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# Climatic Changes in Northern Europe Since Late Glacial Times,
## with special reference to dendroclimatological studies in northern Finnish Lapland

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### *Abstract*

*The oxygen isotope variations in deep Greenland ice cores indicate large and extremely rapid temperature oscillations during the last ice age. Strongly fluctuating climatic conditions also characterized the final few thousand years of the Weichselian glaciation. Temperatures rose to levels near their present values during the Bölling-Alleröd interstadial, but during the Younger Dryas episode (12,700 - 11,500 years ago) the cold ice age conditions came back. The cold spell ended with an abrupt warming about 11,500 years ago and it marked the beginning of the present warm interglacial, the Holocene. The Holocene climate has also varied, but all oscillations in temperatures have been an order of magnitude smaller than those common in the North Atlantic region during the last ice age. Temperatures rose to about present-day values in early Holocene time and generally continued to rise slowly until 6000 - 5000 BP. During the last 5000 years climatic conditions have gradually become cooler and obviously somewhat more unstable. The variations in humidity show some differences in various regions of Fennoscandia, but since 2500 BP, wetness has generally increased. There are different climatic proxy data which can be used for Holocene studies, but the extraction and interpretation of signals of small short-term changes is often difficult because of unavoidable inaccuracies in data and dating. The main trends of Holocene climatic development are relatively well known over northern Europe, but the short-term variations are not known in detail. Tree rings provide a possibility to study high-frequency climatic variability, since annual and even seasonal resolution in dating can be achieved using dendrochronology. The pine tree-ring data collected from the tree-line area of northern Fennoscandia indicate changes in past summer temperatures. The absolute tree-ring curve constructed from subfossil pines (*Pinus sylvestris, *L.) of Finnish Lapland indicates a duration of over 2000 years, extending to 165 B.C. and, after a ca. 200-year gap, the older unbroken part of the chronology extends until about 7500 calendar years before present. Preliminary interpretations of the data suggest that the climatic variability in the study area increased around 5000 BP with a subsequent trend towards cooler and wetter climatic conditions.*

*Key words: climatic proxy data, dendrochronology, Holocene climate, humidity variations, Scots pine, subsfossils, Lapland*

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## 1. *Introduction*

The Earth's climatic system is extremely complicated, with numerous connections and feedback mechanisms which are difficult to explain in detail. To understand modern climate, we need, among many other things, information about the climatic development in the past. Palaeoclimatological data show that numerous changes and fluctuations in climatic conditions have taken place both globally and regionally. Our knowledge of past climatic development is expanding rapidly while at the same time our understanding of the causal mechanisms of changes is increasing. Despite the remarkable advance in these studies, however, many fundamental questions remain unanswered.

The climatic history of the past 2.5 million years (the Quaternary Period), is characterized by extremely strong fluctuations in temperature, which have caused the appearance and disappearance of many ice ages and of warm interglacial periods. A periodicity of about 100,000 years has been found in the interglacial-glacial cycles during the past hundreds of thousands of years and it has been shown that these cycles are linked to the small variations of the Earth's orbit and related changes in solar insulation. The celestial mechanics in the solar system causes small perturbations in the geometry of the Earth's orbit and these result in periodical variations in the regional distribution of solar insulation. According to the Milankovitch theory of climatic change, these variations in solar radiation trigger processeses which lead to enormous climatic and environmental changes. The orbital periodicities include the 41,000-year cycle in the tilt of the Earth's axis of rotation with respect to the plane of the orbit; the 22,000-year precession cycle (moving of solstices and equinoxes along the elliptical orbit); and the 100,000 cycle in the eccentricity of the orbit. Climatic cyclicity at these periodicities has been observed during the Quaternary Period, but for some reason the occurrences of cold ice ages have commonly followed a 100,000-year eccentricity cycle (*Imbrie* and *Imbrie,* 1979; *Berger et al.*, 1984; *Berger,* 1989).

Even though certain cycles can be observed in climatic record, climate will never return to a status which is exactly similar to a previous one. Each ice age and warm interglacial differs in many features from its predecessors and, for example, the temperature development during the past well-known interglacials has been clearly different from the others during each thermomer (Fig. 1).

Continuous, never recurring development can also be seen in the high-frequency variability during shorter sequences of time. The variations in the annual ring-width of trees constitute one good example in this respect. It is possible to correlate two or more tree-ring series with each other, provided the trees have been growing for at least several tens of years at the same time. There are no identical sequences of long ring-width variations. This is the basis for dendrochronologcal cross-dating (*Fritts,* 1976; *Schweingruber,* 1988).

We live in what is a relatively warm period in the context of Quaternary climate development. Fluctuations and changes in temperatures and in humidity have occurred in different regions, but on average the climatic conditions have been comparatively stable since the end of the last ice age.

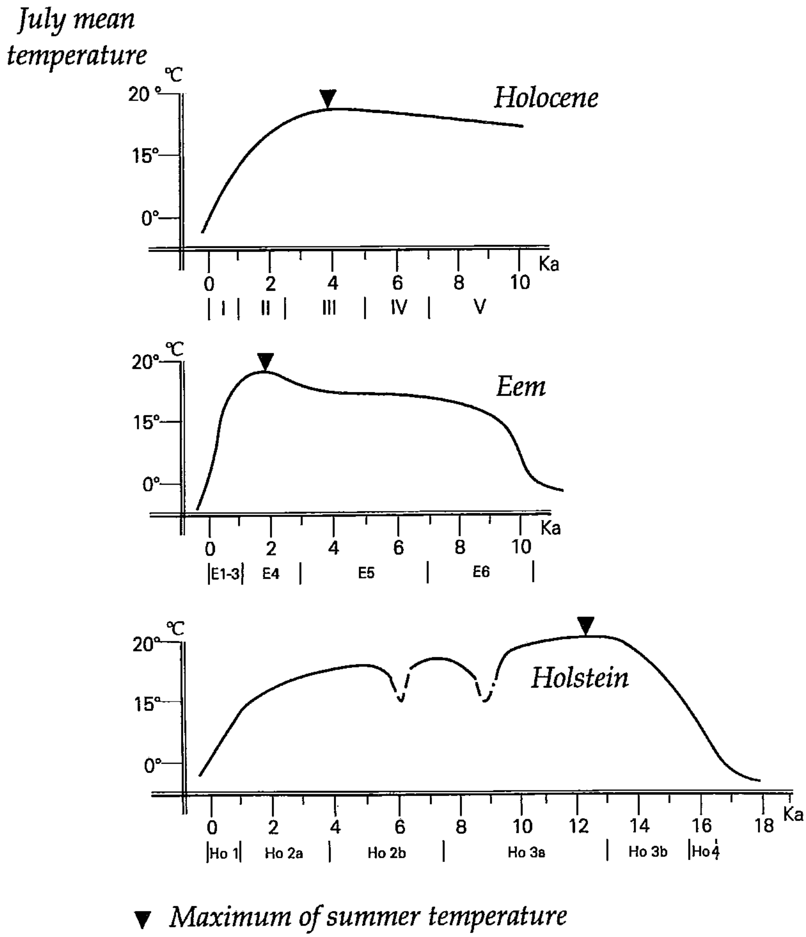


Fig. 1. Comparison of temperature development during three Quaternary interglacials. The estimations are based on mean July temperatures inferred from palaeobotanical evidence. The summer temperatures reached their maximum values at various time intervals during the ongoing Holocene and earlier Eem and Holstein interglacials (after *Zagwijn*, 1992).

## 2. *Climatic changes towards the end of the Weichselian ice age*

The ice ages are by definition cold periods, but it has been known for a long time that remarkable climatic changes also occurred during the ice ages. Still, the magnitude of temperature oscillations found through the studies of deep Greenland ice cores was a surprise, when conclusive evidence on them first began accumulating in the early 1990's (*Johnsen et al.*, 1992; *Dansgaard et al.*, 1993; *Greenland Ice-core Project (GRIP) Members*, 1993). The variations in the oxygen isotope ($^{18}O/^{16}O$) ratio in the 3000 m deep Greenland ice cores indicate that there were numerous abrupt temperature oscillations during the last (Weichselian) ice age. Very cold periods were interspersed with episodes of relatively mild climate (interstadials), which generally lasted from 500 to 2000 years. Temperatures changed by even more than 10 degrees in only a few decades. These changes from cold to cool conditions and vice versa are known as Dansgaard-Oeschger cycles. They were irregular, occurring at differing time intervals. The warmings were abrupt, but the coolings were most clearly more gradual (*Johnsen et al.*, 1992; *Dansgaard et al.*, 1993).

The climatic oscillations were related to iceberg discharges into the North Atlantic (*Bond et al.*, 1993; *Bond* and *Lotti,* 1995). The icebergs must have been produced by increased flow of the ice sheets bordering the North Atlantic during glacial times and these ice discharge occurrences must have had a strong effect on the thermohaline circulation in that oceanic area. The cause and effect relationships linking the climatic, glacial and oceanic effects are, however, still largely unknown at this time. Probably the strongest temperature oscillations during the last ice age took place in the North Atlantic and adjacent areas, but simultaneous climatic changes also occurred in other areas. The winter monsoons increased in China during the North Atlantic cold episodes (*Rutter et al.*, this volume) and traces of contemporaneous iceberg discharge events have been found in the North Pacific area (*Kotilainen* and *Shackleton,* 1995). The ice age climatic fluctuations have thus affected very large areas.

Strong climatic oscillations continued into late glacial time. Between 15,000 and 14,500 calendar years ago an abrupt global warming took place and the ice age seemed to be rapidly nearing its end. Temperatures generally rose to interglacial levels, causing the ice sheets in northern Europe and in North America to melt rapidly. At the same time the forests began to spread northwards to the former steppes and tundras and newly deglaciated areas. (*Björck* and *Möller,* 1987; *Donner,* 1995). This warm phase is known as the Bölling-Alleröd interstadial, which lasted until ca. 12,700 calendar years before present. Even though relatively temperate conditions prevailed, a declining trend in temperatures can be seen in the ice-core data towards the end of the interstadial (Fig. 2). Finally, it terminated with an abrupt cooling.

The ensuing cold stadial, called the Younger Dryas, lasted about 1200 years, ending about 11,500 years ago. It was the most significant rapid climatic oscillation during the Late Weichselian deglaciation. A general steep fall in temperatures can be seen in the Greenland ice-core data (Fig. 2) and in various proxy data in areas adjacent to the North Atlantic, even though there are some regional differences in the details (*Lowe et al.*, 1994). On the Norwegian coast a cooling of 5 to 6 °C occurred probably in a few decades (*Mangerud,* 1987) and in Great Britain the rapid lowering of temperature is recorded in fossil beetle faunas (*Coope* and *Brophy,* 1972; cf. also *Lemdahl,* 1988). A stagnation in the retreat of the ice margin of the Scandinavian ice sheet gave rise to the large Salpausselkä end-moraines in Finland and corresponding formations in Central Sweden and on the coasts of Norway (*Donner,* 1995; *Lundqvist* and *Saarnisto,* 1995).

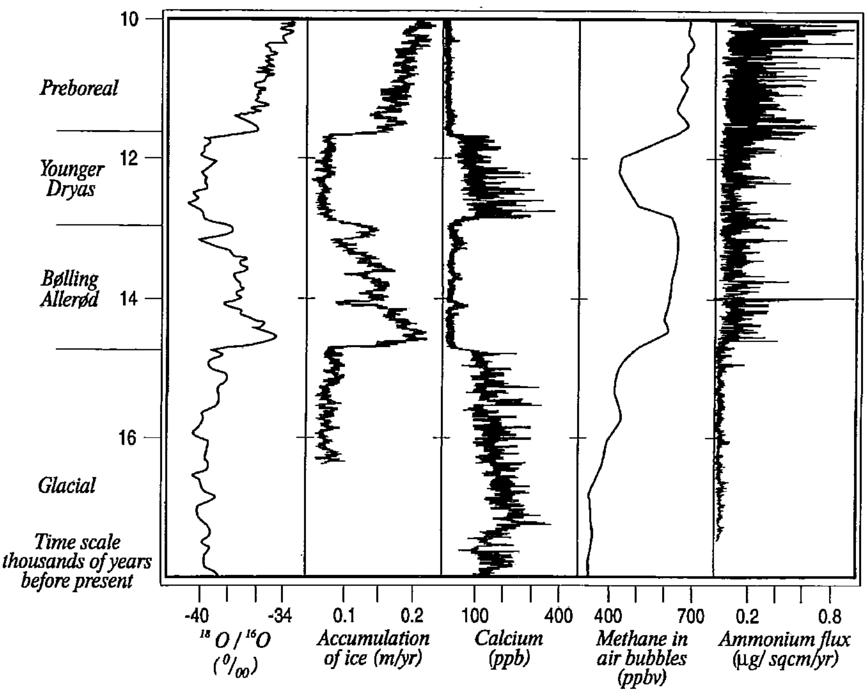


Fig. 2. A view of the period 18,000 to 10,000 (calendar) years ago based on five different parameters studied in the Greenland ice cores; oxygen isotope variations, accumulation rate of ice, amount of deposited calcium in ice, concentration of methane in the air bubbles of ice, and calculated ammonium flux (cf. text). According to Mayewski (1994).

The climatic conditions during the Younger Dryas were similar to those during the earlier glacial times. This can be seen in Fig. 2, which shows the change in five different parameters in the Greenland ice cores. The oxygen isotopes indicate low air temperatures and the low accumulation rate of ice is indicative of dry climatic conditions. The high deposition rate of calcium on the ice sheet shows that the dustiness of air (loess transport and deposition) increased back to the pre-Bölling levels. There also was a drop in the atmospheric concentration of methane, which probably indicates a reduction in the area of tropical and subtropical wetlands. The ammonium flux shows no marked change during the Younger Dryas and it may show that the northern high-latitude continental vegetation was not very widely affected by this cold spell (Fig. 2, *Mayewski,* 1994). The studies on atmospheric $^{14}C/^{12}C$ values in different proxy records show that a major perturbation in the oceanic circulation took place during the Younger Dryas, resulting in changes in the carbon exchange between the atmosphere and the world oceans (*Edwards et al.,* 1993; *Goslar et al.,* 1995). The changes in the carbon dioxide concentration are not included in Fig.2, but it is known from Antarctic ice core data that the concentration of atmospheric $CO_2$ increased markedly towards the end of the last ice age (*Barnola et al.,* 1987).

The Younger Dryas terminated with a warming that, in the North Atlantic area, was even more abrupt than the cooling at the beginning of this episode. Oxygen isotopes measured from the Greenland ice core imply that the rise in temperatures was as much as 7 °C within a period of only 10 years (*Alley et al.*, 1993; *Mayewski,* 1994). Some other calculations have resulted in even considerably larger values for this abrupt warming (*Cuffey et al.*, 1995). Similar rapid warming has been found also in other parts in the North Atlantic sphere. Some glaciological data suggest that in Norway the temperatures rose about 7 °C, probably within a few decades (*Mangerud,* 1987; cf. also *Birks et al.,* 1994) and the warming was also abrupt in Britain, according to the fossil beetle faunas and palaeobotanical evidence (*Atkinson et al.*, 1987; *Walker et al.*, 1994). In southern Scandinavia, however, the late glacial temperature oscillations seem to have been of slightly smaller amplitude and more gradual than in the areas discussed above (*Berglund, B.E. et al.*, 1984, 1994; *Björck* and *Möller,* 1987; *Lemdahl,* 1988; *Berglund, M.,* 1995). It may be that the temperature changes were most pronounced and most abrupt in the North Atlantic and in regions directly bordering it.

## 3. *Main trends of temperature changes during the Holocene*

A major reorganization of the Earth's climatic system took place when the Younger Dryas cold episode ended and the temperatures jumped to interglacial levels at the Pleistocene/Holocene transition about 11,500 years ago. This date for the beginning of the Holocene is based on counting of annual ice layers in the Greenland ice cores (*Dansgaard et al.*, 1993; *Mayewski,* 1994). The age for the beginning of the Holocene is on the basis of radiocarbon dating 10,000 years BP (Before Present, calculated from A.D. 1950). Thus there is a considerable difference between the calendar or sidereal years and the radiocarbon years. Later, in early and mid-Holocene times, radiocarbon chronology gives dates which are hundreds of years earlier than the calendar years. Many important climatic proxy data, are, however, derived from organic deposits, which are commonly dated by the radiocarbon method. Even though there are tables and computer programmes to convert the $^{14}C$ dates to calendar ages (*Stuiver et al.*, 1993), it is necessary to use the conventional radiocarbon ages in the discussion. (The capital letters BP thus indicate that the age is given in radiocarbon years).

Large continental ice sheets still existed in Northern Europe and North America at the beginning of the Holocene, but in the warm climate they began to shrink rapidly. The last remnants of these ice sheets had disappeared in Scandinavia by about 8500 BP and in North America slightly after 7000 BP (*Lundqvist* and *Saarnisto,* 1995). Vegetation, first the herbaceous pioneer plants and then the forests, spread northwards behind the retreating ice margins (*Birks,* 1986; *Webb et al.*, 1993; *Donner,* 1995).

The composition of forest vegetation, reconstructed on the basis of pollen analyses, is one of the most important indicators of past climates. The immigration of forest trees, however, follows the climatic changes with considerable time lags. Other changes in the composition and distribution of forests also lag behind the changes in climate, since some species can survive up to even hundreds of years despite a change to unfavourable growing conditions. Critical climatic thresholds and competition between species can also affect the development of vegetation (*Bradshaw,* 1993; *Lowe,* 1993). Thus it is impossible to distinguish the abrupt warming at the Younger Dryas/Holocene transition in the palaeobotanical data. The early Holocene vegetation does not indicate real climatic conditions in northern areas at that time, because the different tree species could not migrate immediately to the new climatically favourable regions.

The pollen data show that the main vegetation in Finland during the early Holocene Preboreal Chronozone (Fig. 3) was birch woods (*Donner,* 1995; *Huntley* and *Prentice,* 1993). The vegetation, dominated by birch species, is generally associated with cool climatic conditions which prevailed e.g. during some Quaternary interstadials (*Donner,* 1995), but the Preboreal birch woodland in Finland was probably not an expression of a cool climatic phase. Temperatures were still rising relatively rapidly in the early Holocene and it is reasonable to assume that they reached approximately their present levels by 9000 BP. Birch dominated the flora because of the longer migrational lag of other tree species. In southern Scandinavia and in the Baltic region there are components in the past vegetation which indicate that the summers were warmer around 9000 BP than at present (*Huntley* and *Prentice,* 1993). An important factor causing the warm summers at that time was the orbital geometry which brought about almost 8% stronger summer solar radiation than at present (Fig. 4.).

The warming trend continued until about 6000 BP (*COHMAP Members,* 1988; *Huntley* and *Prentice,* 1993), but there were some regional differences (cf. *Zagwijn,* 1994) and certainly short-term oscillations within the main development, as discussed below. The palaeobotanical data indicate that during the mid-Holocene climatic optimum temperatures in northern Europe were about two degrees higher than at present. The distribution of many thermophilous trees and other plant species was considerably more extensive than at present (*Donner,* 1995) and the pine forest in the north extended beyond its present-day limit (*Eronen* and *Huttunen,* 1993). During the latter part of the Holocene temperatures became gradually lower.

## *4. Holocene changes in humidity*

The best proxy data with which to study late Quaternary changes in humidity are the geomorphological features and sediments which show fluctuations of lake levels. These were enormous in arid regions, with amplitudes of tens or even a few hundreds of metres (*Goudie,* 1983; *Smith* and *Street-Perrot,* 1983). Generally speaking, the ice age climate was drier compared to the Holocene (cf. Fig. 1), but the changes in rainfall have not been parallel in different regions. Very impressive long-term fluctuations in humidity occurred in the Sahara Desert area. After dry glacial conditions, the increased strength of monsoonal winds brought more moisture so that in the early Holocene

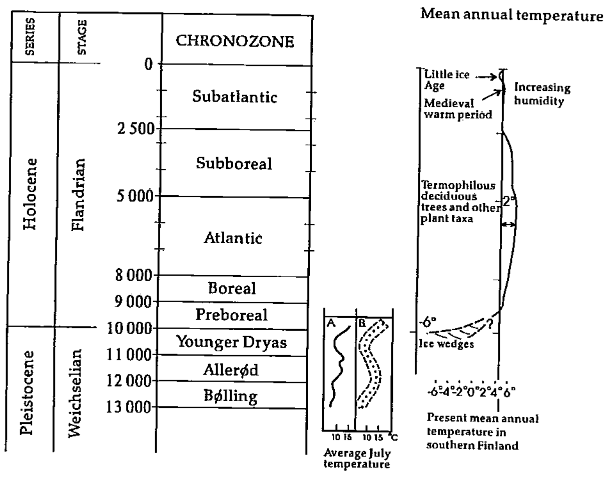


Fig. 3. Chronostratigraphical division of the youngest phases of the Quaternary Period in northwestern Europe (*Mangerud et al.*, 1974). The Late Weichselian and Holocene climatic variations are shown on the right. The lower curves show summer temperature variations in southern Scandinavia during the Late Weichselian deduced from changes in vegetation (A) and from the fossil beetle faunas found in the sediments (B) (*Lemdahl,* 1988). The upper diagram on the right depicts estimated climatic variations in southern Finland, which largely represent the general climatic development in northern Europe during the past 10,000 (radiocarbon) years.

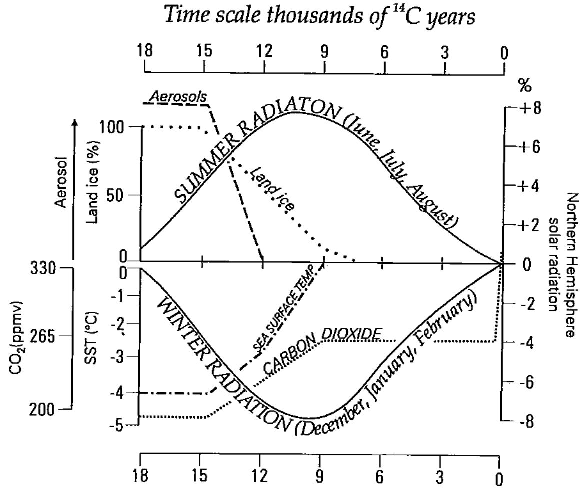


Fig. 4. A scheme showing the boundary conditions used in the COHMAP simulation of climatic development during the past 18,000 (radiocarbon) years. The smooth curves indicate the Northern Hemisphere solar radiation for summer (June, July, August) and winter (December, January, February) expressed as percentage difference from the present. The surface boundary conditions used in the modelling include land ice volume (100% = value for 18,000 BP) and global mean annual sea-surface temperatures (SST). Schemes of the changes in atmospheric aerosol and carbon dioxide concentrations are also given (*COHMAP Members*, 1988; *Kutzbach* and *Ruddiman*, 1993).

savanna vegetation prevailed there. The strong monsoons were caused by high solar radiation during the summer and when this gradually reduced, the monsoon winds decreased and the desert began to expand after 6000 BP (*COHMAP Members,* 1988; *Street-Perrot* and *Perrot,* 1993; cf. also Fig. 3).

In northern Europe the postglacial changes in humidity were small compared with the fluctuations in arid regions. Conditions were humid in northern Europe during the entire Holocene and, consequently, large climatically induced changes in lake levels did not occur. Major changes in the development of large lakes have been caused by land uplift in glaciatiated areas and lake levels have been artificially altered by Man. Because of these and some other interpretational difficulties in studies, the changes in humidity during the Holocene are not particularly well known in Europe (*Yu* and *Harrison,* 1995).

The most detailed investigations of Holocene lake-level fluctuations have been done in southern Sweden (*Digerfeldt,* 1988). According to these studies, the climate was dry and lake levels low at 10,000 to 9000 BP, but afterwards the humidity increased and lakes reached high levels between 8000 and 7000 BP. The levels fell after 7000 BP and conditions remained drier than at present until ca. 2500 BP. After this wet phase lake levels fell by several metres, but then rose again after 1500 BP gradually reaching their present status (Fig. 5).

Evidence of low early Holocene lake-levels has also been found in southern Finland (*Donner et al.*, 1978; *Huttunen et al.*, 1978). Diverse pieces of evidence found in the stratigraphical studies on mires and small lakes suggest that in southern Finland the wetness increased during the Boreal and early Atlantic Chronozones, as in southern Sweden (*Korhola,* 1992a; 1992b; 1995), but in Finland these changes have not been studied in as much detail as in Sweden.

The pattern of humidity changes in the southern parts of Sweden and Finland cannot be generalized over large areas, as shown by recent results from northern Finnish Lapland. Sediment stratigraphical studies in small lakes show that during early and mid-Holocene times, from about 8000 to about 4000 BP, lake levels were low. The water levels of closed-basin lakes rose to their present stands during the past 4000 years (*Hyvärinen* and *Alhonen,* 1994). The development in the north deviated clearly from that in southern Sweden, where the early Atlantic chronozone was wet. The subfossil pines collected for dendrochronological studies from the tree-line area of Lapland corroborate this conclusion. A large number of pine logs have been preserved on the bottom of small lakes and the radiocarbon and dendrochronological dates show that many of the trees grew thousands of years ago. In several cases it is possible to see in the field that the trunks or stumps are submergred *in situ*, which means that pines grew on ground above the water surface at lakesides, which were later inundated. Quantification and accurate dating of the water level rise still remains to be done (*Eronen* and *Huttunen,* 1993; *Eronen* and *Zetterberg*, in press; *Zetterberg et al.*, 1994, 1995).

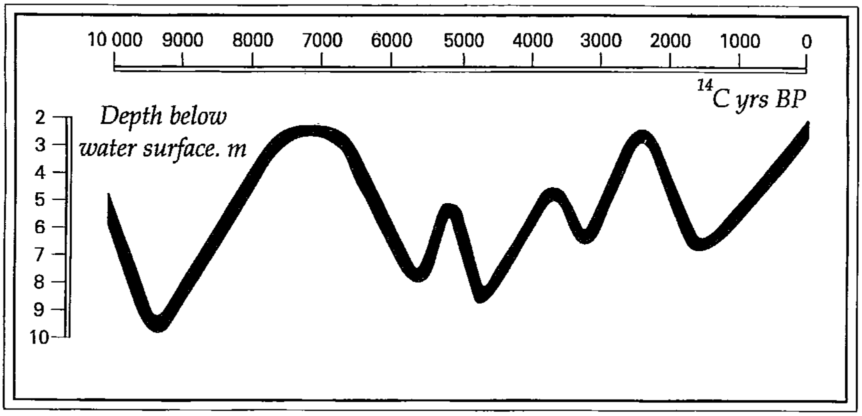


Fig. 5. A curve showing Holocene variations in the upper limit of sediment deposition in a number of small lake basins in southern Sweden. The changes in that "sediment limit" indicate water-level fluctuations of the lakes. After *Digerfeldt* (1988).

In addition to millenial-scale humidity changes, short-term fluctuations in wetness have been studied using peats of ombrotrophic bogs. Their stratigraphy often shows layers of light relatively undecomposed peat alternating with dark humified peat. The lower boundaries of the light weakly decomposed peat layers are called "recurrence surfaces" and they are believed to indicate phases of wet and cool climate. The dark humified layers likely represent warm and dry periods. Radiocarbon dates of these layers, however, differ ages and the recurrence surfaces are in fact often a result of the dynamic growth of raised bogs, and are independent of climatic changes. The vertical growth of *Sphagnum* peat in ombrotrophic mires generally increased after 2500 BP, during the cool and wet Subatlantic chronozone (*Korhola,* 1992ab; *Donner,* 1995; *Seppä,* 1995).

## 5. *Holocene glacier and tree-line fluctuations showing climatic variations*

It is generally concluded that most glaciers in Scandinavia melted by 8000 BP because of the warm climatic conditions. Present glaciers appeared during the latter part of the Holocene as the climate cooled. The climatic reconstructions based on the equilibrium line altitude (ELA) changes at the Jostedalsbre glacier in western Norway indicate that between 8000 and 6000 BP temperatures were, at most, over two degrees higher than at present (*Nesje et al.*, 1991; *Nesje,* 1992). During the Little Ice Age, around A.D. 1750, the west Norwegian glaciers grew to their largest size since the climatic optimum. The glacier expansion can be attributed to the temperature lowering (perhaps by one degree), but increased winter precipitation probably also contributed (*Grove,* 1988; *Nesje,* 1992; *Matthews,* 1977; 1993). The glaciological evidence from northern Sweden suggests glacier development which clearly deviates from the west Norwegian pattern. According to *Karlén* (1988), glaciers persisted there and fluctuated during the climatic optimum, and the Little Ice Age pronounced glacier expansion cannot be distinguished in the North.

Landforms and deposits created by mass movements on the slopes can also be interpreted in terms of climate. The solifluction activity on the Scandinavian mountains was enhanced during the past 5000 years, which can be explained by climatic cooling and an increase in soil moisture (*Matthews et al.*, 1993).

Changes in the limit of pine (*Pinus sylvestris*, L.) provide valuable palaeoclimatic information in Fennoscandia. The postglacial distribution of pine has been studied by pollen analyses and by dating the subfossil pine trunks and stumps found in lakes and peatbogs. Pine spread to Scandinavia from the south and reached its highest limits there very early, between 9000 and 8000 BP. The radiocarbon dates of pine megafossils during this time show that pines grew up to 200 -300 m above the present pine limit in southern Norway (*Aas* and *Faarlund,* 1988) and on the southern Swedish Scandes (*Kullman,* 1993; cf. also *Eronen* and *Huttunen,* 1993). It is possible that, climatically the most favourable time for the growth of pine was from 9000 to 8000 BP., when the astronomical factors provided high solar radiation during the summer time (*COHMAP Members,* 1988; *Kutzbach* and *Ruddiman,* 1993). It must be noted, however, that part of the apparent postglacial lowering of the tree-line was caused by the isostatic uplift, which occurred rapidly in the early Holocene (*Eronen,* 1979).

The pine limit rose to its highest positions in southern Scandinavia before the spread of pine to northern Fennoscandia (*Hyvärinen,* 1975; 1976). In Swedish and Finnish Lapland the maximum occurrence of pine occurred between 6000 and 4000 BP (*Eronen* and *Huttunen,* 1993; *Karlén,* 1993; *Hyvärinen,* 1993). This difference in the history of pine forests in southern and northern Fennoscandia was certainly related to the delayed migration to the north, but it may also reflect a time difference in the optimal climatic conditions between these two areas.

Substantial data on the changes of the pine limit are available from northern Finnish Lapland, where pine megafossils have been collected for dendrochronological studies (Fig. 6). Well over 1000 pine trunks have been dated by dendrochronology (*Eronen et al.,* 1996), which represents a major increase in the number of dated subfossil trees previously recorded in this area. The expansion of data has not, however, changed the earlier conclusion that the time of maximum spread of pine was 6000 to 4000 BP, but it has provided new information about regional differences in the distribution and subsequent retreat of the pine limit. The change in the pine limit was greatest in northwestern Finnish Lapland in the Enontekiö area. The present northernmost pine stands in Enontekiö are located at 400-440 m above sea level, but subfossil pines have been found at a height of 560 m and 60-70 km further to the northwest (*Eronen,* 1979; *Eronen* and *Huttunen,* 1987; *Eronen* and *Zetterberg*, in press). Most of the dated subfossils in this area are from the mid-Holocene and there are only relatively few trees from more recent periods. The subfossils collected from northeastern Lapland (Inari area) are spread temporally over the time span 7500 BP to the present, but the dates from the northernmost part of Lapland (Utsjoki area), are largely concentrated within the last few millenia (Fig. 7).

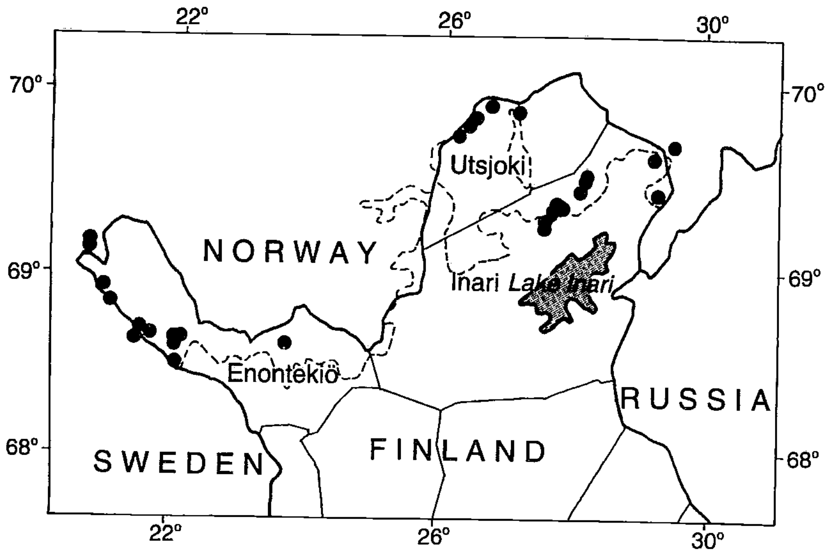


Fig. 6. Map showing the study area and sites of dendrochronological sampling (black dots) of pine subfossils in northern Fennoscandia. The broken line indicates the northern forest-limit of pine in this area.

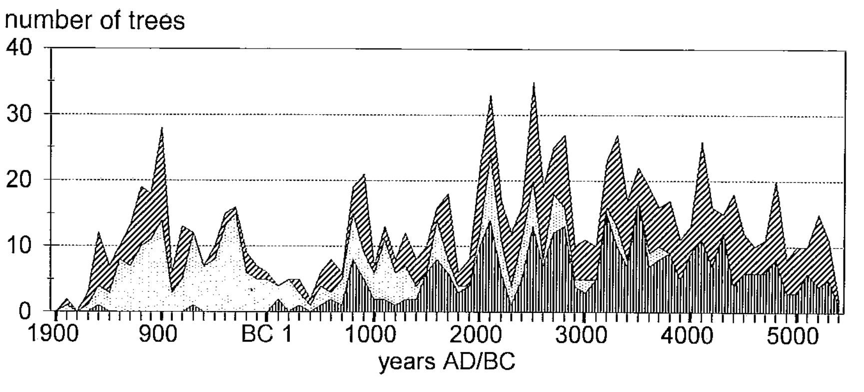


Fig. 7. Age distribution of subfossil pines from northern Fennoscandia dated by dendrochronology. The material is divided into three different geographical groups, which show temporal deviations in the occurrence of pine. The vertical hatching indicates subfossils from western part of the study area (Enontekiö), dotted area is for the northern part (Utsjoki) and slant hatching (on the top) for the eastern part (Inari) of the study area. The samples are put into age classes of hundred years in the calendrical time scale. For each sample the age is the mean value of the beginning and the end of the dated tree-ring series.

Thus the plentiful megafossil dates, from a large area, show that any large-scale oscillations in the tree-line during the past 7000 years cannot be distinguished. The pine limit has retreated and the distribution area of pine contracted, but changes have not been similar in different parts of Finnish Lapland. In fact, an important change occurred in the forest-limit zone in Lapland is the retreat due to the thinning of forest and to the reduction of the outermost stands of pine. There are isolated pines growing at distances of tens of kilometres from the present pine forest limit. We have observed, during field work, that often a few pines grew near lakes, which were sampled for dendrochronological studies. The large number of subfossil trees preserved in such lakes, beyond the present continuous pine limit, indicates that relatively dense forests stands may have existed at those sites in earlier Holocene time.

## *6. Construction of a long pine tree-ring chronology for palaeoclimatic studies in northern Fennoscandia*

Most of the palaeoclimatic proxy data available for studies of Holocene climates do not have high time resolution. The temporal uncertainty is often tens, even some hundreds, of years. Tree rings in this respect are among the most accurate source of information about the past, because they provide annual, sometimes even seasonal resolution. Changes in forest composition, tree-lines, glaciers and other variables are valuable in studies of long-term trends and also for studying some shorter oscillations of climate, but tree rings provide a reliable yearly record .

The tree rings of forest-limit pines in northern Fennoscandia are good indicators of past climates, since the radial growth of pine is, in that ecotone, determined by summer temperatures. This close relationship between pine ring-width variations and summer, especially July temperatures, has been carefully examined and verified by many studies (*Hustich,* 1948; *Aniol* and *Eckstein,* 1984; *Briffa et al.*, 1990, 1992; *Lindholm,* 1995). The mean monthly temperature of July is +12 °C in the westernmost part and +14 °C for the Inari area. For the whole study area it is approximately +13 °C. The August mean temperature varies from +10 °C to +12 °C, being approximately + 11 °C for the whole area (*Atlas of Finland,* 1987). In the study area the range of annual variation for the July-August mean temperature, which is the most important factor for the growth of pine, was approximately 4.5 °C for the period 1876-1975 (*Briffa* and *Schweingruber,* 1992).

By building long unbroken pine tree-ring chronologies it possible to reconstruct seasonal year-to-year temperature variations over thousands of years back in time.

The Finnish data now comprise 1465 pine megafossil samples collected from 42 lakes in Finnish Lapland and adjoining areas of Norway (Table 1). Over 1000 samples have been dated by dendrochronology and assembled within a master chronology for Finnish Lapland (Eronen et al., 1996). The unbroken "absolute" chronology is over 2000 years in duration, extending to the year 165 B.C. The major part of it is made up from data from pine logs collected from Lake Ailigas in Utsjoki (*Zetterberg et al.*, 1994), but there are additional samples from other lakes in the same region and the curve is fixed to present time by means of tree rings from logs from local wooden buildings and from samples cored from living pines (*Zetterberg*, 1990).

Table 1. Sampling sites of subfossil pines in Finnish Lapland. For each site the name, location, elevation, total number of collected samples and number of samples dated by dendrochronology are given.

| Site name | x | y | Latitude | Longitude | Elevation | Samples | |
|---|---|---|---|---|---|---|---|
| | | | | | a.s.l. | collected | dated |
| **Enontekiö** | | | | | | | |
| Ailakkavaaran lompolo | 76645 | 32580 | 68°57'16" | 20°57'21" | 515 | 7 | 6 |
| Vallijärvi | 76330 | 32805 | 68°41'33" | 21°35'01" | 465 | 32 | 28 |
| Eteläinen Haukijärvi | 76255 | 32970 | 68°38'17" | 22°00'13" | 465 | 9 | 8 |
| Peeran suo | 76560 | 32610 | 69°52'53" | 21°03'03" | 500 | 2 | 1 |
| Pohjoinen Haukijärvi | 76265 | 32970 | 68°38'49" | 22°00'06" | 475 | 19 | 18 |
| Tsohkkajärvi | 76275 | 32980 | 68°39'23" | 22°01'27" | 505 | 6 | 4 |
| Paijulaslompolo | 76280 | 33000 | 68°39'45" | 22°04'20" | 505 | 1 | 1 |
| Hattulompolo | 76216 | 33627 | 68°38'35" | 23°37'18" | 385 | 219 | 175 |
| Läntinen Ladnajärvi | 76355 | 32767 | 68°42'42" | 21°29'05" | 481 | 6 | 6 |
| Itäinen Ladnajärvi | 76355 | 32775 | 68°48'44" | 21°30'16" | 487 | 1 | 1 |
| Ainavarppijärven lompolo | 76332 | 32748 | 68°41'23" | 21°26'36" | 440 | 10 | 8 |
| Kelottijärven suo | 76155 | 32691 | 68°32'53" | 22°00'06" | 375 | 30 | 24 |
| **Utsjoki** | | | | | | | |
| Luossakoadneljärvi | 77515 | 34750 | 69°50'24" | 26°21'01" | 110 | 147 | 130 |
| Luossakoadnelj. lomp. | 77516 | 34752 | 69°50'28" | 26°21'19" | 110 | 5 | 5 |
| Kordsamladdu | 77646 | 34879 | 69°57'30" | 26°41'01" | 130 | 2 | 2 |
| Ailigasjärvi | 77593 | 35027 | 69°54'40" | 27°04'13" | 75 | 102 | 100 |
| Koadnelveijavri | 77555 | 34779 | 69°52'34" | 26°25'28" | 109 | 26 | 21 |
| **Inari** | | | | | | | |
| Puollimvarrinlompolo | 77340 | 35808 | 69°40'20" | 29°05'00" | 160 | 13 | 5 |
| Njargavääríjärvi | 77103 | 35845 | 69°27'31" | 29°09'26" | 220 | 16 | 10 |
| Lujapuoli 210 m mpy | 77104 | 35852 | 69°27'34" | 29°10'30" | 210 | 44 | 32 |
| Lujapuoli 220 m mpy | 77107 | 35847 | 69°27'44" | 29°09'45" | 220 | 11 | 8 |
| Tsehajaaurads | 77020 | 35262 | 69°23'47" | 27°40'00" | 197 | 30 | 29 |
| Ooggusjdaurads | 77025 | 35263 | 69°24'03" | 27°40'10" | 160 | 17 | 16 |
| Ulasjärven lompolo | 76990 | 35234 | 69°22'11" | 27°35'41" | 199 | 3 | 3 |
| Sammuttivaaran järvi | 76947 | 35208 | 69°19'53" | 27°31'40" | 207 | 10 | 7 |
| Namatesjavren lompolo | 76908 | 35188 | 69°17'48" | 27°28'34" | 215 | 107 | 87 |
| Annanjärvi | 76990 | 35239 | 69°22'11" | 27°36'27" | 205 | 10 | 7 |
| Namatesjavri | 76907 | 35190 | 69°17'44" | 27°28'52" | 207 | 12 | 11 |
| Loassamlompolo | 77006 | 35275 | 69°23'01" | 27°41'58" | 197 | 35 | 19 |
| Loassamlompolo kapea | 77002 | 35283 | 69°22'48" | 27°43'11" | 207 | 41 | 32 |
| Aulinlompolo | 77005 | 35283 | 69°22'58" | 27°43'11" | 207 | 138 | 112 |
| Kämppälompolo | 77140 | 35405 | 69°30'08" | 28°02'09" | 197 | 12 | 9 |
| Selkäjärvi A | 77168 | 35418 | 69°31'37" | 28°04'13" | 208 | 2 | 2 |
| Selkäjärvi B | 77178 | 35424 | 69°32'09" | 28°05'10" | 208 | 88 | 67 |
| Selkäjärvi C | 77175 | 35418 | 69°32'00" | 28°04'14" | 208 | 95 | 0 |
| Selkäjärvi D | 77177 | 35419 | 69°32'06" | 28°04'24" | 208 | 31 | 0 |
| Vuotkimlompolo | 77011 | 35284 | 69°23'17" | 27°43'21" | 205 | 53 | 41 |
| Pieni Vuotkimlompolo | 77013 | 35285 | 69°23'23" | 27°43'31" | 202 | 20 | 9 |
| **Norway** | | | | | | | |
| Guoppalampi | 77390 | 35925 | 69°42'47" | 29°23'23" | 147 | 7 | 2 |
| Nuvvosmohkki | 77453 | 34698 | 69°47'02" | 26°13'02" | 130 | 2 | 0 |
| Gardebårvarri I | 76880 | 32520 | 69°09'31" | 20°44'51" | 490 | 32 | 29 |
| Gardebårvarri II | 76890 | 32520 | 69°10'03" | 20°44'42" | 490 | 12 | 11 |

## 7. *Palaeoclimatic interpretations from the pine chronology*

In the following discussion of tree-ring chronology, the ages are given in calendar years, because the dendrochronological timescale is more accurate than that based only on radiocarbon dates. The long-term climatic trends are difficult to distinguish in the tree-ring data, because absolute ring-width variations are strongly dependent on the conditions at the local growing site and the age of the trees. Thus a warm period is not always expressed in increased mean width of tree rings. In fact the data from Finnish Lapland show that the mean thickness of the mid-Holocene tree rings does not differ from the late Holocene value, both being approximately 0.6 mm. Tree rings, however, do reveal the inter-annual variability of climate, and this is their particular value in palaeoclimatic analyses.

The palaeoclimatic interpretation of our tree-ring curves are still preliminary, because attempts to bridge the chronological gap prior to the Christian era has prolonged the work. The gap could not be filled despite there being one sample dated (by radiocarbon assay) exactly to the period of the gap and several other samples have been cross-matched with it. It is apparent that weather conditions were very unstable and unfavourable to pine growth during a 200-300 year period representing the gap. Its existence is indicative of an exceptional climatic episode, possibly comparable to the Little Ice Age, occuring approximately 2000 years later. Presumably the irregular growth of pine prior to 165 B.C may have been caused by cold and strongly variable weather and by increased wetness, which worsened the growing conditions on the shores of small lakes (cf. *Zetterberg et al.*, 1994).

The older part of the chronology is over 5000 years in duration, extending about 7500 years before the present (in calendar years), but its precise age is not known, because there is the gap in the curve beyond 165 B.C. (Figs. 8 and 9). The dating of this floating chronology is based on many radiocarbon dates of samples included in the master curve, and it is estimated that the limits of error in the older part are within some tens of years.

The tree-ring curve in Fig. 8 is based on raw measurements which have not been standardized. Some of the fluctuations may thus be dependent on factors other than climate, e.g. the age of trees, edaphic factors at the growing site, but it is certainly possible to draw some palaeoclimatological conclusions from these data. An important turning point in climatic development in mid-Holocene time was identified by the pioneer researchers A. Blytt and R. Sernander, who marked it as the boundary between the Atlantic and Subboreal periods. In the chronostratigraphical division *by Mangerud et al.* (1974) the age of this boundary is 5000 BP. The age is about 3800 B.C. in calendar years, and tree rings indicate a climatic shift at that time (Fig. 8). Before 3800 B.C the ring-width variations show relatively small variability, but after this time the variability increases. The data on ages of pines at death show that trees commonly survived longer at the end of the Atlantic period than in early Subboreal time. Thus the growing conditions during the former period seem to have been stable and favourable

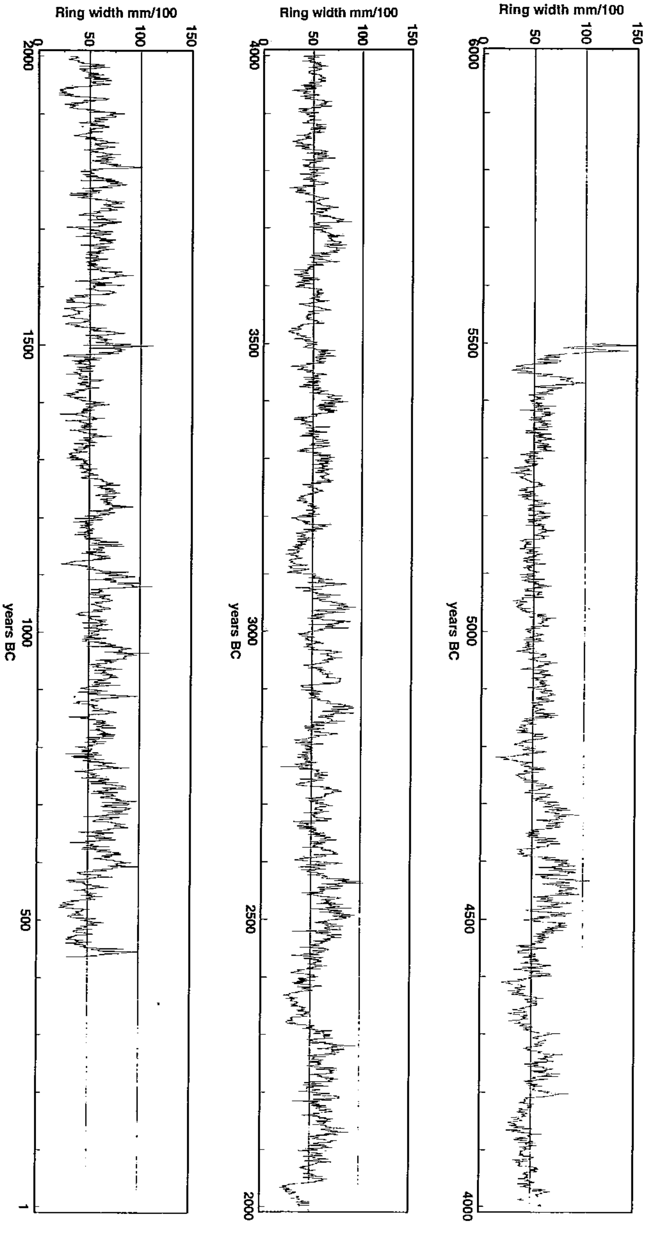


Fig. 8. Preliminary version of Finnish Lapland subfossil pine tree-ring chronology for the time period circa 7500-500 B.C. The curve is based on (unstandardized) raw mean ring-width measurements. Y-axis scale is in 0.01 mm, time scale is calendrical years.

to pine (*Zetterberg et al.*, in press). The shift probably indicates a marked regional change in climatic conditions, because a similar increased variability in ring-widths after 3800 B.C. has been detected in the pine chronology constructed for the Torneträsk area in northern Sweden (*Briffa,* 1994). As discussed above, the present mean ring-width variability corresponds to about 5 °C interannual growing season temperature variations. The mean variability in ring-width before 3800 B.C. indicates also that summer temperatures were relatively stable during the times preceding 3800 B.C, but it is impossible to calculate exactly the difference in relation to recent variability. The palaeoclimatic data generally suggest that summer temperatures during the mid Holocene period were about two degrees higher than at present which must also be taken into account. These variations in ring width took place in warmer climatic conditions than at present, but this cannnot be distinguished in these dendroclimatological data.

The next shift in the climatic pattern after the mid-Holocene occurred somewhat Little Ice Age, but the evidence does not contradict the Torneträsk data which suggest that no persistent multi-centennial period of less than 2000 years later and appears as the gap in the curve described above. That point probably marks the boundary between Subboreal and Subatlantic periods, even though the chronozone boundary defined in radiocarbon years is at 2500 BP (*Mangerud et al.* 1974). The "post-gap" unbroken pine chronology is, for the most part, so well established that it can be regarded as a master curve for northern Finland (Fig. 9, *Zetterberg et al.*, 1994, 1995, in press).

A detailed account of the above "absolute" pine chronology is given *in Zetterberg et al.*, (1994), but some major conclusions of that work can be repeated here. The younger part of the chronology is similar to the older part drawn from the non-standardized measurement data. Thus not all parts of the graph indicate the regional growth variations in exactly the same way, because some parts of the curve are based mainly on old trees and some other parts on younger trees, growing faster than the old on average. There are also variations in the replication (number of samples) in different parts of the chronology. Notwithstanding these deficiences, the curve can be considered regionally well representative, since the ring-width variations for the period from A.D. 500 to the present (time period common for both chronologies) are very much similar to the ones in the (standardized) Torneträsk curve of northern Sweden (*Briffa et al.*, 1990; 1992).

The annual peaks and troughs in the curve indicate higher and lower values of summer temperature, respectively, and, as can be seen in the long-term run, there are often oscillations on decadal time-scales. A significant feature of the curve is the lack of evidence for the Medieval Warm Period (about 900 to 1300 A.D. and the Little Ice Age (about 1550 to 1850 A.D.) (*Lamb*, 1982; *Grove*, 1988). There are pronounced warm periods between A.D. 870 and 1100, but these were interrupted several times by temporary coolings. The replication of the present data is very poor for the time of the cool conditions prevailed during it. A distinct shorter cold period does occur, in accord with the Torneträsk data, from the late 16th to middle 18th centuries (*Briffa et al.*, 1992).

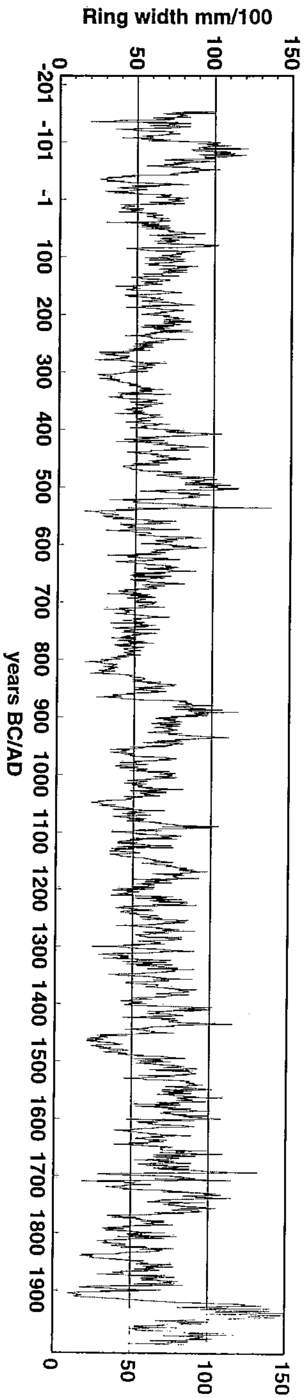


Fig. 9. Finnish Lapland subfossil pine tree-ring chronology based on (unstandardized) raw mean ring-width measurements for the time period from 165 B.C to the present. Scales as in Fig. 8.

It should be noted that the pine tree rings of northern Fennoscandia predominantly indicate summer temperatures, not mean annual temperatures. It may be that late Medieval winters were mild and cold winters contributed strongly to the Little Ice Age cooling in Europe, but the dendroclimatological results from northern forest-limit pines show that further studies, correlations and analyses of proxy data are needed to solve these problems connected with the climatic variability during the last millenium (cf. *Bradley* and *Jones,* 1993).

There are considerable possibilities for improving pine tree-ring chronologies and their climatic interpretations. The available dendrochronological data from northern Finland provide extremely valuable palaeoclimatological information, because the trees come from ecologically sensitive conditions and the annual rings contain an easily intrerpretable temperature signal.

## *8. Conclusions*

Long-term Quaternary climatic variations are driven by orbital changes which slowly alter the seasonal solar radiation at different latitudes. The Milankovitch theory, based on these cyclic variations, explains the variability of climate on timescales of tens and hundreds of thousands of years and, the occurrences of ice ages and warm interglacials. There is, however, climatic variability of shorter time spans as shown by the oxygen isotope studies of the Greenland ice cores. These temperature oscillations were strong and abrupt during the last ice age. The present warm Holocene also began with an abrupt warming about 11,500 years ago. Orbital forcing changes created the conditions of the present warm climatic period, but the slow decreasing trend in solar radiation during the summer in the Northern Hemisphere caused the trend towards a cooler climatic condition. There have also been, however, small climatic variations, caused by different stochastic changes of the earth's system. These small changes and their consequences in natural systems have generally been more or less diachronous and of differing amplitudes in different parts of the world. The explanation of Holocene short-term climatic variability is a demanding task and requires careful multi-proxy mapping of past changes. The palaeoclimatic proxy data are always somewhat inaccurate and need critical evaluation. Precise dating is also often impossible, but, some data, for example tree-ring records, have year-to-year accuracy. Detailed insightful knowledge of past climates and of the dynamic behaviour of the present climatic system, is needed for improved understanding of the complicated causal mechanisms of climate changes. The background climatic variability must be known for evaluating the magnitude and probable effects of the threatening greenhouse warming.

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