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Holocene environmental history and climate of Råtåsjøen, a low-alpine lake in south-central Norway

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Abstract

The Holocene environmental history and climate are reconstructed for Råtåsjøen, a low-alpine lake in south-central Norway. The reconstructions are based on chironomids, diatoms, pollen, plant macrofossils, and sediment characteristics. From plant macrofossil evidence, birch trees (*Betula pubescens*) immigrated ca. 10,000 cal BP. The chironomid-inferred mean July air temperature was high, but may be unreliable during the early stages of the lake's history due to the high abundance of *Chironomus anthracinus* type, a taxon that may include several species. From ca. 9000 cal BP the inferred mean July temperature was lower (ca. 9 °C). Temperatures increased towards 8000 cal BP and pine (*Pinus sylvestris*) reached its upper limit near the lake. July temperature may have become a significant factor controlling long-term pH in the lake, starting shortly after 8000 cal BP. High pH values were associated with periods of warm summers and lower pH values occurred during periods of colder summers. Alkalinity processes within the lake and/or the catchment are possible factors controlling this relationship. A temperature decline at ca. 5400 cal BP separated two 10.6 °C temperature maxima around 6400 and 4500 cal BP. The 1.5 °C decline in July air temperatures from ca. 4400 cal BP was paralleled by a decrease of pH from 7.2 to 6.8. Following the temperature drop, first pine and then birch trees declined and disappeared from the catchment and organic accumulation in the lake increased. The increased organic accumulation rate had a positive effect on diatom production. At ca. 2700 cal BP the temperature reached a minimum (ca. 9.2 °C) and correspondingly a second pH minimum was reached. Temperature decreased again slightly at ca. 400 cal BP during the 'Little Ice Age', before increasing by about 0.5 °C towards the present. Percentage organic carbon as estimated by loss-on-ignition appears to be better correlated with chironomid-inferred July temperatures than organic accumulation rates, at least for the last 9000 years. Accumulation rates of organic sediments are more coupled with catchment-related processes, such as erosion and major changes in vegetation, than is percentage organic carbon.

Introduction

The environment and its flora and fauna are constantly changing, and biota responds in relation to a range of interacting abiotic and biotic factors. An understanding of these natural dynamic changes, both in terms of climate and biotic responses to climate, is important in order to anticipate future climatic and biotic changes. The past is the key to the present and also to the future. In the present, we are unable to measure directly past environment and climate and thus have to use stratigraphical records, such as proxy indicators, to reconstruct them from a temporal sequence, such as lake sediments.

Different environmental proxies have individual strengths and weaknesses. Where one proxy has limitations, another may provide useful information (Ammann et al. 2000; Campbell et al. 2000). By combining and interpreting information from several proxies together, the overall picture of past environment and climate may be more comprehensive than the information gained from the equivalent number of single proxies considered alone (Birks et al. 2000). Similarities between individual results may assist hypothesis generation and may help validate results. Outliers and discrepancies between proxies may initially look like 'noise'. However, the challenge of interpreting such discrepancies may add information about co-influencing variables (Campbell et al. 2000). A thorough knowledge of the mechanisms of biotic response will guide us in the process of understanding biotic changes in light of environmental change. There is, however, a danger of over-simplification and circularity of argument when a few proxies are selected to represent the whole ecosystem and at the same time to reflect climate change. The danger lies in inferring past climate from biota and then explaining changes in the ecosystem from the estimated climate shifts. Biota respond to both climate change and ecosystem processes (e.g., Wetzel 1975). Whether climate or ecosystem change is the controlling agent for the biota may be hard to determine. They may, however, be correlated and one can be used to reconstruct the other indirectly. We should keep in mind that responses are dynamic and that, for example, an animal or plant taxon may have a different environmental optimum in the presence of competitors, parasites, and predators than when in isolation (Davis et al. 1998). There may or may not be a time lag in response time (Birks 1981; Ammann et al. 2000; Lotter et al. 2000) depending on the strength of forcing factors, e.g., the large late-glacial temperature changes may override other ecosystem processes. To obtain the maximum information in a palaeoecological study, a multiproxy procedure is therefore essential (Birks et al. 2000).

Sites at high altitudes or latitudes are increasingly recognised as being potentially sensitive to climate change (Douglas et al. 1994; Kupfer and Cairns 1996; Battarbee et al. 2002a, b). The present low-alpine region of Norway has undergone extensive post-glacial vegetational changes (Eide 2003) and may be considered an ecologically sensitive ecotone (Moen 1999; Odland and Birks 1999). Lake biota in this region are potentially sensitive to Holocene climate changes as the biota are influenced both directly by temperature through physiological constraints (e.g., Rossaro 1991), and indirectly by catchment vegetational changes (Battarbee 2000). The conditions for studying climate changes and biotic responses in alpine regions seem therefore to be favourable. However, for Norway there is a great lack of data on alpine limnological responses to Holocene environmental changes. Based on a multiproxy approach, this study aims at reconstructing the Holocene environmental history and limnological response to vegetational change of Råtåsjøen, a low-alpine lake in south-central Norway. Subfossil pollen and plant macrofossils are used to characterise the catchment changes, and diatoms, chironomids, and sediment characteristics (organic and carbonate content and accumulation rates) are used to study the lake history.

Study area

Råtåsjøen (62°16′N and 9°50′E, Figure 1) is located in the Dovre Mountains, south-central Norway, 1169 m above sea level in the low-alpine vegetation zone above the treeline. The climate is continental (Aune 1999) with a relatively low annual precipitation (450 mm) and mean July and January temperatures of 8.6 and −14 °C, respectively (A. Odland pers. comm.). Glacio-lacustrine sediments that were deposited in large glacier-dammed lakes during deglaciation cover the region

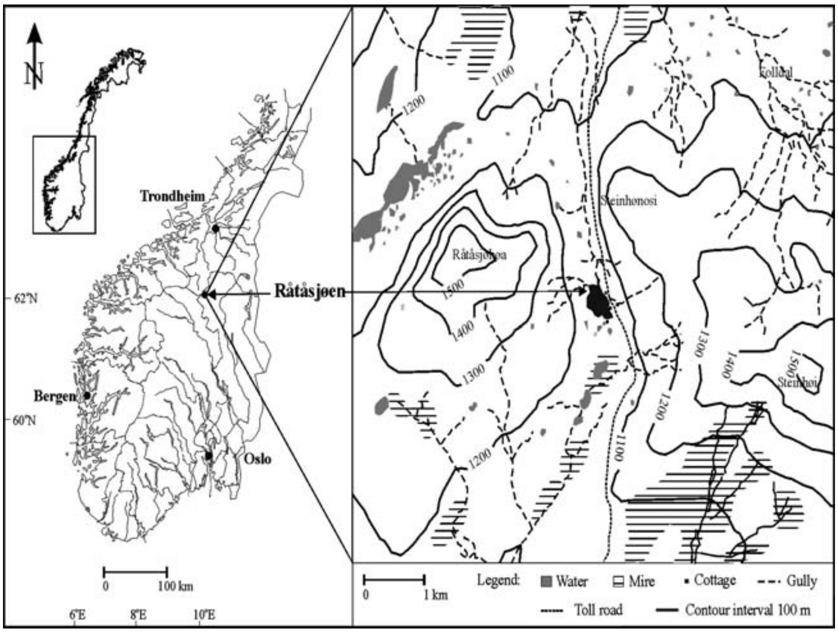

Figure 1. Map of Råtåsjøen located in the Dovre mountains, south-central Norway.

near Råtåsjøen. The chronology and processes of deglaciation in this part of central Norway are poorly understood at present due to a lack of reliable radiocarbon dates relating to deglaciation (Nesje 2002). Lichens and drought-tolerant higher plants dominate the catchment vegetation, reflecting the low annual precipitation, the well-drained glacio-lacustrine sediments, and the relatively high pH of the soils (Dahl 1956; Moen 1999). The lake is 500 m long and 250 m wide with bedrock thresholds, and has a maximum water depth of 8.3 m. The lake is oligo-mesotrophic (total phosphorus = 11.8 μg l^{-1}, $n = 2$) with a lake-water pH of 7.1 ($n = 2$), and a low organic content (TOC = 1.9 mg l^{-1}, $n = 2$). No permanent streams enter the lake, but there is an outflow draining towards the north. There is a small fishing hut on the western side of the lake constructed in the early 1980s. There is no evidence for disturbance to the lake from this hut or its construction.

Methods

Sediment sampling and analyses

Four series of overlapping sediment cores were retrieved from the deepest part (8 m) of the lake using a 5 cm diameter modified Livingstone piston corer (Wright 1967). The cores were wrapped in cling-film, aluminium foil, and heavy polythene and stored in the dark at 4 °C until sub-sampled. Series 3, extending from 801 to 998 cm below the water surface, was chosen for analyses because it provided the most complete and least disturbed sedimentary sequence. The upper 44 cm of unconsolidated sediment was retrieved using a 7 cm diameter Perspex tube sampler (Wright 1980) and extruded in the field. From the Perspex tube, section 801–821 cm was extruded vertically while the section 821–845 cm was solid enough to be extruded horizontally. These cores were sliced into 1 cm intervals and stored in plastic bags. This provided a master sequence of 197 cm (801–998 cm) from which pollen, diatoms, and chironomids were analysed. A 233 cm continuous core was used for plant macrofossil analysis. This core was retrieved from 800 cm water depth, using a 110 mm diameter piston corer (Nesje 1992), stored at 4 °C until opened, and cut into 0.5 cm slices. The master sequence (series 3) of 197 cm, the 233 cm continuous core, and the Perspex tube core were all correlated by sequence slotting (Birks and Gordon 1985; Thompson and Clark 1989) of loss-on-ignition (LOI) profiles aided by visual inspection.

For analysis of LOI, volumetric 0.5 cm^3 samples were taken at 1 cm intervals according to the methods of Heiri et al. (2001). LOI at 550 °C (organic content) and 950 °C (carbonate content) were calculated as percentages of the sediment dry weight and as accumulation rates per unit time (g cm^{-2} yr^{-1}). Bulk density was calculated as fresh weight divided by volume. Sediment lithology was described following (Troels-Smith 1955).

Water sampling and analyses

Water samples for pH, alkalinity, conductivity, and total organic carbon (TOC) were collected twice, in August 1998 and September 2001 from the surface at the centre of the lake. Samples were stored for about six weeks in the dark at 4 °C before being analysed at the Freshwater Fisheries Laboratory at Faskally, Perthshire, Scotland. Water samples for total phosphorus analyses were collected in August 2000 and 2001. The samples were frozen as soon as possible and sent to the Centre for Ecology and Hydrology in Edinburgh, Scotland, for analysis.

Chironomid analyses

Sediments of 1 cm slices at 2–4 cm intervals from 76 levels were sieved through a 100 μm mesh and chironomid larval head capsules were sorted from the residue under 30–50× magnification. All head capsules retrieved were transferred to pure alcohol and mounted in Canada balsam prior to identification. Fifty chironomid head capsules are often considered enough to give stable chironomid-inferred temperatures (Heiri and Lotter 2001; Larocque 2001; Quinlan and Smol 2001). In the present study, on average 94 specimens were identified from each level with the aid of Rieradevall and Brooks (2001), Sæther (1976), and Wiederholm (1983) and by comparison with the chironomid reference collection at the Museum of Zoology, Bergen, and the identifications in the modern Norwegian chironomid calibration set. This calibration set relates the chironomid distribution and abundance in 157 Norwegian and Svalbard lakes to mean July air temperature (July T; Table 1) (Brooks and Birks 2000, 2001, unpublished).

Diatom analyses

Sediment samples of 1 cm were prepared from 62 slices at 2–4 cm intervals. Diatom preparation and concentration estimation followed the method of Battarbee (1986) and Battarbee and Kneen (1982) where sediment samples were treated with 30% hydrogen peroxide and a known concentration of plastic microspheres was added before the valves were mounted on cover slips in Naphrax®. About 500 valves were counted per level. Diatoms were identified using several floras (e.g., Cleve-Euler 1951–1955; Patrick and Reimer 1966, 1975; Krammer and Lange-Bertalot 1986–1991). *Fragilaria pinnata* < 10 μm was separated from the nominate. The fine striated and delicate *Fragilaria construens* var. *venter* in Camburn and Charles (2000: Plate 7, Figures 20 and 21) was separated from the coarser variety. Both taxa were amalgamated with the nominate for pH and total phosphorus (TP) reconstructions, and for analogue matching. Taxonomy and nomenclature were harmonised to the Surface Water Acidification Programme data set (SWAP: Stevenson et al. 1991) for pH reconstruction, as the AL:PE calibration set used also follows the SWAP conventions (Cameron et al. 1999). In the absence of a calibration set for high-altitude Scandinavian lakes, a Swiss calibration set (Lotter et al. 1998) was used for TP reconstruction. This calibration set is available at the European Diatom Database

Table 1. Performance statistics for the calibration data sets employed, assesses by leave-one-out cross validation

Calibration set	Reference	R^2	RMSEP	Maximum bias
Chironomid July T	Brooks and Birks, unpub.	0.91	1.01	0.93
Diatom pH	Cameron et al. (1999)	0.82	0.33	0.74
Diatom TP	Lotter et al. (1998)	0.79*	0.19*	0.33*

* Based on log 10 transformed values.

(EDDI) web page. The SWAP taxonomy conventions were used and taxon codes were changed where necessary when applying the Swiss calibration set. See Table 1 for calibration data set details. The diatom-pH and diatom-TP calibration data sets were chosen because these contain high-altitude lakes with low DOC levels and a median pH of about 6.1, resembling Råtåsjøen. No attempts were made to infer air or water temperature from the fossil diatom assemblages because of the absence of an available calibration set for high-altitude lakes in northern Europe.

Pollen analyses

Samples of 0.5 cm^3 sediment were taken at 2 or 4 cm intervals from 68 levels using a small calibrated brass sampler (Birks 1976) and were prepared using method B of Berglund and Ralska-Jasiewiczowa (1986). The residues were suspended in 2000 cS silicone oil. Weighed tablets of a known concentration (182 grains mg^{-1}) of *Eucalyptus* pollen (batch 903722, Maher 1980) were added to each sample prior to preparation so that fossil pollen concentrations and accumulation rates could be estimated. Identification and counting were at a magnification of 400× with critical determinations at 1000×. Identifications were made with the aid of an extensive modern pollen reference collection, and keys of Birks (1973a), Moore et al. (1991), and Punt et al. (1976–1995). A minimum of 500 determinable terrestrial pollen and spores were counted from each sample.

Plant macrofossil analyses

Using the pollen diagram as a guideline, plant macrofossils were picked from 78 slices of 0.5 cm sediments in the 11 cm diameter continuous core. Contiguous slices were sampled when one slice did not provide enough plant macrofossils for AMS-dating, otherwise the slices were sampled at 10 or 20 cm intervals. The 20 cm intervals were used where there seemed to be little change in sediment or macrofossil composition. The sediments of each sample were submerged in water for volumetric measurement and carefully washed through a 125 μm mesh. To aid deflocculation of humified material, 10% KOH was added for a few hours and the sample washed in water. Plant macrofossils were picked out from the residue, using a stereomicroscope with a magnification of 12–16×. Magnification of 40–66× was used for identification or 100× when a more detailed examination was necessary. Numbers of macrofossils were calculated for 50 cm^3 fresh sediments. Plant nomenclature for pollen, spore, and macrofossil taxa follows (Lid et al. 1994).

AMS-radiocarbon dating

Terrestrial plant macrofossils were extracted for radiocarbon dating from eight levels in the 11 cm diameter core (Table 2). After being washed in water, small pieces of sand or other potential contaminants were removed and the macrofossil samples were air-dried to constant weight. Minimum weight for each dried sample was set to 10 mg. A minimum of 4 contiguous samples was

Table 2. AMS-dates from Råtåsjøen

Type of plant macrofossils	Laboratory number	Sample depth (cm)	Weight (mg)	^{14}C age BP ($\pm 1\sigma$)	^{14}C age cal. BP ($\pm 1\sigma$)
1, 2, 3, 4, 5, 6, 7, 8, 9, 10	Beta-154530	816.0–819.0	10.9	1700 ± 40	1580 ± 40
1, 2, 4, 5, 8, 11	Beta-154531	855.5–858.5	12.0	2780 ± 50	2870 ± 70
1, 2, 3, 4, 5, 8, 9, 12	Beta-154532	889.0–893.0	11.1	3880 ± 50	4290 ± 85
1, 2, 4, 5	Beta-154534	939.0–942.0	10.4	6390 ± 50	7300 ± 35
1, 2, 5, 9, 11, 14, 16	Beta-154535	961.5–963.5	10.1	8000 ± 60	8990 ± 125
2, 8, 10, 11, 14, 16	Beta-149272	973.0–975.0	10.5	8710 ± 40	9680 ± 75
2, 8, 9, 10, 13, 18	Beta-154536	976.0–977.0	11.8	9220 ± 60	10390 ± 120
1, 8, 9, 11, 14, 16, 17, 18, 19	Beta-149271	980.5–981.0	10.1	9380 ± 80	10570 ± 90

Macrofossil types: 1. *Betula nana*, 2. *Betula*, 3. *Bistorta vivipara*, 4. *Selaginella selaginoides*, 5. *Juniperus communis*, 6. *Cassiope hypnoides*, 7. *Phyllodoce caerulea*, 8. *Empetrum nigrum*, 9. *Salix*, 10. *Salix herbacea*, 11. *Dryas octopetala*, 12. Poaceae, 13. *Carex* cf. *rostrata*, 14. Charcoal, 15. *Rumex*, 16. Vaccinium-type, 17. Twig fragment, 18. Carbonised fragments, 19. *Saxifraga oppositifolia.*

needed to provide enough material (Table 2). The dried samples were placed in sterilised (heated to 600 °C for 3 min) glass vials and sent to Beta Analytic Inc. for dating by accelerator mass spectrometry (AMS). The radiocarbon dates were calibrated into calendar years before present (cal BP) using CALIB 4.3 (Stuiver and Reimer 1993), method A, and the INTCAL98 bi-decadal calibration data set (Stuiver et al. 1998). Age-depth modelling from the calibrated dates was done by semi-parametric mixed-effect weighted regression within the framework of generalised additive modelling (Seppä and Birks 2001; Heegaard et al. 2004). Approximate ages for the basal 20 cm of the core were estimated by linear extrapolation of the fitted regression model (Figure 2).

Numerical analyses

Relative abundance, as opposed to absolute abundance, was used in all numerical analyses to preclude potential bias caused by changes in sedimentation rate and processes such as focussing (Birks and Gordon 1985). Pollen, diatoms, chironomids, and sediment variables were divided into local assemblage zones using optimal sum of squares partitioning (Birks and Gordon 1985) in the software ZONE (Juggins 1991) and the number of statistically significant zones was assessed by comparison with the broken-stick model (Bennett 1996).

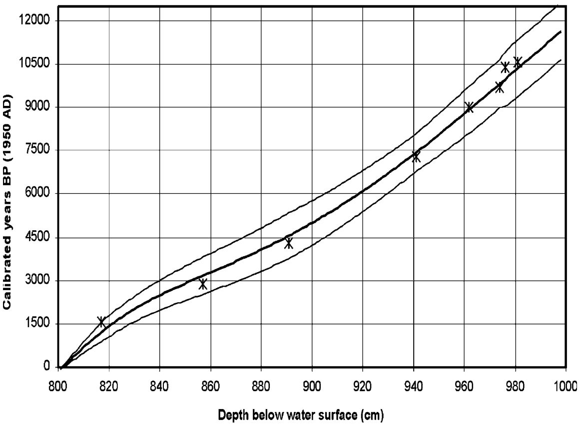

Figure 2. Age-depth model of the sediment sequence based on weighted generalised additive mixed-effect regression of eight calibrated AMS ^{14}C dates from terrestrial plant macrofossils. The thin lines are the 95% confidence intervals for the fitted model (thick line).

July T reconstructions from the fossil chironomid assemblages and pH and TP reconstructions from the fossil diatom assemblages were made by weighted averaging (WA) type calibration methods using transfer functions derived from the calibration sets described above. The transfer functions were established using the unimodal-based technique of weighted averaging partial least squares regression (WA-PLS: ter Braak and Juggins 1993) and the software WAPLS (Juggins and ter Braak 1995). Sample specific prediction errors were estimated by Monte Carlo simulation (Birks 1995). A LOESS smoother (Cleveland 1993) with span 0.18 (2nd order) was fitted through the chironomid-inferred July T and diatom-inferred pH to highlight the major long-term trends using S-Plus. The chironomid-inferred temperatures are not corrected for glacio-isostatic land uplift since the temperatures are discussed in relation to the corresponding environment and vegetation. Analogue matching (Overpeck et al. 1985) using squared chord distance as a measure of dissimilarity and the software MAT (Juggins 1995) with a 5% cut level of the modern data as the criterion for good analogues was used to assess how well the subfossil diatom and chironomid species assemblages are represented in the modern calibration sets. The assemblage diversity was estimated as the effective number of taxa in a sample (Hill's N2: Hill 1973).

Results and discussion

Results

The age-depth model is shown in Figure 2. The most abundant and/or ecologically informative taxa of chironomids are shown in Figure 3, diatoms in Figure 4, and pollen, spores, and macrofossils in Figure 5. Results from the quantitative and sedimentological analyses are summarised in Figure 6 and in the environmental synthesis (Figure 7). The chironomid-inferred temperatures together with the diatom-inferred pH are shown in Figure 8, with non-analogue taxon assemblages and environments marked. The chironomid stratigraphy is divided into 6 statistically significant zones, the diatoms into 4 zones, pollen into 4 zones, and the sediment characteristics into 5 zones (Figure 6). The use of phases (Phase I–VIII)

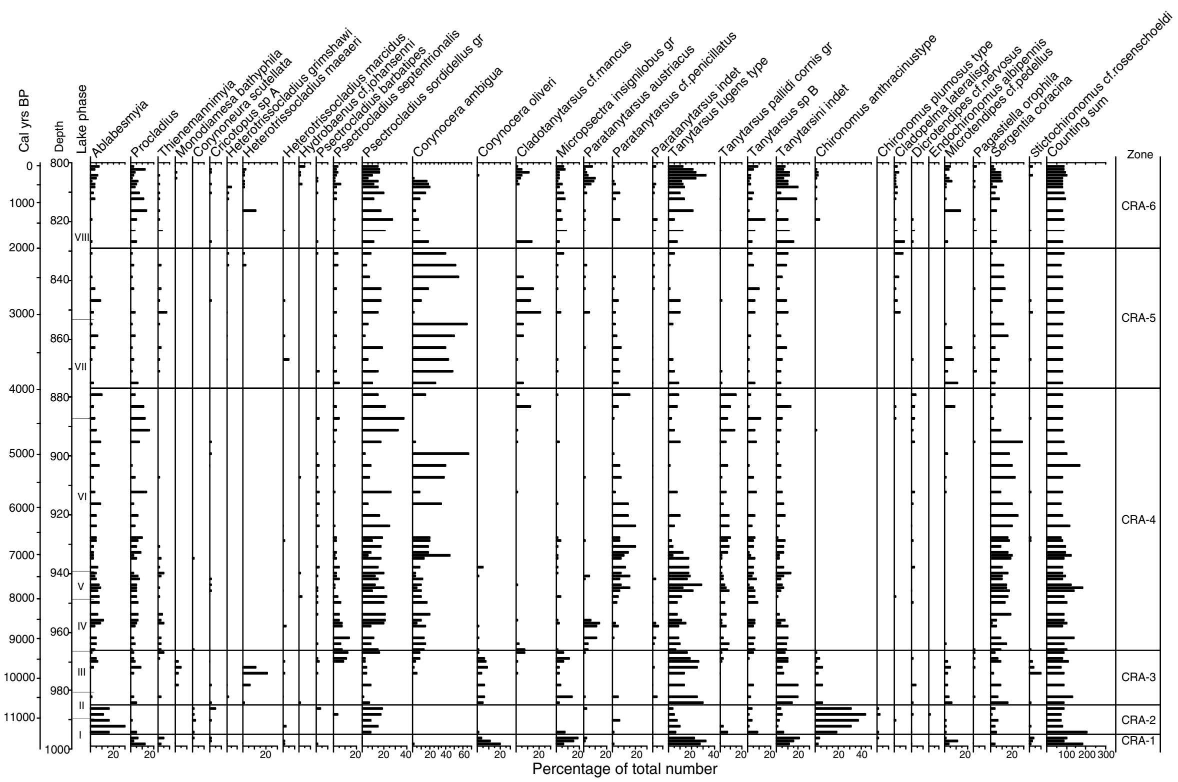

Figure 3. Selected chironomid taxa from Råtåsjøen, expressed as percentages of the total number determinable head capsules. Phases follow the environmental outline in the text.

in the environmental history represents a quantitative interpretation of the environment based on all proxies such that the name of the phase reflects milestones in the prevailing conditions. These phases are not drawn from the proxy zonations. The phases are shown in Figures 3–8. The full pollen, diatom, chironomid, and plant macrofossil data are available on request to the authors and will be deposited in the national NORPAST database.

Environmental history

Deglaciation

According to the age-depth model, the Råtåsjøen basin became open for sedimentation after deglaciation at ca.11,600 ± 1000 cal BP. Till was left at the base of the lake, and glacio-lacustrine sediments were deposited as the glacier retreated. The age of the onset of lake sedimentation seems slightly older than compared to the Younger Dryas-Holocene boundary at Kråkenes lake, about 230 km west of Råtåsjøen, which was dated to 11,530 (+40–60) cal BP (Gulliksen et al. 1998). However, this difference is well covered by the error estimate of ± 1000 cal years at this depth at Råtåsjøen.

Phase I: Pioneer assemblages (ca. 11,600–11,000 cal BP)

Birch and pine dominated the pollen percentage (Figure 5). As the pollen accumulation rate was very low, these pollen were most likely blown in from distant low-altitude sites (Aario 1940; Birks 1973b; Birks and Birks 2000). Pioneer and alpine taxa dominated the scattered and sparse local vegetation, as indicated by the pollen record (Figure 5). These included *Salix*, *Dryas octopetala*, *Saxifraga oppositifolia* type, and *Oxyria digyna*, pioneer taxa indicative of open, base-rich soils.

Inorganic silt was deposited in the lake basin by streams, slope-wash, and/or solifluction, at a time when erosion rates were high and water velocities were sufficient to carry relatively dense materials into the centre of the lake (Figures 6 and 7). These

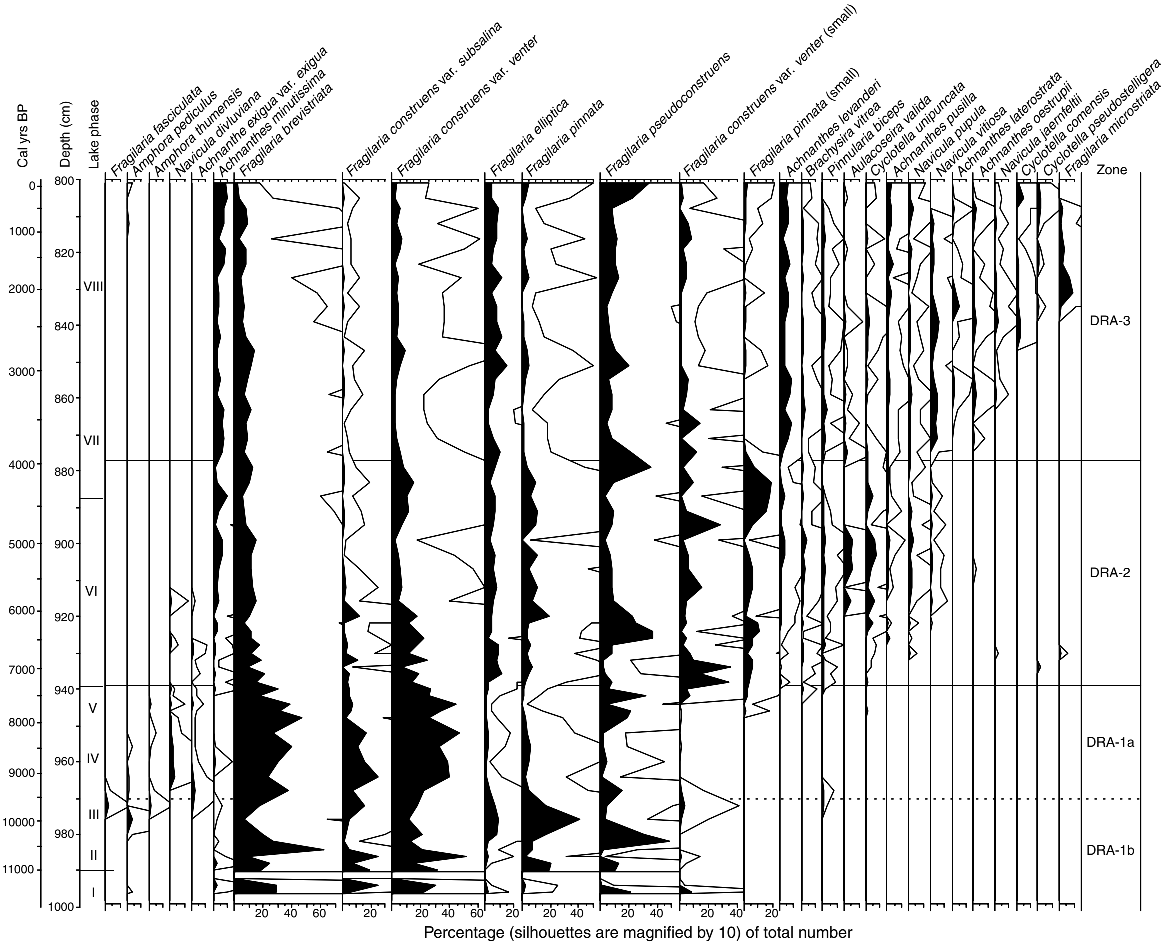

Figure 4. Selected diatom taxa from Råtåsjøen, expressed as percentages of the total number determinable valves. Shaded silhouettes are magnified 10×. Phases follow the environmental outline in the text.

deposits may include an unknown quantity of eolian silts originating from glacial outwash deposits (Birks 1976; Jacobson and Birks 1980). The organic accumulation was low and the carbonate accumulation high (Figure 6). The aquatic flora other than diatoms (*Chara* and *Nitella)* and fauna (*Chironomidae*, *Daphnia*, *Plumatella)* appeared quickly after deglaciation (Figure 7). The high carbonate content is correlated with the presence of *Chara* in the lake. Charophytes utilise carbonates at high pH during photosynthesis, leading to the formation of calcium carbonate. *Chara* spp. are thus indicative of high pH and also low productivity (e.g., Haas 1994). The large proportion of silt indicates that the surrounding soils were immature and unstable. Silt entering in spring following snowmelt, which is a critical time for diatom blooms, may have created turbid conditions that restricted the development of the diatom flora due to the lack of light (e.g., Stoermer and Smol 1999). The diatoms have a low accumulation rate, but a relatively diverse flora (Figures 4 and 6). There were no indications of poor diatom valve preservation. The diatom taxa present are characteristic of oligotrophic lakes with relatively high pH-values (Stevenson et al. 1991; Cameron et al. 1999; Rosén et al. 2000). These taxa included *Achnanthes minutissima* and different *Fragilaria* taxa (Figure 4). The *Fragilaria* taxa varied in relative abundance. However, their ecological demands other than pH are poorly known. Today, analogues to the late-glacial and early Holocene *Fragilaria* assemblages can be found in the moats of high Arctic alkaline ice-covered lakes

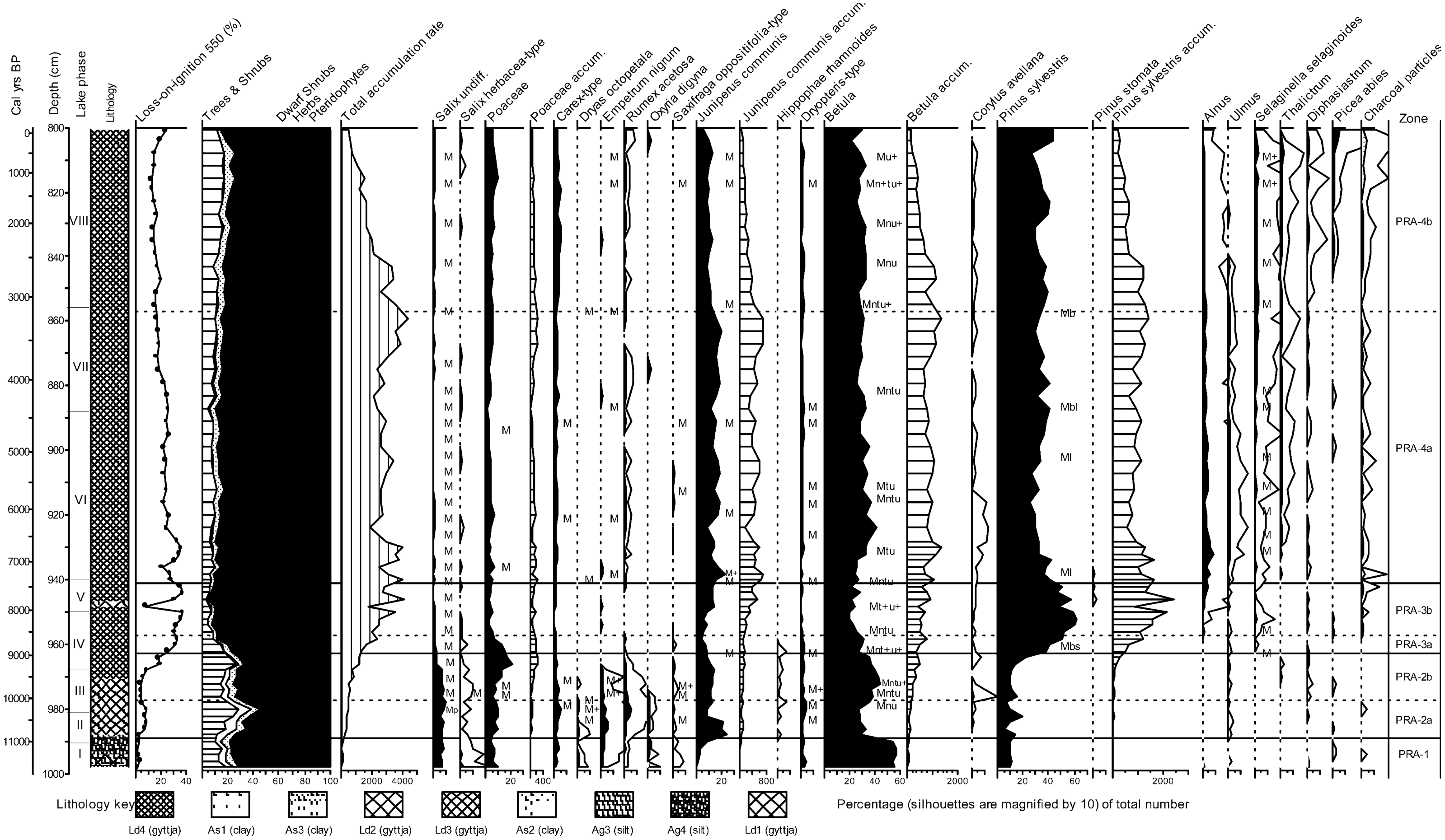

Figure 5. Selected pollen, spore, and plant macrofossil taxa from Råtåsjøen. Pollen and spores are expressed as percentages of the total number of determinable terrestrial taxa. Unshaded silhouettes are magnified 10×. Curves with depth bars are pollen accumulation rates (grains cm^2 yr^{-1}). Letters in diagram denote plant macrofossils. Where the letter is followed by ‘+’ there are 10 or more macrofossils. accum. = accumulation rate. Macrofossil notes: *Saxifraga* are placed by the pollen of *S. oppositifolia*-type, but contains several species. Fern-sporangia are placed by *Dryopteris*-type spores. Cyperaceae is identified to genus level. *Salix* contains branches, buds, budscales, capsules, cuticles, and twigs. *Empetrum*, Poaceae, and *Saxifraga* are seeds. M = macrofossil, n = *Betula nana*, p = *Salix polaris* (leaf), b = bark, l = needle, s = seed, t = tree-type, u = undiff. Phases follow the environmental outline in the text.

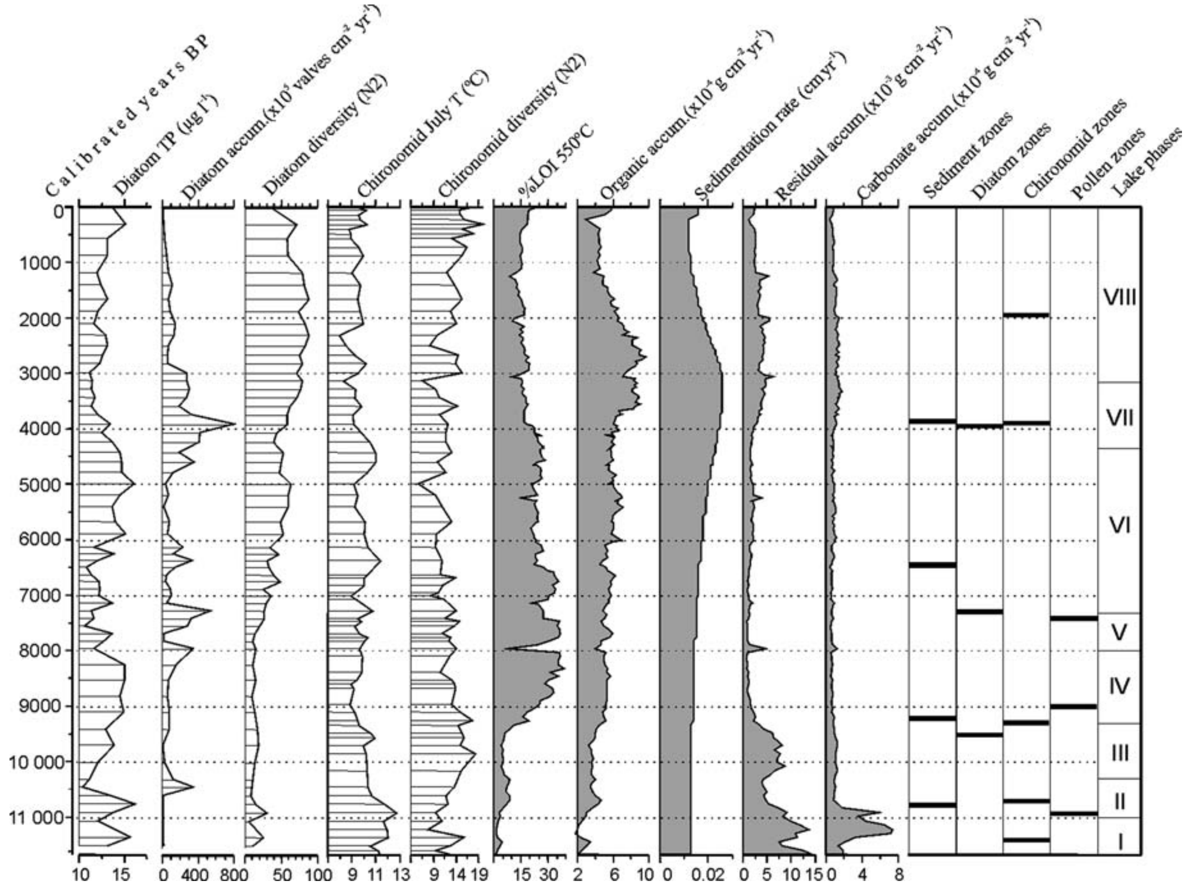

Figure 6. Diatom-inferred total phosphorus (μg l^{-1}), diatom accumulation rates (× 10^5 valves cm^{-2} yr^{-1}), diatom diversity estimated as Hill's N2, chironomid-inferred mean July air temperature (July T, °C), chironomid diversity estimated as Hill's N2, sedimentation rate (cm^2 yr^{-1}), organic content of sediments as percentage loss-on-ignition (LOI 550 °C), organic accumulation rate (×10^{-4} g cm^{-2} yr^{-1}), residual minerogenic accumulation rate (× 10^{-3} g cm^{-2} yr^{-1}), carbonate influx (×10^{-4} g cm^{-2} yr^{-1}), and statistically significant zones for sediment variables, diatoms, chironomids, and pollen. Phases follow the environmental outline in the text. accum. = accumulation rate.

(Smol 1988), which may be a good analogue for Råtåsjøen at this time.

A relatively rich chironomid fauna was established at the beginning of phase I with *Tanytarsus lugens* type, *Corynocera oliveri*, *Procladius*, *Micropsectra insignilobus* gr., and 15 additional taxa (Figures 3 and 6). These taxa suggest well-oxygenated conditions (e.g., Brundin 1949) in line with the unstable soils and enhanced runoff (Figure 7). The chironomid-inferred July T was 11.2 °C, but it may not be reliable as the fossil assemblages lack adequate modern analogues in the calibration set (Figure 8). The pioneer fauna was soon replaced by *Chironomus anthracinus* type. This event resulted in warmer inferred July T (11.8 °C) and a substantial shift in the chironomid assemblage at 11,200 cal BP. Seemingly, the warm inferred July T contradicts the terrestrial pollen and macrofossil records since these show an abundance of *Salix herbacea*, other vascular plants, and bryophytes all typical of open treeless vegetation, cool alpine regions, and unstable ground. The inferred temperatures were mainly influenced by *C. anthracinus* type with up to 44% abundance and a WA-PLS temperature 'optimum' (beta coefficient) of 14.6 °C in the modern Norwegian calibration set. *C. anthracinus* type includes at least three species in alpine regions of mainland Norway (Schnell and Aagaard 1996). The particular specimens found in Råtåsjøen at the time may be a single species adapted to cooler conditions, while the calibration set may include several

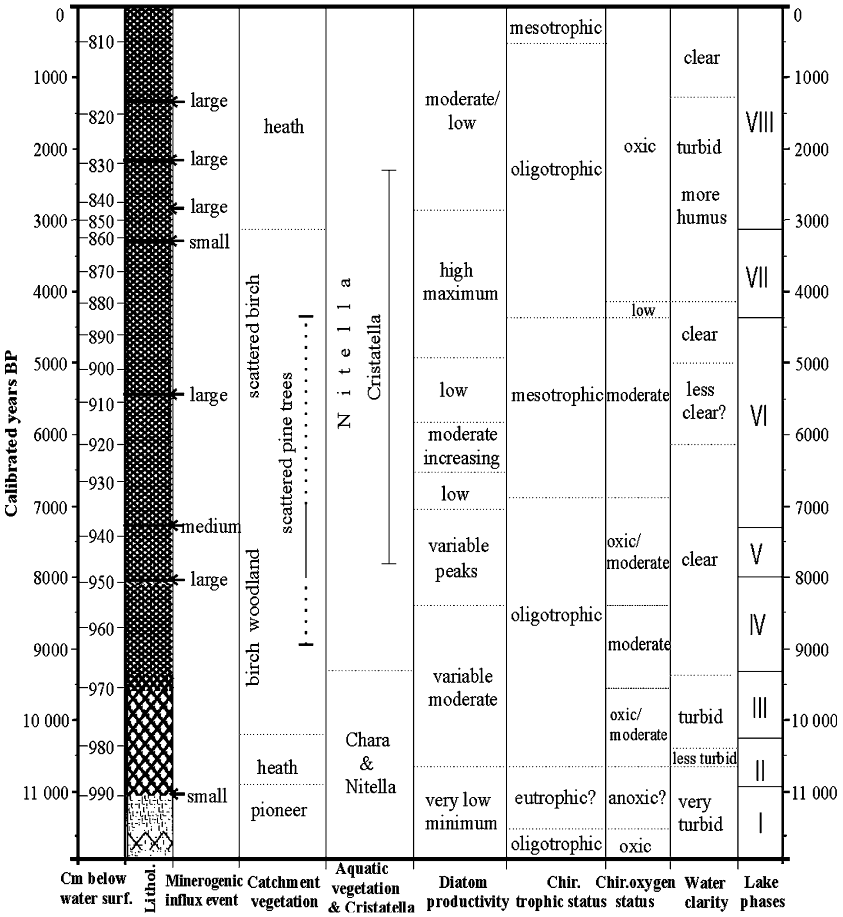

Figure 7. Synthesis of events based on the biota and sediment characteristics. Phases follow the Environmental outline in the text. For lithology key, see Figure 5. Chir. = Chironomid, Lithol. = Lithology.

species, possibly adapted to warmer conditions. This would result in over-estimated temperatures at Råtåsjøen. However, *C. anthracinus* and the genus *Chironomus* are, in general, associated with high temperatures, high production, and/or low oxygen (Wiederholm and Erikson 1979; Johnson and Pejler 1987; Brodersen and Anderson 2002). This suggests a high production in Råtåsjøen at the time, but the concurrent presence of *Chara* in shallow waters, the diatom assemblages, and the low accumulation of organic matter indicate low production. The taxon *M. insignilobus* gr. disappeared as *C. anthracinus* type increased in abundance. *M. insignilobus* gr. is often associated with an oxygen-rich hypolimnion while *C. anthracinus* can tolerate prolonged periods of anoxia (Clerk et al. 2000). *C. anthracinus* is reported to colonise more quickly than other chironomid taxa and benefit from a lack of competition in what may be sub-optimal temperatures for the genus (Brooks 1997). This indicates that the taxon may thrive under sub-optimal temperature conditions as long as competition is low. Therefore, this may either have been a time of high temperatures or oxygen deficiency. The cause for the anoxic conditions may have been prolonged ice-cover leading to temporary stagnation. Until further evidence is available, we cannot draw any firm conclusions whether this was a time of high temperatures or a time of lake anoxia.

Phase II: Salix dominance (ca. 11,000–10,300 cal BP)

Between 11,000 and 10,900 cal BP the pollen assemblage changes, with a sudden decrease in

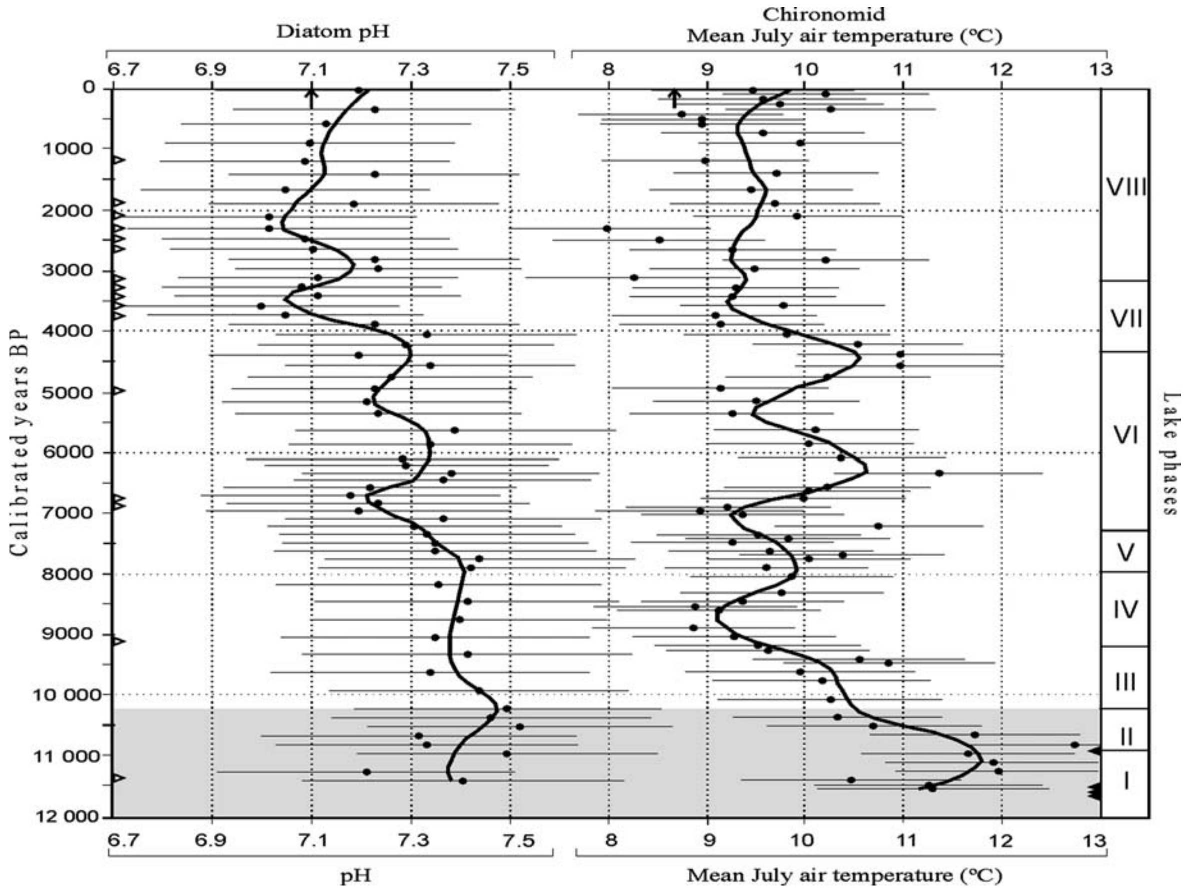

Figure 8. Diatom-inferred pH and chironomid-inferred mean July air temperature with sample-specific prediction errors and LOESS smoother (span 0.18, 2nd order) fitted through the data points. Triangles along the *Y*-axes indicate presence of non-analogue taxon assemblages, open triangles for diatoms and filled triangles for chironomids. Shaded area indicates the occurrence of non-analogue environment. Arrows indicate present-day values for temperature and pH. Phases follow the environmental outline in the text.

birch percentages and an increase in juniper (*Juniperus communis*) (Figure 5). Since lower altitudes may have been ice-free for a longer period of time (Nesje 2002), it is likely that the birch and pine pollen were transported from lower elevations (cf., Birks 1973b). This interpretation is supported by the absence of tree macrofossils (Birks and Birks 2000). The local catchment pollen production was probably very low. Both the sediment characteristics and chironomid assemblage changed around 10,700 cal BP (Figure 6). The sediment changed as the organic content increased and the carbonate content decreased. The fauna changed as *C. anthracinus* type was replaced by more oxygen-demanding and cold-tolerant taxa. Hence, the chironomid-inferred July T quickly decreased to 10.4 °C (Figure 8).

In phase II, the diatom-inferred pH was high and varied from 7.40 to 7.15 (Figure 8), and *Chara* was present (Figure 7). Modern *Chara*-lakes from central Europe have pH values between 7.0 and 8.4 (Haas 1994). In the AL:PE diatom calibration set, pH values range from 4.5 to 8.0, but there are only five lakes with pH above 7.3. Hence, high pH values are underestimated in a leave-one-out jack-knife cross-validation (Cameron et al. 1999). The lake diatom composition did not change significantly with the introduction or disappearance of *Chara*, and the inferred pH does not show higher pH values when *Chara* is present. Instead, the variations in inferred pH may have been caused by the great fluctuations in *Fragilaria* taxa (Figure 4). These taxa have WA pH-optima varying between 6.7 and 7.1 (*F. construens* var. *venter*, *F. pinnata*) in the AL:PE calibration set. Although good modern analogues for the diatom assemblages exist, the calibration set lacks *Chara*-lakes and high pH-lakes, suggesting that the diatom-inferred pH may not be reliable, and most likely is underestimated. The peak in the diatom valve accumulation rate (Figure 6) may be caused by less turbid conditions (Figure 7) that were more favourable for increased diatom growth, or it could be due to an overall longer growing season.

Phase III: Birch immigration and establishment (ca. 10,300–9300 cal BP)

At around 10,000 cal BP, dwarf-birch macrofossils are recorded (Figure 5), and at the same time the open-ground plant species were at a maximum abundance. Tree-birch (*Betula pubescens sensu lato*) macrofossils appeared soon after dwarf-birch, and the increase in the concentration of macrofossils suggests a rapid development of open sub-alpine birch forest around the lake. *Dryas octopetala* decreased, most likely as a result of shading (Chapin et al. 1994). The overall increased pollen accumulation rate suggests a more productive and denser vegetation cover (Figure 5). On drier ridges, juniper and *Empetrum nigrum* probably dominated. In the macrofossil record, the occurrences of *Potentilla palustris* seeds and different species of *Juncus* suggest some development of mires around the lake (e.g., Dahl 1956). Most likely as a consequence of the increased terrestrial vegetation cover, the soils became stabilised and the organic content of the sediments increased (Figure 6). Following the immigration of birch, the chironomid species diversity reached a maximum (Figure 6). The presence of trees in the catchment would have resulted in the input of leaves, etc. in the lake, and in addition there was an increased abundance of macrophytes in the lake. This may have presented a wider range of available niches for the immature stages of the chironomids (Coffman 1995), which could have increased the chironomid diversity. The chironomid-inferred July T gradually decreased (Figure 8).

From ca. 10,000 cal BP, the diatom-inferred pH started to decrease (Figure 8). The pH decrease may have been caused by a reduced supply of calcium carbonate as this was buried in the sediments and was no longer recycled. In addition, catchment stabilisation meant that the supply was not maintained by inwash of detrital carbonate or dissolved calcium carbonate (Mackereth 1965). This small acidification could also be explained as the first step in a long-term acidification trend following lake formation, or due to the decreasing temperature (see section on diatoms below). At ca. 10,000 cal BP, the same *Fragilaria* taxa existed as before, but other less abundant genera decreased (Figure 4). *Fragilaria* taxa are good competitors as they can divide rapidly and probably out-compete other taxa during periods of higher diatom accumulation. This is expressed by the general low diatom diversity (Figure 6). From ca. 9500 cal BP, the dominant *Fragilaria* taxa stabilised in abundance. There is evidence for increased conductivity between about 10,500 and 7500 cal BP from the presence of taxa such as *Achnanthes exigua* var. *exigua*, *Amphora pediculus*, *Amphora thumensis*, *Fragilaria fasciculata*, and *Navicula divluviana*. This could be linked with the increased input of organic acids produced as birch established, and organic soils humus developed facilitating the availability of cations to the lake system (Ford 1990).

Phase IV: Cool conditions (ca. 9300–8200 cal BP)

After ca. 9300 cal BP, the inferred temperature decreased more rapidly and July T reached a minimum just above 9 °C at ca. 8700 cal BP (Figure 8). Both the chironomid assemblage and sediment characteristics changed at ca. 9300 cal BP. The change in chironomid assemblage may reflect a climatic cooling, as indicated by the increase in *Paratanytarsus austriacus*, *Corynocera ambigua*, and *T. lugens* type (Figure 3). In the Norwegian calibration set, these taxa have WA-PLS temperature 'optima' (beta coefficients) of 2.3, 4.3, and 4.4 °C, respectively (Birks and Brooks 2001, unpublished). Several taxa disappeared at the zone boundary (e.g., *Microtendipes* cf. *pedellus*, *C. oliveri*, *M. insignilobus* gr., and *C. anthracinus* type), and these did not only include warm-adapted taxa. The chironomid zone change may therefore also be directly related to changes in sediment organic matter (Olander et al. 1999; Dimitriadis and Cranston 2001), or be caused by habitat instability in the changing sediment regime. From 8700 cal BP, the inferred July T starts to increase. The inferred pH remains stable at 7.4 throughout the phase.

By ca. 8900 cal BP, the relative percentage and accumulation rate of organic matter had increased rapidly to values similar to present-day conditions (Figure 6). As the vegetation-cover increased, the catchment hydrology may have involved more evapo-transpiration than direct run-off (Engstrom et al. 2000). At about 9000 cal BP, the pollen record changed again. This time, plants typical of low-alpine areas (*Salix*, Poaceae, *E. nigrum*, *Rumex acetosa*, *Hippophae rhamnoides*) was replaced by high percentage and accumulation rates of pine. However, the macrofossil record shows only one pine bark fragment and part of a seed around 8900 cal BP suggesting that pine was not locally present, but was probably advancing altitudinally (Figure 5) (Eide 2003). Since *H. rhamnoides* in general is shade-intolerant and a poor competitor, its decrease was probably due to competition by shading with tree-birch (Iversen 1954; Birks 1973a).

Phase V: Pine immigration (ca. 8200–7300 cal BP)

From 8200 cal BP, high accumulation rates of pine pollen, some pine stomata, and a few fragments of pine bark and needles are recorded, suggesting the local occurrence of pine trees (Figure 5). However, it is important to note that small macroscopic fragments like this can be windblown from lower altitudes, especially over snow and ice. So, it is likely that pine was at or near its uppermost limit close to Råtåsjøen (Eide 2003). Around 7500 cal BP *Potamogeton* was recorded in the lake and *Cristatella mucedo* statoblasts, often found in lakes surrounded by trees up to about 1000 m in Norway (Birks, unpublished data), increased (Figure 7). There is a significant pollen zone boundary towards the end of this phase at ca. 7400 cal BP (Figure 6). This time, although a high accumulation rate of pine pollen is maintained, the percentage decreases, probably due to increases in other taxa (mostly *Betula*, but also *Alnus*, *J. communis*, and *Selaginella selaginoides*).

At the beginning of phase V, chironomid inferred July T steadily increased, and at *c.* 7900 cal BP the temperature reached a local maximum with temperatures just below 10 °C (Figure 8). Shortly after 8000 cal BP, some of the trends in inferred pH seem to follow the inferred temperature (see Figure 8 and section on diatoms below). At the end of phase V, diatom taxa indicating higher conductivity levels decreased or disappeared together with the dominant *Fragilaria* taxa (Figure 4). The decreased alkalinity and conductivity led to a shift in diatom composition. Diatom taxa with lower pH optima such as *Achnanthes levanderi*, *Brachysira vitrea*, and *Pinnularia biceps* increased or colonised the lake. The diatom diversity also increased, possibly explained by the shift towards less optimal conditions for *Fragilaria* taxa, which became less competitively dominant.

In phase V, the organic fraction of the sediments reached a plateau of about 35% and sedimentation rates increased, reflecting a higher within-lake production (Figure 6). However, at ca. 8000 cal BP, visible in the core as light bands in the organic gyttja (see lithology, Figure 7), there was a substantial short-time decrease in organic carbon accumulation rate and percentage as well as an increasing minerogenic accumulation (Figure 6). Coinciding with these events, there was an abrupt decrease in the accumulation rate of both pollen and diatom valves (Figures 5 and 6). This might be interpreted as an inwash of sediment-carrying melt water from the north-western slope (Figure 1), burying diatom valves and pollen quickly in fine silts, hence giving the impression of a lowered accumulation rate. Similar, but smaller-scale changes happened at irregular intervals throughout the Holocene (11,000, 6200, 4900, 3300, 2800, 2200, and 1400 cal BP; see minerogenic influx event, Figure 7)

Phase VI: Mid-Holocene temperature maximum (ca. 7300–4400 cal BP)

As reconstructed from the chironomid assemblages, this was a warm period at Råtåsjøen (Figure 8). The July T rapidly increased to 10.6 °C at ca. 6400 cal BP followed by a gradual drop of 1 °C towards ca. 5400 cal BP and a second increase culminating ca. 4500 cal BP, with a July T of 10.5 °C. The inferred temperature peaks are caused by an increase in *P. sordidellus* gr., the presence of warmth-indicating Chironomini like *Dicrotendipes* cf. *nervosus*, *Cladopelma lateralis*, and *Polypedilum* cf. *nubecolusum*, and a decrease in the cold indicators *T. lugens* type and *C. ambigua* (Figure 3). The latter cold-indicators had a general increase during the cooler period separating the two mid-Holocene temperature peaks. The organic content of the sediments was high at the time. The vegetation during phase VI around Råtåsjøen consisted of open birch forest with an understory dominated by abundant *B. nana* and juniper, and some *Empetrum* together with tall herbs represented by *R. acetosa* (Figure 5). Frequent *Salix* macrofossils suggest the occurrence of willow shrubs typical of wet areas in sub-alpine areas (Dahl 1956). Pine accounts for a large proportion of the pollen accumulation, however, the number of pine macrofossils is small (Figure 5). This suggests that pine was still at or near its uppermost limit, bur perhaps rare in the lake catchment. Even at the temperature maximum, the chironomid-inferred July T is still below the approximate thermal limit of pine (ca. 12 °C) in this part of Scandinavia (Eide 2003)

The diatom composition was unique during the mid-Holocene. The smaller species *F. pinnata* (small) (< 10 μm) and *F. construens* var. *venter* (small) characterized this time period and became important (Figure 4). The diatom accumulation rate increased during periods with warm summers (Figure 6), possibly due to increased growth rate and/or longer growing seasons, and decreased during periods with colder summer temperatures. Coinciding with the inferred increase in pH and valve accumulation rate from ca. 4500 cal BP, the diatom diversity dropped, probably as a consequence of increased competition. *Aulacoseira valida* was present between 5000 and 6000 cal BP. This large diatom requires turbulent conditions to remain in suspension in the water column (Smol 1988). This may indicate a time of windier summer conditions. The appearance of *A. valida* could partly be explained by the presence of trees in the Råtåsjøen catchment as it is a characteristic species in forest lakes in northern Finland (Weckström et al. 1997).

Phase VII: Climate cooling, tree decline and instability (ca. 4400–3100 cal BP)

After the warm period, the temperature decreased by almost 1.5–9.2 °C (Figure 8). In the macrofossil record, the last major pine fragments were found at ca. 4400 cal BP (Figure 5). Only one pine bark fragment was found after this time. This suggests that by 4000–4400 cal BP pine had receded from the catchment. From ca. 3200 cal BP the abundance of tree-birch macrofossils decreases (Figure 5) indicating that tree-birch also receded (Figure 5). As tree-birch retreated from the local catchment, sediment organic accumulation and sedimentation rates reached their maxima (Figure 6). Chemical weathering was likely dominant during the phase of denser vegetation (Engstrom and Wright 1984), whereas during phases of open vegetation, physical erosion prevailed, due to less stable upper soils that were no longer protected by tree roots (Mackereth 1965; Ford 1990; Kauppila and Salonen 1997). Interestingly, the organic carbon fraction of the sediments steadily decreased as the trees receded, according to the percentage LOI. However, according to the organic accumulation rate the input of organic carbon reached a maximum, most likely due to the high erosion of terrestrial organic material together with even more minerogenic material.

Concurrently, the diatom and chironomid species assemblages responded to the changing climate and catchment regime with shifts in species composition (Figures 3, 4 and 6). The diatom zone boundary at 3900 cal BP is drawn at a decrease in *Fragilaria* taxa and an increase or colonisation of *Achnanthes* and *Navicula* taxa (Figure 4). Among the colonisers, *A. laterostrata*, *A. pusilla*, and *N. vitiosa* have been found to be more common in alpine or arctic lakes than in lakes in the boreal forest (Rosén et al. 2000; Weckström et al. 1997). These observations support a temperature drop and vegetation change at this time. Following the decrease in the relative abundance of the competitive *Fragilaria*, the diatom diversity increased (Figure 6). Also, inferred pH reached a Holocene minimum (pH 6.8 at ca. 3500 cal BP; Figure 8). However, these values may be unreliable since the AL:PE calibration set lacks taxon assemblages analogous to the Råtåsjøen fossil record with a high abundance of small benthic taxa.

Phase VIII: Modern low-alpine conditions (ca. 3100 cal BP – present)

The landscape opened as tree-birch declined, and concurrently dwarf-birch and *Selaginella selaginoides* megaspores increased and present-day low-alpine heath and grassland vegetation gradually developed. The gradual vegetation change probably reflects a deterioration of climate with decreased growing season and expansion of snowbeds. A maximum pollen accumulation rate during the beginning of phase VIII (Figure 5) possibly reflects the landscape opening as the lake now was more exposed to far-distance transported pollen from trees. The decreasing accumulation rate of pollen towards the present (Figure 5) reflects the decreasing cover of the local vegetation and the increased distance to the treeline. Today, tree-birch grows at an altitude of 170 m below the lake, whereas pine trees grow 370 m below the lake (Eide 2003).

Sedimentation rates decreased through the phase, with 0.026 cm yr^{-1} at the beginning and 0.012 cm yr^{-1} at the top (Figure 6). This may reflect high erosion rates of organic material as the trees retreated, and the subsequent reduced erosion as the ecosystem productivity declined, as indicated by the decreasing organic accumulation rates (Figure 6). In addition, reduced sediment compaction and reduced sediment focusing could partially explain the lower sedimentation rate towards the top (Lehman 1975). As the deeper areas of the lake fill in, sediments are dispersed over a broader area, thereby producing an apparent drop in the sediment accumulation rate at the coring site.

The LOESS smoothed chironomid-inferred July T is generally stable in phase VIII, except for a minor temperature increase towards the present (Figure 8). However, both the inferred temperatures and the chironomid diversity were highly variable, with all major taxa fluctuating in abundance throughout the phase (Figure 3). In general, a more diverse fauna followed high inferred temperatures (Figure 6), which is similar to findings in North America by Palmer et al. (2002). The seemingly fluctuating temperatures may be caused by the effect of the in-filling of the lake-basin on the chironomid assemblage. Lake shallowing may lead to a weakening thermocline that might break up more easily during windy summers, causing a variation in bottom-water temperature. This effect was likely reinforced by wind exposure as the trees declined. The chironomid assemblage would then reflect changes in bottom-water temperature and oxygenation rather than changes in air temperature. At ca. 2050 cal BP there was a significant shift in the chironomid assemblage. This event occurred at a time of slowly increasing smoothed July T, but does not seem to be caused by a warming. There was an increase in the taxa *D.* cf. *nervosus*, *C. lateralis*, and *P. sordidellus* gr., (Figure 3). This indicates that this assemblage change and associated zone boundary may be caused by terrestrial plant material entering the lake, as these taxa are often associated with aquatic macrophytes and plant detritus (Hofmann 1984; Brodersen et al. 1998). After the zone boundary, the chironomid diversity increased (Figure 6). This may reflect a wider range of available niches within the lake. Towards the top of the core, the smoothed July T increased to 9.8 °C, while the inferred value at the top sample is 9.4 °C. The present temperature at Råtåsjøen estimated from nearby meteorological stations is lower (8.7 °C), but well within the sample specific error estimates of the chironomid-inferred July T (Figure 8).

The diatom accumulation rate was near its maximum in the transition to phase VIII and declined steadily towards the present (Figure 6). The diatom diversity reached a peak between ca. 2500–650 cal BP. In general, a longer growing season allows the development of more complex diatom communities with higher diversity (Douglas and Smol 1999). This is not evident at Råtåsjøen during this time period. The development of *Fragilaria* taxa during the warm phase VI may have limited the coexistence of other genera. Other genera colonised when the climate became cooler, the pH decreased, and *Fragilaria* taxa experienced less optimal conditions. The diatom-reconstructed pH of 7.0 from the surface sediment (Figure 8) is close to the measured lake-water pH of 7.1. The error estimates are stable through the reconstructed levels, varying from 0.35 pH units at the top of the sequence to 0.41 pH units towards the oldest sediment levels.

Calibration sets and non-analogues

During some time periods, both inferred pH and temperature may be unreliable, especially during non-analogue phases. Numerically, these phases represent a time when the fossil species assemblage lacks good modern analogues in the training set (Overpeck et al. 1985; Birks 1998). However, we should also be aware of non-analogue environmental and limnological phases (Figure 8). These phases occur when certain environmental conditions or combination of conditions that influence the biota occurred in the past, but such conditions or combinations are absent today from the modern calibration sets and from the study area (Jackson and Overpeck 2000; Jackson and Williams 2004). Around 10,500 cal BP in the present study, the presence of *Chara* suggesting cation leaching of the freshly exposed glacial soils, is an environmental feature not seen in any of the modern calibration-set lakes or in alpine Scandinavian lakes today. Such non-analogue environmental phases can represent a major methodological problem in the inference of past environments based on modern calibration sets (Birks 1998). The inferred environmental conditions for these time-periods should therefore be interpreted with care. This highlights the importance of not only sampling calibration-set lakes along a large gradient of a particular environmental factor, but also of sampling different types of lakes in different environmental settings and including several proxies. This will ensure that there is a wider spectrum of analogues in the calibration set and the environmental reconstructions may be more reliable. For example, it might be valuable to extend the search for sediments in lakes that are in the pioneer phase, or are recently deglaciated (e.g., Birks 1980; Engstrom et al. 2000). Sediments in such lakes formed during or after the 'Little Ice Age' in Scandinavia are rare and are absent or very poorly represented in existing Scandinavian calibration sets.

Climate history

The high chironomid diversity during the earliest Holocene indicates that a delayed succession is not influencing the temperature reconstruction. Still, the chironomid-inferred July T at Råtåsjøen starts with high and maybe unreliable temperatures due to the lack of modern analogues in the calibration data set, both in terms of analogue taxon assemblages and in terms of analogue environmental conditions (Figure 8). These temperatures continue to increase towards 11,000 cal BP, but coincide with pioneer vegetation, unstable soils, and oligotrophic lake conditions. If the temperatures had indeed been as high as the chironomids suggest, the vegetation cover would have been expected to be more continuous with birch and possibly pine forest, more stable soils, and low inorganic inwash. Such a succession occurred within 600 years in the earliest Holocene at Kråkenes, western Norway (Birks et al. 2000). Perihelion occurred at mid-summer during the early-Holocene, and this may have resulted in warm summers and cold winters (Kutzbach and Guetter 1984), also supported by findings of tree megafossils in mid-Sweden (Kullman 1998a, b). However, at Kråkenes neither chironomids nor pollen indicates such a warming event (Birks et al. 2000; Brooks and Birks 2000), nor does the GRIP borehole temperatures in Greenland (Johnsen et al. 2001).

From about 10,300 years cal BP, the non-analogue period ends and the inferred temperatures may be more reliable (Figure 8). At that time, the temperature was about 0.5 °C warmer than at the present and birch became established. According to Tuhkanen (1993) the forest limit of tree-birch (*Betula pubescens*) follows the mean July isotherm of 10 °C, and Odland (1996) shows that the isotherm increases with distance from the coast in western Norway. The establishment of birch around Råtåsjøen from 10,000 years cal BP at about 10.5 °C corresponds well with this. It has been suggested that the early Holocene in north-western Fennoscandia had an oceanic climate humid (snow-rich) winters since tree-birch was dominant at the time (Berglund et al. 1996; Barnekow 1999). From 9000 cal BP, the low inferred summer temperature, the occurrence of sporangia of moisture-demanding ferns in the macrofossil record (Figure 5), and the presence of tree-birch at Råtåsjøen all suggest a snow-rich climate. However, orbital forcing is believed to have caused an increased summer insolation and increased seasonality at high northern latitudes at this time (Kutzbach and Guetter 1984). This disagreement between low chironomid-inferred temperatures and high insolation may be caused by an enhanced westerly air flow from the North Atlantic and increased cloudiness associated with elevated zonal atmospheric flow causing lower summer temperatures (Seppä and Birks 2001; Hammarlund et al. 2002). At 8000 ± 600 cal BP, there was a pronounced drop in LOI 550 and an increase in the residual accumulation rate (Figure 6). Nesje and Dahl (2001) and Nesje (Nesje 2002) found a corresponding LOI-drop in several lacustrine sediment sequences in Norway between 8400 and 8000 cal BP and suggested that these were related to the '8200-cool event' in Greenland (O' Brien et al. 1995). In southern Norway, this time saw glacial advances termed the 'Finse-event' (Dahl and Nesje 1994; Nesje 2002). At Råtåsjøen, neither pollen, macrofossils, diatoms, nor chironomids display the Finse-event at or around the drop in organic accumulation at around 8000 cal BP. The increased residual minerogenic accumulation rate and concurring July T close to present-day values suggest that the LOI-drop found at Råtåsjøen was caused more by increased winter precipitation than by decreased temperature. Most chironomid-inferred temperature records in northern Fennoscandia indicate a cooling prior to 8200 cal BP between 8500 and 9000 cal BP (Rosén et al. 2001; Seppä et al. 2002; Bigler et al. 2003).

After birch immigration around Råtåsjøen, about 1500 years passed before pine immigrated. The temperatures at Råtåsjøen were most likely below the thermal limit for pine during the Holocene. However, the climate cooling around 8700 cal BP with minimum Holocene temperatures may have prevented the altitudinal pine-advance at altitudes below Råtåsjøen. Towards 8000 cal BP, conditions warmed and the altitudinal pine-advance proceeded (Eide 2003). The expansion of pine may have been influenced by an increasingly continental climate and less winter snow, possibly caused by a replacement of the westerly air flow by a more meridional flow pattern with blocking anticyclones and high pressure weather systems (Barnekow 1999; Seppä and Birks 2001). Following the rapid temperature improvement towards 8000 cal BP, the chironomids indicate a cooling around 7000 cal BP. However, most other studies indicate this to be a period of persistent warm temperatures (Johnsen et al. 2001; Rosén et al. 2001; Seppä and Birks 2001; Korhola et al. 2002; Seppä et al. 2002). During the mid-Holocene temperature maximum at Råtåsjøen, there was also a pronounced cooling at around 5500 cal BP. A similar cooling has been reported from northern Fennoscandia (Seppä and Birks 2001; Korhola et al. 2002; Seppä et al. 2002). The temperature decline proceeded rapidly from ca. 4400 cal BP, in agreement with both ice-core studies from Greenland (Johnsen et al. 2001) and chironomid studies from Fennoscandia (Rosén et al. 2001; Seppä et al. 2002; Bigler et al. 2003). Because of the cooling and persistent lower temperatures, first pine and then tree-birch receded from the local catchment. A gradual Holocene cooling may be expected from glacio-isostatic rebound and decreased long-term solar insolation. In addition, Seppä and Birks (2001) suggest that the reduced influence of the blocking anticyclones may have caused lower temperatures.

There is no clear trend in the highly variable temperature signal from 3500 to 500 cal BP, while there seems to be a slight cooling at ca. 500 cal BP prior to the warming towards the present. Numerous studies have demonstrated cooler climates and advancing glaciers around 400 cal BP, a period termed the 'Little Ice Age' (Grove 1988). Nesje and Dahl (2003) have recently suggested that the 'Little Ice Age' glacial advances may have been caused more by increased winter precipitation than by cool temperatures. With only a slight temperature decrease detected, the present study suggests that the 'Little Ice Age' may not have been much colder than today.

Diatoms, pH-development, and sediment accumulation rate

Long-term natural acidification is a typical phenomenon following lake formation, especially in lakes with a catchment consisting of less basic bedrock (Ford 1990; Renberg 1990; Bradshaw et al. 2000). Acidification is often considered to be a coupled process caused by the release of weak organic acids from the catchment vegetation and a reduced input of cations (Whitehead et al. 1986; Ford 1990; Birks et al. 2000). At Råtåsjøen, after 8000 cal BP, low inferred summer temperatures are associated with low inferred pH values and high summer temperatures are associated with high pH values (Figure 8) (linear regression: $R^2 = 0.42$, $p < 0.01$). The relationship between temperature and pH could have been initiated by the steady depletion of available inorganic nutrients within the catchment, leading to lower alkalinity levels. Before 8000 cal BP there were still available inorganic nutrients to maintain a high pH at low temperatures. Temperature change is directly followed by a pH change in lakes with high in-lake alkalinity generation (Psenner 1988; Psenner and Schmidt 1992). Even though the alkalinity processes are unknown in Råtåsjøen, a high-resolution diatom-inferred pH record covering the last 800 years at Råtåsjøen indicate a time lag of about 35 years between inferred temperature and pH (Larsen, unpublished data). This may suggest that in-lake alkalinity generation is not a significant alkalinity process at Råtåsjøen, which may partly explain the time lag between inferred July T and pH. In addition, the diatoms and chironomids are analysed in alternating sediment levels between 6000 and 7000 cal BP, which partly accounts for the time lag during this period. Psenner and Schmidt (1992) report a relationship between pH reconstructed from diatoms and measured air temperatures during the 19th century, and Psenner (1988) links alkalinity levels through water evaporation, water residence time, and biomass production to temperature. For millennial time scales, Wolfe (2002) found that pH follows temperature trends and stressed the importance of ice-cover as a controlling agent. Sommaruga-Wögrath et al. (1997) and Whitehead et al. (1989) suggest that enhanced chemical weathering in the catchment act as the main mechanism for increased pH during warm periods.

The inferred TP was low during the whole Holocene sequence (from 11.6 to 16.2 μg l^{-1}; Figure 6) with all diatom taxa being characteristic of oligotrophic lakes or with a wide tolerance for nutrients. The reconstructed TP from the surface sediment (13.7 μg l^{-1}) is in good agreement with the measured TP (11.8 μg l^{-1}). We therefore assume that nutrient input to the lake was limited during the Holocene. However, it is worth noting that the reconstructed values may be unreliable as all the fossil diatom assemblages lack modern analogues in the TP-calibration set. The diatom accumulation rate was highly fluctuating (Figure 6) and may be caused by factors other than TP since this was generally stable. The processes that caused the change in diatom accumulation rate may not be attributed to single factors. During times of high inferred summer temperatures both inferred pH and diatom accumulation reached maxima (Figures 6 and 8). The increased diatom accumulation rate during periods of warm summers may be due to increased growth rate and/or longer growing seasons. A lower growth rate and shorter growing seasons may have caused a decreased diatom accumulation rate during periods of cooler summers. Similarly, Kaplan et al. (2002) interpret biogenic silica as a direct proxy for palaeoclimate and assume a strong linkage between lake production and air temperature. Typical planktonic taxa may be more abundant during warmer periods when the ice-cover is less extensive (Douglas and Smol 1999). This is in agreement with the present study, as during the mid-Holocene temperature maximum at Råtåsjøen the accumulation of planktonic or tycoplanktonic *Cyclotella* taxa increased and peaked around 4750 cal BP together with a peak in the accumulation of *Nitzschia* taxa. The latest accumulation peak at around 4000 cal BP commenced during warm summers when *Fragilaria* taxa dominated the accumulation, but continued into the climate deterioration when the accumulation of *Achnanthes* and *Navicula* taxa peaked. The accumulation increase of these taxa coincides with an increase in organic accumulation and could be due to increased cation and nutrient supply as birch trees declined from the catchment and the organic material from the catchment mineralised. A similar process was proposed by Whitehead et al. (1973), where an increase in diatom accumulation was caused by the release of nutrients from coniferous leaves as hemlock was replaced by hardwoods.

LOI-profiles, productivity, and temperature

To define changes in aquatic productivity, Wetzel (1975) and Whiteside (1983) state that the organic matter measured as LOI 550 is of little use and argue that organic matter percentage in the sediments reflects the varying intensity of terrestrial erosion and residual minerogenic input. In the present study, the percentage and accumulation rate of organic matter generally correlates (Figure 6). However, there are some inconsistencies that give valuable information about the lake catchment and environment. These inconsistencies seem to coincide with changes in vegetation cover and changes in residual minerogenic accumulation, especially subsequent to the establishment of woodland. After the local woodland was established, the percentage organic matter increased rapidly. This gives the impression that the organic content was mostly based on woodland soils and litter. However, this conclusion is not supported by the organic accumulation rates. The accumulation rates show that the catchment already had a significant production of organic matter prior to the establishment of trees. Further, the increased relative abundance of organic matter was most likely due to the stabilising effect of vegetation leading to a decreased inwash of mineral matter. Similarly, at ca. 8300 cal BP, based on the percentage LOI, one may draw the conclusion that the input of organic carbon reached a maximum. This is in contrast to the organic accumulation rate which indicates that the input of organic carbon reached a maximum 5000 years later as the trees receded from the catchment. Hence, organic matter and nutrients in the catchment were no longer stored in trees and were transported to the lake. At around 8000 cal BP, the rapid decrease in percentage organic matter was more pronounced than the decrease in accumulation. Again, this indicates that the percentage of organic matter was more related to increased residual minerogenic accumulation than decreased production. The increased residual minerogenic accumulation coupled with only a small decrease in organic matter accumulation suggests that this was a time of increased erosion, most likely resulting from increased precipitation.

Willemse and Törnquist (1999) and Battarbee et al. (2002a) suggest a correlation between lacustrine records of percentage organic matter (LOI 550) and temperature. In support of this, at Råtåsjøen, from the time of well established organic production at about 9000 cal BP and towards the present, there is a weak, but significant relationship between sediment percentage organic matter and chironomid-inferred July T (linear regression: $R^2 = 0.30$, $p < 0.01$; Figure 6). Organic matter accumulation, on the other hand, shows no such relationship with inferred temperature. This may indicate that percentage organic matter may roughly reflect temperature and that organic accumulation rates are more related to changes in the local catchment. Diatom accumulation rate is not related to the minerogenic fraction of the sediments (Figure 6). This indicates that the inorganic sediments at Råtåsjøen do not predominantly represent diatomaceous silica, and that the relationship between temperature and organic content is not driven by diatom productivity changes.

Timing of events

Wright (1984) states that each proxy will have a lag in response time depending on the amount and rate of temperature change and different sensitivities and thresholds to changes in the system. For the present study, there are similarities in the timing of statistically significant compositional events in vegetation, diatoms, chironomids, and sediments (Figure 6). However, some of the shifts occur independently for the different proxies and the concurring shifts only occur during major lake events. The first corresponding shift occurred as the local production increased during the early Holocene and the sediment characters changed. The diatoms responded to the commencement of this event seemingly without any time lag, while the chironomid response may have lagged by about 200 years. Significant shifts in chironomids, diatoms, and sediments also occurred during the climate deterioration after the mid-Holocene temperature maximum. This change occurred within 200 years as a consequence of the combined effects of deteriorating climate and gradually changing catchment regime. A time lag of about 200 years is in line with Davis and Botkin (1985), who predict that a compositional biotic change should lag a stepped climate event by 100–200 years, and Williams et al. (2002) who report 100–200 years vegetation responses to climate change.

Conclusions

Interpretations on the Råtåsjøen environmental history were roughly based on 150,000 data entries of pollen, macrofossils, diatoms, chironomids, on changes in the organic and carbonate content of the sediments, on sediment accumulation rates, and on the sediment lithology. Concurrent statistically significant changes in the different proxies only occur during major lake events, like the July T minima around 9000 and 3700 cal BP. This suggests that the proxies at times do not respond to the same environmental variables and/or may be linked to the turnover thresholds of particular taxa in the system. The chironomid fauna of Råtåsjøen was apparently most strongly affected by July temperature, catchment regime, and oxygen levels, while pH, temperature, catchment regime, and turbidity controlled the diatom flora. For both the diatom–pH and chironomid–July T calibration sets, this study has shown that it may be difficult to infer unequivocally past environmental conditions during non-analogue periods. The non-analogue problems at Råtåsjøen are probably a result of the unique combination of environmental conditions in the lake in the early Holocene, conditions that today are very rare in Scandinavia. This condition includes the presence of *Chara* with highly fluctuating sediment parameters in an alpine lake.

From the time of *Betula* tree establishment at *c.* 10,000 cal BP, lake water conductivity and acidification increased. The cause of these limnological changes could include increased inwash of organic acids from soil, decreased minerogenic inwash, decreased groundwater inflow, and depletion of cation input to the lake. From the time of well-established organic production at *c.* 8900 cal BP and towards the present, the inferred July T is weakly, but statistically significantly correlated to percentage organic matter in the sediments. Organic accumulation rates are more related to changes in the local catchment than to July T. In line with Livingstone et al. (1958) this suggests that more comprehensive knowledge on changes in catchment production and erosion will be gained by examining both percentage organic matter and organic matter accumulation rates. Inferred July T became a significant factor controlling inferred pH after about 8000 cal BP where high pH was associated with high July T and low pH was associated with low July T. There is also a strong linkage between diatom accumulation and July T where warmer summers resulted in longer growing seasons and faster growth rates. *Pinus sylvestris* reached its upper limit around Råtåsjøen between ca. 8000 and 4400 cal BP coinciding with maximum inferred temperatures. Subsequent to the deterioration in climate that commenced from ca. 4400 cal BP, trees receded from the catchment and the organic influx to the lake increased. Both the chironomid- and diatom taxon assemblages fluctuated largely before the establishment of woodland in the catchment and after the woodland receded. This may have been related to the stabilising effect of trees on soils and the sheltering effect to wind exposure.

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