Working Report 2006-111

Literature Review on Future Development of the Baltic Sea and Recommendations for Safety Modelling

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December 2006

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ABSTRACT

The report represents the summary of the main factors, which affects the future development and state of the Baltic Sea. The emphasis is on land uplift, shoreline displacement, and physical, chemical and biological characteristics of the sea. In addition, historical evolution of the Baltic Sea after the last ice age and potential impacts of the different climate scenarios are presented.

The Baltic Sea has an important influence on the development of the geosphere-biosphere interface zone at Olkiluoto. Thus, it is important to take account all these factors in the Safety Modelling of the nuclear waste repository. Different models have been used to evaluate land uplift at Olkiluoto and changes in the sea and land area, especially in the interface zone. The main parameters have been monitored satisfactory way at the Olkiluoto offshore. Based on the present sea sediment stratigraphy, overburden, topography and vegetation it has been fairly well estimated coming future land and sea areas of the Olkiluoto Island and its surroundings. According to the results, Olkiluoto will be part of the continent during the next decade. In the shallow shores of Olkiluoto, the amounts of common reed are increasing naturally, resulting in paludification.

The spatial and temporal changes at Olkiluoto can be estimated and modelled more detailed by using the well focused research sites and more accurate results. In addition, more information is needed on development of the watershed areas, lakes, rivers and vegetation and on sedimentation and erosion processes, hydrology, quality and quantity of seabed sediments and stratigraphy, element budgets and recharge and discharge areas, especially at the geosphere-biosphere interface zones.

Keywords: Baltic Sea, shoreline development, climate, sea-bed sediments, geosphere-biosphere interface zone, past and future development

Kirjallisuusselvitys Itämeren kehitykseen vaikuttavista tekijöistä ja suosituksia loppusijoituksen turvallisuusperusteiden mallinnukseen.

TIIVISTELMÄ

Raportissa esitettään yhteenveto Itämeren tulevaan kehitykseen vaikuttavista tärkeimmistä tekijöistä. Pääpaino on maankohoamisen, rannansiirtymisen, ja merialueen fysikaalisten, kemiallisten ja biologisten tekijöiden kuvaamisessa. Lisäksi tarkastellaan Itämeren historiallista kehitystä viimeisen jääkauden jälkeen sekä eri ilmastoskenaarioiden perusteella aiheutuvia potentiaalisia muutoksia.

Itämerellä on merkittävä vaikutus Olkiluodon geosfääri-biosfääri rajapinnan kehitykseen, joten on tärkeää huomioida eri tekijät ydinjätteiden loppusijoituksen turvallisuustodisteisiin liittyvässä mallinnuksessa (Safety Modelling). Erilaisten mallien avulla on arvioitu Olkiluodon maankohoamista ja siitä aiheutuvia muutoksia maa- ja merialueilla sekä erityisesti niiden rajapinnoilla. Olkiluodon merialueelta on monitorointituloksia kohtalaisen hyvin useimpien parametrien osalta. Nykyisen merenpohjasedimenttitiedon sekä saaren maaperän, topografian ja kasvillisuuden kehityksen perusteella voidaan jo tietyssä määrin arvioida Olkiluodon saaren ja merialueen tulevaa kehitystä. Tulosten mukaan Olkiluoto on osa mannerta jo seuraavan vuosisadan kuluessa. Myös ranta-alueiden ruovittuminen ja muun kasvillisuuden lisääntyminen aiheuttaa paikallisesti merkittäviä muutoksia.

Oikein kohdennettujen tutkimuskohteiden ja tarkempien tulosten perusteella voidaan arvioida ja mallintaa alueelliset ja ajalliset muutokset Olkiluodon kehityksessä yksityiskohtaisemmin. Lisätietoa tarvitaan mm. maankohoamisesta johtuvasta valuma-alueiden järvien, jokien ja kasvillisuuden kehityksestä, sedimentaatio- ja eroosioprosesseista, hydrologiasta, merenpohjasedimenttien laadusta ja stratigrafiasta, ainetaseista sekä tulevista ja poistuvista ainemääristä erityisesti geosfääri-biosfääri rajapinnoilla..

Avainsanat: Itämeri, rannansiirtyminen, ilmasto, merenpohjasedimentit, geosfääri-biosfääri rajapinta, historiallinen ja tuleva kehitys

ACKNOWLEDGEMENTS

This study has been carried out by Pöyry Environment Oy and is based on the research contract for Posiva Oy. On behalf of Posiva the study has been supervised by Ari T.K. Ikonen. Besides him, the author wish to thank to experts Jens-Ove Näslund and Ulrik Kautsky from SKB, Martti Hakanen from University of Helsinki, Laboratory of Radiochemistry, Lasse Koskinen and Petteri Pitkänen from Technical Research Centre of Finland and Robert Broed from FACILIA AB, all have reviewed and given useful comments for the report.

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1 INTRODUCTION

The Olkiluoto Island has been selected for the location of a repository for spent nuclear fuel in Finland. SKB is conducting for site investigations at two sites; in southeast of Forsmark in the municipality of Östhammar in northern Uppland and in Oskarshamn area adjacent to the Oskarshamn Nuclear Power Plant in Oskarshamn Municipality. All three sites situate in the coast of the Baltic Sea.

This report represents the summary of available information for the factors, which affect the future development and thus state of the Baltic Sea and should be taken into account in the modelling for safety assessment (Figure 1). The geological history of the Baltic Sea has been diverse resulting in profound changes in the hydrographic conditions and subsequently also in chemical, physical and biological features of the sea and its catchment area. The emphasis is on land uplift, shoreline displacement, future climate and physical, chemical and biological characteristics of the Baltic Sea, e.g. quaternary stratigraphy, salinity, temperature, sea bottom sediments, radioactivity, runoff and the information needed in the modelling of geosphere-biosphere interface zone for the long term safety.

The Baltic Sea is an essential part of the geosphere-biosphere interface and has to be understood and analysed in a safety assessment of a nuclear waste repository, since the consequences of a potential release occur in this system. The future geosphere-biosphere states in the Baltic Sea are mostly determined by land uplift, sea level changes and future climate and climate related changes (Figure 2). Climate changes are caused by factors external to the climate system and by internal dynamics of climate system. In addition, to natural processes, human activities have been identified as potentially significant cause of climate alternations. Changes in climate and geological environment will have a significant effect on local and regional hydrological conditions, e.g. to groundwater flow, salinity, discharge and recharge areas as seen in the past.

Based on the type of current sea bottom sediments, future ecosystem types at Olkiluoto offshore can be forecasted. The current prevailing forest types will prevail also in the future, but somewhat more wetlands and deciduous forests around the water bodies are predicted. On the basis of the terrain and ecosystems forecast for the selected time period the corresponding landscape model have been built using ecosystem-specific biosphere modules taken from earlier assessments and modelling exercises. However, in order to determine those, more detailed data from these critical threshold areas are required along with the data on the present Baltic Sea sediment quality and thickness.

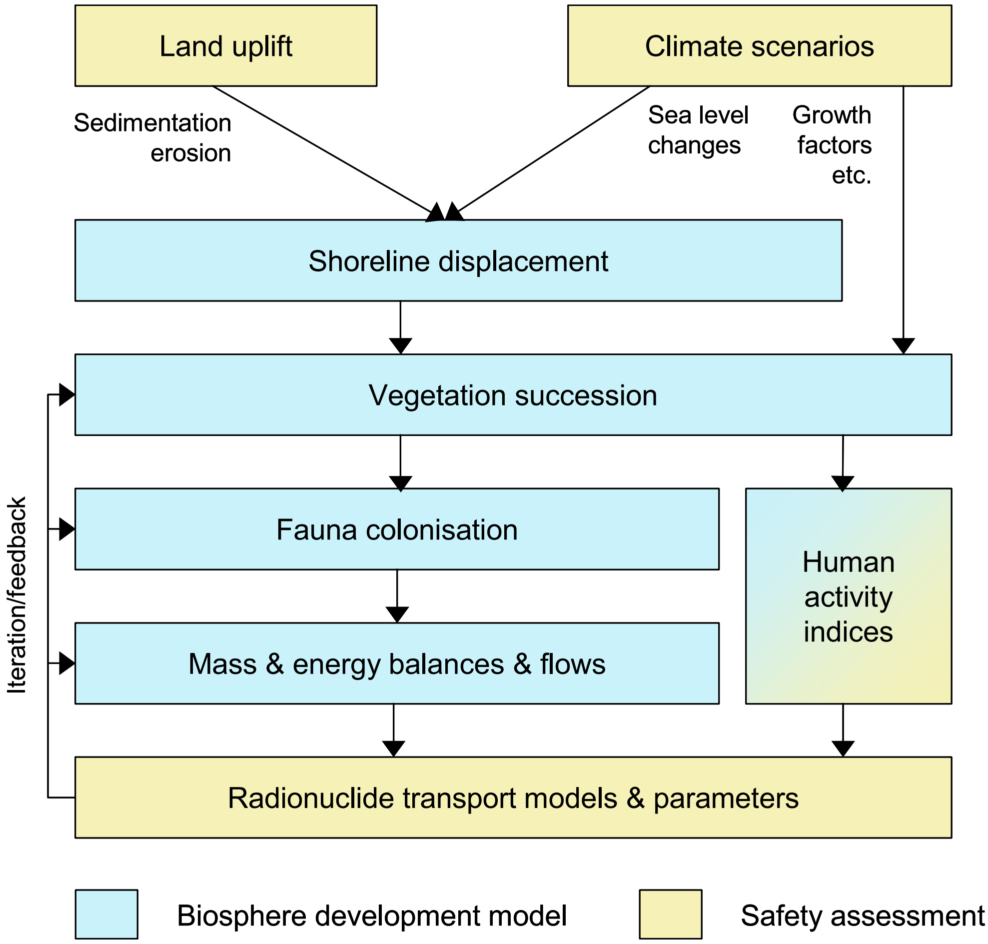


Figure 1. *Stages of the biosphere development modelling approach (Ikonen et al. 2003).*

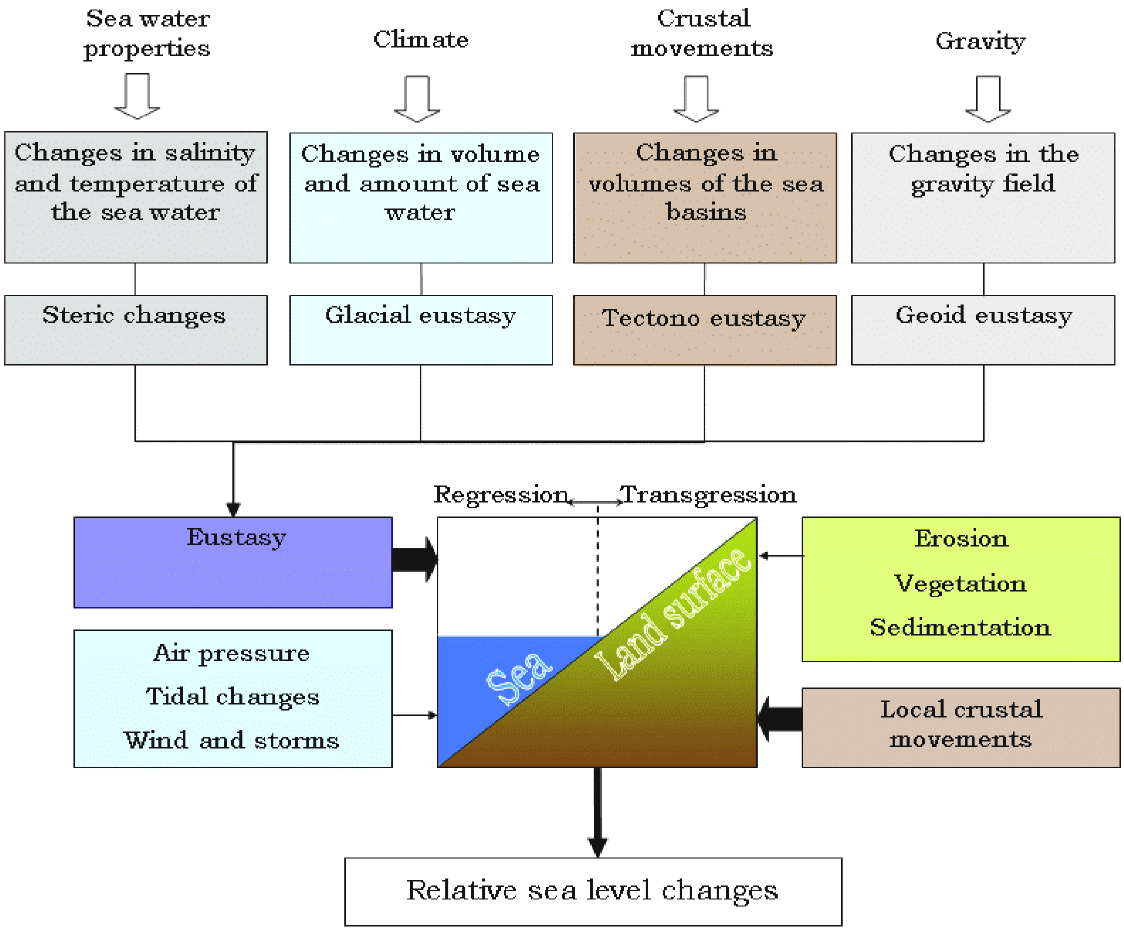


Figure 2. *Factors inducing relative sea level changes and shore displacement. Eustasy is caused primarily by global components whereas the rise or fall of the Earth's surface is more of regional origin. Modified after Mörner (1980) and Mäkiaho (2003) in Mäkiaho (2005).*

2 THE BALTIC SEA

The geological history of the present brackish Baltic Sea has been complicated, resulting in profound changes in the hydrographic conditions and subsequently also in the physical, chemical and biological features of the sea (Voipio 1981, Björk 1995, Eronen 2005). The Baltic Sea basin is formed thousands of millions of years ago as the tectonic plates moved and the bedrock of Fennoscandia was formed. Glaciations have occurred periodically for at least the last 900 000 years. The general topographical outlines of the Baltic Sea area were established long before the beginning of the Pleistocene glaciation. Only when the glacial flow direction coincided with structurally weak zones in the bedrock, considerable deepening and widening of channels and valleys were caused by glacial erosion (Voipio 1981).

All the common forms of glacial deposits, e.g. ground moraines, drumlins and end moraines occur on the bottom of the Baltic Sea. Not only advance and retreat of the ice sheet, but also the melt water streams have caused local erosion of the underlying bedrock. Accumulation of the material transported by melt water took place in the form of eskers and ice-marginal deltas. The finer material, silt and clay was transported to a greater distance and deposited as varved sediments (e.g. Gudelis & Litvin 1976).

Shore displacement in Fennoscandia is the result of two main factors, glacial isostasy and global eustasy. In the past, the isostatic component has been greater than the eustatic component leaving large parts under sea level. The Fennoscandian lithosphere is still undergoing postglacial rebound and the rebound still has 20 000 years to run (Påsse 1996). Uplift can be about considered constant in the timescale of a few centuries. What is unclear is how long the uplift rate will be constant and what will the boundary be (Ekman 1996).

The current surface area of the Baltic Sea is approximately 422 000 km^2 and its volume is 21 000 km^3. The extent of its catchment area exceeds 1 700 000 km^2. Thus, the catchment area is approximately five times as large as the surface area itself (Figure 3). The Baltic Sea is shallow; the average depth of the Baltic Sea is around 55 m, compared to other landlocked seas such as the Mediterranean which has average depth of 1000 m. The greatest depth of the Baltic Sea is only 450 m in the Landsort Deep in the Baltic Proper (Voipio 1981).

The Baltic Sea is connected to the world's oceans by the narrow and shallow waters of the Sound and the Belt Sea which limits the exchange of water with the North Sea. Furthermore, the link with the North Sea is very narrow, the shallowest still being only about 18 m deep. Thus inflows of salt water must be extremely forceful to penetrate and renew the deepest waters of the Baltic proper. This means that some of the water may remain in the Baltic for up to 30 years – along with all the organic and inorganic matter it contains. Salinity levels vary also with depth, increasing from the surface down to the sea-floor. Baltic Proper is mainly replenished by oxygen-rich saltwater flowing in from the North Sea along the sea floor. The salinity of its surface waters varies from around 20 psu (parts per thousand) in the Kattegat to 1–2 psu in the northernmost Bothnian Bay and the easternmost Gulf of Finland, compared to 35 psu in the open oceans (e.g. Gustafsson 2002).

Hundreds of rivers running through the Baltic Sea catchment area bring approximately 450 km^3 of fresh water annually. In addition, 100 km^3 of precipitation in the form of water or snow falls on the surface of the sea itself annually. The same amount, about 100 km^3 evaporates into the atmosphere. The annual fresh water surplus is therefore around 2% of the total volume of the Baltic Sea. Since the volume of the Baltic Sea remains stable in the long term, the surplus runs to the North Sea via the Danish Straits. The Baltic Sea would gradually become a fresh water sea, unless occasional influxes of salty water would occur in the opposite direction. The Baltic Sea consists of a series of sub-basins, which are mostly separated by shallow sills. These basins each have their own water exchange characteristics. Due to the Baltic Sea's special geographical, climatological and oceanographic characteristics, the Baltic Sea is highly sensitive to the environmental impacts of human activities in its catchment area (www.helcom.fi) (Figure 3).

Compared to other aquatic ecosystems, only relatively few animal and plant species live in the brackish ecosystems of the Baltic Sea. Although this limited biodiversity does include a unique mix of marine and freshwater species adapted to the brackish conditions, as well as a few true brackish water species. In the northern and eastern parts of the Baltic Sea salinity level is low; thus fewer marine species can thrive, and marine habitats are dominated by freshwater species, especially in estuaries and coastal waters (www.Balticseaportal.fi, www.helcom.fi).

Nine countries share the Baltic Sea coastline; Sweden and Finland to the north, Russia, Estonia, Latvia and Lithuania to the east, followed by Poland in the south, and Germany and Denmark in the west. About 16 million people live on the coast, and around 80 million in the entire catchment area of the Baltic Sea. The catchment area includes part of Belarus, the Czech Republic, Norway, the Slovak Republic and Ukraine, as some of the rivers find their sources to the Baltic Sea (Figure 4).

The northern part of the Baltic Sea is known as the Gulf of Bothnia out of which the northernmost part is referred to as the Bay of Bothnia. Immediately to the south of it lies the Sea of Åland. The Gulf of Finland connects the Baltic Sea with St. Petersburg. The Northern Baltic Sea lies between the Stockholm area, south-western Finland, and Estonia. The Western and Eastern Gotland Basins form the major parts of the Central Baltic Sea. The three Danish Straits, the Great Belt, the Little Belt and The Sound (Öresund*)* connect the Baltic Sea with the Kattegat Bay and Skagerrak strait in the North Sea.

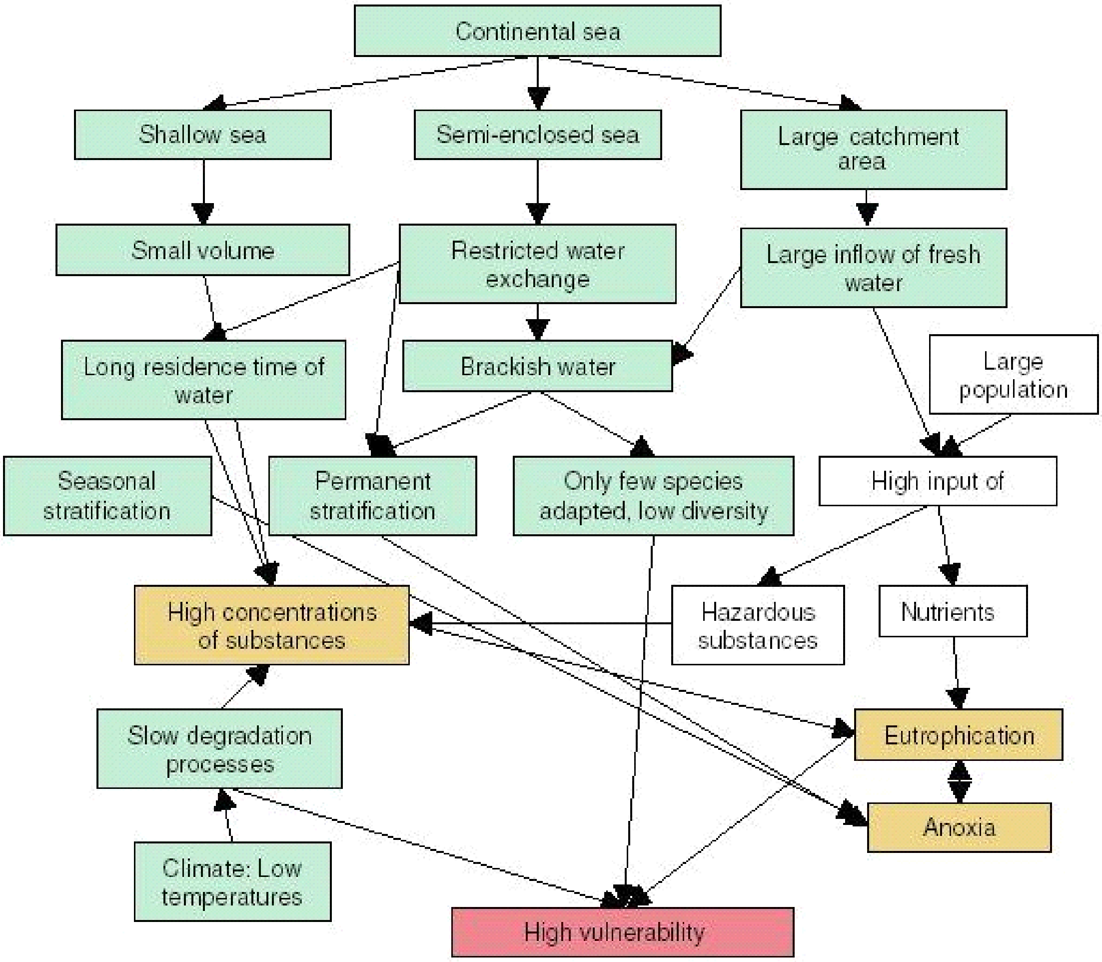


Figure 3. *Specific features and processes which make the Baltic Sea sensitive to environmental and human impacts (green - natural characteristics, white – human impact, yellow – harmful effects (www.helcom.fi).*

Figure 4. *The catchment area of the Baltic Sea (www.helcom.fi).*

OLKILUOTO

The coast of Olkiluoto Island is characterised by shallow bays surrounded by some small islands (Figure 5 and 6). Half a kilometre apart from the shoreline the depth of the water is around 5 meters In general the sea area surrounding the Olkiluoto is shallow, with maximum depth of about 30 m (Kotilainen & Hutri 2003). The shore line is characterized by meadows and swamps (Miettinen & Haapanen 2002). The narrow, only some ten meters wide inlets separate the Olkiluoto Island from the sub-continent. There is more open sea beyond the few rocky inlets at the western end of the island and there are only a few islands to the north, thus the water mixing conditions are favourable. The Rauma archipelago lies to the south. Due to openness of the sea, the winds strongly affect water currents in the area (Posiva 2003).

The quality and biological production of the water are affected by the general state of the Bothnian Sea and the loadings brought by the Eurajoki and Lapinjoki Rivers increasing the concentrations of solids and nutrients, especially at the river mouths. Irregular climatic variation also exerts a considerable influence on the nutrient economy and biological production of the area (Ikonen et al. 2003).

The state of the nearby sea water is partly connected to the nuclear power production activity. The capacity factor of OL1 was 95.1% and that of OL2 96.1% in 2004. The cooling water consumption was 1.81 x 10^9 m^3, and 100 PJ/y of heat was conducted to the sea (Ikonen et al. 2003, Haapanen 2005). Cooling water for the nuclear power plant (on average $60m^3/s$) is taken from sea at the southern side of the island and is discharged to the Iso Kaalonperä Strait to the west. The cooling water intake and discharge of the nuclear power plant significantly affect the temperature and the currents only in their close vicinity. The rise in temperature caused by cooling water has remained local and moderate (Posiva 2003).

Figure 5. *The Olkiluoto offshore (Figure by Posiva/FM-Kartta Oy 16.7.2004) (Posiva 2005).*

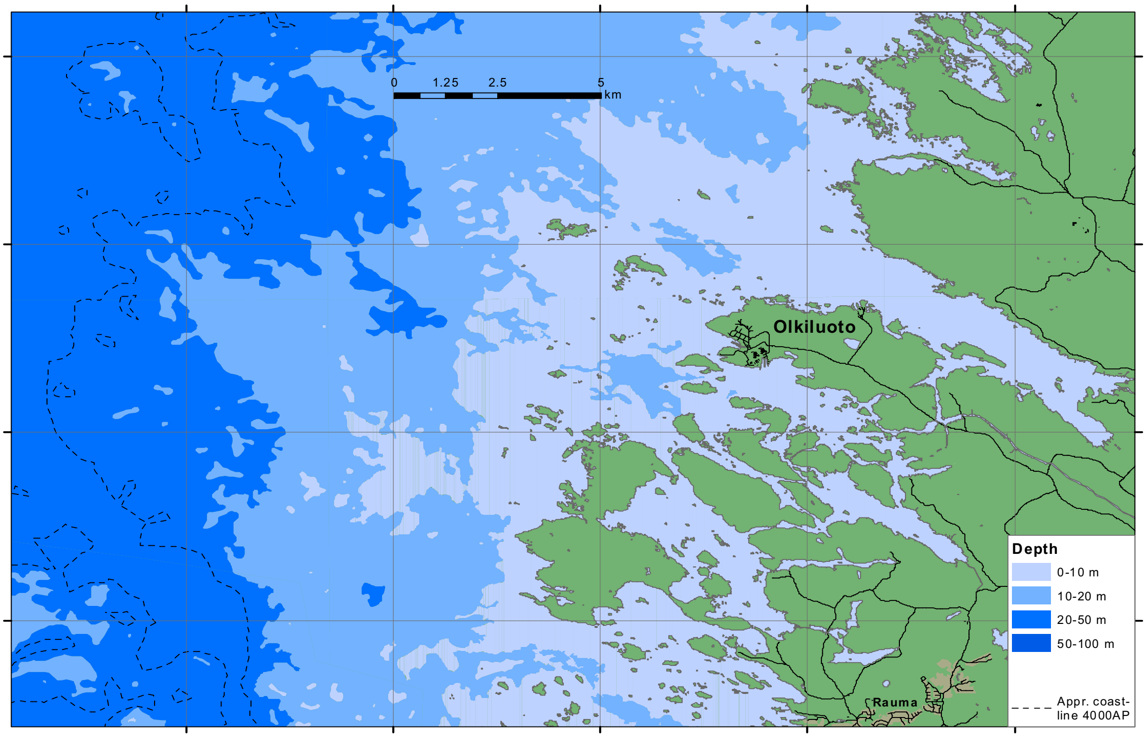


Figure 6. *Coastal area of Olkiluoto. Depth ranges 0-10, 10-20 and 20-50 m are shown in shades of blue, and estimated coastline 4 000 AP, dashed lines by Mäkiaho (2005). The grid size is 5 x 5 km^2.*

The main study area includes the sea around Olkiluoto and extends to 5-6 km distance from the nuclear power plant cooling water discharge site. Description of the sea water monitoring system can be found in, Haapanen (2005), Kirkkala (2005), Kinnunen & Oulasvirta (2005) and Ikonen et al. (2003). In addition, some seawater samples have been taken for hydrochemical characterization and are summarised in the Baseline Report (Posiva 2003). The sea water study locations are shown in Figure 7 (Haapanen 2005).

Averages of the main water quality parameters during the summer for the period of 1990-2001 and the year 2003 north of Olkiluoto, in depths of 0-8 m are presented in Table 1 (Ikonen et al. 2003, Mattila 2004, Haapanen 2005). Water samples from seven observation plots were taken at four instances in February-October 2004 as vertical series with 5 meter distance (Haapanen 2005).

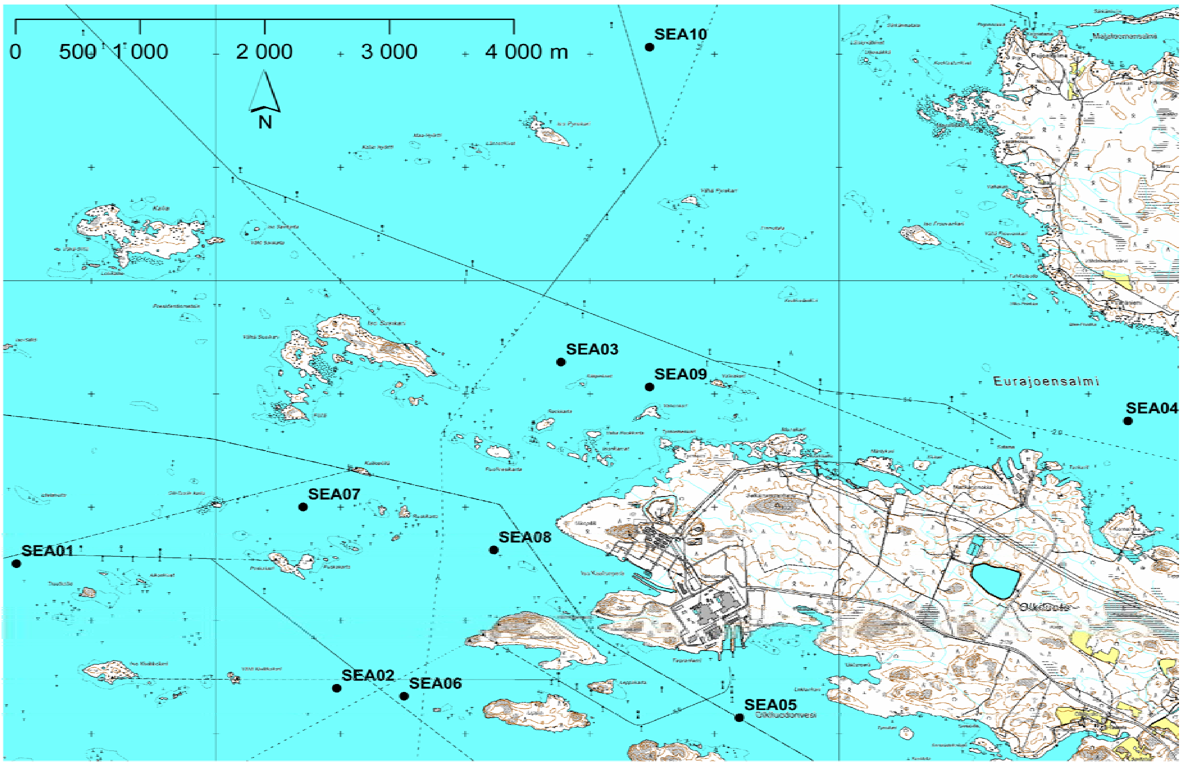


***Figure** 7. Sampling sites at Olkiluoto sea area (Haapanen 2005).*

Table 1. *Averages of the main water quality parameters during the summer for the period of 1990-2001 and the year 2003 north of Olkiluoto, site 480, depths 0-8 m. Corresponding minimum and maximum values and standard deviations (*) are shown in parenthesis (Ikonen et al. 2003, Mattila 2004, Haapanen 2005).*

Parameter	**Dimension**	**1990-2001**	**2003**
pH		8 (7.5-8.3)	7.8 (7.6-8.0)
Temperature	^{o}C	11 (1.9-21)	8.1 (5.0-17)
Oxygen saturation	%	99 (89-117)	96 (4*)
Turbidity	FNU	3.1 (0.4-42)	2.2 (1.3*)
Conductivity	mS/m	937 (310-1060)	911 (750-990)
Suspended solids	$\mu g/dm^3$	5000 (1200-36000)	4000 (2400*)
Total nitrogen	$\mu g/dm^3$	370 (230-900)	390 (200*)
Ammonium nitrogen	$\mu g/dm^3$	13 (1-93)	21 (30*)
Total phosphorus	$\mu g/gm^3$	20 (7-52)	18 (5*)
Salinity	psu	5.5 (1.7-6.2)	5.2 (4.2-5.7)

3 PREVIOUS CLIMATE

The timing and extend of glaciations is largely dependent on small periodic changes in the orbit of the Earth around the Sun resulting changes in the solar radiation on the Earth, known as Milankovitch (1941) theory, even some ambiguities still exist (Wunsch 2004, Forsström 1999).

Prior to the present Holocene epoch, beginning in mid-Pleistocene time approximately 900 000 years ago, what has experienced nine cycles of glaciations and deglaciations; the 100 000 year period of the canonical glacial cycle being characterized by a glaciation phase that has lasted approximately 90 000 years and an interglacial phase that has lasted approximately 10 000 years. At the maximum extent of each of these glacial epochs, sea level has lowered by approximately 120 m in supplying the water of which the great continental ice sheets of the glacial epochs were constructed (Wunsch 2004, Forsström 1999).

Climate data

Our knowledge of past climate is based on measurements of climate variables (temperature and precipitation), which begun at the beginning of the 18th century. It was interpreted proxy data recorded in natural features such as trees, ice sheets, lake and ocean sediments, shoreline terraces and other geomorphological features. Glacial-interglacial climate changes are documented by complementary climate records derived from deep-sea sediments, continental deposits of flora, fauna, loesses and ice cores (e.g. Korhola et al. 2000).

The most continuous sediment formations are found on the ocean floor, where the sedimentation has been continuous for millions of years and almost no erosion has taken place (Petit et al. 1999). Another approach to derive past climate variables is to measure the abundance of different isotopes in deep-sea cores, which were the first physical evidence of a long series of ice ages (Holmgren & Karlén 1998). Most deep-sea cores coversa time span of several 100 000 years. A core covering the last 6 million years was provided by Shackleton et al. (1995a, 1995b) in Holland et al. (1999).

Climatic information can also be retrieved from ice cores. The air composition at the time of freezing is ciphered in the ice. Cores drilled in Antarctica and Greenland (Figure 8), and more recently into smaller ice shields in tropical mountain areas, have provided unrivalled sequences, allowing annual resolution of climatic events in the upper parts of deep ice cores (Ehlers & Gibbard 2003). Ice cores also contain other information of importance for past climate, for instance the content of gases in the atmosphere and the occurrence of volcanic eruptions. The data from a core from Antarctica covering the past 420 000 years was presented by Petit et al. (1999). This core from Antarctica for instance shows that the content of carbon dioxide in the atmosphere has been extremely high in 420 000 year time perspective. The most detailed proxy record comes from the Vostok ice core (Holmgren & Karlén 1998).

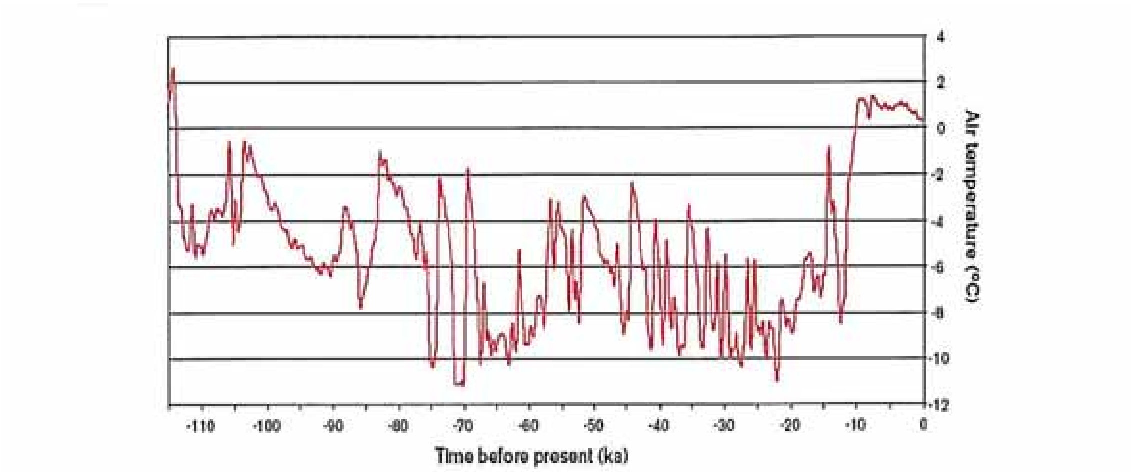


Figure 8. *Temperature reconstructed based on $\delta^{18}O$ in the Greenland ice core (Dansgaard et al. in SKB 2004).*

The launching of meteorological satellites has made climatic monitoring more global. However, the satellite time series span only the last 20-30 years. The satellite observations are central for the monitoring of sea surface temperature and precipitation. Other climatic variables derived from satellite measurements are content of water vapour in the atmosphere, clouds, radiation, surface and atmospheric temperature, snow cover, vegetation and sea level (Holland et al. 1999).

A raw BP, (before present, specifically, before 1950) date cannot be used directly as a calendar date, because the level of atmospheric C-14 has not been strictly constant during the span of time that can be radiocarbon dated. The level is affected by variations in the cosmic ray intensity which is affected by variations caused by solar storms. In addition, there are substantial reservoirs of carbon in organic matter, oceans and ocean sediments and sedimentary rocks. Changing climate may sometimes disrupt the carbon flow between these reservoirs and the atmosphere. The level has also been affected by human activities, e.g. atomic bomb tests in the 1950s and 1960s and release of large amounts of CO_2 used in industry and transportation.

Correct dating of climate changes is of most importance for the understanding of the climate system and causes of climate change. Methods of dating have improved in the last decades, but uncertainties still remain. They grow larger the further back in time we go. Due to different reference sources, somewhat different datings have been used in this report.

3.1 Recent Glaciation

The most reliable evidence for the impact of glacial advances in Fennoscandia is obtained from the recent glaciations, the Saalian and especially the Weichselian. The Saalian begun about 200 000 years BP and it is assumed that the deglaciation at the end of this period was rapid and semi-continuous. Although the climate changes of the late Pleistocene are relatively well known, the events before the Eemian interglacial (130 000-115 000 BP) can only be reconstructed in broad outline (Ehlers & Gibbard 2003). Climate warmed rapidly during the Eemian interglacial period beginning about 130 000 BP. Eemian temperatures were 4 to 5 ^{o}C higher than present-day ones and sea-level is estimated to have been 5 to 6 meters higher. In addition, the humidity was probably higher than present. The saline Eemian Sea had a seawater temperature close to present (Anttila et al. 1999, Forsström 1999).

Climate became colder again about 117 000-115 000 BP as the Weichselian period began (Lindborg 2005). It is characterised by colder phases, interstadials, interrupted by milder interstadials. The model presented by e.g. Fredén (2002), Lundqvist (1992) is often used to illustrate the history of Weichsel (Sveriges National Atlas, www.sna.se) (Figure 9). The Weichselian ice sheet covered eastern and central Europe during 25 000 -12 000 BP and ice cover was at its largest around 2.5 km thick (Anttila et al. 1999, Ahonen et al. 2002).

The Weichselian ice erased most of the evidence of older climatic periods in most areas. However, a nearly continuous sediment sequence covering the last interglacial/glacial cycle had previously been recovered from Sokli carbonatite massive in Finnish Lapland. The dating analyses have indicated the presence of three Weichselian interstadials. Detailed fossil sediment and dating analyses are expected to provide multi-proxy records of late Quaternary climate. The Sokli record can be considered a key site on the northern European continent for ice sheet and climate modelling (Helmens et al. 2005).

Recently Hohl (2005) compiled and evaluated information from Scandinavian (excluding Finland), Northern and Central European geological archives, which record climatic conditions during the Weichselian time period (ca. 115 000-10 000 BP). In that report the reconstructed summer temperatures for southern Sweden during the past interstadials (Brörup 105 000-100 000 BP and Odderade 85 000-75 000 BP) were, in general, 3-6 °C lower than for other investigated European sites (England, Germany, Holland, Poland), while winter temperatures might have differed by 2-20 °C. For the cold period before 14 700 BP reconstructed temperatures for Sweden were about 7 °C lower than those for Germany, as for during the warm period 12 700-11 500 BP records indicate very similar temperatures in these two countries.

The transition from the last glacial, Weichselian to the Holocene (around 13 000 – 12 000 BP) was quite an abrupt event with an increase in air temperatures of about 7 ^{o}C in only 50 years (Dansgaard et al. 1989). Full interglacial conditions were reached after a short cold period in early Holocene at 11 2000-11 050 BP (e.g. Knudsen et al. 1996, Björck et al. 1997). However, the mean annual temperatures dropped 5-7 ^{o}C in a few decades around 11 000 BP with concomitant advance of ice sheets (Anttila et al. 1999).

At about 9 300 years BP, the Baltic Sea became diluted with glacial melt waters. Temperatures at 9 000 years BP were probably close to present and summers were likely warmer (Anttila et al. 1999). More humid and warm conditions with temperatures about 2-3 ^{o}C warmer than today were reflected in the Atlantic chronozone about 7 500-4 500 BP. This warm period is often referred to as the postglacial climatic optimum. It was characterized by deciduous trees growing beyond the northern reaches of the Baltic Sea. The marine area had high organic productivity with algal blooms (cyanobacteria) at least as abundant as seen in the present marine areas. The borehole temperatures directly measured in the Greenland Ice Core Project GRIP and DYE3 boreholes show a long period with temperatures 2-3 ^{o}C warmer than today between 8 000-4 500 BP (Dahl-Jensen et al. 1998). After that period climate became generally colder and more humid about 2 500 years BP (Anttila et al. 1999).

The margins of ice retreat have been well mapped in Fennoscandia but even there remain uncertainties due to different time scales used in different regions (the radiocarbon time scale versus the varve time scales of Sweden and Finland). More important is the lack of more strict observational evidence for the thickness of the ice sheet at any time during the last Glacial Maximum and its retreat. During the retreat phase, the observations of shoreline elevations are quite sensitive to the geometry of the ice load and the spatial pattern of the shorelines can vary significantly from epoch to epoch (Lambeck & Purcell 2003).

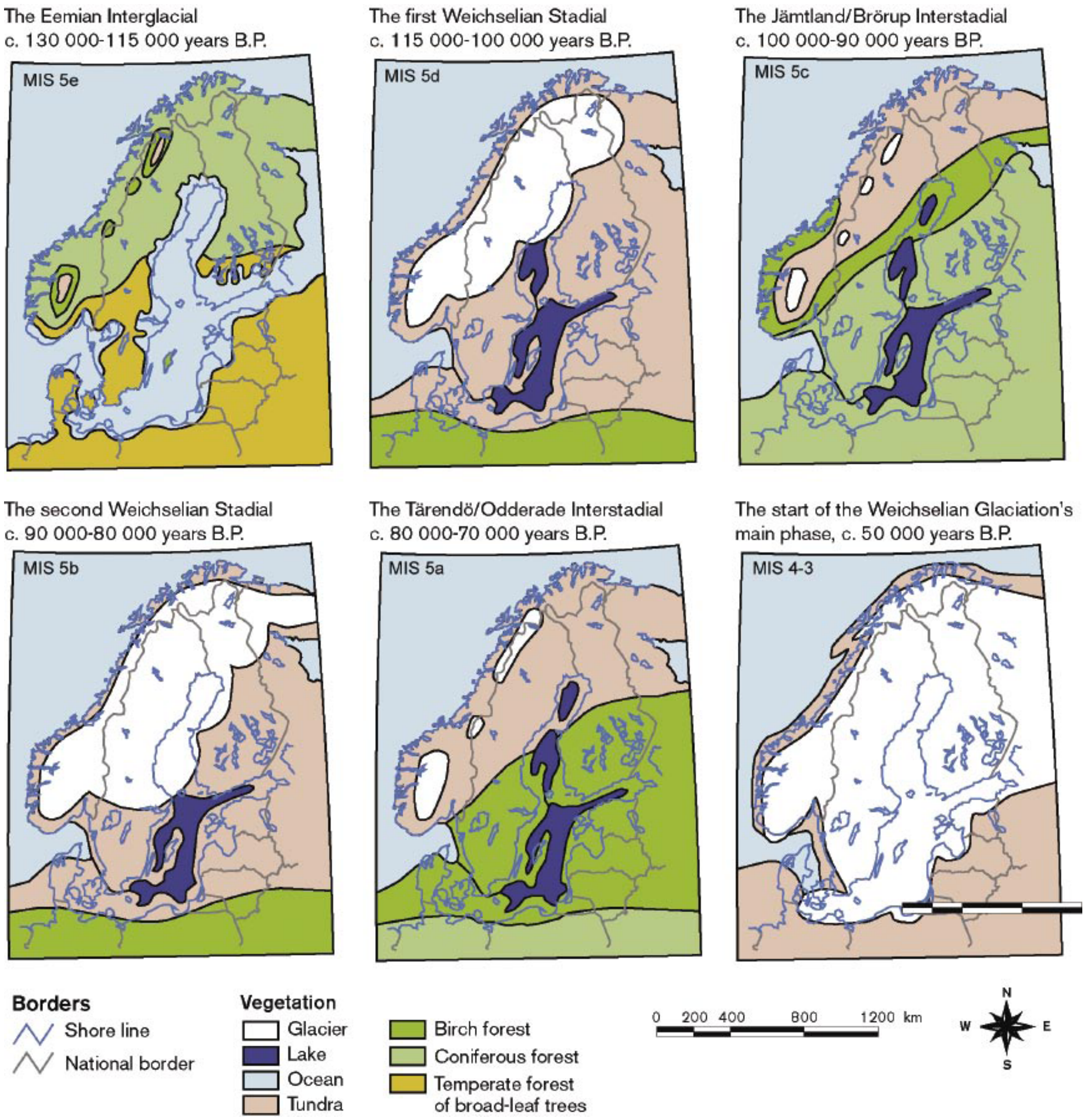


Figure 9. *The development of vegetation and ice cover in northern Europe during the latest interglacial (Eem) and first half of the last ice age (Weichsel). The different periods have been correlated with the Major Isotope Stages (MIS). The maps should be regarded as hypothetical due to the lack of well dated deposits from the different stage (Sveriges National Atlas, www.sna.se).*

Ice sheet thickness and basal conditions

Ice sheet thickness has influences in the conditions below and at the base of the sheet. It is widely recognized that basal conditions are a control on ice sheet dynamics (e.g. Clarke 1987, Alley 1989). The major concern with respect to the safety of a nuclear repository is the basal hydrology and the processes at the bedrock-ice sheet interface that would cause both erosion and lead to the penetration of melt waters into the bedrock.

An ice sheet, or parts of an ice sheet, can be either warm-based or cold-based (frozen-bed). An ice sheet may be warm based initially where an ice sheet forms above an area that does not have permafrost. Alternatively, a cold-based ice sheet may evolve to a warm-based one by several processes (McMurry et al. 2003).

With respect to Olkiluoto, it is uncertain whether warm or cold-based ice sheets prevailed during glacial periods. Based on two different approaches, Forsström & Punkari (1997) and Kleman & Hättestrand (1999) refer to the occurrence of frozen bed conditions in the Fennoscandian ice sheet that covered an area north of Olkiluoto. Results from thermal modelling experiments also suggest that frozen-bed conditions prevail in central ice-sheet areas (Heine & McTigue 1996) and the recent study of Zweck & Huybrechts (2005) seems to confirm it too. Taking this into account, a warm-base ice sheet would have covered Olkiluoto most of the time during the last ice age. Then recharge and preservation of glacial melt water in fractures in the bedrock at Olkiluoto is a relevant issue to consider in repository safety assessments. The processes in cold- and warm-based ice sheets can be similar differing only in the rate at which processes occur:

- Subglacial processes: erosion by abrasion and by quarrying (crack growth; water filled cavities and joints).

- Englacial processes: begin with sediment entrainment and end with deposition. The methods of entrainment are net basal freezing, downward ice intrusion by regelation (Iverson and Semmens 1995) and freeze-on by regelation around obstacles (Weertman 1957). From the three methods, the second one is based on a sound physical theory also supported by experimental studies (Rempel et al. 2001).

The advance of cold-based glaciers has been measured as 1 cm per year in a polar glacier (Meserve Glacier, Antarctica) which has a basal temperature of -17 °C (Cuffey et al. 2000), whereas the advance of warm-based or temperate glaciers (e.g. Alpine glaciers; Black Rapids, Alaska; Russell Glacier, Greenland) can be up to 100 m per year and even more (e.g. Benn & Evans 1998, Knight 1999). Estimates of the maximum Fennoscandian ice sheet thickness during the last ice age range widely (few hundreds to thousands (over 2000 m) of meters (e.g. Koivisto 2004, Eronen & Olander 1990). This uncertainty stems from difficulties in determining the basal temperatures of the ice sheet.

The climate during the present interglacial period, starting about 11 500 years BP, has had varying climate with different warmer and colder periods. The Holocene climatic optimum was a period of warming when the global climate became 0.5-2°C warmer than today. However, the warming was probably not uniform. It began roughly 9,000 years ago and ended about 5 000 years ago. This warmer period changed to a cooler period, which continued until about 2 000 years ago. The period from years 1 500 to 1 850 is known as the Little Ice Age due to lower average temperatures and very uncertain weather. It has been identified two causes of the Little Ice Age from outside the ocean/atmosphere/land systems: decreased solar activity and increased volcanic activity. In contrast, the temperatures have risen throughout most of the twentieth century (e.g. Magny 1993, Karlén & Kuylenstierna 1996, Anttila et al. 1999).

OLKILUOTO

The rapid warming during the Eemian interglacial period, it is estimated that temperatures at Olkiluoto area were 4 to 5 ^{o}C higher than present-day ones and the sea-level is estimated to have been 5 to 6 meters higher. In addition, the humidity has been s probably higher than at present (Eronen et al. 1995).

The early phases and extent of ice sheets during the beginning of Weichselian period are unknown, but Olkiluoto may have been free of ice at least periodically, (Anttila et al. 1999). Estimates of Fennoscandian ice sheet thickness during the last ice age range widely (e.g. Koivisto 2004, Eronen & Olander 1990). Maximum ice thickness over Olkiluoto is estimated to have been about 2 km during the last glacial maximum (Lambeck et al. 1998).

Olkiluoto became free of ice once again about 9 500 BP, and was then covered by the Yoldia Sea. Olkiluoto started to emerge about 3 000-2 500 years BP (Eronen & Lehtinen 1996). The total depression of the area due to glacial loading is estimated to have been 600 m. The uplift is likely to have been 10 times greater during and immediately after deglaciation than at present time (Kahma et al. 2001).

3.2 Permafrost

For permafrost to form, dry and cold conditions; mean annual temperatures below minus 2-3 ^{o}C, are required. Seas, lakes and rivers act as insulators and tend to melt permafrost. Permafrost will restrict such phenomena to the surface environment, while potentially serving to isolate deep groundwater from the surface hydrological regime. Continuous permafrost of over 500 m depth requires tens of thousands of years to develop. For deep continuous permafrost to develop in Fennoscandia, both mean annual precipitation and temperature would have to decrease considerably (Ahonen 2001).

Both the thickness of permafrost and its active layer depend on local climatic conditions, vegetation cover and soil properties. The thickness of permafrost can be altered by changes in the climate or disturbance of the surface. Permafrost thickness is a function of a number of combination factors, including ground surface temperatures and the rate of temperature increase at depth. Because rock deep beneath the earth's crust is hot and molten, the temperature beneath the earth's surface increases with depth, called the geothermal gradient (Figure 10).

It is assumed that during glacial times, permafrost has appeared widely across the Europe. However, there are no physical evidences for this. When Fennoscandia was covered with ice, most proxies have been destroyed by erosion and little is known about the permafrost that may have existed before and during the last glaciation. At present in Finland, palsa mires are the northernmost complex type within the aapa mire zone. Palsas are large peat mounds, up to 7 meters high, containing sporadic permafrost. However, time series of climate domains (glacial, permafrost and temerate/boreal domains, SKB 2003), ice thickness and depth of permafrost and perennially frozen ground, as permafrost is defined by temperature of occurrence of permafrost, does not always mean that the ground is frozen (Figure 11).

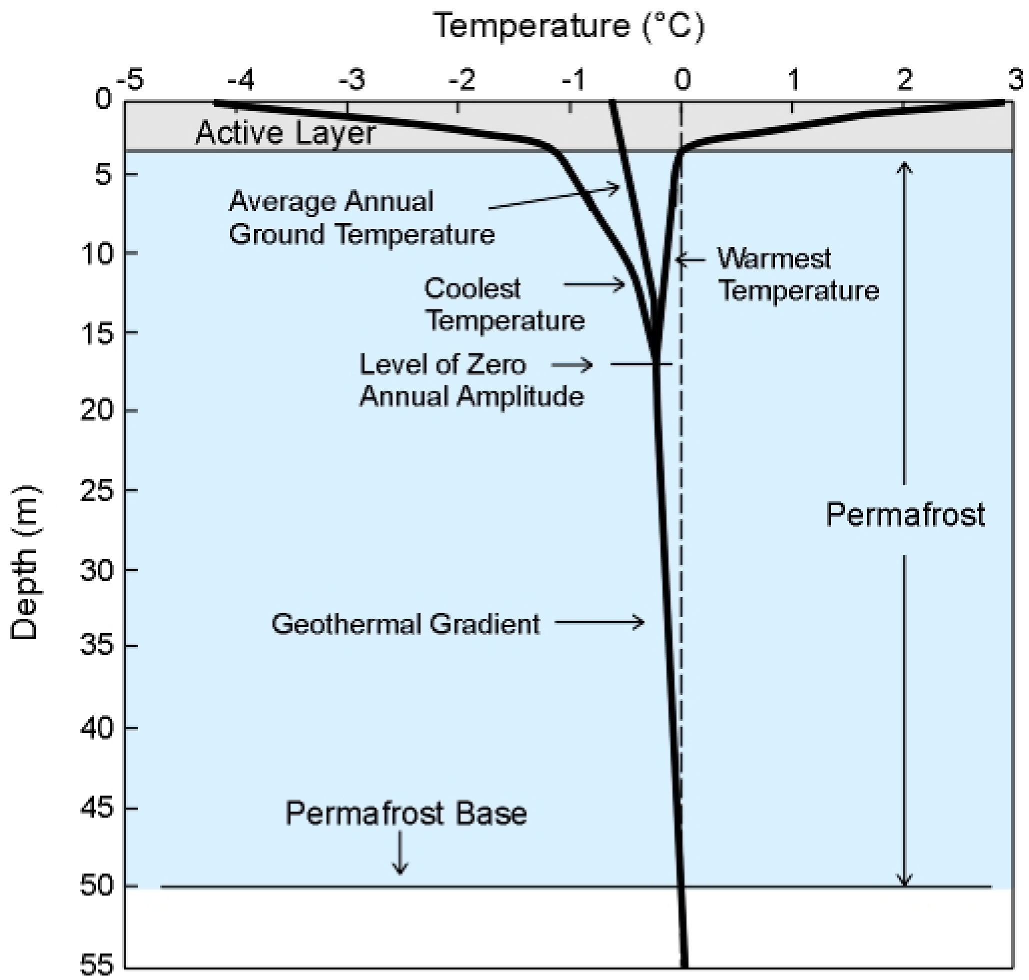


Figure 10. *An illustration of the range in temperatures experienced at different depths in the ground during the year. The active layer (shown in grey) thaws each summer and freezes each winter, while the permafrost layer remains below 0°C (Geological Survey of Canada).*

Under the permafrost layer may exist a more saline groundwater layer (freezing point of saline waters -2 °C). As the ground freezes salts cannot tie themselves to the ice crystals but are pushed deeper and deeper as the freezing proceeds. The more saline layer can be detected using electrical probing methods (e.g. Ruotoistenmäki & Lehtimäki 1997). The segregation of salt from water during freezing has been observed in laboratory experiments, but recent investigations in the Canadian permafrost area at Lupin (Ruskeeniemi et al. 2002, 2004) have not confirmed that the same process happens in natural conditions and neither did the hydrological investigations in Lac de Gras (Northern Canada), where permafrost is 100 m deep under un island (Kuchling et al. 2000). According to Kejonen (2004) estimate of permafrost thickness in Finland might has been during the last ice age 300–1000 m, depending on the geographical area (Posiva 2006).

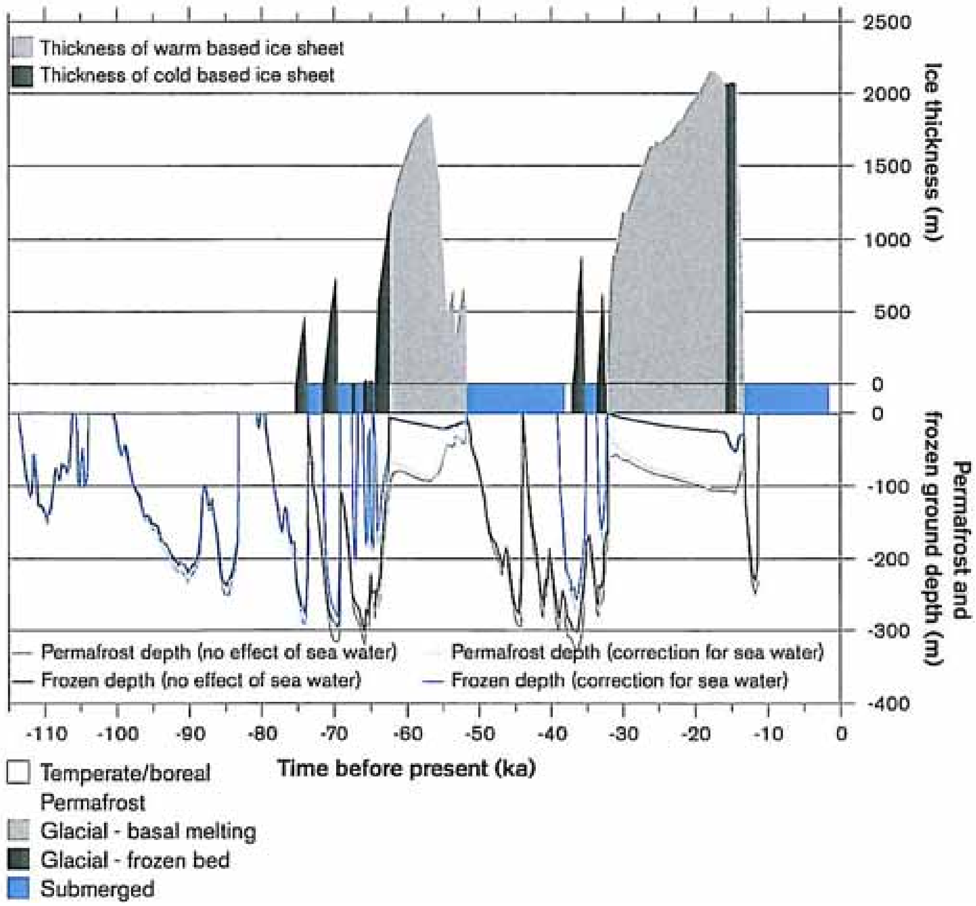


Figure 11. *Time series of climate domains, ice thickness and depth of permafrost and perennially frozen ground /as permafrost is defined by temperature of occurrence of permafrost does not always mean that the ground is frozen (SKB 2004).*

In addition to atmospheric impacts, permafrost is also affected by the precence and absence of ice sheet, the increase of greenhouse gases and global warming in arctic environments. Some areas of discontinuous permafrost throughout Alaska and Siberia are currently thawing from the top and bottom. In a warming climate, carbon and methane trapped in permafrost have a high potential for release into the atmosphere through chemical and biological processes. This can result in a feedback loop and more permafrost thaw; when permafrost thaws and higher levels of CO_2 and CH_4 are released, atmospheric temperature also increases.

OLKILUOTO

To be able to assess the depth and extent of permafrost development in the future at Olkiluoto area, it is essential to know (Cedercreutz 2004):

- If Olkiluoto will be under water, under ice or free of ice and water cover
- For how long the climate in question will last

Temperature of the bedrock at Olkiluoto at a depth of 100-200 m is 6-6.5 °C at 400 m 10.5 °C and at 500 m 12-13 °C (Cedercreutz 2004). Using simple reasoning a permafrost layer reaching down to the repository at 500 m, without the heating effect of the disposal canisters, to develop would require a temperature fall of 12 °C and a time period of tens of thousands of years at that temperature. This would mean a surface annual mean temperature of -6.2 °C (5.8-12 = -6.2), temperatures corresponding to the current region of central Alaska.

At present knowledge the worst scenario is that Olkiluoto will stay above sea level a long time during ice advance elsewhere in Fennoscandia. As the ice masses accumulate in the Fennoscandian mountains there could be a fore bulge in southern Finland, raising Olkiluoto higher above sea-level. Subsidence will be slower in Finland than in Sweden because of the greater distance from the assumed centre of the ice sheet (Cedercreutz 2004).

The most reasonable scenario is that permafrost will indeed develop during the advance of ice sheets, but the most reliable scenario is that Olkiluoto will never be long enough without ice or water cover for permafrost to develop to repository depths (Hartikainen 2005,

Vallander & Erenius, 1991, SKB 2004). This conclusion is based on historical data of the retreat of the Weichselian ice sheet and subsequent postglacial uplift between 10 500 BP and the present (Eronen & Olander 1990).

4 FUTURE CLIMATE CHANGE AND SCENARIOS FOR FINLAND

The variations of the orbital elements affect the amount and distribution of solar irradiation reaching the Earth. According to the astronomical climate theory the orbital variations are the main forcing factor of long-term climate changes (Milankovich, 1921, 1930). Periodic orbital changes for the future can be calculated and predictions regarding future climate made on that basis. This requires modelling of the causality between the orbital driver and the climate change (Boulton et al. 2001).

Model projections of future climate change are based on a simplified simulation of the past global climate and embody large uncertainties. Lack of knowledge and capabilities to simulate the climate system are the main sources of uncertainty for the future climate and ice sheet scenario. The future climate changes cannot be without doubt derived from past variations, mainly because of current changes and poorly unknown dynamics of the major climate-affecting factors such as ocean currents, which cannot be treated in deterministic way. Understanding of how water vapour, clouds and anthropogenic aerosols will influence global warming is still rudimentary and there are uncertainties about the natural variability of the climate (IPCC 2001).

Arrhenius made in 1896 the first attempt to make a climatic change prediction by the means of modelling. He developed a model for the surface – atmosphere radiation budget, and calculated the effect of climate from changes in the carbon dioxide concentration in the atmosphere (Holland et al. 1999). In order to investigate the dynamics of the climate system, coupled models where an atmospheric component is coupled to models of the ocean and the land surface are required. Such models are generally referred to as circulation models or global climatic models (GCL) (Hendersson–Sellers 1996).

The Imbrie & Imbrie (1980) developed a differential model describing the relation between orbital variations and global ice volume (Figure 12) (Ahlblom et al. 1991, King-Clayton et al. 1995, Morén & Påsse 2001). Kukla et al (1981) presented an astronomical climate index (ACLIN) (Figure 13). It is designed to predict the major climate changes in the late and middle Pleistocene and in the near future. Radiometrically dated evidence of climatic changes from three sets of proxy records – pollen, sea level and $\delta^{18}O$ were used to create a climate severity index for the past 130 000 years. As the ACLIN model, the Imbrie & Imbrie model predicts colder conditions during the next glacial cycle than during the Weichselian in spite of week insolation forcing. This may be a consequence of the model's capacity to reproduce the current interglacial.

A model developed during the 1990's at the Lowaine-la Neuve University – the LLN 2D HN, includes a two dimensional model of the northern hemisphere climate system coupled to an ice sheet (Figure 14) (Berger et al. 1996, Gallée et al. 1991, 1992). This model can be used to study the importance of different components of the climate system for the transitions between glacial and interglacial conditions (Morén & Påsse 2001). Surface and subsurface processes that are included in the LLN 2D HN model are: precipitation, evaporation, vertical heath fluxes, surface albedo, oceanic heat transport and oceanic mixed-layer dynamics. Potentially important processes that are not taken into account are variations in water vapour transport, cloudiness, atmospheric dust content and deep-water circulation. Variations in insolation force the model. As the model does not include a carbon cycle, the CO_2 concentration in the atmosphere is also considered as an external forcing. The LLN 2D HN model has been used to simulate the present climate of the Northern Hemisphere and it reproduces the main climatic characteristics well (Gallée et al. 1991, Morén & Påsse 2001).

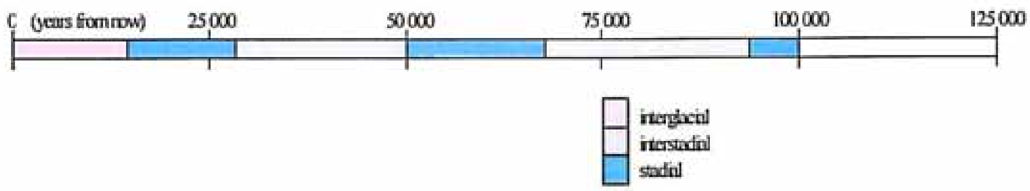


Figure 12. *Future climate according to the Imbrie & Imbrie model. The model simulates 100 000 years into the future (Cedercreutz 2004).*

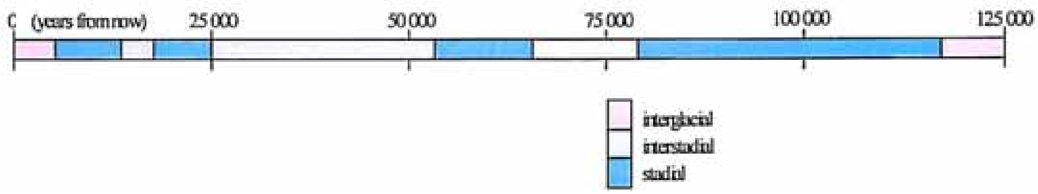


Figure 13. *Climate over the next 125 000 years as simulated by the ACLIN model (Cedercreutz 2004).*

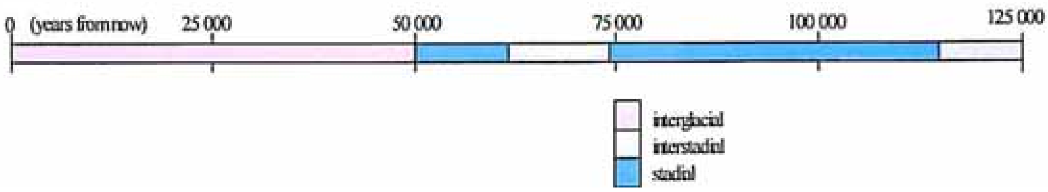


Figure 14. *Northern hemisphere climate over next 125 000 years as simulated by the LLN 2D HN model using no fossil fuel CO_2 contribution (Cedercreutz 2004).*

Morén & Påsse (2001) have presented a scenario for the next 150 000 years describing the evolution of the climate and related environmental changes (Figure 15). They used the results from Imbrie & Imbrie and ACLIN models together with geological reconstructions of past conditions and results from LLN 2D HN model. All these models are based on astronomical climate theory (e.g. Figure 16). Morén & Påsse (2001) assumed that astronomical data is valid, because the models reconstruct the long-term evolution during the Weichselian quite well. The time period 150 000 years was chosen because it is the approximate time period required encompassing an interglacial cycle. The scenario is not a prediction of the future but merely a description of an evolution similar to one that may occur. The scenario can also be seen as a general description of the evolution during any of the glacial cycles of the Quaternary. The scenario for the next 150 000 years is accounted in Table 2.

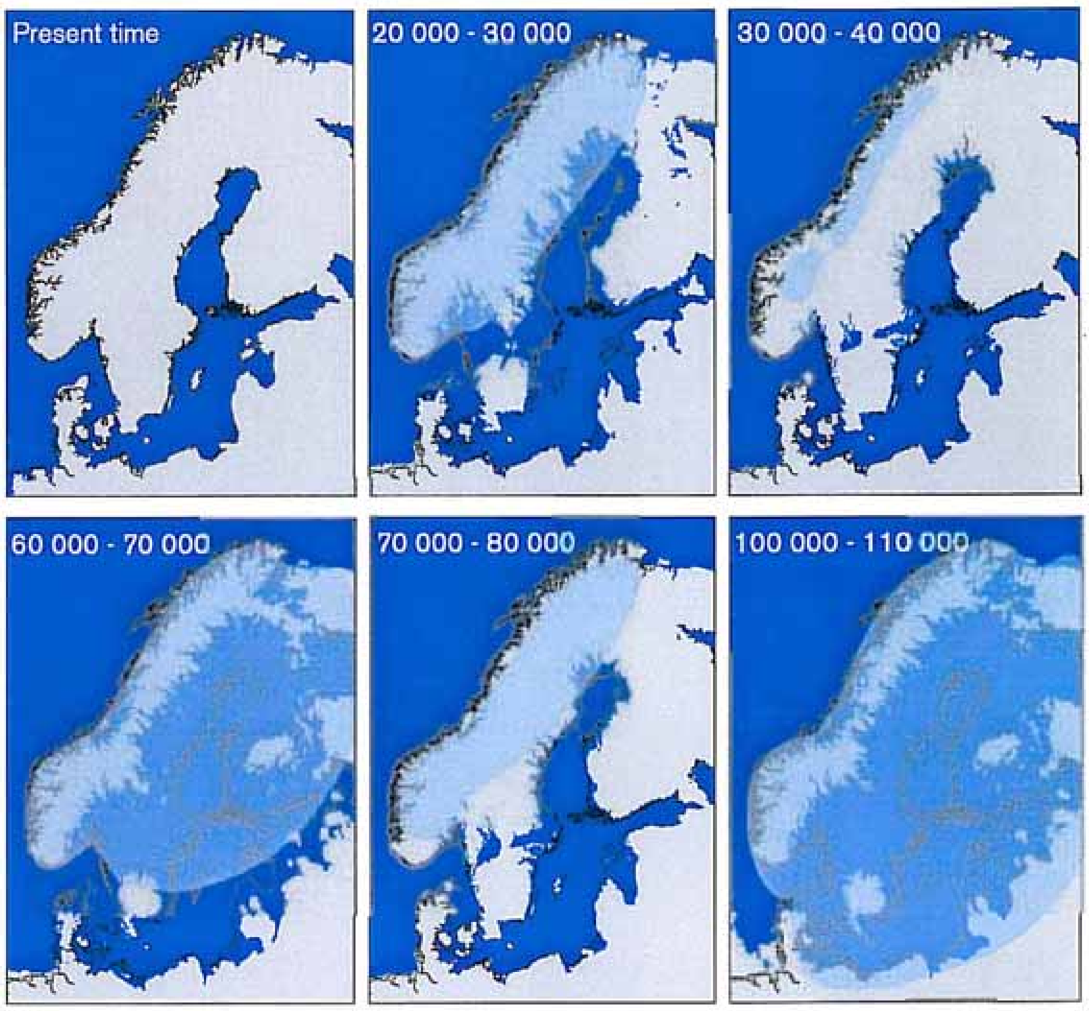


Figure 15*. Scenario of the ice sheet extension for the next 150 000 AP (Morén & Påsse 2001).*

Table 2. *Climate and ice sheet scenario for the next 150 000 years (Morén & Påsse 2001).*

Period (AP)	**Climate and ice sheet**
0-20 000	Gradually colder, an ice sheet is starting to grow in the mountains at about 5 000 years.
20 000-30 000	Stadial, the ice extends to the current Swedish Baltic coast in the east and to the lake Siljan in the south.
30 000-40 000	Interstadial, the ice sheet melts except for the ice caps in the mountain area.
40 000-60 000	The climate is getting colder again and the ice sheet expands to the south-east.
60 000-70 000	Stadial, the ice sheet covers Finland and extends over the lake Vättern in the south.
70 000-80 000	Interstadial, the ice sheet melts away rather quickly to an extension that is little bit less than the 20 000–30 000 years minimum
80 000-100 000	A new stadial is initiated, the climate is getting colder and the ice sheet expand
100 000-110 000	Stadial, the glacial maximum is reached, the ice sheet extends into Russia in the east and into northern Poland and Germany.
110 000-130 000	Interglacial, the ice melts away and the climate gets similar to the present, the warm maximum is reached at about 120 000 years.
130 000-150 000	A new glacial period is initiated, the ice sheet grows and reaches almost as far as during the 60 000–70 000 years maximum at theend of period.

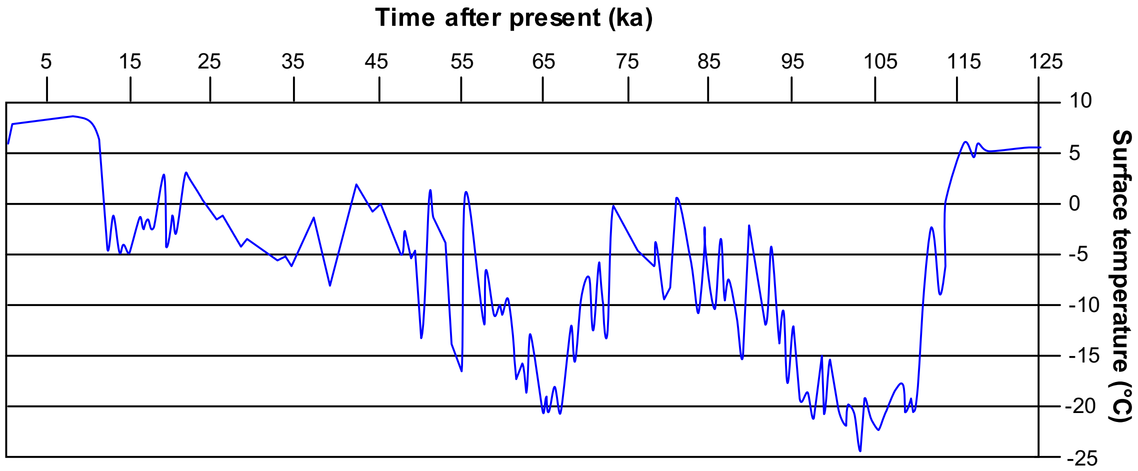


***Figure 16**. Surface temperature 0-125 000 AP for scenario B. Modified to fit a future timescale (SKB 2004).*

Cedercreutz (2004) has presented five different climate scenarios (Figure 17). Two of these, namely scenarios B and D, were selected for further study and renamed as Weichselian-R (R for repetition of the Weichselian) and Emissions-M (M for moderate emissions), respectively. The Emissions-M scenario (D) reaches further into the future opposed to the Weichselian-R scenario (B), which ends at 125 000 AP. The reason for this is that both scenarios reach until the next interglacial, which occurs at different times. These scenarios were selected because there exists data and they can be easily refined to full scenario description (Posiva 2006).

Scenario A is based on the scenarios of Imbrie & Imbrie and ACLIN and LLN 2D NH model. They do not include CO_2 concentrations, except the LLN model takes it as constant < 230 ppmv. The B scenario is taken from SKB (2004). Scenarios C, D and E are based on BIOCLIM scenarios, respectively (BIOCLIM 2001). The BIOCLIM simulations for future global climate change stated at the present day with a present-day simulated ice sheet, i.e. Greenland ice sheet. The BIOCLIM scenarios take into account CO_2 in calculating ice volume and temperature (Figure 18).

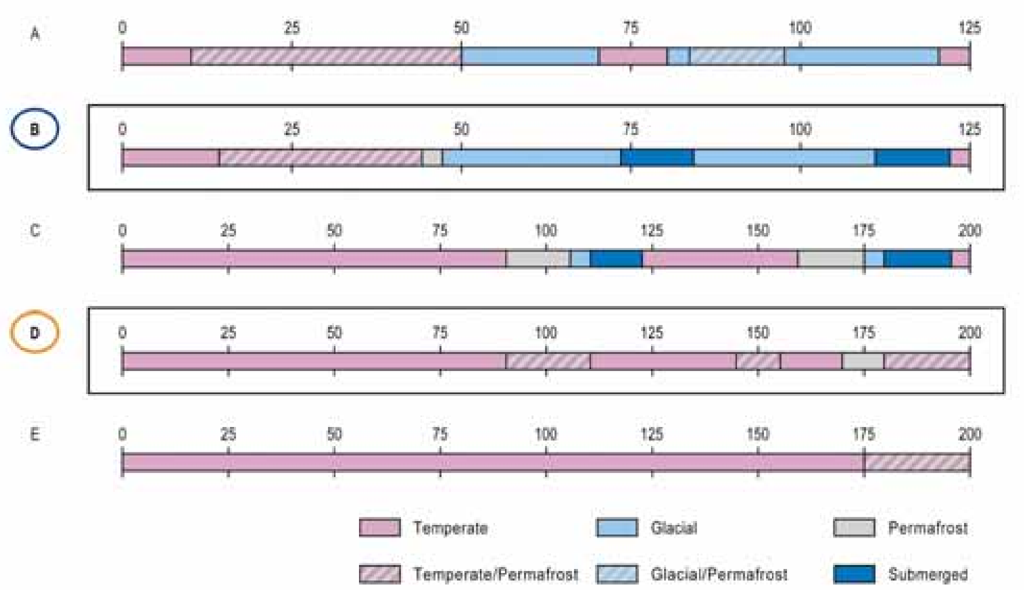


Figure 17. *Five climate scenarios from Cedercreutz (2004) of which scenarios B (Weichselian-R) and D (Emissions-M) have been selected for further study. The timescales are from present to 125 000 AP (A, B) and 200 000 AP (C, D, E).*

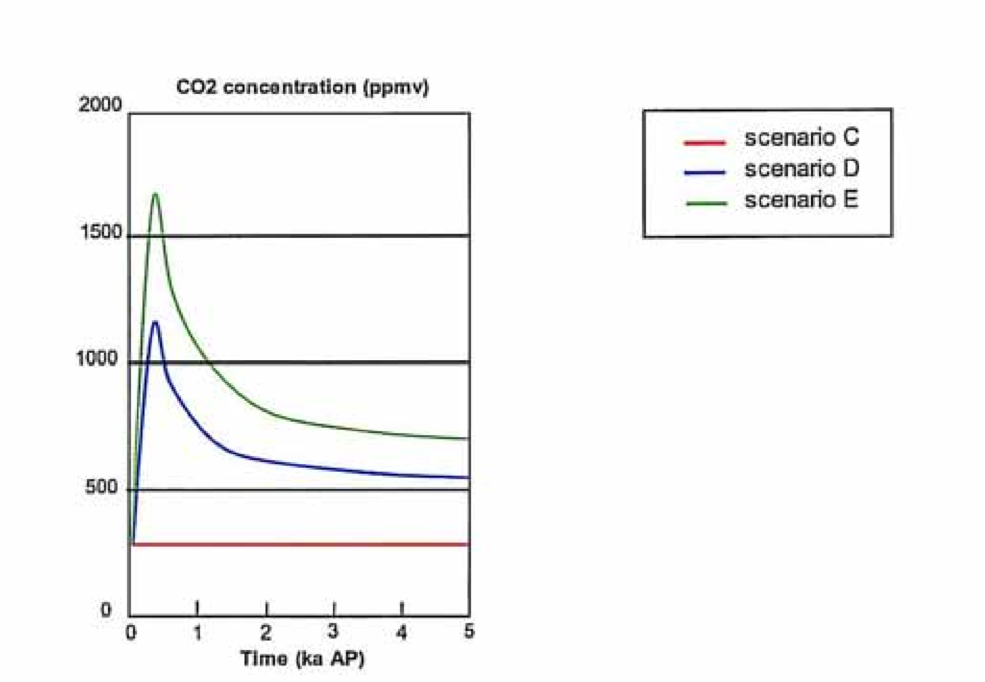


Figure 19. *Graphs relating to scenarios C (red line), D (blue line) and E (green line).* CO_2 *concentrations for the next 5000 years. Figure modified after BIOCLIM (2003) by Cedercreutz (2004).*

The Weichselian-R scenario (B) is based on SKB main scenario (SKB 2004), in which it is assumed that after 10 000 Weichselian glacial cycle will repeat itself (Posiva 2006) (Figure 19). No greenhouse gases emissions induced climate change is taken into account. The presented evolution of climate-related conditions is the result of numerical modelling of the evolution of the Scandinavian ice sheet during the Weichselian (Fastook, 1994, Fastook & Holmlund, 1994). This scenario is based on preliminary modelling results for the Forsmark site in Sweden. Climatic characteristics at Forsmark (SKB 2004a) are closely similar to those at Olkiluoto (Ikonen 2005). Modelling results for the Weichselian glacial cycle at Forsmark are here taken as an analogue for the future climatic evolution over a glacial cycle at Olkiluoto (0-125 000 AP). Geological data (e.g. Eronen & Olander 1990, Lundqvist 1992) have shown that the Fennoscandian ice sheet grew in a fan-like motion reaching Forsmark and Olkiluoto almost simultaneously.

Temperature reconstructions are based on the GRIP ice core from Greenland (Dansgaard et al. 1993). The temperature curve for Forsmark (SKB 2004a) is based on further calibration with temperature data from the Northern European climate archives (Hohl 2005).

According to recent data on the Fennoscandian geothermal heat flux, Olkiluoto lies in an area with geothermal heat flux values of approximately 60 mWm^{-2} (Näslund et al. 2004, SKB 2004a), which is higher than the corresponding value for Forsmark. This implies that Forsmark would probably experience deeper permafrost than Olkiluoto. In Forsmark permafrost depth is estimated to be close to 300 m (SKB 2004a).

At Olkiluoto the first 40 000 are characterized by alternating permafrost and temperate type climate after which, about 50 000 AP, the area will be for the following 25 000 Olkiluoto will be submerged as the glacier retreats and remains submerged until the next glaciation 87 000–112 000 AP. The next interglacial is expected at about 112 000 AP. Similar climatic conditions to the current ones will return in a new cycle at about 125 000 AP. For most of the period (50 000–125 000 AP) Olkiluoto will be covered either by ice or by water (Posiva 2006). Temperatures, precipitation values, Northern Hemisphere ice volume, relative sea level as well as CO_2 concentration for the scenario have been compiled in Table 3.

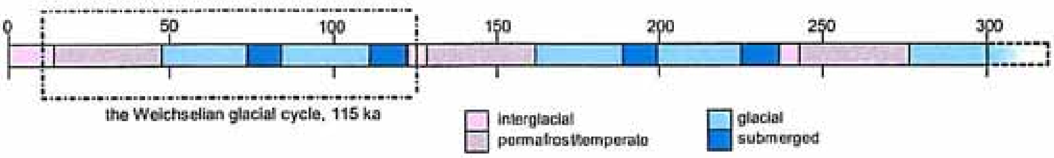


Figure 19. *SKB's scenario – the Weichselian–R scenario- the reconstructed Weichselian climate repeats itself over and over again (Cedercreutz 2004).*

The Emissions-M scenario (D) is based on the simulation scenario B2 of the EC project BIOCLIM (2003b). It assumes that fossil fuel burning will contribute to a moderate increase in the natural CO_2 concentration in the atmosphere (Posiva 2006). Total atmospheric CO_2 will add up to a maximum value of 1100 ppmv at 350 years AP. This increase in atmospheric CO_2 will contribute to a rise of global temperatures. Over the next 50 000 AP variations in insolation will be small (Berger et al. 1996), which reinforces the warming impact of atmospheric CO_2. The simulation started at the present day with a present-day simulated northern hemisphere ice sheet, i.e. the Greenland ice sheet of 3.2×10^6 km^3, and used combined natural CO_2 concentration and low fossil fuel contribution. The fossil fuel contribution will be large over the next 40 000 AP compared to the current annual CO_2 emissions. Therefore the simulated ice volume remains small (less than 30×10^6 km^3, on average) over most of the next 400 000 years, except for short cooler periods at 178 000, 267 000 and 361 000 AP. After 500 000 AP, the fossil fuel contribution will become smaller and the climate can start a recovery to its natural state (BIOCLIM 2001). Temperatures, precipitation values, Northern Hemisphere ice volume, relative sea level as well as CO_2 concentration for the scenario have been compiled in Table 4.

According to Cedercreutz (2004) and Morén & Påsse (2001) it could be predicted that the next 50 000 years will be characterised by small - if any – Northern hemisphere ice sheets. No significant variability in the continental ice volume is expected during that time. The behaviour of the ice sheets after 50 000 years will depend on how much and for how long the atmospheric CO_2 concentration is influenced by human activities. Even if CO_2 concentrations rapidly returns to a natural concentration, the amount of continental ice will only return to the natural values after 100 000 years AP. Figures 20 and 21 present schematic illustrations of the climatic events of the Weichselian-R scenario (B) and the Emissions-M scenario (D), respectively. In the snapshots Olkiluoto is presented as a grey cliff with varying surface covers, temperatures and shoreline conditions.

Table 3. *The Weichselian-R scenario: temperature, precipitation, carbon dioxide, ice sheet thickness, relative sea level and surface cover for Olkiluoto. (MAAT=mean annual air temperature). Data from SKB (2004), Hohl (2005), Ruosteenoja (2003), Ikonen (2005) and Koivisto (2004). (Posiva 2006).*

Time (ka AP)	MAAT (°C)	Temperature (°C		Precip. (mm)	CO_2 (ppmv)	Ice thickness (m)	Relative sea level (m)	Surface cover
		July	February					
Present	6	17	-4.5	532	350	0	0	Mixed forest
0.1	10	19	1,5			0	-0.5	Hardwood trees
10-13	10	19	1.5			0	-40	Hardwood trees
13	-4	10	-16			0	-60	Dwarf birch, permafrost
20	-4	10	-16			0	-90	Dwarf birch, permafrost
20-25	3	10	-8			0	-90	Pine forest
25-40	-4	10	-16			0	-90	Dwarf birch, permafrost
40-45	3	10	-8			0	-90	Pine forest
45-50	-4	10	-16			0	-92	Dwarf birch, permafrost
50-62	-8	10	-20			0-500	150	Unstable, thin, cold-based ice sheet
62	-8	10	-20			650	150	Unstable, thin, cold-based ice sheet
62-66	-15	3	-20			650-1200	450	Stable cold-based ice sheet
66-69	-15	3	-20			1200-1500	480	Stable cold-based ice sheet
69-80	-5	12	-15			1500-0	480-550	Warm-based withdrawing ice sheet
80-81	-8	10	-12			0-800	480	Warm-based growing ice sheet
81-83	-5	12	-8			800-0	480	Unstable, thin, cold-based ice sheet
83-84	-6	9	-15			0-500	480	Unstable, thin, cold-based ice sheet
84-85	0	12	-8			500-0	500	Stable, growing ice sheet
85-87	-6	9	-15			0-650	550	Stable, growing ice sheet
87-91	-8	5	-18			650-1200	580	Warm-based growing ice sheet
91-95	-10	5	-18			1200-1650	600	Warm-based growing ice sheet
95-105	-15	3	-20			1650-2000	600	Warm-based growing ice sheet
105-109	-15	5	-15			2000-1910	600	Warm-based ice sheet
109-115	-5	10	-8			1910-200	150	Warm-based withdrawing ice sheet
115	0	14	-9			200-0	350	Submerged, fresh water
125	6	10	-4.5			0	0	Mixed forest

Table 4. The Emissions-M scenario: temperature, annual precipitation, atmospheric CO2, northern hemisphere ice volume, relative sea level change (from present) and surface cover for Olkiluoto. (MAAT=mean annual air temperature). Data from Ikonen (2005), Ruosteenoja (2003), BIOCLIM (2003a, 2004), Hohl (2005). (Posiva 2006).

Time (ka AP)	MAAT (°C)	T (°C) July	T (°C) February	Precip (mm)	CO_2 (ppmv)	NH ice volume (10^6 m^3)	Relative sea level (m)	Surface cover
Present	6	17	-4.5	532	350	3.2	0	Mixed forest
0.1	10	19	1,5			3.2	0	Hardwood trees
0.35	14	22	6			3.2	0	Hardwood trees
2	14	22	6			0	-2	Hardwood trees
10	14	22	6			0	-29	Hardwood trees
20	14	22	6			0	-80	Hardwood trees
67	10	19	0			0	-75	Hardwood trees
90	8	18	-2			0	-78	Mixed forest
100	2	14	-9			0	-80	Dwarf birch, brush
110	2	14	-9			1.6	-83	Dwarf birch, brush
140	8	18	-2				-84	Mixed forest
145	2	14	-9				-83	Dwarf birch, brush
150	2	14	-9			0.7	-85	Dwarf birch, brush
155	2	14	-9				-83	Dwarf birch, brush
160	8	18	-2				-90	Mixed forest
167	8	18	-2				-90	Mixed forest
170	-2	14	-15				-93	Permafrost, grasses
175	-2	14	-15			17.4	-93	Permafrost, grasses
178	-2	14	-15				-100	Permafrost, grasses
180	-2	14	-15				-100	Permafrost, grasses
185	6	17	-4				-100	Mixed forest
200	6	14	-4				-100	Mixed forest

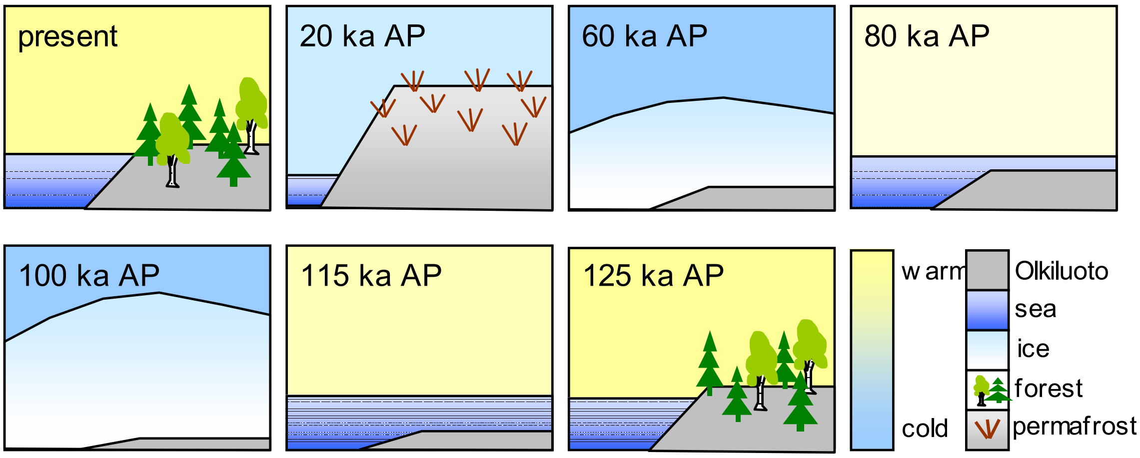


Figure 20. *Schematic presentation of snapshots for the Weichselian-R scenario (B). Climate development at Olkiluoto from present day to the end of next interglacial at 125 000 AP (Posiva 2006).*

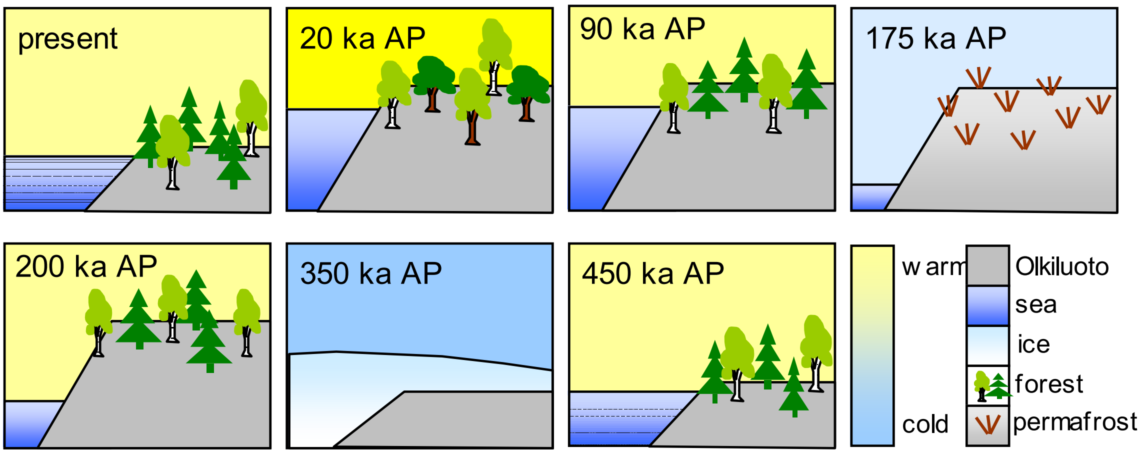


Figure 21. *Schematic illustration of snapshots for the Emissions-M scenario (D). Climate development at Olkiluoto from present time to the next interglacial at about 450 000 AP (Posiva 2006).*

The scenario D adds a low fossil fuel induced CO_2 contribution. The D scenario assumes that fossil fuel burning will contribute to a small increase in the natural CO_2 concentration in the atmosphere. It is predicted that the simulated climate, will be warmer than in the scenario C throughout the simulation. The impact of the fossil fuel contribution is larger at the beginning of the simulation than towards its end (Cedercreutz 2004).

Scenario E combines with a high fossil fuel CO_2 contribution – a total emission of 5160 GtC over the period 2000-2300 (BIOCLIM 2001). It represents a truly drastic impact of human activities on climate. Even at the end of the simulation the fossil fuel contribution is large enough to have a significant impact (Posiva 2006).

The near future

The conditions in the near future are related to the operational phase and the transient period immediately after the closure of the repository and thus relevant for safety issues. The near future climate conditions are the same for the (Weichselian-R and Emissions-M) selected scenarios.

Ruosteenoja (2003) has presented the results for the near future (2010-2350) climate forecasts at Olkiluoto using several emission scenarios of IPCC (2001). An increase in temperature is predicted during the current century. By 2070-2099 winter temperatures are estimated to rise by 3.8-10.4 °C and summer temperatures by 1.6-5.6 °C. High latitude temperature increase is larger than the global average (e.g. a high latitude increase of 8 °C requires only a global rise of 2.5 to 6.7 °C). Relative humidity tends to decrease in all seasons except winter. Climatic warming enhances ice sheet melting in the summer, whereas an increase in wintertime snow precipitation increases ice volume. Occurrences of each climatic phase in the scenarios are presented in Table 5.

Temperature change for the next 100 years is identical for both scenarios. Within the next 100 years mean annual temperatures will rise by 4 degrees reaching +10 °C. After the next century temperatures will continue to rise with a slower rate for another 250 years due to increasing atmospheric CO_2 concentration, reaching a maximum of MAAT (mean annual air temperature) of +14 °C. Mean annual, summer and winter temperature values are shown in Figure 22.

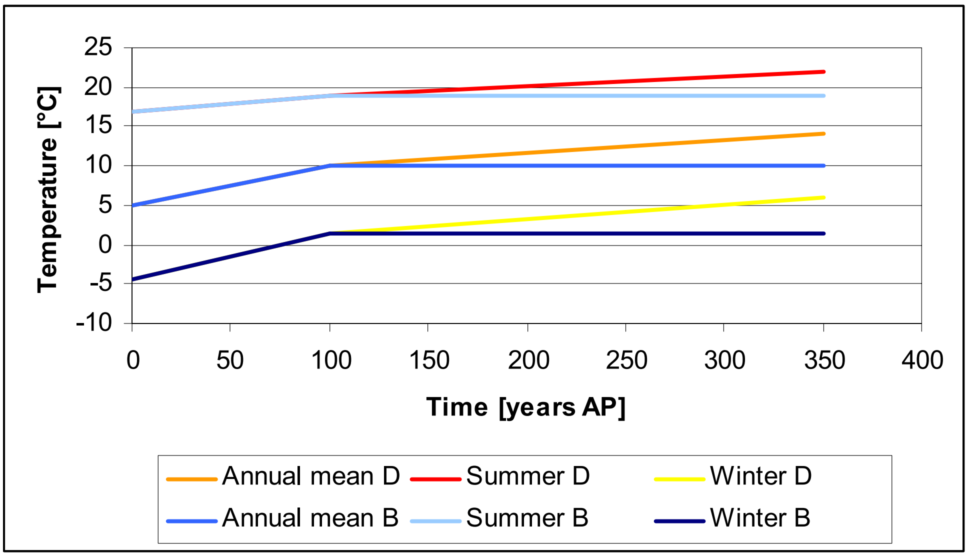


Figure 22. *The selected mean annual, summer and winter temperatures for Olkiluoto in the near future (2010-2350) for both scenarios (B for the Weichselian-R, D for the Emissions-M). The mean annual temperature is the average from winter and summer temperature data in Ruosteenoja (2003).*

Table 5. *Occurrence of each climatic phase in the scenarios (ka AP) (Posiva 2006).*

Climatic phase	**Temperate**	**Temperate/ Permafrost**	**Permafrost**	**Glacial**	**Submerged**
Scenario B	0-13 120-125	13-50		50-74 93-112	74-93 112-120
Scenario D	0-90 110-145 160-170	90-110 145-160 180-200	170-180		

Greenhouse effect and aerosols

In the view of Loutre & Berger (2000), future climate predictions should take into account the inevitable increase in human induced greenhouse gases. It concludes that owing to the warming effect of greenhouse gases, a future glaciation will be delayed and probably be milder.

The atmospheric concentration of CO_2 has increased by 31 % since 1750, with a rate of increase of about 1.5 ppm (0.4 %) per year over the past two decades. The concentrations of methane, CH_4 in the atmosphere has increased by 151 % and those of nitrous oxide (N_2O) by 17 % since 1750. Also the concentrations of halocarbons gases and their substitute compounds have increased. The radiative forcing due to greenhouse gases, from the year 1 750 to 2 000, is estimated to be 4.86 W/m^2, in total (IPCC 2001).

Human impact on the atmosphere has made our current climate significantly different from any past one, so that temperature decline at the end of the last interglacial cannot be taken as an analogue of near-future developments (Kukla et al. 2002). Furthermore, the lower insolation in next 50 000 years, CO_2 will play a relatively strong role as forcing agent in future climate development.

Water-saturated soils, especially mires and aquatic sediments are considered the primary sources of CH_4 in arctic and sub-arctic landscapes during the summer. It has been found that in wet meadow tundra, a 2°C increase in temperature at a depth of 10 cm to 20 cm CH_4 transport increases, or flux to the atmosphere by approximately 120 %. In the winter, when soils are frozen, northern lakes are the main source of CH_4 release, as their sediments maintain a positive temperature and have anaerobic conditions throughout the year. Taliks, which lie under lake sediments, are zones of thawed permafrost and are the places where methane originates in winter. Methane accumulates under the ice and is released through cracks and holes. The role of taliks in the current atmospheric CH_4 balance could be significantly underestimated. Thawed lakes that have aged a few thousands years might cover a layer of thawed permafrost, or talik, by around 100-200m, or more. Vast reservoirs of ancient organic carbon immobilized in permafrost can become available for anaerobic decomposition as the lakes evolve.

Thawing of permafrost under lakes, both onshore and offshore, may be the means by which reservoirs of methane hydrate. During the warming process, the methane gas, trapped in ice, is disturbed and rises to the surface of the lake. Some of the permafrost pockets were formed before the Holocene flooding 10 000 years ago. This is a main way in which ancient methane enters the modern chemical cycle.

Aerosols include sulphates, nitrates, organics, soot, dust and fly ash. Some aerosol particles occur naturally, originating from volcanoes, dust storms, forest and grassland fires, living vegetation and sea spray. About 10 % of aerosols derive from human activities, such as the burning of fossil fuels and alteration of natural surface cover. Anyhow, aerosols are one of the least understood influences on global climate, it is even unclear whether the net effect of aerosols is to warm or cool our planet, although the common opinion is the latter (IPCC 2001).

Scenario from present to period beyond one million years

On the basis of Boulton & Payne (1992) and Ahlblom et al. (1991), Vieno & Nordman (1999) have presented the following scenario (At present called the Weichselian-R scenario, Posiva 2006) for the Island of Olkiluoto (Figure 23).

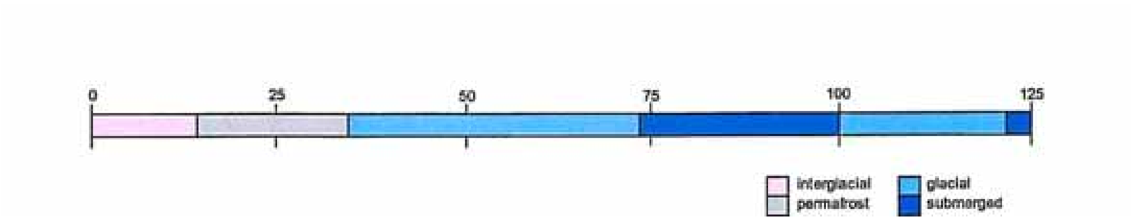


***Figure 23**. Future scenario (The Weichselian-R scenario, ka AP) for Olkiluoto based on Vieno & Nordman (1999) in Cedercreutz (2004).*

First hundred years

Surface conditions at Olkiluoto site will change only slightly in the next hundred years as the climate responds to greenhouse gas emissions and other anthropogenic influences. In Northern Europe the annual average temperature is estimated to increase by 2-3 ^{o}C and precipitation by 10-20 % (Kattenberg et al. 1995). Relative sea-level will remain essentially constant, as any rise in eustatic sea-level as a consequence of global farming will be offset by continued land uplift following isostatic depression during the last glaciation (Vieno & Nordman 1999).

Period 100 to 10 000 years

There is a close correlation between the past climate in the northern hemisphere and the intensity of insolation reaching the upper atmosphere (e.g. Berger 1988). The pattern of astronomical perturbations indicates that insolation in the northern hemisphere has been decreasing since a maximum at about 3 000 years before present. This will lead to a decrease of temperatures. Without the influence of greenhouse gases, the effects would be evident in a few hundred years. With assumed patterns of greenhouse gases emissions, however, there will an initial rise in temperature that will compensate for the decrease in insolation. At Olkiluoto, different climatic and tundra conditions ranging from those prevailing today in Central Europe to the tundra conditions prevailing today in the far north of Finland may occur within the next 10 000 years (Vieno & Nordman 1999).

Period 10 000 to 100 000 years

The global climate will cool during this period and an ice sheet will develop to cover Fennoscandia, with ice thickness of up to two or three kilometers (e.g. Berger et al. 1991, Ahlbom et al. 1991, Boulton & Payne 1992). Significant global cooling will start at about 15 000 years. The temperature and sea-level fall and continuous permafrost will develop. Beyond 20 000 years, continued cooling will eventually lead to the amalgamation of valley glaciers and nucleation of the next Fennoscandian ice sheet. A subsequent period of rapid retreat will lead to a series of freshwater lakes interspersed with marine intrusions (Goodess et al. 1991). This second glacial period will persist until about 120 000 years, at which time there will be a return to the climatic conditions similar to the present day (Vieno & Nordman 1999).

Conditions during the glacial cycle will increase the hydrostatic pressure and change the composition and rate of groundwater recharge. Stress relief following removal of glacial load could cause movements along major fracture zones. However, it has been estimated that shear movements of more than 10 cm in fractures intersecting the deposition hole would be required to break the canister (SKB 1992b). In turn, such shear movements would require an earth-quake of magnitude 8.2 on the Richter scale on a postglacial fault one kilometer away from the repository (La Pointe et al. 1997). The Olkiluoto area is remote from the region in the northern Fennoscandia where conditions are suitable for large postglacial faults to form (Saari 1992, Kuivamäki et al. 1998). This earthquake activity would exceed the maximum activity estimated to have occurred in the region of postglacial faulting during previous glacial retreats (Muir-Wood 1989).

Continuing isostatic uplift over the 100 000 years will cause a fall in relative sea level up to 40 meters at Olkiluoto (Påsse 1996). The isostatic uplift will drive groundwater flow downwards. The brackish sulphate-rich and fresher groundwaters presently above the repository depth will start to replace the brackish saline groundwaters currently at repository depth (Löfmann 1999a-b).

At the coastal site, the more saline and ammonium-rich groundwaters will probably be more corrosive over the long term towards cement-based seals than the groundwaters at the inland sites. However, the cement-based seals will maintain a low hydraulic conductivity for tens of thousands to hundreds of thousands years (Alcorn et al. 1992).

Period 100 000 to 1 000 000 years

Beyond 100 000 years, the timing of changes in climate and near-field evolution become even more difficult to predict. During this period, repeats of the glaciation-deglaciation events related to climate change can be expected. The magnitude of these climatic events will vary. Only some long-term change in the climatic system could bring this cycling to a close, and such a change is not expected in this period.

Infiltration of melt water and sea-level changes associated with future glaciations will introduce further compositional layers of dilute and saline water to the groundwater systems. The exact changes will depend on the position of the sites with respect to the evolving shoreline. Olkiluoto is certain to receive a period of saline groundwater recharge during higher sea levels (Vieno & Nordman 1999).

Period beyond one million years

The growth of mountain ranges, changing oceanic circulation patterns, and/or changes in distribution of oceanic and continental crust could bring an end to the Quaternary climatic pattern of glacial-interglacial cycling (Goudie 1992). Natural fluctuations in atmospheric CO_2 also appear to have an important control on climate in the past (Berger & Loutre 1997). The timing of any change is highly uncertain but, over the next 10 to 100 million years, it is likely that interglacial-glacial cycling will cease (e.g. Wilmot 1993).

Present climate scenarios in Sweden

SKB has one base scenario, which consists two variants (SKB 2006): 1) a repetition of the Weichselian, 2) warming greenhouse scenario and additional scenarios, colder scenario (more favourable for permafrost growth and a scenario with ice sheets thicker than during the Weichselian, resulting a larger maximum hydrostatic pressure at repository depth).

5 QUATERNARY STRATIGRAPHY OF THE SEA-BOTTOM SEDIMENTS

The geology of the bedrock is of utmost importance with respect to the character and distribution of unconsolidated sediments in the Baltic Sea. Besides being a major source of material deposited during and after glaciation, the old bedrock topography contributed much to the distribution and evolution of glacial deposits. A factor which has considerably affected the distribution of late glacial and post-glacial sediments is the differential uplift of the Baltic Sea basin, together with the multiple transgressions and regressions occurring during the various phases of the evolution of the Baltic Sea (Flodén & Winterhalter 1981).

5.1 Late-glacial sediments

Baltic Glacier Lake 13 500 - 10 300 years BP

The Baltic Glacier Lake was dammed up by the ice sheet, and it was first a smallish lake at the southern end of the present Baltic Sea basin but grew as the edge of the ice receded, reaching its greatest extent about 11 500 years ago. Baltic Lake was a freshwater lake built against the retreating ice margin and it had an outlet via Öresund straits (Agrell 1976, Mäkiaho 2005).

With the waning of the ice sheet the intensified crustal uplift together with the still rather slow eustatic sea-level rise must eventually have severed off any possible connections with the world ocean. To the north and northwest the Baltic Glacier Lake was bounded by the retreating ice margin. The melt waters from the ice forced their way westward, possibly through the Danish Straits (Kolp 1965). The subsequent retreat of the ice sheet opened a new outlet across the lowlands of central Sweden at Billingen causing the lake level to drop suddenly 26-29 m establishing a connection with the ocean and the contact with the North Sea changed the lake into a sea. This stage about 11 000 years ago is called the Yoldia Sea, after the then common mussel *Portlandia Yoldia arctica* (e.g. Flodén & Winterhalter 1981).

Yoldia Sea phase 10 400 - 9 500 years BP

The Yoldia phase began when the margin of the continental ice retreated behind the mountain of Billingen in central Sweden. Salt water intrusion through the widened channel across central Sweden rapidly increased the salinity of the Baltic Sea water. Due to the proximity of the ice margin, the northern part of the Yoldia Sea exhibited arctic conditions as witnessed by a very scare arctic biota and the deposition of varved clays (e.g. Flodén & Winterhalter 1981).

Varved silty clay deposited on top of till, sand and gravel. The varve-thickness diminished upwards due to transformation of the depositional environment from proximal to distal relative to the glacier margin. In general, the thickness of the varves could represent the distance to the glacier, i.e. diminution of thickness of varves was in response to the glacier’s retreat (Donner 1977), but the lowering amount of suspended matter alone could cause diminished varve thickness.

Synchronously with the deposition of varved clays in the north, homogenous clays, stained black by amorphous iron sulphides, were being deposited in the southern part of the sea. This is an indication of a somewhat higher production of biogenic matter and more uniform conditions of sedimentation (Winterhalter et al. 1981). The spring and early summer proportion of varves typically consists of fine-grained sand. The material changes sharply upward to silt/clay that was deposited during autumn and winter (Rantataro 2001).

Towards the end of the preboreal Yoldia Sea stage crustal uplift, being more rapid than the eustatic sea-level rise. Restricted inflow of saline water thereby lowered the overall salinity of the water by the end of the Yoldia Sea phase and the ice vanished from Finland but was still present on the mainland in Sweden. The progressing crustal uplift finally superseded the eustatic sea-level rise cutting the oceanic connection. The Baltic Sea basin isolated from the ocean changed into the fresh water lake, the Ancylus Lake (e.g. Flodén & Winterhalter 1981).

Ancylus Lake phase 9 500 - 7 500 years BP

The Ancylus Lake had its first outlet through the Svea River in southern Sweden (Munthe 1927). The Ancylus Lake phase lasted 1 000 - 1 500 years, and the ice sheet margin retreated quickly after the formation of the Salpausselkä end moraines (Björk 1995). Thus part of Finland was ice-free since the beginning of Ancylus time. Due to differential land uplift, greater in the north than in the south, the water level of the Ancylus Lake tilted, transgressing in the south, and finally established a new contact with the rising ocean through the Danish Sounds (e.g. Flodén & Winterhalter 1981).

The Ancylus clay is composed of nearly homogenous sulphide-stained clay deposited on a calm lake environment. Sulphide rich clays indicate anaerobic conditions in the bottom of the sea (Kotilainen & Kohonen 2005). Poorly developed lamination is visible in the proportion of the sediment without sulphide matter. In the lower Ancylus clay sequence, sulphide matter occurs as layers of separated aggregates embedded into laminated clay. The upper part of the Ancylus unit is more homogeneous and is characterized by disseminated sulphide spots and/or thin sulphide concretions (Rantataro 2001, 2002). According to Winterhalter (1992), this uppermost (0.5-1.0 m thick) sub-unit of Ancylus clay is found almost everywhere in the Baltic Sea area. Based on the sounding, the total thickness of the Ancylus clay sequence is generally between 2-3 m, but attains 8-10 m in deep basins (Rantataro 2001).

5.2 Postglacial deposits

Litorina Sea 7 500 - 3 500 years BP

Approximately 8 000 years ago sea levels rose rapidly in the oceans as the ice age was finally drawing to a close and Lake Ancylus became a sea again. This stage is called the Litorina Sea (Figure 24). The exchange of water with the North Sea was more efficient due to wider straits connecting the bodies of water, and the salinity of the Litorina Sea was almost at the same level as in the North Sea.

Sediment cores from various parts of the Baltic Sea indicate that the end of the Ancylus lake stage and the beginning of the Litorina Sea stage is marked by exceptionally sharp litostratigraphic boundary (Jerbo 1961, Ignatius et al. 1968, Blažčšin 1976a). However, the Ancylus/Litorina transition is gradational if both of the Ancylus clay and Mastogloia clay contain black staining with disseminated sulphide aggregates as vague layers. The

Mastogloia stage began about 500-1 000 BP. The thickness of the Mastogloia unit is generally less than 0.5 m (Rantataro 2001).

The transition from the late-glacial Ancylus clay to postglacial clay is quite difficult to specify if the transitional Mastogloia clay between them is absent. A thin and obviously erosion and/or non-depositional transition-zone containing coarser material is occasionally found between the Ancylus and Litorina clay, producing a clear reflection on the sounding profiles. The Mastogloia unit is a poorly identified phase occurring before the actual Litorina Sea period (Ignatius et al. 1981).

Post-glacial deposits, characterizing the Litorina Sea phase, are composed of olive-green muddy clay, deposited 7 500-- 3 500 years BP (Eronen 1974, Ignatius et al. 1981). In the lower part of the Litorina clay, there are two or three sections of laminated, gyttja-banded clays generally found separated from each other by clay layers. In the upper part of the Litorina clay, the laminated texture disappears and the overall appearance is generally homogeneous, although faint macroscopic layering is ubiquitous. Occasionally, pyrite concretions are found in the Litorina clay in the eastern part of Gulf of Finland and also in the Bothnian Bay (Georgala 1980). Based on soundings (Rantataro 2001), the Litorina unit thickness is 5-8 m in the near shore depressions, but attains a thickness of 10 m in the deeper offshore. The investigation cores taken elsewhere in the Baltic Sea show that the Litorina unit has generally lost its upper surface by erosion, thus the original thickness has been greater (Nuorteva 1994).

Sub-recent and /or recent mud around 3 500 BP - to present

The sub-recent material began to deposit about 3 500 years BP (Eronen 1974, Ignatius et al. 1981). The sub-recent and recent sediment consists of organic-rich black mud (gyttja). Very often, biogenic matter is found as un-deposited organic matter. The topmost surface (0.2-1.0 cm thick) is mostly oxidized with reddish-brown colour, but below this layer anaerobic conditions prevail due to the decomposition of organic matter. Sub-recent/recent sediment generally contains a high amount of hydrogen sulphide gas with its strong and characteristic smell. Sub-recent and recent clay/gyttja is deposited in several tranquil areas onto the older sediment after a sharp erosive contact, but in deep offshore basins the gyttja layers are more continuous (Rantataro 2001, 2002). A sub-recent/recent sedimentary unit is deposited on a lag-layer, consisting of sandy mud or fine-grained sand, in the open-sea basin that are deep enough or in the near-coast depressions that are sheltered against erosive currents and waves by island (Rantataro 2001).

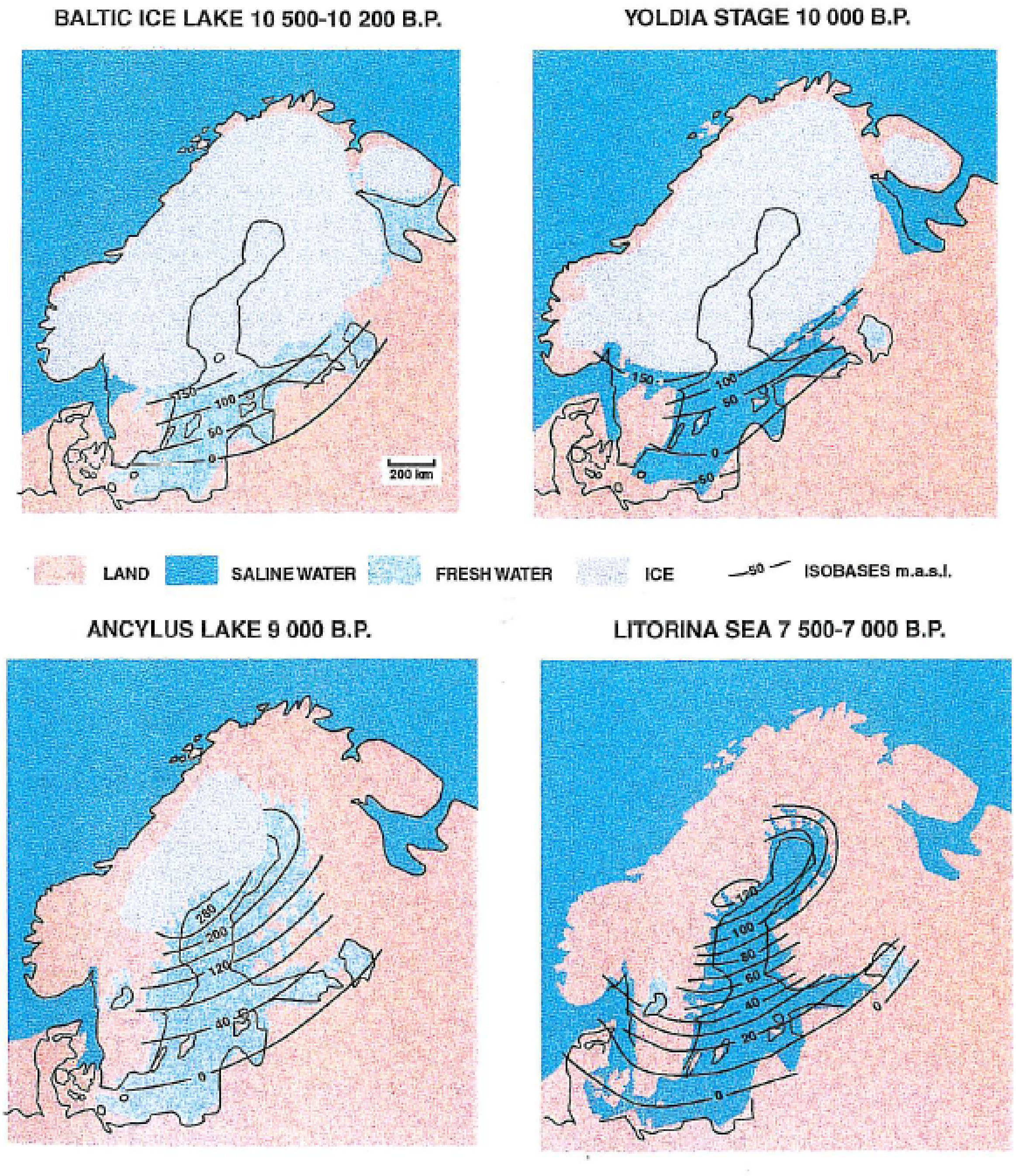


__Figure 24.__ Evolutionary stages of Baltic Sea according to Eronen (1988, 2005).

5.3 Sedimentation

Due to the differential land uplift there is a marked difference in the sedimentation conditions between the southern Baltic Sea and the northern marine area. The sedimentation conditions in the southern part are rather stable, since sea level fluctuations have been rather small for a considerable time span. In the northern part, however, the continuous regression has brought new sea floor areas into the regime of erosion processes (Winterhalter et al. 1981). The distribution of various sediments depends on a number of factors, thus it can differentiate between at least five different zones of sedimentation (e.g. Pratje 1948, Gudelis 1976):

- Coastal and accumulation zone, especially noteworthy in the southern and south-eastern parts of the Baltic Sea

- Relict clastic deposits, e.g. exposed glacial drift, with only minor evidence of reworking of the uppermost layer of sediments

- Zone of retarded sedimentation, also of non-deposition; increasing exposure to wave- and current – induced water motion following crustal uplift or eustatic sea level change

- Zone of erosion, not to be confused with local erosion in deep channels and on the flanks of topographic highs, the latter being a case of local increase in current velocities due to topographic flow restrictions

- Zone of sedimentation (silts, muds) generally located well below the permanent halocline and the "wave-base" level including internal waves in the permanent halocline (50-80 m)

The sediment accumulation rate (SAR) or net sedimentation is an essential parameter in contamination or monitoring studies and in budget calculations (Wulff et al. 1993, Kankaanpää et al. 1997, Vallius & Leivuori 1999, Emeis et al. 2000, Isosaari et al. 2002). The vertical distribution of artificial radionuclides Cs-137 and Pu-239 and Pu-240 and the naturally occurring radionuclide Pb-210 in sediments can be used in sediment dating and SAR estimations (Mattila et al. 2006). These radionuclides have high concentration factors in sediments in the brackish water environment of the Baltic Sea (IAES 1985, Helcom 2003, Ikäheimonen 2003).

Due to heterogeneity of soft sediment deposits, variation in the SAR inside sedimentation basin and between closely-situated positions can be large. Sediment layers formed over thousands of years may vary considerably depending on the bottom topography (Winterhalter 1972, 1992, 2001). The thickness of more recent soft sediments layers inside sedimentation basins may also indicate large differences in accumulation (Kankaanpää et al. 1997, Vallius & Leivuori 1999, Perttilä et al. 2003). In the Baltic Sea the main factors effecting for the erosion, transport and accumulation of the sediments are (Kotilainen & Kohonen 2005):

- Grain size
- Stream velocity
- Water depth
- Topography
- Distance from the mainland
- Water masses, climate
- Ice cover
- Base production
- Bottom fauna and
- Land up-lift

Mattila et al. (2006) have studied an average bulk SAR at 69 stations and 99 cores from the Baltic Sea during the years 1995-2003. SAR values varied widely; between 60-6160 g/m^2/year. The highest SAR values were observed in the northern part of Bothnian Sea, river estuaries and in the eastern part of the Gulf of Finland. At the Bothnian Sea the median SAR values were two, three and seven times higher than at the stations of the Bothnian Bay, Gulf of Finland and Baltic Proper, respectively. Near Olkiluoto basin SAR value 1810 g/m^2/year was estimated. The more exact data on sedimentation rates near the offshore of Olkiluoto is lacking.

The large variation in SAR is due to several factors, although generally the most important factors affecting sediment accumulation in these relatively shallow areas in the Baltic Sea are water depth and bottom topography (Brydsten 1993, Winterhalter 1972, 1992, 2001). There were clear differences between the studied sea areas, such as in the surface areas of accumulation bottoms, the influence of transportation and erosion processes on bottom areas, the loading rivers and the amount of primary production (Voipio 1981, Helcom 2002, 2003, 2004). For example, in the Gulf of Bothnia over 50 % of the bottoms are influenced by transportation and erosion processes (Winterhalter 1972, Brydsten 1993), providing considerable sedimentary materials from shallower areas to accumulation area. According to Mattila et al. (2006) the comparison of different results used, the combination of several estimation methods is recommended in sediment dating and in estimation of SAR, especially in areas with substantial heterogeneity.

5.4 Quaternary stratigraphy of the sea sediments at the Olkiluoto offshore

In general, the sea-floor deposits present a very fragmentary pattern in the surroundings of the Olkiluoto Island. The most common Quaternary sediment on Olkiluoto offshore bottom is till (30-40 %) covered by post-glacial clays, mainly Ancylus clays (Figure 25) (Rantataro 2001). The amount of exposed bedrock or sedimentary rock area on the sea floor is also 30-40 %.

Seven different geological units were interpreted from survey of Olkiluoto sea-floor based on acoustic-seismic sounding data and related calibration sampling (Rantataro 2001, 2002, Posiva 2003). However, caution must be used in interpreting them, because it is well known that acoustic boundaries do not always coincide with stratigraphical boundaries (Thorslund & Axberg 1979). Different sedimentary units (from bottom to top) deposited on Precambrian basement and sedimentary rock are illustrated in Figures 26 and 27 as following:

- Till, sand and gravel
- Washed surfical/erosion remnant sand (lesser extent)
- Proximal varved clay
- Glacial silt or clay (lesser extent)
- Ancylus (varved) clay and sulphide clay
- Litorina clay
- Recent black mud and /or clay

All investigated sediments offshore from Olkiluoto were deposited during (late-glacial) or after (postglacial) the retreat of the last continental ice. The shallow water environment surrounding Olkiluoto has a maximum depth of 30 m, but during the retreat of last Pleistocene ice sheet, about 11 000 years ago, the depth of water was over 150 m due to lack of infilling sediments and depression of ground surface (Kotilainen & Hutri 2003).

The lowermost layer of the Quaternary sequence in Olkiluoto region consists of till, sand or gravel deposited directly on either Precambrian basement or sedimentary rock. The deposition lies conformably on the basement usually as 2-4 m thick unit whereas on the sedimentary rock area, as much as 10-15 m thickness can be found (Rantataro 2001).

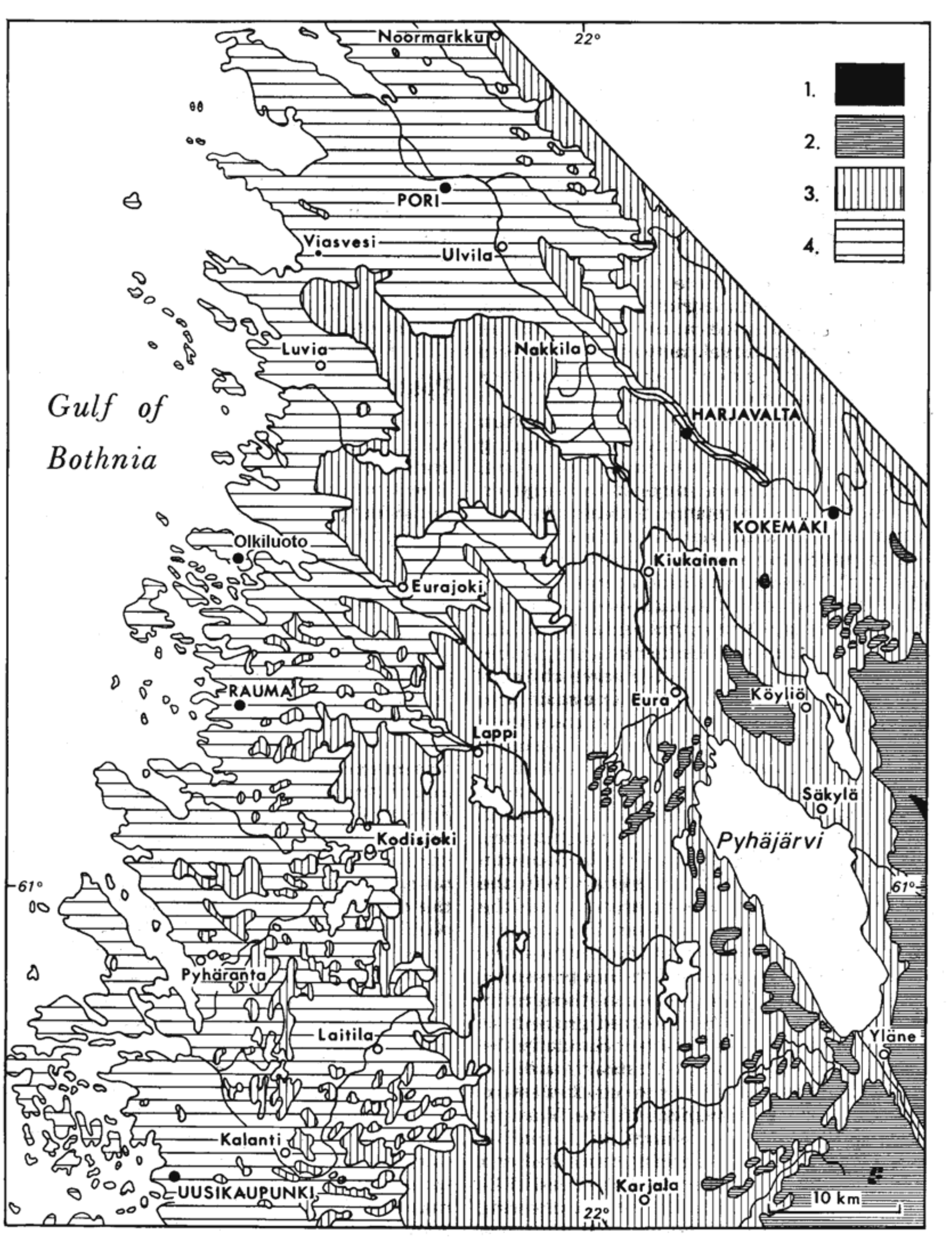


Figure 25. *Holocene shoreline displcement in areas east from Olkiluoto. (1) Areas emerged from the Yoldia Sea, (2) areas emerged during the Ancylus lake phase, (3) areas emerged during the early Litorina phase, (4) areas emerged since 3000 BP (Tikkanen 1981, Mäkiaho 2005).*

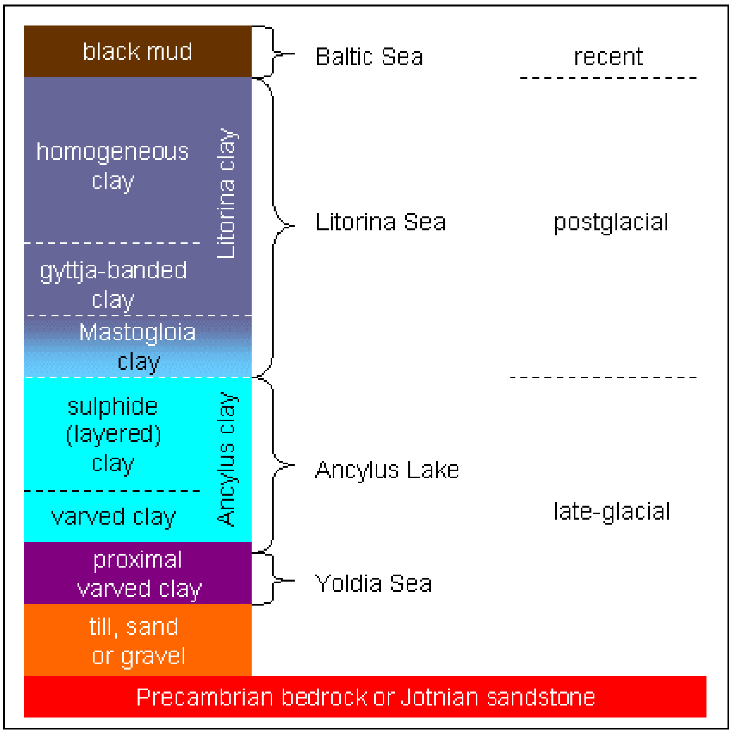


Figure 26. *Conceptualised sea bottom sediment stratigraphy (Posiva 2003).*

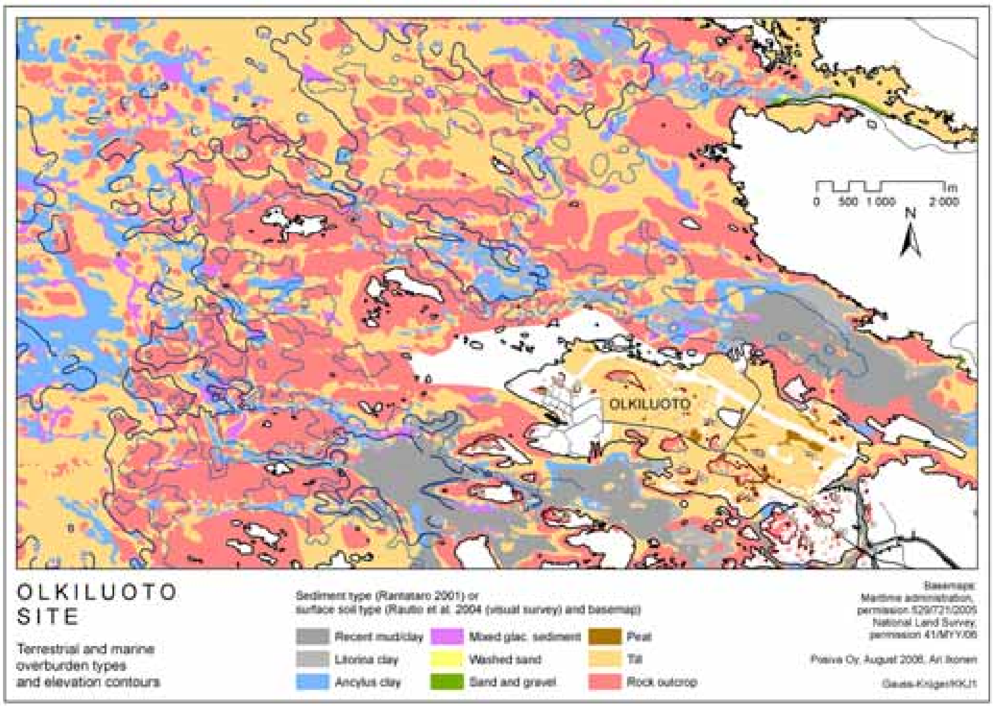


Figure 27. *Topography and terrestrial and marine overburden types in the Olkiluoto Island and surrounding sea areas (after Rantataro 2001 and Rautio et al. 2004).*

Washed surficial/erosion remnant sand, some centimetres thick, accumulates on the top of older sediments due to wave/current action. As a result of land uplift in Olkiluoto area, this unit is continually providing material for the erosion zone (Ekman & Mäkinen 1996). Erosion forces disturb fine-grained particles and re-sedimentation will take place in a calm environment such as a deep basin or sheltered bay. As a consequence of this washing effect, heavier particles stay in place as erosion remnants (Rantataro 2001).

Ancylus phase varved clay is usually deposited conformably on the underlying Precambrian basement and sedimentary rock, till/sand/gravel and proximal clay. The Litorina clay as well as any younger sediment is deposited as a basin-fill type, differing from older clay sediments that are deposited more or less conformably over the substratum (Winterhalter 1992).

Late glacial varved and sulphide-rich Ancylus clays are typically present in substratum basins and outside the archipelago in the open sea areas west of Olkiluoto. Postglacial Litorina clay/mud as well as recent/sub-recent mud/clay covers the sea-floor in sheltered near-shore basins and in the inner archipelago. Present day areas of active sedimentation are situated in the inner archipelago south/southwest and northeast from Olkiluoto (Rantataro 2001).

Acoustic seismic data

Sea-bottom sediments have been studied by acoustic-seismic methods offshore from Olkiluoto from than 350 km and six samples have been taken with vibrohammer core. The distance of the track lines were about 300-500 m in east-west direction. The area of soundings covered about 150 km^2 (Rantataro 2002). The aim of the soundings were to get as good picture as possible of the seafloor conditions in Olkiluoto sea area and to receive by acoustic-seismic methods material in a scale which makes it possible to interpret lineaments in the bedrock. Further, the aim was to collect material of possible leaking areas of the groundwater (Rantataro 2001). In the additional survey, the aim was to get extra knowledge due to earlier work in year 2000 concentrated in immediate surroundings of Olkiluoto (Rantataro 2002). In the Figure 28 is presented the locations of two acoustic seismic lines and in Figures 29 and 30 the stratigraphy of the sea bottom sediments using the aquistic seismic method near and more far of Olkiluoto offshore.

Considering the characterisation of the seabed sediments, uncertainties are associated with underlying units containing clay-rich layers, which result in difficulties in interpreting echo sound surveys. In some basins, and also in the areas where the sea floor is covered by sandy/gravel lag, Ancylus clay can be indistinguishable from other related sediments in such surveys. Sometimes recent mud and sub-recent mud/clay sequences are nearly indiscernible, and are present as a transparent layer with faint stratification; the resulting structure is similar to that shown by the Litorina clay, with which it could be confused (Posiva 2005).

It was not possible to investigate the structure of the Precambrian rock basement at Olkiluoto area with the acoustic-seismic equipment used because of poor penetration of the sound waves into unbroken bedrock. The sound pulse could penetrate some distance into a more porous and soft sedimentary unit, likely the Jotnian sandstone. During the interpretation of acoustic-seismic sounding data it became clear that area covered by sedimentary rock could be much nearer to Olkiluoto, only 5-6 km (Rantataro 2002, Kuivamäki 2001) than previously estimated.

The sedimentary rock areas are generally large and are situated in basement depressions and are thus sheltered from erosive forces. Between these areas, till, sand and gravel is deposited on the bedrock surface. It is likely, that sedimentary rock coverage area was found about 12 km west of Olkiluoto where the surface topography is very gentle. Here the sedimentary rock unit is very thick and internal reflectors ubiquitous. The topography of rock surface varies gently in the areas that are interpreted to be covered by sedimentary rock (Rantataro 2002).

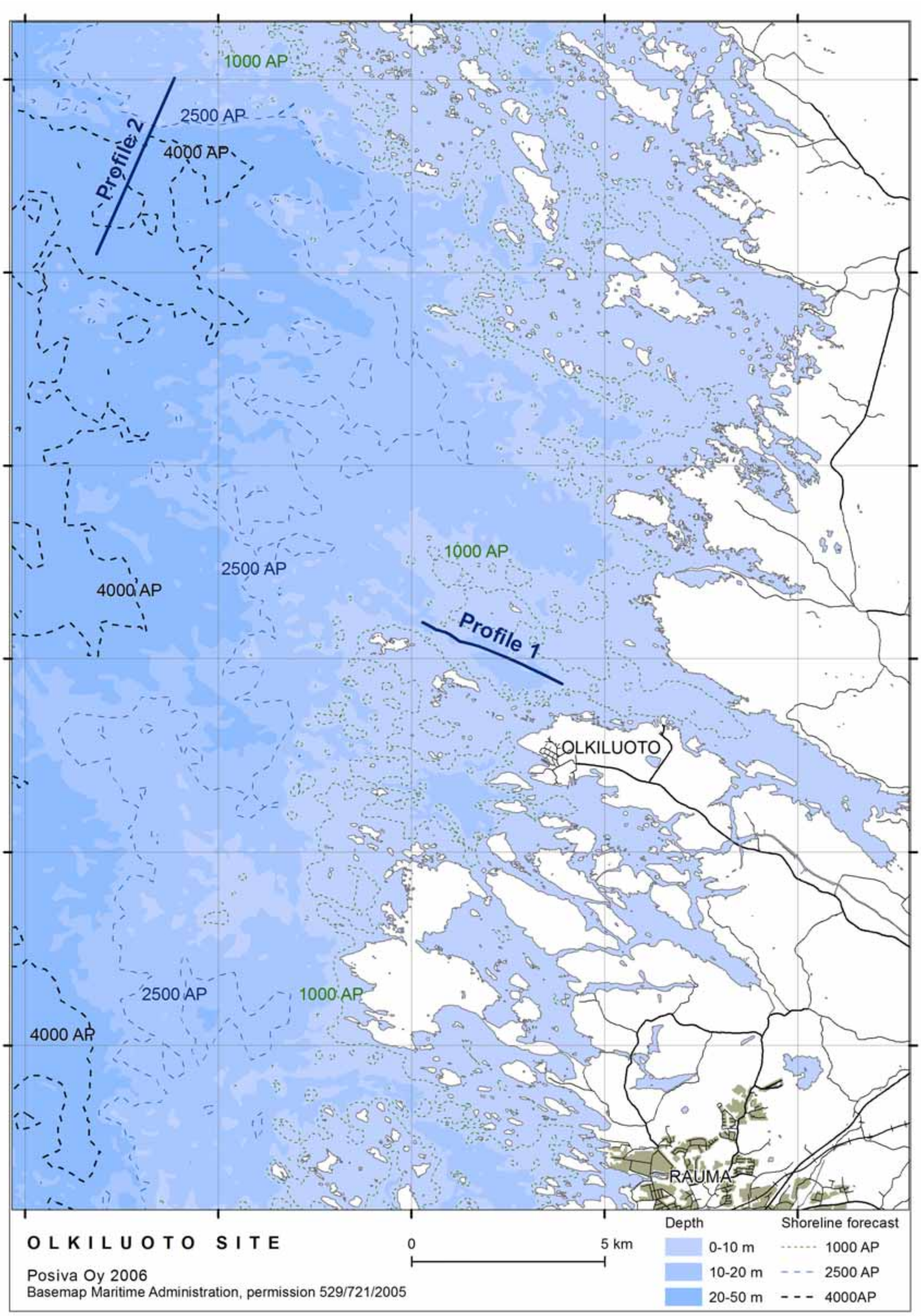


Figure 28. *The locations of the two acoustic seismic lines near (Profile 1) and more far (Profile 2) of Olkiluoto offshore. The shoreline forecast is by Mäkiaho (2005). (Data of the profiles according Rantataro 2001, 2002).*

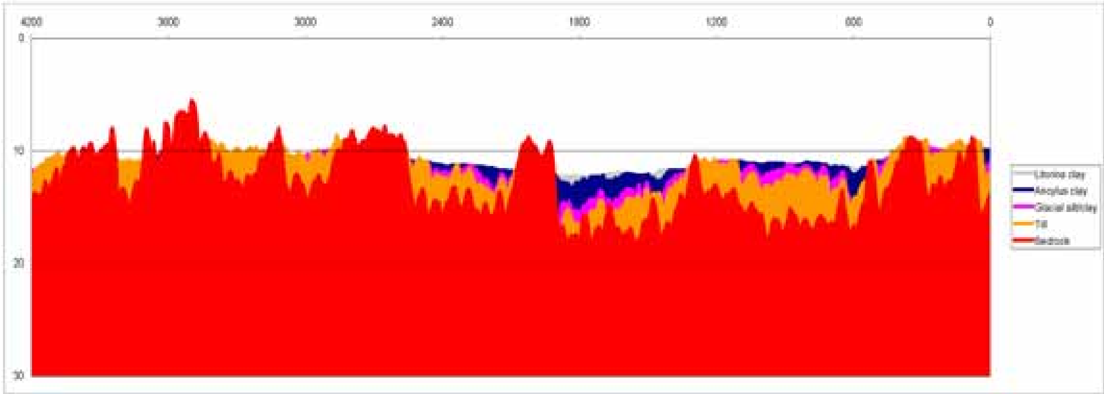


Figure 29. *The stratigraphy of the Profile 1 (Part of the line 11291150, according to Rantataro 2002).*

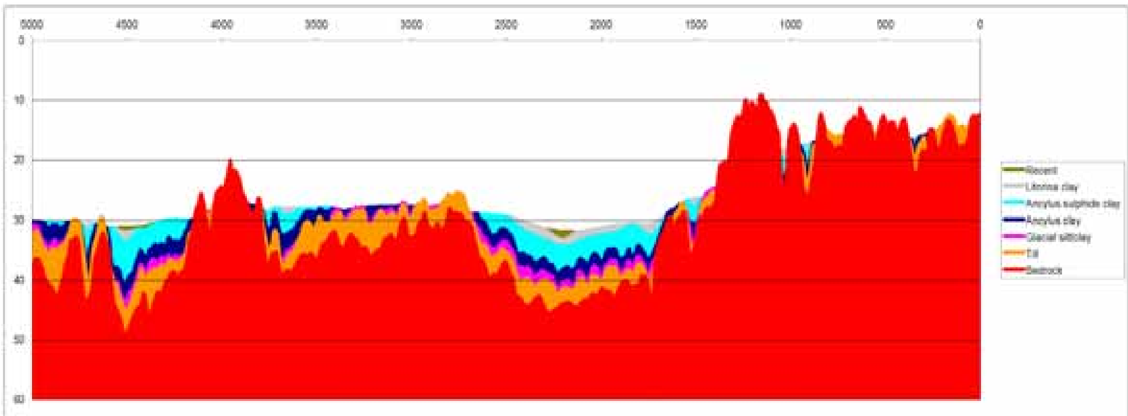


Figure 30*. The stratigraphy of the Profile 2 (Part of the line 19181925, according to Rantataro 2001).*

5.5 Postglacial palaeoseismicity

Kotilainen & Hutri (2003) have used both, high resolution echo-sounding data and sediment cores to define the occurrence of disturbance structures in submarine sediments in the vicinity of the Olkiluoto Island. The survey lines (about 25 were situated approximately 500 m from each other. All echo-sounding profiles, about 350 km, covered approximately 100 km^2.

Regional seismicity in the southern Finland contains two zones of higher seismic activity: the Åland archipelago-Paldis-Pskov zone and the Southern Bothnian Bay-Ladoga zone. Both zones might continue north-westward to Sweden (Saari 1998). Olkiluoto is located near the Åland-archipelago-Paldis-Pskov zone running from Åland Archipelago to Estonia.

The stratigraphy position of the sedimentary faults reveals that these disturbances were caused by palaeoseismicity. The age distribution of the faults suggest one or several earthquakes in a short time about 10 700 BP. The similar distribution pattern as of the faults, a large number and limited time-span failures (e.g. slumps) in the study area supports this assumption. Most of these features occur along south- or southwest-facing slopes, close to or directly above the two bedrock fracture zones. This close spatial relationship between the sediment instability features in the early Holocene deposits and the bedrock fracture zones may suggest that they were formed by reactivation of the fracture zones. In the area, in the form of rapid bedrock block movements, could also have affected slope failures.

Other commonly known triggering mechanism for these features is offshore instability processes and glaciotectonic deformation. Wave and bottom current activity could cause erosion and sediment removal. According to observations made from sediments of the Gulf of Bothnia, wave activity in open sea prevents sedimentation of clay shallower than 50 m (Ignatius et al. 1980). During the suggested neotectonic event(s) (10 700 BP) the study area was submerged more than 150 m, and it is unlikely that the activity erosion was involved with these failure processes. Slope failures could also occur during deposition, when the shear strength of the sediment was exceeded.

Glaciotectonic deformation is also known to cause such features (Hart & Boulton 1991), so that disturbance structures in glacioaquatic sediments may also related to glacial surges (Elverhøi 1984). In general, the varve thickness tends to thin upwards which supports the belief that ice edge was retreated across the region. Other factors might affect the varve thickness too, e.g. wind and bottom topography. In this case, since the disturbance structures occur above or the upper part of these distal varves, it is probable that the edge of Weichselian ice sheet had already retreated form this area (Strömberg 1990). Based on the retreating rates it can be estimated that at the time of neotectonic event(s) the edge of the ice sheet was at least about 50 km away (Donner 1995).

One possibility is that the faults and gravitational disturbance deposits were induced by gas escapes deep from the bedrock. However, no postglacial faults have been found deforming Quaternary deposits in central or southern Finland, although large areas of the land have been below the highest shoreline, so that evidence may have been destroyed by wave erosion. In the study area, deglaciation took place from a SE to NW direction (Atlas of Finland 1990, Stömberg 1990) and the average directions of horizontal stresses measured in boreholes at Olkiluoto have an orientation of about W-E with some variations (Posiva 2005, Ljunggren & Karlsson 1996).

The known earthquakes around the study area are located along lineaments (Saari 2000) parallel to the class II fracture zones (Kuivamäki 2001). The approximated location accuracy of the historical data is generally 10-20 km or worse, for the one instrumentally detected earthquake accuracy is 5-10 km. The recent GPS measurements at Olkiluoto area have even revealed very small movement in one fracture zone which also strikes SE-NW (Ollikainen & Kakkuri 2000). The horizontal crustal velocity vectors in south-western Finland are parallel to this direction (Milne et al. 2001). Thus, the faulting reported here can also be a consequence of the Mid-Atlantic Ridge push (Kotilainen & Hutri 2003).

5.6 Palaeohydrological evolution

Changes in climate and geological environment have had a significant effect on local palaeohydrogeological conditions at Olkiluoto site. The boundary conditions for the palaeohydrological evolution can be determined from the understanding of recent geological history, since the Weichselian glaciation. Based on earlier groundwater geochemical studies (e.g. Pitkänen et al. 1994, 1996, 1999a, b, 2004) it is clear that the Holocene history of the Baltic Sea, in particular, has had a major effect on groundwater compositions at Olkiluoto.

Total Dissolved Solids (TDS) refers to the amount of dissolved solids (typically various compounds of salts, minerals and metals). During the main part of the Litorina stage the TDS in the seawater was about 4 ‰ higher than in the modern Baltic Sea at the Finnish coast (Donner et al. 1999). Since that time the salinity of the seawater has been reduced steadily to its current value of about 6 ‰ off Olkiluoto Island. Table 7 presents estimates of the composition of glacial melt water during the Weichselian deglaciation and Litorina seawater, the latter estimate is based on the compositions of modern sea waters. The estimates for the compositions of present Baltic Sea and mean global ocean water are for surfical seawaters (Posiva 2005).

The interpretation of chemical and isotopic data indicates that there are at least five end-member water types influencing current groundwater composition at the site. They originate from different epochs ranging from modern times, through former Baltic stages to preglacial times (Pitkänen et al. 1999):

<u>Modern</u>

- Meteoric water infiltrated during terrestrial recharge
- Seawater from the Gulf of Bothnia (0-2 500 BP)

<u>Relic</u>

- Litorina Seawater (2 500- 7 500 BP)
- Fresh water prior to the Litorina Sea stage containing glacial melt water (7 500 – 10 000 BP)
- Saline water (brine) intruded and/or formed under the influence of hydrothermal activity (pre-Quaternary, probably early Phanerozoic to Precambrian

Table 6. *Estimated Quaternary glacial melt water and Litorina compositions with inferred Baltic Sea and mean ocean water composition (Pitkänen et al. 1999).*

Parameter	**Glacial water**	**Litorina Sea**	**Baltic Sea**	**Ocean water**
T (^{o}C)	1.0	10.9	8.5	20.0
O_2 (mg/l)	7.2	6.6	7.2	4.3
pH	5.8	7.6	7.7	7.5
Density (g/ml)	1.000	1.008	1.002	1.030
HCO_3 (mg/l)	0.16	92.5	78.7	144.2
SO_4 (mg/l)	0.05	890	450	2540
PO_4 (mg/l)	0.0003	0.06	0.02	0.22
Ntot (mg/l)	0.19	0.27	0.21	0.5
Cl (mg/l)	0.70	6500	3025	19550
F (mg/l)	0.00	0.49	0.27	1.3
Br (mg/l)	0.001	22.2	10.3	67
NO_3 (mg/l)	0.07			
SiO_2 (mg/l)	0.01	1.84	0.58	6.61
Fe_{tot} (mg/l)	0.0001	0.002	<0.01	0.002
Al (mg/l)	0.0001	0.002	<0.01	0.002
Na (mg/l)	0.15	3674	1760	10860
K (mg/l)	0.15	134	66	391
Ca (mg/l)	0.13	151	82	412
Mg (mg/l)	0.1	448	219	1310
Mn (mg/l)	0.0	0.0	<0.01	0.0
Sr (mg/l)	0.0001	2.68	1.20	8.24
Li (mg/l)	0.0	0.07	0.04	0.18
Charge Balance (%)	1.03	0.97	2.13	0.27
δ^{2}H (‰SMOW)	-166.0	-37.8	-60.8	-30.0
δ^{13}C (PDB)	125.0	-1.0	-1.68	-1.0
δ^{18}O (‰SMOW)	122.0	-4.7	-7.55	-4.0
^{3}H (TU)	0.0	0.0	15.4	15.4
^{14}C (pM)	28.0	43.0	115.8	100
^{87}Sr/^{86}Sr		0.70940	0.70945	0.7094

Table 7. *Vertical variation of the main hydrochemical parameters and microbes at Olkiluoto shown against indicative depth ranges. Variation in pH corresponds with calcite equilibrium in groundwaters and follows the variation in carbonate and calcium contents. Vertical lines in redox column depict steady condition (Posiva 2003).*

Depth (m)	Used classification	Water type	Cl (mg/l)	pH	Alkalinity (meq/l)	Redox
0	Fresh HCO_3	Ca-Na-Mg-HCO_3-SO_4	<10	5.5	<0.5	Post-oxic
10		Ca-Na-Mg-HCO_3-(SO_4-Cl)	10	7	3	Sulphidic
150	Brackish HCO_3	Na-(Ca)-Cl-(HCO_3-SO_4)	2000	7.8	4	
200	Brackish SO_4	Na-(Ca)-Cl-(SO_4)	4500	7.5	1.0	
	Brackish Cl	Na-Cl	2700	8.2	0.4	Methanic
450	Saline	Na-Ca-Cl	8000	8	0.2	
600		Ca-Na-Cl	14000	7.8		
			45000	7.5	<0.1	
1000						

Seawater input to groundwater is indicated by increasing SO_4 with marine S-34 signature and higher Mg contents. The HCO_3-rich groundwater type has been formed since Olkiluoto Island rose above sea level about 2 500 years ago, but the general observation of tritium and moderate radiocarbon values (50-60 ppm or higher) in shallow groundwater, indicate most recent recharge, in the last decades (Pitkänen et al. 2004). Schematic representations of interpreted initial and boundary conditions at Olkiluoto since the glacial period are presented in Figure 31.

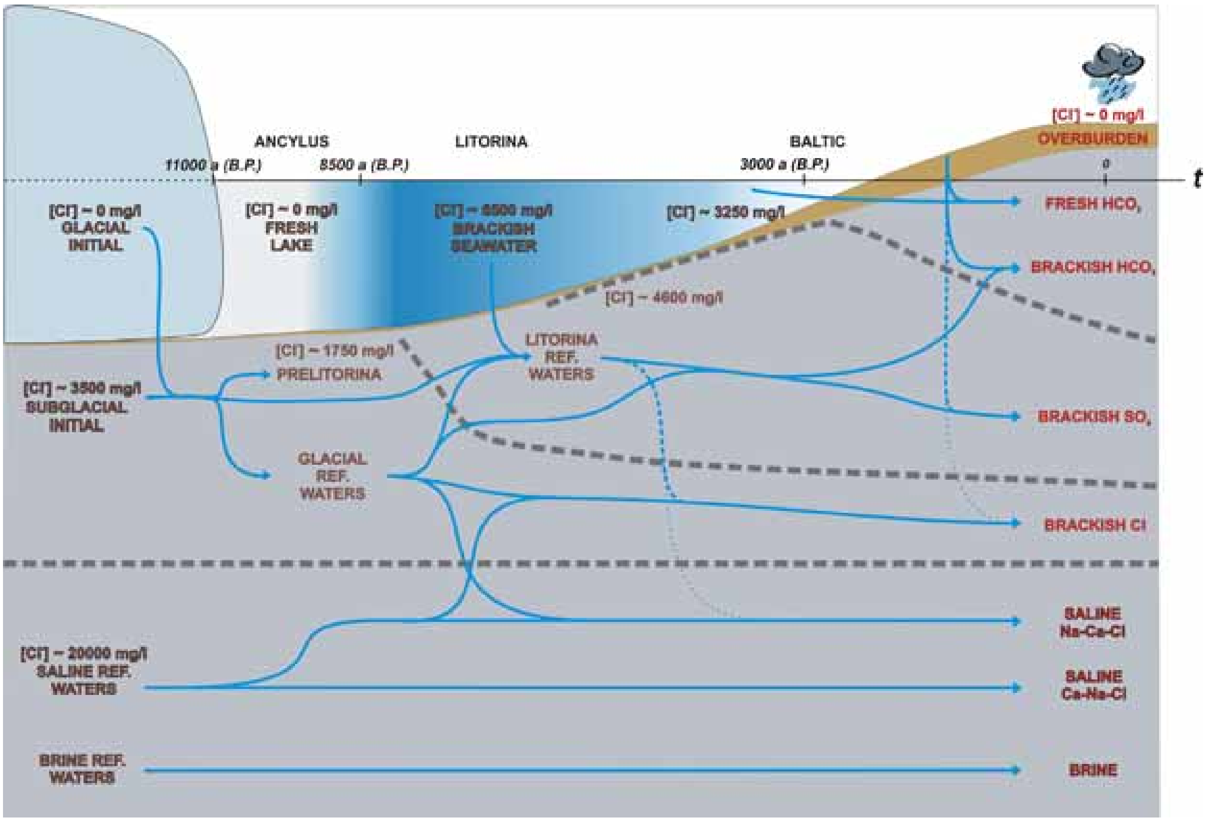


***Figure 31**. Schematical representation of interpreted initial and boundary conditions at Olkiluoto since the glacial period. Potential salinity (as Cl content) is shown for recharge waters (in the upper part along the time line) and bedrock groundwaters (on the left) at the initial stage of the modelling. Saline and brine reference waters are contemporaneous initial waters for studied time period. The generalized hydrochemical mixing hypothesis that is solved in detail using initial waters with mass-balance calculations is presented with blue arrows for the current groundwater types. Dashed lines between arrows imply minor mixing. Major groundwater types are bounded with grey dashed lines (Posiva 2005).*

6 CHARACTERISTICS OF THE BALTIC SEA AT PRESENT

6.1 Shoreline development

Rebound

The rebound model for Fennoscandian presented by Lambeck et al. (1998a) includes a three-dimensional and global model for the Earth's response to changes in surface (ice and water) loading and high-resolution descriptions of ice sheets that are also glaciologically plausible. It is constrained by comprehensive observational data base level information (around 1200 observations for Fennoscandia, Brittish Isles and the North Sea extending from the present to the Last Glacial Maximum).

The model provides realistic predictions of a number of geophysical, geological and geodetic observables, demonstating that it provides good first oder predictions of the response of the Earth to glacial cycles for Fennoscandia. These observations include the age-height relationship of paleo-shorelines, the history of the Baltic Sea and Baltic Lake levels since deglaciation (Lambeck et al. 1998b), and GPS measurements of radial and horizontal crustal displacement (Milne et al. 2001).

Glacial rebound is a global and issue and in Fennoscandia it is not independent of the melting history of the other ice sheets. This complex interaction has led to the development of a process in which the rebound for the different parts of the globe is evaluated iteratively. Thus, the solutions for the Fennoscandia ice sheet may need revision because of improvements made in the Antaric and in the Northern America ice sheets.

Isostatic and eustatic components

The relation between land uplift, eustatic sea level and the water balance of the Baltic determine, whether the sea level is generally rising, maintaining stable or lowering in the present coastlines (Johansson et al. 2001). In the past, the isostatic component has been greater than the eustatic component leaving large parts of Baltic Sea countries under sea level. The Fennoscandian lithosphere is still undergoing postglacial rebound (Påsse 1996) and the rebound still has about 20 000 years to run. Uplift can be considered about constant in the timescale of a few centuries. What is unclear is how long the uplift rate will be constant and what will the boundary be (Ekman 1996). Taipale & Saarnisto (1991) have estimated that the Fennoscandian Global Isostatic Adjustment (GIA) is going to continue for the next 7 000-12 500 years, whereas Kakkuri (1991) gives an estimate of 10 000-12 500 years.

As a result to climatic variations water is stored in the global ice sheets during cool phases and in the oceans during warmer phases, thus the changes of the global sea level, eustatic changes, could be correlated to the extension of the ice sheets. The present shoreline models are based on empirical data and include no physical description of the land evolution process (Påsse 1997) (Figure 32). Empirical data from which the course of glacio-isostatic uplift can be estimated are e.g. investigations of lake tilting (e.g. Påsse 1990b, 1996b, 1998a). Eronen et al. (1995) have studied the shore level changes of the Baltic Sea and the land uplift process in south-western Finland during the last 8 000 years (Figure 33) and the empirical data obtained by studies on sediments in small lakes is consistent with the model by Påsse (1996).

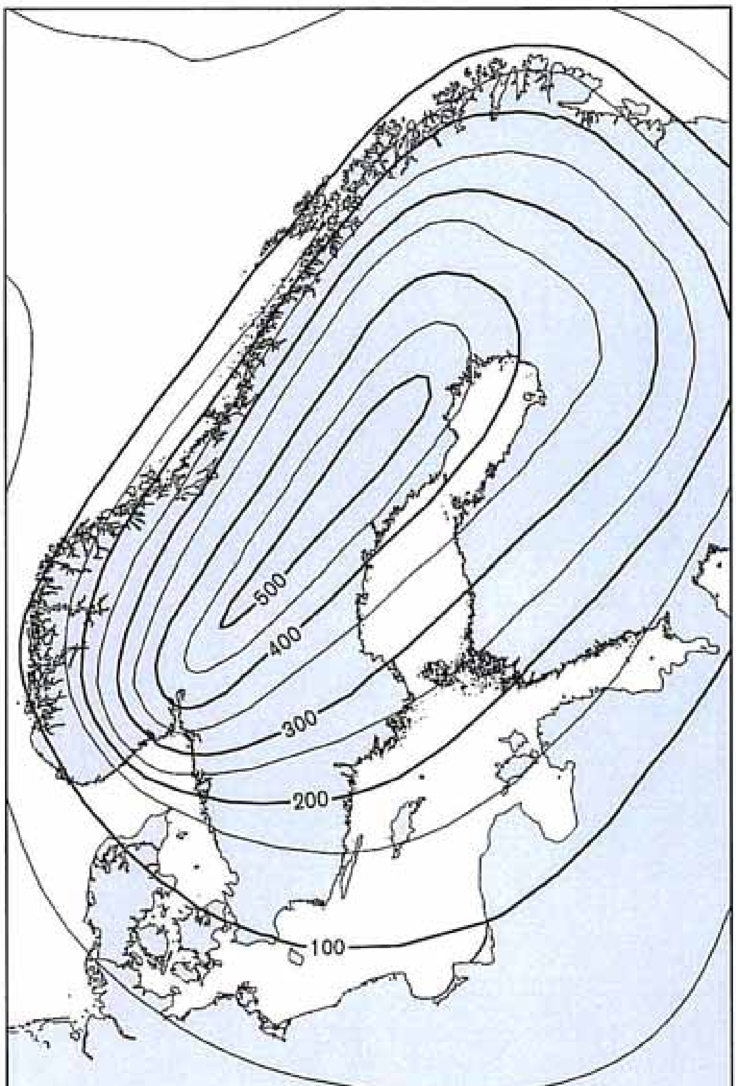


Figure 32. *Isobases for the uplift at 12 5000 BP (Eronen et al. 1995, Påsse 1996 in Morén & Påsse 2001).*

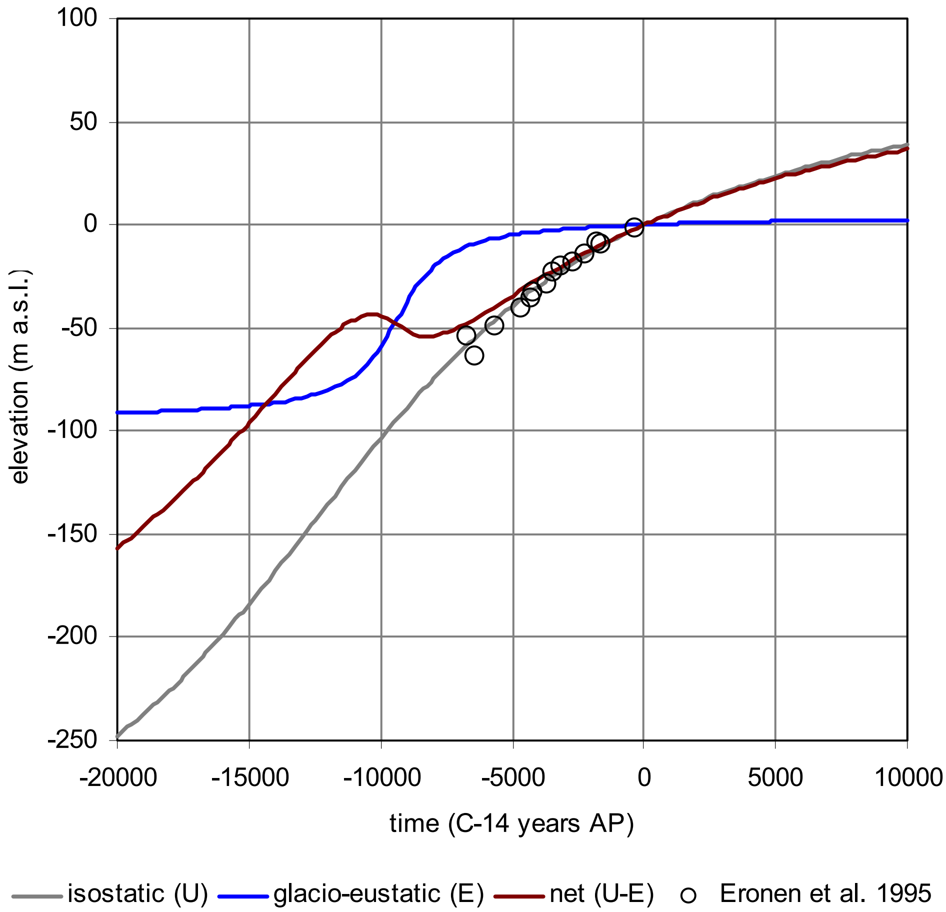


Figure 33. *The postglacial land uplift and the global sea level rise in the Olkiluoto area (Löfman 1999). Curves are based on data by Påsse (1996). Circles denote the empirical data by Eronen et al. (1995). The time range is from 20 000 ago to 10 000 years AP.*

During the last ice age continental environments existed to the east and to the south of a Fennsoscandian ice sheet and oceanic coast conditions prevailed to the west and to the north. These environmental differences make the modelling of shore level displacement in the Fennoscandia quite complex issue (Morén & Påsse 2001).

Currently it is widely agreed that the Earth's climate is warming up and that the sea level is rising. The mean sea-level, which appears to have been steady for the last 3 000-4 000 years, has shown a linear rise of between 1 to 2 mm per year over the last 100 years (IPCC 2001, Johansson et al. 2001). According to ICCP (2001), global average sea level will rise 0.11-0.77 m in the next century. Factors taken into account in this estimate were: thermal expansion (0.09-0.37 m), glaciers and ice caps, Greenland and Antarctic ice sheets and they thawing of permafrost and the effects of sedimentation. A rise of the sea level could have severe impacts on the spatial development of regions around the Baltic Sea, e.g.

- Lowland inundation and wetland displacement
- Shoreline erosion
- More severe storm-surge flooding
- Saltwater intrusion into estuaries and freshwater aquifers
- Altered tidal range in rivers and bays
- Changes in sedimentation patterns

According to the most radical scenario, the Greenland ice sheet will totally melt and raise the sea level by 6 m during the next millennia. Maximum sea level caused by thermal expansion is 4 m. Antarctic temperatures are so low for the ice sheet to significantly melt. An increase in temperature 10-20 ^{o}C would be required, which is beyond present climate-change scenarios (ICCP 2001).

Staudt et al. (2004) have modelled the probable Baltic Sea sea-level rise up to the year 2100. They conclude that future sea level rise in the Baltic Sea declines from south to north due to the ongoing isostatic adjustment, with the German coast experiencing a sea-level rise of 80 cm by the year 2100 and the Bothnian Bay 0 cm. Thus, the lowering trends in relative sea levels will prevail in the northern part of the Baltic for a long time, unless greenhouse warming causes a strongly accelerated rise in world ocean level.

Morén & Påsse (2001) estimated in their model that the shoreline displacement is purely empirical. The model was based on the assumption that the evolution of the isostatic component in time follows the same pattern as observations imply as it did since the late Weichselian. In reality, due to coupled effects of sea-level changes and crustal displacements in response to global loading and rebound, both the global and local sea-levels are complicated to predict (Forsström 1999). The process has also been studied e.g. GIA models (e.g. Milne 2004). Models of future trends should take into account ice load and the physical properties of the crust as realistically as possible.

Thus, geological and other data concerning isostatic/eustatic development during the Weichselian glacial is uncertain. Lundberg & Ford (1994) have presented an intermittent curve based on coral dating. This curve includes hydro-isostatic changes and can not be used to describe shoreline displacement solely caused by eustatic lowering/rise. Shackelton (1987) has form oxygen isotope variations calculated eustatic curve, which is in wide use (Morén & Påsse 2001).

The ice thickness is of great importance for the amount of isostatic depression. However, the viscous flow mechanism also implies that the duration of the glacial load is very important for the depression. A thick ice existing during a short period may produce a small depression, while a thinner ice existing during a long period may produce a similar or even bigger depression (Morén & Påsse 2001).

OLKILUOTO

At Olkiluoto the shoreline migration has affected and will continue to affect subsurface conditions. For Olkiluoto the uplift rate is 6.8 mm/year (Kahma et al. 2001). It can be assumed that Olkiluoto will rise at least for the further 8 000 years and during that time reach an elevation between 48 and 100 m (Cedercreutz 2004).

Due to postglacial uplift and shallow coast areas of sea bottom sediments are emerged from the sea continuously (Figure 34). This in turn, makes primary succession along shores very fast. When comparing the situation at Olkiluoto thousand of years ago to the situation 500 years ago and further to the situation today, it can be seen that already the change in land area has been fast. Even though newly exposed shores are rapidly vegetated (wet land), succession can continue still for thousands of years (Rautio et al. 2004).

According to Starr (1991) the most typical soil type is a podzol after sufficient time and favourable conditions (i.e. Scots pine dominating forests on sandy till soils with vertical drainage). Starr (1999) also observed that at the latitudes of Olkiluoto mere podzol soil formation can take up to 500-1500 years.

The sites mention in text are found in Figure 35. The first summits of Olkiluoto rose above the sea-level about 2 800 years BP and still some eight hundreds of years later there existed about 10 separate small islands, while most of the present land areas lie still under sea-level (Figure 36). At 2000 BP the highest point in the present area of Liiklankallio lain 5.6 m above sea-level (Mäkiaho 2005).

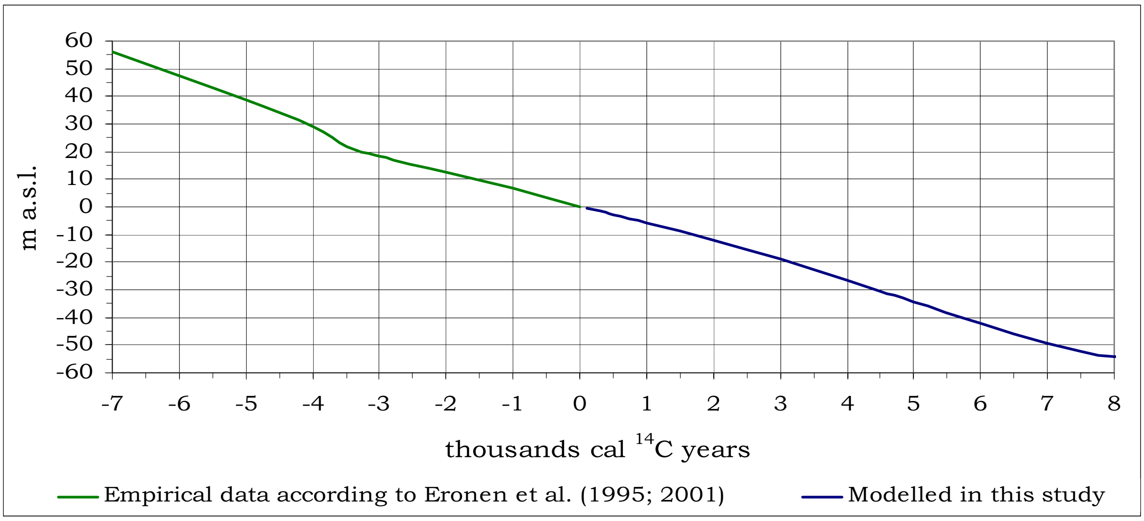


Figure 34. *Shoreline displacement curve for Olkiluoto. Dates Before Present (BP) are adopted from empirical data of Eronen et al. (1995, 2001). The future sea-level curve is derived mainly from the data of Påsse (1996, 1997). The smooth regressive trend of shoreline displacement is expected to continue for thousands of years (Mäkiaho 2005).*

Figure 35. *Sites mention in text (figure modified by Reija Haapanen, Haapanen Forest Consulting, "Basemaps: National Land Survey, Permission 41/MYY/07").*

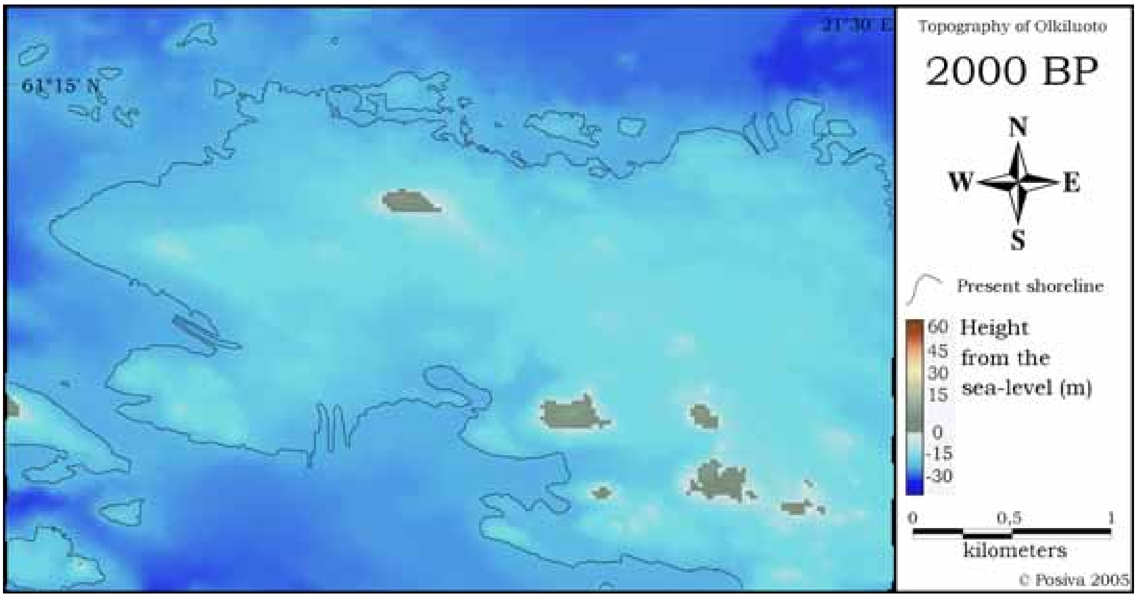


Figure 36. *Topography of Olkiluoto area at 2 000 BP (Mäkiaho 2005).*

The islands emerged from the sea at 1 500 BP are presented in Figure 37. Some or all of these drumlinoils have had a till cover after the declaciation, but strong littoral processes have washed away these glaciogenic sediments, sorting and depositing them to lower elevations and revealing the present rock surfaces (Mäkiaho 2005).

A thousand years ago many of the initially formed small islands have amalgamated into one bigger island and the frame of the present island begins to get its shape (Figure 38). Other small islands expanded and some new summits also rose from the sea and, like in the previous stage, they also lost their likely glacigenic sediments as the waves washed the loose deposits away. Most of these bare rock surfaces are those facing west, which has been a direction of most effective wave action during this stage. This is almost the same direction that has originally had the thinnest soils, since the glacial deposits are thinnest in the direction of the strongest glacial erosion, which in this area are the north-west slopes of the hills (Seppälä 2005). A few small basins were already been disconnected from the sea and formation of bogs that exist in these depressions nowadays started. Altitude of the highest point was just above 11 metres (Mäkiaho 2005).

Figure 37. *Topography of Olkiluoto area at 1 500 BP (Mäkiaho 2005).*

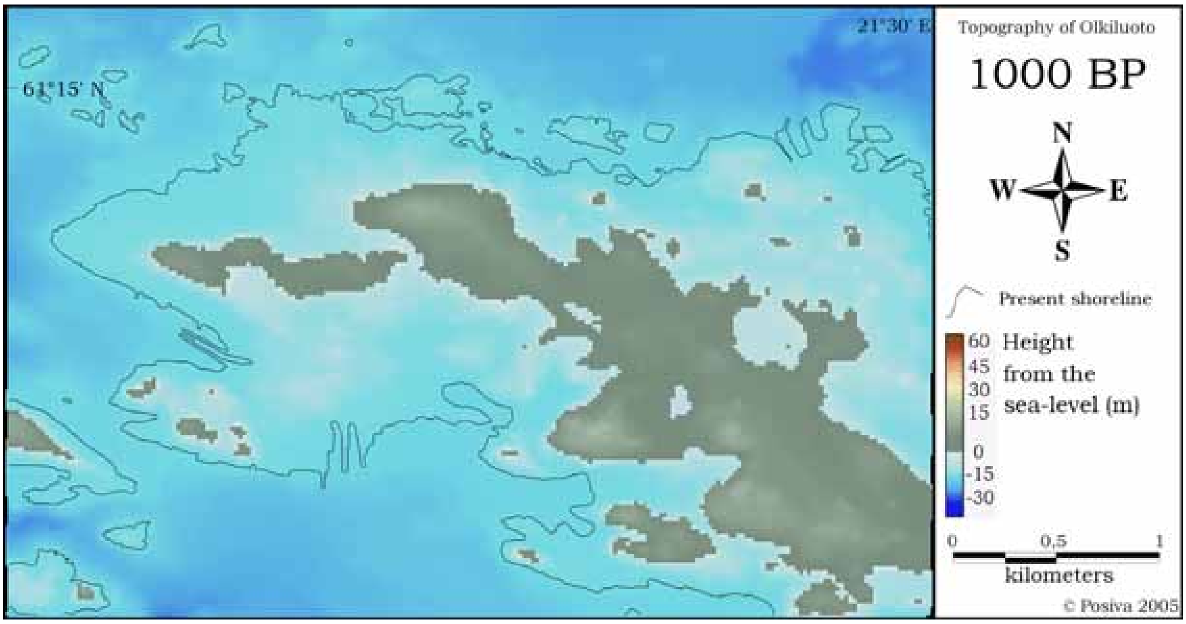


Figure 38. *Topography of Olkiluoto area at 1 000 BP (Mäkiaho 2005).*

Five hundred years ago the Olkiluoto area resembled pretty much that of today (Figure 39). The shape of the island was already recogniseble but the shoreline laid still some hundreds of metres from the present, farthest in the west. The highest point had an altitude of just over 14 metres at Liiklankari

During the next century no dramatic changes will take place (Figure 40). Only the shoals of present Munakari on the north shore of Olkiluoto will become attached to the mainland, which in one part becomes attached to the continent as the narrow strait separating the island from the continent dries up.

Figure 39. *Topography of Olkiluoto area at 500 BP (Mäkiaho 2005).*

Figure 40. *Topography of Olkiluoto area at 100 AP (Mäkiaho 2005).*

After 300 years from present the shoreline will displace a couple of hundreds of metres seawards (Figure 41). Small rocks west from the north-western spit of the peninsula will become a physical part of Olkiluoto. However, probably the most significant event is that the island Kuusisenmaa becomes attached to the westward growing peninsula of Olkiluoto. During this development the bay on the southern side of Olkiluoto, Olkiluodonvesi, will shrink substantially, but it will still be connected to the sea through a relatively wide inlet between the present islands of Kuusisenmaa and Lippo. This inlet seems to exist still 500 AP (Figure 42) and 1 000 AP (Figure 43), but as the land uplift upheaves the threshold between the previously mentioned present islands, a lake with extent somewhat smaller than the sea-areas in Figure 43 will develop just east from the present island Lippo just after 1 000 AP.

Figure 41. *Topography of Olkiluoto area at 300 AP (Mäkiaho 2005).*

Figure 42. *Topography of Olkiluoto area at 500 AP (Mäkiaho 2005).*

The highest point of present Olkiluoto at 1 000 AP will lie at an altitude just above 23 metres. By this time, the marine effect will have decreased substantially as the forming peninsula will become wider and the bays around more isolated.

Figure 43. *Topography of Olkiluoto area at 1 000 AP (Mäkiaho 2005).*

The lake developing eastbound of the present island Lippo is going to have the altitude of the threshold between islands Lippo and Leppäkarta. In Figure 44 it is determined in this neighbourhood a small area that will have altitudes below sea-level. The forming lake will, however, be larger than this observable depression, but its extent depends greatly on the height and material of the threshold.

In the north-eastern corner can be seen a similar depression, which also will very likely be a part of lake forming in that area. However, the continuation of the Eurajoki and Lapinjoki rivers would seem to discharge their waters and sediment loads in and through this depression. These incoming river sediments along with the lakes internal sediment production will have strong effects on the lake development in this vicinity (Ikonen et al. 2003). Other questions arising are the routes of continuations of the previously mentioned rivers. In order to determine those, more detailed bathymetric data from these critical threshold areas are required along with the data on the present sediment quality and thickness (Mäkiaho 2005).

Figure 44. *Topography of Olkiluoto area at 1 500 AP (Mäkiaho 2005).*

Shoreline succession at Olkiluoto offshore

The development of the shore-line will induce changes in the internal conditions, such as biosphere succession, sediment redistribution (sedimentation and re-suspension/erosion) and groundwater flows. These will in turn influence the positions of the potentially discharge and recharge areas.

The shallow shores of Olkiluoto, especially in geolittoral regions, the amounts of common reed are increasing naturally, resulting in paludification of coves and accumulation of organic matter in shallow and nearly-stagnant water (Figure 45). This can result in a faster apparent shoreline displacement than mere land uplift or changes in sea-level would yield. In addition, eutrophication of the Baltic Sea speeds up the process (Miettinen & Haapanen 2002).

In particular climatic and topographical conditions, vegetation succession turns more probably at first to mire ecosystems. In past, most of the peat lands in western Finland were initiated on land uplift shores (Aario 1932, Brandt, 1948, Huikari 1956). Primary mire formations are controlled by the shore displacement of the Baltic Sea and the uplift also segregates bays, which will develop to lakes. Part of them will convert to peat lands filling-in and overgrown by mire vegetation. The primary mire formation and the overgrowth are the starting points of larger mire areas, which will reach their later scale by expanding over adjacent forests (Ikonen et al. 2005).

Figure 45. *Typical common reed colonies causing paludification of shallow bays at Olkiluoto, November 2004. Figure by Ari Ikonen, Posiva Oy. (Posiva 2005).*

The general trend in the vegetation succession of the mires has been from minerotrophic sedge-dominated communities to ombrotrophic shphagnum-dominated communities, which can be found in several bogs in the south-western coast (Aartolahti 1965, Tolonen 1967, Elina 1985, Heikkilä et al. 2001). Transformation from minerotrophic mire to ombrotrophic bogs takes about 2 000 years (Damman et al. 1994). This can be found out in several bogs in Satakunta region; Häädetkeidas bog in Parkano, 80 km east from coast and about 110 km northeast from the Olkiluoto site, which was initiated on land uplift shore about 9 100 BP. When forecasting the coming vegetation types at the Olkiluoto Island the logical supposition is that at the moment prevailing types will be the prevalent ones also in the future, as also proposed in the regulatory guidance (STUK 2001).

Modelled future shoreline positions at Olkiluoto

According to Mäkiaho (2005) the positions of the seashore and crypto-depressions (such inland locations that have elevations below the sea level) can be determined in models. It is more than likely that lakes and mires will develop in most of these crypto-depressions, but from the maps can not determine their extent, but the extent of crypto-depressions.

During the next century land uplift and expand vegetation of the shallow waters will connect the former island of Olkiluoto to the continent and small islands and shoals west of Olkiluoto will become larger and the marine influence at Olkiluoto declines somewhat. Between the 1 000 and 1 500 AP will take place the most dramatic changes in the shoreline displacement at Olkiluoto area. At this time the larger lakes and also some smaller short living lakes and bogs begin to form. However, these events will depend greatly on the routing and sediment transportation of forming continuation of the river Eurajoki on the northern side of present Olkiluoto. Sedimentation rates in forming lakes will also differ depending on the forthcoming discharge direction of the currently bifurcative river Lapinjoki (Mäkiaho 2005).

Mäkiaho (2005) has presented in his report more detailed modelled points of land uplift at Olkiluoto from present to 8 000 AP. The lowest modelled point at 8 000 AP lies about three meters above sea-level, meaning that all the places inside the study area would have emerged from the sea by that point (Table 8).

Table 8. *Minimum distance between the present and future shorelines and modelled height of the highest point of present Olkiluoto (site Liiklankallio) (Mäkiaho 2005).*

Time AP	Minimum distance from the seashore (m)	Highest point of present Olkiluoto (m.a.s.l.)
Present	0	17.5
100	0	18
500	20	20
1000	200	23
2000	6000	29.5
3000	6500	36
4000	10 000	43.5
5000	15 000	51
6000	19 000	59
7000	> 21 000	66
8000	> 21 000	70.5

The model of Ekström & Broed (2006) describes the future terrain and ecosystem development on the basis of sea depth data and on the approximation that 2000 years after present there will be a land uplift of 10 meters above the current depth and elevation values (Rautio et al. 2005). It was assumed that there will be no significant sea level changes during this period. The discharge was assumed to go directly to an area above the repository, which at the time will be covered by forest. The succession of linked biosphere models was identified from maps of the terrain and will also involve lakes, rivers and coastal areas, as illustrated in Figure 46. From the depressions remaining under the sea level, locations of lakes can be estimated. Likely some of the lakes will be larger, or smaller, than the depression, but for testing the tools and modelling methods these estimates were judged to be adequate. Based on the type of current sea bottom sediments, future forest types on the area can be forecasted (Rautio et al. 2005). The current prevailing forest types will prevail also in the future, but somewhat more wetlands and deciduous forests around the water bodies are predicted. On the basis of the terrain and ecosystems forecast for the selected time period the corresponding landscape model was built using ecosystem-specific biosphere modules taken from earlier assessments and modelling exercises. Also the project called "NICELAND" will give more detailed information on future shoreline positions and ecosystem succession (Broed 2006, in preparation).

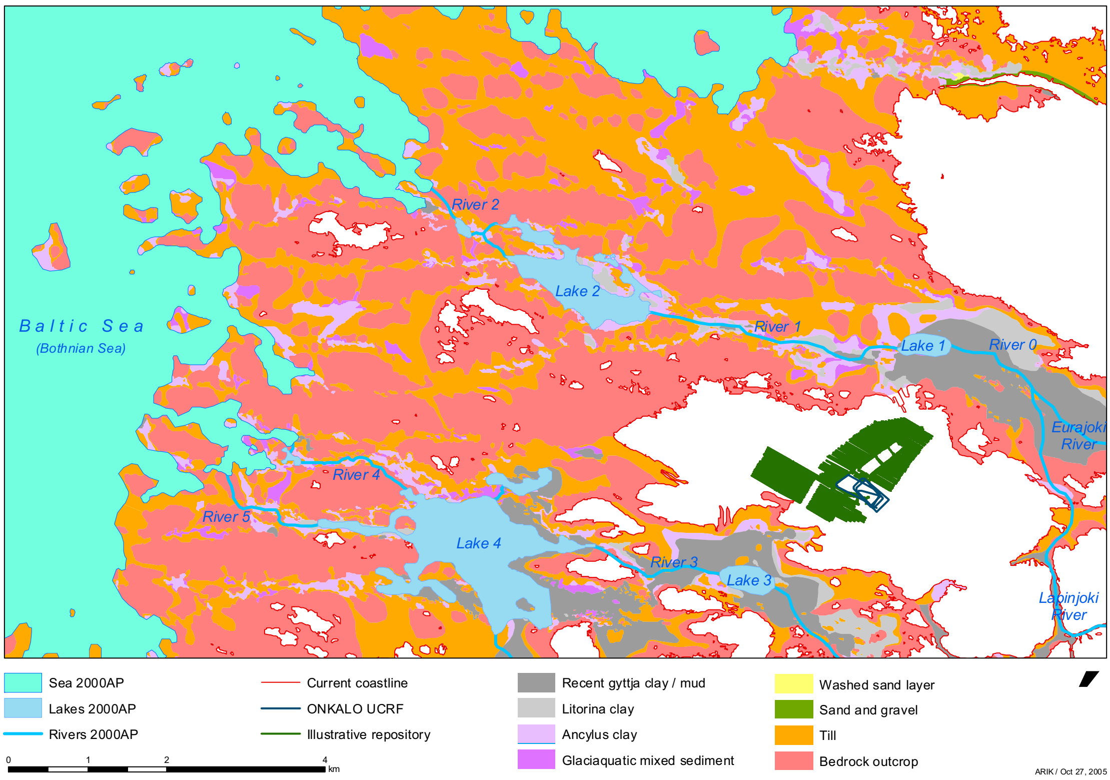


Figure 46. Surroundings of Olkiluoto at about 2000 AP (modified from Rautio et al. 2005 in Ekström & Broed 2006).

Helcom (1996) project has suggested that due to the ongoing isostatic adjustment, the Kvarken area (between Umeå and Vaasa) there will be a rise at a rate of 8-9 mm/y and around 1 km^2 of new land rises from the sea every year. According to Gustafsson (2004) the Bothnian Bay will gradually become less saline and finally a lake. During the 4 000 years the Baltic Sea will be a land bridge, which will be formed between Finland and Sweden, with only a river joining the "Bothnian Lake" with the remaining brackish Baltic Sea (The GEONAT project, GTK 2004) (Figure 47*)*.

Figure 47. *The Baltic Sea about 4 000 AP based on the elevation model of Seifert et al. 2001 and uplift isobases of Ekman 1996 (after Breilin et al. 2004; image copyright by the Geological Survey of Finland/Hanna Virkki).*

6.2 Sea-bottom sediments

Sediments contain records of the historical events from the Baltic Ice Lake to the present post-Litorina Sea stage, and past inputs into the aquatic system. A vast number of processes tend to immobilize records of sediments, while others, such as bioturbation, disturb the record of input. Often the reactions are slow, and reflect biotic processing as well as chemical transformations and they are greatly influenced by the redox conditions in the sediments e.g. variations in sediment redox conditions can be linked to historical observations of salt-water ingressions into the Baltic Sea from the North Sea (Leivuori 1998). At the offshore of Olkiluoto data was not available on sea sediments and their chemical, physical and biological analyses.

In the Baltic Sea the sedimentary environment has changed from detrital rich sediment in the past to more organic rich sediments partly due to the increased eutrophication, which also has an influence on distribution of nutrients and harmful substances. This can be seen e.g. as an increase in the total organic carbon, zinc, mercury, cadmium (Figure 48), PCB and DDT in the recent Baltic Sea sediments. Salinity corrected distributions of cadmium and mercury in sediments in the Baltic Sea are presented in Figure 49 (Leivuori 1998).

Eutrophication of the Baltic Sea is mainly caused by the excessive nitrogen and phosphorus loadings from land-based sources as (Figures 50 and 51). About 75 % of nitrogen and at least 95 % of phosphorus enters the Baltic Sea i.e. via rivers or as direct discharges. Nitrogen and phosphorus loads vary considerably from year to year depending mainly on hydrological conditions. In the periods of high runoff nutrients are abundantly leached from soil increasing the loads originating from diffuse sources and natural leaching.

High concentrations of heavy metals in biota of the Baltic Sea are mainly caused by loading from land-based sources. About 50 % of mercury, 60-70 % of lead, and 75-85 % of cadmium enters the Baltic Sea as waterborne. Another significant contributor to the total heavy metal load is atmospheric deposition. In the Baltic Sea measured concentrations of heavy metals have been even an order of magnitude higher than concentrations in the North Sea (Table 9). The main reason behind the high concentrations in the Baltic Sea is large catchment area with an intense industrial activity, high amount of population, and above all long renewal time of water.

Quantified annual information on waterborne inputs of heavy metals is needed for evaluation of long-term changes of heavy metal concentrations in the state of marine environment. Due to the incomplete data on heavy metals, a quantitative picture of the loads entering the Baltic Sea can not be given sufficiently. Shortcomings in national monitoring programs and the lack of proper laboratory equipment in some countries mean that heavy metal figures can not obtain in many cases, or the loads reported have not been not fully reliable (e.g. no harmonized detection limits). Also different calculation methods have been used in the Baltic Sea countries if the measured concentrations have been below the detection limit. The data sets of unmonitored rivers and coastal areas are even more incomplete.

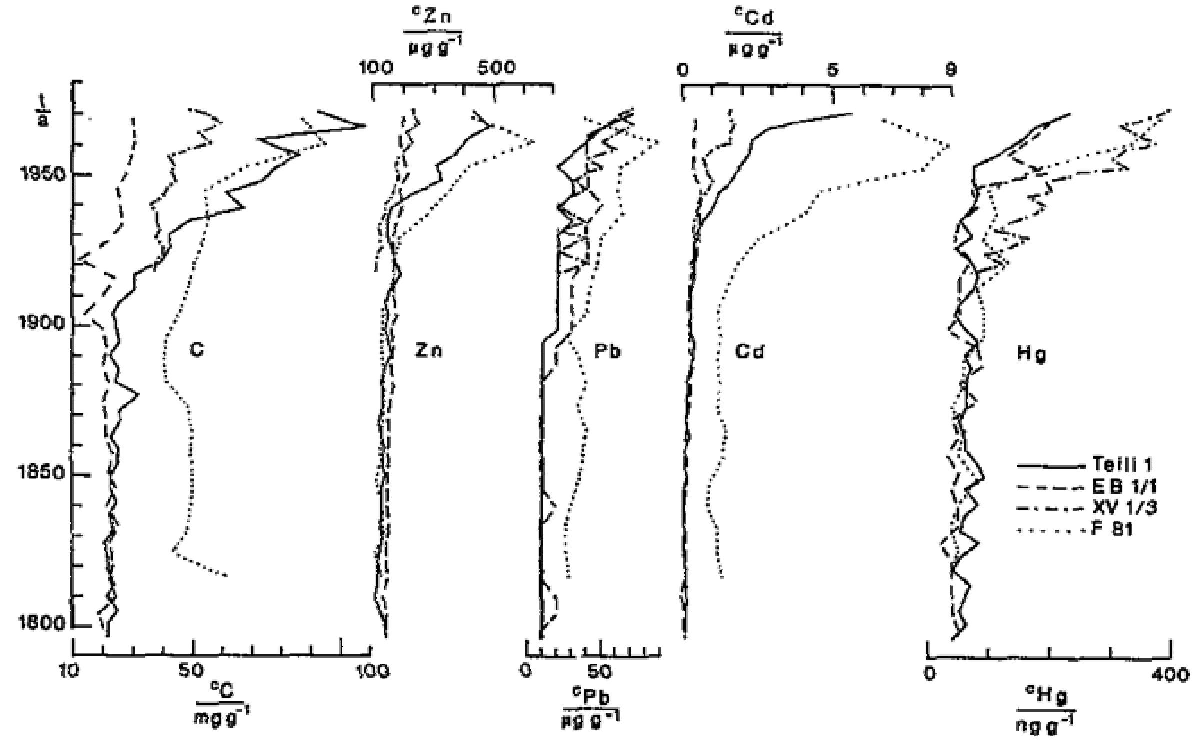


Figure 48. *Vertical distribution of carbon, zinc, lead, cadmium and mercury in dated sediment cores at stations Teili-1 (Northern Baltic Proper), EB1 (central Bothnian Sea) and XV-1 (Eastern Gulf of Finland) (Leivuori 1998, www.balticseaportal.fi, from Niemistö & Tervo 1981).*

Table 9. *Concentrations of dissolved trace metals (mg/kg) from the North Atlantic and the Baltic Sea (www.balticseaportal.fi).*

Element	**North Atlantic**	**Baltic Sea**
Hg	0,15-0,3	5-6
Cd	4 (+-2)	12-16
Pb	7 (+-2)	12-20
Cu	75 (+-10)	500-700
Zn	10 (+-75)	600-1000

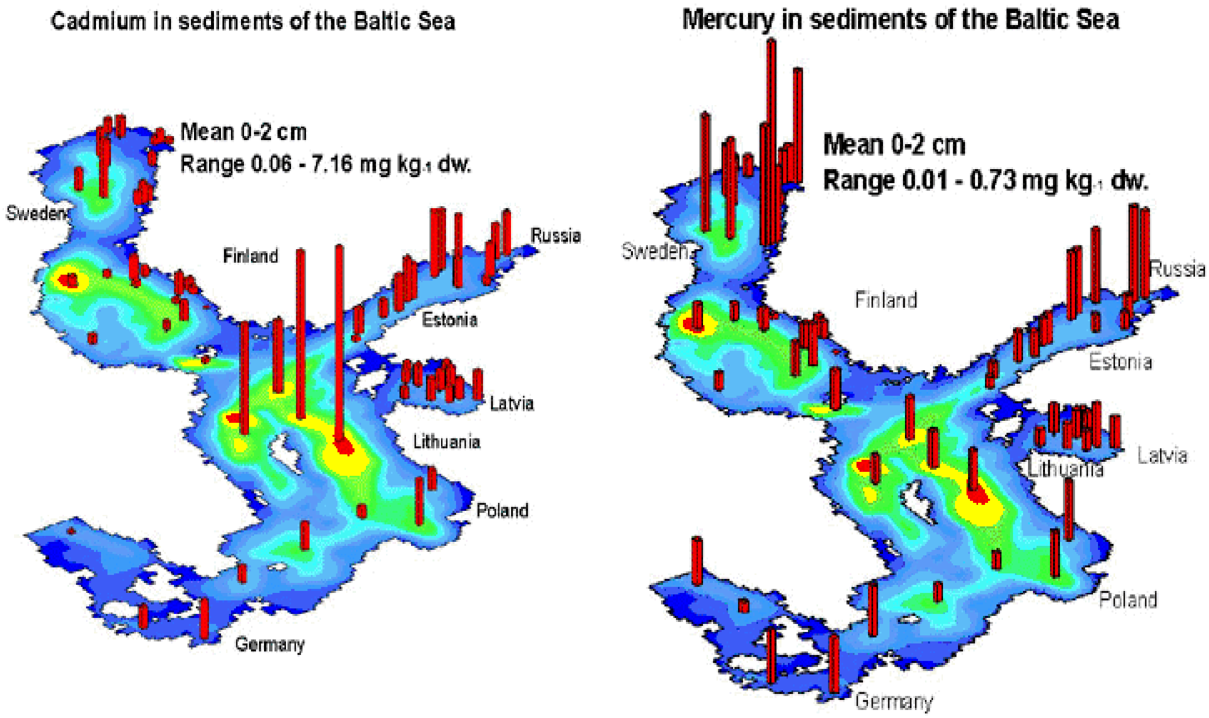
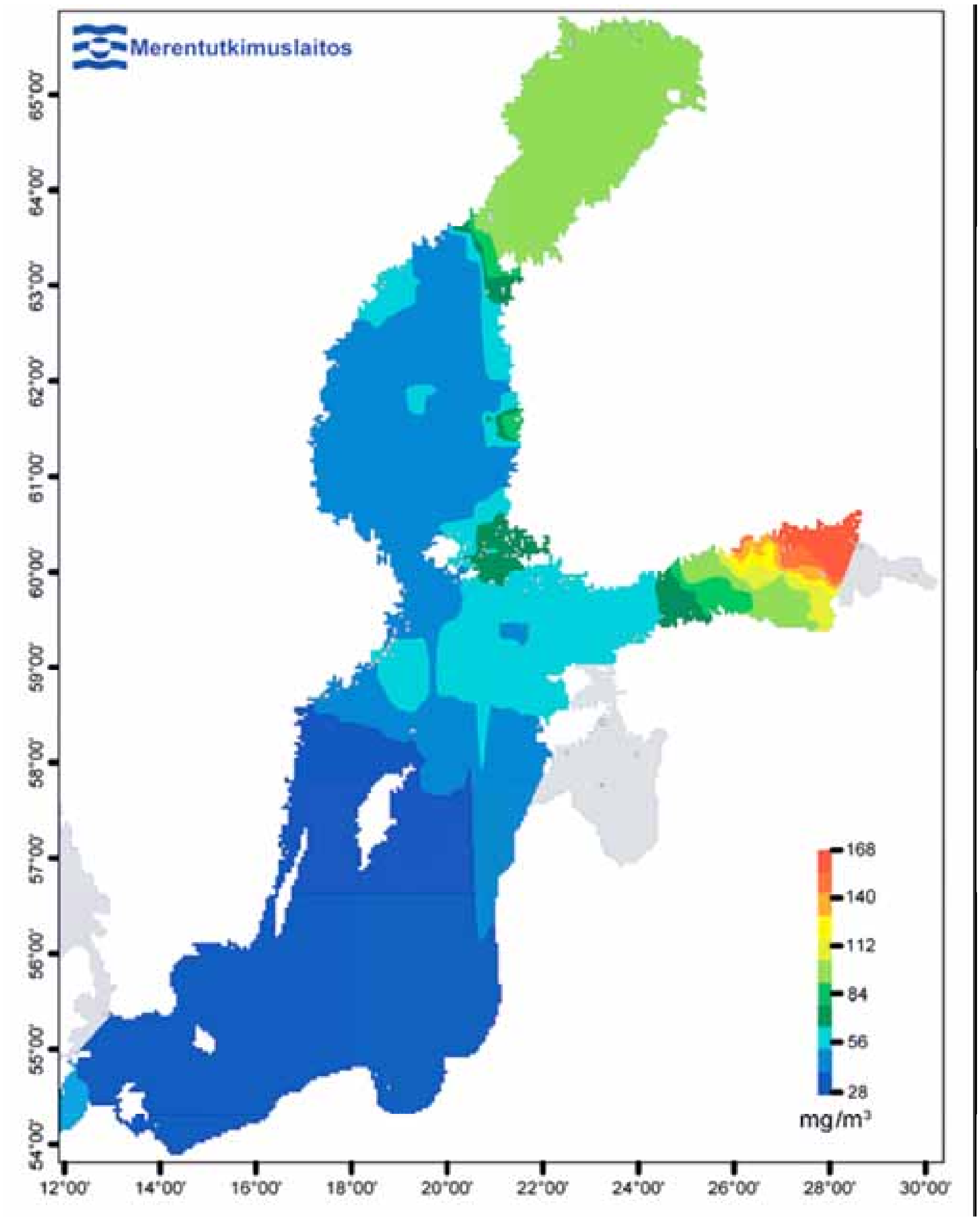
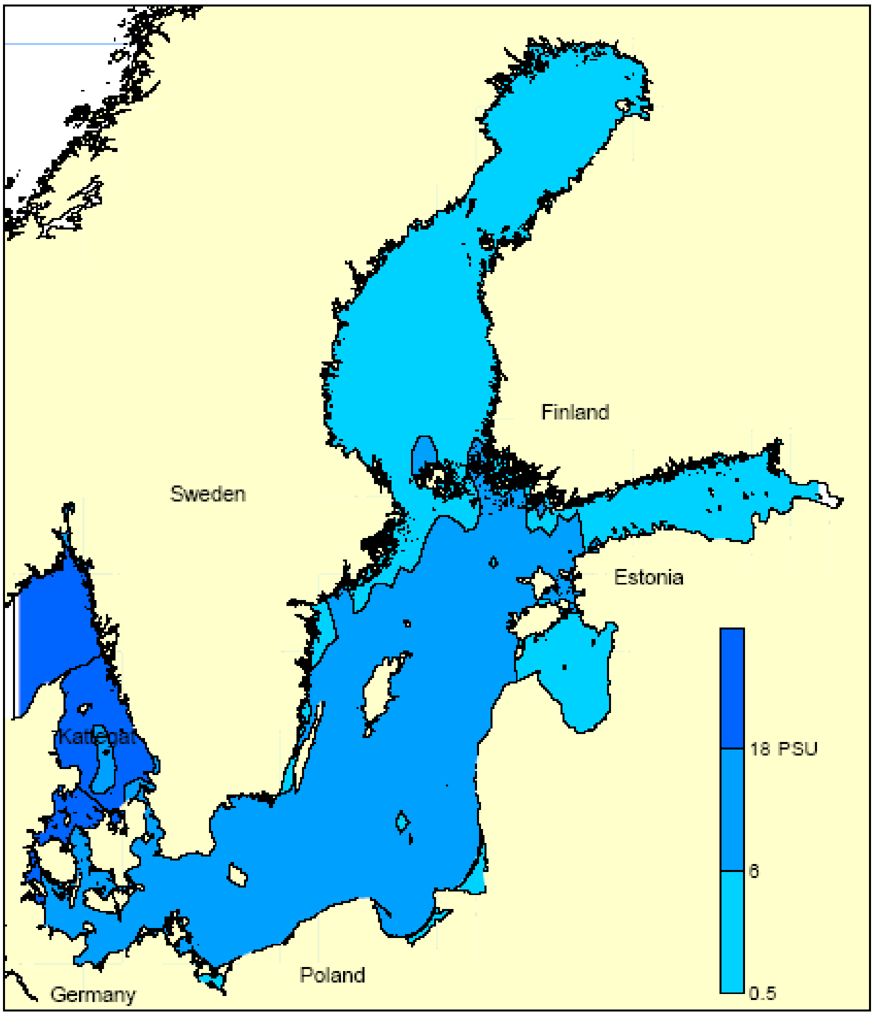


Figure 49. *The salinity corrected distributions of cadmium and mercury in sediments of the Baltic Sea (Leivuori 1998, www.balticseaportal.fi, data partly combined from Perttilä 1998).*



(Merentutkimuslaitos=Finnish Institute of Marine Research)

Figure 50. *Nitrate concentrations (mg/m^3) in the surface water (Kotilainen & Kohonen 2005).*



(Merentutkimuslaitos=Finnish Institute of Marine Research)

***Figure 51**. Phosphate concentrations (mg/m^3) of the surface sea water in the winter 2004 (Kotilainen & Kohonen 2005).*

OLKILUOTO

Suspended solids

The concentrations of suspended solids in the sea water are typical in the area of Olkiluoto to those of the Bothnian Sea coastal waters and are at same level as those of the more southerly areas off Rauma and Pyhäranta, the reference area about 26 km to the south (Posiva 2005). Intermitted variation appears in concentrations of suspended solids, especially during windy periods when fine matter rises up to the water mass from the bottom. The concentrations of suspended solids have increased slightly from the outset of the 1990s, at which time the average summer season concentrations were approximately 2-3 mg/l, whereas they are currently approximately 3-5 mg/l, on average (Ikonen et al. 2003).

Similarly, the turbidity values rose particularly during the summer periods at the beginning of the 1990s. Due to increase in water turbidity, Secchi depths (visibility of the water) have decreased on average approximately 0.5 m from the outset of the 1990s. During recent years, Secchi depths have been approximately 2.5-3 m (Ikonen et al. 2003).

Nutrients

The phosphorus concentration during the growth period rose during the 1980s and 1990s, as they did elsewhere in the coastal waters. The N concentrations at most of the observation plots were 20-40% lower in winter 2004 than at corresponding time in 1995-2001, in average. The background concentration in coastal waters of Bothnian Sea was in average 350 mg/m^3 in 1997-1999. Highest concentrations, with unusual amounts of total N, were measured from the surface water of Eurajoensalmi. This may be caused by the weather conditions, rains after long dry period.

The concentrations of total N at the nearby waters of Olkiluoto during the open water season 2004 varied between 260-310 mg/m^3. The N concentration of open sea locations did not much differ from background values. The amount of inorganic N compounds was low year around. The ammonium N concentration varied between < 3-12 mg/m^3, except at the mouth of Eurajoki river, where the concentrations raised to 170 mg/m^3 at maximum (Haapanen 2005).

The phosphorus concentration during the growth period also rose during the 1980s and 1990s. In the winter 2004 the P concentrations) were between 12-21 mg/m^3, which were clearly below previous winter concentrations but at the same level with values from outer sea regions. The background concentration in coastal waters of Bothnian Sea was in average 22 mg/m^3 in 1998-2000. The growing season 2004 average P concentrations of nearby waters corresponded to the averages of early 1990s, but were higher than in 2003 due to rains. There has been cyclic, irregular variation in the phosphate concentrations during the monitoring period 1979-2004 (Haapanen 2005, Kirkkala 2005).

Eurajoki River and Eurajoensalmi Strait

The Eurajoki River represents second largest river system in the south-western Finland after the Kokemäenjoki River, by reason of which its water quality has been monitored as regular measure for decades. Eurajoki River belongs to the river basin of Eurajoki, which surface area is 1.336 km^3 with lakes comprising 13 % of the same area. In the Eurajoki riverside catchment area (excluding Lake Pyhäjärvi catchment area), the proportion of fields is approximately 28 % and forests (primarily peat land) make almost 70 % of the surface area. The early sulphide clay areas of Litorina Sea are within the catchment area of the river. Wastewater processes by industry and the municipalities are discharged into the river (Ekholm 1992) ((Table 10).

The water flowing from Lake Pyhäjärvi determines the water quality of the upper course of the Eurajoki River. Due to compensating influence of the lake basin, seasonal variation in upper course water quality is relatively minimal. The Eurajoki River has been classified as clean. Intermittently, turbid additional waters rich in nutrients flow into the river from ditches and tributaries. Diffuse pollution is the most important factor affecting the lower course of the Eurajoki River for most of the year, but particularly during minimal stream flow periods, the wastewaters of riverside municipalities and industry have an effect. The water close to bottom has been nearly as salty as the seawater farther out from Olkiluoto. The Eurajoensalmi Strait has in recent summers been more eutrophic than previously Considering essential variables, no statistically significant trends are indicated (Ikonen et al. 2003).

On average, the stream flow of the Eurajoki River has been 9 m^3/s during 1990-2001 with variation from 5.9 to 11.8 m^3/s. During larger flood periods, maximum flow has been 44 m^3/s. The loading to the sea close Olkiluoto brought by the Eurajoki and Lapinjoki Rivers is shown in Table 11. The river waters increase the concentrations of nutrients and solids and affect the opacity of water at the mouth of the Eurajoki River and in the Olkiluodonvesi waters south of Olkiluoto. In general, the nutrient and solid content is characteristic of coastal waters of the Bothnian Sea (Posiva 2005).

Table 10. *Municipal and industrial wastewater loads (kg/d) in the Eurajoki River during the years 1996-2000 (Ikonen et al 2003).*

Parameter	1996	1997	1998	1999	2000
BOD_{7ATU} *	256	246	212	318	407
Total phosphorus	5.3	9.4	6.7	10.5	8.8
Total nitrogen	238	220	231	252	267

* BOD_{7ATU} analysis reflects the biological oxygen demand in the water

Table 11. *Averages of the loading (metric tons) to the sea by the Eurajoki and Lapinjoki Rivers for the period 1990-2001 (Ikonen et al. 2003).*

Parameter	Eurajoki	Lapinjoki
Suspended solids	7000	2000
Total nitrogen	530	250 – 300
Total phosphorus	17	5 - 6

Sea fauna and flora

In region of Olkiluoto affected by cooling waters the currents and rise in temperature have influenced phytoplankton production, but the changes have been rather mild and local. The lengthening of the growing season, especially during the spring, has increased production. The differences between the cooling water discharge site and other sea area have increased during the 1990s, as the total amount of plankton algae has increased in the entire area as compared to the latter part of the 1980s (Ikonen et al. 2003).

The vegetation off Olkiluoto varies from algae-dominant community in the outer archipelago to the vascular-plant-dominant community in the more sheltered Olkiluodonvesi area. In transect studies eutrophication was discernible in the cooling-water-affected area during 1990s. The species composition have clearly changed, and the perennial species to primarily annual filamentous algae (Ikonen et al. 2003).

The amount of benthic fauna and species composition has varied considerably. The differences have arisen in part from the quality of the bottom, but also from eutrophication. The larvae and polychaetes, which tolerate lack of oxygen, have proliferated during the close of the 1990s. At the same time, the decline in the Baltic clam has decreased the total biomass. Especially in the cooling water discharge area, the status of the bottom has deteriorated (Ikonen et al. 2003).

Reference areas

In the sea area off Olkiluoto, there are many characteristic features limiting the location of the reference areas:

- Between the mainland and open sea, only the narrow archipelago zone remains, so the influence of the open sea is strong
- The area is shallow (< 10 m) and rocky
- The loading from rivers and wastewaters is rather small; thus considerable loading sources cannot be situated in the reference area

To the south of Olkiluoto, off Rauma, the sea area has been followed by obligatory monitoring studies (Figure 52). The sea area off Rauma is different in its depth ratios from that off Olkiluoto, and additionally a considerably amount of wastewater loading comes to the area. In going south from Rauma, the archipelago zones broaden further, so suitable reference areas cannot be found (Ikonen et al. 2003).

The coast continues to be shallow and stony on the north side of Olkiluoto. Off the city of Pori, the present southernmost sampling sites in the sea area are at Luvia to the southwest side of Lankoori Cape. The impact of loading from Kokenmäenjoki River and Pori area is no longer clearly distinguishable in Lankoori. In the coast area remaining between Olkiluoto and Lankoori, there are fish farms whose nearby waters have been studied with separate monitoring programmes. Locations appropriate as reference areas with respect to natural conditions and loading would evidently be possible to find south of Olkiluoto off the Nurmes and Aikonmaa islands or in the north between the Pujonkulma and Lankoori capes. Unfortunately, rather minimal material is available from both areas (Ikonen et al. 2003).

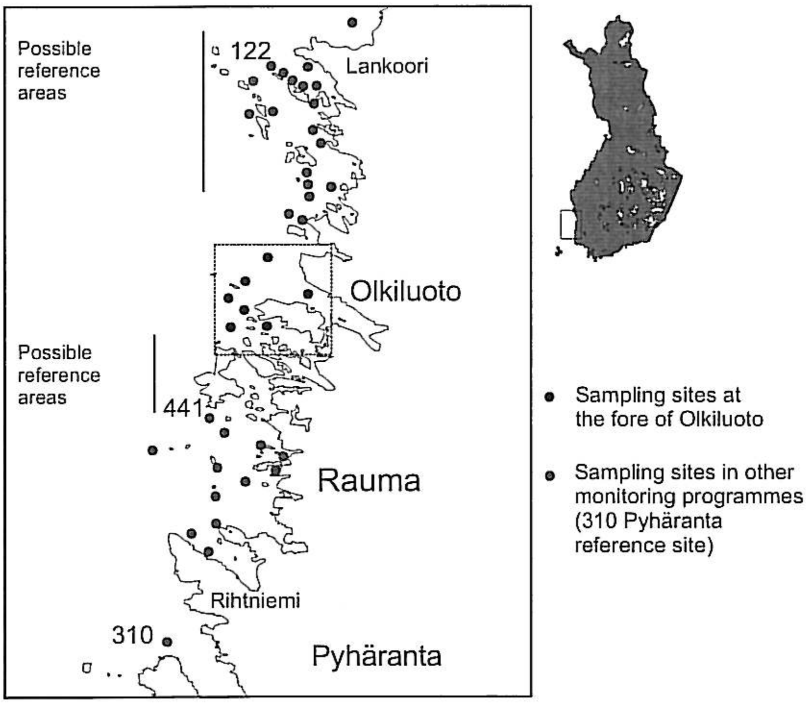


Figure 52. *Location of potential reference areas (Ikonen et al. 2003).*

6.3 Coastal water system

Ongoing geological processes are shaping the coast; long term data show annual to decadal fluctuations of physical, chemical and biological variables, which are mainly caused by climatic features and land uplift. Changes of the coastline are linked to processes of erosion, transport and accumulation. For all kinds of coasts it is important to know the regularity of the circulation of water and sediments in the coastal zone (www.fimr.fi, www.helcom.fi, www.balticseaportal.fi).

The functions of the coasts zone in the ecosystem:

- Acting as a filter and purification plant for nutrient discharges
- Acting as a nursery and spawning are for fish
- Potential for aquaculture

Coastal sensitivity - the sensitivity of coasts depends on:

- Water turnover - affects the concentration of substance
- Bottom-dynamic conditions - effects depend on the type
- Morphometry - affects the dispersal and/or sedimentation of fine material. The coastal morphometry determines how the energy is processed within the area

All of these factors influence the dispersal, sedimentation and recirculation of different substances, e.g. nutrients in the coastal zone.

Bottom types:

- Erosion bottom - found in near-shore areas, coarse materials domination, water content < 50 %
- Transport bottom - characterised by discontinuous deposition of fine matter, water content 50 - 80 %
- Accumulation bottom - characterised by continuous deposition of fine matter, water content > 75 %

Archipelago:

- From crustal rise (ca. 4 mm/year) along shorelines with a hilly outcrop terrain archipelago between the Åland Islands and the Finnish mainland and the Stockholm archipelago, one of the largest in the world with its more than 11,000 km^2 and 24,000 islands and islets

6.4 Present water level variations in the Baltic Sea

The main factors influencing water level in the Baltic Sea are atmospheric pressure, wind, currents through the Danish straits and, during winter, the extent of ice cover and its effects. Water flow in and out of the Danish straits changes the total volume of Baltic water and thus to the water level all around the Baltic. Currents are caused by water level differences between the Atlantic and the Baltic and strong winds in the area.

Tidal forces cause only few centimetres of water level variation in Finnish coastal areas. Wind piles up water to certain areas of the Baltic, especially inner bays, and the highest amplitudes of water level can be found in these areas. Wind can also have local effects. High atmospheric pressures push water surfaces down. A density gradient of one millibar equals to one centimetre water, normal changes in atmospheric pressures can thus shift water levels tens of centimetres.

There is an annual cycle in the Baltic water levels caused by a parallel cycle in atmospheric pressures and thus winds. The mean water level is highest during December and lowest during April-May. Changes in water levels are strongest during November-January and weakest during May-July. Single years can vary a lot and during some years the kind of cycle described is not present.

The severity of ice winter can be measured by the maximum extent of ice cover at a given year. There is a clear relationship between ice cover and water level; more extensive ice cover means lower water levels. Ice cover influences also to the amplitude of water level changes; more ice means less variation (www.fimr.fi, www.helcom.fi, www.balticseaportal.fi).

6.5 Water balance and water circulation

Factors influencing the hydrography characteristic (e.g. temperature and salinity) of the Baltic Sea include the narrow and shallow connections with the Atlantic through Danish Straits, the great volume of incoming freshwater through river runoff and precipitation as well as the depth and form of the Baltic basins. The large volume of less dense freshwater entering the Baltic create a low-salinity water layer on top of the saltier Atlantic water over the whole sea and not only in estuarine areas as in most marine basins. Since the water level of the Baltic Sea on the average remains the same, there is a balance between the influx and the outflow of water:

$$V_f + V_p + V_{in} = V_e + V_{out}$$

The water flow from the catchment area (V_f), the precipitation (V_p) and the water introduced through the Danish straits (V_{in}) should be as large as the evaporation (V_e) and the water flowing out into the North Sea (V_{out}). The water balance of the Baltic Sea is positive: the amount of fresh water flowing into it is greater than evaporation. The excess flows through the Danish straits into the North Sea. The average water exchange between the Baltic Sea and the North Sea is only about 400 cubic kilometres per year, and an average, completely theoretical time for water to remain in the Baltic Sea (including both fresh and seawater sources) is therefore 22-30 years (www.fimr.fi). Helminen et al. (1998) have simulated the flow directions and water circulation in the Baltic Sea (Figure 53).

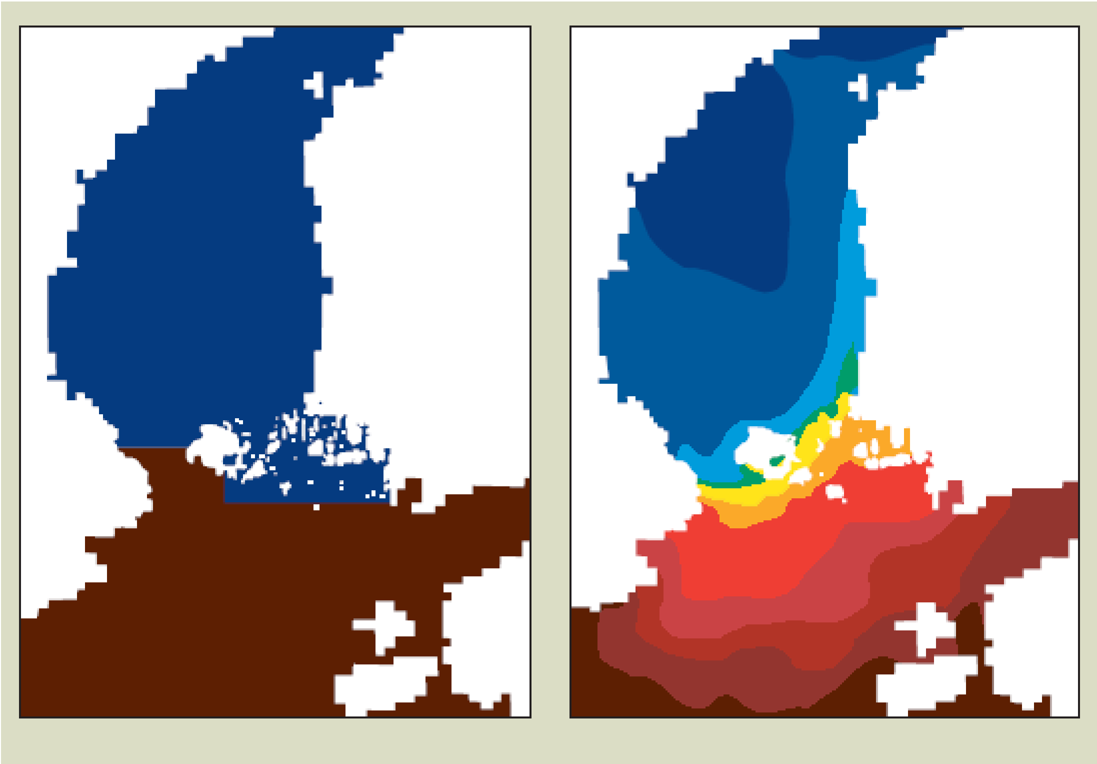


Figure 53. *Water mixing simulated by flow direction models in the Baltic Sea and Gulf of Finland (dark brown) and in the Gulf of Bothnia (blue). In the west; initial state, in the right; after simulation. The light blue figure represents the water flow from the archipelago to the Gulf of Bothnia (Helminen et al. 1998).*

6.6 Salinity and water mixing

Litorina Sea salinity

The salinity of the Baltic Sea has changed during past ice ages and it is expected to do so also in the future. During the time span the Baltic Sea salinity has varied substantially, increasing from zero to 10-15 ‰ (Table 12). Around 6 000 – 5 000 years the salinity seems to have been about the same as now. The duration and salinity level of the maximum salinity phase, the Litorina Sea, has been discussed extensively during over 100 years (e.g. De Geer 1889, Munthe 1894, 1910, 1931, 1940, Sedercrantz 1896, Sernander 1911, Segerståle 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, Tavast 1996). Dating of salinity changes during the Litorina Sea stage have been based mainly on interpolations of shore-displacement curves (Munthe 1910, 1940, Kessel & Raukas 1979, Hyvärinen et al. 1988, Punning et al. 1988).

The extensive investigation is that of Munthe (1910, 1940) on the Island of Gotland (Westman et al. 1999). 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 ‰. Furthermore, the 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. However, both earlier and later estimates which are based on fossil assemblages 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).

The estimates based on occurrence of cyanobacterial blooms during the entire Litorina Sea phase (Bianchi et al. 1998) suggested 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 stage, and this probably resulted in a much less pronounced salinity gradient than at present (Westman et al. 1999). 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).

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

Table 12. *Maximum salinity of the Litorina Sea stage by comparison with present day values (Westman et al. 1999).*

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	Win 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 Depth, bottom water	*c.* 8	12	Huckriede et al. 1996
Baltic Proper	> 12	7-8	Bianchi et al. 1998

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

Present salinity and mixing conditions

Due to the considerable load of freshwater and the low water exchange, the Baltic Sea is intensively stratified. The water in the Baltic Sea consists roughly of two layers; surface and bottom waters. Salinity changes abruptly in the depth of 50-80 m from the more fresh surface waters to the saltier deepwater. This depth of rapid change in salinity is called a halocline and it is a permanent phenomenon in the Baltic. Like a lid, the halocline limits the vertical mixing of water. In the Gulf of Bothnia, the halocline is very weak or absent.

In summer a thermocline – a distinct layer of water where the temperature changes rapidly – divides surface waters into two layers; a wind-mixed surface layer down to a depth of 10–25 m, and a deeper, denser and colder layer extending down to the sea-bed or the halocline (Figure 54). Salinity increases slowly towards the bottom below the halocline. Deepwater oxygen is thus minimally replenished from atmospheric oxygen diffusing to the surface waters and requires saltwater pulses from the Atlantic for replenishment. Irregular oceanographic events, salinity pulses are created by extraordinary strong and long lasting winds pressing surface waters through the Danish Straits from the Atlantic. The significant saline pulses of the last century arrived in 1913, 1921, 1951, 1976, 1993 and 1994. The saline pulse easily takes six months, often longer, to reach the Baltic Sea (www.balticseaportal.fi).

However, there is a high-frequency exchange of water going on all the time, but it has almost no effect on the Baltic Sea, as the same water just goes back and forth. Only during very exceptional conditions does the influx last long enough (over two weeks) to reach far enough into the Baltic Sea not to recede again. During such a significant pulse, the Baltic Sea receives between 200 and 400 cubic kilometres of salty ocean water within a few weeks, but it is only very slowly mixed with Baltic Sea water.

The salinity in the Baltic Sea is in general lower than 10 psu compared to the global oceanic where the average salinity is of 35 psu. Due to freshwater runoff surface salinity increases towards the bottom in most Baltic basins but the salinity difference between surface and bottom decreases from the Danish Straits towards the Bay of Bothnia (Figure 55). Bottom salinity is about16 psu in the Southern Baltic Proper, 10-12.5 psu in the Gotland Deep and only 6.5-7 psu in the Bothnian Sea. The halocline of the Baltic Proper locates at 60-80 depth and below this the salinity increases to 11-13 psu (Samuelsson & Stigebrandt 1996, Gustafsson 2004). In Figure 56 is presented stratification and water residence time in selected inshore area of the Baltic Sea (Schernewski & Wielgat 2004).

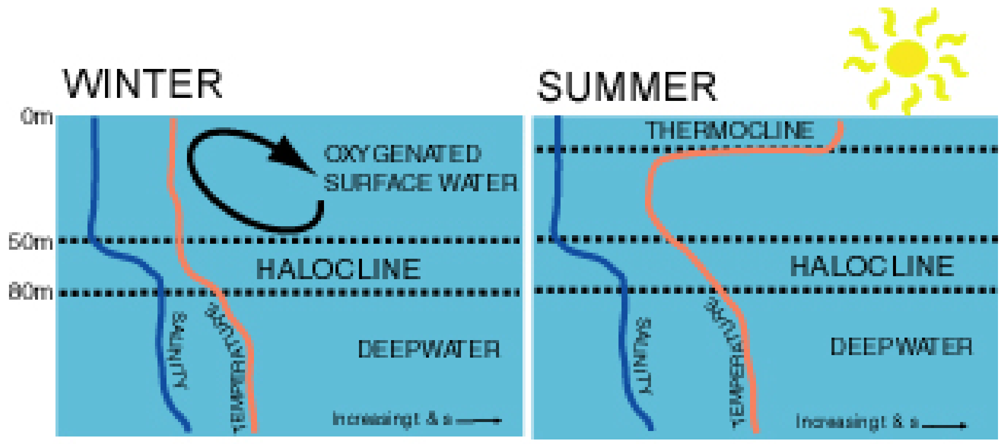


Figure 54. *Annual variation in salinity and temperature (www.balticseaportal.fi).*

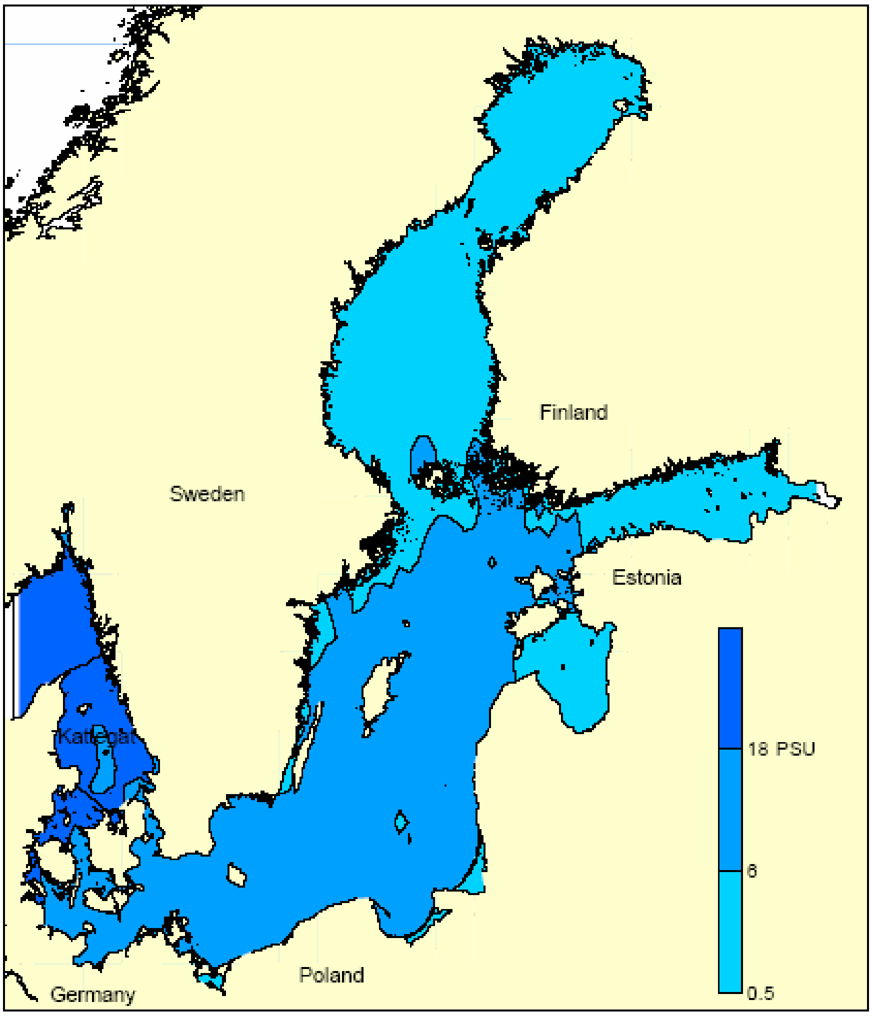


Figure 55. *Distribution of salinity in surface Baltic Sea waters up to 5 meters depth (Schernewski & Wielgat 2004).*

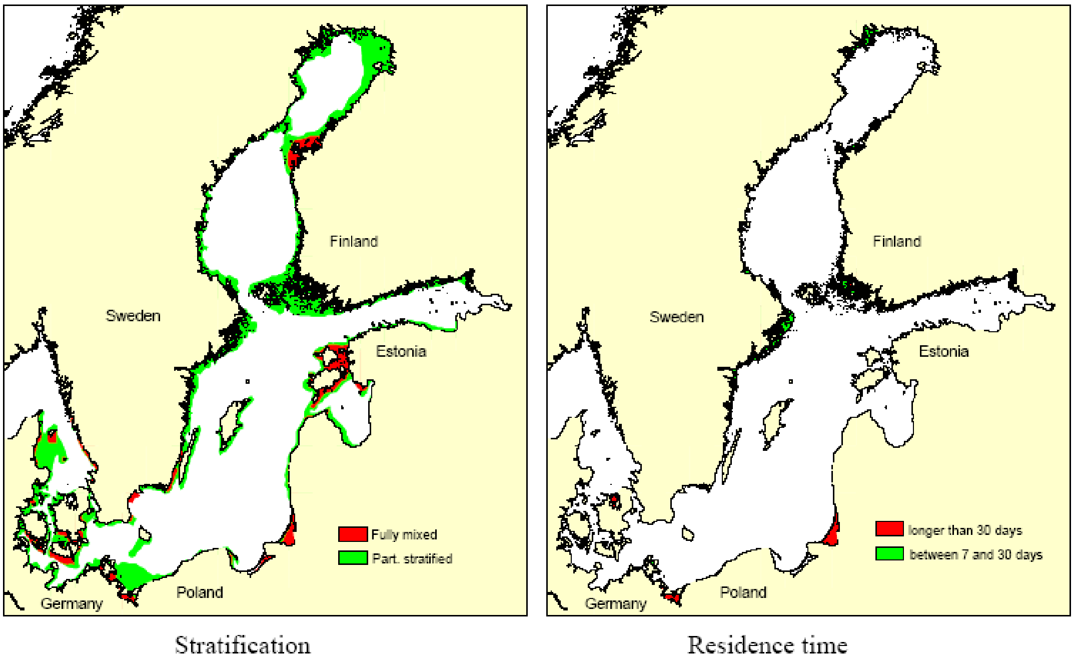


Figure 56. *Stratification (left) and water residence time (right) in selected inshore areas of the Baltic Sea (Schernewski & Wielgat 2004).*

Several investigations have quantified the sensitivity of the salinity to changes in freshwater supply and oceanic exchange in the Baltic Sea (Gustafsson 1997a, Gustafsson 1997b, Gustafsson 2000b, Gustafsson & Westman 2002, Omsted et al. 2000, Stigebrandt 1983, Westman et al. 1999). However, these investigations dealt with relatively small changes in environmental conditions. Rodhe & Winsor (2002), Stigebrandt (2003) and Gustafsson (2003) have exploited the freshwater needs to be to force the Baltic Sea to become a freshwater lake. These two investigations gave quite diverging results; where Rodhe & Winsor (2002) found that the salinity would become negligible already for a 25 % increase of the supply, while Stigebrandt and Gustafsson (2003) found that the supply needs to increase by a factor of 4. The latter figure has also been confirmed by Meier & Kauker (2003) (Gustafsson 2004).

Gustafsson (2004) has presented computation results of the Baltic Sea salinity for different sea levels and freshwater supplies. Salinities for Baltic Proper, Bothnian Sea and Bothnian Bay are computed for global sea-level changes from -10 m to +10 m and freshwater supplies from 0 to 60 000 m^3/s. The results showed that the change in salinity is almost twice as large for a lowering of sea level as for a rise. A global sea level lowering about 5 m or an increase of freshwater income by a factor of three would be necessary to force the northern Baltic into a limnic state.

Typology and spatial type distribution in coastal waters

Salinity thresholds used to differentiate between types were chosen in line with Water Framework Directive System A and CIS Working Group Guidance ranges and according to the well accepted Venice system. In the Baltic Sea there are three salinity classes; from oligohaline to polyhaline waters. The Baltic typology subdivided according to the Finnish typology is presented in Figure 57 (Schernewski & Wielgat 2004).

- Freshwater < 0.5 psu
- Oligohaline waters 0.5 – 6 psu
- Mezohaline waters > 6 – 18 psu
- Polyhaline waters > 18 – 30 psu

The present classification of types within the Baltic Sea is based on three main factors:

- Surface salinity
- Water residence time which separates open coast from semi-enclosed bays/inshore areas are delimited as separate geographical units
- Depth, which corresponds to the mixing of the water column

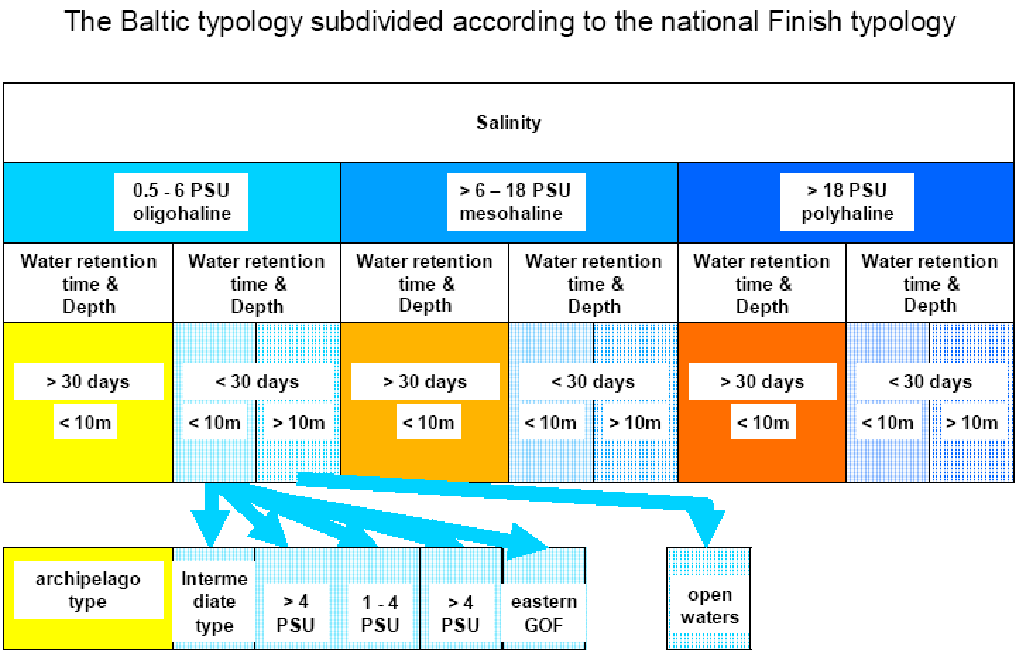


Figure 57. *The Baltic typology subdivided according to the Finnish typology (Schernewski & Wielgat 2004).*

6.7 Temperature

The northern Baltic Sea has considerable wide annual temperature ranges; between < 0 ^{o}C and 20 ^{o}C. The Bothnian Bay is normally completely covered by ice by January and in generally, ice coverage also occurs in coastal zone down to Åland Sea and along the Gulf of Finland (Figure 58).

During the winter the temperature in the layer above the permanent halocline is only a few degrees above freezing. In general, ice forms in marine waters when temperatures are below 0^{o}C. Exact freezing temperature depends on the salinity of the water; more saline water freezes at lower temperatures. Seawater freezes around at -0.20 ^{o}C in the Bothnian Bay, but at -0.45 ^{o}C in the more salty Baltic Proper.

During the spring water warms and turns denser and sinks deeper due to density maximum of freshwater is at 4 °C and for the brackish Baltic seawater at 2.3-3.5 °C. From this point on warming decreases the density and the thus heat is transported to the deeper layers by the forces of wind and waves and summer thermocline forms dividing the upper water layer in two. Temperatures may drop 10 °C within a few meters in thermocline depths, shallower in the spring but at around 15-20 m during the end of August.

The varying Baltic currents influence horizontal and vertical distributions of seawater temperatures and salinity. Average water transportation is counter clockwise. This can be observed in the slightly lower surface salinities along the Finnish Bothnian Sea coast compared to the Swedish coast at similar latitudes. During summer the phenomenon called up-welling can induce a drastic drop in coastal surface temperatures. Up-welling is caused by wind pushing water off the coast. Cooler and saltier deepwater flows up to replace it.

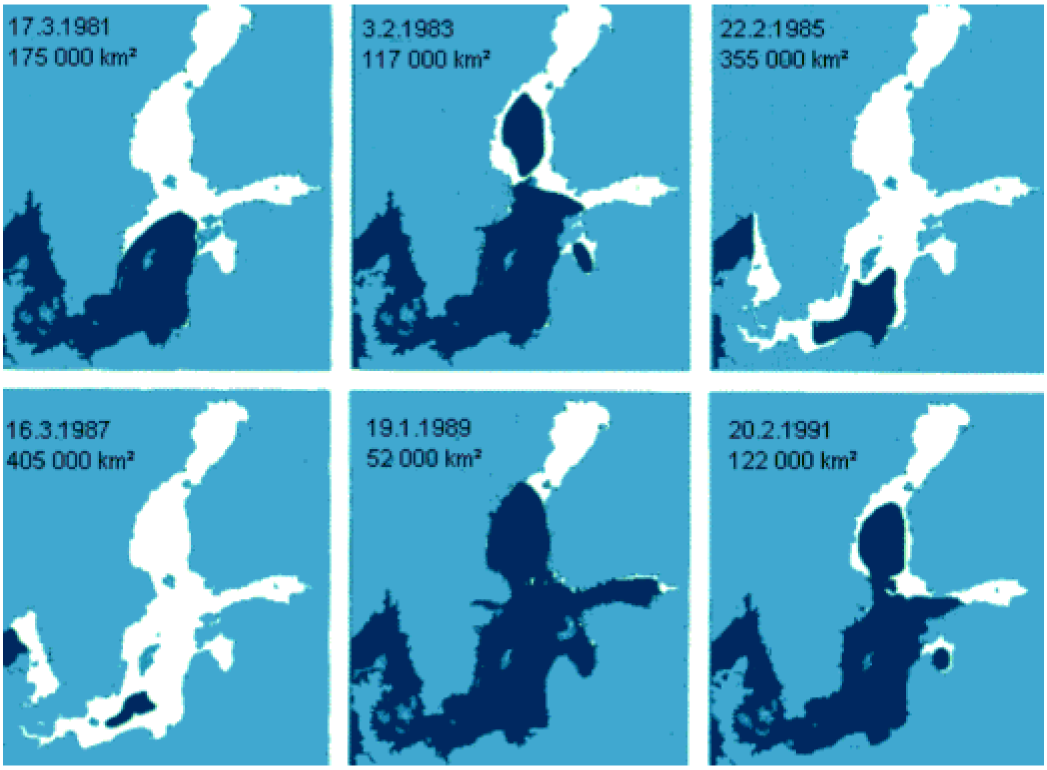


Figure 58. *Ice situation and surface water temperatures in the Baltic Sea differ significantly per year. The ice cover ranged from 52 000 km2 to 405 000 km2 in 1981 1991 (www.fmri.fi)*

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During the winter, the cooling waters mix with the surface layer of the sea area, and the local rise in temperature is noted with in distance of 3-5 km. The temperature of the surface layer at the cooling water discharge area, i.e. near Kaalonperä Cove, generally increases 5-7 ^{o}C and further out generally 0.5-2 ^{o}C from the background level. The local rise has significance for fish, the ice situation of the immediate waters of Olkiluoto and algal production in non-frozen water area. During the open water season, the rise in seawater temperature remains quite local, and temperatures does not deviate from the normal values fro the Bothnian Sea coastal waters. A considerable (over 2 ^{o}C) increase in temperature is limit to the surface layer of the cooling water discharge area, and a moderate rise in temperature appears within 3-4 km (Ikonen et al. 2003).

In the monitoring performed in February 2004, the nearby waters were mostly free of ice cover. Some pack ice, was, however disturbing the sampling. The surface layer temperature at Kaalonperä was 6.3-10.5 ºC, which can be directly translated to temperature difference compared to outer sea. Close to bottom the rise in temperatures was under 2 ºC. The effect of cooling waters could be observed in other locations as well, but it was mostly mild (0.7-5.7 ºC). In the open water season the rise of water temperature caused by the cooling waters was greatest in June. Significant (over 3 ºC) rise was limited to the 0-2 m layer of the very discharge area. Slight rise (0.1-3 ºC) could be seen as far as 3-4 km from the discharge site. No clear changes have been observed in the impact of cooling waters to the sea water temperature in recent years. The mixing of cooling waters in the sea varies with wind conditions and water stratification (Kirkkala 2005).

6.8 Oxygen

Oxygen depletion in the deep basins of the Baltic Sea is a natural occurrence due to limited water exchange between the Baltic and North Seas. The input of freshwater runoff into the Baltic Sea and salinity differences prevents deeper waters from mixing with oxygen-rich surface waters (Figure 59). Eutrophication induced by excessive nutrient input has considerably worsened oxygen depletion, further threatening e.g. marine ecosystems and biodiversity.

Oxygen enters the bottom through almost regular spring and autumn circulations, and there is no oxygen deficiency. In the winters the oxygen conditions have been good in the whole open sea area and also in most parts of the coastal waters. Reduced saturation values (below 70 %) have been observed only in some deep parts of the open Gulf of Finland and very restricted coastal areas, mostly small bays and estuaries. In the summers the conditions have been clearly poorer in the deep open, where mean saturation values from 30-60 % were common. The deep lowering of oxygen has usually followed by increase of salinity conditions. Oxygen conditions have been poorest under the depth of 60-70 meters, in halocline layer (www.balticseaportal.fi, www.helcom.fi).

Due to the increased sedimentation, and decomposition of organic matter, oxygen could be used up and anoxic conditions can occur in the water column, especially in the bottom layers. Oxygen depletion is widely used as an indicator for the indirect effects of nutrient enrichment. While oxygen levels above 4.5 ml/l are considered to cause no problems for macroscopic animals, levels below this cause increases stress to most organisms. Shortage of oxygen will increase the production of hydrogen sulphide. When oxygen concentrations fall below about 1 ml/l, bacteria start to use anaerobic processes, producing hydrogen sulphide (Figure 60). Hydrogen sulphide is toxic, and its concentration is described in terms of negative oxygen (e.g. www.balticseaportal.fi, www.helcom.fi).

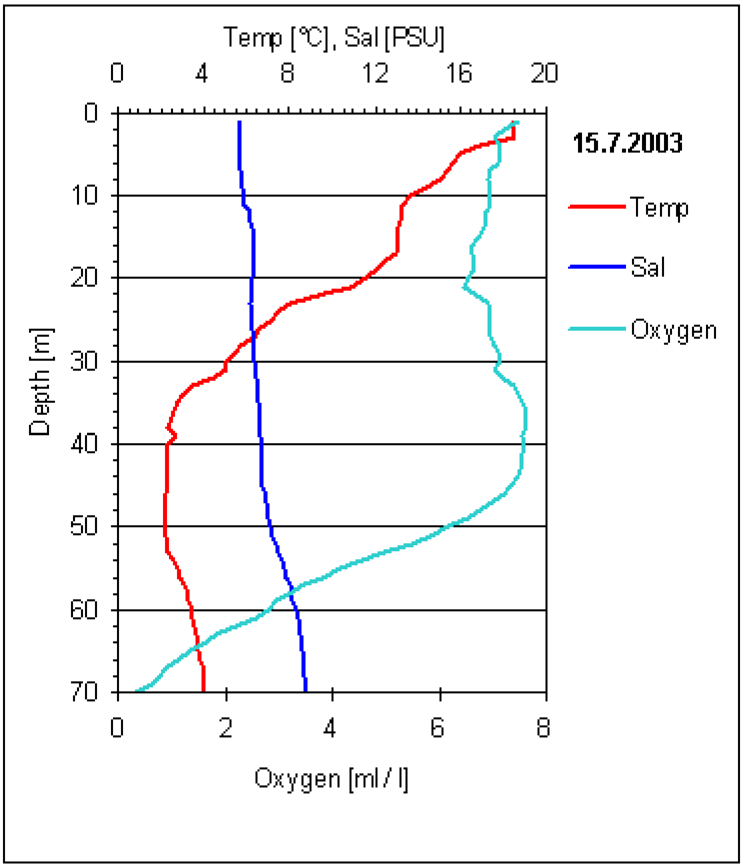
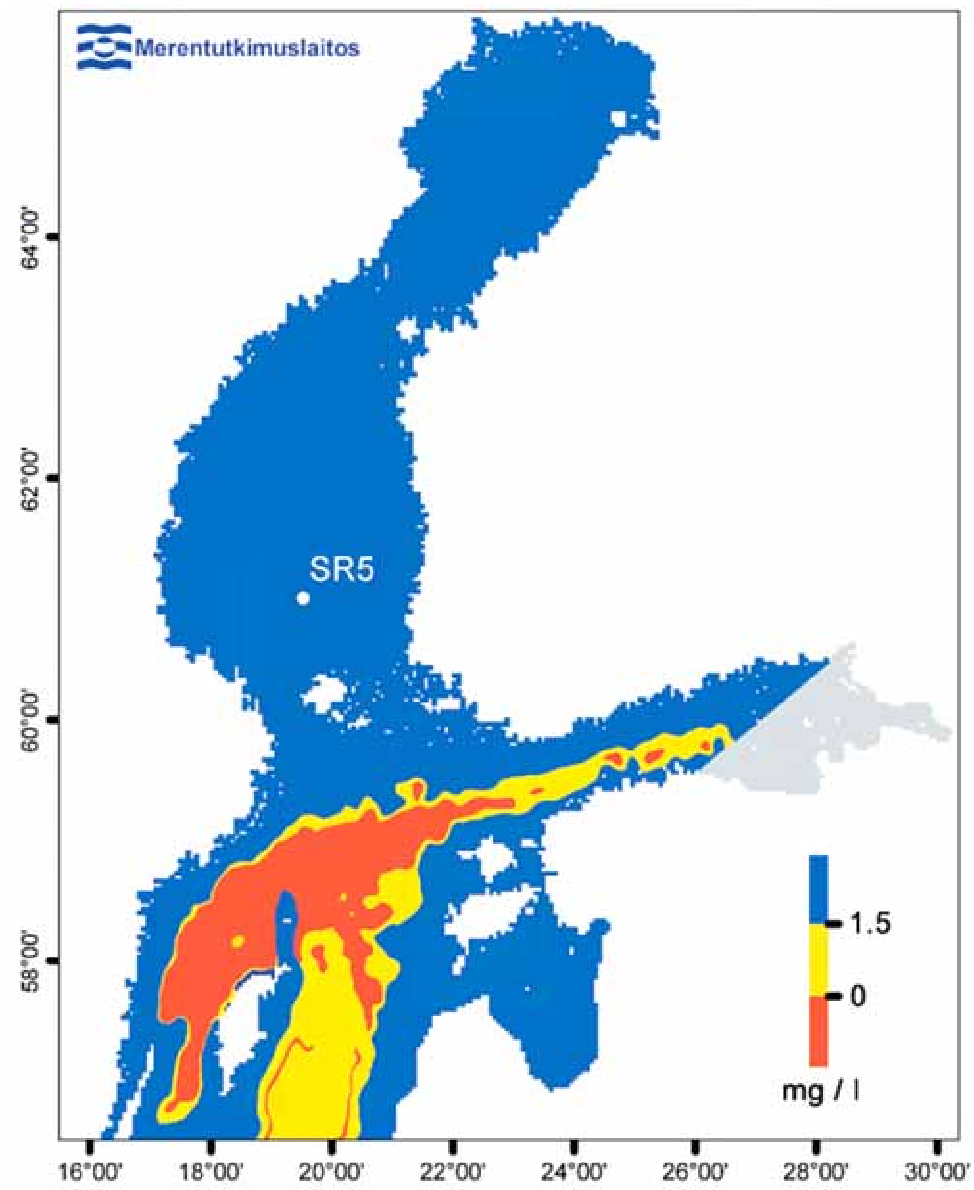


Figure 59. *Temperature (red), salinity (blue) and oxygen concentration (turquoise). Data: Finnish Institute of Marine Research, Aranda 15 July 2004 (www.fimr.fi).*



(Merentutkimuslaitos=Finnish Institute of Marine Research)

Figure 60. *Oxygen concentrations (mg/l) in the bottom surface water in summer 2003. The negative value represents concentration of hydrogen sulphide (Kotilainen & Kohonen 2005).*

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The oxygen balance in the sea water was mainly undisturbed during the entire study period 1990-2001 (Figure 61) (Ikonen et al. 2003). Only two sampling sites had the oxygen concentrations intermittently fallen below 70 % saturation, which is insufficient for salmon, among other fishes. Disturbances have, however, been random and short in duration for the time being.

The oxygen saturation of sea water at the observation plots in 2004 is presented in Table 13. The oxygen concentrations were in line with the background values in late winter (February). At the beginning of May there was oversaturation of oxygen in the surface water due to the production maximum of plankton, as in previous years. The summertime oxygen balance was more stabile than normally due to cold weather and weak stratification; even the waters close to deeps had saturation values of 85-104 %. In the autumn (September) the oxygen conditions were excellent (94-103 %), and in accordance to background values of other coastal waters of Bothnian Sea.

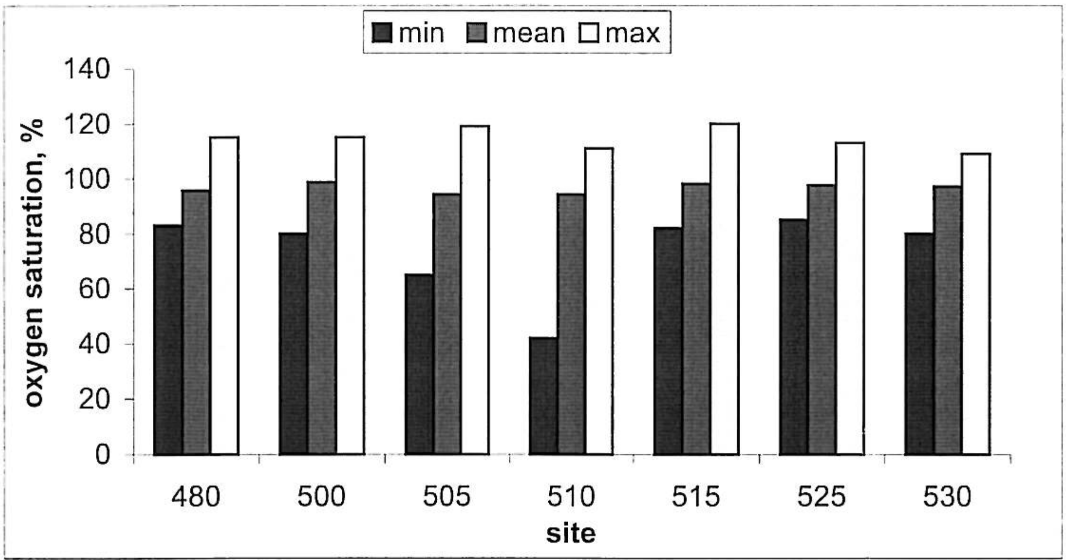

Figure 61. *Variation in oxygen saturation in water layers close to bottom (bottom +1 m) at various stations. The figure presents the average value of each station as well as measured minimum (min) and maximum (max) values for the years 1990-2001 (Ikonen et al. 2003).*

Table 13. *Oxygen saturation (%) of water in winter and open water season 2004 (see Figure 7). Standard deviation is shown in brackets. Table after Kirkkala (2005) in Haapanen (2005).*

Location	Depth, m	Winter	Open water season
SEA05	0-5.5	92 (9.2)	100 (5)
SEA06	0-10	103 (2.1)	101 (6)
	12-13		96 (10)
SEA03	0-10	97 (1.4)	102 (6)
SEA09	0-8	95 (2.1)	101 (6)
SEA10	0-10	-	107 (5)
	13	-	101 (10)
SEA08	0-8.5	106 (8.7)	101 (8)
SEA07	0-7	98 (2.0)	101 (5)

6.9 Runoff

Runoff carries nutrients and other element supply due to varying climate and climate changes and it is a sum of direct river and diffusive runoff. Changes in sea-level migrations and land-use also influence on runoff. To evaluate the change of pressure on element supply to the Baltic region it is necessary to know the variability of runoff and normalise for this natural variability. In all sub-regions a strong seasonal, annual and decadal variability can be distinguished. Especially wet and dry periods are significantly affecting on the runoff.

Geographically, the runoff is of about the same size in the Gulf of Bothnia, Gulf of Finland and the Baltic Proper (Figure 62). During the last four years runoff has fallen from a high to a very low annual value, which could lead to a slight increase in surface layer salinity and a lower nitrogen concentration in the Baltic Sea sub-basins. The most dramatic decrease took place in the runoff to the Gulf of Bothnia and to the Gulf of Finland. The runoff during 2003 to the Bothnian Sea was the lowest since 1921 (e.g. www.balticseaportal.fi, www.helcom.fi).

The annual fresh water inflows to the Baltic Sea are about 14 000 m^3/s. During the period 1950-2003 total runoff to the Baltic Sea area has not shown long-term trends (Figure 63). On the other hand, this time period was characterised by dry and wet periods lasting for a couple of years to a decade and largely following the NAO index (North Atlantic Oscillation Index, e.g. Hurrel 1995). From a regional point of view, the amount of runoff which enters to the Gulf of Bothnia is 6 000 m^3/s, the Gulf of Finland and the Baltic Proper 3500 m^3/s and to a lesser extent in Gulf of Riga and in Kattegat 1 000 m^3/s, on average (Bergström & Carlsson 1994).

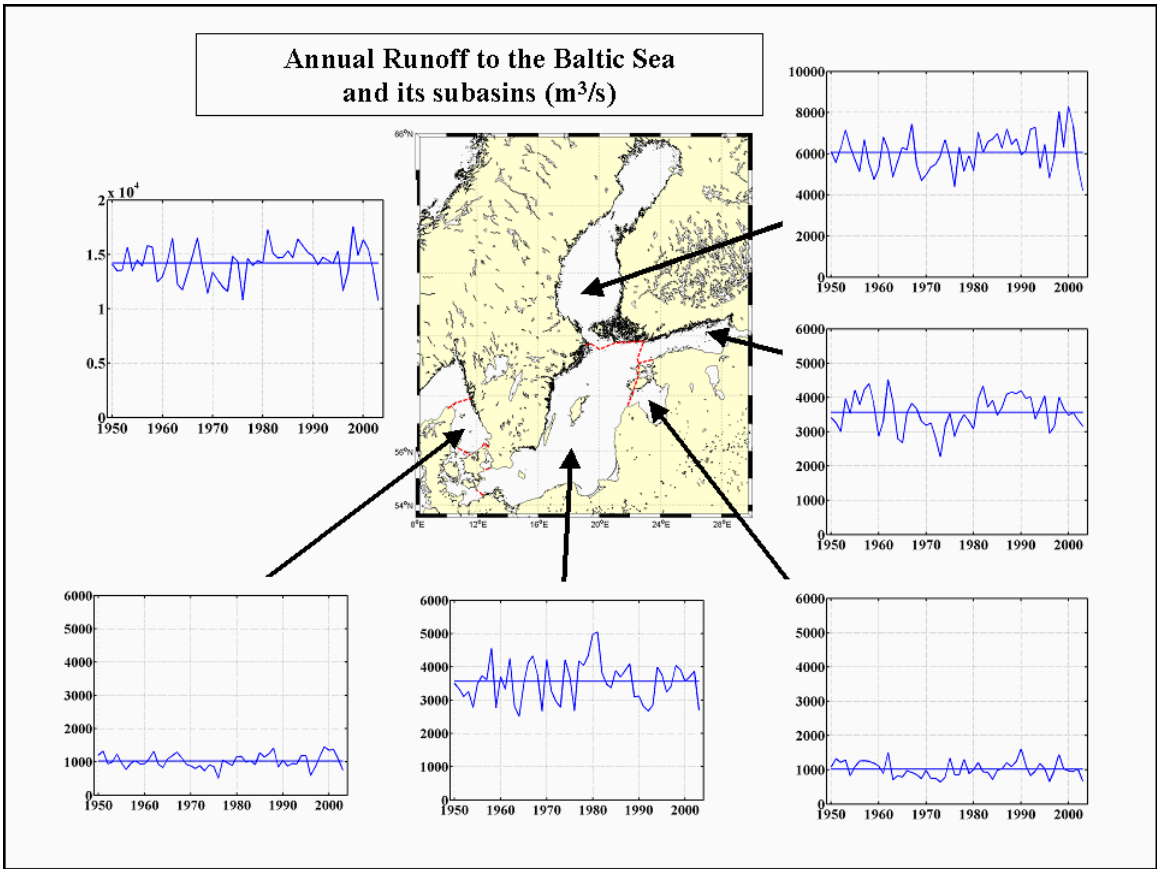

Figure 62. *Annual runoff of the Baltic Sea and its sub-basins (m^3/s) (www.fimr.fi).*

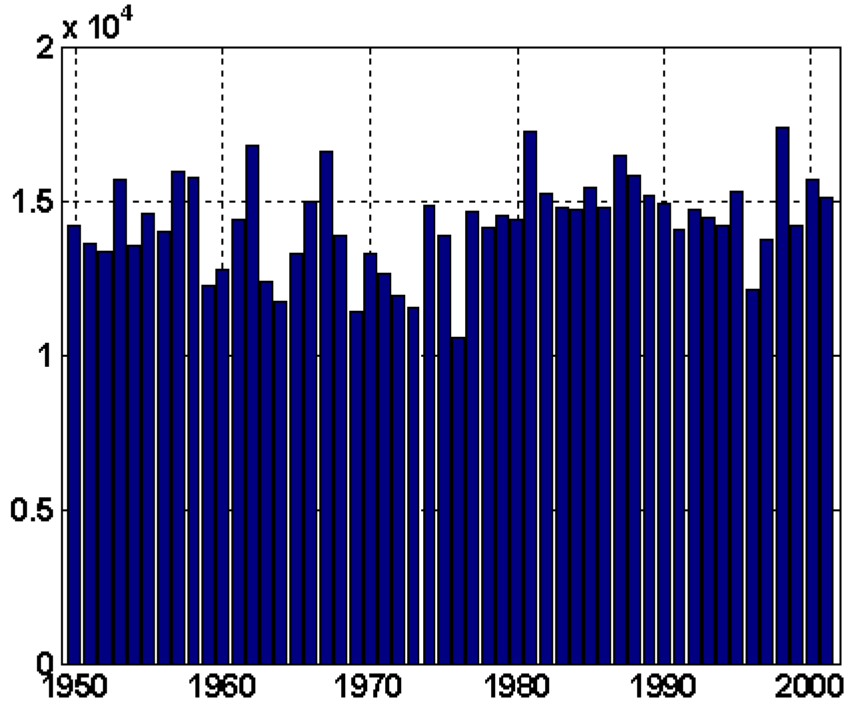

Figure 63. *Total runoff (m^3/s) to the Baltic Sea based on annual mean from 1950 to 2001 (www.helcom.fi). During the last 50 years, annual runoff into the Baltic Sea has remained approximately stable. While long-term cyclical fluctuations with alternating wet and dry periods are observable, year-to year variations are dominant along with the annual runoff cycle (www.fimr.fi).*

6.10 Radioactivity

The Baltic Sea has had intensive radioecological monitoring programmes early as late 1950's. Samples have been taken from water, sediments, fish, aquatic plants and benthic animals. Radionuclides are naturally present and due to human activities. The nuclear weapon tests carried out mainly in the 1950s and 1960s, as well as nuclear power plant accidents have caused radionuclide releases, for example. The radioisotopes iodine, caesium and strontium are the main radionuclides originating from these sources. H-3, S-14 and radioisotopes of iodine are important substances contained in nuclear power plants and spent nuclear fuel reprocessing plants. Contamination from all these sources is primarily in the form of fallout (Greger 2004).

Cs-137 is the most important radionuclide released in the environment in the Chernobyl nuclear power plant accident in 1986. The fallout came down very unevenly in the Baltic drainage basin. The eastern Gulf of Finland and the Bothnian Sea received the heaviest loads (Figure 64). The highest total amounts of Cs-137 in sea bottom sediments occurred in these gulfs, and the scattered nature was further emphasized as a consequence of river discharges, sea currents and varying sedimentation rates on erosion and soft sedimentation bottoms. Levels of Sr-90 and Cs-137 are higher in the Baltic compared with other seas.

Overall levels of radioactivity in the Baltic Sea water, sediments and biota have shown declining trends since the Chernobyl accident. The total input of Cs-137 from Chernobyl to the Baltic Sea was estimated at 4700 TBq. In 1998 the total inventory of Cs-137 in the Baltic Sea sediments was estimated at 1900-2000 TBq. The amounts of Cs-137 in sediments are so low that they do not cause any hazard to human or aquatic organisms. Radioactivity is now slowly transported from the Baltic Sea to the North Sea via Kattegat. Minor amounts of radioactivity from Sellafield are transported in the opposite direction.

The dose rates are predicted to have peaked in 1986 at a value of 0.2 mSv/y, which is below the dose limit of 1 mSv/y for the exposure of the general public set out in the EU Basic Safety Standards (1996). It is unlikely that any individual has been exposed from marine pathways at a level above 1 mSv/y dose limit considering the uncertainties involved in the assessment. Doses to man due to liquid discharges from nuclear power plants in the Baltic Sea area are estimated at or below the levels mentioned in the EU Basic Safety Standards to be of no regulatory concern (individual dose rate of 10 μSv/y and collective dose of 1 manSv). In Finland, the average radiation dose from other sources is estimated to be 3.7 mSv, half of which from breathing radon in air. It should be noted that the assumptions made throughout the assessment were chosen to be realistic and not conservative. Consequently, this also applies to the estimated radiation doses to man (Ilus et al. 1999, Ilus et al. 2000, Ilus et al. 2003, Ikäheimonen & Muslow 2003).

The total collective dose from man-made radioactivity in the Baltic Sea is estimated at 2600 manSv of which about two thirds (1700 manSv) originate from Chernobyl fallout, about one quarter (650 manSv) from fallout from nuclear weapons testing, about 8% (200 manSv) from European reprocessing facilities, and about 0.04% (1 manSv) from nuclear installations bordering the Baltic Sea area (www.helcom.fi).

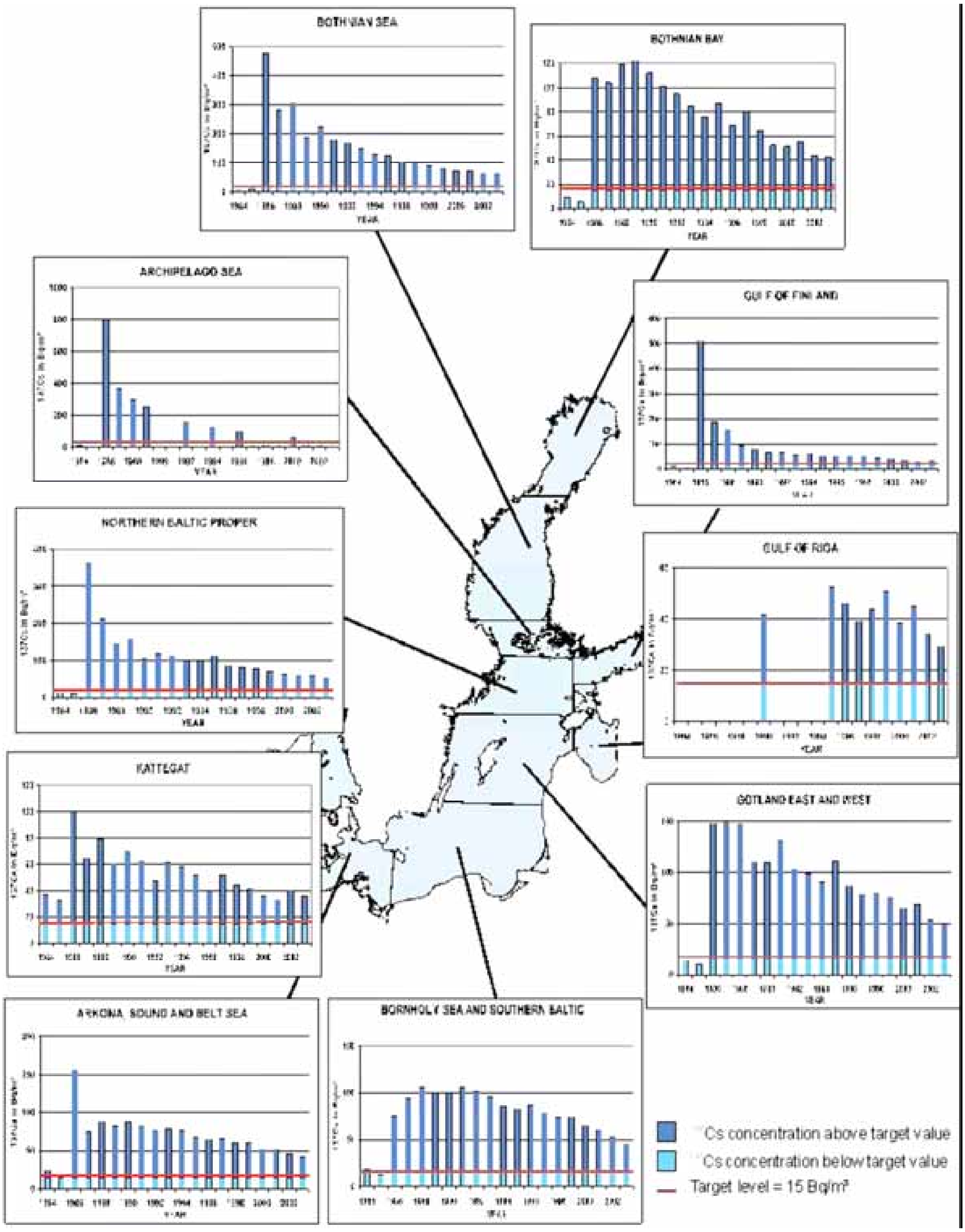

Figure 64. *Ceasium-137 concentrations (in Bq/kg wet weight) in surface water (sampling depth < 10 m) in 1984-2003, as annual mean values by basin. Target value has been calculated as average of pre-Chernobyl (1984-1985) concentrations. (Note: variable scales in the graphs) (Helcom 2006).*

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At Olkiluoto, the nuclear power plant have had an extensive surveillance programme for radionuclides in the sea ecosystems (e.g. STUK 1992, 1995, 1997a, 1997b, Ikäheimonen et al. 1995, Ilus et al. 1987, 1992, 1993, Klemola et al. 1991, 1998, Ikonen 2003, Roivainen 2005, Roivainen 2006). Samples are taken at varying intervals from seawater, suspended matter, and sea bottom sediments, some bio-indicators (bladder wrack, green alga, Baltic clam, and blue mussel) and from the main species of fish (Roivainen 2005). In addition, the surveillance programme includes some species of bottom vegetation quite near the cooling water outlet, representing the sea volume directly affected by the licensed discharges from the nuclear power plant (Ikonen et al. 2003).

Due to fallout from the Chernobyl accident Cs-137 concentration in seawater and sea sediments are still more than three times higher than the concentrations prior to accident. The number of nuclides originating in Chernobyl and not usually detected in the environment samples can be even 20 or more, from which majority is short-lived. This gives some data on the behaviour of certain less studied nuclides, but on the other hand, expands the data greatly for a relatively short period time. The description of overall situation and trends can be found in Ikonen et al. (2003) and Roivainen (2005).

The natural abundance, e.g. Be-7 and K-40 are found almost in every sample, whereas other nuclides are rarer. However, the concentrations of the natural nuclides represent natural variations in the ecosystem and also reflect the quality of measurement. In addition, the behaviour of these elements and nuclides is relatively well known enabling e.g. model adjustments against the measurements. Thus, also these nuclides are important to be reported (Ikonen et al. 2003).

A new kind of a contamination route has to be assessed in connection with the geological disposal of spent nuclear fuel, as the radionuclides potentially released from the repository are carried in the groundwater to the forest food cycle (Avila 2006). At the area in many cases in Olkiluoto, releases from the facilities will probably not take place until in the far future (Koch-Steindl & Pröhl 2001) when the discharging contaminated groundwater can be converted into water available to organisms at the boundary surface of the top bedrock layer and the soil and the sediments of the sea bottom (Posiva 2005). Contamination carried in groundwater could have two probable routes; the radionuclides that migrate from groundwater to the sediments of the sea bottom end up in mire and forest cycle as a result of land uplift. However, in a future ecosystem development will change in several stages due to the land uplift. The other probable contamination route is directly via groundwater. Groundwater flows towards areas located on low ground and in these areas groundwater often erupts as springs or directly into soil. The groundwater erupted as springs can migrate further in brooks, rivers, lakes and sea, or filtered back into soil (Roivainen 2006).

7 CONCLUSIONS

The geological history of the Baltic Sea has been very diverse resulting in profound changes in the hydrographic, chemical, physical and biological conditions of the sea and its catchment are. The future geosphere-biosphere state in the Baltic Sea is mainly determined by land uplift, sea level changes and climate and climate related changes. In addition, to natural processes, human activities are to be considered potentially significant cause of ecosystem alternations.

Two different climate scenarios have been selected, namely Weichselian-R and Emissions-M for the Safety Case and modelling. In the Weichselian-R scenario it is assumed that previous Weichselian glacial cycle will repeat itself and it is in line with the SKB's main scenario. Greenhouse emissions induced climate change is not taken account. In the Emissions-M scenario a moderate increase of greenhouse emissions will contribute to the natural CO_2 concentrations in the atmosphere and thus will rise the global temperatures.

At present the Fennoscandian lithosphere is undergoing postglacial rebound and a rebound still is going on. Different models have been used to describe land uplift at Olkiluoto area from present to near future. Shoreline displacement will effect on many processes, e.g. it displaces coastlines, shifts watersheds, causes differentiation in sedimentation and erosion rates and amounts and affects the local climate conditions by changing topographic conditions and the local ration of land and sea. In the shallow shores of Olkiluoto, the amounts of common reed are increasing naturally, resulting in paludification of coves and accumulation of organic matter in shallow and nearly-stagnant water. These can result in a faster apparent shoreline displacement than mere land uplift or changes in sea-level would yield. In addition, eutrophication of the Baltic Sea speeds up the process. During the time span present Olkiluoto will be connected to the Finnish continent.

The future scenarios predict that that the climate is warming up and the sea level will raise in near future, but however, the estimations varies. At Olkiluoto, it seems that land uplift will exceed the eustatic rise also in the future. Only in the most dramatic scenarios could the extremely rapidly rising sea-level exceed the land uplift for a time period of some hundreds years.

Based on the data of current sea bottom sediments, future ecosystems at Olkiluoto area can be forecasted, at least to some extent. It can be estimated that current prevailing forest types will also prevail in the future, but more wetlands and deciduous forests around the water bodies are predicted.

The models need still more data by involving a sedimentation-erosion model with data on the seabed quality and depths and margins between future ecosystems (e.g. land and aquatic environments). The additional and more precise and reliable data will make it possible to model the evolution of topography more exact scale and make estimations about the spatial and temporal extents and routes of lakes, rivers and mires. All these future changes will have a significant effect on local and regional hydrological conditions, e.g. to groundwater flow, salinity, discharge and recharge areas as seen in the past.

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APPENDICES

APPENDIX 1: List of main monitored parameters at the Olkiluoto offshore.

APPENDIX 2: Physical, chemical and biological features, which increase or may increase the vulnerability of the Baltic Sea ecosystem to anthropogenic chemicals compared to marine of freshwater environments addressed within the OSPAR Commission and EU framework.

APPENDIX 1.

List of main monitored parameters related to the Olkiluoto offshore.

PARAMETER	UNIT	OLKILUOTO	REFERENCE
Climate			
Average temperature 1992–2004	^oC	-4,4–17.1	Ikonen 2003, Ikonen 2005
Extreme temperature 1992–2004	^oC	-27.1–31.6	Ikonen 2003, Ikonen 2005
Precipitation total 1992–2004	mm	410–685	Ikonen 2003, Ikonen 2005
Precipitation day max 2002-2004	mm	0.2-21.5	Ikonen 2003, Ikonen 2005
Average relative humidity 1992-2004	%	79–92	Ikonen 2003, Ikonen 2005
Average air pressure 1992-2004	hPa	1008.7–1011.9	Ikonen 2003, Ikonen 2005
Average snow thickness 2002-2004	cm	0-24.8	Ikonen 2003, Ikonen 2005
Main wind direction (Olkiluoto) 1992–2004		South	Ikonen 2003, Ikonen 2005
Wind	m/s	3.0–5.2	Ikonen 2003, Ikonen 2005
Sea Water			
Salinity 1990–2001	‰	5.4 (1.7–6.2)	Ikonen 2003, Ikonen 2005
Salinity 2003	‰	5.2 (2.2–5.7)	Ikonen 2003, Ikonen 2005
Temperature 1990–2001	^oC	11 (1.9–21)	Ikonen 2003, Ikonen 2005
Temperature 2003	^oC	8.1 (5.0–17)	Ikonen 2003, Ikonen 2005
pH 1990–2001		8 (7.5–8.3)	Ikonen 2003, Ikonen 2005
pH 2003		7.8 (7.6–8.0)	Ikonen 2003, Ikonen 2005
Turbidity 1990–2001	FNU	3.1 (0.4–42)	Ikonen 2003, Ikonen 2005
Turbidity 2003	FNU	2.2 (1.3*)	Ikonen 2003, Ikonen 2005
Conductivity 1990–2001	mS/m	937 (310–1060)	Ikonen 2003, Ikonen 2005
Conductivity 2003	mS/m	911 (750–990)	Ikonen 2003, Ikonen 2005
Oxygen saturation 1990–2001	%	99 (83–117)	Ikonen 2003, Ikonen 2005
Oxygen saturation 2003	%	96 (4*)	Ikonen 2003, Ikonen 2005
Density in front of Olkiluoto 2002	g/ml	1,003	Paaso 2003
Oxygen in front of Olkiluoto 2002	mg/l	8.5–9.0	Paaso 2003
HCO_3 in front of Olkiluoto 2002	mg/l	78–80	Paaso 2003
DOC in front of Olkiluoto 2002	mg/l	4.6–5.0	Paaso 2003
DIC in front of Olkiluoto	mg/l	12.9–14.3	Paaso 2003
TDS in front of Olkiluoto 2002	mg/l	5400–5630	Paaso 2003

PARAMETER	UNIT	OLKILUOTO	REFERENCE
Chloride in front of Olkiluoto 2002	mg/l	2910–3050	Paaso 2003
Tot S in front of Olkiluoto 2002	mg/l	140	Paaso 2003
SO_4 in front of Olkiluoto	mg/l	430–440	Paaso 2003
K in front of Olkiluoto 2002	mg/l	55–59	Paaso 2003
Ca in front of Olkiluoto	mg/l	85–90	Paaso 2003
Mg in front of Olkiluoto	mg/l	210–220	Paaso 2003
Na in front of Olkiluoto	mg/l	1610–1730	Paaso 2003
Suspended solids 1990–2001	µg/dm^3	5000 (1200–36000)	Ikonen 2002, Ikonen 2005
Suspended solids 2003	µg/dm^3	4000 (2400*)	Ikonen 2002, Ikonen 2005
Sediment accumulation rate	g/m^2/year 1995	1810	Mattila et al. 2006
Depth offshore Olkiluoto	m	5 (0-30)	Kotilainen & Hutri 2003
Nutrients			
Tot N 1990–2001	µg/dm^3	370 (230–900)	Ikonen 2003, Ikonen 2005
Tot. N 2003	µg/dm^3	390 (200*)	Ikonen 2003, Ikonen 2005
NH_4 1990–2001	µg/dm^3	13 (1-93)	Ikonen 2003, Ikonen 2005
NH_4 2003	µg/dm^3	21 (200*)	Ikonen 2003, Ikonen 2005
Tot P 1990–2001	µg/dm^3	20 (7-52)	Ikonen 2003, Ikonen 2005
Tot. P 2003	µg/dm^3	18 (5*)	Ikonen 2003, Ikonen 2005
Isotopes in front of Olkiluoto 2002			
Rn-22	Bq/l	<1	Paaso 2003
^{18}O	‰SMOW	-7,5	Paaso 2003
^{2}H	‰/MOW	-57,3	Paaso 2003
^{3}H	TU	13.5–16.4	Paaso 2003
$^{34}S(SO_4)$	‰CDT	19.41–20.10	Paaso 2003
$^{18}O(SO_4)$	‰SMOW	9.16–9.29	Paaso 2003
^{13}C	‰PDB	-2,3	Paaso 2003
^{14}C	%pM	107.5–112,0	Paaso 2003
$^{87}Sr/^{86}Sr$		0.7094–0.7095	Paaso 2003
Cl-37	mBq/l	0.9–0.17	Paaso 2003
$^{238}U(H_2O)$	mBq/l	8.05–10.10	Paaso 2003
$^{234}U/^{238}U(H_2O)$		1.09–1.36	Paaso 2003
^{238}U (filter)	mBq/l	0.13–0.37	Paaso 2003
^{238}U (filter)	µg/l	<0,01–0,14	Paaso 2003
$^{234}U/^{238}U$ (filter)		0.60–2.43	Paaso 2003
Radionuclides: Seawater			
Cs-137	Bq/m^3	59 (5*)	Roivainen 2005
K-40	Bq/m^3	2000 (6*)	Roivainen 2005
Sr-90	Bq/m^3	19 (5*)	Roivainen 2005
Co-60	Bq/m^3	<3	Roivainen 2005

PARAMETER	UNIT	OLKILUOTO	REFERENCE
Radionuclides: Sea sediments			
Cs-137	Bq/m^3	620 (4*)	Roivainen 2005
K-40	Bq/m^3	940 (5*)	Roivainen 2005
Sr-90	Bq/m^3	1.8 (7*)	Roivainen 2005
Co-60	Bq/m^3	79 (4*)	Roivainen 2005
Radionuclides: Suspended material North of Olkiluoto			
Cs-137	Bq/m^3	400 (3*)	Roivainen 2005
K-40	Bq/m^3	660 (5*)	Roivainen 2005
Co-60	Bq/m^3	5.2 (10*)	Roivainen 2005
Radionuclides: Sea sediments North of Olkiluoto			
Cs-137	Bq/m^3	480 (3*)	Roivainen 2005
K-40	Bq/m^3	780 (4*)	Roivainen 2005
Co-60	Bq/m^3	5.7 (8*)	Roivainen 2005
Suspended solids			
Eurajoki	metric tons	7000	Ikonen 2005
Lapinjoki	metric tons	2000	Ikonen 2005
Tot. N Eurajoki	metric tons	530	Ikonen 2005
Tot. N Lapinjoki	metric tons	250–300	Ikonen 2005
Tot. P Eurajoki	metric tons	17	Ikonen 2005
Tot. P Lapinjoki	metric tons	6	Ikonen 2005

APPENDIX 2

Physical, chemical and biological features, which increase or may increase the vulnerability of the Baltic Sea ecosystem to anthropogenic chemicals compared to marine of freshwater environments addressed within the OSPAR and EU framework.

Feature	Consequence	Implication for exposure or effects of chemicals
Physical features		
Semi-enclosed sea	Slow exchange of water Minimal tides, low sediment circulation	Trapping of chemicals Stocking up of chemicals in anoxic, deep sediments, occurrence of stable hot spot area (sedimentation areas)
Large, densely populated catchment area	High inflow of freshwater High atmospheric deposition of anthropogenic contaminants	High input of hazardous substances
Shallow compared to the Atlantic Sea	Small water volume compared to seas and hence smaller dilution of hazardous substances compared to seas	High concentration of chemicals
Ice and snow cover	Inhibition of photodegradation and volatilisation	Higher concentrations of photodegradable and volatile chemicals
Short day conditions in autumn and winter	Inhibition of photodegradation	Higher concentrations of photodegradable chemicals
Permanent stratification of water because of halocline. Temporary stratification of water because of thermocline	Inhibition of exchange of water and dissolved substances as well as particulate matter across halocline or thermocline	Concentrations of chemicals
Hydrodynamic fronts, e.g. in the eastern Gulf of Bothnia	Selective sedimentation of metals)	Affects the proportions of chemicals present in different compartments
Chemical features		
Brackish water, salinity range from 0-20 %	Salinity affects speciation of metals	Toxicity of metals in inversely related to the salinity (metals appear in more toxic forms in the low saline water compared to the seawater
Low calcium concentration compared to oceans	Increased permeability of cell membranses	Increased uptake of metals compared to the seawater
Anoxic and hypoxic sediments		Hazardous substances such as metals, PCB and PAH, are often bound in sediments under hypoxic or anoxic conditions. The improved oxygen conditions may temporary increase the mobilisation of some metals from sediments

Feature	Consequence	Implication for exposure or effects of chemicals
Biological features		
Short history of the Baltic Sea (The current salinity has existed about 3000 years)	Organisms are not fully adapted to live in the Baltic Sea → low biodiversity	One consequence of low biodiversity is that the Baltic Sea has only few key species, i.e. species that have an important ecological role in the ecosystem. If these species would decline, there are now species taking over their functions
Species living in the Baltic sea are originally marine or freshwater species and thus live close to their physiological tolerance limits regarding the ambient salinity		Hypothesis: Species living in the Baltic Sea are more vulnerable to chemicals compared to marine or freshwater species) There are very little studies to verify whether brackish water species are more, less or equally sensitive to chemicals compared to marine and freshwater organisms. Different type of water may also affect the toxicity of chemicals
High sedimentation rates compared to oceans	Efficient input of particle-bound contaminants to sediments	Hypothesis I: Increased sedimentation reduces the bioavailability of pollutants and increases bioludition Hypothesis II: The sediment may function as a source of hazardous substances, when the input to the sea stops. Due to equilibrium partitioning will substances end up in the water from the sediment, if the concentration in the water gets low enough