2. Past climate variability

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2.1. Introduction and summary *(5 pp = 7 Word pages without figures)*

Introduction

Quantitative past climate data obtained using new methods and technology have improved the testing and tuning of numerical climate models. In addition, this data helps understanding causes and mechanisms of climate variations. The IPCC Report (2007) includes paleoclimatic materials as an individual Chapter where it is emphasized that historical climate data might be very important when studying global climate system sensitivity to various internal and external factors. In practice, almost all future climate scenarios are based on the gradual, though relatively fast, increase in global temperature due to greenhouse gases concentration growth in the atmosphere, primarily, carbon dioxide and methane.

However, analyses of the paleoclimatic records show that relatively long-term past warm periods were followed by a shift to much colder weather, and vice versa, cooling periods changed to rapid (of the order of a few decades) warming. These abrupt climate changes are assumed to be related to nonlinear processes in the climate system. At some threshold values the climate system can transit “jump wise” from one stable condition to another one, almost “instantaneously” within first few decades. Chemical analyses of air trapped in Greenland and Antarctica ice cores indicate that the periods of abrupt warming correlate with high concentrations of greenhouse gases in the atmosphere and cooling with their low values. The mechanism of these abrupt climate events is likely to depend on a massive influx of fresh glacial meltwater into the ocean and intensified hydrological cycle under global warming. In high latitudes, abrupt climate changes are most noticeable, especially in the areas adjoining the continental ice sheet. The Baltic Sea Basin is believed to be a key region in studying the causes of the past abrupt climate change. Independent tree-ring, ice-core, clay-varve, tephra ash layers,

^{14}C, ^{10}Be and other chronologies have been used to develop a detailed timescale with the accuracy of up to few years for Lateglacial-Holocene climate events.

The first timescale of climate events for this period was established for the Baltic Sea Basin about 100 years ago. For many years it has been used as a classic timescale to analyze paleodata from other regions of the Northern Hemisphere. Analyses of multiproxy data made it possible to quantitatively estimate the amplitude of climatic variations in different parts of the Baltic Sea Basin. As a result, an assessment has been made of environmental implications of these variations and their influence on the early human migration in this region. Studying these processes based on paleodata analysis is directly related to developing long-term future climate forecasts for next decades. Of particular interest is the climate change data over the past thousand years.

Summary

History of the basins that had existed on the territory of the modern Baltic Sea since the last deglaciation has been already studied for more than a hundred years. Nevertheless, the issues are still questionable of both the chronologies of transgression-regression phases of the pre-historic Baltic basins and their spatial characteristics. Greenland ice cores with annual layers, abundant radiocarbon dates obtained from studies on lake and continental sections, varve chronologies and other data allow us to obtain both a detailed time frame of climate events for the past 14,000-15,000 years and to determine the age of warming-cooling boundaries revealed from spore-pollen records and tied to the classic Scandinavian scale. Tree-ring chronologies are very important in reconstructing the Lateglacial-Holocene events. Attempts have been made to extend tree-ring Holocene chronologies to the Lateglacial by updating them with the "floating" sections of the dendrochronological scale. The final deglaciation stage in the Baltic Sea Basin is supposedly related to an abrupt warming at the Bølling/Allerød interstadial which started about 14,600 cal yr BP. At that time arboreal vegetation first appeared in ice-free regions of the Baltic Sea Basin. The warming was interrupted with a series of cold events. The strongest cooling of approximately 1000 years long (the Younger Dryas) has begun at about 12,700 cal yr BP and ended about 11,600-11,500 cal yr BP, at the Holocene boundary. About 11,200 years ago, a cold interval– the Preboreal Oscillation – occurred which then shifted to an extended warming of 2000 years long and the arboreal vegetation expanded rapidly throughout the Baltic Sea Basin. The subsequent two cooling periods about 9,300 and 8350 – 8150 are indicated in Greenland ice cores, North Atlantic marine sediments, clay-varve chronologies, and other proxy data.

The second 8.2 ka cool event of about 160 to 200 years long has been known long ago. This cold episode correlates with mountain glaciation in the Alps and Scandinavia and reduction in warm-loving tree species in pollen spectra from different parts of the Baltic Sea Basin. These cold

episodes are theorized to be related to a massive melt water surge outflow to the North Atlantic from Ice Lakes Superior and Agassiz that were formed after the destruction of the Laurentide Ice Sheet. The 8.2 ka event was the last cooling episode of the Early Holocene followed by a stable and relatively warm climate period with summer temperatures of 1-2^{0}C higher than the present ones.

The period between 7,500 and 5,500 cal yr BP ago was the warmest one for the entire Baltic Basin area, though the times of maximum temperatures were not synchronous in different parts of the region. In Sweden territory, pollen and chironomids chronologies show that the times of maximum temperatures changed in parts of Sweden between 7,900 and 5,700 cal BP with the amplitude varying from 0.8 to 1.0^{0}C and higher. In the Northern Finland, maximum temperatures took place between 7,500 and 7,000 cal BP, and in north-western Russia, about 5800 to 5000 cal BP. The negative Northern Hemisphere temperature trend and increased climate instability are typical of the Late Holocene interval. The Baltic Sea region cooling about 5000 to 4500 cal yr BP coincided with decreased summer solar radiation incoming to the earth's surface due to astronomical factors. Greenland ice core data indicates significant oscillations in the concentration of greenhouse gases in the atmosphere, in particular, methane. Chironomids and pollen proxies allow us to reconstruct a two-stage nature of air temperature decrease in the Late Holocene. The first stage occurred between 5,000 and 4,500 cal yr BP and the second one between 4,300 and 3,300 (2,800) cal yr BP. During each period the temperature drop was not less than 1^{0}C. A complicated climate changes in the Baltic Sea region in the Lateglacial and the Holocene was reflected in lake levels' status, vegetation changes, and in the formation of a complex hydrographical network. These environmental changes affected the stages of ancient man migration in this territory.

2.2. Climate changes during the Holocene (the last 10.000 yr)

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It is still preparating (about 3 pages)
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ANNEX
Ax Climate and environmental changes during the Lategglacial time in the Baltic Sea Basin (15,000 – 11,500 cal yr BP) (*about 10 – 15 Word pages without figures*)
It is still preparating

2.2.1. Methods and sources of the paleoclimatic reconstructions for the Holocene time

2.2.1.1. Pollen and plant macrofossils analysis. The use of paleobotanic data for palaeoclimatic reconstructions assumes that flora of a particular region, or the composition of plant species growing there, experiences direct influences of the natural environment as a whole and climate in particular. In reconstruction of palaeoclimatic conditions during the Late Quaternary time, the method for denominating changes in bio-or zoocenoses is most frequently used. This method is based on revealing ecologically indicative species of land flora or marine microfauna most dependent on summer or winter air or water temperatures. The ecological-palaeontological methods are most widely applied to estimate the thermal and moisture regimes. There are three ways of using palaeontological and palaeobotanical data:1. Method of species or genetic indicators; 2. Method of morphological indicators; 3. Method of changing dominants in biocenoses.

The method of species or genetic indicators is based on interpolation into the past of ecological requirements of modern organisms that are widely distributed in terms of geology and have distinct areas of existence. There are a few modifications of this method: 1. Arealographic, proposed by Iversen and Shafer and elaborated by Grichuk (1985), 2. Landscape-phytocenological, suggesting transition from definite plant communities (tundra, forest, steppe, etc.) to quantitative climate indicators, 3. Information-statistical, uniting a few approaches, including usage of different "transfer“ functions.

The arealographic method is described as having two approaches: first is based on cartographical summing up of areas with modern species detected in fossil pollen composition (or leaf imprints), which reveals the region of their joint growth thus representing an analogue of past climatic conditions and second is based on superimposing the climatograms constructed using data from meteorological stations located in the area containing fossil pollen. Then climatograms of all kinds are superimposed, and after that the plot common to the entire fossil flora complex is determined.

The classic work on using the arealographic method is that by Iversen (1944), who examined the present distributions of *Ilex aquifolium* (holly), *Hereda helix* (ivy), and *Viscum album* (mistletoe) in North Europe. By plotting modern climatic parameters at sites where fossil remains of these vegetations have been found, Iversen was able to estimate the temperature in the Lateglacial and Holocene time. This approach is still applied to quantitative palaeoclimatic reconstructions (e.g. Zagwijn, 1994).

V.P. Grichuk (1969, 1985) and Grichuk et al., (1984) developed method for reconstructing main climatic indicies from fossil plant data using a concept by Szafer (1946), who proposed locating modern analogue of a palaeoflora by comparison of present-day ranges of plant components of flora. V.P. Grichuk has shown that as a certain group of plant species is limited in its geographical spread by definite climatic conditions, the presence of this group of species in the fossil flora indicates that the climatic values at the time of its existence were within the climatic limits of a region where this group of species grows at present. The region currently inhabited by the species of a certain fossil flora derived from palynological and plant macrofossil data is determined cartographically by superimposing the modern ranges of these species (the method of arealograms). This method has been widely used by Russian researchers for reconstructing the Lateglacial and Holocene climate. The accuracy of determining summer and winter temperatures is ± 1°C, and annual precipitation is ± 50 mm (Borisova, 1990,1997).

The landscape-phytocenological, or zonal method, of palaeoclimatic reconstruction based on the law of natural zonality determines modern and past vegetation and soil distribution (Khotinsky, 1977; Borzenkova, 1992). When reconstructing the pattern of natural zonality in the past by pollen-spore or some other proxy data, it is possible to reconstruct the climatic indicators typical of each landscape zone. In spite of the fact that this method yields a far lower accuracy of climatic parameters compared to the arealographic and information-statistical ones, it enables an accurate reconstruction of landscape-climatic zone boundary shifts.

In the last few years alternative methods for climate reconstruction from pollen data have been developed. These methods are based on multivarious statistical techniques. A number of different mathematical techniques can be used for developing a "transfer“ function, ranging from multiple regression to canonical correlation analysis. The generation of "transfer" functions for pollen data requires a collection of modern surface pollen samples mathematically correlated to modern climate parameters to produce a calibration model. The transfer functions have been developed for different parts of the Baltic Sea Basin and can be applied to climate reconstructions in the Lateglacial – Holocene time (Antonsson, 2006; Antonsson et al., 2006; Antonsson, Seppä, 2007; Birks, 1995; Heikkilä, Seppä, 2003; Seppä et al., 2004, 2005; Sepä, Birks , 2001; Sepä, et al., 2002, , Seppä et al., 2008, etc). A pollen-climate calibration model has been recently developed to quantitatively reconstruct the Holocene annual mean and summer and winter air temperatures in northern Europe (Seppä et al., 2008; Birks, Seppä, 2010). The model input has been derived from modern pollen dataset from Finland, Estonia, and southern and central Sweden. This north-European pollen-climate response model, or versions thereof, has been used to reconstruct Holocene temperature and climate patterns in northern boreal and southern boreal regions of the Baltic Sea Basin (Antonsson et al., 2006; Heikkilä, Seppä, 2010).

2.2.1.2. Fossil insects (Coleoptera and Chironomid) data. Arealograms and information-statistical methods are successfully used to process data on fossil insects (Coleoptera records and Chironomid evidence) (Coope et al., 1998; Walker, 2001). Fossil Coleoptera (beetles) can serve as a good proxy tool in the reconstruction of Lateglacial and Holocene climate changes in a portion of the Baltic Sea Basin. Most paleoclimatic studies have been carried out in this region, where modern temperate species of Coleoptera fauna were replaced with altered boreal or polar assemblages in glacial time, and with more southern warm–loving species during the interstadials. Beetles react rapidly to climate change, which results in minimal climate-fossil data lags. Coleoptera fauna has been widely employed to generate both quantified temperature estimates and thermal gradients during the Lateglacial in Northern Europe. Chironomids (non-biting midges) are of special interest in palaeoclimatology because their strongly sclerosal larval head capsules are stratigraphically preserved in lake sedimentary deposits. By using inference models that link present distribution and abundance of chironomids to contemporary climate, the past climate can be quantified from fossil assemblages. Inference models based on chironomids are increasingly used to quantify past environments, and have been developed as a tool to track past changes in air and water temperatures. Chironomid populations respond rapidly to climate change occurring in a wide variety of environmental conditions (Antonsson, 2006; Brooks, Birks, 2000, 2001; Larocque et al., 2001; Luoto, 2009 a,b; Velle et al., 2005; Walker, 2001). Statistical comparison between modern chironomid assemblages and July temperature values enables training sets to be developed, which provides a basis for quantitative estimation of summer temperature during the Lateglacial (Antonsson, 2006; Velle et al., 2005). Finally, by comparing pollen inferred temperatures with independent proxy records, (e.g. chironomids and oxygen isotopes), it is possible to create a more comprehensive picture of past climatic pattern in the Baltic Sea Basin.

2.2.1.3. Dendroclimatology evidence. Dendroclimatical data is most successfully used to reconstruct climate in those regions where interannual variability is a determining factor (northern limit of the forest and vegetation upper boundary). Various dendrochronological indicators are used as indicators of climate change. They are tree-ring width and density, and the availability of frozen rings (Briffa et al., 2001). The stable isotopes of hydrogen and carbon stored in tree-rings‘ cellulose can be used for reconstruction of temperature and humidity during the warm period of a year. Analysis of radiocarbon (^{14}C) from tree-rings has been used to establish high-resolution radiocarbon chronologies. The density of tree rings determined by X-ray densitometry is a much more informative characteristic of past climate compared to the conventional data on tree-ring width. In recent times, this technique has been widely accepted in many countries, where the data on tree-ring density of coniferous (predominantly pine, black fir, and oak trees) are used to reconstruct the summer air temperature. Long Hohenheim Holocene chronologies consist of oaks and pines tree – ring data from the southern part of Germany and cover large portions of Lateglacial, extending into the Younger Dryas back to about 12,000 BP (Friedrich et al.,. 1999).

2.2.1.4. Isotopic and geochemical methods. The isotopic paleothermometry method is based on the assumption that there is an equilibrium exchange between water and carbonate oxygen isotopes (^{16}O and ^{18}O) (Dansgaard, 1993; Beerling et al., 1995; Ruddiman, 2008; Stauffer et al., 2002). Actually, this assumption is fully realistic only for inorganic (chemogenic) sediments. The oxygen-isotopic paleothermometry method is being widely applied not only to organic carbonates of marine origin but also to other natural objects such as fresh-lake sediments (algae, lake lime), inorganic carbonate sedimentation of caves (stalactites and stalagmites), and continental glaciers (mountain glaciers, polar ice sheets) (Baldini et al., 2002; Hummerlund et al., 2003).

The oxygen-isotopic studies of lake sediments and ice cores appear to be promising. The ice cores obtained by drilling the continental ice sheets in Greenland and the Antarctic are the richest sources of different palaeoclimatic information. Along with conventional information about air temperature variations, data has been obtained on atmospheric gas composition (CO_2, methane and other greenhouse gases), the intensity of explosive type volcanic eruptions and troposphere aerosol concentration in the past (Schmincke, 2004; Zielinski et al., 1996; Zielinski, 2000). Data on atmospheric gas composition in the Late Pleistocene - Holocene is of great scientific and practical value both for understanding the mechanism of past climate change and for substantiating future possible climate changes influenced by anthropogenic growth of atmospheric carbon dioxide and other greenhouse gases.

2.2.1.5. Other evidence. Among other indications of past climate change, of wide use is the diatomic analysis of lake sediments, data on varve width depending on summer water temperature, and archaelogical artefacts. Archaelogical artefacts (ancient man tools, land cultivation traces, ceramics) primarily indicate to improved climatic conditions, in particular to increased summer and winter air temperatures and humidity in the Baltic Basin area (Dolukhanov et al., 2007, 2009 a,b; 2010; Pazdur, 2004, Poska et al., 2004, 2008) . To restore past landscapes and climate, the entire set of proxy evidence is used. This makes these reconstructions more accurate.

2.2.1.5. Dating control. In palaeogeography determining the age of different events has always been an important task. Many dating methods have been developed for the Baltic Basin, e.g. varve chronology started by De Geere in the early 1900s. Developing dating methods leads to obtaining a great data massif for different sections and to the necessity of correlating different regional scales among eath other.

Different methods have been used to create the chronology of Holocene events. They can be groupped as follows. The first group unites methods that date sediments according to the content of different isotopes with the known period of disintegration. For the Holocene, it is primarily isotope ^{14}C. The second group includes methods based on studies on characteristics of some biotic and abiotic elements related, for instance, to annual formation of tree rings, annuals layers of the lakes sediments, seas, as well as annual ice layers in modern ice sheets. Specific is tephrochronology based on identifying interlayers of volcanic ash in different sediments located at large distances from the source vulcano, which allows correlation of sea, lake, land, and ice sections.

Each dating method has its own limitations. For example, determining primary content of ^{14}C isotope in a sample is among the basic problems. In the atmosphere, there are variations of ^{14}C isotope with different periods depending on oscillations of Earth's magnetic field, variations of solar activity, and changes in ocean circulation.

In addition to these planetary causes resulting in increased or decreased ^{14}C content in the atmosphere, its primary amount in a sample can change because of local and regional factors. The global variations in the primary ^{14}C isotope content can be corrected when calibrating radiocarbon datings. Taking into account local variations is of special consideration. Thus, for instance, lake sediments represent a good source of palaeoclimatic information and can be dated directly with the radiocarbon method.

However, presence of carbon sediments on a watershed can decrease ^{14}C isotope content in water due to dissolving and outcropping of ancient carbon with ground water, which thus can increase the sediment age.

There is a lot of time scales based on the methods for calculating annualy formed layers and measuring their characteristics such as thickness, density, etc. The Swedish scale is based on calculations of changing annual layer pairs in varve clay deposited on the Baltics Sea Basin after ice sheet retreat. It covers more than 13,000 years. Data massif on changing thickness and density of tree rings of pines and oaks in Germany allowed construction of continuous dendrochronological scale covering almost 12,000 years (Friedrich et al., 1999). Dendrochronological data served as the base for developing the calibration system for radiocarbon data on the Holocene that allowed transformation of instrumental age of a sample (always differring from true age) into its calendar age. It is worth mentioning that chronological scales based on calculating varves or tree rings can have gaps and thus not being able to form a continuous scale.

Ice cores from wells drilled in modern ice sheets are the most detailed source of palaeoclimatic information. Analysis of variations in the air isotope content in annual ice layers issues the data on basic climate trends of Earth for the last several hundred thousand years. The temperature estimate at the YD/PB boundary (Grachev and Severinghous, 2005) based on the ratio of different isotopes of nitrogen and argon is 10 ± 4ºC. Detailed isotope data from Greenland cores has been recently used to create a detailed dating scale of climatic events of the Late Pleistocene (Rasmussen et al., 2008).

In addition to limitations typical of every dating method there are common problems for instance in the extent of time coincidence of natural changes reflected in sediments. Therefore, in recent decades, among the tasks in Quaternary studies, of most importance is the interregional correlation of data obtained from sediments of different types with different methods (Lowe et al., 1994; 2008).

One of the methods for correlating different type sediments is the tephrachronology. Climatic changes in the Late Pleistocene – Holocene fixed with biostratigraphic or geochemical methods were not simultaneuos on a regional scale. In case of inability of dating a sediment stratum of interest with the traditional methods, the tephra method can be applied to determine the age of the given stratum if a volcanic ash is present. With the available datings tephra interlayers represent an additional independent instrument justifying or correcting the age determined by means of direct dating methods. Iceland tephra is proving to valueble chronological tool for climate and eviromental reconstruction in the Northen – Western Europe (Haflidson et al., 2000; Hummerlund et al., 2003).

In Europe, during the Pleistocene-Holocene, a few volcanic regions were active. In the territory of Iceland, there are 32 volcanic systems that were active during the Holocene (Haflidason et al., 2000). Bottom sediment cores from North Atlantic and Norvegian Sea show volcanic ash interlayers *silicic* composed of material from more than 50 large eruptions occurring in Iceland and Yan-Maien during the last six million years (Lacasse, Garbe-Schönberg, 2001). In the Baltic Region, ash layers from Iceland and Aifel (West Germany) are mostly found in lake and swamp sediments. As many as ten ash interlayers beginning with the age of 14,400 cal yr BP have been identified in the territory of Sweden (Wastegård, 2005). Ash Vedde whose interlayers with the age of 10,300 ^{14}C cal ago or 12,120 cal ago are widely distributed in North Europe and can serve as an example of using tephrachronology as a stratigraphy method. Glass of this ash has been found in sea, lake, swamp and ice cores in the area from Greenland to North-Western Russia and from Norway to Switzerland (Davies et al, in press). Location of Vedde layers in sediments formed at YD time are indicators of opposite trends of increasing/decreasing pollen *Artemisia* and *Chenopodiaceae* in sections depending on sublatitude disposition, which indicates to non-synchronous response of natural objects to climatic changes even within North Europe.

Other evidentiary materials (artifacts, ceramics, and traces of ancient land-use) are of particular importance for reconstructing moisture conditions in arid regions where all other kinds of information are almost unavailable.

2.2.2. Climate Variability during the Holocene in the Baltic Sea Basin

The Annexes present a more detailed analysis of climate and environment in the Baltic Sea Basin during the Lateglacial (See Supplement, p.)..

2.2.2.1. The Early Holocene oscillations

Complex studies of ice cores from central Greenland with time resolution of the first decades allowed the developing of detailed scale of climate events during the Lateglacial-Early Holocene. These studies are also useful in determining the calendar age of the events and estimating the climate change speed and amplitude (Björck, 1995; Dansgaard et al., 1989; Grachev, Severinghaus, 2005; Hoek, Bohncke, 2001; Hoek et al., 2008, Kobashi et al, 2008; Steffensen et al., 2008; Stuiver et al., 1995, 1998). The dendrochronology data is important to specify the climate event chronology between 15,000 and 10,000 years ago (Wohlfarth, 1996; Friedrich et al., 1999; Schaub et al., 2008). Also of importance is the determining of the chemical composition and age of ice and tephra interlayers in lake sedimentation. Subsequent interregional correlation of these data allows the obtaining of time scales with a year resolution (Hoek et al., 2008). Analysis of cosmogenic isotopes ^{10}Be and ^{14}C in ice cores and lake sediments has recently been successfully used as an independent method for dating the marginal moraine of the Lateglacial (Rinterknecht et al., 2004, 2006).

Fig.2.2.1. Lateglacial and Early Holocene oxygen isotope record plotted on a calendar scale.

Figure 2.1 depicts the timescale of climatic events between 15,000 and 7,000 years ago based on the comparison of an event age by using different timescales (Hoek, Bos, 2007; Björck et al., 1996, 2001). The two warmest phases ~14,700 and ~11,500 cal years BP are more explicit.

Between these time intervals climate was most variable with cooling of first hundreds years duration alternating with warming. During a warm episode ~14,000 cal years BP air temperature in ice-free regions of the Baltic Sea Basin increased to values close to modern ones. This is seen from data on changing vegetation (Hoek, 2001) and from modeling results by the general circulation model ECHAM4 (Renssen, Isarin, 2001). Summer air temperatures in ice-free regions reached 13-15°C being evident in findings of pollen of sea-buckthorn (*H. rhamnoides*) and *Typha latifolia*. This warming was interrupted with a series of cooling episodes: the Oldest Dryas (GL-1d), the Older Dryas (GL-1b), and the Younger Dryas (GS-1). (See Figure 2.2.1.). During the latter the cooling lasted 700 to 1000 years and the expansion of arboreal vegetation that started during the Bølling (GL-e) and Allerød (GL-1c) warmings (the Bølling and the Allerød in the continental sections) was interrupted and replaced again with tundra-steppe vegetation typical of the glacial time. In the Late Dryas, the sea level was 60 to 70 meters lower than at present. An episode of drastic slowness in sea level rise corresponded to this cooling stage. The Baltic Sea basin was filled with fresh-water. Actually it was the Baltic Glacial Lake with the dried area of islands and straits around the Zealand Island and a shelf strip 50 to 80 kilometers wide along the modern coast of North-German and Pomorze lowlands. Coasts of the Gulfs of Riga and Finland as well as Lake Ladoga basin being subjected to a strong glacio-isostasy were submerged under lake waters (Björck, 1995; Mangerud et al., 2007). In the Late Dryas, the Scandinavian ice shield covered even a greater part of Fennoscandia. Cooling of that time has been correlated with the terminating phases of glacial advance resulting in the formation of marginal moraine ridges Salpausselka I and II with the dating of 12,250 cal years and 11,600 cal years ago, respectively (Subetto et al., 2002; Subetto, 2006, 2009; Rainio et al., 1995; Rinterknecht et al., 2004, 2006). During the formation of the margins the Salpausselka northern lowland part of the Karelian Isthmus was the bottom of the Baltic Glacial Lake (BGL) filling with its waters the Baltic and Ladoga basins (Subetto, 2009). At that time the periglacial vegetation type emerged on weakly developed soils formed under the conditions of severe Arctic climate. On the Karelian Isthmus, cold and dry climatic conditions and grass-bush (tundra-steppe) associations predominated until 11,000 cal years ago. On the shores of the Baltic Glacial Lake in the territories of modern Lithuania, Latvia and Estonia open tundra landscapes spread with predomination of *Betula, Betula nana, Pinus, Salix, Artemisia, Poaceae, Cyperaceae Chenopodiaceae,* and *Dryas* (Ozola et. al., 2010).

Analysis of relic cryogenic forms indicates that permafrost developed in an ice-free portion of the Baltic Sea Basin in the Late Dryas (Mangerud, 1987; Isarin et al., 1997). In addition evidence is found of activation of aeolian processes (Isarin et al., 1997; Kasse, 2002) and changes in river channel morphology (Starkel, 2002), lake level and lake sedimentation composition (Subetto, 2009). Air temperature has been estimated by applying the arealographic method based on using palaefloristic data (pollen-spore diagrams and macrofossil data) (Borisova, 1990; 1997). The results show that during the coldest phase of the Young Dryas deviations of the January temperature means from the modern ones were the largest in northwestern Europe, i.e., in the Baltic Sea Basin. The January temperature there was 10 to 13°C below the modern one. Like at present, this temperature decreased from –14 down to –20°C from west (the Jutland Peninsula) to east (coast of Gulf of Finland). Temperature deviations of the warmest month (July) from modern values were much less being -2°C on the southern coast of the Baltic Glacial Lake. For the most part of this region the July temperature was about 13-14°C in the Late Dryas.

In northern Germany, July temperatures were about 12°C, in central Germany and Poland about 13°C (Borisova, 1990). In Finland Karelia, they were 7-10°C, southern Sweden about 10°C, and western Poland about 12°C. In Poland, January temperatures were not higher than - 20°C (Walker, 1995).

The calculations based on using the transition function between modern climate and vegetation show that the mean annual temperature of 14°C during the coldest time of the Younger Dryas (about 10,500–10,700 BP) was approximately 6^0C lower than the present annual temperature (Arslanov et al., 2001). At the beginning of Holocene the mean annual temperature was about 2°C lower than that at present.

Analysis of the contents of stable isotopes in sediments of Lake Gościąż located in the mid-current of the Vistula River shows that in the Late Dryas the July temperature was 10-13°C (Starkel, 2002). An independent reconstruction based on insect fauna composition (*Chironomidae)* indicates that the July temperature in southwestern Norway in the vicinity of the southwestern border of the Scandinavian ice shield (Velle et al., 2005) was 5-6°C vs. present 11°C. In southeastern Karelia east of Onega Lake (beyond the limits of the Baltic Sea Basin) in the Late Dryas the minimum July temperature was estimated at 4°C by pollen and macrofossils *Betula nana* data vs. modern July temperature of 14°C (Wohlfarth et al., 2002, 2007).

The modern Greenland core chronology (GRIP, NGRIP, and Dye-3) places the Late Dryas/Holocene boundary at about 11,653 cal years ago. By ^{10}Be and ^{14}C isotope content in tree rings this boundary is dated at about 11,573 cal years ago, and by tree-ring chronology at ~11,590 cal years ago. The latter from (Kobashi et al., 2008) is considered to be the most accurate. Mörner (1980) was the first to quantitatively estimate air temperature changes at the DR3/Holocene boundary by oxygen-isotope analysis of lake carbonates from southern Sweden. According to this data the air temperature increased by about 9°C at the Late Dryas/Holocene boundary. In geological terms this increase was instantaneous. Later this result was fully confirmed by oxygen-isotope data from Greenland core (Dansgaard et al., 1989). Independent estimates on the $^{29}N/^{28}N$ and $^{40}Ar/^{36}Ar$ ratio in Greenland core show the temperature increase at the DR3/Holocene boundary was 10± 4°C within less than 50 years (Grachev, Severinghaus, 2005). In the Early Holocene a cold and relatively dry climate was drastically changed to a warmer and humid one. The summer temperature in northwestern Russia increased from 4° to 10-12°C (Wohlfarth et al., 2007). In the very first warm phase of the Preboreal around 11,530 – 11,500 cal BP as a result of an abrupt warming (within 50 years or less) the arboreal vegetation started spreading fast in ice-free regions, however climate was unstable (Bos et al., 2007). Tundra-steppe vegetation with predomination of shrubs and grass typical of the Late Dryas changed to vegetation of open forest associations. In the early Preboreal, birch tree started spreading, pine spread but less, though periglacial complexes still composed a large percentage in individual parts of the Baltic Sea Basin. In the early Holocene during the first warming wave, the broad-leaved vegetation appeared in ice-free regions, where *Tilia, Ulmus, Corylus, and Fraxinus* were first found. By using AMS ^{14}C dating ~11,600 cal years BP from Lake Hańcza (northeastern Poland) sediments it was found that in the Early Holocene there was shift from clastic-detrital deposition to an autochthonous sedimentation dominated by biochemical calcite precipitation **(**Lauterbach et al. 2011).

In the territory of the present Lithuania the gradual amelioration of the environmental situation started at about 11,500 cal BP while it was less pronounced in the territory of Lithuania (Stančikaitė et al., 2004, 2008) as well as in the northwestern Russia (Wolfhart et al., 2007) compared with that recorded in the North Atlantic region (Björck et al., 1996). Talking about the environmental situation of that time early expansion of *Picea*, even before 11,500 cal BP, in the local vegetation should be stressed. Despite the long-lasting discussion based on pollen data, macrofossil finds and modern genetic information, Lateglacial and early Holocene history of *Picea* in this part of Europe is still under debates (Giesecke and Bennett, 2004; Latałowa, van der Knaap, 2006). *Picea* pollen and macrofossil data suggests the local presence of this tree during the earliest stages of the Holocene in the northeaster and northern Lithuania (Stančikaitė et al., 2009, Gaidamavičius et al., 2011, Kabailiene et al., 2009; Kisielienė, pers. information). Moreover the presence of *Picea abies* suggests a warm continental climate with warm summers (~+10-13ºC) and moist soil conditions (Giesecke and Bennett, 2004) predominated shortly before 11,500 cal BP in area. However, vegetation composition and aquatic regime indicate the presence of at least two short-lasting cold climate episodes between 11,500 and 11,100 cal BP in the northeastern Lithuania (Stančikaitė et al., 2009).

The Holocene boundary warming at about 11,530 and 11,500 cal BP was interrupted by a short cold the Preboreal oscillation dated from ice cores at 11,430-11,270 cal years (Kobashi et al., 2008). A relatively short cooling (about 200-250 years) occurred approximately 250 years after the Holocene boundary warming (Rasmussen et al., 2006). During this cooling methane (one of the greenhouse gases) decreased by 8% (60 ppbv) as compared to its increase during the preceding warming. The coldest part of the Preboreal oscillation is dated at ~11,430-11,350 cal BP. The birch expansion that started during the warming in the beginning of the Early Holocene was interrupted by a dry continental phase with open grassland vegetation (Rammelbeek Phase) (Bos et al., 2007). In the territory of the present Lithuania The most prominent climate cooling recorded shortly before 11,100 cal BP may have been correlated with the climate event, termed the Preboreal Oscillation (PBO, 11,300–11,150 cal BP), described as a humid and cool interval in northwestern and central Europe.Only after 11,100 cal BP ongoing forestation of the territory by open birch-pine predominant forest suggests climatic improvement that could be interpreted as a delayed reaction to the Pleistocene/Holocene warming

At the start of the Preboreal cooling the sea level remained significantly lower than at present (by approximately 50 m). Then a rapid step-wise rise took place due to influx of great volumes of fresh water into the North Atlantic from melted continental ice shields (Laurentian and Scandinavian). The Preboreal cooling was explicitly prominent in all natural objects (sea and ice cores, dendroclimatic records, and lake sedimentation). Spore-pollen chronology indicates a well-expressed climate instability in the Early Holocene. During its coldest phase forest-tundra and open forest landscapes were established over a greater part of the Baltic Sea Basin which remained in some regions until 10,700-10,600 cal years ago. Pazdur (2004) characterizes the Preboreal climatic conditions as cold and dry in the territory of Poland.

At the start of the Late Preboreal time, between 11,270 and 11,210 cal BP, a sudden shift to warmer and more humid climate occurred and forest vegetation expanded again. Expansion of pine occurred in the later part of the Late Preboreal. By the GISP2 core isotope data in the end of the Preboreal time at the Boreal time boundary, air temperatures increased by 4 ±1,5°C (Grachev, Severinghaus, 2005). At the onset of the Boreal time (between 10,770 and 10,700 cal BP), dense woodland with pine started to develop in the southern part of the Baltic Sea Basin (Bos et al., 2007), on the Karelian Isthmus and in northwestern Russia boreal forest with *Pinus, Picea, Betula, Alnus incana* was present at the lower altitudes (Subetto et al., 2002).

At that time in Southern Finland open landscapes typical of the Preboreal cooling changed to poplar-pine-birch closed vegetation (Alley, 2000). In the south of modern boreal belt in Finland beginning with 11,000 cal years, glacial boundary retreated fast and at about 10,700 cal years the region was free of ice, though the glaciation boundary was still very close. According to estimates in (Hiekkilä, Seppä, 2003) between 10,700 and 10,500 cal years the annual temperature means were as much as –3.0 – 0.0°C. These temperatures correspond to the modern temperature at 70°N. Temperature started rising after 10,000 cal years BP and lasted until the beginning of cooling at 8.2 years BP. Over the greater part of southern Sweden, in Estonia, Latvia and Lithuania open boreal woodlands and sparse birch vegetation were established. As compared to the DR3 cooling summer air temperatures rose by 7-10°C. However major environmental changes in area started after 10,300-10,200 cal BP when deciduous trees including thermophylous species established in the region.

In northeastern Poland according to Lake Hańcza (11,600 cal BP) data in the beginning of the Holocene shrub pollen decreased and a shift occurred from clastic-detrital deposition to an autochthonous sedimentation dominated by biochemical calcite precipitation (Lauterbach et al., 2011). A stable improvement of climatic conditions continued and between 10,000 and 9,000 cal BP pollen spectra show increased pollen of broad-leaved forms and lake sediments depict an increased organic content (Lauterbach et al., 2011).

Between 11,000 и 9,500 cal years temperatures rose but slowly reaching the values of ~0.5°C lower than the modern ones. Prominent warming started about 9,000 cal years ago and lasted until the thermal optimum (HTM) about 8,000-4,500 cal years with temperatures of 2.5-3.5°C above the modern ones. Davis et al. (2003) based on analysis of spore-pollen data from more than 500 sections in the territory of the entire West Europe obtained quantitative characteristics of climate for the past 12,000 years. In northern Scandinavia summer air temperatures close to the modern ones have been obtained based on both pollen data (Seppä, Birks, 2001) and independent estimates based on chironomids and diatoms (Rosén et al.,2001). The temperature estimates based on macrofossils (tree-lines position) indicate the rapid and stable warming after 10,000 cal BP in entire northern Scandinavia (Seppä, Birks. 2002). Climate reconstruction for central Sweden by using spore-pollen spectra from Lake Gilltjärnen sedimentation shows the summer temperatures to fast increase from 10.0 до 12.0°C between 10,700- 9,000 cal BP. This stable positive trend remained until cooling at 8,600 cal BP (Antonsson et al., 2006).

Figure 2.2.2. depicts analysis of data on air temperature changes during the past 12,000 years in different regions of Northern Europe.

2.2.2.2. The "8.2 ka cool" event The 8.2 ka cooling had been known comparatively long time by changes in oxygen-isotope composition in ice cores from the Summit well. The cooling has been estimated by changes in air temperature by 6 ± 2°C in central Greenland (Dansgaard, 1993; Alley et al., 1997; Alley, Áqústsdóttir, 2005; Muscheler et al., 2004; Rohling et al., 2005). Later an independent method was developed to estimate both air temperature change amplitude and the duration of this cooling. This method is based on using $\delta^{15}N$ in N_2 as a calibration tool (Leuenberger et al., 1999; Kobashi et al., 2003, 2007; Thomas et al., 2007). Kobashi et al. (2007) used the GISP2 ice core with the time resolution of nitrogen isotope of ~10 year corresponding to a gas age near 7,600-8,600 BP. This data showed a complicated time structure of this event. The coldest phase of about 70 years in duration occurred in the middle of this interval. In the beginning and end of this event the climate was milder. Air temperature decrease of 3 ± 1.1°C occurred within less than 20 years and the entire cooling episode lasted approximately 150 years. Independent estimates of changes in air temperature during the 8.2 ka event have been obtained by using isotope data from the four Greenland cores: NGRIP, GRIP, GISP2, and Dye3. The first three are from central Greenland, and Dye3 is from southern Greenland (Thomas et al., 2007). By the GRIP core data the event has been dated at c. 8,190 cal BP. The GISP2 cores data revealed two comparatively warm peaks: c. 8,220 and c. 8,160 cal BP. The initial stage of cooling is more explicit in the NGRIP core data. Duration of the entire cooling event is about 160.5 ± 5.5 years, the coldest phase being 69 ± 2 years. Independent empirical data such as sea cores with high resolution, lake sediments, clay varves, speleotem isotope data, spore-pollen diagrams and others indicate that the cooling took place in the entire the Baltic Sea Basin, except for the most northern regions above 67°N. The mountain glaciation increased in all mountainous regions. In central and southern Norway, the snow line (equilibrium line altitude) lowered by 200 m (Nestje, Dahl, 2001; Nestje et al., 2001; Nestje, 2009). The altitude of the upper boundary of arboreal vegetation also lowered (Seppä et al., 2001, 2007, 2008). In other mountainous areas of the Baltic Sea Basin ice movement was also recorded (Denton, Karlen, 1973). This cooling was also observed by changes in lake sediment accumulation as a hiastus data (Veski et al., 2004).

Spore-pollen diagrams from lake and continental sediments contain independent evidence of this cooling. Davis et al., (2003) based on analysis of about 500 pollen sites from West Europe over the past 12,000 years indicated with confidence the 8.2 ka cooling episode over the entire territory with the lowest air temperatures in areas adjoining the North Atlantic (Seppä et al., 2008; Rasmussen et al., 2009). Pollen data from Estonia and southern Fennoscandia indicate a 1°C air temperature drop. In the northern most regions there was either no cooling or a slight warming. In Estonia near the Rõuge lake, air temperature declined by 1,8°C (Veski et al., 2004), which agrees with estimates of cooling in central Europe (Wiersma, Renssen, 2006). Lake sediments from Estonia south of 61°N (Seppä, Birks, 2007) indicate that pollen of warm-loving forms *Corylus* and *Ulmus* decreased from 10-15% in the Early Holocene to 5% between 8,250-8,050 cal years BP.

In south-eastern Latvia in the Early Holocene, relatively warm and stable climate was interrupted by cooling at about c. 8,350-8,150 cal BP when mean annual air temperatures dropped by 0.9-1.8°C. This was reflected in vegetation species change, decrease in broad-leaved vegetation productivity, and pollen increase of arboreal species (birch and fir-tree) (Heikkilä, Seppä, 2010). Re-establishment of *Picea* in the eastern Baltic region (in the territory of present Lithuania) dated back to 8,600–8,000 cal yr BP. This short climatic deterioration may have limited the expansion of deciduous trees, providing space for the spruce. The expansion of spruce may have been a response to a wet and cool climate with colder and snowier winters (Seppä and Poska, 2004).

Studies on sediments from Lake Kalksjön in western part of central Sweden by means of varve chronology, radiocarbon dating and by using dendrochronological dates provided a detailed climate chronology and the dating of "the 8.2 ka event" between 8,066 ± 25 и 7,920 ± 25 cal BP. This interval contains about 400 varve layers. During this cooling the speed of snow accumulation grew. The duration of the event is estimated at 150 years (Snowball et al., 2002, 2010; Zillen, Snowball, 2009). In Estonia (Seppä, Poska, 2004; Veski et ai., 2004), this event lasted longer. This indicates a longer period of vegetation re-establishment after a decrease in annual air temperature by at least 1,5- 2,0°C.

In Germany, by changes in isotope composition of ostracod valve carbonate the duration of the cooling is estimated at 200 years (von Grafenstein et al., 1998). Almost all data indicate that the decrease in air temperature in winter was significantly larger as compared with the warm season. This promoted an earlier formation and later thawing of ice on sea and lakes .

Causes of the cooling during "the 8.2 ka" event and other cooling episodes in the Lateglacial/Early Holocene were of interest for researchers a long time (Alley et al., 1997; Alley, Áqústsdóttir, 2005; Barber et al., 1999; Broecker et al., 1988, 2010; Clark, 2001, 2002; Clarke et al., 2004; Denton et al., 2005; Leuenberger et al., 1999; Kobashi et al., 2007; Thomas et al., 2007). Although individual scientists ascribe "the 8.2 ka" cooling to changes in the incoming solar radiation (Bos et al., 2007), most of them relate this cooling and other cooling events of the past 13,000 years to changes in the circulation of surface and deep water in the North Atlantic as the continental ice sheet melted. Clark (2001, 2002) guessed that freshening of the sea surface layer not only disturbed circulation in the surface layer it also impeded the formation of deep water thus affecting the intensity and position of the Atlantic "conveyer" itself.

Drainage of glacial lakes the Agassiz and Ojibway as a result of the Laurentide ice sheet melting c. 8,470 cal BP (~ 7,700 ^{14}C yr BP) during which about $2 \cdot 10^{14}$ m^3 fresh lake water could have been released within less than 100 years could exert a serious impact on the formation of sea ice thus significantly changing the term of their formation in autumn and melting in spring. Sea ice has a higher albedo, therefore, due to a feedback mechanism a longer period of sea ice presence contributes to an additional cooling increasing the existing one (Barber et al., 1999; Clark, 2001; Clarke et al., 2004; Fisher et al., 2002).

2.2.2.3. The Atlantic warming. After the last most significant Holocene cooling about 8,2 ka warm period occurred when air temperature and annual precipitation values were above the present ones. At that time forest vegetation with warm-loving species flourished. In northern Finland, the warmest climate of the Holocene occurred between 8,200-5,700 cal BP with temperature maxima at c.7,950-6,750 cal BP. Based on spore-pollen diagrams of lake sediments by means of the "transition function" July temperatures were estimated at 13°C. They dropped by 1°C at about 5,750 cal BP only. This temperature drop is evident from increased *Pinus sylvestris* in pollen spectra and decreased lycopod and fern spore values (Seppä, Birks, 2001; Seppä et al., 2007,2008). In southern Finland, the earlier part of the Atlantic period is characterized by maximum values of nemoral deciduous tree taxa, with *Corylus, Ulmus*, and *Tilia* reaching their highest percentages. Maximum summer temperatures were between 8,000 and 5,800 cal yr BP. At the same time by the spore-pollen spectra data with those high temperatures *Picea* started to increase at ca 6000 cal yr BP. The highest temperatures occurred at 8,000– 5,800 cal yr BP. (Heikkilä, Seppä, 2003; Seppä et al., 2004).

In northwestern and northern Russia, forest vegetation with fir-tree and deciduous plants predominated (*Carpinus, Fagus, Ulmus, Tilia, Quercus)*. Between 8,000 - 4,500 cal yr BP summer temperatures were 2.0 -2.5°C and precipitation was 100 -150 mm/year above the present-day values (Arslanov et al., 1999, 2001). In northern European part of Russia (the Kola Peninsula), the warm period occurred between c. 8,000 and 4,200 cal BP with maximum summer temperatures at c. 6,000 and 5,000 cal when July temperature was 13.6°C (Solovieva et al., 2005).

By using transfer function and based on a combined Fennoscandian pollen surface sample data-set and spore-pollen data from three lakes (two are in central and western Sweden and one is in southern Finland) Giesecke et al. (2008) have reconstructed winter (January) and summer air temperatures (July) for the past 10,000 years. Summer temperatures were reconstructed by using varve thickness data. The winter temperatures maxima were found between 6,000 and 5,500 years ago, and the largest summer warming between 7,000 and 6,000 years ago.

Reconstructions of mean annual air temperatures by pollen data from Lake Trehörningen in southwestern Sweden showed that the temperature maxima dated at 7,000 to 4,000 cal yr BP were 2.5-3.0°C above the modern ones. In northern and eastern Scandinavia, mid-Holocene temperatures were somewhat lower, however they were 1.5-2.5°C above the present-day ones (Antonsson, Seppä , 2007).

In central Sweden at about 7,000 cal BP, warm-loving vegetation (oak and linden-tree) expanded. This vegetation needs high summer temperatures and is relatively resistant to low winter temperatures. During the warmest time of the Holocene (between 7,000 and 4,000 the linden-tree boundary advanced 300 km north of its modern position. At that period precipitation dropped and lake levels were low

. Maximum annual temperature means of 6.0-7.0°C took place at about 6,500 cal years as compared to 5.0°C at present near Lake Gilltjärnen. (Antonsson et al., 2006). Reconstructions of summer air temperatures in central Sweden by chironomid subfossil showed that these temperatures changed within wide limits from -0.8 °C to +0.8 °C vs. its modern values (Brooks, Birks, 2000, 2001; Velle et al., 2005). In present Estonia, *Ulmus* and *Tilia* spread maximally between 7,000 and 5,000, which coincides with maximum summer air temperature period. At that time the area with these species was about 40% vs. 1% at present (Saarse, Veski, 2001).

Reconstructions made by using sediment data from Lakes Holzmaar and Meerfelder in the Westeifel Volcanic Field (Germany) showed that the temperatures were 1°C above the modern ones. Temperatures started lowering after 5,000 cal BP (Litt et al., 2009).

In modern Latvia, deciduous forest first appeared later than in Estonia. *Ulmus* started spreading in the first half of the Boreal period, *Tilia* at about 7,000-5,000 cal. yr. BP (Heikkilä, Seppä, 2010). With the beginning of the Atlantic warming deciduous forest with *lmus, Tilia, Alnus, Corylus*, and *Picea* spread (Murniece et.al., 1999; Ozola et. al., 2010). In Latvia, the Holocene Thermal Maximum (HTM) was recorded between 8,000 and 4,000 cal yr BP. During HTM summer temperatures were ~2.5–3.5 °C higher than the modern values (Heikkilä, Seppä, 2010), and subsequently declined towards present-day values. Light-demanding *Quercus* and *Fraxinus* reached their maxima later, c. 5500–3000 cal yr. BP, in concert with the expansion of *Picea*, when climatic cooling induced forest restructuring.

However, in Lithuania, the major environmental changes started after 10,300-10,200 cal BP when deciduous trees, including thermophylous species, were established in the region. In eastern Lithuania, the expanding deciduous taxa, e.g., *Corylus* (*ca.*10,200–10,000 cal yr BP), *Alnus* (8,200–8,000 cal yr BP), and broad-leaved species with *Ulmus* (*ca*. 10,000 cal yr BP), *Tilia* (7,700–7,400 cal yr BP), and *Quercus* (5,200 cal yr BP), formed a dense mixed forest where *Picea* re-appeared at 7,300–6,800 cal yr BP (Gaidamavičius et al., 2011). At the same time in northwestern Lithuania, migration of *Corylus* was dated back to 7,600–7,200 cal BC, *Alnus* – back to 7,300–6,900 cal BC, and *Ulmus* – back to 8,100–7,500 cal BC, respectively

(Stančikaitė et al., 2006). Since about 9,800-9,900 cal BP expansion of deciduous species was accompanied by development of *Cladium mariscus* (L.) Pohl. suggesting minimum mean July temperature increased up to +15.5ºC even in the northern part of the country. Development of deciduous forest requires moist and fertile soil and high humidity. Therefore, its expansion confirmed that such ecological conditions pre-existed there. The Holocene Thermal Maximum, *ca*. 8,000 – 4,500 cal yr BP, with dryer and warmer summers indicating an increased continentality (Seppä, Poska, 2004) was generally responsible for expansion of deciduous trees to the eastern Baltic region. This taxa continued to flourish between about 7,500 to 4,200 cal yr BP (Giesecke, Bennett , 2004; Giesecke et al., 2008; Gaidamavičius et al., 2011; Stančikaitė et al., 2003, 2004, 2006, 2008,2009).

In central Poland, *Tilia* and *Quercus* appeared about 9,300 years ago, however they were not widely spread until 8,000-7,000 years ago. At the same time *Alnus* spread widely, and in mountainous regions – *Picea*. Since 6,500 years ago the first deforestation began because of the start of land cultivation. Then at about 5,000 years ago, when the Sub-Boreal period started, climate became wetter and colder (Ralska-Jasiewiczowa, Starkel,1994; Pazdur, 2004; Lauterbach et al., 2011).

Davis et al. (2003) show that in northern Europe, the explicit Holocene maximum is dated c. 6,000 cal BP. The temperature maxima of 1.5- 2.5°C above the present-day ones were in the northwest, and of about 1.0°C northeast of the region with anomalies being of 1.5-1.0°C. After reaching its maximum value the temperature started decreasing during the rest of the Holocene. In both regions, temperatures rose until 6,500 BP. Seppä and Birks (2001) have accomplished the reconstructions of temperature based on pollen data. Rosén et al., (2001) used different proxy data, including pollen, chironomids, diatomic, and others.

Figure 2.2.3. presents air temperature changes in different parts of the Baltic Basin obtained from different proxy data.

2.2.2.4. The Late Holocene cooling. As seen from Fig.2.2.3., climate was very unstable in the entire Baltic basin during the past 4,500 years. At the background of global cooling trend, there were relatively short warming alternating with short cooling.

Different proxy data allow us to reconstruct a two-stage nature of air temperature decrease in the Late Holocene. The first stage occurred between 5,000 and 4,500 cal yr BP and the second one between 4,300 and 3,300 (2800) cal yr BP. During each period the temperature drop was at least 1^0C. A warming c. 3,200 and cooling c.2,800 cal BP are revealed on detailed palaeoclimatic reconstructions. Although the general trend of the Late Holocene cooling is undoubtedly related to decreased summer solar radiation due to astronomical factors, the causes of these oscillations need further studies.

In northwestern and northern Russia, climate cooling started at about 4,500 coinciding with other regions of the Baltic Basin. A warming at about 3,500 cal BP and cooling at about 2,500 cal BP are revealed on the general cooling trend (Arslanov et al., 2001; Subetto et al., 2006).

. In southern Sweden, based on analysis of Lake Igelsjön sedimentation, an abrupt hydrological shift to cooler and/ or wetter conditions has been revealed at around 4,000 cal yr BP. (Jessen et al., 2005). A cooling trend remained to the end of the Holocene with a series of short rapid warm fluctuations. Pollen diagrams show changes in vegetation composition and a distinct decrease in warm-loving forms (primarily, *Corylus*). Relatively warm conditions remained until 4,700 cal years BP. About 2,000 cal years ago the temperatures were close to modern values (Jessen et al., 2005).

In Poland during the Late Holocene, lake levels elevated and peat bogs developed.. Between 5,000 and 4,200 cal years BP climate was relatively warm and dry. However, by the end of the period it changed for a more humid one. Dominants in the composition of the forest associations slightly changed. In mountains, fir-trees spread and *Carpinus* and *Fagus* spread from the southeast. About 3,500 years ago *Abies* spread from the southwest and became a forest-forming species in southern Poland (Kulesza et al., 2011). *Picea* spread to the northeast and *Fagus* spread from the west. A. Pazdur (2004) showed that the Sub-Atlantic period started about 2,800 years BP. This cooling is revealed in spore-pollen spectra, including other parts of the Baltic Sea Basin.

In Latvia in the Late Holocene, the summer temperatures dropped by ~3 °C (Heikkilä, Seppä, 2010). Fir-alder trees have dominated during the past 3,000 years. *Pinus* and *Betula* were the dominating tree species in the area, distribution of *Picea* considerably decreased.

Thus analysis of different proxy data for the past 12,000 cal years has revealed the following three stages of natural climate oscillations in the Baltic Sea region:

1. The 200-year long short-term cooling episodes related to deglaciation occurred between 11,000 and 8,000 years ago at the background of a stable positive temperature trend stipulated by a higher total summer solar radiation due to astronomical factors.

2. The period of warm and relatively stable climate took place between 8,000 and 4,500 cal years ago with air temperatures of 1.5-2.5°C above modern.

3. The Late Holocene (past 5,000-4,500 years) is characterized by the negative temperature trend and increased climate instability.

2.2.3. Causes of Climate Change in the Lateglacial - Holocene Time

2.2.3.1. Solar radiation forcing during the Holocene (astronomical, solar activity)

2.2.3.2. Volcanic eruptions forcing

2.2.3.3. Greenhouse gas forcing

2.2.3.4. Modes of variability

Three climate-forming factors (solar radiation, atmosphere gas composition, and Earth's surface-atmosphere albedo) determine the multiversity of past and present climates. Solar radiation at the upper atmosphere boundary is subjected to both long-term (of thousands to millions years) and short-term (multi-centennial) variations due to astronomical factors and changes in luminosity of the Sun as a star. Solar radiation is absorbed when passing through the atmosphere and partially dissipated back to space under the influence of microscopic aerosol composing a permanent layer in the stratosphere with the concentration changing by more than an order of magnitude depending on powerful and catastrophic explosive-type volcanic eruptions. Injections of volcanogenic materials into the stratosphere decrease the incoming solar radiation resulting in an air temperature drop at the Earth's surface. According to various climate models volcanic eruptions contribute as much as 30% of air temperature variability.

Studies on atmosphere gas composition are based on proxy calculations and direct measurements in Greenland and Antarctic ice cores. According to these studies concentrations of major greenhouse gases (carbon dioxide, methane and nitrogen dioxide) influencing air temperature varied naturally in the past. At present atmospheric greenhouse gases concentrations have noticeably increased because of burning oil, natural gas, and coal causing surface air temperature rise (IPCC, 2007).

Changes in albedo of underlying surface and Earth-atmosphere system exerted a considerable effect on radiation and thermal conditions in the past when continental and sea-ice area and the land-sea area ratio dramatically changed at different latitudinal zones. At present the air surface temperature effect of Earth-atmosphere albedo depends on cloud amount variations, in particular in the upper cloud level. Volcanic ash coming to the stratosphere during powerful explosive-type volcanic eruptions exerts a certain effect on the system albedo.

The global climate effect of earth's structure (including mountain formation) has been widely discussed in palaeoclimatic literature (Borzenkova, 1992; Zubakov and Borzenkova, 1990). According to recent model estimations the air temperature effect of this factor is significantly less than that from changing carbon dioxide concentration (Borzenkova, 2002).

There are other factors such as the changing speed of Earth's rotation, shifting geographical and magnetic poles, continent disposition in high latitudes, for example. Studies have recently been made on relationships between atmosphere processes and Earth's rotation instability that can affect circulation and thus the thermal regime and hydrological cycle.

2.2.3.1. Solar radiation forcing during the Holocene (astronomical, solar activity). The mean annual heat income to the Earth's surface depends on the solar radiation amount at the upper atmosphere boundary with an average Earth's distance from the Sun. This value, the so-called astronomical solar "constant ", undergoes long-term and short-term oscillations related both to physical processes on the Sun and astronomical factors. Sun's luminosity is believed to increase very slowly over time, at least 5% over a billion years. Short-term variations in the astronomical solar "constant" depend on these in Sun's luminosity resulting from solar spot and flare formation. An 11-year cycle of these variations has been most studied though scientists have recently discovered a 200-year periodicity. Scientists have been trying to find a relationship between the number of solar spots or solar flares and Earth's surface temperature for a long time. Satellite observations of Sun's disk brightness made it possible to start solar radiation monitoring at the upper atmosphere boundary (Foukal et al., 2004, 2006; Lean, 2010; Fröhlich, Lean**,** 2004; Solanki et al., 2004; Scafetta, 2010). Recent measurements of astronomical solar "constant" accomplished within the framework of the SORCE Project (Solar Radiation and Climate Experiment) in 2003/2004 have resulted in 1361 Wt/m^2, which is somewhat lower than that of 1367±7 Wt/m^2 by WMO. As satellite data analysis (IPCC, 2007) shows, there is not yet an accepted alternative for the WMO solar "constant" estimate. Satellite measurements show that the present comparatively small changes in solar radiation income between the minimum and maximum solar activity periods related to the 11-year cycle of solar activity can be attributed more to measurement errors than to actual changes in Sun's luminosity (IPCC, 2007).

Van Loon et al. (2004) indicate insignificant changes in incoming solar radiation of not more than 0.1% between the solar activity peaks corresponding to an average global forcing of ~ 0.25 Watt/m^2. These solar radiation variations can lead to an air surface temperature change of not more than 0,1°C. These variations are smoothed by thermal ocean inertia not exerting significant effect on climate change. This conclusion has recently been confirmed in (Foukal et al., 2004, 2006; Lean, 2010). The biggest changes in the "astronomical solar constant" and air surface temperature can be attributed to periodic variations in Earth's orbit elements. Considerable changes of the astronomical solar "constant" are related to orbital factors (changes in obliquity, precession, and eccentricity). Orbital forcing involves a significant redistribution of solar energy, both seasonally and latitudinally (Berger, 1978; Bradley, 2003; Crowley, 2002; Hünicke et al., 2010; Wanner et al., 2008; Weber et al., 2004).

The orbit eccentricity characterizing its deviation from the circle changes from 0.0007 to 0.0658 (at present it equals 0.017). Eccentricity oscillates with periods of 100, 415-420 and 1200 thousand years, the Earth's obliquity changes from 22°06,8´ to 24°56,8´(the present value is 23°45), the obliquity change periods are 23, 41 and 200 thousand years. The periodic changes in these factors cause relatively small changes in the total seasonal solar radiation at different latitudes resulting in air surface temperature variations. In the Holocene, precession was a major astronomical factor in changing solar radiation. Due to changes in precession solar radiation reached its maximum at the upper boundary of the atmosphere in the Northern Hemisphere about 12,000 cal years ago, then it gradually decreased. Maximum changes in solar radiation income (positive anomaly of up to 40 Watt/m^2) took place in high latitudes, whereas at the equator the positive anomaly was not above 25 Watt/m^2. In some studies (Gauthier, 1999), global climate oscillations have been detected to correspond to different fractions (1/2, ¼, 1/8, and 1/16) of the 23-thousand-year precession cycle of Earth's orbit. Global climate oscillations of this scale are found to be non-periodic; therefore the above cycles show only the tendency to correspond to them on the average. Data on such proxy climate indicators as tree rings and anomalies of oxygen isotopic composition (δ^{18}O) in Greenland ice cores and Davis Hole calcite (Nevada, USA) illustrate the presence of a united structure of climate variability. Variations in both orbit parameters and solar activity determine the nature of the variability in question. Direct (in terms of the observed sunspot number) and indirect (in the form of cosmogenic isotopes such as ^{10}Be and ^{14}C) proxies can provide longer term histories of solar activity. Conversely, higher radionuclide production rates can in principle be associated with lower solar activity (Lockwood, 2006).

Figure 2.2.4. depicts the dependence of changes in air temperature in high latitudes of the Northern Hemisphere, reconstructed by oxygen-isotopic ratio in the Summit core from Greenland (GISP2), on solar radiation at the upper boundary of the atmosphere ("astronomical solar constant") and on concentration of greenhouse gases CO_2 and CH_4 (Alley et al., 2007; Flückiger et al., 2002; Hünicke et al., 2010; Kobashi et al., 2003, 2007, 2008; Thomas et al., 2007). In the Lateglacial-Holocene solar radiation at the upper boundary of the atmosphere in the Baltic region ("astronomical solar constant") between 16,000 and 8,000 cal BP differed significantly from the modern one. Maximum radiation took place about 12,000-11,000 cal yr BP when at 50°N summer radiation sums were 34 Watt/m2 above the modern ones. At the same time winter radiation decreased by 11Watt/m2. However, winter radiation affects air temperature of the Earth's surface but much less than the summer one due to a shorter day duration and larger cloudiness. About 6,000-5,000 cal BP summer radiation in the Baltic region was 5% above the modern one. It decreased fast to the values close to modern ones 2000 years ago.

A rapid growth of carbon dioxide (CO_2) and methane (CH_4), major greenhouse gases, provided an additional rise of air temperature in the Lateglacial-Early Holocene. As seen from Fig. 2.2.4, relatively small changes in air temperature between 8,000 and 4,000 years ago on shorter timescales (decadal to inter-annual) are most likely related to changes in solar radiation due to changing “meteorological solar constant” as a result of emission of products of volcanic eruptions to the upper layers of the atmosphere (Robock, 2000; Hünicke et al., 2010).

2.2.3.2. Volcanic eruptions forcing So, at the upper boundary of the atmosphere solar radiation experiences long-term oscillations (from tens and hundreds years to thousands years) causing long-term Earth’s climate variations of this scale. Then solar radiation passing through the stratosphere is partially absorbed and dissipated back into space due to microscopic aerosol particles layer permanently available in the stratosphere, the so-called Junge Layer or the Stratospheric Aerosol Layer (SAL).

Aerosol particles in this layer consist mostly of sulfuric acid and water whose concentrations vary depending on both natural and anthropogenic aerosols injected into the upper atmosphere. Maximum changes (by an order of magnitude or more) in aerosol concentration occur with powerful and catastrophic eruptions of explosive-type volcanoes delivering volcanic ash and gas directly to upper troposphere and stratosphere. Such aerosol injections into the atmosphere decrease the incoming solar radiation and the surface air temperature drops. The upper troposphere boundary radiation determining climate change at the Earth’s surface is called “the meteorological solar constant” (Budyko, 1986, 1988).

Changes in stratospheric aerosol concentrations cause significant short-period variations in “the meteorological solar constant” value even with relatively stable “astronomical solar constant” during thousands and tens of thousands years. Stratospheric aerosol concentration depends largely on natural aerosol injected into the upper layers of the stratosphere as volcanic gases and products of sea plankton vital functions. A portion of stratospheric aerosol is of anthropogenic origin produced from burning fossil fuel. There is an obvious relationship between stratospheric aerosol concentration and radiation change at the upper boundary of the troposphere and on Earth’s surface.

Aerosol optical thickness **(τ)** is one of the most important characteristics of atmosphere transparency and aerosol layer status in the troposphere and stratosphere. With no large volcanic eruptions the **(τ)** mean changes within 0.01-0.02 for the visible portion of spectrum depending on aerosol concentrations in the upper troposphere and in the stratosphere. The maximum aerosol optical thickness **(τ)** of 0.4 and more occurs in the troposphere and stratosphere after powerful explosive volcanic eruptions as Tambora, Krakatoa, and Pinatubo.

Zielinski (2000) has reconstructed the value of aerosol optical thickness for the most powerful and catastrophic eruptions over the past 2100 years based on the relationship between total sulfuric acid aerosol mass and optical thickness obtained by Robock (2000). As initial information, Zielinski (1995) used detailed acidity data on Greenland ice from the Summit well (GISP2 Project) to estimate the aerosol mass. After the most powerful volcanic eruptions the value of optical thickness has been found to vary within 0.2 to 0.4. For individual events considered as catastrophic the value (τ) appeared to be 0.6-0.8 increasing in some cases up to 1.5 as, for instance, during the event of 1259.

After strong and powerful volcanic eruptions, an increased concentration of the (SAL) decreases the absorption of solar radiation at the Earth's surface by more than 10-15%. At the same time scattered radiation increases. Climate impact of explosive type volcano eruptions is determined by the height of volcanic gas column, sulfur content, latitudinal disposition of the volcano, and climatic conditions during the eruption. Equatorial and tropical volcanoes affect global temperature the most because their discharge contains the greatest amounts of sulfurous compounds. The surface air temperature can change during 2 to 4 years after individual powerful volcanic eruptions, however no considerable change occurs in the global temperature trend sign.

C. Hammer and his colleagues (1980) proposed a new method for estimating the contents of volcanic sulfur-acid aerosol in the stratosphere by using ice cores data from the Greenland and Antarctic ice sheets. The greater part of volcanic gases injected into the stratosphere is assumed to consist of sulfates being oxidized in the atmosphere before falling down with precipitation. Information about past volcanism shows that in a few years after eruptions ice layers contain higher concentrations of dissolved admixtures. The indicator of their availability could be a higher electrolytic conductivity or excess sulfate in the ice layers.

Fig. 2.2.3.5. Holocene record of volcanic sulfate (anomalies from background variations) recorded in ice core Summit, Greenland (GISP2) (Zielinski et al., 1994).

Figure 2.2.3.5 shows changes in concentration of sulfurous aerosol in the stratosphere for the past 12,000 years. In this Figure, most significant peaks correspond to the powerful and catastrophic eruptions that led to changes in aerosol concentration of stratospheric aerosol layer increases by 1 or 2 orders of magnitude, as compared with the background one. At that time the most the powerful eruptions, according to acid signal in the cores, have been dated at 7,910, 7,810, 7,700, 7,640 (the most powerful), 7,500, 7,240, 7,090, 6,230, 5,470, 3,800, 2,690,2,800, and 1,100 (Hekla), and 4,400 (Mazama) (Robock, 2000; Schmincke, 2004).

A large number of micro layers of volcanic ash of the Lateglacial ages has been found in lake and bog sediments in Scandinavia. Studies on tephra chronologies in the areas at large distances from volcanic areas began in Scandinavia in the 1960s. Perrson was a first who study number of volcanic ash interlayers in bog sediments on the Faeroes Islands, Norway, and Sweden (Perrson, 1971). Tephra dating was carried out by radiocarbon analysis of peat from upper and lower layers and their identification by the refraction index of volcanic glass. Tephra interlayers described by Perrson include eruptions of Askja (1875), Oraefajokull (1362), Hekla

(1104), "Layer G" (850-900 AD), Hekla - 3 (~2800 years ago), Hekla - 4 (~3800 years ago). The Katla volcanic system (an active volcano in Southern Iceland) is the most active in that period. There were more than 150 Katla eruptions in the Holocene (both of explosive and effusion type, Vedde ash) producing volcanic rocks of basalt composition (Larssen et al., 2001). Visible Vedde tephra interlayers (age 10,300 ^{14}C years ago or 12,000 cal years ago) have been discovered in Norway, Scotland, in sea cores from the North Atlantic and the North Sea, Southern Sweden, North-West Russia (the Karelia Isthmus), and in the Greenland core GRIP (Greenland ice cores).

2.2.3.3. Greenhouse gas forcing Carbon dioxide (CO_2), methane (CH_4) and other "greenhouse" gases (or trace gases) are transparent for short-wave solar radiation (visible light) and prevent the long-wave energy (infrared radiation) outgoing from the Earth's surface from escaping back into space. As this effect is analogous to that of a greenhouse glass roofs or walls, it has been named as the "greenhouse" effect. In the late 1970s, the first empirical data on CO_2 variations in the Late Pleistocene was obtained in studies on the chemical composition of air bubbles in Greenland and Antarctic ancient ice. Atmospheric CO_2 was shown to be about 180-190 ppmv (parts per million by volume) during the Last Glacial maximum, and to increase up to 275-290 ppmv during the Last Interglacial (the Holocene).

A stomata-based method of palaeo-CO_2 estimation has recently been applied to a temporally detailed sequences of macrofossils (leaves) from high-latitude lake sediments in Northern Scandinavia (Beerling et al., 1995; Rundgren, Beerling, 1999). This method is based on the relationship between leave stomata density (for example, *Salix herbacea* L.) and atmospheric CO_2 concentration. High concentration of CO_2 values during the Allerød warming decreased rapidly in 150-200 years at the start of the Younger Dryas cooling. They then increased steadily through the Younger Dryas, reaching typical interglacial values once more (*ca.* 275 ppmv) in the Holocene. The rapid late Allerød decrease in CO_2 concentration preceded the Younger Dryas temperature drop, possibly by several decades.

As seen from Fig. 2.2.3.4, CO_2 concentration grew with relatively insignificant variations between 16,000 and 11,000 years ago. At the same time concentrations of methane (CH_4) changed very significantly. By radiation forcing methane is the second gas after carbon dioxide, though methane concentrations in the atmosphere are significantly less than CO_2 content but the radiation effect of methane is 25 times stronger than that of carbon dioxide. Figure 2.2.3.4 shows a close correlation between methane concentrations and temperature variations. A time shift of 50 years is seen between air temperature changes and methane concentrations. As seen, temperature changes followed those in CH_4 and CO_2. Most dramatic changes in CH_4 concentration occurred during a cooling of the Late Dryas (DR3), although its noticeable changes took place also during relatively insignificant coolings, for instance during the "8.2" ka event. By fine time resolution data from the Greenland cores NGRIP, GISP2, and DYE3 methane concentration decreased from 635 ppbv at the start of the cooling to 555± 18 ppbv during the maximum phase of temperature decrease. By model estimation methane emission decreased by 32 ± 14 Tg (from 220 to 188 ± 10 Tg) a year (Kobashi et al., 2007, 2008; Thomas et al., 2007).

Along with natural atmospheric CO_2 variations in the ice core data, there are also variations in concentrations of other greenhouse gases (in particular, methane CH_4 and N_2O) that constitute the so-called small gas contaminants and create "the greenhouse effect". Methane concentration in the atmosphere during the glacial periods was found to be half the concentration during the interglacial periods. During glacial periods it was 300-320 ppbv (parts by billion by volume) and during warming periods (interglacials) - 750-800 ppbv. Methane could play a major role in past climate change during Pleistocene glacials and interstadials due to melting underground glaciers and increasing bog formation processes during interglacials and releasing large amounts of preserved methane. Apparently Northern Hemisphere contribution into the planetary methane balance both at present and in the past has been larger than that of southern due to a bigger area of underground glaciers (Flückiger et al., 2002).

MacDonald et al. (2006) have shown that a substantial contribution to peak values of CH_4 during the Early Holocene originated from peat-bog emissions in the high northern latitudes following the retreat of ice sheets during the deglaciation period. CH_4 emission from high latitudes at the start of the Holocene occurred in the context **of maximum northern hemisphere insolation, minimum precession and maximum obliquity** producing warming in high latitudes. All of greenhouse gases (CO_2, CH_4 , N_2O) show a generally positive trend over the past 10,000 years (Flückiger et al.,2002).

The "greenhouse" gas concentrations (CO_2, CH_4, and N_2O) obtained from ice core show a significant radiation contribution of these gases into an air temperature change in glacial-interglacial cycles of the Late Pleistocene (Borzenkova and Trapeznikov, 2005). A direct radiation effect from greenhouse gases has been supposed to increase due to feedbacks with water vapour, changes in sea ice volume (via albedo) and in cloudiness. According to Petit et al. (1999) changes in greenhouse gas concentrations could be responsible for about 80% air temperature variability during the transition from glacial to the Holocene. Lorius et al. (1985) showed that a change in CO_2 concentration could contribute as much as 50% of air temperature variability (about 2-3°C) during the glacial-interglacial cycles of the Pleistocene.

2.2.3.4. Albedo variations as a climate forcing factor The conditions of solar radiation reflection from the Earth's surface are defined by the value of albedo equal to the ratio of reflected and incoming radiation. Empirical data shows a great difference in solar radiation absorption on land

and water surfaces. The greatest albedo values of 0.90-0.95 are observed for clean and dry snow. In the absence of snow, the largest land-surface albedo of up to 0.50 is observed in desert regions. Damp-soil albedo is usually less than that of a similar dry soil. Albedo of natural surfaces with a dense vegetation cover varies within comparatively narrow limits of 0.10 to 0.20-0.25. Albedo of the Earth-atmosphere system is of a more complicated nature than that of the Earth's surface. Its value is about 0.28-0.30 with no ice sheet in high latitudes. Albedo of the earth-atmosphere system without clouds is usually greater than that of the Earth's surface. With cloud presence, it is also usually greater than the Earth's surface albedo, except for the cases when the surface is covered with more or less clean snow. As compared to the modern epoch, the greatest global air temperature effects of the earth-atmosphere albedo took place during the non-glacial periods and during the large glaciations of the Earth.

2.2.3.5. Modes of variability In addition to the above factors that can be considered as "external" (external forcing) there are those (internal forcing) that lead to climate variability on multi-centennial timescales caused by extremely complicated and non-linear atmosphere-ocean interactions (Forthcoming climate changes , 1991).

Studying the variability of the Atlantic Overturning Circulation (AMOC) is of particular interest in understanding the causes of multi-centennial to multi-millenium timescales climate change in the Lateglacial-Early Holocene. This type of climatic variability was first discovered in studying deglaciation in the Northern Hemisphere and climatic changes in northern and temperate latitudes during the Lateglacial-Early Holocene. Proxy data from the areas adjoining North Atlantic shows rapid climatic coolings related to upper layer sea water freshening due to influx of large volumes of meltwater from disintegrated continental ice sheets (Scandinavian and Laurentide). These can cause a weakening or even shutdown of the meridional overturning circulation (THC). Analysis of data in the above chapter shows that these processes took place in reality in the Baltic Sea Basin during the Early Holocene. General circulation models have shown that similar AMOC disturbances can take place in the future with developing modern global warming and changing hydrological cycle due to increased precipitation in high latitudes (Ágústsdóttir, 1998; Alley, Ágústsdóttir, 2005; Ganopolski, Rahmstorf, 2001; Renssen et al., 2001a,b)

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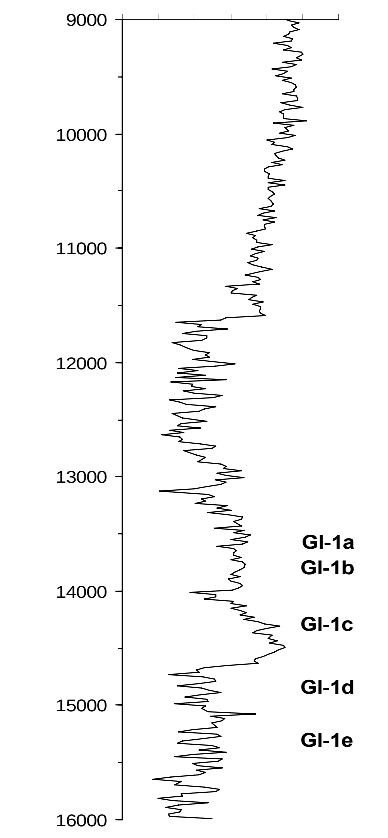
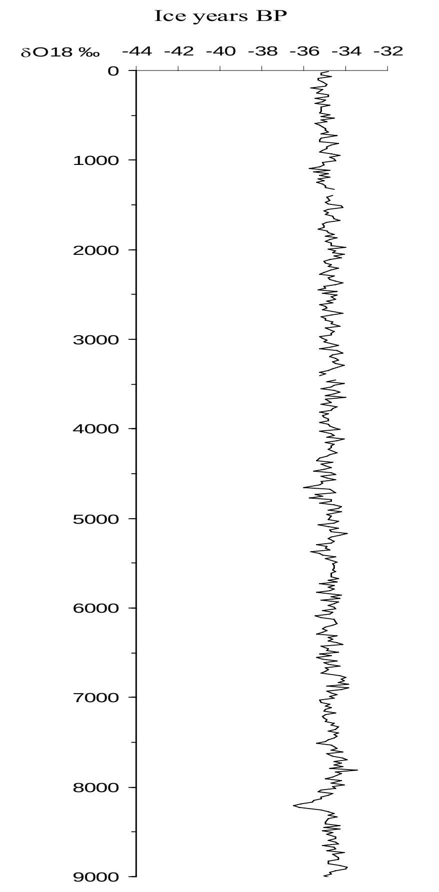


		9000
PB4		
		9300
PB3	Lengholtz g.a. (the Alps), Yoldia 2 tr. in the Baltic Sea **Saksunarvatn tephra layer**	9600
PB2	Piottino g.a.(the Alps), Eidford cold phase (Sweden)- Rammelbeek (Netherlands)- Pereyaslavl (Russian Plain)- (k)	
		9800
PB1	Frisland Europe)-Polovetsk (Russian Plain) (t), - Yoldia I tr. of the Baltic Sea **Vedde tephra layer**	10300
DR3	Ra-OS g.a. (the Alps) Salpausselka (Fennoscandia)- Tromso-Lyngen (Norway)- **Laacher See tephra layer**	11000
AL	Alleröd (Europe (t)	
		11800
DR2	Earlier Dryas (Europe) –Neva (Russia)- Skarpms (Norway) g.a.	
		12300
BO	Bolling-Windermeer	
		12800
DR1	Earliest Dryas-Luga g.a.	
		13300
RA	Raunis-Makkinow-Kokorevo (t)	
		13800
PD	Krestay-Port Brus g.a.	
		15300
LA	Lascaux-Tab-Yakha-Erie (t)	
		16000

Figure 2.2.1. Climatic events in the Baltic Sea Basin between 16,000 and 9,000 years ago.



g.a. - glacial advances, tr - transgressions, rg - regressions, t - thermochrones, k - cryochrons		Approximated age of bound areas by ^{14}C
SA4	Warming of the XIX-XXth centuries	170(130)
SA3	"Little ice age", g.a. Fernau in the Alps	800
SA2	"Viking warming", Dyunkerk and Nimphey trs. (the Black Sea), Derbent rg.(the Caspian Sea)	1200
SA1	Goshenen g.a. in the Alps, Phanagoriya rg. in the Black Sea	2500
SB4	Subboreal warming 2	2900?
SB3	Lobben-Zimming g.a. (the Alps)	3500?
SB2	Subboreal warming 1	4600?
SB1	Piora-Rotmoos g.a. in the Alps, Raphael g.a. (the Chilean Andes)	5300
AT5	"Late Atlantic optimum" tr. Littorina 3 (the Baltic Sea, tr.)- Fromentain-Nouakchott tr.(Afrika)	6200
AT4	Larstig-Frosnits 2 g.a. in the Alps	6400
AT3	Littorina 2 tr. in the Baltic Sea	6800
AT2	Misox-Frosnitz 1 g.a.(the Alps)	7500
AT1	Littorina 1 tr. (the Baltic Sea), Tyrreli in the Canada	7900
BO4	Frobisher Bay, Cochrane I (8.2 kyrBP), Cochrane II (8.0 kyrBP)-Cockburn g.a. (N. America), Novosanchugovskoe (Siberia)	8300
BO3	Ancylus II tr. in the Baltic Sea	8500
BO2	Gothard g.a. (the Alps), Gousan tr.(Caspian Sea)	8700
BO1	"Early Boreal optimum", Ancylus 1 tr. of the Baltic Sea	

Figure 2.2.2. Chronology of the climatic events in the Baltic Sea Basin during the last 9,000 yeras

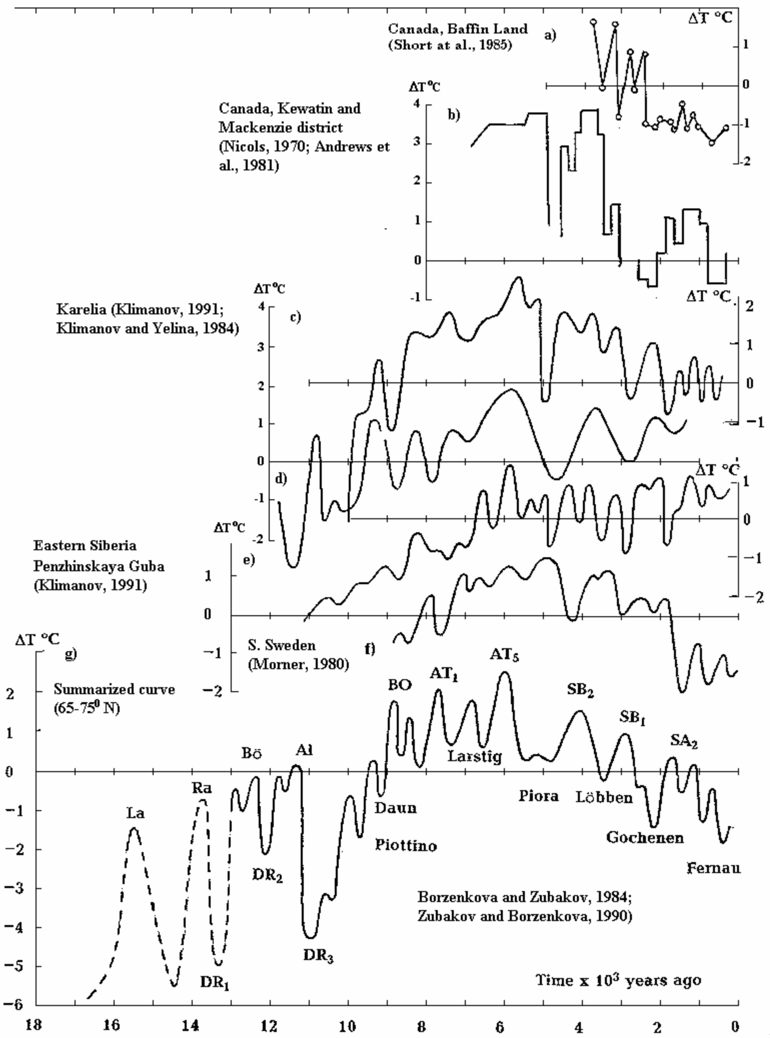


Figure 2.2.3. Deviation of the summer air temperature from modern one in the different regions closed o the Baltic Sea Basin

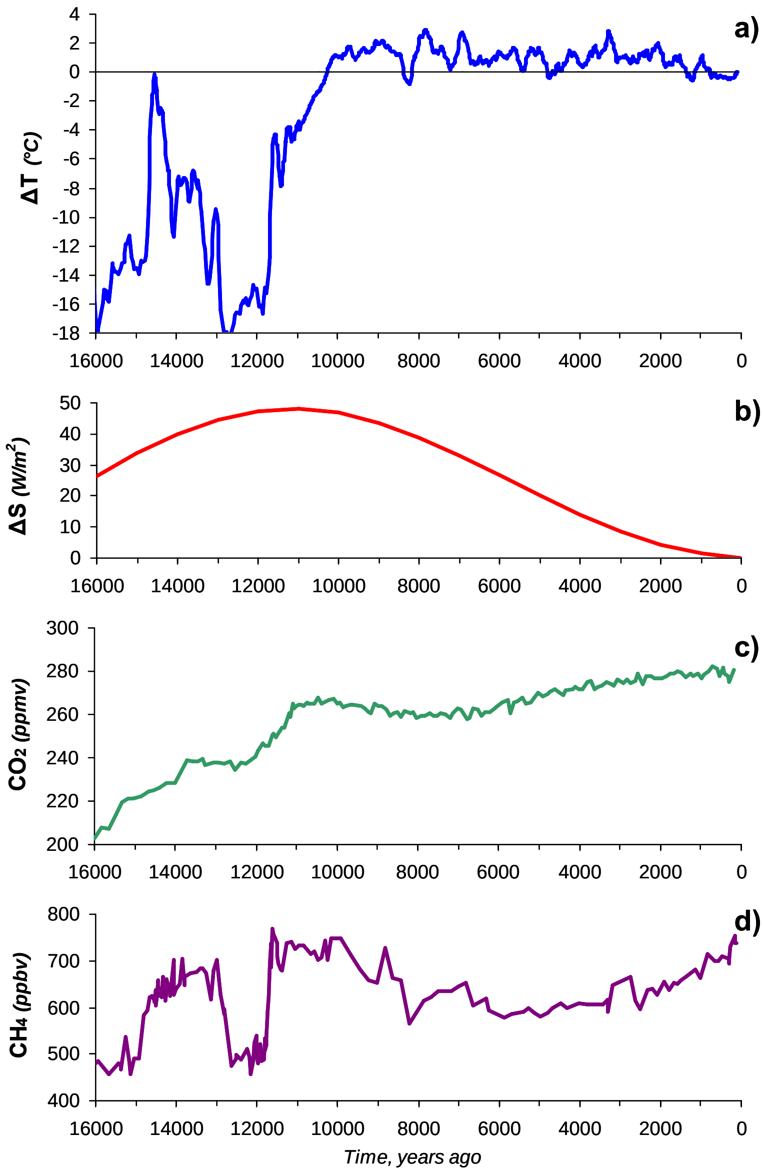


Figure 2.2.4.

a) The air temperature changes at the high latitudes obtained by ice cores (Greenland, Summit, GISP2), anomaly of the solar radiation incoming to the upper atmosphere due to astronomical factors (b) and greenhouses gases concentration CO2 (c), CH4 (d) during the last 16,000 years