with reference to a whole assemblage of relicts of forms and sediments of former coastal zones, sometimes identifiable on the recent seafloor.

As already mentioned in the Introduction, the problem of the Baltic Sea shoreline displacement has been recognised at the earliest, and largely solved, in Sweden, Finland and in Estonia. In those countries, former shorelines are frequently located far away from the present coast and situated at substantial altitudes. The term "shoreline" is used often, but its meaning is most frequently restricted to a set of more or less transformed forms associated with the coastal zone. In the case of submerged shores, the scale of changes in those coastal zone formations that became submerged is much more extensive. The basic question is what, if any, forms could persist and are identifiable on the seafloor. The question is extremely important for the analysis of the southern Baltic coast development, because the former shorelines there are located exclusively below the present sea level.

The relevant literature has initially focused at relicts of cliff shores (Gudelis, 1965; Kolp, 1965, 1976, 1979a, b; Rosa, 1967, 1968; Charin, 1987; Litvin, 1987). The so-called erosional terraces were interpreted as cliff shore relicts; the terraces were assumed to have evolved at the cliff base when the sea level stabilised. Such interpretations were usually derived from bathymetric charts developed by interpolation of depth data from navigational charts, with very scant information on the seafloor-forming sediments. A wider application of echo-sounding, at first, and seismoacoustic equipment, later on (e.g. Rosa, 1987; Mojski ed., 1989–1995; Przezdziecki, Uścinowicz, 1989; Pikies in: Mojski *et al*., eds., 1995; Przezdziecki, 2004) demonstrated that flattened erosional seabed forms (terraces) are much less frequent that hitherto thought. The structure of those terraces is not uniform, either. The erosional surfaces may be covered only by a thin layer of lag deposits or by, occasionally a few metres thick, younger sediments. In such cases, spatial correlation of terraces, based only on their bathymetric position, is impossible. Some flattenings and erosional terraces, earlier interpreted as typical relicts of cliff shores or downright as submerged cliffs (e.g. Rosa, 1967; Charin, 1987; Litvin, 1987), are, as demonstrated by seismic profiles, structure-dependent erosional formations (Fig. 41). Completely disregarded was the impossibility of dating erosional terraces or of finding corresponding forms in the morphology of recent cliff coasts, as well as determining their relation to the mean sea level at the time of their formation.

In the light of these remarks, it is assumed for the purpose of this work, that cliff shores could have occurred in areas that show at present coarse grained deposits, a residue of washed off glacial and fluvioglacial sediments. Such areas are delimited from maps of seabed sediment (Mojski ed., 1989–1995; Kramarska in: Mojski *et al*., eds., 1995). The seafloor surface in such areas can be flat (erosional terraces and plains) or to have diverse irregular 3–5 m high relief (Pikies in: Mojski *et al*., eds., 1995).

Progress in the seafloor exploration, and particularly the wide use of seismoacoustic profiling, has led to recognition of relicts of various accumulating formations. The earliest known among the accumulating forms in the Gulf of Gdańsk was the submerged part of the Vistula Delta (Ejtminowicz, 1982; Wypych *et al*., 1982). The formation was developing on the recent seafloor from the Late Glacial until the end of the Littorina Sea (Atlantic) transgression and persisted in spite of numerous transgression and regressions owing to a large amount of sedimentary material discharged by the Vistula. When the sea level fell, the delta front developed more intensively, outlet fans emerged at still lower and lower levels, and the delta plain formed water distribution channels incised deeper and deeper. Transgressions enhanced the emergence of sand barriers (spits) in the delta front and of lagoons in the hinterland, within the deltaic plain. The barrier migration rate depended on the amount of material supplied by the river and the sea level rise rate. Agradation of distribution channels resulted in accretion of clastic and organic sediments on the delta plain. All those processes are recorded in the deltaic sediments forming extensive surfaces of the Gulf of Gdańsk bottom. Unfortunately, the problem has not been looked into with a sufficient detail. Similar deltaic forms should have been formed by the Odra in the southern part of the Arkona Basin, and by the Nemunas in the northeastern part of the Gdańsk Basin. Those deltas, provided they do exist, have not been so far detected and described. Formations associated with mouths of numerous smaller rivers emptying into the Baltic Sea have not been preserved due to their small size; they were destroyed by transgressions. Relationships of various deltaic formations and sediments with the sea level prevailing during their emergence are much clearer than those pertaining to erosional formations of the former cliff shores. The location of proximal parts of the delta fronts (outlet fans) or of a deltaic plain is closely associated with the mean sea level. The depth of erosional incisions is also related to the erosional base. Analysis stratification sequence allows to approximate the relative age of the layers. When determining the sea level, later

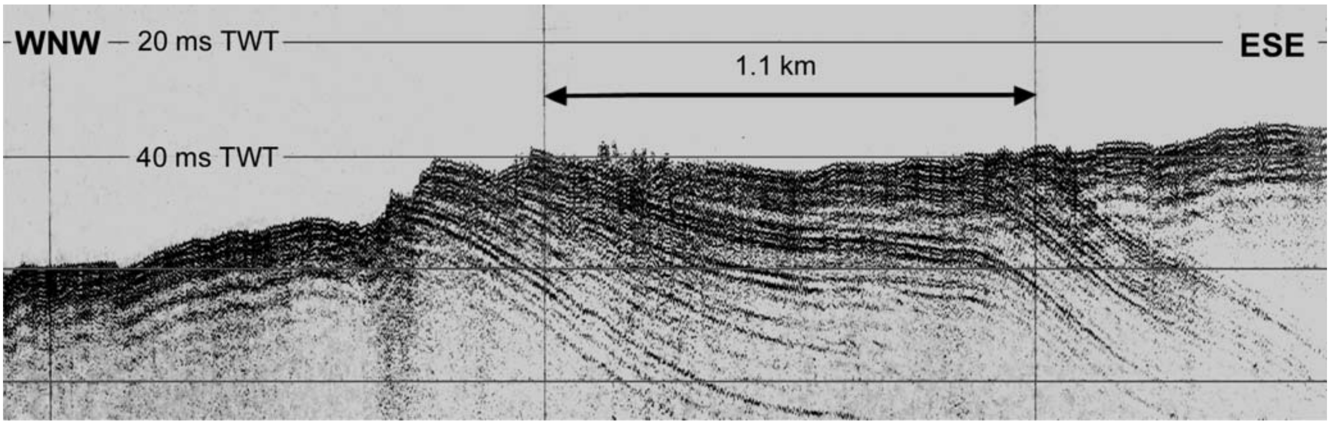


Fig. 41. Outcrops of Tertiary (Palaeogene) deposits offshore the Sambian Peninsula; an example of seabed relief governed by geological structures

transgressive transformations of the delta surface should be, of course, taken into account.

The last to be recognised on the seafloor were relicts of sand barriers (spits). Spit relicts were investigated in, i.a., the western Baltic (Jensen, Stecher, 1992; Jensen *et al*., 1997, 1999; Bennike *et al*., 2000; Schwarzer *et al*., 2000). In the southern Baltic, sand barrier relicts were identified in the southwestern part of the Bornholm Basin, in the eastern part of the Gdańsk Basin, in the southern part of the Southern Middle Bank and on the Słupsk Bank, on the Czołpino Shallow and in some places of the present coastal zone at depths of about 12–15 m (e.g. Kramarska *et al*., Uścinowicz in: Mojski *et al*., eds., 1995; Przezdziecki, 2004). The sand barrier relicts occur in two forms: regressive and transgressive. The relicts of barriers formed during regressions cover extensive areas, extend across wide depth ranges and show a large-scale oblique bedding, indicative of progradational processes (Figs. 17, 28, 29). The relicts of barriers formed during transgression are smaller, occupy limited spaces and extend across a narrower depth range. Both generation of barrier formations developed in areas of the Pleistocene sand layers, and both are in their top parts cut flat by the subsequent erosional processes. The regressive barrier formations could have remained, despite the subsequent erosion, due to their considerable size and thickness. The transgressive barrier relicts were retained only when the shoreline location changed as a result of a sudden catastrophic event. That could have happened at extremely severe storms, when the barriers were broken and the hinterland was low. During periods of a rapid sea level increase, the transgressive barriers were probably relatively weak, thus enhancing occasional catastrophic shoreline displacement. Relicts of both types of barriers, formed during regressions or transgressions, mark only the location of the actual shoreline. The lack of information on the subsequent erosion allows the location of the actual sea level to be approximated only. Determination of the relative age of a formation is possible through sequence analysis, while the absolute age can be determined only when macrophyte remains and/or malacofauna are found and dated.

In the present work, determination of the age of submerged shorelines, or more exactly, of relicts of formations associated with barrier and deltaic shores, and hence determination of their relationships with the sea level curve was in some cases possible with pollen and malacofauna-based analyses as well as by radiocarbon dating of the associated sediments. More often, age determination was assisted by the analysis of sequence of structures and of barrier or deltaic sediments, supported occasionally by bio- and chronostratigraphic techniques. Location of the shoreline, or — more strictly — of the cliff shore relicts in a given period of time was traced by correlating them spatially with relicts of accumulational shores and by correlating their present location and bathymetry with the relative sea level curve.

SHORELINE DISPLACEMENT

Shoreline location depends primarily on vertical crustal movements and eustatic sea level changes. Superposition of the two phenomena is manifested on the coast as a relative change of sea level. Depending on interactions between and rates of eustatic sea level changes and vertical crust movements, various scenarios of the change in the shoreline — or, more generally, the coastal zone — location are possible. The rate of relative change (rise or fall) of the sea level is a very important, but not the only, factor affecting the coastal zone location. Important are also morphological and geological characteristics of the coast, lithological properties of rocks making up the seafloor and the coast, as well as sedimentary material erosion, transport and accumulation. Occasionally, those characteristics and properties may very substantially affect the rate, and sometimes also the direction, of changes taking place on the shores subjected to changes of the sea level.

THE SOUTHERN BALTIC SHORELINES IN LATE PLEISTOCENE

Evolution of the southern Baltic coast in Late Pleistocene depended mainly on very dynamic changes in the sea level, resulting from interactions between deglaciation dynamics and glacio-isostatic rebound, duration of the connection with the ocean, and on eustatic changes in the ocean level.

Before the Baltic emerged: deglaciation and ice-dammed lakes (14.0–13.0 ka BP)

Until *c.* 14 ka BP, the southern Baltic area was covered by the ice sheet. Shortly before that time the ice sheet melted in the area of the present Pomeranian Bay (Uścinowicz, 1996, 1999a). Deglaciation of the southern Baltic area began with the ice sheet retreat from the Gardno moraines *c.* 14 ka BP (Rotnicki, Borówka, 1995b; Mojski, 1993; Mojski in: Mojski *et al.*, eds., 1995). The ice sheet margin extending from Puck to Darłowo was located slightly south of the present shoreline. Between Lake Kopań and Darłowo, the ice sheet margin crossed the present-day shoreline to run north of Kołobrzeg and the Odra Bank. Location of the marginal formations of that phase on the Baltic seafloor is most probably marked by coarse-grained deposits (boulders, cobbles and gravels) north of lakes Bukowo and Jamno and north-west of Kołobrzeg (Uścinowicz, 1996, 1999a). That line may be correlated to the range of the Halland–West Skåne Phase in southwestern Sweden (Lagerlund, Houmark-Nielsen, 1993). In the east, the Gardno Phase has its equivalent in the Middle-Lithuanian Phase (Gudelis, 1976; Raukas *et al.*, 1995). During the Gardno Phase, the ice sheet was still filling the Gdańsk Basin and covered the present Vistula Delta. Deglaciation, similarly to that occurring earlier in the Pomeranian Coastal Area adjacent on the south, was, at least at first, of a similar aerial character. As the ice sheet receded from the southern Baltic and the northward tilted surface was uncovered, more and more numerous and increasingly larger ice-dammed lakes were formed. Initially, they were relatively small and shallow water bodies that formed a system of proglacial lakes. Sediments of such lakes are known from several sites on the Polish coast, i.a., at Wytowno and Sławno (e.g. Rosa, 1994; Mojski, 2000).

As deglaciation proceeded, small local ice-dammed lakes disappeared to make way to a large dammed lake between the present shore and the Słupsk Bank. Sediments of that lake were identified within an area extending from the western flanks of the Słupsk Bank to Lake Wicko in the west, to the Puck Bay in the east, at depths of about 10–40 m. Grain size distribution of those sediments is diverse; they are mostly silty clays and clayey and sandy silts, laminated, occasionally with sandy bands and locally silty sands and fine sands. Post-sedimentation deformational structures are frequent. The ice-dammed lake sediment thickness depends on the basement relief and ranges from several tens of centimetres to 25 m (Kramarska, 1991a, b; Uścinowicz, Zachowicz, 1991a, b). The dammed lake was most probably at its largest when the ice sheet margin became stationary on the Słupsk Bank *c.* 13.5 ka BP. The stationing is marked by the presence of elevations in the relief of the northern and northwestern parts of the Słupsk Bank, with the southern Baltic largest relative heights and steepest slopes as well as the presence of boulders. South-west of the Słupsk Bank, an ice sheet margin lobe made an incursion into the southern part of the Bornholm Basin where semicircular relicts of end moraines are still visible in the seafloor relief. Lag deposits remaining from the Słupsk Bank marginal formations as well as a series of relicts of marginal zone formations from the southern part of the Bornholm Basin can be correlated to the location of the ice sheet margins in Scania from *c.* 13.5 ka BP (Lundqvist, 1994). The ice sheet lobe covered also the Gdańsk Basin and left a series of mounds in the western part of the Gdańsk Basin which can be related to the North-Lithuanian Phase (Luga) in the north-east, dated to about 13.2 ka BP (Raukas *et al.*, 1995). The ice-dammed lake on the ice sheet forefront was most probably more extensive than indicated by the present location of its deposits. It extended along the ice sheet margin from Sambia through the northern part of the Vistula delta and the southern part of the Gulf of Gdańsk to reach Scania. It is possible that the ice-dammed lake in question was connected, south of the Bornholm Basin, with extensive wetlands and bogs occurring then in the present-day area of the Odra Bank and the Pomeranian Bay, fed by rivers discharging from the south. As opposed to the damned lake deposits containing no macro- and microfaunal remains, the Pomeranian Bay wetland sediments harbour abundant malacofauna, particularly freshwater snails and bivalves (Kramarska, Jurowska, 1991; Krzymińska, 2001). This shows that the connections, if present, were episodic at most.

The knowledge on the location, nature and changes of the shoreline of the large ice-dammed lake under discussion is very scant. The lake was probably large enough to make it possible, in spite of cold climate and long periods of ice cover, for the shoreline to be affected by wave action. It is possible that the so-called Nowęcin rampart (5 km east of Łeba), thermoluminescence-dated to Late Pleistocene (Rosa, 1987), is a formation that marks the dammed lake's shore. Should that be the case, the maximum water level of the lake would by higher than the present sea level by about 1.0–1.5 m. At that time, the Vistula rolled from Bydgoszcz northwards (Mojski, 1990; Starkel, Wiśniewski, 1990) and discharged into the ice-dammed lake in the northern part of the present Vistula Delta. This is perhaps why there are no typical dammed-lake deposits there, the oldest arms of the Vistula Delta being formed within the lake itself. It is not known what the southern shores of that lake looked like in the western part of the Polish coast as no traces of the dammed lake deposits were preserved there. The dammed lake shores could move north of the present shoreline and the lake's southern border could be formed by moraine elevations occurring north of Koszalin and Kołobrzeg. The presence of such elevations there is evidenced by extensive coarse-grained sediment fields covering the present seafloor. The moraine elevations could have been eroded by the dammed lake water, draining along the ice sheet margin towards the Öresund. If they were not totally destroyed in Late Pleistocene, they eventually succumbed to erosion during the Atlantic (Littorina Sea) transgression. The intensive erosional processes during deglaciation are indicated by the absence of deposits from the last glaciation north of the Odra Bank. Marine sands there are directly underlain by the Grudziądz Interstadial (Interpleniglacial) deposits (Kramarska, 1998). The dammed lake water level was controlled by relief sills, the first of which being located in between the Adlergrund (Eagle Bank) and the Odra Bank and the next one in the Öresund. The intensive erosional activity of meltwater on the sill separating the Adlergrund and the Odra Bank is indicated by the very thin bed of Pleistocene sediments there overlie Cretaceous deposits (Lemke, 2002, personal comm.). If the range of the ice-dammed lake north-west of Koszalin and Kołobrzeg had been more extensive than described in earlier publications (Uścinowicz in: Mojski *et al*., eds., 1995; Uścinowicz, 1999a) and the lake's shore had been close to the present coast, the lake's sediments could have been wiped out by later erosion.

The lake was bordered in the north by the ice sheet margin. Glaciofluvial deltas formed then within the Słupsk Bank area (Uścinowicz, 1996, 1999a), while blocks of dead ice could have remained the southern parts of the Gdańsk and Bornholm basins, south of the ice sheet margins (Mojski, 1993, 1995b, 1997, 2000; Mojski in: Mojski *et al*., eds., 1995).

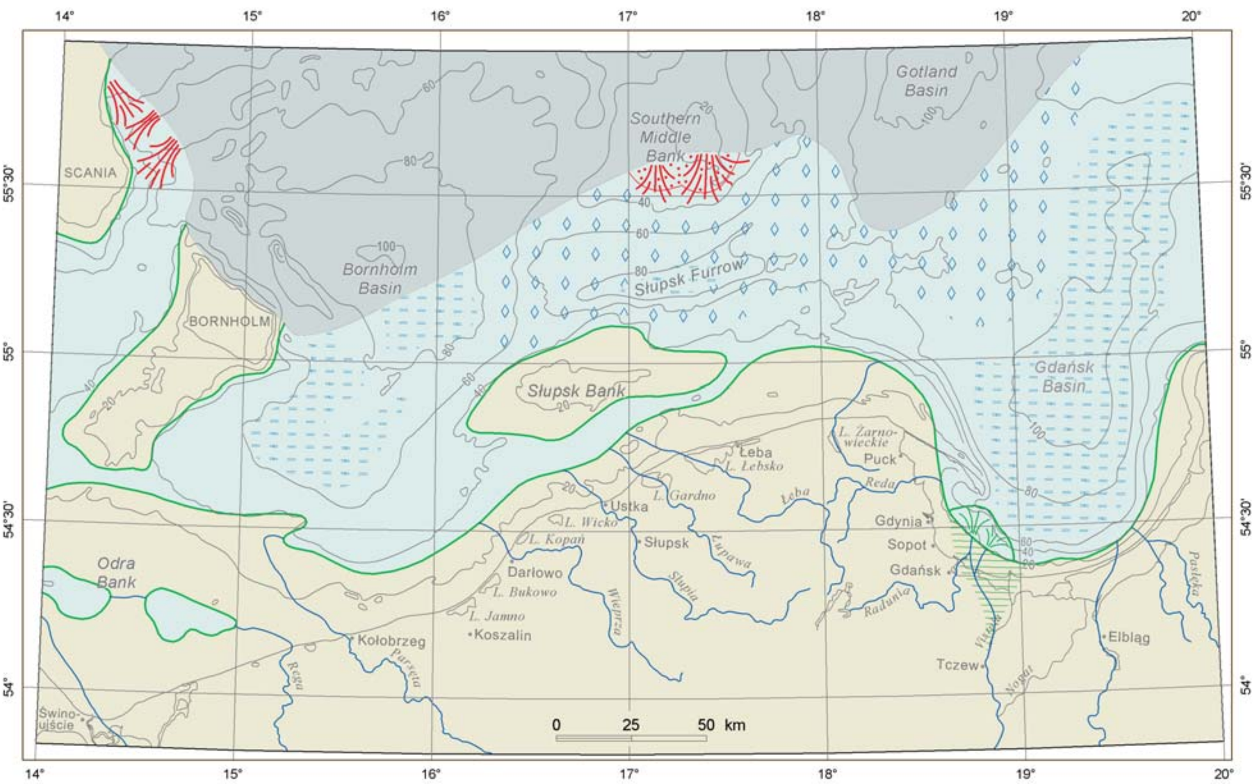
The description given above concerns the largest possible extent of the ice-dammed lake. It is possible that, at different stages of the evolution, as deglaciation proceeded, the lake changed its size and extent to form a series of interconnected smaller ice-dammed lakes.

Origins of the Baltic: the Baltic Ice Lake (13.0–10.3 ka BP)

Within 13.5–13.0 ka BP, the ice sheet margin retreated from the Słupsk Bank to the Southern Middle Bank. The extent of lobes in the Gdańsk and Bornholm Basin declined as well. At that time, the southern Baltic area experienced intensive glacio-isostatic uplift and a large amount of meltwater produced intensive erosion of the sills controlling the discharge of water from the southern Baltic depression. The fast melting of the ice sheet in the southern Baltic depression and sill erosion resulted in lowering the dammed lake water level and emergence of areas located south-east and south-west of the Słupsk Bank, i.e. in the present shallow-water zone. The ice-dammed lake water drawdown was most probably accompanied by intensive erosion in areas of increased water flow. At the same time, the ice-dammed lakes in the southern areas of the Gdańsk and Bornholm basins, freed from under the ice sheet, increased in size. The deglaciation was becoming more and more clearly a subaqual one. The role of dead ice at that stage of deglaciation most probably diminished. As the deglaciation proceeded and deeper and deeper reservoirs in front of the ice sheet emerged, the ice sheet margin could have formed an ice cliff in deeper parts of the basin, varved clays being deposited on the cliff forefront. The ice sheet dome could have become reduced not only by surficial melting, but also by iceberg calving (Malmberg-Persson, Lagerlund, 1990; Lundqvist, 1994).

The key role in the formation of the Baltic Ice Lake was played by the ice sheet meltdown in the Słupsk Furrow. It produced a connection between the dammed lakes of the Gdańsk and Bornholm basins and the formation, in the ice sheet forefront, of a reservoir extending from the Latvian coast to the Öresund, regarded as the beginning of the Baltic Ice Lake (Fig. 42). Deglaciation processes in the Słupsk Furrow were complex and have not been entirely explained. That period of deglaciation is associated with formation of the so-called subaqual till in the Słupsk Furrow and in the sills that separated it from the Bornholm Basin, that separated the Gotland Basin from the Gdańsk Basin, and that developed in the southern part of the Gotland Basin. Since the ice sheet retreat from the Słupsk Bank, those areas most probably harboured large blocks of dead ice that melted in the presence of large amounts of water, some of them melting underwater and some remaining as floating icebergs. Complex sedimentation processes were in operation at that time. Sedimentation of suspended clay particulates was accompanied by gravity flows of the depositional material from the dead ice as well as precipitation of sediment from the floating icebergs. The subaqual till thickness ranges widely and reaches about 40 m in the sill separating the Bornholm Basin from the Słupsk Furrow. The subaqual till is soft-plastic, light brown to grey-brown in colour, with clear pale blue and moderate red irregular lenses and lumps of silt. It is highly inhomogenous in terms of its granulometry. Most often it is silty or clayey till, in places clay or silty clay with an admixture of sandy and gravelly fractions, occasionally places with single sharp-edged rock fragments up to a few cm in diameter. The subaqual till has diverse structure, orderless in places, frequently laminated and irregularly varved. Deposits of a similar origin and lithology are known from southwestern Scania as the so-called Lund diamicton (Lagerlund, 1987, 1995; Berglund, Lagerlund, 1981; Malmberg-Persson, Lagerlund, 1990) or Öresund diamicton (Lagerlund, Houmark-Nielsen, 1993). By analogy, the subaqual till described above can be termed the Southern Baltic diamicton. At the same time, in the ice-free areas of the Bornholm and Gdańsk basins, varved clays were deposited. Varved clays have not been found on the subaqual till in the Bornholm Basin, Słupsk Furrow and Gdańsk Basin, hence the conclusion that they are contemporaneous facial varieties of sediments deposited in the marginal ice-dammed lakes.

The progressing deglaciation was slowed down again when the ice sheet margin stopped at the Southern Middle Bank line (Fig. 42). The Polish Exclusive Economic Zone includes a fragment of the Bank only. Its shallowest parts, at depths of about 15–20 m, shows the presence of tills covered by a thin layer of coarse lag deposits, while the southern and southeastern slopes are covered by sandy and sandy-gravelly sediments of glaciofluvial deltas (Fig. 12). The ice sheet margin was at that time located along a line connecting the Southern Middle Bank south-west towards the central part of the Bornholm Basin, turning north-west towards Hanö Bay. Off the coast of Scania, the line can be correlated to the lines determined by Björck and Möller (1987) and by Lundqvist (1994), dated to 13.0–12.9 ka BP. East of the Southern Middle Bank, the ice sheet margin made an incursion, most likely as a lobe, into



Explanations for Figures 42–47 and 50

Palaeogeography

- ice sheet
- land area
- water area
- Vistula River delta
- aluvia cones
- rivers
- fluvioglacial delta
- coast: a – cliff, b – barrier, c – undefined

Recent extent of various stages of the Baltic Sea deposits

- sands and gravels of fluvioglacial deltas
- subaquatic till (Southern Baltic diamicton)
- varved clays
- clayey deposits of the Baltic Ice Lake
- barrier sands of the Baltic Ice Lake
- clayey deposits of the Yoldia Sea
- clayey deposits of the Ancylus Lake
- barrier sands of the Ancylus Lake
- muds and clays of the Littorina and Post-Littorina seas
- sands and gravels of the Littorina and Post-Littorina seas
- areas of limited deposition (outcrops of older deposits) in the Littorina and Post-Littorina seas
- present erosional boundaries of fluvioglacial deltas and sediments from various stages of the Baltic Sea evolution

Fig. 42. The early phase of the Baltic Ice Lake (*c.* 13.0 ka BP) (after Uścinowicz in: Mojski *et al*., eds., 1995; modified)

the southern part of the Gotland Basin and then through the Gotland–Gdańsk Sill and the Klaipeda Bank was directed towards the Latvian coast where the ice sheet stationing is marked by moraines of the Otepää Phase (Raukas *et al*., 1995), also called the North-Latvian Phase (Gudelis, 1976).

As already mentioned, when the ice sheet margin stabilised in the area of the Southern Middle Bank, the Baltic Ice Lake water level was by about 30 m lower than at present. The northern coast of that earliest Baltic Sea stage was formed by the ice sheet margin, most probably ice cliffs stretching along considerable lengths. Large, complex systems of glaciofluvial deltas formed on the Southern Middle Bank. Glaciofluvial deltas were most probably formed simultaneously in the western part of the Hanö Bay as well (Andrén, Wannäs, 1988). Equally scant is information about the southern coasts of the emerging Baltic Ice Lake. Most probably, the shoreline was located about 70 km north of the present Pomeranian Bay coast, about 15–20 km north of the present central Polish coast and about 10 km north of the present Gulf of Gdańsk coast. Depending on changes in the water level, the Słupsk Bank was either connected with the land mass by a narrow isthmus or was an island separated from the land mass by a narrow and shallow strait (Fig. 42). The marginal formations, remnants of the ice sheet stationing in the northern part of the Słupsk Bank predisposed the area to develop cliff shores. The oldest part of the Vistula delta was formed in the Gulf of Gdańsk (Fig. 10).

The ice sheet receded from the Southern Middle Bank *c.* 13.0–12.7 ka BP. This may be inferred from correlating the ice sheet retreat from the Bank to the retreat from the south--west of Sweden (Björck, Möller, 1987).

The extent and level of the Baltic Ice Lake water varied greatly. Within about 11.2 to 10.4–10.3 ka BP a rapid regression and a somewhat slower transgression occurred (Figs. 6, 8). Sea level variations reached 25–30 m. The water level could have fallen at a rate of 100–200 mm/a, the rise rate approaching about 40–45 mm/a. Such extensive and fast changes of water level must have brought about catastrophic changes in the shoreline location. Within about 100 years, the shoreline moved at first by about 30–40 km north to retreat, at a slightly lower rate, back south. Most probably, this is the period that produced progradational deltaic structures in the southeastern part of the Gulf of Gdańsk, barrier structures on the western slopes of the Gdańsk Basin (Figs. 16, 17) and erosional surfaces of tills on the western slopes of the Bornholm Basin and the Słupsk Furrow (Fig. 18). As the Baltic Ice Lake evolved and the water level rose following the first regression, the clay sedimentation range extended. Lithological characteristics of the deposits were changing as well: looking from the lowermost to the uppermost layers, they change from varved into microvarved and then into homogenous. Sedimentation was of a glacio-marine (glacio-lacustrine) type. Clayey sediments contain sandy inclusions and single grains of gravel. The total thickness of varved, microlaminated and homogenous layers overlying the tills is about 10 m in central parts of the deep basins. Characteristic for those sediments is a low organic matter content, usually less than 1.5%, and a considerable (up to 20%) content of carbonates. The top parts of the Baltic Ice Lake clays may be carbonate-free, with occasional black laminae or spots marking accumulations of iron sulphides, the colour changing from light brown to brown-grey. The Baltic Ice Lake clays in the southern Baltic usually contain no pollen or diatoms. Those sediments usually reach the 45 m (locally 35 m) depth contour on the Bornholm Basin slopes. On the slopes of the Southern Middle Bank, too, those sediments are found at depth larger than 40 m. This pattern emerged at the final phase of the Baltic Ice Lake when the lake attained its maximum size and the water level was by about 25–26 m lower than at present (Figs. 6, 43). The erosional nature of the boundaries of the Baltic Ice Lake's clayey sediment is related to later processes. Within *c.* 13.0–10.3 ka BP, the northern boundary of the Baltic Ice Lake moved north, the ice sheet margin retreating north at that time by about 350–400 km. From the north and north-west, the Baltic Ice Lake was bounded by the receding ice sheet margin. At the terminal phase of the Baltic Ice Lake, *c.* 10.5–10.3 ka BP, central-Swedish and Salpausselkä moraines were formed along with their continuation on the present seafloor of the northern Baltic Proper (Södeberg, 1988).

In the southern part of the Baltic Ice Lake, the coast migrated from the present shore line by about: 60 km in the Pomeranian Bay, 20 km near Kołobrzeg and 10 km in the central coast and the Gulf of Gdańsk (Fig. 43). The Southern Middle Bank and the Słupsk Bank were large islands. Bornholm was connected, via a narrow isthmus, running across the Rønne Bank and Adlergrund (Eagle Bank), with the land mass that was at that time present in the area of Odra Bank and Pomeranian Bay. It is possible that, towards the end of the Baltic Ice Lake, at the highest water level, the isthmus between Odra Bank and Adlergrund was flooded. Despite the smaller size of the lake and colder climate, the coastal processes on the southern shores were similar to those prevailing at present. The western part of the Southern Middle Bank and northern flanks of the Słupsk Bank were cliffs. Cliffs extended also along the stretch of the Baltic Ice Lake north of lakes Jamno and Kopań. Barrier shores occurred on the southern and southeastern coasts of the Southern Middle Bank and on the southern shores of the Słupsk Bank; they developed there on glaciofluvial deltas. In addition, long stretches of barrier shores extended north of the present Pomeranian Bay and Kołobrzeg as well as from Ustka to the Gulf of Gdańsk. In those areas, Pleistocene deposits were an abundant source of sand; possibly, that material was also supplied by the Eemian Sea sands. The deposits of that age, occurring at a similar hypsometric level, are known from the Vistula Spit (e.g. Makowska, 1986; Mojski, 1988; Tomczak *et al*., 1989). The barrier coast stretch between Ustka and the Gulf of Gdańsk was separated by a short cliff shore near the Stilo Bank. The Puck Bay (western part of the Gulf of Gdańsk) was at that time a land and the Hel Peninsula had not yet emerged. In the southern part of the Gulf of Gdańsk, the Vistula mouth was developing vigorously. During the first phase of the Baltic Ice Lake, the delta front progradation structures were emerging fast. The development was stopped by the first Baltic Ice Lake regression *c.* 11.2–11.1 ka BP, which resulted in a rapid formation of erosional incisions within the delta. The next water level rise induced accumulation of sediments carried in by the Vistula, most probably in an area east of the oldest washover fan. It may be suggested that in

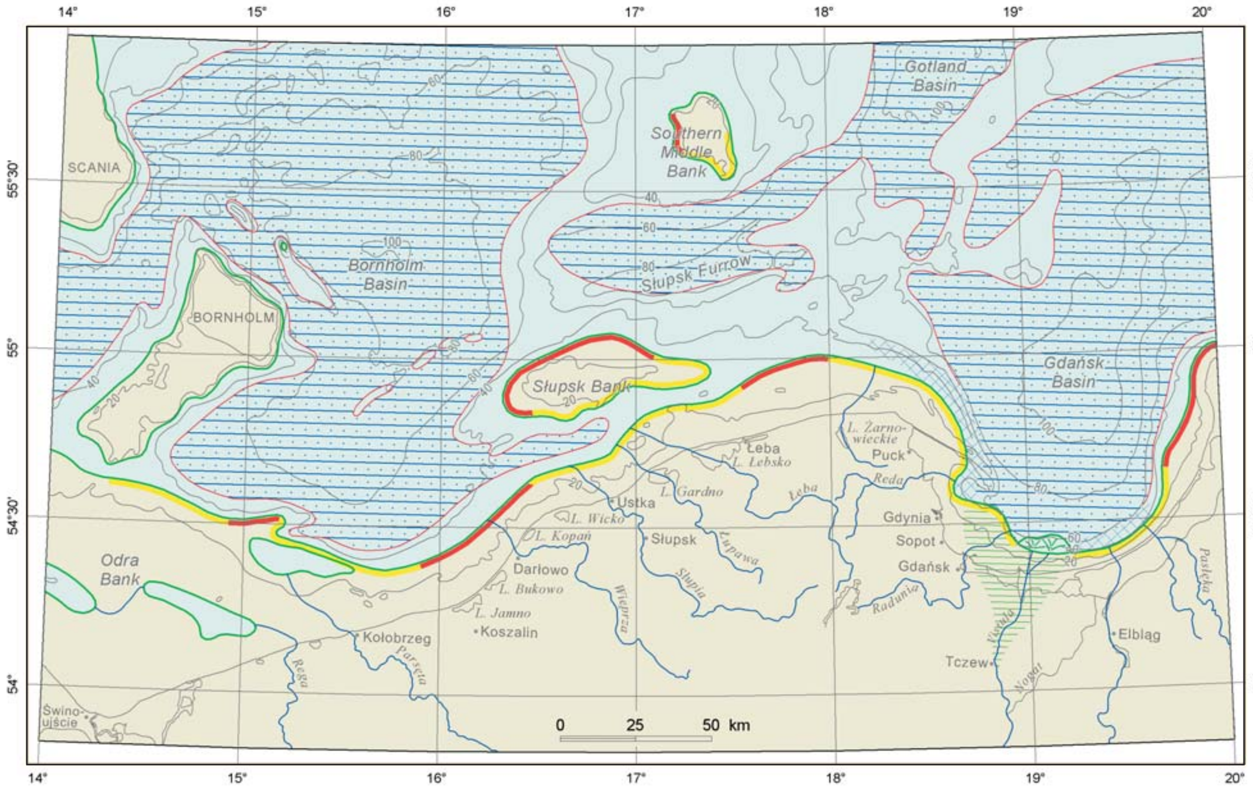


Fig. 43. The final phase of the Baltic Ice Lake (*c.* 10.3 ka BP) (after Uścinowicz in: Mojski *et al.*, eds., 1995; modified)

For explanations see Figure 42

the cold climate of the Younger Dryas, with the erosional base located several tens of metres lower that today, the Vistula was transporting more sediment, the emerging mouth fans being large and accreting fast. Deltaic structures situated at depths of 30–60 m b.s.l., which showed a fast accretion as well, are much larger than the younger, higher situated deltaic forms making up the bottom of the Gulf of Gdańsk. This is clearly seen on seismoacoustic profiles.

The termination of the Baltic Ice Lake was a catastrophic event. The Baltic Ice Lake drained within a period estimated at a few years (e.g. Svensson, 1989, 1991; Björck, 1995) to about 90 years (Strömberg, 1992). The water level was falling at a rate of about 0.3 to 3 m a year, the shoreline moving north over 30–40 km. Depending on the true drainage rate and local tilting of the seafloor, the land mass accreted at a rate of 0.3 to 4 km a year.

SOUTHERN BALTIC SHORELINES IN HOLOCENE

Dynamics of the southern Baltic coast evolution in Early Holocene was still dependent on glacio-isostatic rebound. The importance of differences in uplift rates in the northern and southern areas of the Baltic was particularly emphasised when the Baltic was isolated from the ocean. Effects of vertical crustal movements on evolution of the coast diminished in Middle and Late Holocene. Due to the presence of the permanent connection of the Baltic with the ocean, the dominant role was taken on initially by the glacio-eustatic sea level rise. In Late Holocene, the importance of glacio-eustasy in the Baltic declined along with the reduction of its importance at the global scale.

End of Late Pleistocene and Early Holocene: the Yoldia Sea and the Ancylus Lake (10.3–8.5 ka BP)

Recession of the ice sheet margin and opening of a strait near Mt. Billingen in central Sweden (the Narke Strait) (Fig. 20) in Late Pleistocene resulted in regression of the Baltic Ice Lake and dropping of the sea level to that of the ocean, i.e. falling by about 50–52 m below the present level. This initiated the next stage in the Baltic history, called the Yoldia Sea from the name of the bivalve *Yoldia arctica* (Gray). The bivalve is known to occur in sediments of central Sweden, but has never been found in the southern Baltic. Initially, the southern coast of the Yoldia Sea was located far north from the present shoreline (Fig. 44). The area covering the Pomeranian Bay and Odra Bank connected, via a wide isthmus, with the Eagle Bank (Adlergrund), Rønne Bank and Bornholm. South-east of Bornholm there were two islands: Christiansø, much larger than that known to-

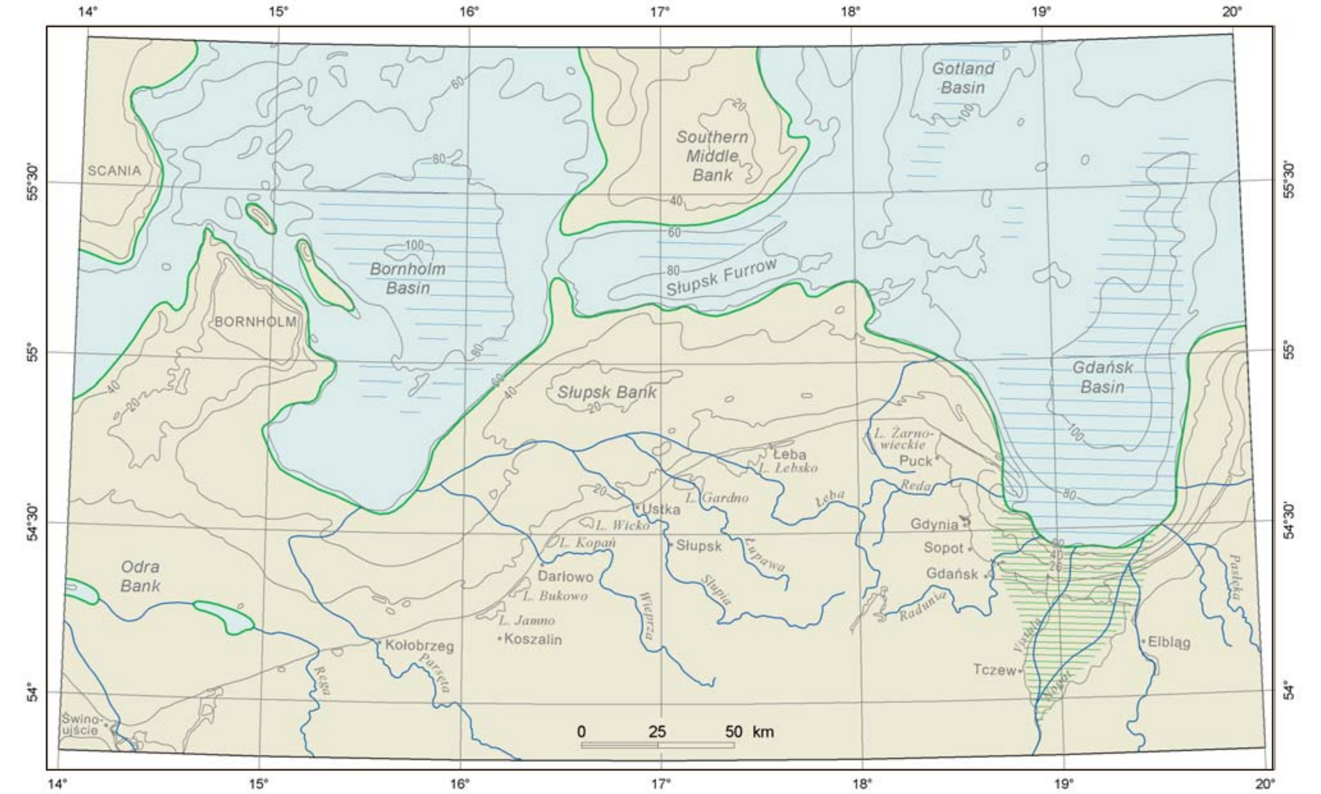


Fig. 44. The early phase of the Yoldia Sea (*c.* 10.0 ka BP) (after Uścinowicz in: Mojski *et al.*, eds., 1995; modified)

For explanations see Figure 42

day and the other one in the area of the present shoals south-east of it. The Southern Middle Bank together with the Northern Middle Bank formed a large island separated from the southern Baltic coast by the Słupsk Furrow. Water exchange through the Furrow was hampered by a sill bordering the Bornholm Basin. The Słupsk Bank was connected with the land mass, the shoreline between Kołobrzeg and the Gulf of Gdańsk being located by about 30–50 km away from the present-day coast. In the Gulf of Gdańsk, the shoreline run north of the present coast, removed from it at a distance of from 20 km in the west to 10 km in the east. It was at that time when, at the lowest Baltic sea level in Holocene, the deepest known erosional incision in the area of the today's Vistula delta, deeper than 40 m near Krynica Morska (Mojski, 1988), emerged. It most likely provided a route, north of the present-day Vistula Lagoon, for the Vistula discharge into the Gulf of Gdańsk.

Between 10.2 and 9.6 ka BP, the Yoldia Sea water level rose by about 10–12 m at a rate of about 15–20 mm/a. The sea level rise rate was only slightly slower than the eustatic ocean level rise. At that time, the southern Baltic area was still subjected to residual uplift. The nature of the coast at that time was most probably completely different than that at present. After a rapid fall of the water level, the Yoldia Sea shores began to evolve on clayey sediments of the Baltic Ice Lake and the erosional boundary of those sediment was located at the depth of 45 m (locally at 35 m) below the present sea level. So far, no direct evidence of sandy sediments from the Preboreal that would be at present located 40–60 m b.s.l. have been found. However, seismoacoustic records indicate that sandy shores could have been present within the today's Puck Bay (western part of the Gulf of Gdańsk) and north of the Hel Peninsula. Only in the deepest parts of the Baltic deeps, where sedimentation proceeded all the time, were the Yoldia Sea deposits retained. Those are brown-grey clayey-silty. As a rule, they contain no carbonates, show the presence of iron sulphide accumulations and a small (up to 2%) amount of organic matter, and form no clearly defined lithostratigraphic level.

The closure, due to uplift, of the connection between the Yoldia Sea and the ocean and transformation of the Yoldia Sea into the Ancylus Lake, named after the freshwater snail *Ancylus fluviatilis* Müller (not found so far in the southern Baltic either) was reflected on the southern coast only by accelerated sea level rise rate. In deep basins with continuing sedimentation, the brown-grey Yoldia Sea deposits grade smoothly, without any clear separation, into grey and pale grey clays of the Ancylus Lake. These contain numerous accumulations or laminae of iron sulphide. Apart from the absence of carbonates, characteristic is a higher content of organic matter, compared to deposits from previous stages of the Baltic evolution. The clayey Ancylus Lake sediments form a continuous layer only in sedimentation basins at depths larger than 65 m. On slopes of the basins, between the 55 and 65 isobaths, particularly in the area between the Słupsk Furrow and the Gotland Basin, the Ancylus Lake deposits — rather than forming a continuous

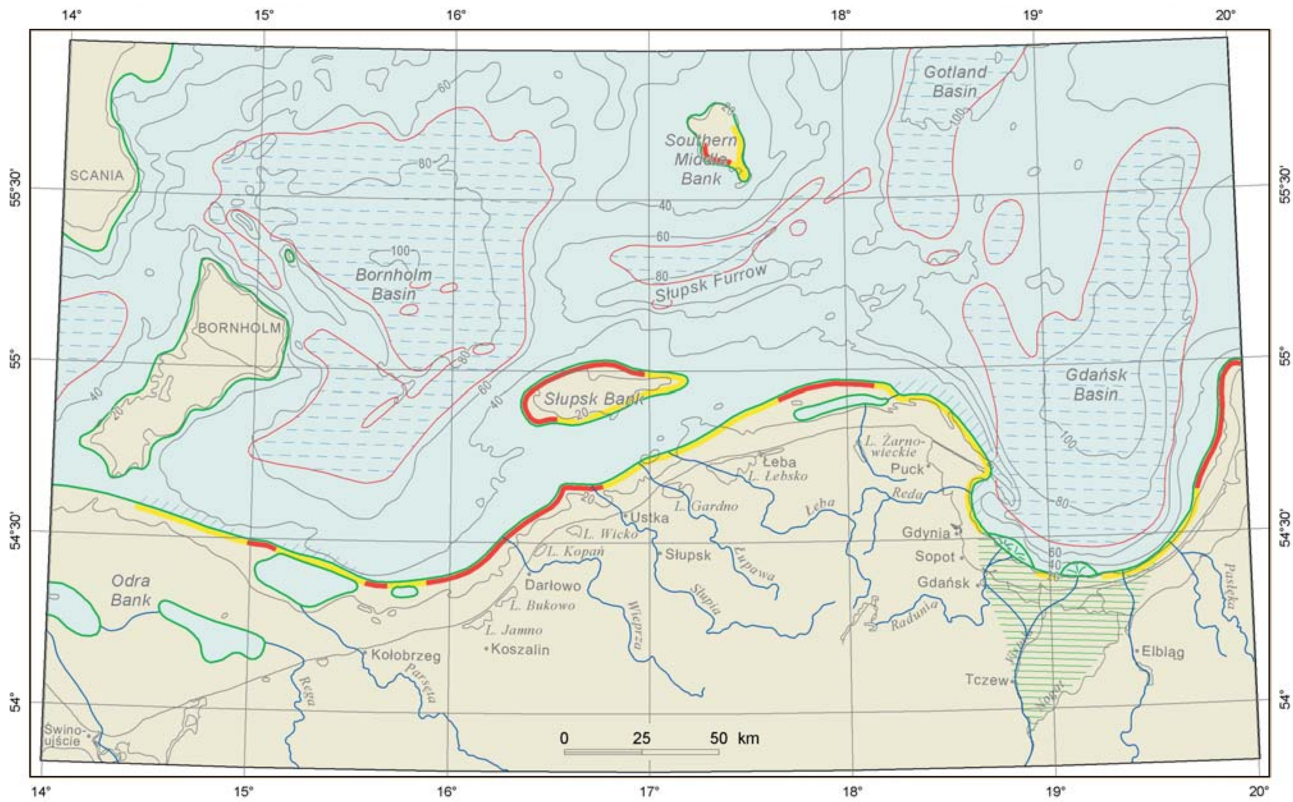


Fig. 45. The maximum extent of the Ancylus Lake (*c.* 9.2 ka BP)
(after Uścinowicz in: Mojski *et al*., eds., 1995; modified)

For explanations see Figure 42

layer — frequently occur as isolated patches filling depressions in an older base. The thickness of those sediments is difficult to determine due to their lithological similarity to the underlying Yoldia Sea clays and continuous sedimentation durig those two phases. The total thickness of the Yoldia Sea and Ancylus Lake deposits ranges from 3 to 8 m.

The high rate of sea level rise on the southern shores of the Ancylus Lake within 9.6–9.2 ka BP, reaching up to 35–45 mm/a, resulted in the predomination of destruction/erosion processes. The Baltic Ice Lake clayey deposits, located at depths shallower than 40 m, were almost completely eroded away. Eroded was also the top of tills occurring at depths of 25–40 m b.s.l. Repetitive erosional processes (the original ones were associated with regression and transgression of the Baltic Ice Lake during 11.2–10.3 ka BP) led to an almost complete flattening of the original relief and to the emergence of extensive erosion surfaces, slightly tilted northwards. At the final stage of the Ancylus Lake transgression *c.* 9.2 ka BP, the water level was about 25–26 m lower than at present (Fig. 45). The shoreline migrations, in spite of the very fast sea level rise, were relatively insignificant. Over 300 years, the shore moved south by not more than about 10 km in the southern part of the Bornholm Basin and by about 5 km in the Gdańsk Basin. That was caused by relatively steep northward tilting of the slopes of those basins. Catastrophic migrations of the shoreline occurred at that time south of the Słupsk Bank. The shores of a gulf formed off the Bornholm Basin side were moving east at a rate of about 150–200 m a year. Over 300 years, the shoreline moved by about 50–60 km, cutting the Słupsk Bank off the land. The connection between the land mass and the Bornholm and Eagle Bank (Adlergrund) was cut off at a somewhat slower rate. At the terminal stage of the Ancylus Lake transgression, the sea level was close to the maximum level of the Baltic Ice Lake *c.* 10.3 ka BP. The outline and nature of the coast was in all likelihood similar, too, the only modification being caused by the repeated erosion under the weakening uplift (Fig. 45).

The Southern Middle Bank and the Słupsk Bank became islands again (Fig. 45). Bornholm, together with the Rønne Bank and Eagle Bank (Adlergrund), was also a large island, separated with a narrow strait only from the land mass reaching north beyond the Odra Bank. The southern coast was farther away from the present shoreline by about 60 km in the Pomeranian Bay. East of the Odra Bank, near Kołobrzeg, the coast was located about 20 km north relative to the present shoreline, while in the central coast and in the Gulf of Gdańsk it moved north by about 10 km. As before, the cliff shore was evolving on the western flanks of the Southern Middle Bank and on the northern flanks of the Słupsk Bank as well as on the section of the shore between lakes Jamno and Kopań. A short stretch of a cliff coast occurred also near the Stilo Bank. It is possible that the bank was also an island separated from the shore by a shallow and narrow strait. Barrier coast was present on the southern and southeastern flanks of the Southern Middle Bank and on the southern margins of the Słupsk Bank. Additionally, barrier coast extended along long stretches north of the present Pomeranian Bay and Kołobrzeg and from Ustka to the Gulf of Gdańsk. The barriers in those areas emerged as a result of transformation of the earlier barriers of the Baltic Ice Lake an abundant source of materials for which were the Pleistocene deposits. Such an outline of the coast persisted for about 800 years, i.e. for a larger part of the Ancylus Lake stage. A small regression that occurred at that time resulted in the emergence of sand spit structures in the southern part of the Bornholm Basin (Figs. 28, 29) and possibly also in the western and southern parts of the Gdańsk Basin. The shoreline recession resulted most likely in termination of the erosion along considerable stretches of the cliff shore. A possibility that the Stilo and Słupsk Banks could have been periodically connected with the land mass cannot be ruled out, either. The first half of the Boreal was a period of a relative stabilisation of the southern coast of the contemporaneous Baltic. The Ancylus Lake water level was then higher than that of the ocean. The relative stabilisation of water level on the southern coast was maintained by the equilibrium between the inflow of meltwater and river discharge on the one hand and the outflow through the Danish Strait on the other. Vertical crust movements associated with the glacio-isostatic rebound on the southern coast within 9.2–8.5 ka BP were already very weak, their importance for the coastal zone evolution being very limited.

End of Early Holocene and Middle Holocene: Mastogloia Sea and Littorina Sea (8.5–5.0 ka BP)

Incursions of saline waters into the Baltic through the Danish Straits began about 8.5–8.4 ka BP or perhaps somewhat earlier, when the water level was by about 28 m lower that at present. The incursions were a consequence of the eustatic rise of the World Ocean's level. A new transgression, termed in the Baltic countries the Atlantic or Littorina transgression and corresponding to the North Sea transgression called Flandrian transgression in western Europe, began on the southern Baltic coast. Although enjoying a long-established tradition, the name "Atlantic" or "Littorina transgression" is not entirely appropriate. The sea level rise in the southern Baltic area began *c.* 8.5–8.4 ka BP, i.e. in mid-Boreal, with the onset of the Mastogloia Sea. Thus the *c.* 400–500 years-long transgression proceeded during the Boreal rather than the Atlantic, while the 1000-year-long transgression proceeded during the Mastogloia rather than the Littorina Sea. Those remarks are made on the assumption that the newest findings on the evolutionary stages of the Baltic and their duration (e.g. Björck, 1995; Eronen, 1988; Hyvärinen, 1988, 2000; Svensson, 1991) are correct. During the first millenium of the transgression (Mastogloia Sea), the water level rose by about 13 m. The Littorina Sea level increased during 7.5–5.0 ka BP by a further 12 m or so. Terminology, however, is not the most important issue for palaeogeographic considerations. Important is the rate and extent of the transgression. In the Late Boreal, the water level increased at a rate of about 14 mm/a, faster than that of the eustatic ocean level rise. During the first millennium of the Atlantic, the mean rate of water level rise in the southern Baltic dropped to about 11 mm/a, but was still slightly higher that the rate of eustatic ocean level rise. The reason was the subsidence of the southern Baltic coast at that time (Fig. 37). The water level rise rate on the southern Baltic coast approached that of the ocean *c.* 6–5 ka BP.

The fast sea level rise during the first millennium following the Baltic Sea connection with the ocean via the Danish Straits resulted in pronounced shoreline changes. The transgression affected those areas of the shallow Baltic seafloor which, beginning from Late Pleistocene, had been developing under terrestrial conditions. The transgression was accompanied by erosional processes that wiped out wide expanses of both Late Pleistocene and Early and Middle Holocene terrestrial deposits. Thermoluminescence datings of the till top (Kramarska, Tomczak, 1988; Kramarska *et al.*, 1990) showed the tills formed during the last glaciation to have succumbed to erosion as well. The transgression left its mark on the erosional surfaces, commonly present in the top of older sediments, frequently overlain by a layer of sandy-gravelly deposits forming erosion pavement. In the Bornholm, Gdańsk and Gotland basins, particularly in their central parts where sedimentation was continuous, no marked and abrupt changes in the sediment profiles are visible. The pale grey colour of the Ancylus Lake clays grades into pale-grey with olive-green hue, the organic matter content increasing gradually as well. The silty-clayey sediments of the transition stage (Mastogloia Sea) are frequently laminated. Saline inflows, emergence of water column haline stratification, and periodic internal waves in the Baltic initiated erosional processes on the outskirts of the sedimentary basins and on the sills separating them (Kramarska *et al.*, 2002). Those areas lack any Mastogloia Sea deposits. The Ancylus Lake sediment surface is erosional, the sediments being completely eroded locally.

Within 8.5–7.5 ka BP, the shore receded south by about 15–20 km in the Odra Bank, by about 5–10 km along the central coast and by less than 5 km in the Gulf of Gdańsk. In the Gulf, the cause of the small extent of shoreline migration despite the 13 m water level rise, was a relatively steep tilt of the seafloor. After the water level rose to become lower by about 15 m than at present, the island made up by Bornholm, the Rønne Bank and the Eagle Bank (Adlergrund) disintegrated into three parts, Bornholm remaining the largest of them (Fig. 46). The Słupsk Bank island reduced its area as well. Its northern coast, with marginal formations of the last glaciation, turned into intensively eroded cliffs. In the southern part of the Słupsk Bank and south of it, within the present-day Czołpino Shallow, barriers the relicts of which are still discernible today were forming. Barrier shores were also present on the northern slopes of the Odra Bank and along an area from Ustka to the Gulf of Gdańsk inclusive. All those areas show thick sand covers, seismoacoustic profiles showing locally

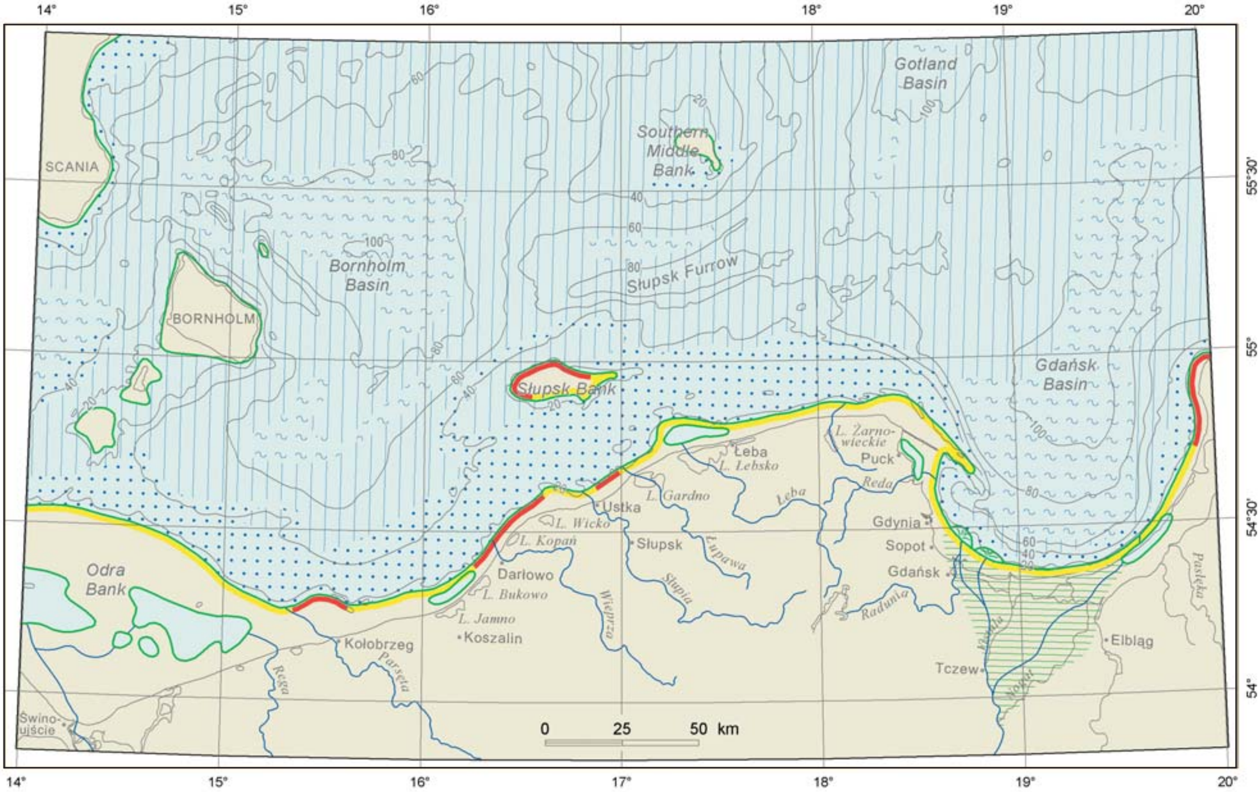


Fig. 46. The early phase of the Littorina Sea (*c.* 7.5 ka BP) (after Uścinowicz in: Mojski *et al*., eds., 1995; modified)

For explanations see Figure 42

the presence of submerged barrier relicts. Most of the cliffs known today had not emerged yet. Cliffs were forming north of Kołobrzeg and north of lakes Jamno, Kopań and Wicko. All those areas show the presence on the seafloor of extensive coarse (including boulders) placer deposits which most probably are relicts of eroded moraine elevations. In the Gulf of Gdańsk, cliffs were present only on the slopes of the Sambia Peninsula, the cliffs forming on Tertiary deposits. At that time, cliffs on the western shores of the Gulf of Gdańsk and north of the present cliff coast between Jastrzębia Góra, Rozewie and Władysławowo had not been formed yet. About 3–5 km north of that fragment of the coast, a barrier shore marking the budding Hel Peninsula began to develop. As the sea level rose, the barrier migrated toward south-west, the alongshore bedload transport from north-west to south-east supplying sands for the emerging Hel Peninsula to be formed south-east of Jastarnia, on the Gulf of Gdańsk slope. Most similar to the present coast was the shore in the eastern part of the Gulf of Gdańsk. At 14–15 m b.s.l. on the Vistula Spit (core Piaski 2a) and Krynica Morska (core K-5) (Fig. 7; Appendix 1), peats dated to 7560 ±90 and 7590 ±70 years BP are present. In the Vistula Lagoon, near Piaski, at the level of about –15 m, there are Early Atlantic muds the pollen spectrum of which indicates the presence of a shallow water body (Zachowicz, 1985), a predecessor of the Vistula Lagoon.

The on-going sea level rise along with changing atmospheric circulation patterns, still generated by glacio-eustasy, resulted *c.* 7.5 ka BP in an intensified water exchange through the Danish Straits. Salinity of the Baltic water increased to a level higher than that at present. The increase in salinity and temperature resulted in the immigration of new species of flora and fauna into the Baltic. Diatoms requiring more saline water appeared and the dominants were planktonic eu- and mesohalobous species. The shallow areas in the southern and central Baltic were inhabited by the gastropod *Littorina littorea* (Linnaeus), the name of which gave the name to that period in the Baltic history (the Littorina Sea). At present, the gastropod occurs in the western Baltic only. On the Polish coast, the shells of *L. littorea* were found at a single site only, in the Czołpino drilling well (Brodniewicz, Rosa, 1967).

Accumulation of silty-clayey sediments that started during the Mastogloia Sea, was continuing. The Littorina Sea deposits are grey and dark grey with an olive-green hue, laminated at places. They are organic matter-enriched, the organic content reaching 10%, and contain no carbonates. As the sea level rose and transgression progressed, erosion continued and erosional surfaces emerged; those surfaces are frequently overlain by a thin layer of sandy-gravelly deposits. Formation of the upper member of the Middle and Late Holocene Baltic sediments, i.e. a sandy cover made up primarily by fine and locally medium, sands began.

As already mentioned, the Littorina Sea water level between 7.5–5.0 ka BP rose from about –15 m to about –2.5 m. The sea level rise rate was constantly decreasing (Table 1). Although the mean rate within 7.5–7.0 ka BP was still about 10 mm/a, it decreased to as little as about 3.0 mm/a during 6.5–6.0 ka BP. The shoreline migrated at a steadily decreasing rate. During the Middle Atlantic, there were still most probably small islands, apart from Bornholm, within the Eagle Bank (Adlergrund) and the Słupsk Bank where the last fragments of moraine elevations were being eroded away.

About 6.2–5.8 ka BP, at the sea level lower than at present by about 5.0–6.5 m, the southern Baltic coast witnessed a series of events that resulted in the shoreline being very similar to that known at present. At that time, the Pomeranian Bay and Odra Bank were still a part of the land mass, although the sand barrier running along the northern slopes of the Odra Bank south-east towards Kołobrzeg, after a rapid transgression in the first half of the Atlantic, was poorly developed. The fact that the relicts of that barrier are located at present at depths of 8–10 m does not have to be taken as evidence of its original height. A substantial part of the barrier could have been eroded later on. The central part of the Odra Bank has still the depth of about 5 m. Depressions south of the barrier contained lakes that were gradually transforming into brackish lagoons (e.g. core R 74; Fig. 32). The Szczecin Lagoon area was covered by marshes and lakes. About 6.2 ka BP, the Szczecin Lagoon experienced the onset of seawater incursion. The inflow is marked as an extensive layer of sands, up to 1.5 m thick, containing marine fauna, including shells of *Cardium* (*Cerastoderma*) *glaucum* Bruguiére (Borówka *et al*., 2001a, b). The intrusion of sea water into the Szczecin Lagoon could have been rapid. Lagoon sedimentation does not, as a rule, begins with such a well-developed sand series. A rapid inflow of marine water into the lower lying areas of the present-day Pomeranian Bay and Szczecin Lagoon could have been a result of the sandy barrier in the Odra Bank being broken, whereby a catastrophic migration of shoreline by about 20–40 m southward ensued. The Świna Spit, connecting the moraine hills on the Wolin and Uznam islands had not emerged yet (Prusinkiewicz, Noryśkiewicz, 1966). The inflowing marine water could penetrate, unhampered, into the Szczecin Lagoon area. Rosa (1963) had already argued that this could have been a pathway along which that part of the coast evolved.

A similar situation, whereby *Cardium* (*Cerastoderma*) sp. shells-containing sands underlie lagoonal muds, is encountered on the Gardno–Łeba Lowland (Rotnicki, 1999; Rotnicki *et al*., 1999), within Lake Sarbsko (Miotk, Bogaczewicz-Adamczak, 1987) and in the northeastern part of the Puck Lagoon (Kuźnica Hollow) (Kramarska *et al*., 1995). The hypsometric location of the marine sand layer in all those areas is similar. The sand series base occurs usually at the ordinate 9.5–12 m b.s.l. The age of that series has been most accurately determined on the Łeba Spit (Barrier) where a number of radiocarbon datings of the *Cardium* sp. shells collected at depths of 9.3–11.3 m b.s.l. placed them within 6.8 do 6.66 ka BP (Rotnicki *et al*., 1999). After including the reservoir effect, the datings correspond very well to the period of emergence of a similar layer in the Szczecin Lagoon.

Similar events, albeit proceeding perhaps less rapidly, were at that time taking place in other parts of the coast. The formation of Lake Jamno has been dated to the Middle Atlantic. *C.* 6.2 ka BP, deposition of muds with *Cardium* sp. shells began in the lake (Dąbrowski *et al*., 1985). The base of halophilous diatom-containing muds in lakes Bukowo and Jamno occurs at the depth of about 4.5–5.0 m b.s.l. The muds directly overlie peats. The peats of the overlying lacustrine sediments occur at present also at the forefront of the Bukowo and Jamno barriers (Wypych, 1973). Hence it may be inferred that these lagoons developed in their present place by migration accompanying the sea level rise and displacement of the barriers separating them from the sea.

At the same time, in the Gulf of Gdańsk area, the Vistula Lagoon began expanding westwards. The Lagoon's water ingressed into the northern part of the Vistula Delta and stopped peat formation. The youngest peat was dated to 6.33 ka BP (Mojski, 1988; Tomczak *et al*., 1989). The onset of halophilous diatom-containing mud deposition in Lake Druzno has been dated to the similar period of time, *c.* 6.44 ka BP. The lake was then a part of the Vistula Lagoon (Zachowicz, Kempińska, 1987; Zachowicz *et al*., 1982). The Vistula Spit at that time was located at its today's position. The highest dunes, so-called yellow dunes, had not been formed yet (Tomczak, 1995a; Tomczak in: Mojski *et al*., eds., 1995). Thus the Spit was much narrower, lower, and consequently less resistant to storm wave action, so it could have been broken during storms. That this was indeed taking place is indicated by sands of storm surge washover fans intercalating with lagoonal muds (Zachowicz, Uścinowicz, 1997). It was then that amber-bearing sands accumulated in the southern part of the Gulf of Gdańsk at the boundary of the Vistula Spit and the Delta. Between 7.0–6.0 ka BP, on the Sambia Peninsula, at the sea level lower by 5–10 m than at present, the Eocene amber-bearing deposits were washed away. Amber was transported, along with sand, towards south-west by the longshore currents along the Vistula Spit. During storms and wave overflow over the low barrier, sediments of washover fans were accumulating, amber bearing deposited in the distal parts of the fans along with plant detritus. A layer of sand mixed with plant detritus, containing amber accumulations, occur along the southern coast of the Gulf of Gdańsk at depths of about 4–10 m b.s.l. Lagoonal deposits and marine sands occur also on the western coast of the Gulf, north of Gdańsk. The location of the cliff, inactive at present, indicates that at that time the sea ingressed over the present-day shoreline by about 0.5 km near Sopot to about 1 km near Gdańsk. The southern Baltic water level was still lower than the present one by about 5 m.

Towards the end of the Atlantic, when the transgression was weakening, most of the barriers on the southern Baltic coast were narrow and low. When the barriers were weak and extensive lowlands formed in the hinterland, the barriers were being frequently broken by the surging sea water. At those areas, the lagoons were poorly isolated from sea water incursions and it was then that marine fauna-containing sand layers were deposited. Where the barriers were stronger and/or there were no extensive lowlands in the hinterland, lagoons developed gradually. As the sea level rose, the peats were transformed into muddy peats and peaty muds to eventually become lagoonal muds.

The cause of coincidence in a series of phenomena should not be ascribed chiefly to the rapidity of transgression, associated with glacio-eustasy, because the transgression slowed down. The main cause was the poor development of sandy barriers.

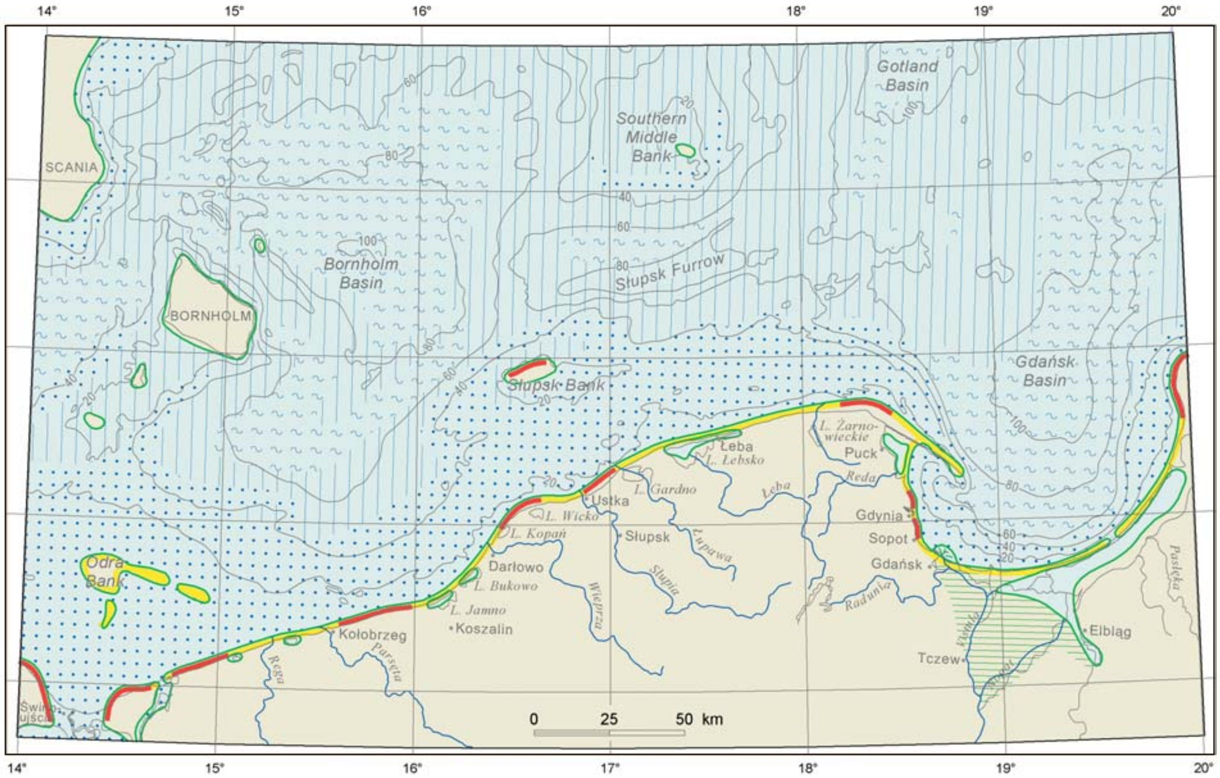


Fig. 47. The middle phase of the Littorina Sea (*c.* 6.0 ka BP)

For explanations see Figure 42

The other factor was the physiography of the area. The hinterland of the barriers in the Pomeranian Bay, Gardno–Łeba Lowland and Vistula Delta formed depressions prone to marine inflows. The third factor, perhaps the most important one, could be the climate. At that time, barographic and wind-driven surges were most likely frequent and high, heavy storms occurring often as well. A change in the North Atlantic climate (e.g. Mörner, 1976, 1980a) resulted in one of the most substantial episodes of water level oscillation at that time. In Blekinge, Berglund (1971) found evidence of a eustatic oscillation (the so-called third Littorina Sea transgression) during 5.9–5.6 ka BP, with the sea level rising by about 1.2 m (at a rate of about 4 mm/a). According to Digerfeldt (1975), the transgression during 6.1–5.7 ka BP left its mark in Skåne by a sea level rise by about 1 m (2.5 mm/a). The largest and fastest eustatic oscillation occurred, according to Mörner (1976, 1980a), during *c.* 6.3 to 5.8 ka BP in the Kattegat area where the sea level rose at that time by about 2.5–2.7 m at a rate of about 6–7 mm/a.

Following the events described above, the shoreline was located very close to its present position, but it was more diverse. In the open sea, the islands within the Eagle Bank (Adlergrund) and the Słupsk Bank finally perished. An island was presumably still existing in the Odra Bank area (Fig. 47). The cliff and barrier shores occurred at locations close to those found at present. However, the picture was not exactly identical. Depending on configuration and lithology of the deposits subjected to marine transgression, fragments of both cliff and barrier shores could have been moved forward or backward relative to the present shoreline. Along its considerable length, the shoreline migrated north of its present position, by, however, not more than about 1–2 km. On the western coast of the Gulf of Gdańsk, between Gdańsk and Sopot, the shoreline was still moving back relative to its present position.

No major changes occurred during the last 1000 years of the Atlantic. The sea level was rising at a mean rate of about 2.5 mm/a to reach, 5 ka BP, the level by about 2.5 m lower than at present. It brought about only slight changes in relation to the picture described above. Areas of, i.a., the present Lake Gardno (Bogaczewicz-Adamczak, Miotk, 1985a; Zachowicz, Zaborowska, 1985) and a considerable part of the Puck Lagoon (Kramarska *et al*., 1995; Uścinowicz, Miotk-Szpiganowicz, 2003) were a land mass. The shore between Władysławowo and Chałupy was a barrier located somewhat north-west of the present one. The Hel Peninsula emerged in the vicinity of today's Chałupy (most of the Puck Lagoon was a land) and, up to Kuźnice, was a barrier separating the emerging Puck Lagoon from the open sea. That section of the Hel Peninsula was a narrow and weak barrier, frequently broken during storms. Towards the end of the Atlantic the Hel Peninsula reached beyond Jastarnia and Jurata, as evidenced by the age of peats. The peat top at Jastarnia, located at 2.5 m b.s.l., was dated to 5.37 ka BP (Bogaczewicz-Adamczak, 1982); at Jurata, it occurs at 0.8 m b.s.l. and was dated to 5.64 ka BP (Tomczak in: Mojski *et al*., eds., 1995).

Late Holocene (from 5.0 ka BP to present): end of the Littorina Sea and emergence of the Post-Littorina Sea

Before the coast evolution during the Subboreal and Subatlantic will be presented, it is necessary to discuss some terminological issues. In the earlier chapters, the terms "sea level rise" and "transgression" were treated as synonyms. Similarly, synonymous were the terms "sea level fall" and "regression". Transgression (Latin *transgressio* = movement forward) is understood as submergence of the land by the sea, i.e. backing up of the shoreline. Regression (Latin *regressio* = backing up) means ebbing of the sea. To use the synonyms was justified because both the rise and the fall of the sea level usually results in the shoreline displacement. This is unquestionable if the sea level changes are extensive and fast. When the sea level rises or falls slightly and slowly, the situation is not always as clear-cut. A frequent cause of misunderstanding is the term "maximum of transgression" without a caveat describing what is actually meant, the maximum sea level or the maximum extent of the sea. The two do not always mean the same. There are situations when, for example, at a stable sea level or during a slight and slow rise a barrier is broken, as a result of which the shoreline moves towards the low-lying hinterland. When that is the case, it seems more appropriate to use the term "ingression" (Latin *ingressio* = entering), althought the etymology of "transgression" does not exclude it being used in this case, either. "Ingression" seems to be an appropriate term to describe the emergence of lagoons with. One cannot, however, use the term "regression" or "recession" (Latin *recessio* = backing up, moving backwards) to describe the shoreline seaward migration at a stable sea level. Shoreline seaward migration occurring as a result of sediment accumulation in the coastal zone (e.g. accretion of a barrier or a delta front) is also possible at a slow sea level rise. At that case, one should use appropriate descriptive phrases. The remarks above are becoming important because towards the end of Middle Holocene, and particularly during Late Holocene, the rate with which the sea level was changing decelerated considerably, while shoreline migrations were occasionally very fast (bearing in mind the time scale under consideration) and quite extensive. Terms such as transgressive phases, regressive phases, or transgressive–regressive phases, frequently used in the literature, refer to the relatively short periods of time (at the Holocene time scale) and small-amplitude sea level changes. It seems that, in the light of the remarks above, a greater precision would be achieved by using the terms "oscillations" or "fluctuations" of the sea level.

The Littorina Sea stage did not terminated at the end of the Atlantic. Climatic changes at the turn of the Atlantic and Subboreal did nor result in synchronous changes of the Baltic hydrology. Salinity remained higher than that at present for at least a few hundred years. A gradual change of the hydrologic regime is not recorded in the sediments as a well-marked boundary (e.g. Eronen, 1988; Andrén *et al*., 2000). The changes may be important for detailed palaeoecological reconstructions or when attempting to analyse details of the Baltic sediment biostratigraphy. For sedimentation processes, detailed geochemical considerations apart, and particularly for the processes affecting the shoreline formation, slight changes in water temperature and/or chemistry are not important.

The Baltic deeps, below the pycnocline, were continuously accumulating muddy-clayey sediments. Thickness of the Middle and Late Holocene (Littorina and Post-Littorina Sea) muddy-clayey sediments is 4–5 m, and reaches as much as 6 m locally. The Post-Littorina Sea sediments differ from those of the Littorina Sea only slightly; they are homogenous and jelly-like in consistency. Reducing conditions were more frequent in the Post-Littorina Sea. The shallow areas above the pycnocline continued to be affected by differentiation of the sedimentary material and by deposition of the sandy-gravelly facies marking the erosion, and particularly of the fine sand subfacies making up the accumulation cover. The sand layers in question are usually not thicker than 2 m. It is only where relicts of accumulation shores occur that the sands are about 6 m thick.

Somewhat more extensive changes were taking place at that time on the coast. After a considerable deceleration of the sea level rise, processes resulting in shore levelling, cliff erosion, enlargement of the Atlantic lagoons and formation of new ones were prevalent. The spits (barriers) were stabilising and strengthening.

During the Subboreal, the sea level rose, to be from about 2.5 to about 0.6–0.7 m lower than at present. A relatively largest and fastest rise occurred during the first millennium of the Subboreal, with the sea level rising at a mean rate of about 1.2–1.4 mm/a to a level lower than that at present by about 1.1–1.3 m. During the subsequent 1500 years, until the end of the Subboreal, the mean sea level rose by a further 0.4–0.7 m. As the sea level rise rate decreased, the shoreline migration slowed down as well. The slow-down of the migration, however, was not uniform along the entire shoreline and depended on the slope of the area and the type of sediments making up the shore.

As already mentioned in the preceding chapter, cliffs and barriers occurred on the shore in locations they occupy at present. The cliffs did not extended further than by about 1.0–1.5 km from the present shoreline, as indicated by the extent of coarse placer deposits (boulders and gravelly sands) in the forefront of today's cliffs (Mojski ed., 1989–1995, 1:200,000 Geological Map of the Baltic Sea Bottom). The extent is much smaller than assumed in some publications (e.g. Subotowicz, 1982). Cliff erosion did not proceeded at an equal rate along the entire coast, but changed presumably both in time and in various cliff stretches of the coast. The mean rate of cliff recession was doubtless slower than that observed at present and did not exceed 0.2–0.3 m/a. There were but a few exceptions from the generalised pattern described above. In some places along the coast, cliffs could have been projecting into the sea by more than 1.5 km, relative to the present shoreline. Conversely, since the Atlantic, the cliffs in the southwestern part of the Gulf of Gdańsk moved landwards, relative the present shoreline. As the sea level rose, upland slopes located more and more northwards became activated, the southern parts of cliffs becoming inactive due to intensively developing accumulation platforms at the cliff's feet. In the southernmost part, near Gdańsk, the cliff was covered by younger sediments deposited on the accumulation platform (Rosa, 1963; Mojski, 2000) to become a fossil cliff. Amber accumulations found

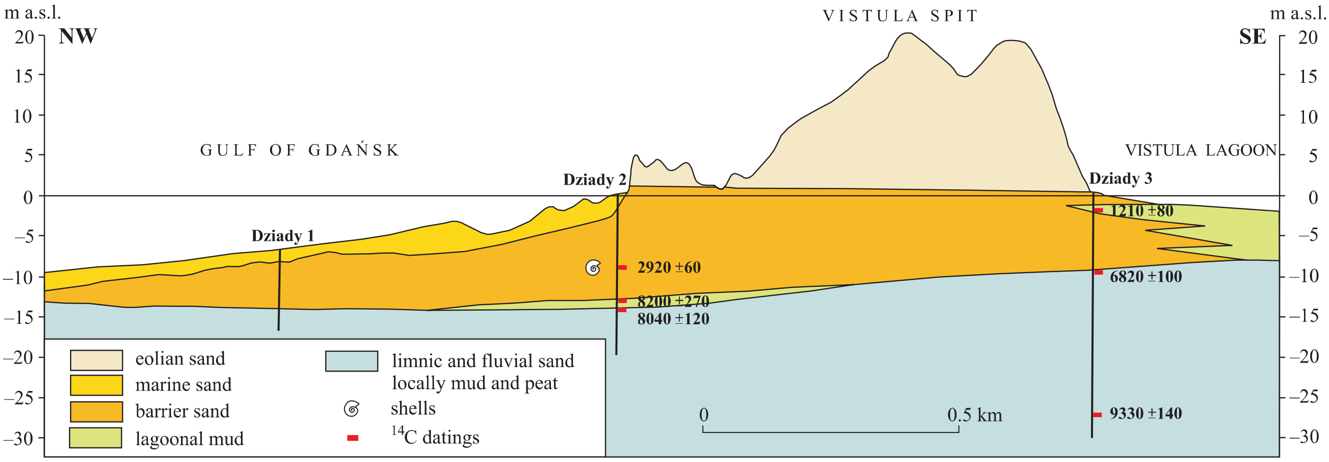


Fig. 48. The Vistula Spit (Dziady) transect

within the accumulation platform sands indicate the sedimentary material to have been derived not only from the eroded cliffs, but to be also transported from the east.

Marine erosion of the morainic upland slopes and cliff development on the western coast of the Puck Lagoon started at the beginning of the Subboreal also. At present, this is the only one lagoon on the Polish coast that has active cliff shores.

Much more extensive changes were taking place during the Subboreal on the barrier shores. Once the barriers had begun to emerge towards the end of the Atlantic, they were intensively evolving. During the Subboreal, two basic barrier types, differing in morphology and the processes governing their evolution, can be discriminated: stationary barriers (accreted) and landward-migrating ones. There are, of course, intermediate formations between the two basic types.

The first barrier type, wide and carrying a well-developed dune system, emerged in areas where, due to peculiarities of the physiography, relatively deep marine ingressions were taking place in the second half of the Atlantic, and sandy material was supplied in sufficiently large amounts. During the Subboreal, at the decelerating sea level rise and with a large supply of sandy material from the eroded cliffs, river mouths, or from the adjacent seafloor, the barriers remained stationary. They were accreting both vertically and seawards. This barrier type is represented by the barriers (spits) of the Świna, Łebsko, Sarbsko and Vistula, which were the fastest to evolve and strengthen. An example is provided by the Vistula Spit where, near Dziady, shells of *Cerastoderma* sp. dated to 2.92 ±0.06 ka BP are found today at about 9 m below the beach surface (Figs. 25, 48). A similar situation occurs near Krynica Morska (core 2, Appendix 1). Intensive sand accumulation occurred not only on the shore, but also on the underwater slope of the Vistula Spit. Accumulations of *Cerastoderma* sp. and *Macoma* sp. shells dated to 4.03 ±0.07 and 3.38 ±0.05 ka BP, respectively, were found near Przebrno and Sztutowo (cores Przebrno 1 and Sztutowo 1; Fig. 7) at ordinate from 6.5 to 8 m below the seafloor located at the depth of 10 m b.s.l.

The other type of barriers, narrow and having poorly developed dune systems, frequently limited to a fore dune only, evolved in areas in which, due to the shore configuration, marine ingressions were not extensive and which experienced a deficiency of sandy material. Such barriers are represented by those of lakes Jamno and Bukowo as well as that part of the Hel Peninsula that separates the Puck Lagoon from the open sea (Fig. 49). These barriers developed as transgressing barriers, entering the lagoons in their hinterland. Subboreal lagoonal deposits dated to 4.1 ±0.14 ka BP were found on the Hel Peninsula near Chałupy, about 6.4 m below the beach surface, while in the Peninsula forefront, at the seafloor depth of 13.7 m outcrops of Atlantic lagoonal sediments were revealed (core R 9; Figs. 33, 49). The same profile (Fig. 49) shows a relict of a submerged barrier from the Littorina Sea transgression, located at present at the depth of 13–15 m b.s.l. Shells of *Mya truncata* Linnaeus and *Thracia papyracea* (Poli), present in the sediments of the submerged barrier (core R 7; Fig. 33) show the sediments to have been formed when the salinity was higher than that in the present southern Baltic. Today, the species mentioned live in western Baltic where the salinity exceeds 15 psu.

The barrier stabilisation at the beginning of the Subboreal is indicated by the development of dunes, the onset of peat formation in depressions between the dunes, and the development of soil-forming processes. The oldest, so-called brown, dunes began to emerge on the Świna and Vistula Spits at the beginning of the Subboreal, their most intensive development occurring during 4.3–3.5 ka BP (Tomczak in: Mojski *et al.*, eds., 1995). The first dune-forming phase on the Łeba Spit is dated to a comparable time, *c.* 4.0–3.8 ka BP (Borówka, 1995). The oldest Świna Spit peats formed in depressions between dunes are dated to 4.81 ±0.365 ka BP, those on the Vistula Spit being dated to 3.92 ±0.06 ka BP. The oldest fossil soils in the Łeba Spit were dated to 3.34 ±0.13 ka BP (Borówka, Tobolski, 1979). Towards the end of the Middle Subboreal, *c.* 3.0 ka BP, the Wicko and Kopań barriers were placed in their respective present positions, the lagoons forming in their hinterlands. The Wicko barrier peats, the top of which is situated at the present sea level, were dated to 3.13 ±0.07 ka BP. Similar (3.24 ±0.05 ka BP) is

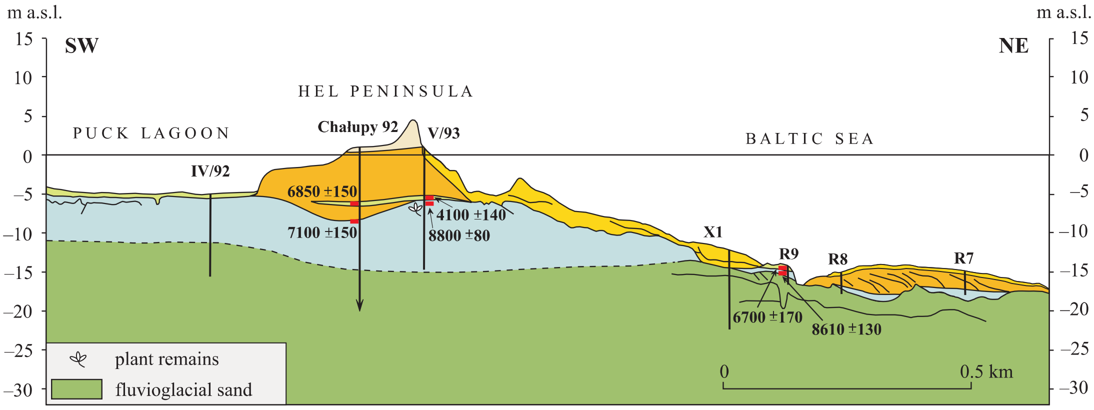


Fig. 49. The Hel Peninsula (Chałupy) transect

For other explanations see Figure 48

the age of the top of the Kopań barrier peat, situated at the present sea level as well. Due to the sea level rise, the Puck Lagoon area became extended northwards so that the Hel Peninsula had its origin near Władysławowo. A narrow barrier separating the Lagoon from the open sea emerged along the stretch connecting Władysławowo, Chałupy and Kuźnica. The Peninsula continued to accrete towards the south-east as a result of the longshore sand transport. The peats occurring about 30 km away from the Peninsula base, at the level of 0.3 m above sea level (a.s.l.) were dated to 3.0 ±0.06 ka BP (Tomczak in: Mojski *et al*., eds., 1995) and are indicative of the Peninsula length at the time of their formation.

The onset of mud deposition differed somewhat between the lagoons present in the hinterland of the barriers and depended mainly on the depositional palaeosurface configuration. Deposition of lagoonal muds began at the earliest, during the Early Atlantic, in the northern part of the Vistula Lagoon (Zachowicz, 1985). In most lagoons, mud deposition began towards the end of the Atlantic or at the turn of the Atlantic and Subboreal. In the Szczecin Lagoon, the onset of mud deposition was dated at about 5 ka BP (Borówka *et al*., 2001a, b). Shells of *Cardium* (*Cerastoderma*) sp. from the mud base beneath the Łeba Spit were dated to 5.37 ±0.28 ka BP (Rotnicki *et al*., 1999). During the first half of the Subboreal, the Puck Lagoon expanded from Kuźnica Hollow to an area the extent of which is close to that covered by the lagoon today (Kramarska *et al*., 1995; Uścinowicz, Miotk-Szpiganowicz, 2003). The lagoonal mud base in Kuźnica Hollow, occurring at 12 m b.s.l., has been pollen-dated to the beginning of the Subboreal (Kramarska *et al*., 1995). The halophilous diatoms-containing base of muddy-sandy sediments (5.15–5.25 m b.s.l.) in the southern part of the Puck Lagoon (core ZP18, Fig. 7) was dated to 5.48 ±0.13 ka BP (Witkowski, Witak, 1993). The mud base (7.85–7.95 m b.s.l.) in the Vistula Lagoon (core ZW 6) was dated to 5.18 ±0.15 ka BP (Uścinowicz, Zachowicz, 1996; Zachowicz, Uścinowicz, 1997). The latter dating was confirmed by pollen analysis. In somewhat shallower areas of the Vistula Lagoon, too, deposition of lagoonal muds began at a similar time. The lagoonal mud base at the depth of 3.25 m b.s.l. was pollen-dated to the turn of the Atlantic and Subboreal (Bogaczewicz-Adamczak, Miotk, 1985b). Some, and most probably most, ^{14}C datings referred to may be overestimates of a varying degree, because of the reservoir effect in the case of shells and additionally because of possible admixtures of older organic matter in the case of lagoonal muds.

Near the end of the Subboreal, Lake Gardno, the youngest southern Baltic lagoon, emerged (Bogaczewicz-Adamczak, Miotk, 1985a; Zachowicz, Zaborowska, 1985). At that time the southern Baltic showed a whole system of coastal lagoons, from the Saaler Boden on the present German coast to the Kuronian Lagoon on the present Lithuanian shore. There is ample geological evidence that at the turn of the Subboreal and Subatlantic, the southern Baltic lagoons were larger than their area today, although the sea level was still somewhat lower than it is at present. Reconstruction of the Vistula Lagoon extent, based on geological evidence (e.g. Mojski, 1988) is supported by archaeological and hydrogeological data. The areas covered at that time by the Lagoon contain no Neolithic archaeological sites (Majewski, 1969), while the first groundwater level is of a young relict nature and shows an increased chloride content and electric conductivity (Nałęcz, 1999).

The Subatlantic, i.e. the last 2500 years of the southern Baltic coast evolution, was characterised by a slow approach of the shoreline to its present location (Fig. 50). The mean sea level was undergoing slight changes only, increasing by about 0.6–0.7 m at a mean rate of 0.25–0.30 mm/a. Coastal cliffs were still evolving, like they did at the beginning of the Subboreal. There were only slight changes involving activation or inactivation of various fragments of the cliff shores. The mean rate of cliff recession was still slow.

At the beginning of the Subatlantic, the barriers continue to grow and stabilise. The largest dunes were at that time formed on the Świna Barrier (Świna Gate), Łeba Barrier and Vistula Spit. The most intensive development of the so-called yellow dunes was ascribed to within 2.2–1.8 ka BP (Tomczak in: Mojski *et al*., eds., 1995). *C.* 1.8 ka BP, another dune-forming phase occurred on the Łeba Barrier, whereby the largest crescent and parabolic dunes were formed (Borówka, 1995). Other

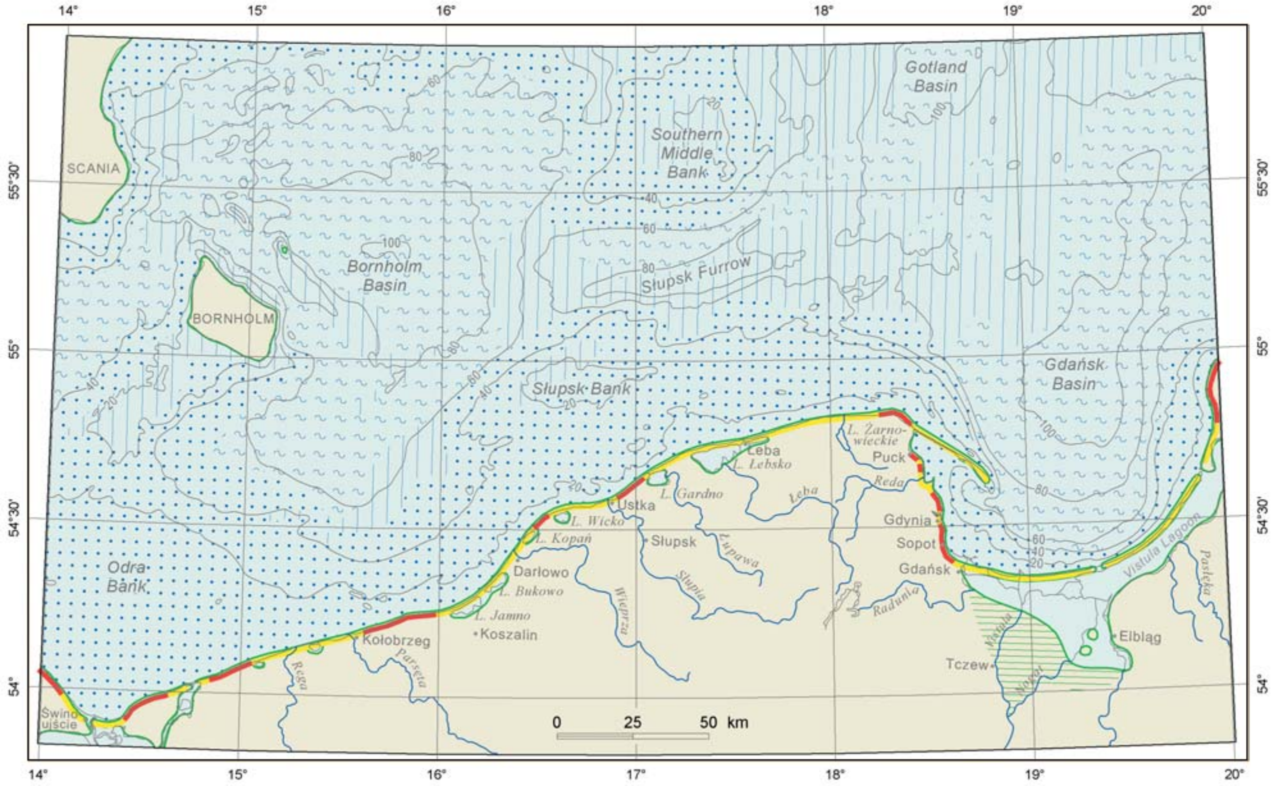


Fig. 50. The southern Baltic area during the Post-Littorina Sea (*c.* 2.5 ka BP) (after Uścinowicz in: Mojski *et al*., eds., 1995; modified)

For explanations see Figure 42

large dune fields on the southern Baltic coast emerged most probably at that time, including the one on the Lake Sarbsko barrier, west of Ustka (the so-called Modelskie dunes and dunes of the Lake Wicko barrier) and west of Kołobrzeg, on the barriers of lakes Resko and Liwia Łuża. Fossil soils on the Lake Sarbsko barrier were dated to 1.36 ±0.06 ka BP, those occurring in dunes between Lake Wicko and Ustka being dated at 1.03 ±0.05 and 1.05 ±0.05 ka BP (Tomczak in: Mojski *et al*., eds., 1995). The barriers grew not only on their landward side. Intensive sand deposition was taking place also on their submerged slopes. Shells of *Cerastoderma* sp. found on the Łeba Barrier forefront, at the sea depth of 10 m, dated to 1.43 ±0.05 ka BP were found at the depth of 7.5 m below the seafloor level (Appendix 1, core Madwiny 1; Fig. 7).

Different was the behaviour of the smaller, less developed barriers of lakes (lagoons) Kopań, Bukowo and Jamno. Most often, the absence of any source of sand in the area resulted in the presence, e.g. during the Subboreal, of a single bulge of relatively low fore dunes. They were frequently broken by storms, whereby washover fans were formed in the lagoons (lakes). As a resulted of those processes, at a slight sea level rise, the barriers ingressed onto lagoonal deposits. Similar processes of transverse barrier migration occurred from Władysławowo to Kuźnice on the Hel Peninsula. On the Puck Lagoon side, the Peninsula shore runs along and undulating line formed by deposition of storm washover fans (Fig. 51). The Peninsula continued to elongate in its southeastern part, the accretion rate rapidly, however, decreasing. Firstly, sands transported by longshore currents were deposited at progressively larger depths. Secondly, as the cliffs north-west of Władysławowo were eroded and the Peninsula base moved south, and its north-

Fig. 51. Hel Peninsula and Puck Lagoon: a satellite image

Landsat ETM+, May 2001, composite image of bands 1, 2, 3, resolution 30 m western part migrated south-west, the southeastern part of the Peninsula was not only elongated, but also broadened by accreting towards north-east. At a distance of about 31 km from the Peninsula base, in a depression between dunes, a peat deposit dated to 1.76 ±0.1 ka BP was revealed at the ordinate *c.* 0.6 m a.s.l. (Tomczak in: Mojski *et al.*, eds., 1995). This indicates how long the Peninsula was at that time: it could have been shorter by about 2–3 km than at present.

Certain larger changes in the shoreline outline were taking place in lagoonal areas, particularly within the Vistula Lagoon. At the beginning of the Subatlantic, the Vistula Lagoon was largest in its history (Fig. 50) extending beyond Elbląg down to Lake Drużno in the south and to Gdańsk in the west and covering almost all the present delta plains area lying below the sea level. Despite the slow sea level rise, the Lagoon was gradually shrinking in size, due to both accumulation of sedimentary material brought in by the Vistula the many branches of which discharged into the Lagoon, and because shallow inshore areas were being vegetated by aquatic plants. The Vistula Lagoon shrinkage was accelerated in the Middle Ages due to anthropogenic effects. Construction of flood dams and polders restricted deposition of the sedimentary material on the delta plain and increased its direct supply into the Lagoon and the Gulf of Gdańsk. During the Subatlantic, the Lagoon shoreline in the northern part of the present Vistula delta moved north by about 10–15 km. The process is particularly well evidenced by charts and maps from the last 300 years. In the eastern part of the Vistula delta, the Lagoon shoreline retreated at that time by about 5–8 km, both due to natural processes of riverine sediment accumulation, and because of polder construction (Majewski, 1969). Small rivers discharging into the Lagoon exerted a much weaker effect on the shoreline changes. In addition to the Vistula, relatively important was also the Pregoła discharging into the northeastern part of the Lagoon, in its present Russian part. In the Polish part, the Pasłęka delta increased in size on the southeastern coast.

Deltas were also formed by other rivers feeding the southern Baltic coastal lagoons. In the Subatlantic, deltas of the Reda, Łeba and Łupawa, opening into the Puck Lagoon, Lake Łebsko and Lake Gardno, respectively, increased in size. In addition to river-built deltas in lagoons and coastal lakes, storm deltas emerged in outlets of the channels connecting the lagoons with the sea. The largest storm delta formed in the Szczecin Lagoon, at the Świna discharge (Fig. 52). Development of the storm deltas demonstrates the significant role of storm surge-driven sea level fluctuations in shoreline evolution.

Changes in the size of the lagoons were also a result of overgrowing by vegetation and of peat formation. Those processes played a particularly important role on landward, shallow and gently sloping shores of the lagoons. The most extensive changes in the lagoon size, brought about by peat formation, took place on the Gardno–Łeba Lowland around Lakes Łebsko, Sarbsko and Gardno, and also on the central coast between lakes Bukowo and Jamno. The latter two lakes were, at the beginning of the Subatlantic, a single lagoon, as indicated by the presence of lagoonal deposits underneath a peat layer in the area in between the present lakes. Presumably lakes Łebsko and Sarbsko, too, were a single lagoon then.

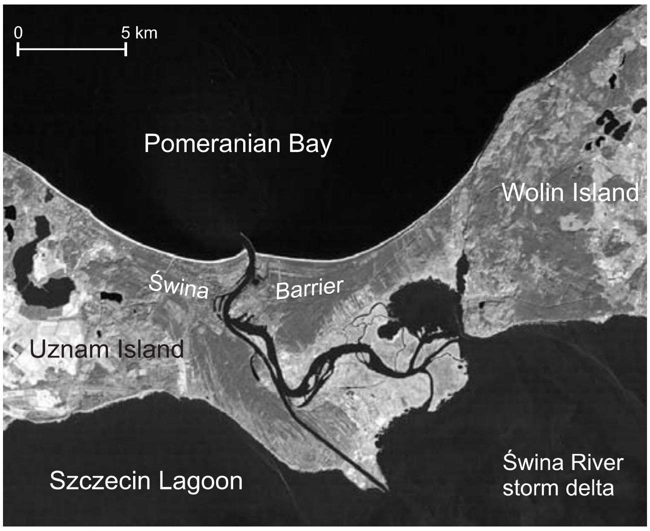


Fig. 52. Storm delta of River Świna in the Szczecin Lagoon: a satellite image

Landsat ETM+, 22 September 1999, panchromatic band, resolution 15 m)

In the 18th, and particularly in the 19th and 20th centuries, as a result of expanding port facilities and breakwaters being built in mouth areas of numerous rivers, i.a. in Świnoujście, Dziwnowo, Kołobrzeg, Darłowo, Ustka, Rowy, Łeba and Gdańsk, natural lithodynamic processes in the coastal zone and shoreline evolution were disturbed. In many places, most often on the western side of the breakwaters, zones of forced sediment accumulation emerged and the shore moved seawards. On the eastern side of the breakwaters, the shore was intensively eroded. Changes in natural evolutionary tendencies were taking place not only in the immediate vicinity of the man-made constructions, but also, to a smaller or larger degree, affected the entire southern Baltic coast.

At present, the southern Baltic coast seems to enter a new developmental stage. For about 50 years, the entire coast has been showing an accelerated sea level rise as well as an increasing frequency and force of storms (e.g. Dziadziuszko, Jednorał, 1996; Rotnicki, Borzyszkowska, 1999). Those phenomena bring about accelerated erosion of the shores, both cliffs and dunes, and have been treated in numerous publications. Important and open question is whether we are witnessing consequences of another natural climatic oscillation, similar to those known from the past, or it is a new process triggered by human activities.

SUMMING UP AND PROBLEMS THAT MERIT DISCUSSION

Geological evidence, primarily the 314 radiocarbon datings of terrestrial and marine sediments, backed up by numerous pollen-, diatom-, micro- and macrofaunal analyses as well as the analysis of seismic profiles and sediment cores have made it possible to reconstruct the history of relative sea level changes and to reconstruct vertical crustal movements and changes in the shoreline location in Late Pleistocene and Holocene in the southern Baltic.

The history of sea level changes in the southern Baltic may be divided into three stages. The first stage, in Late Pleistocene and Early Holocene, was characterised by large and rapid sea level changes, dependent on the differential glacio-isostatic movements in the southern and northern parts of the Baltic, the eustatic ocean level rise and opening and closing of the Baltic basin's connection with the ocean. Between 13.0–8.5 ka BP, the sea level changes within an amplitude as wide as 25–30 m. In some extreme cases, the sea level could have fallen at a rate of about 100–300 mm/a, the sea level rise rate reaching up to about 40–45 mm/a. In Late Pleistocene and Early Holocene, there were three transgressions: within 12.0–11.2 ka BP, 11.0–10.3 ka BP (the Baltic Ice Lake) and 10.2–9.2 ka BP (the Yoldia Sea and the Ancylus Lake); there were also three regressions, setting on 11.2, 10.3 and 9.2 ka BP.

The second stage, in Middle Holocene (the Mastogloia Sea and Littorina Sea), was characterised by a constant sea level rise, initially fast, up to *c.* 15 mm/a, and slower later on. This stage includes also the Late Boreal as the Baltic attained the permanent connection with the ocean about 8.5 ka BP. Within 8.5–5.0 ka BP, the sea level rise in the southern Baltic depended primarily on glacio-eustatic sea level changes. In the Late Boreal, until the onset of the Atlantic, the sea level rose from about 28 m to about 21 m b.s.l. Within 8.0–7.0 ka BP, the sea level rose to 10 m b.s.l. at a rate of about 11 mm/a. Until the end of the Atlantic, the sea level rose to 2.5 m b.s.l., the rate of rise slowing down, within about 6–5 ka BP, to about 2.5 mm/a.

The third, Late Holocene, stage showed a narrow range of mean sea level changes. In the fist Subboreal millennium, the sea level rose from about 2.5 to about 1.1–1.3 m below the present sea level, to be — at the end of the Subboreal — about 0.6–0.7 m lower than at present. During the Subatlantic, the mean sea level was changing only slightly. The slow rise of the mean sea level at the end of the Atlantic, particularly within the last 5000 years, was most probably overlain by fluctuations associated with climatic changes. The whole pattern could have been complicated also by neotectonic movements, not related to glacio-isostasy.

A comparison between the Southern Baltic relative sea level curves and the eustatic ocean level ones made it possible to reconstruct the glacio-isostatic rebound. The restrained rebound phase began *c.* 17.5 ka BP to last until *c.* 14.0 ka BP. The southern Baltic area was at that time uplifted by about 20 m. The uplift rate within 17–16 ka BP ranged from about 1.0 to about 3.0 mm/a. The basic post-glacial uplift proceeded from *c.* 14.0 to *c.* 11.0 ka BP, the southern Baltic area being uplifted by about 85 m. The uplift proceeded at the maximum rate of about 45 mm/a within 12.4–12.2 ka BP. The residual uplift began *c.* 11.0 ka BP to terminate *c.* 9.2–9.0 ka BP, the crust being uplifted by about 15 m. The southern Baltic uplift rate *c.* 10.0 ka BP slowed down to *c.* 5.5 mm/a and the uplift stopped *c.* 9.2–9.0 ka BP. The rapid termination of uplift within 11.0 to 9.0 ka BP was presumably due to a "brake" effect produced by hydro- and sedimento-isostasy. Forebulge migration across the southern Baltic area took place within *c.* 9.0–7.0 ka BP, a slight subsidence occurring within *c.* 7.0–4.0 ka BP. As of *c.* 4.0 ka BP, the Earth crust regained its equilibrium.

Evolution of the southern Baltic coast in Late Pleistocene and Holocene is tightly coupled with relative sea level changes. This is primarily a result of an interplay between glacio-isostatic movements and eustatic sea level changes. To some extent, the coastline evolution depended also on the geological setting as well as on sediment erosion and accumulation processes and on climate oscillations. The coastline evolution was divided into three stages as well: the Late Pleistocene–Early Holocene stage of several rapid and extensive shoreline displacements; the Middle Holocene stage, initially characterised by a rapid shoreline migration, the coast stabilising within 6–5 ka BP; and the Late Holocene stage, with a narrow range of shoreline displacement and a domination of coast levelling processes as a result of which the shoreline approached its present location.

In this work, models of sea level changes and glacio-isostatic rebound in the southern Baltic area have been presented and evolution of the shoreline has been discussed. As new evidence accumulates, many parts of the picture painted here will inevitably be refined, both spatially and temporally. The true history of sea level changes in Holocene, particularly in the Atlantic, Subboreal and Subatlantic was doubtless more complex than that presented by a smoothed relative sea level curve. In Late Holocene in particular, against the background of a slight and slow rise of the mean sea level, regional eustatic fluctuations associated with climatic changes could have substantially affected coastline evolution. Neotectonic movements were another factor, not associated with glacio-isostasy, that produced regional differences in relative sea level changes, and hence affected the shoreline evolution at that time.

Numerous and detailed studies on variability of oceanic circulation and climate in Holocene, analyses of Greenland and Antarctic ice cores, oceanic sediment cores and dendrochronologic reconstructions have revealed the presence of cyclic changes in climate and oceanic circulation. Particularly important for changes in the southern Baltic area climate and hydrology were variations in the atmospheric and oceanic circulation in North Atlantic. Cyclic changes in the North Atlantic oceanic circulation within about 1500 years were shown by, i.a. Bond *et al.* (1997) and Bianchi, McCave (1999). Chapman and Shackelton (2000) demonstrated cyclic variability in the North Atlantic Deep Water circulation within 1650, 1000 and 550 years and argued that the 1500-yr cycles reported by Bond *et al.* (1997) and Bianchi and McCave (1999) and the 1650-yr ones are statistically indistinguishable. Climatic cycles having periodicity of 1000 and 550 years, associated with variations in solar activity, are the best documented ones (e.g. Stuiver, Braziunas, 1993; Stuiver *et al*., 1995; Chapman, Shackelton, 2000). There is no doubt that variability in the atmospheric and oceanic circulation and the associated climate changes do produce changes in the mean sea level.

Research on periodicity of Baltic sea level changes dates back to 1937, i.e. to Iversen's (1937) publication on Littorina transgression in Denmark in which four transgression phases were distinguished within the Atlantic and Subboreal. The best documentation was provided by Berglund (1964, 1971) for Blekinge area, Digerfeldt (1975) for Scania and Mörner (1976, 1980a) for the Kattegat. All those studies demonstrated the presence of transgressive–regressive phases in the Atlantic and Subboreal. Periodicity of the transgressive–regressive phases identified in southern Sweden (Berglund, 1964, 1971; Digerfeldt, 1975; Mörner, 1976, 1980) was set at 1600–500 years and was in good agreement with recent findings from North Atlantic (e.g. Chapman, Shackelton, 2000). Differences in the number and timing of transgression phases were most probably caused by local specificity of the sites studied, supply of materials to be dated, and neotectonic effects.

Effects of regional eustatic oscillations in the southern Baltic coastal area are poorly known and still offer problems in need of solving. This is true of both the first half of Holocene, when the sea level rise rate was high, and the second half, with but small changes in the sea level. The main cause of difficulties in identification and documentation of periodic eustatic oscillations on the southern Baltic coast in the Preboreal, Boreal and Atlantic is a considerable extent and the fast rate of changes which masked effects of changes with relatively shorter periodicity and much smaller amplitudes. The main cause of difficulties in inferring changes taking place in the Subboreal and Subatlantic is the magnitude of barographic and storm surges, which — in extreme cases — were up to about 2 m high on the southern Baltic coast (Majewski, 1987). Sea level amplitudes associated with extreme phenomena are larger than those of eustatic oscillations. Erosion and sedimentation, proceeding during the extreme events, could alter the pattern of coastal zone formations and sediments more than oscillations of the mean sea level with periodicity of hundreds or thousands years. Because of that, reconstruction of the Holocene history of regional eustatic oscillations in the southern Baltic still remains an unsolved problem.

Another problem that remains to be solved is a question, associated with neotectonic movements, if and to what extent the movements resulted in differences in the history of sea level changes between different parts of the southern Baltic in Middle and Late Holocene. In the past, as at present, neotectonic movements varied along the southern Baltic coast. At the present stage it is, however, impossible to solve this problem by a backward projection of the present trends. The history of neotectonic movements in the southern Baltic area once glacio-isostasy had ceased was doubtless much more variable and complex.

The problems outlined above have not been tackled in this work. Solving them requires specialised and detailed studies. This work presents smoothed models of sea level changes and glacio-isostatic rebound of the southern Baltic area, made possible by the existing set of data, primarily the collection of radiocarbon datings. The data set analysed showed a high temporal concordance in timing of certain events taking place throughout the southern Baltic coast. No discrepancies in or disagreements between data from various parts of the coast were found. This does not mean that there was no regional variability in eustatic oscillations and/or neotectonic movements. The reason they were not detected may be attributed to their scale, smaller than amplitudes of both seasonal changes in the mean se level and, most of all, the extent of extreme phenomena such as storm surges and wave action. Natural dynamics of coastal processes as well as the research methods and scope did not make it possible to unambiguously identify effects of both neotectonic and eustatic oscillation phases on the southern Baltic shoreline evolution. Nevertheless, despite all the difficulties, the problems outlined above should be targeted by further, comprehensive research. This research is particularly important for forecasting the future evolution of the coast in the light of climate changes and accelerated sea level rise, observed at present. It is necessary to gain knowledge on the frequency and amplitudes of changes occurring naturally and to determine the extent of the greenhouse effect on the climate and sea level rise.

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Appendix 1

Radiocarbon datings of samples

No.	Site/core	Coordinates		Altitude (m a.s.l.)	Sample position in core (m)	Sample position (m b.s.l.)	^{14}C age BP	Lab. code	Material dated	Supporting analyses	References	Locality
		ϕ	λ									
1	Piaski A 1	54°25.20’	19°33.50’	1.50	14.00–14.20	12.50–12.60	7710 ±170	Gd-2685	lagoonal mud		Tomczak *et al.*, 1989*	Vistula Spit
2	Piaski 2	54°26.34’	19°36.31’	0.50	12.40–12.45	11.90–11.95	7620 ±290	Gd-15319	lagoonal mud		MGB PGI Archives	Vistula Spit
3	Piaski 2a	54°24.45’	19°30.98’	0.50	14.00–14.04	13.50–13.55	8170 ±230	Gd-15318	lagoonal mud		MGB PGI Archives	Vistula Spit
4	Piaski 2a	54°24.45’	19°30.98’	0.50	14.50–14.53	14.00–14.03	7560 ±90	Gd-12372	peat, plant remains		MGB PGI Archives	Vistula Spit
5	Dziady A 1	54°24.65’	19°31.50’	1.50	15.10–15.20	13.60–13.70	8120 ±140	Gd-4145	peat		Tomczak *et al.*, 1989*	Vistula Spit
6	Dziady 2	54°24.51’	19°32.08’	0.50	8.90–9.10	8.40–8.60	2920 ±60	Gd-12369	*Macoma* and *Cerastoderma* shells		MGB PGI Archives	Vistula Spit
7	Dziady 2	54°24.51’	19°32.08’	0.50	13.42–13.47	12.92–12.97	8200 ±270	Gd-15302	lagoonal mud		MGB PGI Archives	Vistula Spit
8	Dziady 2	54°24.51’	19°32.08’	0.50	14.05–14.10	13.55–13.60	8040 ±120	Gd-15317	peat		MGB PGI Archives	Vistula Spit
9	Dziady 3	54°24.58’	19°32.86’	1.30	2.30–2.40	1.00–1.10	1210 ±80	Gd-15313	lagoonal mud		MGB PGI Archives	Vistula Spit
10	Dziady 3	54°24.58’	19°32.86’	1.30	10.00–10.10	8.70–8.80	6820 ±100	Gds-298	peat		MGB PGI Archives	Vistula Spit
11	Dziady 3	54°24.58’	19°32.86’	1.30	27.30–27.40	26.00–26.10	9330 ±140	Gd-15304	peat		MGB PGI Archives	Vistula Spit
12	Krynica Morska I K-1	54°22.83’	19°25.90’	9.00	24.50–25.00	15.50–16.00	7970 ±90	Gd-5141	lagoonal mud, plant remains		Tomczak *et al.*, 1989*	Vistula Spit
13	Krynica Morska I K-5	54°22.80’	19°26.00’	1.00	16.00–16.50	15.00–15.50	7590 ±70	Gd-5154	peaty mud		Tomczak *et al.*, 1989*	Vistula Spit
14	Krynica Morska 2	54°23.15’	19°26.35’	0.50	11.00–11.25	10.50–10.75	2760 ±60	Gd-12359	*Cerastoderma* shells		MGB PGI Archives	Vistula Spit
15	Krynica Morska 3	54°22.56’	19°26.74’	0.90	15.50–15.52	14.60–14.62	7960 ±160	Gd-15310	peat		MGB PGI Archives	Vistula Spit
16	Przebrno E-5	54°21.70’	19°20.30’	2.00	17.00–18.00	15.00–16.00	6680 ±130	Gd-2696	lagoonal mud	malacofauna	Tomczak *et al.*, 1989*	Vistula Spit
17	Przebrno E-7	54°21.40’	19°20.30’	1.00	14.20–15.20	13.20–14.20	7060 ±140	Gd-2684	lagoonal mud		Tomczak *et al.*, 1989*	Vistula Spit
18	Przebrno 1	54°22.47’	19°19.82’	–10.00	7.00–8.00	17.00–18.00	4030 ±70	Gd-15300	*Macoma* and *Cerastoderma* shells		MGB PGI Archives	Vistula Spit
19	Sztutowo 1	54°21.64’	19°11.08’	–10.00	6.50–6.65	16.50–16.65	3380 ±50	Gd-12367	*Macoma* and *Cerastoderma* shells		MGB PGI Archives	Vistula Spit
20	Jantar E-13	54°19.60’	18°59.90’	0.70	9.50–10.00	8.80–9.30	6330 ±60	Gd-1408	peat		Tomczak *et al.*, 1989*	Vistula Spit
21	Komary K-2	54°20.10’	18°54.30’	2.00	11.30–11.50	9.30–9.50	6740 ±80	Gd-1407	peaty mud		Tomczak *et al.*, 1989*	Vistula Spit
22	Komary H-3	54°20.30’	18°54.10’	2.00	11.40–12.00	9.40–10.00	6900 ±70	Gd-1399	peaty mud		Tomczak *et al.*, 1989*	Vistula Spit
23	Komary H-5	54°20.10’	18°54.10’	2.00	10.50–11.00	8.50–9.00	7260 ±90	Gd-1419	peaty mud		Tomczak *et al.*, 1989*	Vistula Spit
24	Komary H-7	54°19.90’	18°54.20’	1.30	9.80–10.50	8.50–9.20	7340 ±70	Gd-1398	peaty mud		Tomczak *et al.*, 1989*	Vistula Spit
25	Komary F-1	54°20.60’	18°53.90’	2.00	9.80–10.50	7.80–8.50	7580 ±95	Gd-1404	peaty sand		Tomczak *et al.*, 1989*	Vistula Spit
26	Komary F-3	54°20.60’	18°53.80’	2.00	11.20–11.50	9.20–9.50	8020 ±130	Gd-1405	peaty mud		Tomczak *et al.*, 1989*	Vistula Spit
27	Komary F-16	54°19.20’	18°53.30’	1.50	9.50–10.50	8.00–9.00	6500 ±65	Gd-1409	peaty sand		Tomczak *et al.*, 1989*	Vistula Spit
28	Komary D-5	54°20.40’	18°53.50’	2.00	10.00–10.50	8.00–8.50	7390 ±90	Gd-1418	peaty mud		Tomczak *et al.*, 1989*	Vistula Spit
29	IIIa	54°23.40’	19°41.70’	–1.00	2.50–2.75	3.50–3.75	6735 ±65	Gd-1238	peat	pollen and diatoms	Bogaczewicz-Adamczak, Miotk, 1985b	Vistula lagoon
30	IIIa	54°23.40’	19°41.70’	–1.00	3.00–3.27	4.00–4.27	7120 ±100	Gd-1237	peat	pollen and diatoms	Bogaczewicz-Adamczak, Miotk, 1985b	Vistula lagoon
31	IIIa	54°23.40’	19°41.70’	–1.00	3.90–4.15	4.90–5.15	7600 ±95	Gd-1239	peat	pollen and diatoms	Bogaczewicz-Adamczak, Miotk, 1985b	Vistula lagoon
32	IIIa	54°23.40’	19°41.70’	–1.00	4.30–4.50	5.30–5.50	9390 ±110	Gd-1240	peaty mud	pollen and diatoms	Bogaczewicz-Adamczak, Miotk, 1985b	Vistula lagoon
33	IIIa	54°23.40’	19°41.70’	–1.00	4.50–4.70	5.50–5.70	11,090 ±150	Gd-772	peaty mud	pollen and diatoms	Bogaczewicz-Adamczak, Miotk, 1985b	Vistula lagoon
34	IIIa	54°23.40’	19°41.70’	–1.00	4.70–4.95	5.70–5.95	11,240 ±110	Gd-773	limnic mud	pollen and diatoms	Bogaczewicz-Adamczak, Miotk, 1985b	Vistula lagoon
35	ZW 3	54°20.80’	19°16.70’	–1.80	12.25–12.35	14.05–14.15	10,200 ±300	Gd-9444	limnic mud	diatoms	Zachowicz, Uścinowicz, 1997*	Vistula lagoon
36	ZW 3	54°20.80’	19°16.70’	–1.80	14.25–14.35	16.05–16.15	11,620 ±390	Gd-9446	peat	diatoms	Zachowicz, Uścinowicz 1997*	Vistula lagoon
37	ZW 3	54°20.80’	19°16.70’	–1.80	14.70–14.85	16.50–16.65	12,270 ±200	Gd-10246	peat	diatoms	Zachowicz, Uścinowicz, 1997*	Vistula lagoon
38	ZW 6	54°19.50’	19°25.80’	–2.40	2.65–2.75	5.05–5.15	2580 ±130	Gd-9450	lagoonal mud	pollen and diatoms	Zachowicz, Uścinowicz, 1997*	Vistula lagoon
39	ZW 6	54°19.50’	19°25.80’	–2.40	2.75–2.85	5.15–5.25	2010 ±140	Gd-9451	lagoonal mud	pollen and diatoms	Zachowicz, Uścinowicz, 1997*	Vistula lagoon
40	ZW 6	54°19.50’	19°25.80’	–2.40	5.45–5.55	7.85–7.95	5180 ±150	Gd-9453	lagoonal mud	pollen and diatoms	Zachowicz, Uścinowicz, 1997*	Vistula lagoon
41	ZW 7	54°21.10’	19°26.90’	–1.80	13.87–13.93	15.67–15.73	7820 ±120	Gd-10245	peat		Zachowicz, Uścinowicz 1997*	Vistula lagoon
42	ZW 10	54°19.70’	19°31.30’	–1.80	4.03–4.11	5.83–5.91	7240 ±80	Gd-11187	peat		Zachowicz, Uścinowicz, 1997*	Vistula lagoon
43	ZW 10	54°19.70’	19°31.30’	–1.80	4.37–4.48	6.17–6.28	5950 ±180	Gd-10248	peaty mud		Zachowicz, Uścinowicz, 1997*	Vistula lagoon
44	ZW 10	54°19.70’	19°31.30’	–1.80	4.90–5.00	6.70–6.80	8590 ±120	Gd-10241	peaty mud		Zachowicz, Uścinowicz, 1997*	Vistula lagoon
45	ZW 10	54°19.70’	19°31.30’	–1.80	5.20–5.35	7.00–7.15	9360 ±90	Gd-11185	peaty mud		Zachowicz, Uścinowicz, 1997*	Vistula lagoon

Appendix 1 continued

No.	Site/core	Coordinates		Altitude (m a.s.l.)	Sample position in core (m)	Sample position (m b.s.l.)	^{14}C age BP	Lab. code	Material dated	Supporting analyses	References	Locality
		ϕ	λ									
46	ZW 14	54°21.10'	19°40.20'	–2.50	3.60–3.69	6.10–6.19	6200 ±120	Gd-10244	peat		Zachowicz, Uścinowicz, 1997*	Vistula lagoon
47	ZW 14	54°21.10'	19°40.20'	–2.50	5.00–5.08	7.50–7.58	8080 ±80	Gd-11186	peat		Zachowicz, Uścinowicz, 1997*	Vistula lagoon
48	ZW 15	54°26.40'	19°43.30'	–4.30	2.76–2.90	7.06–7.20	9700 ±210	Gd-10249	limnic sand, plant detritus		Zachowicz, Uścinowicz, 1997*	Vistula lagoon
49	32 T	54°19.60'	19°31.70'	1.50	8.00–8.12	6.50–6.62	8910 ±90	Gd-10301	peat		Zachowicz, Uścinowicz, 1997*	Vistula lagoon
50	J. Druzno 1a	54°6.20'	19°28.60'	–0.80	6.25–6.35	7.05–7.15	6440 ±50	Gd-1131	lagoonal mud	pollen and diatoms	Zachowicz *et al*., 1982	Vistula lagoon
51	J. Druzno 1a	54°6.20'	19°28.60'	–0.80	6.79–6.88	7.59–7.68	7050 ±70	Gd-1132	peat	pollen and diatoms	Zachowicz *et al*., 1982	Vistula lagoon
52	J. Druzno 1a	54°6.20'	19°28.60'	–0.80	7.12–7.22	7.92–8.02	8995 ±75	Gd-1136	peat	pollen and diatoms	Zachowicz *et al*., 1982	Vistula lagoon
53	J. Druzno 1a	54°6.20'	19°28.60'	–0.80	8.58–8.66	9.38–9.46	11,290 ±105	Gd-1137	peat	pollen and diatoms	Zachowicz *et al*., 1982	Vistula lagoon
54	2 EL 96	54°45.22'	19°11.31'	–101.50	1.80–1.90	103.30–103.40	3040 ±120	Gd-6574	marine mud	pollen and diatoms	Uścinowicz, Zachowicz, 1993c*	Gulf of Gdańsk
55	2 EL 96	54°45.22'	19°11.31'	–101.50	3.80–3.90	105.30–105.40	6870 ±130	Gd-6308	marine mud	pollen and diatoms	Uścinowicz, Zachowicz, 1993c*	Gulf of Gdańsk
56	2 EL 96	54°45.22'	19°11.31'	–101.50	4.10–4.32	105.60–105.82	7590 ±140	Gd-6575	marine mud	pollen and diatoms	Uścinowicz, Zachowicz, 1993c*	Gulf of Gdańsk
57	2 EL 96	54°45.22'	19°11.31'	–101.50	4.52–4.66	106.02–106.16	8130 ±160	Gd-6310	marine mud	pollen and diatoms	Uścinowicz, Zachowicz, 1993c*	Gulf of Gdańsk
58	1 EL 32	54°38.47'	19°12.19'	–85.50	1.50–1.70	87.00–87.20	5450 ±120	Gd-6307	marine mud	diatoms	Uścinowicz, Zachowicz, 1993a*	Gulf of Gdańsk
59	1 EL 38	54°38.16'	19°12.54'	–88.00	0.60–0.80	88.60–88.80	1640 ±90	Gd-4778	marine mud	pollen and diatoms	Uścinowicz, Zachowicz, 1993a*	Gulf of Gdańsk
60	1 EL 38	54°38.16'	19°12.54'	–88.00	2.60–2.80	90.60–90.80	4250 ±130	Gd-6576	marine mud	pollen and diatoms	Uścinowicz, Zachowicz, 1993a*	Gulf of Gdańsk
61	1 EL 38	54°38.16'	19°12.54'	–88.00	4.60–4.80	92.60–92.80	6020 ±120	Gd-4775	marine mud	pollen and diatoms	Uścinowicz, Zachowicz, 1993a*	Gulf of Gdańsk
62	1 EL 38	54°38.16'	19°12.54'	–88.00	5.60–5.80	93.60–93.80	8750 ±170	Gd-6313	marine mud	pollen and diatoms	Uścinowicz, Zachowicz, 1993a*	Gulf of Gdańsk
63	EL 1	54°25.86'	19°19.68'	–52.50	0.40–0.70	52.90–53.20	9000 ±260	Gd-4632	limnic (Anc. L.) clayey silt	ostracods	Uścinowicz, Zachowicz, 1993a*	Gulf of Gdańsk
64	EL 1	54°25.86'	19°19.68'	–52.50	5.80–6.00	58.30–58.50	10,650 ±160	Gd-4833	limnic (BIL) sandy silt	ostracods	Uścinowicz, Zachowicz, 1993a*	Gulf of Gdańsk
65	Stogi 1s	54°2475'	18°45.22'	–13.40	1.60–1.70	15.00–15.10	5620 ±90	Gds-245	deltaic mud		MGB PGI Archives	Gulf of Gdańsk
66	Brzeźno 2	54°24.99'	18°37.20'	0.50	7.80–8.00	7.30–7.50	3930 ±120	Gd-15333	deltaic mud		MGB PGI Archives	Gulf of Gdańsk
67	1 ZG 54	54°31.00'	18°52.52'	–67.50	4.90–5.00	72.40–72.50	12,200 ±240	Gd-4634	prodelta sandy mud	malacofauna and ostracods	Uścinowicz, Zachowicz, 1992b*	Gulf of Gdańsk
68	ZG 1	54°30.73'	18°48.02'	–31.00	1.20–1.30	32.20–32.30	9220 ±140	Gd-4777	deltaic mud	malacofauna	Uścinowicz, Zachowicz, 1994*	Gulf of Gdańsk
69	9/93	54°33.08'	18°45.93'	–54.00	0.64–0.74	54.64–54.74	380 ±180	Gd-9318	marine mud		MGB PGI Archives	Gulf of Gdańsk
70	9/93	54°33.08'	18°45.93'	–54.00	2.20–2.30	56.20–56.30	1610 ±260	Gd-9310	marine mud		MGB PGI Archives	Gulf of Gdańsk
71	9/93	54°33.08'	18°45.93'	–54.00	2.71–2.90	56.70–56.90	9460 ±280	Gd-9326	limnic (Anc. L.) muddy sand		MGB PGI Archives	Gulf of Gdańsk
72	12/93	54°21.96'	18°59.99'	–14.50	0.32–0.42	14.82–14.92	1440 ±130	Gd-9315	prodelta mud		MGB PGI Archives	Gulf of Gdańsk
73	12/93	54°21.96'	18°59.99'	–14.50	2.45–2.55	16.95–17.05	1850 ±170	Gd-9330	prodelta mud		MGB PGI Archives	Gulf of Gdańsk
74	23/93	54°27.27'	19°36.31'	–16.00	2.40–2.53	18.40–18.53	800 ±110	Gd-10127	*Macoma* and *Cerastoderma* shells		MGB PGI Archives	Gulf of Gdańsk
75	VIII/92	54°42.70'	18°34.20'	–1.00	13.50–13.80	14.50–14.80	9540 ±80	Gd-7271	limnic sand, plant detritus		Kramarska *et al*., 1995*	Puck Lagoon
76	VIII/92	54°42.70'	18°34.20'	–1.00	13.80–14.00	14.80–15.00	10,230 ±80	Gd-7272	peat		Kramarska *et al*., 1995*	Puck Lagoon
77	XIII/92	54°46.30'	18°27.90'	–0.90	2.80–3.00	3.70–3.90	6460 ±100	Gd-4958	lagoonal mud	malacofauna	Kramarska *et al*., 1995*	Puck Lagoon
78	34/92	54°43.80'	18°31.10'	–2.00	1.12–1.18	3.12–3.18	5850 ±120	Gd-6804	peat		Kramarska *et al*., 1995*	Puck Lagoon
79	34/92	54°43.80'	18°31.10'	–2.00	1.30–1.40	3.30–3.40	6070 ±20	Gd-6802	peat		Kramarska *et al*., 1995*	Puck Lagoon
80	34/92	54°43.80'	18°31.10'	–2.00	1.76–1.86	3.76–3.86	7910 ±140	Gd-6803	peat		Kramarska *et al*., 1995*	Puck Lagoon
81	SW 1	54°45.00'	18°26.00'	–3.80	0.00–0.05	3.80–3.85	6160 ±220	Gd-9544	peat		MGB PGI Archives	Puck Lagoon
82	W 2	54°39.25'	18°31.60'	–5.00	0.00–0.05	5.00–5.05	8320 ±80	Gd-11244	peat		MGB PGI Archives	Puck Lagoon
83	W 2	54°39.25'	18°31.60'	–5.00	1.90–2.00	6.90–7.00	11,920 ±80	Gd-7744	limnic gyttja	malacofauna and ostracods	MGB PGI Archives	Puck Lagoon
84	W 3	54°39.92'	18°31.70'	–3.70	2.00–2.05	5.70–5.75	9180 ±70	Gd-7759	peat		MGB PGI Archives	Puck Lagoon
85	W 3	54°39.92'	18°31.70'	–3.70	2.95–3.00	6.65–6.70	9440 ±70	Gd-7760	peat		MGB PGI Archives	Puck Lagoon
86	KUZ 2	54°42.26'	18°32.30'	–1.70	5.00–5.05	6.70–6.75	8040 ±130	Gd-10409	peat		MGB PGI Archives	Puck Lagoon
87	KUZ 2	54°42.26'	18°32.30'	–1.70	5.95–6.00	7.65–7.70	8360 ±160	Gd-10401	peat		MGB PGI Archives	Puck Lagoon
88	ZP15	54°44.00'	18°25.50'	–3.00	0.15–0.35	3.15–3.35	8860 ±150	Gd-4839	limnic gyttja	diatoms	Witkowski, Witak, 1993	Puck Lagoon
89	ZP18	54°41.20'	18°30.10'	–5.00	0.15–0.25	5.15–5.25	5480 ±130	Gd-4831	lagoonal mud	diatoms	Witkowski, Witak, 1993	Puck Lagoon
90	ZP18A	54°40.20'	18°30.10'	–4.00	0.45–0.67	4.45–4.67	11,300 ±160	Gd-4838	limnic limy gyttja	diatoms	Witkowski, Witak, 1993	Puck Lagoon
91	ZP19	54°39.10'	18°29.70'	–4.00	0.15–0.25	4.15–4.25	2890 ±120	Gd-4830	lagoonal mud	diatoms	Witkowski, Witak, 1993	Puck Lagoon
92	M1	54°36.80'	18°30.7'	0.80	6.45–6.55	5.65–5.75	9200 ±440	Gd-13016	peat		MGB PGI Archives	Puck Lagoon

No.	Site/core	Coordinates		Altitude (m a.s.l.)	Sample position in core (m)	Sample position (m b.s.l.)	^{14}C age BP	Lab. code	Material dated	Supporting analyses	References	Locality
		φ	λ									
93	Reda 2A	54°37.50’	18°28.3’	1.00	0.20–0.25	(–0.8)–(–0.75)	2850 ±80	Gd-5027	peat		Tomczak, 1995b*	Puck Lagoon
94	Reda 2B	54°37.50’	18°28.3’	1.00	1.75–1.80	0.75–0.8	10,010 ±110	Gd-5029	peat		Tomczak, 1995b*	Puck Lagoon
95	Beka 2/2	54°39.23’	18°28.4’	–0.38	0.02–0.07	0.40–0.45	1225 ±100	Gd-15250	peat	pollen and diatoms	MGB PGI Archives	Puck Lagoon
96	Beka 2/2	54°39.23’	18°28.4’	–0.38	0.64–0.68	1.02–1.06	3335 ±155	Gd-17064	peat	pollen and diatoms	MGB PGI Archives	Puck Lagoon
97	Gizdepka 2/3	54°39.85’	18°28.1’	–0.45	0.26–0.31	0.71–0.76	3840 ±185	Gd-15261	peat	pollen and diatoms	MGB PGI Archives	Puck Lagoon
98	Gizdepka 2/3	54°39.85’	18°28.1’	–0.45	0.40–0.45	0.85–0.90	5290 ±95	Gd-1522	peat	pollen and diatoms	MGB PGI Archives	Puck Lagoon
99	Gizdepka 4/2	54°41.68’	18°27.9’	0.05	0.11–0.16	0.06–0.11	235 ±100	Gd-15255	peat	pollen and diatoms	MGB PGI Archives	Puck Lagoon
100	Gizdepka 4/2	54°41.68’	18°27.9’	0.05	0.26–0.31	0.21–0.26	655 ±115	Gd-17073	peat	pollen and diatoms	MGB PGI Archives	Puck Lagoon
101	R 1	54°41.57’	18°28.50’	0.50	2.22–2.35	1.72–1.85	5520 ±70	Gd-7698	peat	pollen	Uścinowicz, Miotk-Szpiganowicz, 2001	Puck Lagoon
102	R6/96	54°41.60’	18°28.50’	0.80	1.70–1.80	0.90–1.00	2390 ±130	Gd-10836	lagoonal sand, plant remains	malacofauna	Uścinowicz, Miotk-Szpiganowicz, 2001	Puck Lagoon
103	R6/96	54°41.60’	18°28.50’	0.80	2.90–2.95	2.10–2.15	5770 ±170	Gd-14023	peat		Uścinowicz, Miotk-Szpiganowicz, 2001	Puck Lagoon
104	Rzucewo 2/1 (Rz 2/1)	54°41.84’	18°28.05’	0.66	1.67–1.72	1.01–1.06	5620 ±100	Gd-15231	peat		Uścinowicz, Miotk-Szpiganowicz, 2001	Puck Lagoon
105	Rzucewo 3/1 (Rz 3/1)	54°41.82’	18°28.05’	1.18	1.42–1.52	0.24–0.34	3435 ±30	GdA-169	*Cerastoderma* shells	malacofauna	Uścinowicz, Miotk-Szpiganowicz, 2001	Puck Lagoon
106	Rzucewo 3/1 (Rz 3/1)	54°41.82’	18°28.05’	1.18	1.53–1.58	0.35–0.40	3560 ±35	GdA-171	*Cerastoderma* shells	malacofauna	Uścinowicz, Miotk-Szpiganowicz, 2001	Puck Lagoon
107	Rzucewo 3/1 (Rz 3/1)	54°41.82’	18°28.05’	1.18	1.53–1.58	0.35–0.40	5830 ±45	GdA-159	peat		Uścinowicz, Miotk-Szpiganowicz, 2001	Puck Lagoon
108	Rzucewo 3/1 (Rz 3/1)	54°41.82’	18°28.05’	1.18	1.72–1.77	0.54–0.59	5780 ±120	Gd-15209	peat		Uścinowicz, Miotk-Szpiganowicz, 2001	Puck Lagoon
109	45/95	54°43.80’	18°23.80’	0.50	0.50–0.55	0.00–0.05	3950 ±110	Gd-11432	peat		MGB PGI Archives	Puck Lagoon
110	Płutnica 1/1	54°44.30’	18°23.82’	0.18	0.11–0.27	(–0.07)–(–0.09)	720 ±25	GdA-166	*Cerastoderma* shells	pollen and diatoms	MGB PGI Archives	Puck Lagoon
111	Płutnica 2/2	54°44.30’	18°23.84’	–0.25	0.12–0.20	0.37–0.45	835 ±25	GdA-167	*Cerastoderma* shells	pollen and diatoms	MGB PGI Archives	Puck Lagoon
112	Płutnica 2/2	54°44.30’	18°23.84’	–0.25	0.12–0.20	0.37–0.45	945 ±25	GdA-168	*Hydrobia* shells	pollen and diatoms	MGB PGI Archives	Puck Lagoon
113	P 1	54°43.13’	18°25.40’	0.70	2.64–2.80	1.94–2.10	10,150 ±110	Gd-10300	limnic sandy mud	malacofauna	MGB PGI Archives	Puck Lagoon
114	Swarzewo 1	54°46.20’	18°25.00’	0.20	0.35–0.38	0.15–0.18	2030 ±90	Gd-2735	peat		Tomczak, 1995b*	Puck Lagoon
115	Swarzewo 2	54°46.20’	18°25.00’	0.20	0.70–0.75	0.50–0.55	5350 ±70	Gd-5203	peat		Tomczak, 1995b*	Puck Lagoon
116	Swarzewo 3	54°46.20’	18°25.00’	0.20	1.10–1.15	0.90–0.95	6940 ±120	Gd-4167	peat		Tomczak, 1995b*	Puck Lagoon
117	Władysławowo 1/2	54°47.50’	18°25.50’	0.50	1.70–2.00	1.20–1.50	7640 ±90	Gd-5599	peat		Tomczak, 1995b*	Hel Peninsula
118	H 2	54°47.42’	18°27.58’	0.50	3.55–3.65	3.05–3.15	4220 ±160	Gd-10405	lagoonal sandy mud		MGB PGI Archives	Hel Peninsula
119	Chałupy 92	54°45.30’	18°31.70’	1.00	7.00–7.50	6.00–6.50	6850 ±150	Gd-9012	lagoonal muddy sand	diatoms	Tomczak, 1995b*	Hel Peninsula
120	Chałupy 92	54°45.30’	18°31.70’	1.00	9.50–9.60	8.50–8.60	7100 ±150	Gd-9011	peaty mud		Tomczak, 1995b*	Hel Peninsula
121	Chałupy V/93	54°45.27’	18°31.75’	1.00	6.37–6.43	5.37–5.43	4100 ±140	Gd-9362	lagoonal sandy mud		MGB PGI Archives	Hel Peninsula
122	Chałupy V/93	54°45.27’	18°31.75’	1.00	7.00–7.05	6.00–6.05	8800 ±80	Gd-7578	limnic sand, plant remains		MGB PGI Archives	Hel Peninsula
123	Chałupy R 9	54°45.79’	18°32.76’	–13.70	0.55–0.60	14.25–14.30	6700 ±170	Gd-10839	lagoonal mud		MGB PGI Archives	Hel Peninsula
124	Chałupy R 9	54°45.79’	18°32.76’	–13.70	0.83–0.88	14.53–14.58	8610 ±130	Gd-11424	peat		MGB PGI Archives	Hel Peninsula
125	Kuźnica X 2	54°44.13’	18°36.60’	–9.80	1.90–2.00	11.70–11.80	8290 ±290	Gd-9542	lagoonal mud	malacofauna	MGB PGI Archives	Hel Peninsula
126	Kuźnica 92/1	54°43.80’	18°36.30’	1.00	9.00–9.20	8.00–8.20	6640 ±120	Gd-6874	lagoonal mud	diatoms	Tomczak, 1995b*	Hel Peninsula
127	Kuźnica 92/2	54°43.80’	18°36.30’	1.00	11.80–11.90	10.80–10.90	6930 ±120	Gd-9009	lagoonal mud	diatoms	MGB PGI Archives	Hel Peninsula
128	Kuźnica 92/3	54°43.80’	18°36.30’	1.00	11.90–12.00	10.90–11.00	7240 ±130	Gd-6875	lagoonal mud	diatoms	Tomczak, 1995b*	Hel Peninsula
129	Kuźnica 92/5	54°43.80’	18°36.30’	1.00	32.70–33.00	31.70–32.00	8980 ±140	Gd-9015	limnic muddy sand		MGB PGI Archives	Hel Peninsula
130	Kuźnica IV/93	54°43.75’	18°36.13’	2.50	12.05–12.15	9.55–9.65	5720 ±170	Gd-9366	lagoonal mud	malacofauna	MGB PGI Archives	Hel Peninsula
131	Kuźnica K/96	54°44.30’	18°34.01’	1.00	11.15–11.20	10.15–10.20	1940 ±370	Gd-14006	lagoonal mud	malacofauna	MGB PGI Archives	Hel Peninsula
132	Kuźnica K/96	54°44.30’	18°34.01’	1.00	17.70–17.75	16.70–16.75	10,250 ±150	Gd-10822	peat		MGB PGI Archives	Hel Peninsula
133	4/95	54°47.81’	18°34.55’	–20.00	1.30–1.35	21.30–21.35	9200 ±260	Gd-14010	peat		MGB PGI Archives	Hel Peninsula
134	X/92	54°42.50’	18°36.60’	–9.80	10.30–10.80	20.10–20.60	5570 ±120	Gd-6807	marine mud		Kramarska *et al.*, 1995*	Hel Peninsula
135	X/92	54°42.50’	18°36.60’	–9.80	12.10–12.50	21.90–22.30	6660 ±120	Gd-4951	marine mud		Kramarska *et al.*, 1995*	Hel Peninsula
136	XII/92	54°43.50’	18°36.20’	–1.20	9.60–9.90	10.80–11.10	3170 ±140	Gd-4955	marine mud		Kramarska *et al.*, 1995*	Hel Peninsula
137	XII/92	54°43.50’	18°36.20’	–1.20	11.00–11.30	12.20–12.50	7720 ±70	Gd-7273	marine organic sand		Kramarska *et al.*, 1995*	Hel Peninsula
138	Jastarnia 10 S/1	54°41.50’	18°42.20’	1.50	1.45–1.70	(–0.05)–0.20	1030 ±70	Gd-11053	peat		Tomczak, 1990*	Hel Peninsula
139	Jastarnia 10 S/2	54°41.50’	18°42.20’	1.50	1.80–1.90	0.30–0.40	2090 ±50	Gd-11055	peat		Tomczak, 1990*	Hel Peninsula

Appendix 1 continued

No.	Site/core	Coordinates		Altitude (m a.s.l.)	Sample position in core (m)	Sample position (m b.s.l.)	^{14}C age BP	Lab. code	Material dated	Supporting analyses	References	Locality
		ϕ	λ									
140	Jastarnia	54°41.90'	18°41.40'	1.50	4.00–4.10	2.50–2.60	5370 ±95	Gd-1027	peat		Bogaczewicz-Adamczak, 1982	Hel Peninsula
141	Jurata 9 S	54°40.90'	18°42.70'	1.30	1.20–1.35	0.10–0.05	3150 ±100	Gd-10178	peat		Tomczak, 1990*	Hel Peninsula
142	Jurata	54°40.70'	18°44.50'	0.20	1.00–1.20	0.80–1.00	5640 ±120	Gd-4058	peaty soil		Tomczak, 1995b*	Hel Peninsula
143	II/93	54°40.20'	18°42.50'	–2.00	13.60–13.65	15.60–15.65	7130 ±190	Gd-9360	marine mud		MGB PGI Archives	Hel Peninsula
144	II/93	54°40.20'	18°42.50'	–2.00	15.40–15.50	17.40–17.50	5560 ±150	Gd-9365	marine mud		MGB PGI Archives	Hel Peninsula
145	III/93	54°37.87'	18°45.27'	–4.00	14.85–14.90	18.85–18.90	1120 ±150	Gd-9354	marine mud		MGB PGI Archives	Hel Peninsula
146	III/93	54°37.87'	18°45.27'	–4.00	18.57–18.65	22.57–22.65	1500 ±110	Gd-10154	marine mud		MGB PGI Archives	Hel Peninsula
147	III/93	54°37.87'	18°45.27'	–4.00	18.57–18.65	22.57–22.65	3080 ±120	Gd-9371	marine mollusc shells		MGB PGI Archives	Hel Peninsula
148	III/93	54°37.87'	18°45.27'	–4.00	18.95–19.00	22.95–23.00	3710 ±150	Gd-9374	marine mud		MGB PGI Archives	Hel Peninsula
149	III/93	54°37.87'	18°45.27'	–4.00	19.62–19.68	23.62–23.68	4080 ±140	Gd-9367	marine mud		MGB PGI Archives	Hel Peninsula
150	III/93	54°37.87'	18°45.27'	–4.00	21.92–21.98	25.92–25.98	4620 ±160	Gd-9364	marine mud		MGB PGI Archives	Hel Peninsula
151	III/93	54°37.87'	18°45.27'	–4.00	22.50–22.56	26.50–26.56	5000 ±170	Gd-9372	marine mud		MGB PGI Archives	Hel Peninsula
152	III/93	54°37.87'	18°45.27'	–4.00	22.85–22.91	26.85–26.91	4890 ±140	Gd-9378	marine mud		MGB PGI Archives	Hel Peninsula
153	III/93	54°37.87'	18°45.27'	–4.00	26.96–27.00	30.96–31.00	4830 ±170	Gd-9373	marine mud		MGB PGI Archives	Hel Peninsula
154	Hel 10	54°37.80'	18°48.00'	1.50	0.90–1.00	(–0.60)–(–0.50)	1760 ±100	Gd-6112	peat		Tomczak, 1995b*	Hel Peninsula
155	Hel 11	54°37.90'	18°47.30'	1.00	0.70–0.80	(–0.30)–(–0.20)	3000 ±60	Gd-5597	peat		Tomczak, 1995b*	Hel Peninsula
156	Hel OGM 2/89	54°36.80'	18°48.00'	1.00	50.00–56.00	49.00–55.00	5540 ±120	Gd-4468	marine mud	pollen and diatoms	MGB PGI Archives	Hel Peninsula
157	Hel OGM 1/89	54°36.80'	18°48.00'	1.00	80.00–83.00	79.00–82.00	6170 ±180	Gd-4535	marine mud	pollen and diatoms	MGB PGI Archives	Hel Peninsula
158	Bór	54°38.50'	18°46.70'	3.00	33.90–34.20	30.90–31.20	6080 ±110	Gd-10094	marine mud	pollen and diatoms	Tomczak, 1995b*	Hel Peninsula
159	Bór	54°38.50'	18°46.70'	3.00	34.10–34.15	31.10–31.15	5840 ±170	Gd-9284	marine mud	pollen and diatoms	Tomczak, 1995b*	Hel Peninsula
160	Bór	54°38.50'	18°46.70'	3.00	38.80–39.00	35.80–36.00	5400 ±160	Gd-10081	marine mud	pollen and diatoms	Tomczak, 1995b*	Hel Peninsula
161	Bór	54°38.50'	18°46.70'	3.00	40.10–40.30	37.10–37.30	7150 ±100	Gd-10090	marine mud	pollen and diatoms	Tomczak, 1995b*	Hel Peninsula
162	Bór	54°38.50'	18°46.70'	3.00	45.70–45.90	42.70–42.90	6100 ±190	Gd-9271	marine mud	pollen and diatoms	Tomczak, 1995b*	Hel Peninsula
163	Bór	54°38.50'	18°46.70'	3.00	49.30–49.50	46.30–46.50	6310 ±200	Gd-9274	marine mud	pollen and diatoms	Tomczak, 1995b*	Hel Peninsula
164	Bór	54°38.50'	18°46.70'	3.00	60.10–60.30	57.10–57.30	5690 ±120	Gd-9266	marine mud	pollen and diatoms	Tomczak, 1995b*	Hel Peninsula
165	Bór	54°38.50'	18°46.70'	3.00	62.50–62.60	59.50–59.60	3560 ±60	Gd-7527	marine mollusc shells	pollen and diatoms	Tomczak, 1995b*	Hel Peninsula
166	Bór	54°38.50'	18°46.70'	3.00	62.50–62.60	59.50–59.60	5030 ±110	Gd-10092	marine mud	pollen and diatoms	Tomczak, 1995b*	Hel Peninsula
167	Bór	54°38.50'	18°46.70'	3.00	70.20–70.30	67.20–67.30	3730 ±80	Gd-11010	marine mollusc shells	pollen and diatoms	Tomczak, 1995b*	Hel Peninsula
168	Bór	54°38.50'	18°46.70'	3.00	70.20–70.30	67.20–67.30	3740 ±60	Gd-11008	marine mollusc shells	pollen and diatoms	Tomczak, 1995b*	Hel Peninsula
169	Bór	54°38.50'	18°46.70'	3.00	71.90–72.00	68.90–69.00	7510 ±200	Gd-9280	marine mud	pollen and diatoms	Tomczak, 1995b*	Hel Peninsula
170	Bór	54°38.50'	18°46.70'	3.00	76.00–76.20	73.00–73.20	7420 ±310	Gd-9278	marine mud	pollen and diatoms	Tomczak, 1995b*	Hel Peninsula
171	Bór	54°38.50'	18°46.70'	3.00	77.50–77.70	74.50–74.70	7810 ±130	Gd-9260	marine mud	pollen and diatoms	Tomczak, 1995b*	Hel Peninsula
172	Bór	54°38.50'	18°46.70'	3.00	79.90–80.10	76.90–77.10	7800 ±130	Gd-9263	marine mud	pollen and diatoms	Tomczak, 1995b*	Hel Peninsula
173	Ostrowo 7/2	54°50.20'	18°15.30'	1.00	1.10–1.20	0.10–0.20	6090 ±80	Gd-5298	peat		Tomczak, 1993*	central coast
174	Ostrowo 7/2	54°50.20'	18°15.30'	1.00	1.18–1.23	0.18–0.23	6200 ±50	Gd-5299	peat		Tomczak, 1993*	central coast
175	Ostrowo 6/4	54°49.90'	18°15.30'	1.00	1.14–1.20	0.14–0.20	5540 ±120	Gd-4256	peat		Tomczak, 1993*	central coast
176	Ostrowo 6/3	54°49.90'	18°15.30'	1.00	6.45–6.55	5.45–5.55	10,070 ±120	Gd-5297	peat		Tomczak, 1993*	central coast
177	Karwia K 1	54°49.95'	18°14.10'	0.50	7.00–7.05	6.50–6.55	7410 ±260	Gd-10392	peat		MGB PGI Archives	central coast
178	Karwia K 1	54°49.95'	18°14.10'	0.50	8.00–8.10	7.50–7.60	7760 ±100	Gd-9547	peat		MGB PGI Archives	central coast
179	Karwia K 1	54°49.95'	18°14.10'	0.50	21.90–22.00	21.40–21.50	10,240 ±350	Gd-9549	limnic gyttja	ostracods	MGB PGI Archives	central coast
180	Lubiatowo U3	54°48.59'	17°50.78'	6.20	8.00–8.05	1.80–1.85	2090 ±90	Gd-11416	peat		MGB PGI Archives	central coast
181	Stilo S 1	54°47.78'	17°44.10'	0.50	11.95–12.00	11.45–11.50	8020 ±60	Gd-7765	peat		MGB PGI Archives	central coast
182	Mierzeja Sarbska A-B III	54°47.00'	17°40.60'	2.70	3.90–3.95	1.20–1.25	5480 ±90	Gd-2067	peat	pollen and diatoms	Miotk, Bogaczewicz-Adamczak, 1987	central coast
183	Mierzeja Sarbska A-B III	54°47.00'	17°40.60'	2.70	12.30–12.70	9.60–10.00	7590 ±100	Gd-2068	lagoonal limy mud	pollen and diatoms	Miotk, Bogaczewicz-Adamczak, 1987	central coast
184	Łeba Ł 1	54°46.18'	17°35.50'	3.00	3.00–3.05	0.00–0.05	1180 ±50	Gd-7756	peat		MGB PGI Archives	central coast
185	Łeba Ł 1	54°46.18'	17°35.50'	3.00	3.45–3.50	0.45–0.50	1670 ±110	Gd-10406	peat		MGB PGI Archives	central coast
186	Łeba Neptun	54°46.20'	17°34.40'	2.25	1.80–1.85	(–0.45)–(–40)	2450 ±140	Gd-416	peat		Tobolski, 1979	central coast

No.	Site/core	Coordinates		Altitude (m a.s.l.)	Sample position in core (m)	Sample position (m b.s.l.)	^{14}C age BP	Lab. code	Material dated	Supporting analyses	References	Locality
		ϕ	λ									
187	Łeba Neptun	54°46.20’	17°34.40’	2.25	2.05–2.10	(–0.20)–(–0.15)	4610 ±250	Gd-415	limnic gyttja		Tobolski, 1979	central coast
188	Łeba południe p. 16	54°45.20’	17°33.70’	1.50	4.00–4.30	2.50–2.80	7210 ±70	Gd-3148	peat		Morawski, 1989	central coast
189	Mierzeja Łebska	54°45.40’	17°29.00’	0.00	0.50–0.55	0.50–0.55	2910 ±70	Gd-5066	peat		Borówka, Rotnicki, 1988	central coast
190	Mierzeja Łebska	54°45.40’	17°29.00’	0.00	10.00–10.05	10.00–10.05	9840 ±110	Gd-5067	peat		Borówka, Rotnicki, 1988	central coast
191	Łeba 93A -172	54°39.78’	17°16.07’	2.50	7.90–8.20	5.40–5.70	4860 ±380	Gd-9273	*Cerastoderma* shells	malacofauna	Rotnicki *et al*., 1999	central coast
192	Łeba 93A -172	54°39.78’	17°16.07’	2.50	8.20–8.40	5.70–5.90	4880 ±170	Gd-9285	*Cerastoderma* shells	malacofauna	Rotnicki *et al*., 1999	central coast
193	Łeba 93A	54°39.78’	17°16.07’	2.50	8.80–8.90	6.30–6.40	5570 ±120	Gd-10096	*Cerastoderma* shells	malacofauna	Rotnicki *et al*., 1999	central coast
194	Łeba 93A	54°39.78’	17°16.07’	2.50	9.30–9.40	6.80–6.90	6640 ±120	Gd-10097	*Cerastoderma* shells	malacofauna	Rotnicki *et al*., 1999	central coast
195	Łeba 172	54°39.78’	17°16.07’	2.50	9.30–9.40	6.80–6.90	5750 ±110	Gd-10088	*Cerastoderma* shells	malacofauna	Rotnicki *et al*., 1999	central coast
196	Łeba 172	54°39.78’	17°16.07’	2.50	9.40–9.65	6.90–7.15	6700 ±160	Gd-9292	*Cerastoderma* shells	malacofauna	Rotnicki *et al*., 1999	central coast
197	Łeba 93A	54°39.78’	17°16.07’	2.50	9.80–9.90	7.30–7.40	5370 ±280	Gd-9282	*Cerastoderma* shells	malacofauna	Rotnicki *et al*., 1999	central coast
198	Łeba 93A	54°39.78’	17°16.07’	2.50	9.90–10.35	7.40–7.85	6660 ±180	Gd-9289	*Cerastoderma* shells	malacofauna	Rotnicki *et al*., 1999	central coast
199	Łeba 93A	54°39.78’	17°16.07’	2.50	10.35–10.40	7.85–7.90	6000 ±330	Gd-9287	*Cerastoderma* shells	malacofauna	Rotnicki *et al*., 1999	central coast
200	Łeba 93A -172	54°39.78’	17°16.07’	2.50	10.80–11.30	8.30–8.80	6800 ±180	Gd-9272	*Cerastoderma* shells	malacofauna	Rotnicki *et al*., 1999	central coast
201	Łeba 93A	54°39.78’	17°16.07’	2.50	11.30–11.40	8.80–8.90	7850 ±140	Gd-10017	lagoonal mud		Rotnicki *et al*., 1999	central coast
202	J. Łebsko D 1a	54°45.20’	17°28.70’	–1.50	8.32–8.37	9.82–9.87	9700 ±110	Gd-4784	peat		Wojciechowski, 1995	central coast
203	J. Łebsko C 4	54°44.40’	17°31.20’	–1.70	2.38–2.43	4.08–4.13	6620 ±70	Gd-7021	peat		Wojciechowski, 1995	central coast
204	J. Łebsko C 4	54°44.40’	17°31.20’	–1.70	2.79–2.84	4.49–4.54	7100 ±70	Gd-7024	peat		Wojciechowski, 1995	central coast
205	Madwiny 1	54°45.64’	17°22.80’	–10.00	7.50–8.00	17.50–18.00	1430 ±50	Gd-12357	*Cerastoderma* shells		MGB PGI Archives	central coast
206	Kluki 74	54°40.00’	17°19.00’	2.10	2.40–2.45	0.30–0.35	3865 ±70	Gd-563	peat	pollen	Tobolski, 1987	central coast
207	Kluki 74	54°40.00’	17°19.00’	2.10	2.75–2.80	0.65–0.70	4675 ±55	Gd-1330	peat	pollen	Tobolski, 1987	central coast
208	Kluki 74	54°40.00’	17°19.00’	2.10	3.10–3.15	1.00–1.05	4910 ±90	Gd-1315	peat	pollen	Tobolski, 1987	central coast
209	Kluki 74	54°40.00’	17°19.00’	2.10	3.40–3.45	1.30–1.35	5050 ±45	Gd-1331	peat	pollen	Tobolski, 1987	central coast
210	Kluki 74	54°40.00’	17°19.00’	2.10	3.80–3.85	1.70–1.75	5305 ±75	Gd-562	peat	pollen	Tobolski, 1987	central coast
211	Kluki 74	54°40.00’	17°19.00’	2.10	4.30–4.35	2.20–2.25	6055 ±120	Gd-855	peat	pollen	Tobolski, 1987	central coast
212	Kluki 74	54°40.00’	17°19.00’	2.10	4.60–4.65	2.50–2.55	6670 ±120	Gd-1323	peat	pollen	Tobolski, 1987	central coast
213	Kluki 74	54°40.00’	17°19.00’	2.10	5.05–5.10	2.95–3.00	7400 ±60	Gd-1587	peat	pollen	Tobolski, 1987	central coast
214	Kluki 74	54°40.00’	17°19.00’	2.10	5.40–5.45	3.30–3.35	8130 ±85	Gd-1322	peat	pollen	Tobolski, 1987	central coast
215	Kluki 74	54°40.00’	17°19.00’	2.10	5.75–5.80	3.65–3.70	8370 ±115	Gd-547	peat	pollen	Tobolski, 1987	central coast
216	Kluki 74	54°40.00’	17°19.00’	2.10	6.30–6.35	4.20–4.25	9110 ±70	Gd-1329	peat	pollen	Tobolski, 1987	central coast
217	Kluki 74	54°40.00’	17°19.00’	2.10	6.55–6.60	4.45–4.50	9855 ±315	Hv-9104	peat	pollen	Tobolski, 1987	central coast
218	Kluki 74	54°40.00’	17°19.00’	2.10	6.60–6.35	4.50–4.25	9865 ±105	Gd-548	limnic gyttja	pollen	Tobolski, 1987	central coast
219	Mierzeja Łebska otw. 35	54°40.60’	17°14.00’	3.00	3.40–3.60	0.40–0.60	13,800 ±270	Gd-6117	peat		Rotnicki, Borówka, 1995a, b	central coast
220	Mierzeja Łebska otw. 101	54°40.50’	17°5.00’	3.50	13.00–13.00	9.50–9.50	14,310 ±150	Gd-4476	ice marginal lake clay		Rotnicki, Borówka, 1995a, b	central coast
221	J. Gardno A 3	54°39.00’	17°8.90’	–2.00	0.55–0.60	2.55–2.60	9090 ±70	Gd-1286	peat		Wojciechowski, 1990	central coast
222	J. Gardno A 3	54°39.00’	17°8.90’	–2.00	1.50–1.56	3.50–3.56	10,280 ±120	Gd-1288	peat		Wojciechowski, 1990	central coast
223	J. Gardno D 2	54°38.30’	17°7.50’	–1.10	0.25–0.30	1.35–1.40	5380 ±55	Gd-1285	peat		Wojciechowski, 1990	central coast
224	J. Gardno D 2	54°38.30’	17°7.50’	–1.10	1.35–1.41	2.45–2.51	10,140 ±150	Gd-827	peat		Wojciechowski, 1990	central coast
225	J. Gardno D 7	54°39.40’	17°7.60’	–1.80	1.59–1.66	3.39–3.46	9210 ±100	Gd-1550	peat		Wojciechowski, 1990	central coast
226	J. Gardno D 7	54°39.40’	17°7.60’	–1.80	1.96–2.01	3.76–3.81	10,420 ±110	Gd-1551	peat		Wojciechowski, 1990	central coast
227	J. Gardno G 1	54°38.10’	17°6.40’	–1.00	0.49–0.56	1.49–1.56	6400 ±80	Gd-1555	peat		Wojciechowski, 1990	central coast
228	J. Gardno G 1	54°38.10’	17°6.40’	–1.00	2.16–2.22	3.16–3.22	9360 ±100	Gd-1556	peat		Wojciechowski, 1990	central coast
229	J. Gardno H 1	54°38.30’	17°5.90’	–0.90	0.20–0.25	1.10–1.15	6080 ±130	Gd-923	peat		Wojciechowski, 1990	central coast
230	J. Gardno H 1	54°38.30’	17°5.90’	–0.90	2.28–2.33	3.18–3.23	10,110 ±80	Gd-1435	peat		Wojciechowski, 1990	central coast
231	J. Gardno I 4	54°39.20’	17°5.20’	–1.20	1.77–1.83	2.97–3.03	6350 ±90	Gd-1553	peat		Wojciechowski, 1990	central coast
232	J. Gardno I 4	54°39.20’	17°5.20’	–1.20	2.63–2.69	3.83–3.89	8550 ±90	Gd-1554	peat		Wojciechowski, 1990	central coast
233	J. Gardno I 7	54°40.00’	17°5.20’	–1.00	1.67–1.72	2.67–2.72	6280 ±80	Gd-1658	peat		Wojciechowski, 1990	central coast

Appendix 1 continued

No.	Site/core	Coordinates		Altitude (m a.s.l.)	Sample position in core (m)	Sample position (m b.s.l.)	^{14}C age BP	Lab. code	Material dated	Supporting analyses	References	Locality
		ϕ	λ									
234	J. Gardno I 7	54°40.00'	17°5.20'	–1.00	1.94–1.99	2.94–2.99	6710 ±60	Gd-1657	peat		Wojciechowski, 1990	central coast
235	J. Gardno I 10	54°37.30'	17°5.20'	1.50	5.40–5.50	3.90–4.00	10,190 ±110	Gd-1660	peat		Wojciechowski, 1990	central coast
236	J. Gardno K 4	54°37.30'	17°4.50'	–0.90	2.33–2.40	3.23–3.30	7000 ±80	Gd-1443	peat		Wojciechowski, 1990	central coast
237	J. Gardno K 4	54°39.20'	17°4.50'	–0.90	2.97–3.02	3.87–3.92	7960 ±70	Gd-1437	peat		Wojciechowski, 1990	central coast
238	J. Gardno L 1	54°39.20'	17°3.80'	0.60	1.26–1.31	0.66–0.71	2390 ±90	Gd-2136	peat		Wojciechowski, 1990	central coast
239	J. Gardno L 1	54°38.60'	17°3.80'	0.60	1.59–1.64	0.99–1.04	6230 ±80	Gd-1655	peat		Wojciechowski, 1990	central coast
240	J. Gardno L 1	54°38.60'	17°3.80'	0.60	1.99–1.99	1.35–1.39	6610 ±80	Gd-1656	peat		Wojciechowski, 1990	central coast
241	Wicko 3	54°34.50'	16°47.30'	0.00	0.00–0.10	0.00–0.10	3130 ±70	Gd-5630	peat		Tomczak, 1995b*	central coast
242	Mierzeja Kopań A/1	54°31.80'	16°30.90'	0.00	0.00–0.05	0.00–0.05	3240 ±50	Gd-5387	peat		Tomczak, 1995b*	central coast
243	Łazy-2	54°18.00'	16°13.00'	2.00	6.20–6.25	4.20–4.25	5580 ±90	Gd-6092	peaty gyttja		Filonowicz, 1990*	western coast
244	Łazy-3	54°18.00'	16°13.00'	2.00	6.80–6.90	4.80–4.90	6250 ±70	Gd-5561	limnic gyttja		Filonowicz, 1990*	western coast
245	Łabusz 1	54°16.60'	16°11.50 '	3.70	2.70–2.80	(–1.00)–(–0.90)	3690 ±60	Gd-5340	peat		Filonowicz, 1990*	western coast
246	Łabusz 2	54°16.60'	16°11.50'	3.70	3.60–3.70	(–0.10)–0,00	6730 ±80	Gd-5345	marine mollusc shells	malacofauna	Filonowicz, 1990*	western coast
247	Grzybowo A-2	54°10.50'	15°31.30'	4.00	7.20–7.30	3.20–3.30	6210 ±100	Gd-2945	peat		Tomczak, 1995b*	western coast
248	Mierzeja Resko 4T	54°9.40'	15°22.00'	0.30	3.53–3.60	3.23–3.30	6210 ±60	Gd-7846	peat		MGB PGI Archives	western coast
249	Mierzeja Resko 4T	54°9.40'	15°22.00'	0.30	3.90–4.00	3.60–3.70	6480 ±60	Gd-7845	peat		MGB PGI Archives	western coast
250	Mrzeżyno 5 L	54°8.50'	15°19.00'	1.50	11.00–11.05	9.50–9.55	7620 ±60	Gd-7847	peat		MGB PGI Archives	western coast
251	Pogorzelica 4 L	54°6.30'	15°9.40'	6.00	14.20–14.90	8.20–8.90	7060 ±90	Gd-11296	peat		MGB PGI Archives	western coast
252	Niechorze 4 N	54°5.70'	15°6.10'	0.50	5.30–5.40	4.80–4.90	3690 ±200	Gd-9671	lagoonal mud	diatoms	MGB PGI Archives	western coast
253	Niechorze 4 N	54°5.70'	15°6.10'	0.50	6.90–7.00	6.40–6.50	4860 ±110	Gd-10534	lagoonal mud	diatoms	MGB PGI Archives	western coast
254	Niechorze 4 N	54°5.70'	15°6.10'	0.50	12.10–12.20	11.60–11.70	7500 ±40	Gd-7854	peat	diatoms	MGB PGI Archives	western coast
255	Niechorze 4 N	54°5.70'	15°6.10'	0.50	13.85–14.00	13.35–13.50	8050 ±240	Gd-9659	lagoonal mud	diatoms	MGB PGI Archives	western coast
256	Wolin II	53°51.20'	14°37.20'	1.50	1.43–1.47	(–0.07)–(–0.03)	4130 ±60	Gd-5264	peat		Latałowa, 1992	western coast
257	Wolin II	53°51.20'	14°37.20'	1.50	1.63–1.68	0.13–0.18	4930 ±60	Gd-5392	peat		Latałowa, 1992	western coast
258	Wolin II	53°51.20'	14°37.20'	1.50	2.03–2.07	0.53–0.57	5860 ±110	Gd-2789	peat		Latałowa, 1992	western coast
259	Wolin II	53°51.20'	14°37.20'	1.50	2.43–2.48	0.93–0.98	6340 ±70	Gd-4225	peat		Latałowa, 1992	western coast
260	Wolin II	53°51.20'	14°37.20'	1.50	2.70–2.75	1.20–1.25	7320 ±520	Gd-1229	peaty muddy sand		Latałowa, 1992	western coast
261	Mierzeja Świny p. 4	53°54.80'	14°25.80'	1.00	2.60–2.70	1.60–1.70	4460 ±300	Gd-47	peat	pollen	Prusinkiewicz, Noryśkiewicz, 1966	western coast
262	Mierzeja Świny p. 5	53°53.20'	14°20.30'	1.00	1.05–1.10	0.05–0.10	1403 ±230	Gd-49	peat	pollen	Prusinkiewicz, Noryśkiewicz, 1966	western coast
263	Mierzeja Świny p. 3	53°50.10'	14°19.90'	1.00	0.90–1.00	(–0.10)–0.00	2910 ±445	Gd-46	peat	pollen	Prusinkiewicz, Noryśkiewicz, 1966	western coast
264	Mierzeja Świny p. 2	53°50.20'	14°17.30'	1.00	1.45–1.50	0.45–0.50	3035 ±285	Gd-50	peat	pollen	Prusinkiewicz, Noryśkiewicz, 1966	western coast
265	Mierzeja Świny p. 1	53°53.50'	14°13.40'	1.00	2.50–2.60	1.50–1.60	4810 ±365	Gd-45	peat	pollen	Prusinkiewicz, Noryśkiewicz, 1966	western coast
266	Z 13	53°45.10'	14°21.00'	–5.35	1.10–1.18	6.45–6.53	5386 ±300		peat		Wypych, 1980	Szczecin Lagoon
267	Z 30	53°41.90'	14°24.30'	–4.75	1.23–1.41	5.98–6.16	6420 ±340		peat		Wypych, 1980	Szczecin Lagoon
268	Z 30	53°41.90'	14°24.30'	–4.75	1.56–1.64	6.31–6.39	8944 ±400		peat		Wypych, 1980	Szczecin Lagoon
269	R 74	53°55.92'	14°23.22'	–10.00	1.20–1.35	11.20–11.35	7240 ±150	Gd-1043	peat		Jurowska, Kramarska, 1990*	Pomeranian Bay
270	R 74	53°55.92'	14°23.22'	–10.00	1.35–1.50	11.35–11.50	7695 ±120	Gd-1044	peat		Jurowska, Kramarska, 1990*	Pomeranian Bay
271	R 74	53°55.92'	14°23.22'	–10.00	1.50–1.78	11.50–11.78	8090 ±105	Gd-1042	peat		Jurowska, Kramarska, 1990*	Pomeranian Bay
272	W-4	54°15.70'	14°44.20'	–17.30	9.21–9.23	26.51–26.53	13100 ±300	Gd-4336	peat	malacofauna	Kramarska, Jurowska, 1991*	Pomeranian Bay
273	W-4	54°15.70'	14°44.20'	–17.30	9.69–9.72	26.99–27.02	13350 ±270	Gd-4335	peat	malacofauna	Kramarska, Jurowska, 1991*	Pomeranian Bay
274	W-4	54°15.70'	14°44.20'	–17.30	9.75–9.80	27.05–27.10	13490 ±190	Gd-2929	peat	malacofauna	Kramarska, Jurowska, 1991*	Pomeranian Bay
275	W-4	54°15.70'	14°44.20'	–17.30	10.42–10.50	27.72–27.80	14060 ±220	Gd-2928	peat	malacofauna	Kramarska, Jurowska, 1991*	Pomeranian Bay
276	R-86	54°5.00'	14°38.00'	–13.00	2.50–2.60	15.50–15.60	12010 ±100	Gd-10048	peat		Kramarska, 1998*	Pomeranian Bay
277	R-88	54°5.00'	14°38.00'	–12.50	1.75–1.85	14.25–14.35	9880 ±140	Gd-10049	peat		Kramarska, 1998*	Pomeranian Bay
278	51-I	54°19.90'	14°35.30'	–16.40	2.40–2.50	18.80–18.90	9500 ±50	Gd-7505	peat		Kramarska, 1998*	Pomeranian Bay
279	P IX 2	54°18.40'	14°41.40'	–18.40	0.75–0.85	19.15–19.25	8070 ±240	Gd-9309	limnic mud	ostracods	Kramarska, 1998*	Pomeranian Bay
280	P IX 4	54°18.40'	14°42.10'	–18.00	0.83–0.90	18.83–18.90	4510 ±50	Gd-7553	*Cerastoderma* shells		Kramarska, 1998*	Pomeranian Bay

Appendix 1 continued

No.	Site/core	Coordinates		Altitude (m a.s.l.)	Sample position in core (m)	Sample position (m b.s.l.)	^{14}C age BP	Lab. code	Material dated	Supporting analyses	References	Locality
		ϕ	λ									
281	P IX 4	54°18.40’	14°42.10’	–18.00	0.90–1.00	18.90–19.00	6630 ±140	Gd-10126	lagoonal mud		Kramarska, 1998*	Pomeranian Bay
282	95	54°16.80’	14°55.40’	–18.50	0.47–0.50	18.97–19.00	5190 ±130	Gd-6515	*Cerastoderma* shells		Kramarska, 1998*	Pomeranian Bay
283	P VI 82	54°22.90’	14°40.00’	–11.30	0.72–0.80	12.02–12.10	5100 ±200	Gd-9313	plant detritus		Kramarska, 1998*	Odra Bank
284	P VI 82	54°22.90’	14°40.00’	–11.30	0.72–0.80	12.02–12.10	6770 ±120	Gd-10135	lake marl		Kramarska, 1998*	Odra Bank
285	P VI 89	54°23.20’	14°39.50’	–11.50	0.50–0.55	12.00–12.05	4410 ±110	Gd-10137	*Cerastoderma* shells		Kramarska, 1998*	Odra Bank
286	P VI 89	54°23.20’	14°39.50’	–11.50	0.62–0.67	12.12–12.17	8410 ±120	Gd-10136	lake marl		Kramarska, 1998*	Odra Bank
287	P VI 91	54°23.20’	14°40.20’	–12.30	0.75–0.85	13.05–13.15	9500 ±580	Gd-9327	limnic muddy sand		Kramarska, 1998*	Odra Bank
288	116	54°20.00’	14°40.80’	–9.60	0.15–0.20	9.75–9.80	1400 ±70	Gd-5979	*Cerastoderma* shells		Kramarska, 1998*	Odra Bank
289	76	54°19.50’	15°1.50’	–16.50	0.20–0.25	16.70–16.75	350 ±200	Gd-6521	*Cerastoderma* shells		Kramarska, 1998*	Odra Bank
290	50	54°22.70’	14°51.70’	–13.90	0.17–0.21	14.07–14.11	200 ±150	Gd-6520	*Cerastoderma* shells		Kramarska, 1998*	Odra Bank
291	47	54°24.40’	14°48.00’	–13.50	0.10–0.17	13.60–13.67	0 ±	Gd-5980	*Cerastoderma* shells		Kramarska, 1998*	Odra Bank
292	14 097	54°55.66’	17°8.52’	–20.70	2.68–2.73	23.38–23.43	8950 ±70	Gd-3229	peat	pollen and malacofauna	Uścinowicz, Zachowicz, 1991b*	Słupsk Bank
293	14 097	54°55.66’	17°8.52’	–20.70	2.73–2.80	23.43–23.50	9850 ±80	Gd-1947	peat	pollen and malacofauna	Uścinowicz, Zachowicz, 1991b*	Słupsk Bank
294	14 097 B	54°55.67’	17°8.59’	–21.10	0.70–0.85	21.80–21.95	9240 ±90	Gd-5216	wood		Uścinowicz, Zachowicz, 1991a*	Słupsk Bank
295	14 097 B	54°55.67’	17°8.59’	–21.10	2.95–2.97	24.05–24.07	9320 ±150	Gd-4190	peat	pollen	Uścinowicz, Zachowicz, 1991a*	Słupsk Bank
296	14 097 B	54°55.67’	17°8.59’	–21.10	3.22–3.30	24.32–24.40	9320 ±90	Gd-5217	peat	pollen	Uścinowicz, Zachowicz, 1991a*	Słupsk Bank
297	14 097 B	54°55.67’	17°8.59’	–21.10	3.35–3.40	24.45–24.50	9620 ±160	Gd-4191	peat	pollen	Uścinowicz, Zachowicz, 1991a*	Słupsk Bank
298	14 097 B	54°55.67’	17°8.59’	–21.10	3.50–3.55	24.60–24.65	9720 ±160	Gd-2752	peat	pollen	Uścinowicz, Zachowicz, 1991a*	Słupsk Bank
299	14 097 B	54°55.67’	17°8.59’	–21.10	3.55–3.58	24.65–24.68	10060 ±130	Gd-2755	peat	pollen	Uścinowicz, Zachowicz, 1991a*	Słupsk Bank
300	14 097 B	54°55.67’	17°8.59’	–21.10	3.58–3.62	24.68–24.72	10510 ±170	Gd-4187	peat	pollen	Uścinowicz, Zachowicz, 1991a*	Słupsk Bank
301	R 10 (192)	55°34.70’	17°25.90’	–20.00	1.05–1.10	21.05–21.10	10220 ±100	Gd-10304	peat		MGB PGI Archives	S. Middle Bank
302	11 K 02/A	55°2.21’	15°44.00’	–85.50	3.10–3.20	88.60–88.70	7750 ±140	Gd-6316	marine mud	diatoms	Zachowicz, 1995	Bornholm Basin
303	11 K 02/A	55°2.21’	15°44.00’	–85.50	3.70–3.80	89.20–89.30	8680 ±150	Gd-6324	marine mud	diatoms	MGB PGI Archives	Bornholm Basin
304	11 K 02/A	55°2.21’	15°44.00’	–85.50	3.80–3.90	89.30–89.40	8800 ±150	Gd-6317	marine mud	diatoms	Zachowicz, 1995	Bornholm Basin
305	12 K 02/A	55°1.69’	15°44.00’	–85.50	2.90–3.05	88.40–88.55	7310 ±150	Gd-6314	marine mud		MGB PGI Archives	Bornholm Basin
306	12 K 02	55°1.68’	15°44.03’	–85.50	2.90–3.10	88.40–88.60	6700 ±90	Gd-2906	marine mud	pollen	Uścinowicz, Zachowicz, 1993d*	Bornholm Basin
307	12ŁB625	55°11.26’	17°2.6’	–80.20	2.93–3.03	83.13–83.23	7680 ±140	Gd-6309	marine mud	pollen	MGB PGI Archives	Słupsk Furrow
308	GT 5/398	55°28.79’	18°19.19’	–85.50	1.66–1.81	87.16–87.31	6570 ±150	Gd-4633	marine mud		Pikies, 1995a	Gotland Basin
309	GT 5A/82	55°21.95’	18°46.56’	–87.80	1.80–1.90	89.60–89.70	7920 ±150	Gd-10005	marine mud	pollen	Zachowicz, 1995	Gotland Basin
310	GT 5A/82	55°21.95’	18°46.56’	–87.80	2.20–2.30	90.00–90.10	8160 ±420	Gd-9173	marine mud	pollen	Zachowicz, 1995	Gotland Basin
311	GT 5A/82	55°21.95’	18°46.56’	–87.80	2.30–2.40	90.10–90.20	8620 ±200	Gd-9174	marine mud	pollen	Zachowicz, 1995	Gotland Basin
312	GT 5/382	55°28.40’	18°20.60’	–86.00	2.92–3.01	88.92–89.01	5850 ±170	Gd-9170	marine mud	pollen	Zachowicz, 1995	Gotland Basin
313	GT 5/362	55°27.90’	18°22.30’	–87.00	0.45–0.55	87.45–87.55	6290 ±560	Gd-9189	marine mud		MGB PGI Archives	Gotland Basin
314	PŁ 3/327	55°26.72’	17°39.64’	–67.00	1.25–1.40	68.25–68.40	7190 ±100	Gd-6304	marine mud		Pikies, 1995a	Gotland Basin

MGB PGI Archives — Archives of Polish Geological Institute, Marine Geology Branch, Gdańsk

* — also in MGB PGI Archives