Technical Report

TR-99-38

Salinity change in the Baltic Sea during the last 8,500 years: evidence, causes and models

Per Westman, Stefan Wastegård, Kristian Schoning
Department of Quaternary Research, Stockholm University

Bo Gustafsson
Oceanus Havsundersökningar, Göteborg

Anders Omstedt, SMHI, Norrköping

December 1999

Svensk Kärnbränslehantering AB
Swedish Nuclear Fuel
and Waste Management Co
Box 5864
SE-102 40 Stockholm Sweden
Tel 08-459 84 00
+46 8 459 84 00
Fax 08-661 57 19
+46 8 661 57 19

Salinity change in the Baltic Sea during the last 8,500 years: evidence, causes and models

Per Westman, Stefan Wastegård, Kristian Schoning
Department of Quaternary Research, Stockholm University

Bo Gustafsson
Oceanus Havsundersökningar, Göteborg

Anders Omstedt, SMHI, Norrköping

December 1999

Keywords: biosphere, salinity, landrise, surface ecosystem, palaeo salinity

Abstract

The salinity influences which ecosystems will dominate in the coastal area and what property radionuclides have. Salinity is also an important boundary condition for the transport models in the geosphere. Knowledge about the past salinity is important background to evaluate the hydrology and geochemistry in the rock and further to assess the radiological consequences of possible releases from a radioactive repository.

This report concerns the salinity in the Baltic Sea during the last 8500 calendar years BP. Shore-level data for the inlet areas and proxy (indirect) data for the palaeosalinity and the climate are reviewed. These data is furthered used in a steady-state model for the salt exchange between the Baltic Sea and Kattegat. This will then be extended to a model of the future development of the salinity in the Baltic Sea.

We conclude that the changes in the inlet cross-section areas together with a 15% to 60% lower net freshwater input compared to the present input can explain the higher salinity in the Baltic Sea during earlier times.

Sammanfattning

Saliniteten är avgörande för vilket ekosystem som dominerar i kustområden och för radionuklidernas uppträdande. Den utgör också ett viktigt gränsvillkor för transportmodeller i geosfären. Kunskapen om salinitetsförändringar i ett längre tidsperspektiv utgör en viktig bakgrund vid för utvärderingen av hydrogeologin och geokemin i berggrunden och vidare för att kunna uppskatta konsekvenserna av ett möjligt utsläpp från ett radioaktivt lager.

Denna rapport behandlar salinitetsvariationer i Östersjön under de senaste 8 500 kalenderåren, dvs den tid som Östersjön varit ett brackvattenhav (Litorinahavet). Huvudsyftet var att sammanställa och korrelera befintliga data för salthaltsförändringar och att beräkna hur stor roll tröskelarean i de danska sunden (Öresund eller Drogden och Darss trösklarna) och klimatet spelat för dessa förändringar. Förändringar i tröskelarean användes i en matematisk modell med 500 års upplösning för saltutbytet mellan Östersjön och Kattegatt (Gustafsson, 1997). Modellvärdena jämfördes sedan med proxydata (indirekta data) för salthalt, klimat och syresättning av djupvattnet i Östersjön. Trots att kunskapen om salinitetsförändringar i Östersjön fortfarande är bristfällig har vi försökt att summera vad vi anser vara välgrundade åsikter om detta nedan. Genom hela rapporten anges åldrar som kalenderår före 1950 (years BP) och salinitetsangivelserna refererar till medelsalthalten i egentliga Östersjöns (söder om Ålandshav) ytvatten om inget annat anges.

- Brackvatten började flöda in i Östersjön genom de danska sunden cirka 8 500 BP. Saliniteten ökade mer eller mindre konstant till cirka 6 000 BP. Mellan cirka 6 000 och 4 500 BP var salthalten förhållandevis konstant.

- Öresund öppnades ca 8 500 BP. Tvärsnittsarean var som störst ca 5 500 BP (dubbelt så stor som dagens). Efter 5 500 BP minskade arean och nådde sitt nuvarande värde cirka 2 000 BP.

- Sundet vid Darsströskeln öppnades 9 000–8 500 BP. Tvärsnittsarean ökade fram till cirka 6 000 BP, då den nådde det nuvarande värdet. Från 6 000 BP till nutid har tvärsnittsarean varierat obetydligt.

- Analyser av fossil, $\delta^{18}O$ och spårelement indikerar att salinitetsmaximum under Litorinastadiet låg mellan 10 och 15 ‰ (idag cirka 7 ‰). Enstaka molluskfynd pekar dock på att saliniteten kan ha nått 20 ‰.

- Under Holocen (de senaste 11 500 kalenderåren) har ett antal perioder med "klimatförsämringar" kunnat påvisas : cirka 8 200–7 800, 6 000–5 600, 4 400–4 000, 2 900–2 500, 1 500–1 100 och 500–100 BP. Det är troligt att dessa perioder då fuktigare och kallare klimat rådde kan ha ökat nettotillskottet av sötvatten med påföljande sänkning av saliniteten i Östersjön.

- Perioden med den högsta saliniteten (6 000–5 000 BP) under Litorinastadiet sammanfaller med den period då tvärsnittsareorna i sunden var som störst. Det varma klimatet under denna period var antagligen också en bidragande orsak till den höga saliniteten.

- Det är troligt att det instabila klimatet efter 4 500 BP har varit den starkast bidragande orsaken till den observerade salinitetsminskningen 2 000–1 500 BP.

- Enligt den model för saltutbyte vi använt (Gustafsson 1997) kan förändringen i tvärsnittsareorna tillsammans med en variation i nettotillskott av sötvatten (jämfört med det nutida tillskottet) mellan 15 % (om salinitets maximum sätts till 10 ‰) och 60 % (om salinitets maximum sätts till 15 ‰) förklara salinitetsförändringen i Östersjön under de senaste 8 500 åren.

Summary and conclusions

The salinity influences which ecosystems will dominate in the coastal area and what property radionuclides have. Salinity is also an important boundary condition for the transport models in the geosphere. Knowledge about the past salinity is important background to evaluate the hydrology and geochemistry in the rock and further to assess the radiological consequences of possible releases from a radioactive repository.

In this report the past changes of salinity during Holocene are described and modelled. This will then be extended to a model of the future development of the salinity in the Baltic Sea.

The objective of this study was to find a mathematical expression for describing the relative importance of changes in the cross-section areas of the Öresund strait and Darss sill compared to direct climatic influences (temperature and net freshwater input) on the salinity of the Baltic Sea. We have used shore-level data and proxy data for the palaeosalinity and the climate. These data were then used in a steady-state model for the salt exchange between the Baltic Sea and Kattegat (Gustafsson, 1997). Although the knowledge of absolute salinity changes in the Baltic Sea is still incomplete, we have tried to summarise below what we believe are well-founded opinions. Throughout this study ages are given in calendar years BP (before 1950) and the salinities given refer to the mean surface salinities in the Baltic proper (the Baltic Sea south of the islands of Åland) unless otherwise stated.

- Brackish water started to intrude into the Baltic Sea at c. 8,500 BP through the straits in the south. The salinity increased more or less continuously until c. 6,000 BP and remained stable until c. 4,500 BP.
- The Öresund strait area opened 8,500 BP and the cross-section area reached a maximum c. 5,500 BP (double of the present value). After 5,500 BP it started to decrease and reached its present value at c. 2,000 BP.
- The Darss sill area opened 9,000–8,500 BP and the cross-section area increased to the present value at c. 6,000 BP; since then it has been more or less stable.
- Two phases of decreasing salinity have been recognised after the salinity maximum. The first phase is c. 5,000–4,000 BP, and the second is c. 2,000–1,500 BP during which the Baltic Sea reached its present salinity.
- The maximum salinity during the Litorina Sea stage was between 10 and 15‰ (today c. 7‰) as indicated by fossil assemblages, $\delta^{18}O$ analysis and trace-element content in sediments, even though single findings of molluscs indicate that the salinity may have reached 20‰.
- A number of periods with climate deterioration have been recorded in the Baltic Sea area and around the North Atlantic during the Holocene. It is possible that periods of increased wetness and reduced evapotranspiration may have increased the freshwater input to the Baltic Sea and decreased the salinity. Cold and possibly wet phases have been recorded at c. 8,200–7,800, 6,000–5,600, 4,400–4,000, 2,900–2,500, 1,500–1,100 and 500–100 BP.

- The highest salinity of the Litorina Sea (6,000–5,000 BP) coincides with a period when the cross-section areas of the inlets (Öresund and Darss) were at maximum. The climate at this time (warm and humid) also favoured a high salinity in the Baltic Sea.
- It is possible that the unstable climate after c. 4,500 BP may have caused some of the observed salinity changes, for example at 2,000–1,500 BP. It is not possible, however, to evaluate from the present study exactly how climate changes in the drainage area have affected the conditions in the Baltic Sea.
- According to the circulation model used (Gustafsson, 1997) the salinity variations during the last 8,500 years can be explained by the changes in cross-section areas of the inlets together with a variation in net freshwater input (compared to the present value) between 15% (if a maximum salinity of 10‰ is assumed) and 60% (if a maximum salinity of 15‰ is assumed).

Content

1 Introduction

1.1 Project background and objective

In order to be able assess the radiological consequences of possible releases from a radioactive repository it is necessary to understand and model radiological transport in surface ecosystems in a timeframe of several thousands of years. The premises for the long-term development of surface ecosystems are controlled to a great extent by the climate and the distribution between land and water. The development of the Holocene climate is reviewed by (Holmgren and Karlén, 1998) and a model for the shoreline displacement which affects the distribution of land has been described in (Påsse, 1997). The climate and shoreline displacement in the threshold areas probably affects the salinity of the Baltic Sea. The salinity influences which ecosystems will dominate in the coastal area and what property radionuclides have. Salinity is also an important boundary condition for the transport models in the geosphere. Knowledge about the past salinity is important background to evaluate the hydrology and geochemistry in the rock.

In this report the past changes of salinity during Holocene are described and modelled. This will then be extended to a model of the future development of the salinity in the Baltic Sea.

The Baltic Sea has experienced four major stages with altering freshwater and brackish-water during the lasting interglacial (the Holocene; 11,500 years BP to present) namely; the Baltic Ice Lake, the Yoldia Sea, the Ancylus Lake and the Litorina Sea *sensu lato* (in a wide sense). The Litorina Sea stage has further been divided into four substages: the Mastogloia Sea, the Litorina proper, the Limnea Sea and the present-day Mya Sea stage.

This report focuses on the causes for salinity changes within the Litorina Sea stage (8,500 years BP to present) in order to test and calibrate a mathematical model for the circulation and salinity exchange between the Baltic Sea and the Kattegat (Gustafsson, 1997).

The salinity changes in the Litorina Sea stage *sensu lato* have been discussed since the turn of the century. Estimates of the changes were mainly based on the occurrences of mollusc findings in raised beach deposits. During the last c. 20 years new methods have been applied (e.g. $\delta^{18}O$ in calcareous fossils and rhodochrosite, and trace-element content in sediments and ostracode shells). Estimates of the maximum salinity vary between 10 and 20‰ for the surface water of the Baltic proper (for references see Chapter 2). The changes have been attributed to changes in, climate, tidal range in the inlets and cross-section areas of the inlets, but no attempt has hitherto been made to correlate and quantify the different factors.

The objectives of the present study were:

- to compile and correlate climate data for northern Europe, palaeosalinity data for the Baltic Sea and shore-displacement data for the inlets for the last 8,500 years.
- to run an oceanographic model for the salt exchange between the Baltic Sea and the Kattegat with 500-year intervals in order to compare model values with proxy data for the Litorina Sea stage.

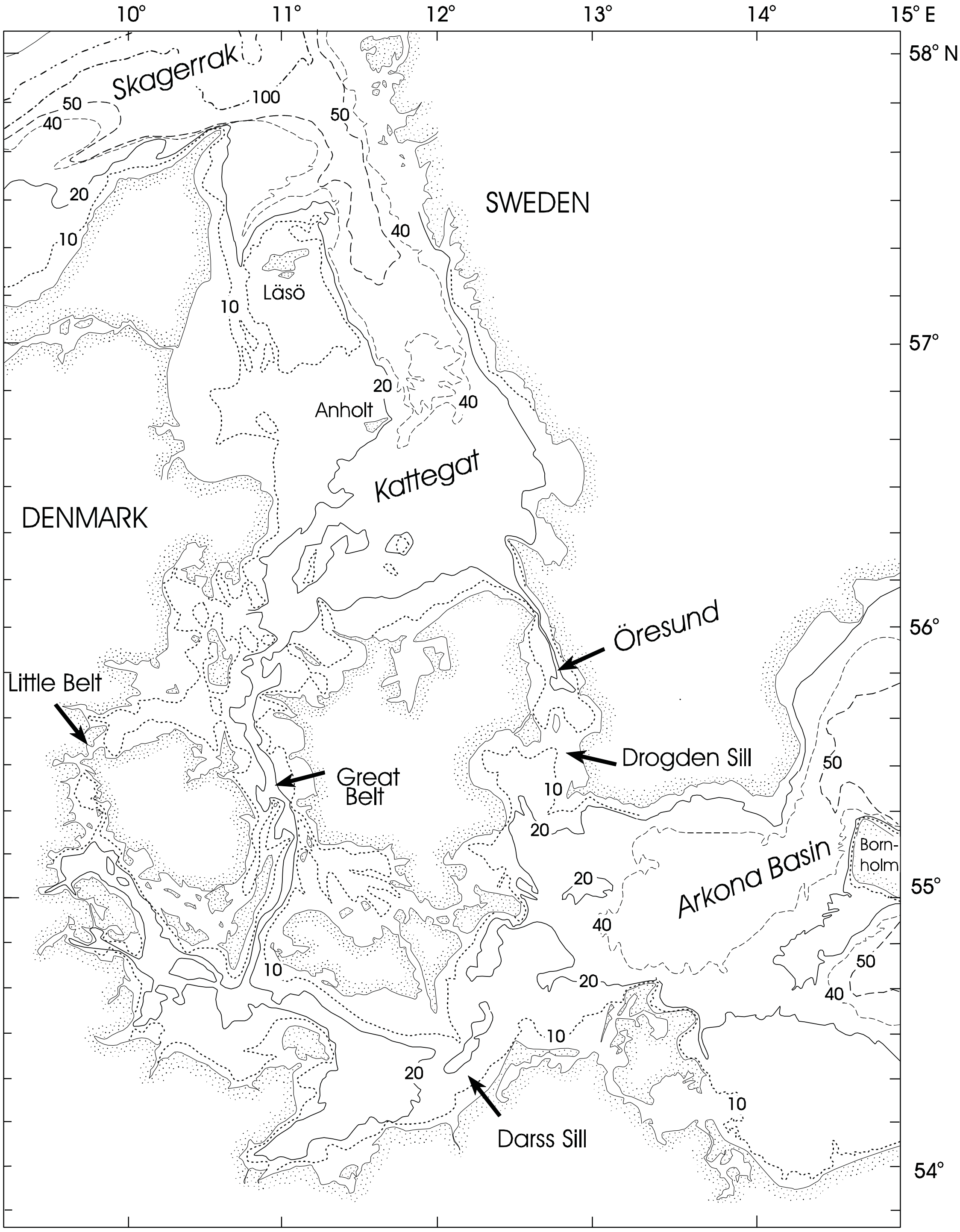


Figure 1-1. *Map showing the inlet area of the Baltic Sea with surroundings.*

2 Oceanography of the Baltic Sea during the Holocene

2.1 Outline of the Holocene history of the Baltic Sea

Finds of faunal elements that are now extinct in the Baltic Sea around the middle to the end of the last century (e.g. Lyell, 1834; Törnebohm, 1862; Lindström, 1886) made geologists aware that the basin had undergone considerable changes in salinity and temperature during its Holocene history. The single most important find was probably that of the mollusc *Ancylus fluviatilis* in raised beach deposits in Estonia by Fredrich W. Schmidt in 1867 (Munthe, 1887). Henrik Munthe made corresponding finds on the island of Gotland, which led him to construct a model for the Holocene history of the Baltic Sea with alternating freshwater and brackish-water stages (Munthe, 1887, 1894, 1902, 1910). These stages are the Baltic Ice Lake (fresh), the Yoldia Sea (brackish), the Ancylus Lake (fresh) and the Litorina Sea (Brackish). Munthe also divided the Litorina Sea into three substages (Munthe, 1894, 1910): the Litorina Sea proper, named after the molluscs *Littorina littorea* and *Littorina saxatilis*, the Limnaea Sea, named after the mollusc *Limnaea ovata*, and the Mya Sea, or the present-day Baltic Sea, named after the mollusc *Mya arenaria*.

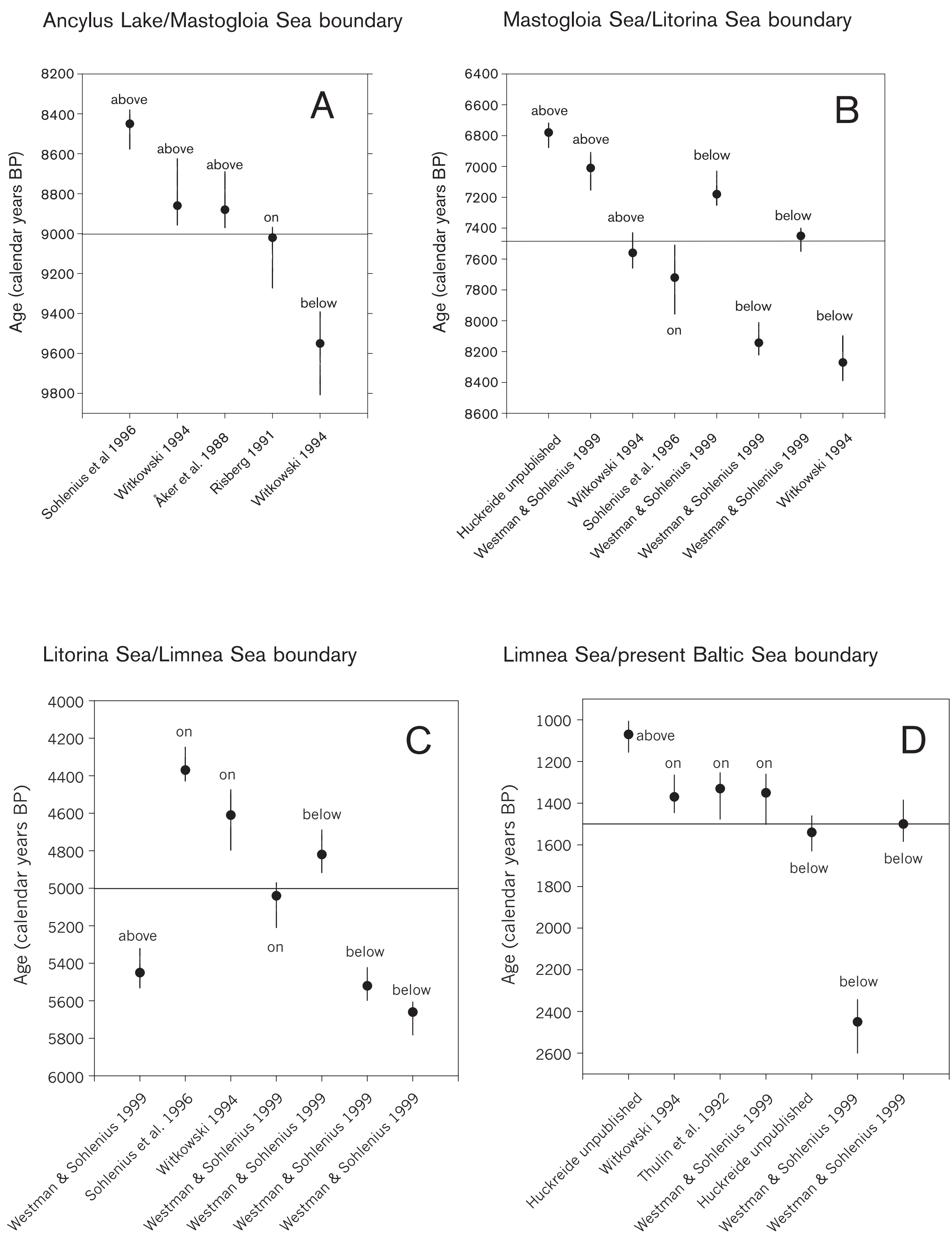
The end of the Ancylus Lake stage has earlier been set at 8,000 ^{14}C (years) BP (see compilation in Hyvärinen et al., 1988), but there are indications that brackish water conditions have prevailed, at least in the southern Baltic, since around 8,500 ^{14}C BP (e.g. Eronen et al., 1990; Winn et al., 1986; Berglund and Sandgren, 1996).

2.2 The Litorina Sea stage (*sensu lato*)

2.2.1 Dating methods (^{14}C)

The ^{14}C activity in the different reservoirs (e.g., atmosphere, ocean and Baltic Sea) has varied over time, which thus has resulted in differences in the initial activity of the samples dated. To overcome this problem calibration with CALIB 4.0 software (Stuvier et al., 1998) has been performed on the offshore material in Figures. 2.1a–d. The marine calibration method has been used and a reservoir effect of 200 years has been assumed (cf. Berglund, 1964, 1971). Dates derived from Witkowski (1994) are assumed to be given as conventional ^{14}C ages and no correction for $\delta^{13}C$ has therefore been performed on them. Datings from Huckriede (unpublished data) were performed on fishbones and have been given a $\delta^{13}C$ of –15‰ according to Stuvier et al. (1998).

The shore-displacement curves from Denmark are mainly based on datings of terrestrial macrofossils and therefore do not need any reservoir correction. They are presented here as curves (Fig. 2.2). The original calibrated data are presented in Christensen (1982, 1993, 1996 and 1998). Original data for the ^{14}C dating of the shore displacement from the German side of the Darss sill area (Hoika, 1972; Ernst, 1974; Klug, 1980) are published only in diagrams without any indications of calibrations. A comparison with Christensen (1998) shows that the dates probably are too young. The shore-



Figures 2-1 a–d. *Dating of the salinity changes in Baltic Sea offshore sediments (except Risberg 1991) of the Baltic Sea. The designations above, on and below refer to the relative position of the dated material to the boundary. The boundaries are defined by changes in diatom composition (Åker et al. 1988, Witkowski 1994, Sohlenius et al. 1996 and Westman and Sohlenius 1999) or changes in $d^{18}O$ (Huckriede unpublished).*

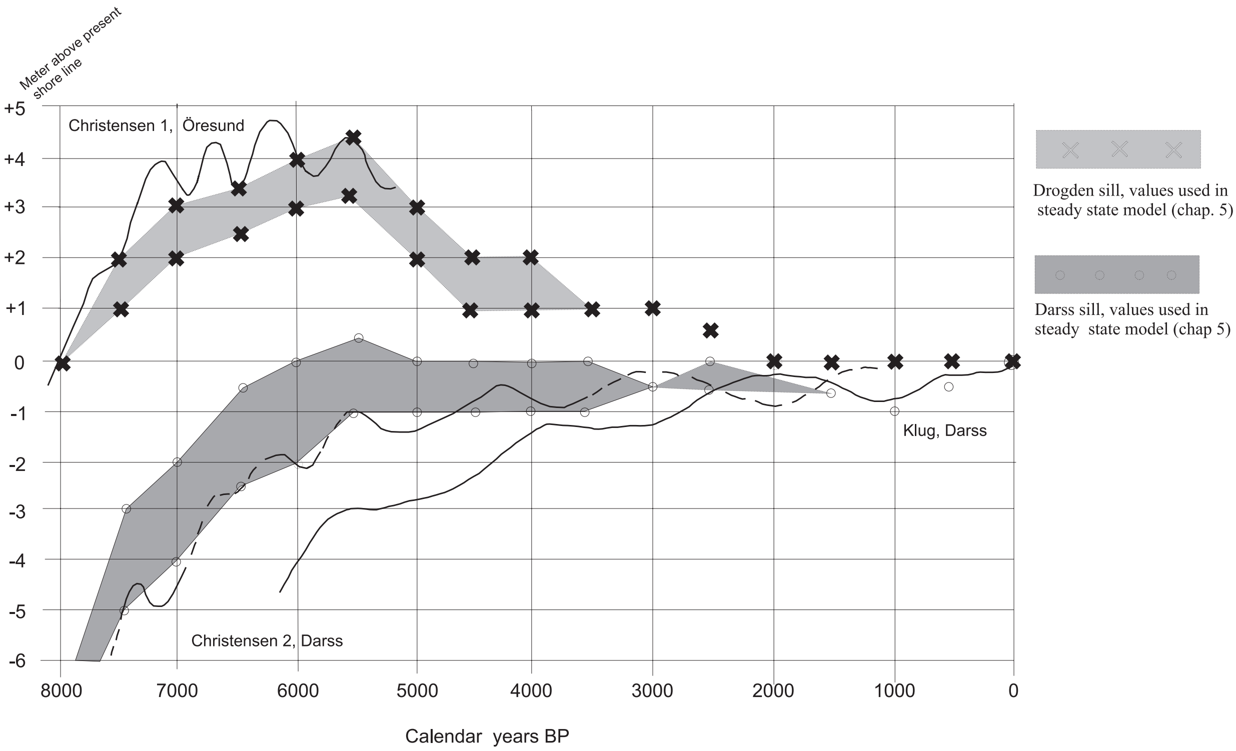


Figure 2-2. *Shore-displacement curves for the Baltic Sea inlet areas (designated: Christensen 1 and 2, and Klug) and intervals of shore displacement used in the steady-state model in chapter 5 (shaded areas, crosses and rings). The curves from Christensen are based on calibrated ^{14}C dates on macrofossils. The curve from Klug is based on uncalibrated ^{14}C dates. The dashed part of Christensen 2 after 3000 BP is based on an 800-year displacement of the curve from Klug (for explanation see text). The values used in the steady-state model (Figs 2.3 and 5.2-5.3) are based on a compilation from; Troels-Smith (1939); Mörner (1969); Wellinder (1972); Hoika (1972); Ernst (1974); Digerfeldt (1975); Klug (1980) and Christensen (1982, 1996 and 1998). Redrawn from Klug (1980) and Christensen (1998).*

displacement curves from the German side of Darss sill have been adjusted by c. 800 years to fit with the macrofossil-dated curve from Denmark (Figure 2.2). This part of Denmark has the same isostatic rebound rate as the southern side of the Darss sill (Christensen, 1998).

2.2.2 Absolute salinity changes

The duration and salinity level of the maximum salinity phase, the Litorina Sea, has been discussed extensively during the last 100 years or so (e.g., De Geer, 1889; Munthe, 1894, 1910, 1931, 1940; Segercrantz, 1896; Sernander, 1911; Segerstråle, 1927; Ekman, 1953; Lundqvist, 1965; Kessel & Raukas, 1979; Miller & Robertsson, 1981; Winn et al., 1986; Punning et al., 1988; Witkowski, 1994; Huckriede et al., 1996 and Tavast, 1996 see also Table 2.1-2.2). The most extensive investigation is still that of Munthe (1910, 1940) on the island of Gotland. He stated that the maximum surface salinity around the island was between 10‰ and 13‰, whereas at present it amounts to between 7‰ and 8‰. Both earlier and later estimates based on fossil assemblages nevertheless point to a maximum salinity of up to 20‰ in the surface water of the Baltic proper (e.g., De Geer, 1889; Ekman, 1953; Witkowski, 1994). On the other hand, recent estimates based on $\delta^{18}O$ measurements of molluscs (Punning et al., 1988), foraminifera (Winn et al., 1986) and rhodochrosite (Huckriede et al., 1996) support Munthe's opinion (Table 2.2). Recent estimates based on the occurrence of Cyanobacterial blooms during the entire Litorina Sea phase (Bianchi et al., 1998) even suggest that the salinity in the Baltic proper did not exceed 12‰. The sills separating the various sub-basins of the Baltic Sea lay much deeper during the early Litorina Sea stage, and this probably resulted in a much less pronounced salinity gradient than at present. The surface salinity in the inner parts of the Gulf of Finland and Gulf of Bothnia was probably more than twice that recorded at present (Munthe, 1894; Segercrantz, 1896; Fromm, 1965; Sjöberg et al., 1984)

Table 2.1: Duration in ^{14}C years BP of the substages of the Litorina Sea *sensu lato*. Note that the dates are from different areas of the Baltic Sea, which especially affects the timing of the onset of the "Mastogloia" and the "Litorina *sensu stricto*" stages. For compilation of older studies see Miller & Robertsson 1979 and Hyvärinen *et al.* 1988.

"Mastogloia Sea stage"	Salinity maximum ("Litorina Sea stage sensu stricto")	"Limnea Sea stage"	Present Baltic Sea	Area	Reference
–	5,500–?	–	–	Bothnian Sea	G. Lundqvist 1963
7,600–7,200	7,200–4,200	4,200–2,500/	2,500/1,500–	Estonian coast	Kessel & Raukas 1979
–	7,000–?	–	–	Bothnian Sea	Miller & Robertsson 1979
–	?–3,000	3,000–1,500	1,500–	Stockholm area	Miller & Robertsson 1981
8,500/8,000–7,000	7,000–4,000	4,000–1,500	1,500–	Compilation	Hyvärinen et al. 1988
–	7,000–3,700	–	–	Gulf of Gdansk	Witkowski 1994
–	?–5,000	–	–	Compilation	Karlsson et al. 1996
–	–	4,000–1,500	1,500–	Estonian coast	Tavast 1996
8,800?/8,200–7,000	6,000–4,000	–	–	Southern Baltic	Berglund & Sandgren 1996

Table 2.2: Maximum salinity of the Litorina Sea stage by comparison with present-day values.

Water mass	Maximum salinity ‰	Present salinity ‰	References
Surface water, Baltic proper	10-15	6-8	Munthe 1894, 1910
Gulf of Finland, Viborg	*c.* 8	3	Segercrantz 1896
Baltic Sea	2 times present		Ekman 1953
Estonian coast	8-15	6-7	Kessel & Raukas 1979
Bothnian Bay, bottom water	*c.* 13*	4	Sjöberg et al. 1984
Kiel Bay	24 ± 2	20	Winn et al. 1986
West coast of Estonia	9-11	6-7	Punning et al. 1988
Baltic proper	15-20	7-8	Hyvärinen et al. 1988
Gdansk deep	>15	7-8	Witkowski 1994
West coast of Estonia	8-10	6-7	Tavast 1996
Gotland deep, bottom water	*c.* 18	12	Huckriede et al. 1996
Baltic proper	< 12‰	7-8	Bianchi et al. 1998

*Sjöberg *et al.* 1984 assume that a more pronounced halocline was present in the Bothnian Bay during the salinity maximum, and they give a probable surface salinity value of 8-10‰, which is in accordance with mollusc data from the area (Munthe 1894; Fromm 1965).

2.2.3 Absolute dating of salinity change

Datings of salinity changes during the Litorina Sea stage have been based mainly on interpolations of shore-displacement curves (Munthe, 1910, 1940; Kessel and Raukas, 1979; Hyvärinen et al., 1988; Punning et al., 1988 and Table 2.1).

Here we present calibrated ^{14}C dates (denoted as years BP) of the changes as indicated by changes in fossil assemblages and $\delta^{18}O$ in offshore material (Fig. 2.1a-d). Most datings are performed just above and below the boundaries to ensure that long-lasting hiatuses do not bias the result. The Ancylus Lake /Mastogloia Sea boundary is difficult to date in offshore material as the organic carbon content is generally low. Therefore we have included a dating performed in a lake basin (Risberg, 1991). There are several datings of the Ancylus Lake/Mastogloia Sea and Mastogloia/Litorina Sea boundaries in the Belt strait and Mecklenburg Bay areas (e.g., Eronen et al., 1990; Winn et al., 1986, 1988). They are not included here as they lay outside the Darss sill area and do not necessarily indicate changes within the Baltic Sea proper. Important datings within the Baltic Sea proper has also been left out due to the lack of information on what kind of material that was dated (e.g., Berglund, 1971; Berglund and Sandgren, 1996).

2.2.4 Changes in circulation within the Baltic Sea

The vertical circulation within the Baltic Sea has also changed during the Litorina Sea stage. Two major periods of deep-water anoxia have been identified c. 7,500–5,000 and 2,000–1,500 years BP (Sohlenius et al., 1996; Sohlenius and Westman, 1998; Westman and Sohlenius, 1999). The first period correlates well with the most saline substage (the Litorina Sea *sensu stricto*) and the second with the Limnea/Mya Sea boundary.

2.2.5 Changes in the cross-section areas of the inlets (Darss and Drogden sills)

Due to isostatic rebound and eustatic sea level changes the depth, and hence the cross-section areas of the sounds has changed during time (Figures 2.2 and 2.3).

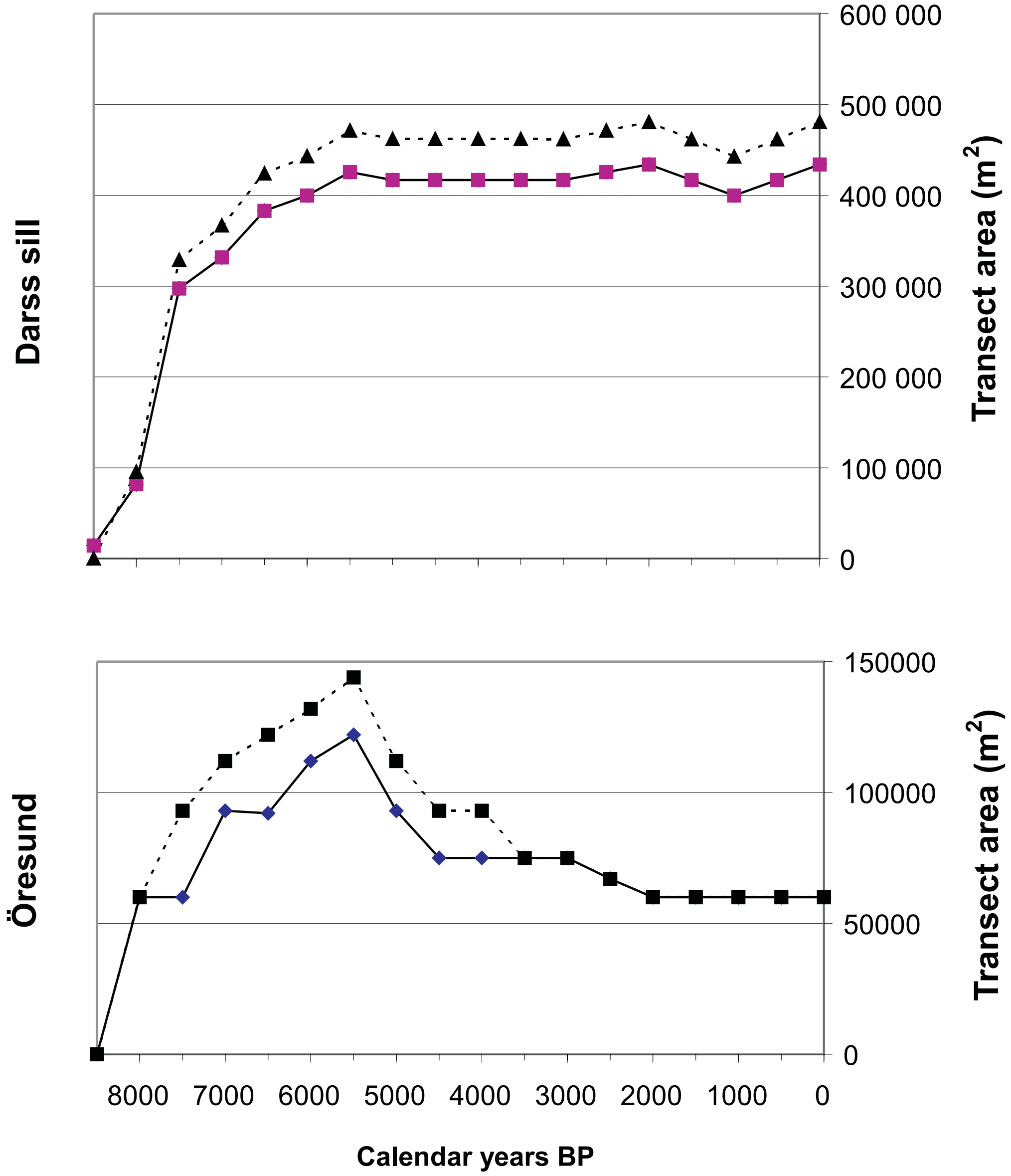


Figure 2-3. *Cross-section area of the Baltic Sea inlet areas from 8500 BP to present. The curve for Darss Sill is based on data from Hoika, 1972; Ernst, 1974; Klug, 1980, and Christensen 1982, 1996, 1998. The calculation of the transect area for the Öresund is based on Christensen 1982 and 1996.*

The Darss sill opened c.9,000–8,500 BP and reached its present depth around 6,000 BP and has since then changed only marginally (see for ref. Klug, 1980 and Christensen 1998).

The Drogden sill opened c. 8,500 BP and reached its present depth for the first time c. 8,000 BP. The depth continued to increase until c. 7,000 BP. Between 7,000 and 5,000 BP the shoreline was about 3.5–5 m above the present level (Troels-Smith, 1939; Mörner, 1969; Wellinder, 1972; Digerfeldt, 1975; Christensen, 1982). According to Digerfeldt (1975) there was a drop in the sea level to about 1 m above the present level 5,000–4,500 BP and the level since then has not reached more than this (Troels-Smith, 1939; Digerfeldt, 1975). Finally c. 2,000 BP the shoreline reached the present level again (Digerfeldt, 1975).

3 Climate conditions in the Baltic region during the Holocene

The climate during the present interglacial period (the Holocene epoch, c. 11,500 calendar years BP to present) is often regarded as extremely stable, especially when compared to earlier interglacials. This view is confirmed by the records of stable oxygen isotopes and other parameters in the icecores from the summit of the Greenland inland ice, GRIP and GISP2 (e.g., Johnsen et al., 1992; Alley et al., 1997), which show only minor fluctuations throughout the Holocene. In these records only one major cold event dated to ca 8,200 cal. years BP has been recorded. This event has also been recorded in marine and terrestrial records from northern Europe (e.g. Alley et al., 1997; Klitgaard-Kristensen et al., 1998; Snowball et al., 1999).

The transition from the last glacial, the Weichselian to the Holocene at c. 11,500 cal. years BP (or ca 10,000 ^{14}C years BP) was an abrupt event with an increase in air temperatures of about 7° in only 50 years (Dansgaard et al., 1989). Full interglacial conditions were reached after a short cold phase in the early Holocene at ca. 11,200–11,050 BP (e.g. Knudsen et al., 1996; Björck et al., 1997). The second part of the Preboreal and the beginning of the Boreal chronozone was characterised by a warm and dry climate with decreasing water levels in lakes in southern Sweden (e.g., Digerfeldt, 1988; Yu & Harrison, 1995). More humid and warm conditions with temperatures about 2–3° warmer than today are reflected in the Atlantic chronozone c. 7,500–4,500 BP. This warm period is often referred to as the Postglacial climatic optimum. Recently published borehole temperatures directly measured in the GRIP and Dye3 boreholes show a long period with temperatures ca 2–3º warmer than today between ca 8,000 and 4,500 BP (Dahl-Jensen et al., 1998; Figure 3.1).

Less stable conditions with decreasing temperatures and increasing precipitation are recorded from ca 4500 BP to present (the Subboreal and Subatlanticum). This pattern is also recorded in the GRIP borehole data (Dahl-Jensen et al., 1998).

The causes for the climate instability during the Holocene are not fully understood, but the general opinion is that variations in the solar output play a major part (e.g. Magny, 1993; Karlén & Kuylenstierna, 1996), probably in connection with other parameters such as increased volcanic activity.

Holocene climatic variability has been the focus of intense research interest during the last few years. Climate archives that have been studied include e.g. peat bogs, lake levels, speleothems and biological parameters. Lake-level fluctuations (e.g. Digerfeldt, 1988) and glacier fluctuations (e.g. Denton and Karlén, 1973) have been studied for a long time in Sweden . Studies of variations in tree-line limits in the Swedish mountains have also provided information about the climate during the Holocene (Karlén and Kuylenstierna, 1996).

A number of parameters showing periods with increasing precipitation and/or colder climate around the North Atlantic are shown in Figure 3.2. The diagram also shows transgressive phases in the Baltic Sea during the Litorina Sea stage (e.g. Risberg, 1991). All ages are given in calendar years BP, and radiocarbon datings have been calibrated according to CALIB 4.0 (Stuiver et al., 1998). Several records indicate synchronicity or close to synchronicity, although the dating control varies among the different methods.

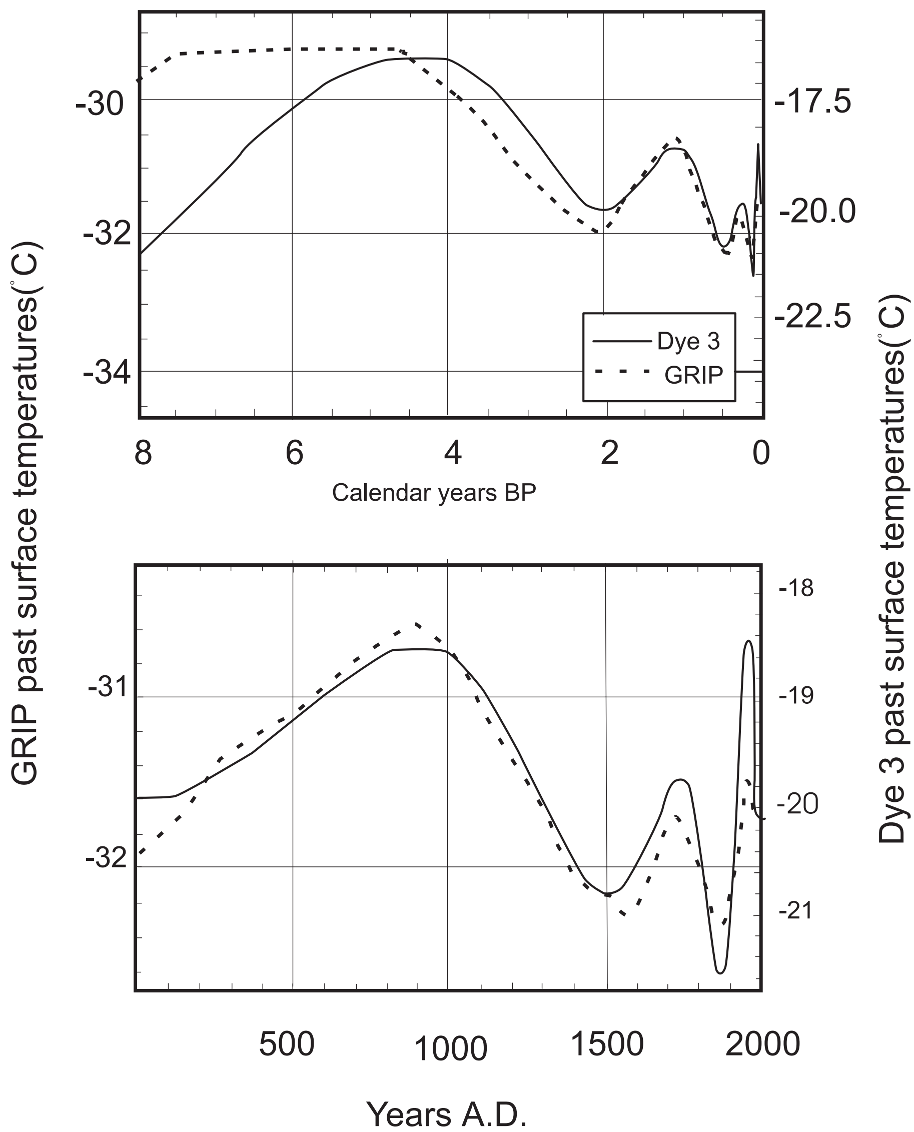


Figure 3-1. *Reconstructed temperatures for Greenland ice cores Dye3 and GRIP. (A) The temperature from 8000 BP to present; (B) the temperature from AD 0-2000 (after Dahl-Jensen et al., 1998). The observed difference in amplitude between the two cores is a result of their different geographic location in relation to the variability in atmospheric circulation.*

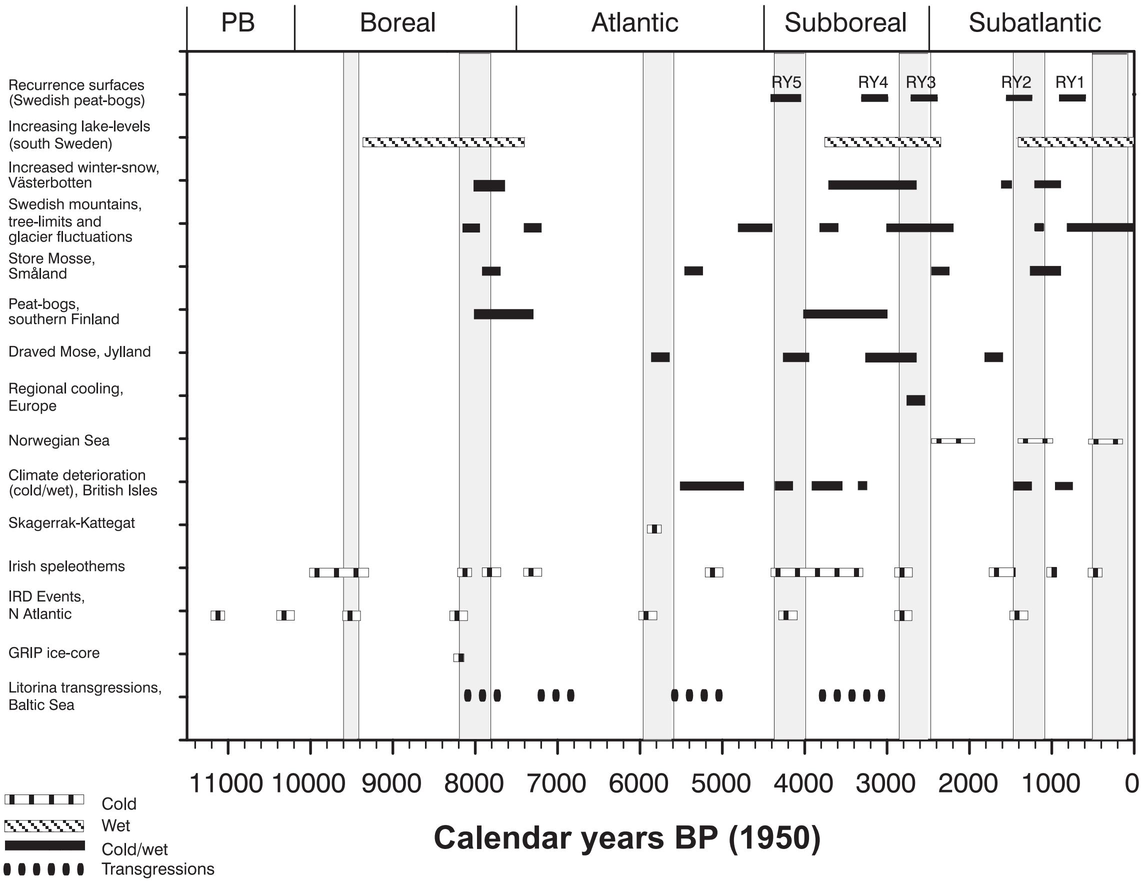


__Figure 3-2.__ Periods with increasing precipitation and/or colder climate around the North Atlantic during the Holocene (11,500-0 cal y BP). Data from peat bogs in middle and south Sweden (Granlund, 1932; Svensson, 1988 for Store Mosse); lake-level fluctuations in south Sweden (e.g. Digerfeldt, 1988; Yu and Harrison, 1995); laminated lake sediments (increased winter snow; Snowball et al., 1999); Swedish mountains (tree-limit fluctuations and glacier fluctuations; e.g. Karlén and Kuylenstierna, 1996; Karlén, 1998); peat bogs in Finland (e.g. Korhola, 1994, 1995); Draved Mose, Denmark (e.g. Aaby & Tauber, 1974); European continent (e.g. van Geel et al., 1996); Norwegian Sea (e.g. Koc et al., 1999); British Isles (e.g. Blackford & Chambers, 1995; Anderson et al., 1998; Caseldine et al., 1998); Skagerrak-Kattegat (Jiang et al., 1997); Irish speleothems (McDermott et al., 1999); GRIP ice core (e.g. Johnsen et al., 1992); North Atlantic (IRD events; Bond et al., 1997); and transgressions in the Baltic Sea during the Litorina Sea stage (dotted lines; e.g. Risberg, 1991). Data given in radiocarbon years BP have been calibrated to calendar years using CALIB 4.0 (Stuiver et al., 1998).

Some variations that seem to be out of phase may be synchronous, but different dating methods, reservoir effects, andvariations in $\delta^{14}C$ may give discrepancies in ages of several hundred years in the ages. Cold and possibly wet phases are recorded around c. 9,600–9,400, 8,200–7,800, 6,000–5,600, 4,400–4,000, 2,900–2,500, 1,500–1,100 and 500–100 BP. The last three of these are probably the best documented, both from historic records and from palaeoecological investigations. A long period with a high degree of variation is seen during the Subboreal at c. 5,000-2,400 BP. In contrast, the period between c. 2,400 and 1,600 BP is characterised by a relatively warm and stable climate. The transgressions in the Baltic Sea during the Litorina Sea stage at c. 8,200–7,700 (L1), 7,300–6,800 (L2), 5,600–5,000 (L3) and 3,600–3,000 (L4) BP seem to occur during periods with few observed climatic deteriorations, which suggests that these high levels of the Baltic Sea occur during warm phases. The reasons behind the transgressions during the Litorina Sea stage are not fully understood, but it is reasonable to assume that the climate plays an important role in connection with changes in the inlet areas.

Peat bogs offer a source of information about the climate during the Holocene. Peat humification and other parameters were studied by Granlund (1932). Granlund described five so-called recurrence surfaces (in Swedish rekurrensytor = RY) in peat bogs in southern and middle Sweden. These appear to indicate changes in former precipitation and can be seen in the peat bogs as a change in peat humification from high to low. The recurrence surfaces were numbered from RYV to RYI and were dated to c. 4,250, 3,150, 2,550, 1,550 and 750 BP. Although the datings are uncertain, a correlation with other data showing cold and wet conditions seems possible (cf. Figure 3-2).

In southern Finland, two phases of more intensive peat initiation were recognised at ca 8,000–7,300 cal y BP and between 4,300 and 3,000 BP (Korhola, 1994, 1995). These phases have been correlated with periods with increasing lake levels in southern Sweden (Digerfeldt, 1988; Yu & Harrison, 1995).

Fluctuating lake levels have been recorded in many lakes in southern Sweden during the Holocene (e.g. Digerfeldt, 1988). The reasons for this may be either increasing precipitation, a decrease in evaporation or a combination of these. A distinct lowering in lake level culminated at c. 10,700–10,300 cal y BP, followed by increasing levels throughout the Boreal and early Atlantic chronozones (Digerfeldt, 1988). A long period of increased dryness began at c. 7,500 BP followed by a number of lake-level fluctuations suggesting a fluctuating climate until c. 2,700 cal y BP. Another lowering is recorded between c. 1,700 and 1,300 cal y BP. It is plausible that some of these changes are related to changes in the land-sea geography of the southern Baltic area. During the early Litorina Sea Stage, when the Baltic Sea was considerably larger than today, the summer contrast between a large, cool water body and the warmer land is likely to have favoured blocking of anticyclones (Yu & Harrison, 1995).

4 Oceanography of the present Baltic Sea

4.1 Hydrographic conditions

4.1.1 The Baltic Sea

The Baltic Sea can be regarded as a fjord estuary as the parallel sills in the Danish sounds are very much shallower than both the average depth and the maximum depth. The river runoff plus precipitation strongly exceeds evaporation. The river runoff is rather accurately known; the average river discharge to the Baltic Sea from1950–1990 was 14,151 m^3/s plus an additional 1,159 m^3/s to the Belt Sea and Kattegat according to Bergström & Carlsson (1994). However, there are uncertainties about both precipitation on and evaporation from the Baltic Sea. An investigation by Omstedt et al. (1997) points to a long-term average precipitation minus evaporation of some 2,000 m^3/s.

Because of the large freshwater supply and the limitation of the water exchange,due to shallow sills and channels with large frictional resistance in the Öresund and the Belt Sea, the surface salinity is only 7‰ to 8‰ in the Baltic proper and even less in Bothnian Sea, Bothnian Bay and Gulf of Riga. The halocline of the Baltic proper is located at some 60–80 m depth and below this the salinity increases to 11‰ to13‰, see e.g. Samuelsson (1996) for a discussion on the salinity variations during the last 40 years. The source of salt for the Baltic Sea is inflows of salty water through the Danish Sounds and since the halocline is much deeper than the sill depth there is no possibility of the Baltic deep water returning directly through the Danish Sounds. Due to the long residence time in the deep water in combination with the high rate of deposition of organic material, the supplied oxygen is rapidly consumed. Long periods of anoxic conditions are common below 150 m (e.g. Fonselius, 1969 and Stigebrandt, 1995). However, at depths above about 120 m the water exchange is more continuous and not restricted to extreme events of high saline inflows, as it is deeper down in the water column, and anoxic conditions very seldom reach as high as 120 m (Stigebrandt and Wulff, 1987).

4.1.2 The Danish sounds

As stated above, the water exchange through the Danish Sounds is very important for the hydrographic conditions in the Baltic Sea. The net water flow through the Sounds is equal to the net freshwater supply to the Baltic drainage basin, i.e., about 16,000 m^3/s. However, the instantaneous flows across the sills are an order of magnitude greater. The variability of the flows is of crucial importance for the effective salt transport across the sills. The Belt Sea and Öresund are all usually very strongly salt stratified because of the large supply of low saline water from the Baltic Sea and supply of highly saline waters from the Kattegat for the lower layers (see Figure 4.1). The momentary conditions in the vicinity of the sills are strongly dependent on the direction, magnitude and duration of the flow. During outflows from the Baltic the low saline water will propagate northward through the Öresund and the Belt Sea stabilising the stratification. Eventually the fronts will reach the openings of the mouths of the Great Belt and the Öresund and enter into the open Kattegat. The mixing in the Belt Sea is greater than in the Öresund so the salinity of the water reaching the Kattegat has increased to at least 14‰ for flow through the former compared to ca. 10‰ for the latter strait (e.g. Pedersen, 1993).

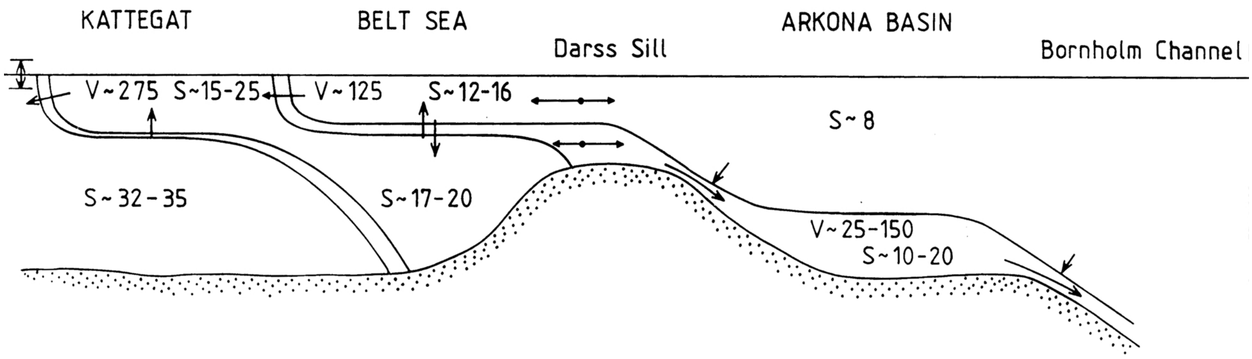


Figure 4-1. Conceptual picture of the stratification in the Kattegat, Belt Sea and Arkona Basin showing typical volumes V *(in km³) and salinities* S *of the different layers. From Stigebrandt (1995).*

4.1.3 The Kattegat

The hydrography of Kattegat is characterised by a very sharp halocline at a depth of about 15 m depth dividing the surface mixed layer with salinity 15‰ to 25 ‰ from the deep water with salinity 32‰ to 35 ‰ (see, e.g., Figure 4-1; Svansson, 1975). The deep-water layers are filled in the north with Skagerrak water that is successively emptied by wind-forced vertical entrainment into the surface layer. The flow of surface water from the Kattegat to the Skagerrak mainly occurs in the salinity interval 23‰ to 27 ‰ (Andersson and Rydberg, 1993).

The Kattegat-Skagerrak front puts a clear marking line between the seas. The position of the front is somewhat variable, but it appears that the most common position is for it to take off at Skagen heading to the northeast (e.g., Gustafsson and Stigebrandt, 1996). It is seldom found south of the island Läsö (Andersson and Rydberg, 1993), but may occasionally expand far into Skagerrak during northeasterly winds (Aure and Sætre, 1981). In general, frontal dynamics force the outflow from the surface layer of the Kattegat to enter the Skagerrak along the Swedish coast.

4.2 Review of process studies

4.2.1 Basic concepts of the water exchange

The strong stratification connected to the front in the surface layer of the northern Kattegat constitutes an effective protection of the Kattegat surface waters from direct influence from the Skagerrak. The properties with respect to salinity, temperature and any other state variables north of the front will be the properties imported to the deep water of the Kattegat. The coupling between the Baltic Sea and the Kattegat/Belt Sea is very tight and it is not possible to find a section closer to the Baltic proper where a simple decoupling is possible. Thus, the existence of the Kattegat-Skagerrak front is of utmost importance for the Baltic Sea system and has to be included in every model of the long-term water exchange of the Baltic Sea. The general assumption is that a geostrophic balance (a balance between the pressure gradient and the Coriolis force; see Engqvist, 1997) as suggested by Stigebrandt (1983) maintains the front although an assumption of a non-rotating hydraulic control gave good results in the model of Pedersen & Møller (1981). Jakobsen (1997) has shown that the geostrophic assumption is consistent with observations. This assumption has been used in time-dependent models not only by Stigebrandt (1983) but also by Omstedt (1987 and 1990) and Gustafsson (2000a and 2000b), and in a steady-state model by Gustafsson (1997).

Effects on the vertical stratification of the Baltic Sea

On timescales of less than a few years, the Baltic Sea surface salinity is almost independent of the salt-flow variations through the Danish Sounds due to the long residence time. The variations are instead due to mixing and variations of freshwater supply over the year, e.g., Stigebrandt (1985) and Eilola (1997). Heating-cooling of the surface layer of the Baltic Sea is also of great importance for the near-surface stratification.

The vertical circulation of the Baltic is maintained by an advection-diffusion balance, where the source for advection is dense gravity currents (i.e., currents of dense water following the bottom steeps, see e.g., Stigebrandt, 1987b), which emerge from the overflow of saline waters across the sills of the Danish Sounds. In order to explain the variability of the dense gravity currents of the Baltic Sea, some additional information about the Kattegat/Belt Sea must be included.

The flow through the Danish sounds depends essentially on the sea-level difference between the Kattegat and the Baltic Sea and not on the stratification in the area, i.e., the flow is barotropic (see e.g., Engqvist, 1997). However, the correlation between the barotropic flow and the salinity at the sills is very strong. For the Öresund, it has been shown that most of the variance in the surface salinity is due to frontal movements (Mattsson, 1996b); that is, when the barotropic flow is directed towards the Kattegat, low-saline water from the Baltic will move northward. During southward flow the front will move back, but also some water from below the sill depth will also cross the sill. The salinity at the sill will increase with time and the strength of the flow as the surface layer is emptied. The water and salt flow across the Drogden sill in the Öresund has been measured directly during the latest years, see Jakobsen and Castejon (1995) and Jakobsen and Lintrup (1996).

At the Darss sill a two-layer stratification is usually present which makes the dynamics of the water exchange across this sill somewhat different compared to that across the Drogden sill. Thus, there exists a more or less continuous flow of high-saline water from below the halocline of the Belt Sea to the Arkona Basin. Most probably the flow is regulated by geostrophy (Stigebrandt, 1983; Lass et al., 1987). Further, it appears that vertical mixing is a much more important factor in determining the stratification of the Belt Sea than it is in the Öresund north of the sill. However, fronts are also prominent in the Belt Sea, and the duration and strength of an inflow to the Baltic will affect the salinity of the inflowing water as can easily be understood from a study of Figure 4-1. If an inflow carries very large amounts of water the stratification of the Kattegat will become important as the surface layers of the Öresund and Belt Sea are emptied. Therefore, to model these processes one needs a time-dependent model that resolves the horizontal movements of fronts and describes the mixing processes correctly.

After crossing the sills, high-saline inflowing water will sink and form a dense pool at the bottom of the Arkona Basin (see Figure 4.1). In Stigebrandt (1987a) it was argued that the flow from this pool into the Bornholm Channel is determined by a geostrophic control in the Arkona Basin. From this assumption it was possible to reconstruct the statistical properties of the upstream condition of the gravity current using salinity observations in the Arkona Basin. The hypothesis has been successfully used in models for the Baltic Sea circulation (e.g. Stigebrandt, 1987b; Omstedt, 1987, 1990). Only very recently a comprehensive set of field observations confirmed the hypothesis of a geostrophic control (Liljebladh and Stigebrandt, 1996). The statistics presented by Stigebrandt (1987a) show that the dense gravity current into the Baltic is almost always present, but that events with large inflows are extremely rare, as also shown by the direct observations of the deep-water flow in the Bornholm Channel presented by Walin

(1981). The statistics of rare events have been analysed by Matthäus and Franck (1992) using data from the
Darss sill. However, since the high-saline water that flows across the sills is stored in the Arkona Basin for a considerable time, it may be exposed to strong wind mixing entraining it into the surface layer and by that be preventing it from forming new deep water in the Baltic proper.

The major inflow to the Baltic in 1993 is the first inflow for which it is possible to quantify the transports by direct measurement and it has been subject to several thorough investigations, Håkansson et al. (1993), Matthäus et al. (1993), Huber et al. (1994), Jakobsen (1995), Matthäus and Lass (1995) and Liljebladh and Stigebrandt (1996).

The effect of a gravity current is twofold; it transports saline water down to depths greater than the sill depth and it entrains water from the ambient stratification. The latter process results in significant mixing of the deep waters in the Baltic Sea (e.g., Kõuts and Omstedt, 1993). Axell (1998) has recently presented a study on the deep-water mixing in the Baltic Sea based on historical data. The vertical diffusivity in a deep basin is probably due to breaking internal-waves (cf. Stigebrandt, 1987b) although the main forcing of the internal-wave activity in the ocean, i.e., tides, is almost absent in the Baltic Sea. Thus the most probable forcing of the internal-wave activity is the wind, but the mechanisms of energy transfer from wind forcing to internal waves and subsequently to turbulence and diffusion are not yet understood.

4.3 Review of sensitivity studies

4.3.1 Climate change impact on salt exchange

Stigebrandt (1983) made prognostic simulations of the water and salt exchange in the Baltic entrance area in a manner that was suitable for climate impact analysis. His time-dependent model reproduced the stratification of the Kattegat and the Belt Sea quite accurately as a function of observed sea level, wind stress and freshwater supply. The model featured several new insights into the dynamics of the entrance area. The most fundamental was the idea that both the outflow from the surface layer of the Kattegat and from the lower layer of the Belt Sea into the Baltic are in geostrophic balance. This dynamic constraint at the Kattegat-Skagerrak front, which later was confirmed observationally by Jakobsen (1997), has been shown to be a key factor of in the whole Baltic Sea water exchange. Further, Stigebrandt (1983) presented functional dependencies between the Baltic Sea surface salinity and climate variations in freshwater supply, local wind stress and Skagerrak salinity. Stigebrandt's model concept was later refined and extended by Omstedt (1987, 1990), Omstedt and Nyberg (1996), Omstedt and Axell (1998) and Gustafsson (1999a, 1999b). His findings concerning the sensitivity of Baltic Sea salinity have been confirmed by Gustafsson (1997, 2000b). The sensitivity of the Baltic Sea surface salinity to changes in freshwater inflow, wind, and sea-level fluctuations in the Kattegat are illustrated in Figure 5-1. As can be seen from the figure a 20% change in the freshwater inflow has a strong impact on the surface salinity. Similar results have also been discussed by Omstedt et al. (1997) in relation to estimating the net precipitation over the Baltic Sea.

The efforts to numerically model the area have increased rapidly during the last few years, and include several investigations using three-dimensional models, for example, Winkel-Steinberg et al. (1991), Huber et al. (1994), Lehmann (1995), Svendsen et al.

(1996), Sayin and Krauss (1996) and Schrum and Backhaus (1999). One problem with the grid models of the area is the enormous computational effort needed to resolve the short lateral scales of the baroclinic currents and straits. We can, however, expect more progress in the field as computer capacity increases and new, more physically correct, parameterisations of unresolved processes are developed.

4.3.2 Climate change impact on the vertical stratification

The vertical structure of the Baltic Sea salinity is strongly dependent on inflow on saline-water inflow, freshwater inflow and turbulent mixing. Climate change in any of these factors can thus have a large impact on the vertical stratification. Under present climate conditions the Baltic Sea has a typical halocline depth of 60 m. There are however many processes that could change this fact. For example with an increased inflow of ocean water the halocline depth could easily decreased, forcing the Baltic Sea towards conditions similar to the Skagerrak or the Hudson Bay, with a thin brackish layer over a deeper saline layer. In the other extreme with very reduced inflows, the Baltic Sea could become a lake with negligible salinities.

Stigebrandt (1987b) has studied some other important aspects of climate change impacts. He showed that the vertical salinity profile should become quite different if the inflow were constant instead of being variable, as it is during the present climate conditions, see Figure 4-2.

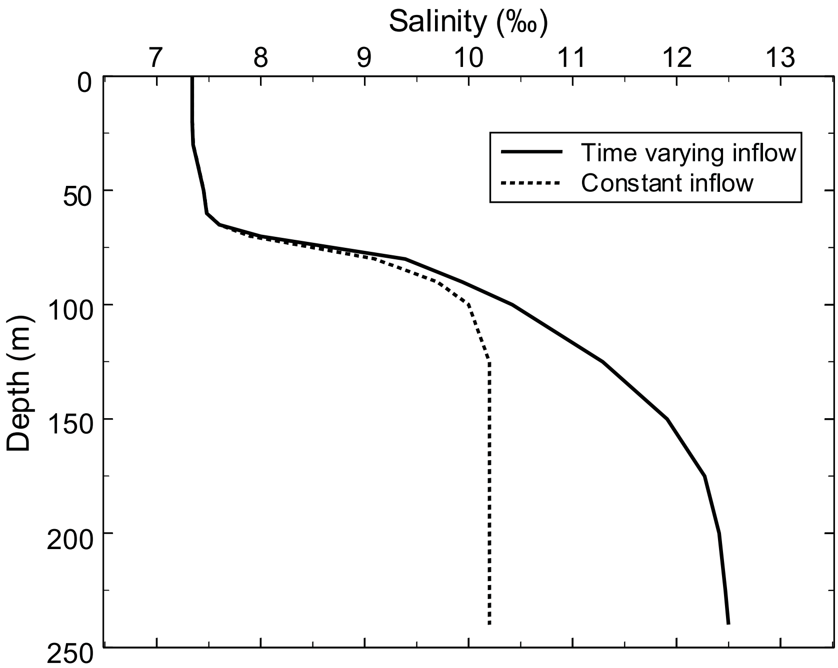


Figure 4-2. *Simulated Baltic proper vertical salinity profiles for time-varying and constant inflow of saline water from the Arkona. From Stigebrandt (1987b).*

4.3.3 Dominating timescales

The coupling between the atmosphere and the sea-surface properties, as temperatures and ice, is strong and therefore the atmospheric timescale governs. This has been illustrated in several investigations. Omstedt and Chen (1999) analysed the interannual variation of maximum ice extent and found a very low auto-correlation with a lag of only 1 year. Leppäranta and Seinä (1985) analysed long-term trends in ice thickness along the Finnish coast and found that the consecutive winters were practically independent of each other. Omstedt and Rutgersson (1999) modelled the heat cycle of the Baltic Sea. They showed that the Baltic Sea responds thermodynamically as a lake and that the large interannual variations of the atmosphere forcing shown were mainly balanced by the annual change in the total heat content of the water body and only very little heat exchange occurred across the sills.

The response time of the salinity of the Baltic Sea to changes of the inflow should be comparable to the flushing time. The flushing time is defined as the volume of the Baltic Sea, V, divided by the sum of all inflows:

$$\tau \equiv \frac{V}{Q_f + (P - E)A_S + Q_{in}} \qquad (4:1)$$

Where Q_f is the river runoff, P and E are the precipitation and evaporation rates respectively, A_s the surface area of the Baltic Sea and Q_{in} is the inflow from across the sills. Using salinities typical for inflows and outflows across the sills one can deduce that the inflow is about equal to the sum of river runoff and net precipitation. Thus, the flushing time can be estimated according to:

$$\tau = \frac{V}{2(Q_f + (P - E)A_S)} \qquad (4:2)$$

For values according to present climate condition the timescale is typically 20 years, but as can seen from the formula the flushing time can easily change if the inflows or the volume of the system change.

The stratification spin-up time for the Baltic proper, was examined by Omstedt and Axell, (1998) by modelling the so-called lock-exchange experiment. This experiment examines the exchange between two basins with different salinities after the lock between the basins has been opened. Starting from the initial condition that the Baltic Sea was almost fresh (0.5‰) and that the Kattegat was saline (34‰), the transient development of stratification was calculated. The results are illustrated in Fig. 4.3. The saline Kattegat water enters rapidly into the Baltic Sea as a dense bottom current. The deep-water mixing then slowly mixes the saline water with the freshwater on top. As the deep-water mixing rate is low, the timescale for reaching surface salinities close to the present climate conditions becomes quite long, or about 100 years.

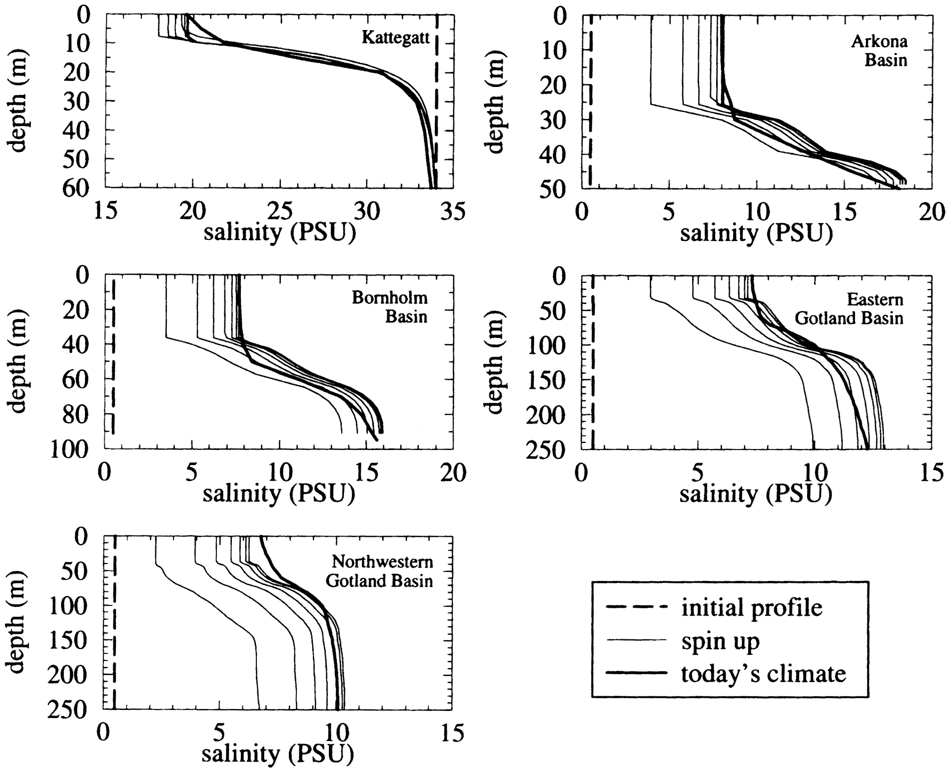


Figure 4-3. *Time evolution from a simulation of the establishment of the stratification of the Baltic Sea. Initial profiles are shown as dashed lines and model results sampled each 15 years up to 105 years after start are shown as thin lines. Also shown are the observed mean profiles (thick lines). From Omstedt and Axell (1998).*

5 Model simulations

5.1 Model formulation

5.1.1 The steady-state salinity model

The timescale for changes in the overall salinity in the Baltic Sea is roughly 25 years. Stigebrandt (1983) used a time-dependent model to examine the sensitivity of the Baltic Sea salinity to changes in the barotropic flow through the Danish Sounds, the freshwater supply to the Baltic Sea and mixing in the Kattegat-Belt Sea. Pedersen and Møller (1981) used a steady-state model of the Baltic Sea and the Baltic entrance area to investigate the sensitivity of the stratification to changes in the freshwater supply. In this section the mechanisms of the long-term salt balance of the Baltic Sea are illustrated with a steady-state model that is even simpler than this. The model was presented in Gustafsson (1997) and the following description closely follows this paper.

In this model we only take into consideration the upper layer of the Kattegat and Belt Sea which are treated as a single water mass (box), while the lower layer is treated as a dynamically passive source of water and salt. In a steady state the volume and salt conservation of the box yields:

$$-Q_G + Q_E + Q_F = 0 \qquad (5.1)$$

$$-SQ_G + S_2Q_E + S_0Q_F + S_0Q_B - SQ_B = 0 \qquad (5.2)$$

Equation (5.1) expresses volume conservation of the Kattegat/Belt Sea box, where Q_G is the geostrophic outflow to the Skagerrak, Q_E is the entrainment flow from the lower layer and Q_F is the net supply from the Baltic Sea (freshwater input). Equation (5.2) expresses the conservation of salt in the upper layer of salinity S, where S_2 is the salinity of the lower layer, S_0 is the salinity of the Baltic, and Q_B is the effective fluctuating flow across the sills. Conservation of salt in the Baltic Sea is also required:

$$S_0Q_F + S_0Q_B - SQ_B = 0 \qquad (5.3)$$

Since both the freshwater supply and the fluctuating flow are due to external forcing. Equation (5.3) gives a relation between S and S_0. If S_0 is eliminated from equation (5.2) and volume conservation, equation (5.1), is used to eliminate the entrainment velocity the following expression is reached:

$$(S_2 - S)Q_G - S_2Q_F = 0 \qquad (5.4)$$

The geostrophic flow and the entrainment velocity, w_E, are calculated from:

$$Q_G = \frac{g\beta(S_2 - S)h^2}{2f} \tag{5.5}$$

$$w_E = \frac{2m_0 u_*^3}{g\beta(S_2 - S)h} \tag{5.6}$$

where h is the depth to the pycnocline in the Kattegat-Belt Sea, f is the Coriolis parameter, g is the constant of gravity, b the expansion coefficient of seawater due to salinity, m_0 is the efficiency of turbulent mixing with respect to work against the buoyancy forces and u_* is the friction velocity.

The hypsographic function of the Kattegat-Belt Sea shows that the horizontal area $A(h)$ decreases approximately linearly with depth and at a depth of about 15.5 m depth the area is 50% of the area at the sea surface. Thus:

$$A(h) = A_0\left(1 - \frac{h}{2h_m}\right) \tag{5.7}$$

where h_m = 15.5 m is the topographical scale height. The entrainment flow is the entrainment velocity times the area at the depth of the pycnocline.

$$Q_E = \frac{2m_0 u_*^3 A_0}{g\beta(S_2 - S)h}\left(1 - \frac{h}{2h_m}\right) \tag{5.8}$$

From equations (5.4) and (5.5), we see that the freshwater height in the Kattegat-Belt Sea:

$$F = \frac{S_2 - S}{S_2}h,$$

is solely determined by the freshwater supply and the lower layer salinity; thus:

$$F = \sqrt{\frac{2fQ_F}{g\beta S_2}} \tag{5.9}$$

This is a direct consequence of the assumption of a geostrophic front as was pointed out in Stigebrandt (1987a). Note that the entrainment velocity, equation (5.6), is inversely proportional to the freshwater height and does therefore only vary as a function of wind stress and freshwater supply and not independently on pycnocline depth and salinity. However, the variation of the area of the pycnocline with depth makes the expression for entrainment flow more complex, see equation (5.8).

If the flow parametrisations are inserted into the volume conservation, equation (5.1), the following expression for the pycnocline depth *h* is derived.

$$h = \frac{h_m(2m_0u_*^3A_0 + \sqrt{2fg\beta S_2 Q_F^3})}{g\beta S_2 h_m Q_F + m_0 u_*^3 A_0} \tag{5.10}$$

The upper-layer salinity is given by

$$S = S_2\left[1 - \frac{g\beta S_2 h_m Q_F + m_0 u_*^3 A_0}{h_m(2m_0u_*^3A_0 + \sqrt{2fg\beta S_2 Q_F^3})}\sqrt{\frac{2fQ_F}{g\beta S_2}}\right] \tag{5.11}$$

Finally the Baltic Sea salinity is given by

$$S_0 = \frac{SQ_B}{Q_F + Q_B} \tag{5.12}$$

or

$$S_0 = \frac{S_2 Q_B}{Q_F + Q_B}\left[1 - \frac{g\beta S_2 h_m Q_F + m_0 u_*^3 A_0}{h_m(2m_0u_*^3A_0 + \sqrt{2fg\beta S_2 Q_F^3})}\sqrt{\frac{2fQ_F}{g\beta S_2}}\right] \tag{5.13}$$

Using the parameter values in Table 5.1, the pycnocline depth becomes 16.9 m, the surface salinity 25.9‰ and the Baltic Sea surface salinity 7.5‰. The geostrophic outflow from the Kattegat becomes approximately 70,000 m^3/s, which is also in close agreement with observations. However, the major drawback of the model is that the effective flow fluctuations, Q_B, have to be set as low as 6,500 m^3/s in order to get the right salinity for the Baltic Sea, which is very much smaller than the actual average of the flow fluctuations. From a simulation of the daily barotropic flow with the channel model by Stigebrandt (1992) for the period 1970–1976, it is found that the average magnitude of the oscillating part of the flow was 43,000 m^3/s. The explanation for the discrepancy is twofold; the lack of a horizontal gradient in salinity causes the salinity of the Belt Sea and Öresund to be highly overestimated and the exclusion of the effect of moving fronts hampers the salt exchange across the sills. It has been demonstrated that the effective flow is about 10,000–14,000 m^3/s (Stigebrandt, 1983). However, as long as the horizontal salinity gradient within the Belt Sea and the Kattegat does not change dramatically, the response to changes in the magnitude of the oscillating flow should be correct to the lowest order. As this is a key issue in the present investigation the oscillating flow and its associated dispersive salt transport will be discussed further in Section 5.1.2.

Note that the fluctuating flow does not have any influence on the long-term average stratification in the Kattegat-Belt Sea; see eqations (5.10) and (5.11). The reason for this is that salt comes into the model layer only through the entrainment of deep water and this is solely wind-driven. However, the fluctuating flow has a prominent influence on the salinity of the Baltic Sea (Figure 5-1). The reason for this is that the net salt flux through the system is required to be zero. Thus, a change of the fluctuating flow will be compensated by the salinity of the Baltic Sea in such a way that the salt flow remains

Table 5-1. Parameter values used to compute the long-term exchange between the Baltic and the Kattegat/Belt Sea

Parameter	Value	Parameter	Value	Parameter	Value
F	$1.2 \cdot 10^{-4}$ s^{-1}	b	$8 \cdot 10^{-4}$	h_m	15.5 m
G	9.8 m s^{-2}	u_*	0.013 m s^{-1}	s_2	33.5 psu
m_0	0.6	A_0	42,000 km^2	Q_F	16,000 m^{3}s^{-1}

zero. Increasing the freshwater supply has a radical influence on the whole system, decreasing the salinity of the Baltic Sea (Figure 5-1) and the pycnocline depth and salinity of the Kattegat-Belt Sea (not shown). Changes of the mixing winds gives a direct influence of the stratification both in the Kattegat-Belt Sea and in the Baltic Sea. However, the absolute change in the salinity of the Baltic Sea is less than in the Kattegat-Belt Sea. Stigebrandt (1983) also drew these conclusions.

5.1.2 The barotropic water exchange

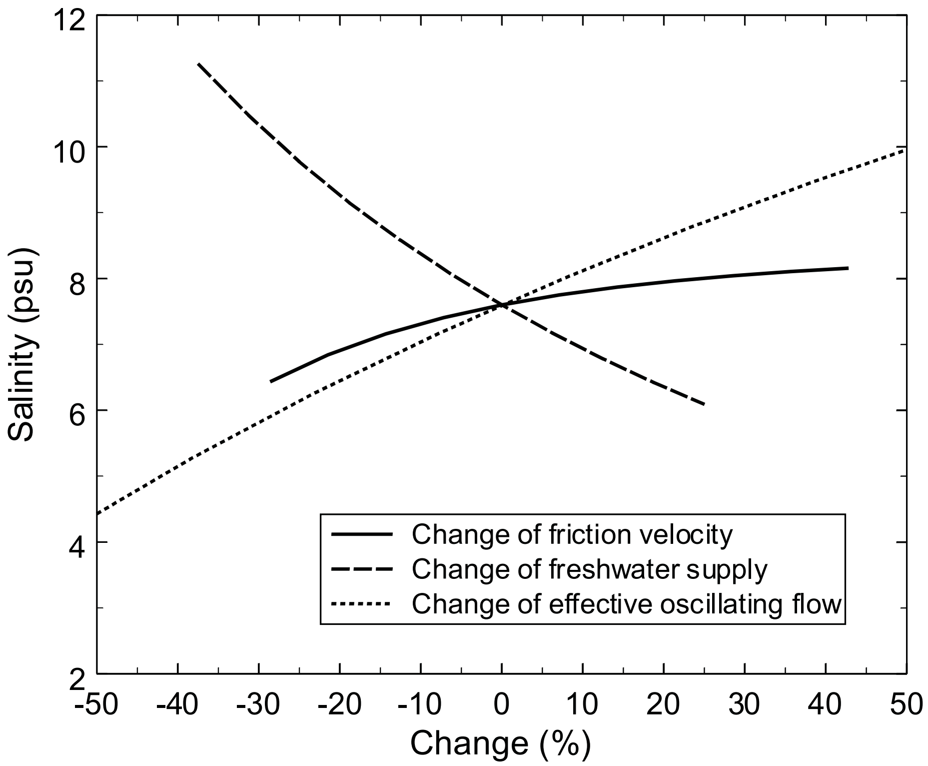


Figure 5-1. Model results for the response of Baltic Sea salinity (S_0) with respect to changes in wind stress (solid line), freshwater input (dashed line) and effective flow oscillations (dotted line).

In order to analyse how topographic changes in the Danish Sounds, primarily in the Öresund, affect Baltic Sea salinity it is necessary to calculate the changes of the oscillating flow. In this Section we review some basics of channel flow and discuss the influence of topographic changes.

The flow of homogeneous water through a channel, such as the Öresund, is primarily forced by the pressure gradient caused by a sea-level difference between the adjacent basins. During steady-flow conditions, the pressure gradient force is balanced by frictional forces against the walls and bottom of the channel and by acceleration in contractions (Stigebrandt, 1980,1992). The latter is actually caused by turbulence downstream from the contraction that dissociates the flow. In the case of a stratified channel there might be an additional pressure gradient due to differences in density in the adjacent basins (Mattsson, 1996a) and additional resisting forces due to transfer of energy to internal waves (Stigebrandt, 1999). The effects of stratification are detectable but still quite small and in the present investigation uncertainties about the stratification are so large that the inclusion of the effects of stratification on the water flow through the Danish Straits seems superficial.

Following, e.g., Stigebrandt (1992), for example, the current speed through a channel reads,

$$u|u| = \frac{g\Delta h}{\frac{1}{2} + C_d L \frac{W+2H}{WH}} \tag{5.14}$$

Where g is the acceleration of gravity, Δh is the sea-level difference between the adjacent basins, C_d is a drag coefficient, L is the length of the channel, W is the width of the channel, and H is the depth of the channel. The flow, Q, is easily found by multiplying with the cross-sectional area as the equation is already averaged across the channel.

$$Q = C_T \frac{\Delta h}{\sqrt{|\Delta h|}} \tag{5.15}$$

Where the transmission coefficient is given by

$$C_T = WH\sqrt{\frac{g}{\frac{1}{2} + C_d L \frac{W+2H}{WH}}} \tag{5.16}$$

In the case of present-day flow across the Darss and Drogden Sills, the transmission coefficients have been found by tuning equation (5.15) towards observations (e.g., Stigebrandt, 1992; Mattsson, 1995, 1996a; Carlsson 1998). The most recent values are

$$C_T = \begin{cases} 0{,}63 \cdot 10^5 m^{5/2} s^{-1} & \text{Öresund} \\ 1{,}67 \cdot 10^5 m^{5/2} s^{-1} & \text{Belt Sea} \end{cases} \tag{5.17}$$

By comparing the empirical value in equation (5.17) for the Öresund and the value obtained by inserting the dimensions of the channel into equation (5.16) one finds, first, that the frictional force is an order of magnitude larger than the resistance due to contraction and second, that the friction must be of considerable magnitude over the whole length of the Öresund, not only at the sill. Further, since the width of the Öresund is considerably larger than the depth and that the frictional forces are dominant, equation (5.16) can be approximated with with good accuracy by the following,

$$C_T = WH\sqrt{\frac{gH}{C_d L}} \tag{5.18}$$

By assuming that the channel length and the drag coefficient are constant, variations of the oscillating barotropic flow can be calculated from equations (5.15) and (5.18) based on time-series of topographic changes. For simplicity a time-varying topographic factor can be defined so that

$$C_T = \alpha C_{T0} \tag{5.19}$$

where C_{T0} is the present flow resistance, as in equation (5.17), and the topographic factor α is defined as

$$\alpha = \frac{WH^{3/2}}{W_0 H_0^{\;3/2}} \tag{5.20}$$

where index 0 denotes present-day values.

A complication is that changing of the flow resistance in one of the straits will be somewhat compensated for by flows through the other. Therefore, it is necessary to evaluate the net change in the magnitude of the oscillating flow by using a simple time-dependent model of the Baltic Sea sea level. Following Stigebrandt (1992), the Baltic Sea sea level is given by

$$Y\frac{dh}{dt} = Q_{Darss} + Q_{Drogden} + Q_F \tag{5.21}$$

Here Y (=370,000 km^2) is the surface area of the Baltic Sea, h the sea level of the Baltic and Q_F (=16,000 m^3/s) is the freshwater supply to the Baltic. The flows across the two sills are given by equations (5.15) and (5.19). The sea-level model defined by equation (5.21) is forced by observed sea-level variations in the Kattegat for a length of time sufficient to give accurate statistics of the flow variability.

5.2 Results

The steady-state model described is used here to quantify the effect on Baltic Sea salinity of the topographic changes at the sills during the past 8,500 years. The topographic changes have mainly affected the Drogden sill in this period. The changes in topography have been used to calculate the net change in oscillating flow for each 500-year band using the model described above in Section 5.1.2. Sea-level observations from 1961–1995 have been used as forcing. Thus, no changes from present-day forcing are assumed. The magnitude of the oscillating flow is defined as the absolute deviation for of the daily flows. The relative change in magnitude of oscillating flow is presented in Figure 5-1.

It is assumed that the effective oscillating flow of the steady-state model (Q_B) is changed in proportion to the magnitude of the oscillating flow. Therefore a time-series of the Baltic Sea salinity changes due to topographic changes in the straits can be calculated from equation (5.13) assuming that all other effects are unchanged. The resulting simulated salinity variations are shown in Figure 5-2. The topographic change has resulted in a maximum an increase in salinity of about 1.5‰ at 5,500 BP. Going further back in time the salinity decreases as a consequence of narrowing of the sills.

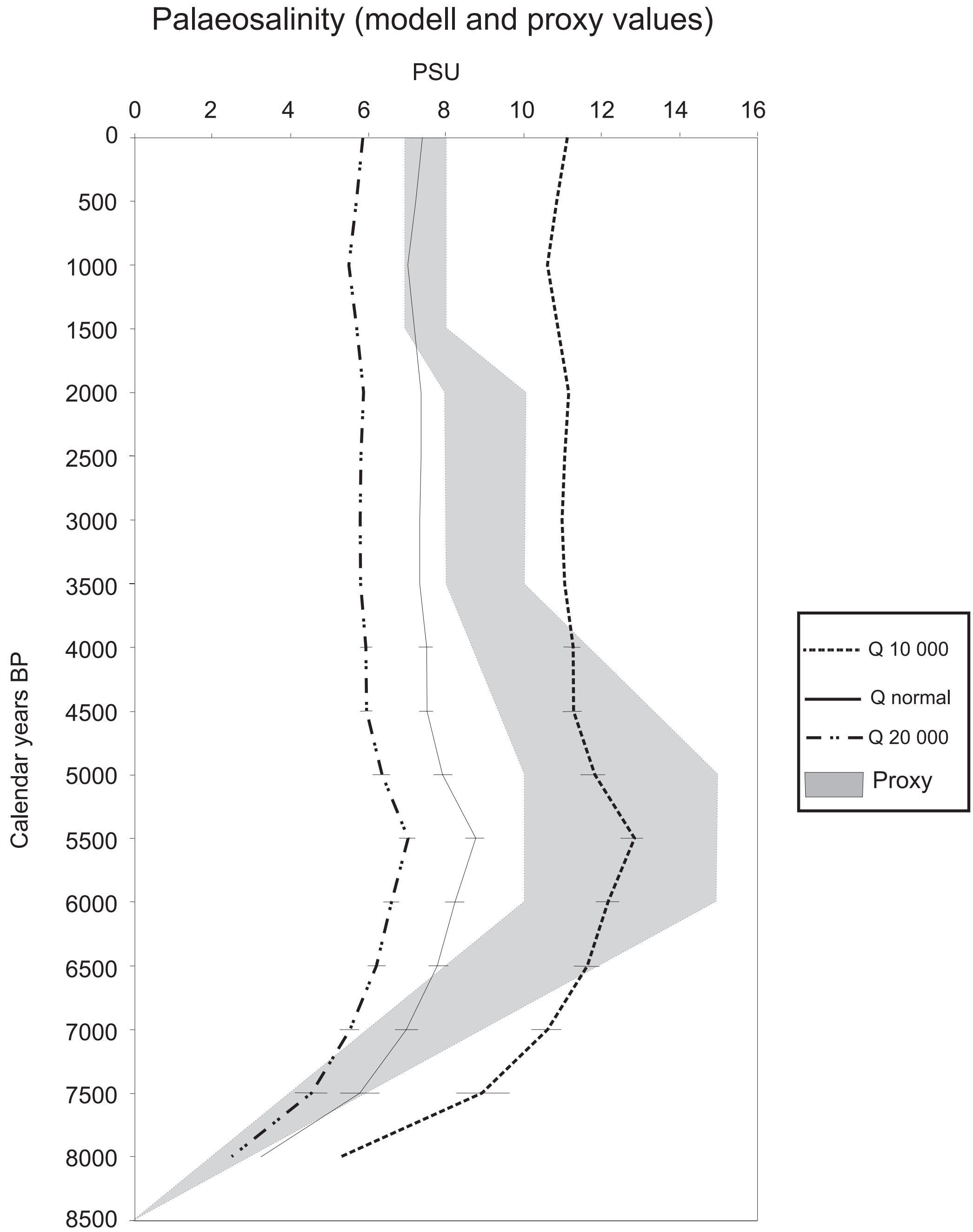


Figure 5-2. *Estimated salinity for the Litorina Sea stage based on proxy data (shaded area) and model values (lines). The horizontal lines represents uncertainties due to different estimates of the cross-section area of the inlets (see Fig. 2.2). Q designates net freshwater input to the Baltic Sea.*

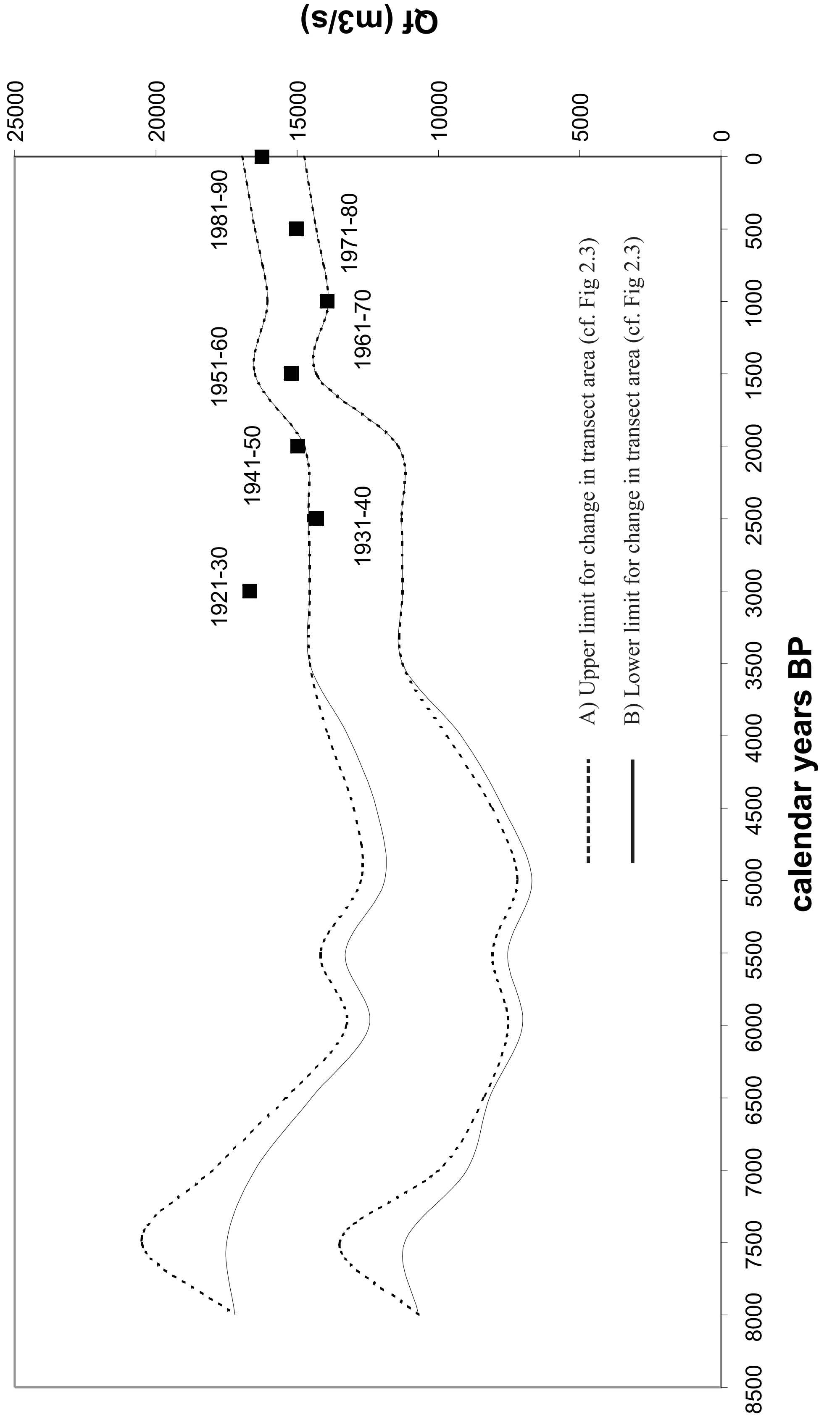


Figure 5-3. *Modelled freshwater supply to the Baltic Sea by assuming observed changes in strait topography and Baltic salinity. The area between the two curves corresponds to the range due to the uncertainty in observed salinity. Calculations are done for upper (a) and lower (b) limits of the change in the topography of the straits.*

6 Discussion

The summary diagram (Fig. 5.2) is based on a compilation of proxy data presented in chapter 2 (shaded area) and model values from Chapter 5 (lines).

The break points of the shaded area has been set at the average age of absolute datings of the boundaries between the Litorina Sea substages (Figures 2-1 a–d). The Ancylus Lake/ Mastogloia Sea boundary has been set to 8,500 years BP as datings of the shoreline displacement indicate that the sounds were not submerged before that time (Christensen, 1982, 1996, 1998). The maximum salinity has been set to 15‰, as there is no evidence of higher salinity being recorded except for a single finding of mollusc shells (*Rissoa* sp.) (De Geer 1889). The seemingly continuous blooms of cyanobacteria (Bianchi et al., 1998) even points to an upper limit of even 12‰. The minimum salinity of the most saline phase has been set to 10‰ based on the occurrence of *Dictyocha speculum* (Witkowski, 1994; Westman, 1998) and the finding of several mollusc species that demand at least 10‰ to survive (e.g. Munthe, 1910; Kessel and Raukas, 1979). The upper salinity limit for the Limnea Sea stage (c. 4,000–2,000 BP) is set to 10‰ based on the disappearance of *Dichtyocha speculum* from the Baltic proper (Witkowski, 1994; Westman, 1998) and changes in the mollusc fauna. The lower salinity limit for the Limnea Sea stage is set to 8‰ as species of the mollusc genus *Litorina* were still present in the Baltic proper (Kessel and Raukas, 1979; Tavast, 1996).

During the most saline stage (c. 6,500–5,000 BP) the climate was warmer than it is today. This coincides with a period of increased dryness recorded in southern Sweden, beginning c. 7,500 BP (Digerfeldt, 1988). There are indications of colder periods coinciding with the most saline phase, but these are considered to be less pronounced than the later cold episodes starting c. 4,000 BP (e.g. Dahl-Jensen et al., 1998). According to the circulation model the changes in cross-section areas of the straits together with a 15% to 60% lower freshwater input during this stage can explain the changes in the salinity between the Litorina Sea stage and the present.

During the following Limnea Sea stage (4,500–2,000 BP) the climate became more unstable but was probably still warmer and dryer than during the present stage of the Baltic Sea (Digerfeldt, 1988; Korhola, 1994, 1995). The cross-section areas of the Öresund and Darss were both almost the same as the present. A 15 to 50% lower net freshwater input (depending on the assumed salinity) compared to the present could explain the higher salinity.

According to Omstedt et al. (1997) the evaporation over the Baltic Sea changes with c. 5% for every degree celsius. The same is also true for the evapotranspiration over the swedish part of the drainage area ($E = 30.4*T+221$, Tamm 1959). As the mean annual temperature during the most saline stage probably was 2-4 °C higher than today (e.g. Dahl-Jansen et al. 1998), the change in net freshwater input caused by temperature was c. 10–20%.

Other factors such as increased tidal water range (Mikkelsen, 1949) and changes in the barotropic-driven salt exchange due to shifts in the polar front (Th. Andrén and E. Andrén, 1999) or a different land-sea configuration in the Baltic Sea (Yu and Harrison, 1995) have been suggested as causing the salinity variations in the Baltic Sea. Both of these factors will, however, have the same effect as changes in the cross-section area and as long as no conclusive evidence for coupling between those factors and the major changes in the Baltic Sea salinity is presented we have to rely on shore-level displacement and changes in freshwater input when modelling the palaeosalinity of the Baltic Sea.

References

Aaby B, Tauber H, 1974. Rates of peat formation in relation to degree of humification and local environment, as shown by studies of a raised bog in Denmark.
Boreas 4: 1–17.

Åker K, Eriksson B, Grönlund T, Kankainen T, 1988. Sediment stratigraphy in the northern Gulf of Finland. In B. Winterhalter (ed.): The Baltic Sea.
Geological Survey of Finland, Special Papers 6; 101–117.

Alley RB, Mayewski PA, Sowers T, Stuiver M, Taylor KC, Clark PU, 1997: Holocene climatic instability: A prominent, widespread event 8,200 years ago.
Geology 25: 483–486.

Anderson DE, Binney HA, Smith M, 1998. Evidence for abrupt climatic change in northern Scotland between 3,900 and 3,500 calendar years BP.
The Holocene 8: 97–103.

Andersson L, Rydberg, 1993. Exchange of water and nutrients between the Skagerrak and the Kattegat.
Estuarine, Coastal and Shelf Science 36: 159–181.

Andrén Th, Andrén E, 1999. Large-scale climatic influence on the Holocene history of the Baltic proper, In: Andrén, E., Holocene environmental changes recorded by diatom stratigraphy in the southern Baltic Sea.
Meddelanden från Stockholms universitets institution för Geologi och Geokemi, Thesis. Paper V, 12pp.

Aure J, Sætre R, 1981. Wind effects on the Skagerrak outflow, In: Sætre, R. & Mork, M., The Norwegian Coastal Current.
University Press, Bergen, Norway, 263–293.

Axell LB, 1998. On the variability of Baltic Sea deep-water mixing.
Journal of Geophysical Research 103, C10: 21,667–21,682.

Berglund BE, 1964. The post-glacial shore displacement in eastern Blekinge, south-eastern Sweden.
Sveriges Geologiska Undersökning C 599: 47 pp.

Berglund B, 1971. Littorina transgressions in Blekinge, south Sweden. A preliminary report.
Geologiska Föreningen i Stockholm Förhandlingar 93: 625–652.

Berglund B, Sandgren P, 1996. The early Littorina Sea environment in Blekinge – chronology, transgressions, salinity and shore vegetation.
Geologiska Föreningens i Stockholm Förhandlingar 118: A64–A65.

Bergström S, Carlsson B, 1994. River runoff to the Baltic Sea: 1950–1990.
Ambio, 23/4–5: 280–287.

Bianchi T, Westman P, Rolff C, Engelhaupt E, Andrén T, Elmgren R, 1998. Cyanobacterial blooms in the Baltic Sea: Natural or Human-induced? In: P Westman:

Salinity and throphic changes in the north-western Baltic Sea during the last 8500 years as indicated by microfossils and chemical parameters in the sediment, Quaternaria A:5,. Thesis. Department of Quaternary Research, Stockholm University, paper IV, 13 pp.

Björck S, Rundgren M, Ingólfsson Ó, Funder S, 1997. The Preboreal oscillation around the Nordic Seas: terrestrial and lacustrine responses.
Journal of Quaternary Science 12: 455–465.

Blackford JJ, Chambers FM, 1995. Proxy climate record for the last 1,000 years from Irish blanket peat and a possible link to solar variability.
Earth and Planetary Science Letters 133: 145–150.

Bond G, Showers W, Cheseby M, Lotti R, Almasi P, deMenocal P, Priore P, Cullen H, Hajdas I, Bonani G, 1997. A pervasive millennial-scale cycle in North Atlantic Holocene and glacial climates.
Science 278: 1257–1266.

Carlsson M, 1998. A coupled three-basin sea level model for the Baltic Sea. Continental Shelf Research 18: 1015–1038.

Caseldine C, Hatton J, Huber U, Chiverrell R, Woolley N, 1998. Assessing the impact of volcanic activity on mid-Holocene climate in Ireland: the need for replicate data.
The Holocene 8: 105–111.

Christensen C, 1982. Haveniveaœndringar 5500–2500 f.Kr. i Vedbœkområdet, NØ-Sjœlland.
Dansk geologisk Forening, Årsskrift for 1981: 91–107.

Christensen C, 1993. Land and Sea in S. Hvass and B. Storgaard eds, Digging into the past, 25 years of Archeology in Denmark.
The Royal Society of Northern Antiquaries, 20–23.

Christensen C, 1996. The littorina transgressions in Denmark in Fischer, A. (ed.), Man & Sea in the mesolithics, coastal settlements above and bellow present sea level.
Oxbow Monographs 53, Oxbow Books, Oxford: 15–22

Christensen C, 1998. Spodsbjerg – en yngre stenalders boplads på Langeland, In Sørensen, H. (ed.) Spodsbjerg: a younger Stone age settlement on Langeland, Rudkøbing, Langeland museum: 7–27.

Dahl-Jensen D, Mosegaard K, Gundestrup N, Clow G D, Johnsen S J, Hansen A W, Balling N, 1998. Past temperatures directly from the Greenland ice sheet.
Science 282: 268–71.

Dansgaard W, White J W C, Johnsen S J, 1989. The abrupt termination of the Younger Dryas climatic event.
Nature 33: 532–534

De Geer G, 1889. Om förekomsten av *Rissoa parva* DA COSTA på Åland.
Geologiska Föreningens i Stockholm Förhandlingar 11: 205–08.

Denton G H, Karlén W, 1973. Holocene climatic variations – their pattern and possible cause.
Quaternary Research 3: 155–205.

Digerfeldt G, 1975. A standard profile for Litorina transgressions in western Skåne, south Sweden.
Boreas 4: 125–142.

Digerfeldt G, 1988. Reconstruction and regional correlation of Holocene lake-level fluctuations in lake Bysjön, south Sweden.
Boreas 17: 165–182.

Eilola K, 1997. Development of a spring thermocline at temperatures below the temperature of maximum density with application to the Baltic Sea.
Journal of Geophysical Research 102/C4: 8657–8662.

Ekman S, 1953. Zoogeography of the Sea. Sidgwick and Jackson. London. 417 pp.

Engqvist A, 1997. Water exchange estimates derived from forcing for the hydraulically coupled basins surrounding Äspö island and adjacent coastal water.
SKB TR 97-14, Svensk Kärnbränslehantering AB, Stockholm, Sweden.

Ernst Th, 1974. Die Hochwachter Bucht. Morphologische Entwicklung einer Küsten-landschaft Ostholsteins.
Schriften des Naturwissenchaftlichen Vereins für Schleswig-Holstein. 44: 47–96.

Eronen M, Ristaniemi O, Lange D, 1990. Analysis of a sediment core from the Mecklenburg Bay, with a discussion on the early Holocene history of the southern Baltic Sea.
Geologiska Föreningens i Stockholm Förhandlingar 112: 1–8.

Fonselius S H, 1969. Hydrography of the Baltic Deep Basins III. Fishery Board of Sweden.
Series Hydrography, Report 23: 97 pp.

Fromm E, 1965. Beskrivning till jordartskartan över Norrbottens län nedanför lappmarksgränsen. Sveriges Geologiska Undersökning Ca 39: 312 pp.

Granlund E, 1932. De svenska högmossarnas geologi.
Sveriges Geologiska Undersökning C 373: 193 pp.

Gustafsson B, Stigebrandt A, 1996. Dynamics of the freshwater-influenced surface layers in the Skagerrak.
Journal of Sea Research 35: 39–53.

Gustafsson B, 1997. Interaction between Baltic Sea and North Sea.
Deutsche Hydrographische Zeitschrift 49, 2/3: 165–183.

Gustafsson B, 2000a. Time-dependent modeling of the Baltic Entrance Area. Part 1: Quantification of circulation and residence times in the Kattegat and the Straits of the Baltic Sill. In press. Estuaries.

Gustafsson B, 2000b. Time-dependent modeling of the Baltic Entrance Area. Part 2: Water and salt exchange of the Baltic Sea. In press. Estuaries.

Hoika J, 1972. Süssau ein neolithischer Wohnplatz an der Ostsee.
Archäologisches Korrespondenzblatt 2: 17–19.

Holmgren K, Karlén, W, 1998. Late Quaternary changes in climate.
SKB TR-98-13, Svensk Kärnbränslehantering AB, Stockholm, Sweden.

Huber K, Kleine E, Lass H-U, Matthäus W, 1994. The major Baltic inflow in January 1993 – Measurements and modelling results. D. Hydr. Z. 46/2: 103–114.

Huckriede H, Clasen S, Meischner D, 1996. Hydrographic and climatic changes recorded in Holocene sediments of the central Baltic Sea.
Baltica 9: 76–91.

Hyvärinen, H., Donner, J., Kessel, H. and Raukas, A., 1988: The Litorina and Limnaea Sea in the northern and central Baltic. *In* Donner, J. & Raukas, A. (eds): Problems of the Baltic Sea History.
Annales Academiæ Scientiarum Fennicæ A 148: 25–35.

Håkansson B, Broman B, Dahlin H, 1993. The inflow of water and salt in the sound during the Baltic major inflow event in January 1993.
Ices Statuary Meeting, C.M. 1993/C:57.

Jakobsen F, 1995. The major inflow to the Baltic Sea during January 1993.
Journal of Marine Systems 6: 227–240.

Jakobsen F, Castejon S, 1995. Calculation of the discharge through Øresund at the Drogden Sill by measurements at two fixed stations.
Nordic Hydrology 26: 237–258.

Jakobsen F, Lintrup M J, 1996. The exchange of water and salt across the Drogden Sill in Øresund September 1993–November 1994.
Nordic Hydrology 27: 351–368.

Jakobsen F, 1997. Hydrographic investigation of the Northern Kattegat front.
Continental Shelf Research 17: 533–554.

Jiang H, Björck S, Knudsen K L, 1997. A palaeoclimatic and palaeoceanographic record of the last 11 000 ^{14}C years from the Skagerrak-Kattegat, northeastern Atlantic margin.
The Holocene 7: 301–310.

Johnsen S J, Clausen H B, Dansgaard W, Fuhrer K, Gundestrup N, Hammer C U, Iversen P, Jouzel J, Stauffer B, Steffensen J P, 1992. Irregular glacial interstadials recorded in a new Greenland ice core.
Nature 359: 311–313.

Karlén W, 1998. Climate variations and the enhanced greenhouse effect. Ambio 27: 270–274.

Karlén W, Kuylenstierna J, 1996. On solar forcing of Holocene climate: evidence from Scandinavia.
The Holocene 6: 359–365.

Karlsson S, Risberg J, Miller U, 1996. South-Central Sweden, Type region S-h. Fissure valley landscape of Södermanland and Uppland. *In* Berglund, B. E., Birks, H. J. B., Ralska-Jasiewiczowa, M. & Wright, H. E. (eds): Palaeoecological events during the last 15 000 years. Regional syntheses of Palaeoecological studies of lakes and mires in Europe.
John Wiley and Sons, Chichester: 254–265.

Kessel H, Raukas A, 1979. The Quaternary History of the Baltic, Estonia. *In* Gudelius, V. & Königsson, L-K. (eds): The Quaternary history of the Baltic. Acta Universitatis Upsaliensis, Symposia Universitatis Upsaliensis Annum Quingentesimum Celebrantis 1, 127–146.

Klitgaard-Kristensen D, Sejrup H P, Haflidason H, Johnsen S, Spurk M, 1998. A regional 8200 cal. yr BP cooling event in northwest Europe, induced by final stages of the Laurentide ice-sheet deglaciation?
Journal of Quaternary Science 13: 165–169.

Klug H, 1980. Der Ansteig des Ostseespiegels im Deutchen Küstenraum.
Eiszeitalter und Gegenwart 30: 237–252.

Knudsen K L, Conradsen K, Heier-Nielsen S, Seidenkrantz M-S, 1996.
Palaeoenvironments in the Skagerrak-Kattegat basin in the eastern North Sea during the last deglaciation.
Boreas 25: 65–77.

Koç N, Hevroy K, Andersen C, Jansen E, 1999. Late Holocene circulation variations in the Norwegian Sea with decadal to century scale resolution. Journal of Conference Abstracts 4: 169.

Korhola A, 1994. Radiocarbon evidence for rates of lateral expansion in raised mires in southern Finland.
Quaternary Research 42: 299–307.

Korhola A, 1995. Holocene climatic variations in southern Finland reconstructed from peat-initiation data.
The Holocene 5: 43–58.

Kõuts T, Omstedt A, 1993. Deep water exchange in the Baltic Proper.
Tellus, 45A: 311–324.

Lass H U, Schwabe R, Matthäus W, Francke E, 1987. On the dynamics of water exchange between Baltic and North Sea.
Beiträge zur Meereskunde, Berlin 56: 27–49.

Lehmann A, 1995: A three-dimensional baroclinic eddy-resolving model of the Baltic Sea.
Tellus 47A: 1013–1031.

Leppäranta M, Seinä A, 1985. Freezing, maximum annual ice thickness and break-up of ice on the Finnish coast during 1830–1984.
Geophysica 21: 87–104.

Liljebladh B, Stigebrandt A, 1996. Observations of the deepwater flow into the Baltic Sea. Journal of.
Geophysical Research 101:C4: 8895–8911.

Lindström G, 1886. Om postglaciala sänkningar af Gotland.
Geologiska Föreningens i Stockholm Förhandlingar 8: 251–281.

Lundqvist G, 1963. Beskrivning till jordartskarta över Gävleborgs län.
Sveriges Geologiska undersökning Ca 42: 74–78.

Lundqvist G, 1965. C 14 dateringar från Gotland.
Sveriges geologiska undersökning C 602: 94 pp.

Lyell C, 1834. On the proofs of a gradual rising of the land in certain parts of Sweden.
Philosophical Transaction of the Royal Society of London I: 1–39.

Magny, 1993. Solar influences on Holocene climatic changes illustrated by correlations between past lake-level fluctuations and the atmospheric ^{14}C record.
Quaternary Research 40: 1–9.

McDermott F, Frisia S, Huang Y, Longinelli A, Spiro B, Heaton T H E, Hawkesworth C J, Borsato A, Keppens E, Fairchild I J, van der Borg K, Verheyden S, Selmo E, 1999. Holocene climate variability in Europe: evidence from ´18O, textural and extension-rate variations in three speleothems. Quaternary Science Reviews 18, 1021–1038.

Matthäus W, Franck H, 1992. Characteristics of major Baltic inflows – a statistical analysis.
Continental Shelf Research 12, 12: 1375–1400.

Matthäus W, Lass H U, 1995. The recent salt inflow into the Baltic Sea.
Journal of Physical Oceanography., 25: 280–286.

Matthäus W, Lass H U, Tiesel R, 1993. The major Baltic inflow in January 1993.
Ices Statuary Meeting, C.M. 1993/C:51.

Mattsson J, 1995. Observed linear flow resistance in the Öresund due to rotation.
Journal of Geophysical Research 100:C10: 20779–20791.

Mattsson J, 1996a. Some comments on the barotropic flow through the Danish Straits and the division of the flow between the Belt Sea and the Öresund.
Tellus, 48A: 456-464.

Mattsson J, 1996b. Analysis of the exchange of salt between the Baltic and the Kattegat through the Öresund using a three-layer model. Journal of.
Geophysical Research 101, C7: 16571-16584.

Mikkelsen V, 1949. Prästö Fjord. The development of the post-glacial vegetation and a contribution to the history of the Baltic Sea.
Dansk botanisk Arkiv 13/5. 171 pp.

Miller U, Robertsson, A-M, 1979. Biostratigraphical investigations in the Anundsjö region, Ångermanland, northern Sweden. Early Norrland 12.
Kungliga Vitterhets Historie och Antikvitets Akademien: 1–76.

Miller U, Robertsson, A-M, 1981. Current biostratigraphical studies connected with archeological excavations in the Stockholm region.
In L-K Königsson & K. Paabo (eds), Florilegium Florinis Dedicatum Striae 14: 167–173.

Mörner N-A, 1969. The late Quaternary history of the Kattegatt Sea and the Swedish West Coast. Deglaciation, Shorlevel displacement, Isostasy and Eustasy.
Sveriges geologiska undersökning. C 640: 487 pp.

Munthe H, 1887. Om Postglaciala aflagringar med Ancylus fluviatilis på Gotland.
Översigt af Kongliga vetenskaps-Akademiens Förhandlingar 4: 719–732.

Munthe H, 1894. Preliminary report on the physical geography of the Litorina-Sea.
Bulletin of the Geological Institution of the University of Upsala 2: 1–38.

Munthe H, 1902. Beskrivning till det geologiska kartbladet Kalmar.
Sveriges Geologiska Undersökning Ac 6: 119 pp.

Munthe H, 1910. Studier över Gottlands senkvartära historia.
Sveriges geologiska undersökning Ca 4: 213 pp.

Munthe H, 1931. Litorinahavet, Clypeushavet och Limneahavet. En orienterande översikt.
Geologiska Föreningens i Stockholm Förhandlingar 53/ 2: 159–186.

Munthe H, 1940. On the late Quaternary development and the Stone Age settlement of north Europe, primarily the Baltic.
Kungliga Svenska Vetenskapsakademiens Handlingar 19/1: 242 pp.

Omstedt A, 1987. Water cooling in the entrance of the Baltic Sea.
Tellus 38 A: 254–265.

Omstedt A, 1990. Modelling the Baltic Sea as thirteen sub-basins with vertical resolution.
Tellus 42 A: 286–301.

Omstedt A, Axell L B, 1998. Modelling the seasonal, interannual, and long-term variations of salinity and temperature in the Baltic proper.
Tellus 50 A: 637–652.

Omstedt A, Chen D, 1999. Influence of atmospheric circulation on interannual variation of maximum ice extent in the Baltic Sea. In manuscript.

Omstedt A, Nyberg L, 1996. Response of Baltic Sea ice to seasonal, interannual forcing and climate change.
Tellus 48A: 644–662.

Omstedt A, Meuller L, Nyberg L, 1997. Interannual, seasonal and regional variations of precipitation and evaporation over the Baltic Sea.
Ambio 26: 484–492.

Omstedt A, Rutgersson A, 1999. Closing the water and heat cycles of the Baltic Sea.
Contribution to Atmospheric Physics, in press.

Påsse T, 1997. A mathematical model of past, present and future shore level displacement in Fennoscandia.
SKB TR 97-28, Svensk Kärnbränslehantering AB, Stockholm, Sweden.

Pedersen F B, 1993. Fronts in the Kattegat: The hydrodynamic regulating factor for biology.
Estuaries 16, 1: 104–112.

Pedersen F B, Møller J S, 1981. Diversion of the river Neva, How will it influence the Baltic Sea, the Belts and Cattegat.
Nordic Hydrology 12: 1–20.

Punning J-M, Martma T, Kessel H, Vaikmäe R, 1988. The isotopic composition of oxygen and carbon in the subfossil mollusc shells og the Baltic Sea as an indicator of palaeosalinity.
Boreas 17: 27–31.

Risberg J, 1991. Palaeoenvironment and sea level changes during the early Holocene on the Södertörn peninsula, Södermanland, eastern Sweden.
Stockholm University, Department of Quaternary Research. Report 20. 27 pp.

Samuelsson M, 1996. Interannual salinity variations in the Baltic Sea during the period 1954–1990.
Continental Shelf Research 16: 1463–1477.

Sayin E, Krauss W, 1996. A numerical study of the water exchange through the Danish Straits. Tellus 48A: 324–341.

Schrum C, Backhaus J O, 1999. Sensitivity of atmosphere – ocean exchange and heat content in North Sea and Baltic Sea – a comparative assessment. Tellus in press.

Segercrantz W, 1896. Några förekomster af postglacialt skalgrus i Finland.
Fennia 12/8: 1–6.

Segerstråle S G, 1927. Skalmärgelfyndigheterna i Finland.
Fennia 47/ 8: 52.

Sernander R, 1911. Om den postglaciala värmetiden och det baltiska havets fauna.
Geologiska Föreningens i Stockholm Förhandlingar. 33: 365–374.

Sjöberg L E, Georgala D, Rickard D T, 1984. Origin of interstitial water compositions in postglacial clays (northeastern Sweden).
Chemical Geology 42: 147–158.

Snowball I F, Sandgren P, Petterson G, 1999. The mineral magnetic properties of an annually laminated Holocene lake sediment sequence in northern Sweden.
The Holocene 9: 353–362.

Sohlenius G, Westman P, 1998. Salinity and redox alternations in the northwestern Baltic proper during the late Holocene.
Boreas 27/2: 101–14.

Sohlenius G, Sternbeck J, Andrén E, Westman P, 1996. Holocene history of the Baltic Sea as recorded in a sediment core from the Gotland Deep.
Marine Geology 134: 183–201.

Stigebrandt A, 1980. Barotropic and baroclinic response of a semi-enclosed basin to barotropic forcing from the sea. In. Freeland, : H. J, Farmer, D. M. & Levingsm C. D. (eds.).
Fjord Oceanography, Plenum, New York, 151–164.

Stigebrandt A, 1983. A model for the exchange of water and salt between the Baltic and the Skagerrak.
Journal of Physical Oceanography 13/3: 411–427.

Stigebrandt A, 1985. A model for the seasonal pycnocline in rotating systems with application to the Baltic proper.
Journal of Physical Oceanography 15/11: 1392-1404.

Stigebrandt A, 1987a. Computations of the flow of dense water into the Baltic Sea from hydrographical measurements in the Arkona Basin.
Tellus 39A: 170–177.

Stigebrandt A, 1987b. A model for the vertical circulation of the Baltic deep water.
Journal of. Physical Oceanography 17: 1772–1785.

Stigebrandt A, 1992. Bridge-induced flow reduction in sea straits with reference to effects of a planned bridge across Öresund.
Ambio 21: 130–134.

Stigebrandt A, 1995. The large-scale vertical circulation of the Baltic Sea. In: Omstedt, A. (editor) First study conference on BALTEX, Conference proceedings, International Baltex secretariat, Pub. 3.

Stigebrandt A, 1999. Resistance to barotropic tidal flow in straits by baroclinic wave drag.
Journal of Physical Oceanography 29: 191–197.

Stigebrandt A, Wulff F, 1987. A model for the dynamics of nutrients and oxygen in the Baltic proper.
Journal of Marine Research 45: 729–759.

Stuiver M, Reimer P J, Bard E, Beck J W, Burr G S, Hughen K A, Kromer B, McCormac G, van der Plicht P, Spurk M, 1998. INTCAL98 Radiocarbon Age Calibration, 24,000–0 cal BP.
Radiocarbon 40: 1041–1083.

Svansson A, 1975. Physical and chemical oceanography of the Skagerrak and the Kattegat, 1. Open sea conditions. Fishery Board of Sweden.
Marine Research Institute Report 1: 1–88.

Svendsen E J, Berntsen M, Skogen B, Ålandsvik, Martinsen E, 1996. Model simulation of the Skagerrak circulation and hydrography during Skagex.
Journal of Marine Systems 8: 219–236.

Svensson G, 1988. Bog development and environmental conditions as shown by the stratigraphy of Store Mosse mire in southern Sweden.
Boreas 17: 89–111.

Tamm O, 1959. Studier över klimatets humiditet i Sverige. Kungliga skogshögskolans skrifter 32. 48 pp.

Tavast E, 1996. Subfossil mollusc investigations of the Baltic: a case study from Estonia. Estonia Geographical studies.
Estonian Academy Publishers, Tallinn 1996: 121–132.

Thulin B, Possnert G, Vuorela I, 1992. Stratigraphy and age of two postglacial sediment cores from the Baltic Sea, Geologiska Föreningens i Stockholm Förhandlingar 114/2: 165–179.

Troels-Smith J, 1939. Stenalderbopladser og Strandlinier paa Amaager.
Medd dansk geologisk forening 9: 489–508.

Törnebohm A E, 1862. Några ord till upplysning om bladet Södertälge.
Sveriges Geologiska Undersökning Aa 4: 51 pp

Walin G, 1981. On the deep water flow into the Baltic. Geophysica 17: 75–95.

van Geel, B., Buurman, J. And Waterbolk, H.T., 1996: Archaelogical and palaeoecological indications of an abrupt climate change in The Netherlands, and evidence for climatological teleconnections around 2650 BP. Journal of Quaternary Science 11: 451–460.

Wellinder S, 1972. The landscape development around the Ulamossen site and the Löddesborg gravefield. Medelanden från Lunds Universitets Historiska Museum 1971-72: 32–38.

Westman P, 1998. On the possible use of the siliceous microalgae; [Dictyocha speculum] and [Ebria tripartita] as biomarkers and palaeoecological indicators in Holocene Baltic Sea sediments. *In*, Westman, P.: Salinity and throphic changes in the north-western Baltic Sea during the last 8,500 years as indicated by microfossils and chemical parameters in the sediment, Quaternaria A:5, paper III, 12 pp. Thesis, Department of Quaternary Research, Stockholm University, Stockholm.

Westman P, Sohlenius G, 1999. Diatom stratigraphy in five off-shore sediment cores from the north-western Baltic implying large scale circulation changes during the last c. 8,500 years. Journal of Paleolimnology 22: 53–69.

Winkel-Steinberg N, Backhaus J O, Pohlmann T, 1991. On the estuarine circulation within the Kattegat. In: Prandle, D. (editor): Dynamics and exchanges in estuaries and the coastal zone. Springer-Verlag, New York, 231–251.

Winn K, Averdieck F R, Erlenkeuser H, Werner F, 1986. Holocene Sea Level Rise in the Western Baltic and the Question of Isostatic Subsidence. Meyniana 38: 61–80.

Winn K, Werner F, Erlenkeuser H, 1988. Hydrography of the Kiel Bay, western Baltic, during the Littorina transgressions. Meyniana 40: 31–46.

Witkowski A, 1994. Recent and fossil diatom flora of the Gulf of Gdansk, Southern Baltic Sea. Origin, composition and changes of diatom assemblages during the Holocene. Bibliotheca Diatomologica 28; 315 pp.

Yu G, Harrison S P, 1995. Holocene changes in atmospheric circulation patterns as shown by lake status changes in northern Europe. Boreas 24: 260–268.