AMS ^{14}C measurements and macrofossil analyses of a varved sequence near Pudozh, eastern Karelia, NW Russia

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The laminated sediments at Pudozh in eastern Karelia are generally assumed to have been deposited between 13 000 and 16 000 ^{14}C yr BP and have been used to date the recession of the active ice margin. However, 17 AMS ^{14}C measurements performed on terrestrial plant macrofossils contained in these sediments show that deposition began during the late Allerød, when the ice margin had already receded to the northern part of Lake Onega. Based on an age model, we assume that the 1933-year-long varved sequence covers the time period between *c.* 12 900 and 11 000 calendar years BP. During this period, which comprises the later part of the Late Weichselian and the early Holocene, the local vegetation consisted of open, tree-less dwarf shrub heaths. Increased soil erosion may have occurred before 12 550 calendar years BP.

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Reconstruction of the timing of the ice recession in Russian Karelia has been based on clay-varve chronological studies (Sauramo 1926; Markov 1931), bulk-sediment radiocarbon dates from lacustrine sediments (summarized in Ekman & Iljin 1991) and on a correlation of palaeomagnetic curves established at varved clay sections (see Fig. 1) (Bakhmutov & Zagniy 1990). However, in these latter studies, the number of varves in each section and their thickness was only estimated (Bakhmutov & Zagniy 1990; Ekman & Iljin 1991) and varve diagrams or varve diagram correlations are not available. Following these investigations, the areas east, south and west of Lake Onega became ice-free during the so-called Vepsovo-Krestets and Luga stages (16 000 and 13 000 ^{14}C years BP, respectively), which are commonly correlated with the Oldest Dryas Chronozone (Bakhmutov & Zagniy 1990; Ekman & Iljin 1991) (Fig. 1). During the successive deglaciation, the ice margin receded to the Neva ice marginal line, which is thought to have formed during the Older Dryas Chronozone, i.e. 12 500 to 11 800 ^{14}C years BP. The two younger ice marginal lines, the Rugozero/Salpausselkä I (11 300–10 850 ^{14}C years BP) and the Kalevala/Salpausselkä II (10 600–10 200 ^{14}C years BP) stages, are associated with the Younger Dryas Chronozone (Ekman & Iljin 1991).

Recently, Hang (1997) reinterpreted the old clay-varve correlations of Markov & Krasnov (1930) and estimated, in comparison with the Swedish and Finnish clay-varve chronology, that the northern part of Lake Onega had already become ice-free between 13 940 and 13 140 varve years BP. However, he did not take into account an error of several hundred years which is present in the Swedish varve chronology (Wohlfarth 1996; Björck *et al.* 1996). When these years are added, the timing of the deglaciation would be equivalent to approximately 14 800 and 14 000 calendar years BP and would correspond to the Bølling period (Björck *et al.* 1996). Although such an early ice retreat is in contrast to the generally assumed deglaciation chronology (Ekman & Iljin 1991), Hang's (1997) estimate of about 800 years for a deglaciation from the middle to the northern part of the Lake Onega basin agrees well with the *c.* 1000 years estimated from ongoing studies in Lake Onega (Saarnisto *et al.* 1995). Apart from Hang's (1997) compilation of Markov & Krasnov's (1930) earlier studies, no other clay-varve chronological studies, which would allow assessing the timing of the deglaciation in greater detail, are presently available for eastern Karelia.

The occurrence of varved clays, possibly older than 13 000 ^{14}C years BP (Ekman & Iljin 1991), led to the initiation of a research project in the eastern part of Russian Karelia, with the objective of reinvestigating these old varved-clay sections and to AMS ^{14}C date terrestrial plant macrofossils from these clays. Here, we present AMS ^{14}C measurements and macrofossil analyses performed on the 1933-year-long varved section at Pudozh, which is located east of Lake Onega (Fig. 1). Our results are compared to the generally

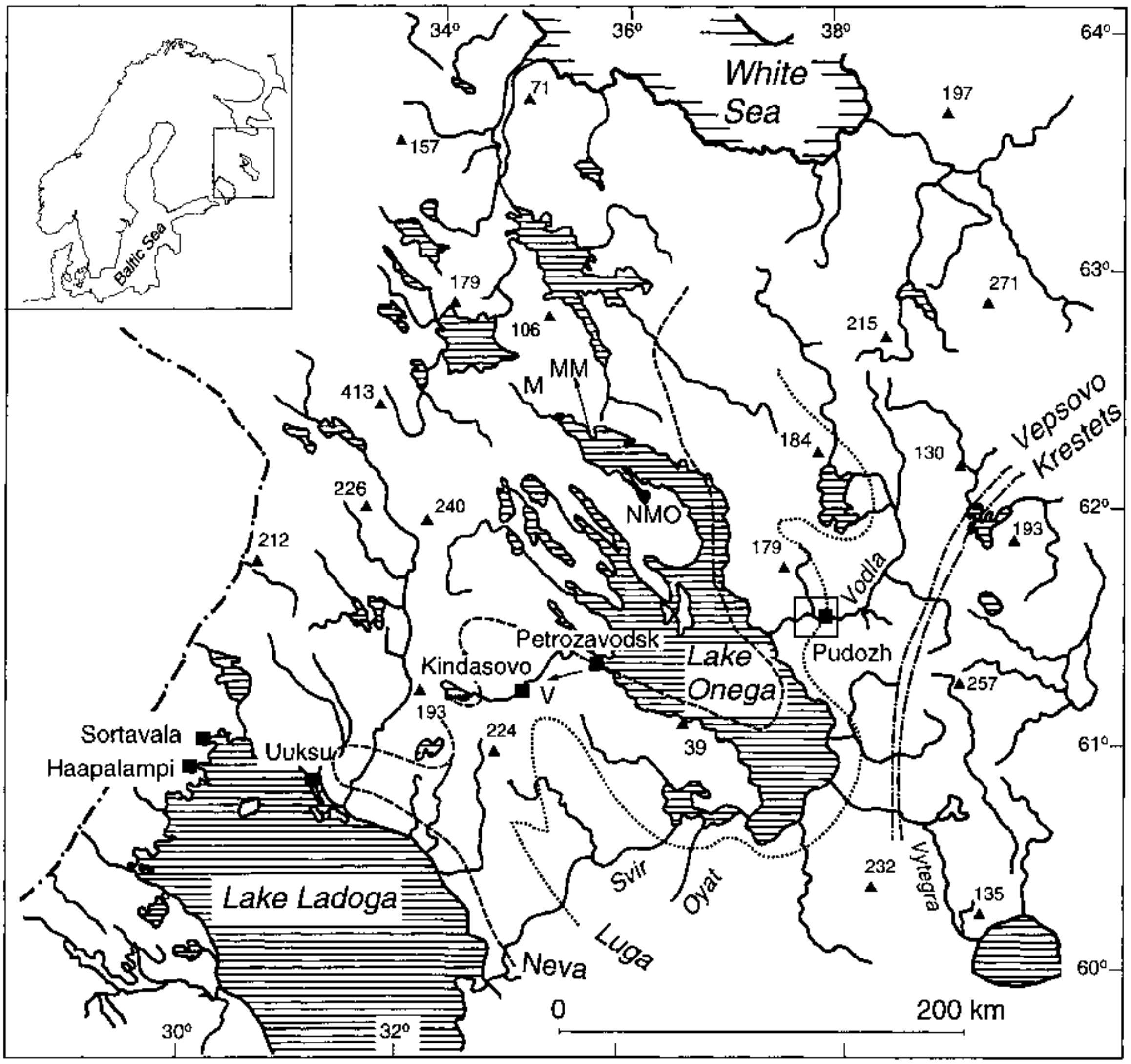

Fig. 1. Location of the study area in eastern Karelia, Russia. The age assignment of the ice-marginal formations Vepsovo–Krestets (16 000–14 000 ^{14}C yr BP), Luga (14 000–13 000 ^{14}C yr BP and Neva (12 500–11 800 ^{14}C yr BP) is based on a correlation of palaeomagnetic curves established at the sites of Pudozh, Petrozavodsk, Kindasovo, Uuksu, Sortavalla and Haapalampi (Bakhmutov & Zagniy 1990). Ice-marginal lines are according to Ekman & Iljin (1991). The investigated section at Pudozh is marked with an open square. MM = Morskaya–Maselga threshold, M = Maselga threshold, V = Vieljärvi threshold, NMO = Lake Nizhnee Mjagrozero. Arrows indicate the different drainage paths during the deglaciation.

assumed deglaciation chronology (Ekman & Iljin 1991), to palaeoenvironmental and palaeoclimatic studies in southeastern Karelia (Elina & Filimonova 1996) and to isolation studies in Lake Onega (Saarnisto *et al.* 1995).

Late Weichselian and early Holocene lake-level changes in the Lake Onega basin

Deglaciation of the southeastern part of the Onega lowland, which according to Kvasov (1979) occurred during the Luga stage (*c.* 13,000 ^{14}C years BP), led first to the formation of the small ice-dammed lakes Vytegorskoe and Vodlinskoe, which were situated east and south of present-day Lake Onega (Poryvkin 1960; Kvasov 1976, 1979). During further deglaciation these lakes gradually merged to form a 'South-Onega Ice Lake', with a water level of *c.* 95 m a.s.l. This lake drained first to the east across Lake Volotskoe to the Onega river basin, and later across a lower threshold via the River Svir into the Lake Ladoga basin (Kvasov 1976, 1979; Saarnisto *et al.* 1995) (Fig. 1). The deposition of varved clays, sands and silts in the lower and middle part of Vodla River (below *c.* 55–65 m a.s.l.)

is taken as evidence for the existence of a deep, eastward extending fjord (Demidov, unpublished), which may have served as an additional drainage towards the Onega river catchment (Kvasov 1976, 1979).

When the northern part of Lake Onega had become ice-free, outlets towards the White Sea (Morskaya–Maselga and Maselga thresholds) and possibly towards Lake Ladoga (Vieljärvi threshold) became active (Kvasov 1976, 1979; Saarnisto *et al*. 1995). Studies performed in the northern part of Lake Onega show such a drainage path to the White Sea between *c*. 11 000 and 9500 ^{14}C years BP, and indicate that the western drainage to Lake Ladoga via Vieljärvi may have become abandoned around 10 000 ^{14}C years BP (Saarnisto *et al*. 1995). After the emergence of the northern Maselga threshold due to land uplift, the River Svir, which connects the southwestern part of Lake Onega with Lake Ladoga, became the main outlet (Fig. 1). The initiation of this southwestern outlet at around 9500 ^{14}C years BP caused a lake lowering to the present-day level of Lake Onega, i.e. to 33 m a.s.l. (Saarnisto *et al*. 1995).

The direction of the isobases of the present-day land uplift are more or less orientated NNE–SSW and the tilt axis of Lake Onega runs from the outlet of River Svir to the mouth of River Vodla (Zemljakov 1936; Kakkuri 1993) (see Fig. 1). Owing to the differential land uplift, the shoreline diagram for the time period 10 100–9 400 ^{14}C years BP shows a shoreline at 125 m a.s.l. in the northern part, at *c*. 80 m a.s.l. in the central part and at *c*. 30 m along the tilt axis (i.e. the River Vodla area) (Saarnisto *et al*. 1995). Based on these estimates, Saarnisto *et al.* (1995) calculated an uplift rate of *c*. 90 m in the northern part, of *c*. 50 m in the central part and a transgression of *c*. 15 m in the southeastern part since *c*. 9500 ^{14}C years BP. However, following Devyatova (1986), Younger Dryas shorelines were recognized at slightly more than 57 m a.s.l. in the River Vodla area and at 66–61 m a.s.l. in the northern and northwestern part of Lake Onega. Her estimates for the Preboreal range between 45 m and 50 m a.s.l. at the Vodla river mouth and between 55 m and 59 m a.s.l. in the northern part of Lake Onega.

Site description

The studied section is exposed in a steep cliff, between 45 m and 37 m a.s.l., on the southern shore of River Vodla, southwest of the town of Pudozh (61°48′20″ N; 36°28′E) (Fig. 1). Vodla, which is one of the largest rivers in Russian Karelia, originates in the northeastern part of the province and drains west of Pudozh into Lake Onega. This meandering river is surrounded by a vast flood plain with numerous abandoned channels and oxbow lakes. Geotechnical corings around Pudozh have shown that laminated sediments (clay and sandy silt) occur frequently between 35 m and 55 m a.s.l. in the area west and northwest of Pudozh along the river.

Pudozh belongs to the middle taiga zone and has a moderate continental climate with long winters and short summers (Antipin *et al.* 1987). The annual average air temperature is 1.6°C (Elina & Filimonova 1996), with mean January and mean July temperatures around −1.5°C and 16°C, respectively. Annual precipitation is given as around 600 mm/yr (Elina & Filimonova 1996). The present-day natural forests are mainly composed of spruce and pine, with some stands of alder and birch. However, large parts of these forests have been cut and are now inhabited by birch and alder.

Ice-marginal deposits associated with the Vepsovo-Krestets stage can be observed east of Pudozh, where they follow approximately in a NW–SE direction the administrative borders between the provinces of Karelia, Vologda and Archangelsk (Fig. 1). These terminal deposits have been mapped as moraines, hummocky moraines and glaciofluvial material (Niemelä *et al*. 1993). The next younger ice-marginal stage, the Luga terminal belt, has been mapped as end moraine ridges and hills which traverse the town of Pudozh and extend along the southern and western shore of Lake Onega (Ekman & Iljin 1991). The Neva ice-marginal line west of Pudozh consists of fluvioglacial material (Niemelä *et al*. 1993).

Methods

The outcrop at Pudozh was excavated in September 1996 and divided into seven subsections (Table 1). The surface was cleaned carefully before the laminated silty sand and clay couplets were measured. From top to bottom, each silty sand-clay couplet was first marked with a needle and then measured by fixing a paper strip to the cleaned surface and by marking the thickness of each couplet on the paper strip. Sediment samples comprised between 20 and 70 couplets and also included non-laminated sandy gyttja lenses which were intercalated in the lower part of the section. Sample mass varied between *c*. 1 kg (in the organic lenses) and *c*. 3 kg (in the laminated part).

In the laboratory, the samples were wet sieved through meshes with a diameter of 0.5 mm. The material retained on the sieves was analysed under a dissecting microscope and divided into terrestrial, telmatic and limnic plant and animal remains (Table 2). The macrofossils were not counted, but put into frequency classes. Seventeen samples from Pudozh contained enough terrestrial plant macrofossils for radiocarbon dating (Table 3a). Leaves, nutlets, catskin scales, flowers and seeds of *Dryas octopetala, Betula nana* and *Salix* spp. as well as twigs of *Alnus* were generally selected for AMS ^{14}C measurements. The macrofossils were dried overnight on aluminium foil at

Table 1. Lithostratigraphy and number of couplets measured in each subsection at Pudozh.

Sub-section	Thickness (cm)	Lithostratigraphic description	No. of couplets
1	134	Light-brown, weathered clayey silt	Not counted
	33.5	Couplets of brown-yellow silt and brown to greyish-blue clay laminae	1–66
2	82.8	Couplets of brown-yellowish silt and brown-greyish to grey-bluish clay laminae Fe/Mn precipitates around silt laminae in lower part, especially between couplet nos. 150 and 270. Couplets 140, 151, 187 and 210 contain thick (14–19 mm) fine sand laminae	67–270
3	29	Couplets of brown-yellowish fine sand and brown-reddish clay laminae. Couplets 283 and 300 contain thick (13 and 15 mm) fine sand laminae. Fe/Mn precipitates around fine sand-silt laminae	271–340
	1.4	Disturbance	341–343 (estimated)
	82.3	Couplets of brown-yellowish fine sand and brown-reddish, partly grey-bluish clay laminae. Couplet 390 contains a thick (15 mm) fine sand lamina. Fe/Mn precipitates around fine sand-silt laminae and in single fine sand-silt laminae	344–520
4	45.9	Couplets of beige-yellowish fine sand and brown-reddish clay laminae. Fe/Mn precipitates around fine sand-silt laminae	521–642
	31	Couplets of grey-orange-beige fine sand and brown-reddish clay laminae	643–717
	64.9	Couplets of light-grey to orange coloured fine sand and grey-bluish to dark grey clay laminae. Fe/Mn precipitates around fine sand-silt laminae. Couplet 858 contains a thick (13 mm) fine sand lamina	718–884
	c. 15	Couplets difficult to count; ?disturbance	885–906 (estimated)
5	137.7	Couplets of light-grey to orange coloured fine sand and grey-bluish to dark grey clay laminae. Fe/Mn precipitates around single fine sand laminae. Couplets 972, 1136, 1218, 1242, 1298 contain thick (13–19 mm) fine sand laminae	907–1371
	3.2	Indistinct couplets	1337–1343 (estimated)
6	98	Couplets of light-grey to orange coloured fine sand and grey-bluish to dark grey clay laminae. Fe/Mn precipitates around fine sand laminae, especially in couplets 1513–1516 and 1520 and FeS stains in clay laminae between couplets 1520 and 1538. Couplets 1480, 1517, 1520–1521, 1527, 1572 contain thick (13–24 mm) fine sand laminae	1372–1623
	30–35	Folded and faulted layers of beige sand, grey-bluish clay and dark brown organic-rich silty sand with brown organic clay balls and twigs	1624–1630
	38	Couplets of light-grey to orange coloured fine sand and grey-bluish to dark grey clay laminae. Fault structures in couplets 1631–1640	1631–1692
	0–17	Dark-brown to reddish-brown clay with sand lenses, organic debris and rounded grey clay balls.	1693–1697
	55	Couplets of light-grey to orange coloured fine sand and grey-bluish to dark grey clay laminae. Fe/Mn precipitates around fine sand-silt laminae	1698–1821
	0–2	Brown-greyish, organic-rich clay with light grey fine sand laminae and rounded dark-brown clay particles	1822–1823
	4.8	Couplets of light-grey to orange coloured fine sand and grey-bluish to dark grey and dark brown clay laminae. Fe/Mn precipitates around fine sand laminae. Rounded clay particles are intercalated in single fine sand laminae	1824–1830
7	2.5	Rippled fine sand lamina	1831
	3.7	Couplets of light-grey to orange coloured fine sand and grey-bluish to dark grey clay laminae	1832–1839
	1–5	Dark brown organic-rich clay with light grey fine sand lenses	1840
	38.2	Couplets of light-grey to orange coloured fine sand and grey-bluish to dark grey clay laminae. 1841–1933	
	56	Bluish-grey stiff clay with folded and faulted laminae of beige orange-coloured sand and silt lenses	Not counted

70°C before the samples were submitted. The pretreatment in the AMS laboratory included 1% HCl (6 h below boiling) and 0.5% NaOH (1 h at 60°C). A complete pretreatment was performed on all samples, except Ua-11536 and Ua-11501, which were pretreated with only 1% HCl.

Results

Lithostratigraphy

The excavated section comprised approximately 10 m and terminated *c.* 2–3 m above the river. A lithostratigraphic description is given in Table 1. The bottom of the sequence in subsection 7 is characterized by a stiff, bluish-grey clay with folded and faulted laminae and lenses of beige and orange coloured sand and silt which grades upwards into distinctly laminated fine sandy silt and clay layers. The uppermost 134 cm of the section consists of a light brown, weathered clayey silt with indistinct laminations. Each of the measured 1933 couplets is made up of fine sand/silt and clay laminae with distinct boundaries. FeS stains and Fe/Mn precipitates around the sand laminae are common and often the sandy parts are orange-yellow in colour. Some sand laminae show internal laminations.

In the upper part of subsection 7 and in the lower and middle parts of subsection 6, four dark-brown organic-rich clay and sand layers are intercalated in the laminated sequence (Table 1). The uppermost of these four layers, which is composed of folded and faulted layers and lenses of beige sand, grey-bluish clay and dark-brown organic-rich silty sand, caused erosion and dislocation of the underlying laminations. However, the lowermost two organic-rich and clayey horizons did not disturb the substratum, because the laminations below and above these horizons continue horizontally.

Fig. 2 shows the thickness variation of each couplet and the intervals with disturbances and intercalated organic-rich lenses. The latter are restricted to the oldest part of the sequence between couplets 1600 and 1900. Apart from disturbances around couplets 1343–1337, 906–885 and 340–345, where the number of laminae was based on an extrapolation of the sedimentation rates of the underlying and overlying couplets, the middle and upper parts of the sequence seem undisturbed and the laminations were easy to measure. Mean laminae thickness (clay and silty sand laminae), calculated for 100-laminae intervals, ranges between 3.5 and 4.4 mm for couplets 1100–1933, except for a period of thinner laminae between couplets 1400 and 1500. A distinct decrease in thickness to $\sim$2.9 mm is visible between couplets 900 and 1100. This is followed by a two-step increase in thickness to $\sim$4 mm between couplets 800 and 500 and to $>$4 mm from couplet 500 upwards.

Macrofossil content

Macrofossils were present in samples 1–3 and 13–42 (subsections 1–2 and 4–7), but they were absent in subsection 3 (samples 4–12) (Table 2). Although the highest abundance of macrofossils could be observed in an organic lens (sample 32), which was intercalated in the laminated sequence, there is little variation between samples. Therefore, no zonal subdivision is proposed and the fossil assemblage is described as a single unit.

A total of 25 taxa were identified, but many of these taxa obviously comprise different species. Nevertheless, the total diversity is surprisingly low, with mainly terrestrial plant macrofossils and a rare occurrence of telmatic and limnic species (Table 2). *Betula nana, Dryas octopetala* and mosses are common, while remains of *Salix* spp., *Arctostaphylos alpina, Juniperus communis, Alnus* cf. *glutinosa, Potentilla nivalis, Melandrium angustiflorum, Poaceae, Carex* spp. and *Cenococcum geophilum* are sparse. The poor limnic fauna comprised midge larvae, caddis fly larvae and water fleas. Rodentia droppings were observed in a few samples.

The macrofossils are generally well preserved. The organic lens (sample 32) contained, for example, many *Dryas octopetala* twigs with entire leaves still attached, which could not have been transported very far. A few achenes of *Dryas octopetala* were also present. In contrast to the leathery leaves of *Dryas octopetala*, achenes are rarely preserved in sediments and their occurrence here is probably due to a rapid sedimentation.

It is remarkable that only a few remains of limnophytes are present (Table 2). *Ranunculus* sect. *Batrachium* grew in shallow water, either along the shore of the lake basin or in ponds in the drainage area. Remains of land plants are common, and we suggest that these give a good picture of the local vegetation of the area. Remains of *Betula nana* and *Dryas octopetala* are by far the most common, indicating that dwarf shrub heaths existed, thus pointing to dry and moist soils (Table 2). As far as we know, *Dryas octopetala* has not previously been reported from Russian Karelia, although this could reflect the lack of macrofossil studies. The species has been recorded from a number of late-glacial sites at the bottom of the Gulf of Finland (Tralau 1962) and from Finnish Karelia (Bondestam *et al*. 1994), so its occurrence at Pudozh is not surprising. Today, the nearest habitats for this shade-intolerant species are in northernmost Russia, but isolated finds occur in Kunsamo, eastern Finland at a latitude of 66°N. The same applies to *Arctostaphylos alpina* of which a few endocarps are present, and *Melandrium angustiflorum*, which is represented by a single seed. The finds of grasses and sedges could indicate that plants grew on wet soil, and this also applies to several of the bryophytes. However, *Distichium* sp. and *Ditrichum* sp. are characteristic of open plant communities on well-drained soil without shading trees. These taxa, as well as *Dryas octopetala*, indicate calcareous soils.

The macrofossil analysis thus points to an open and treeless local landscape with dwarf shrub heaths dominated by *Betula nana* and *Dryas octopetala* on well-drained soils. Poorly drained soils housed a vegetation with sedges, grasses and bryophytes. The finds of droppings, similar to lemming droppings, show that small rodents were present in the area.

AMS ^{14}C measurements

The majority of the radiocarbon dates range between

Table 2. Macroscopical plant and animal remains from the sequence at Pudozh (1 = terrestrial; 2 = telmatic; 3 = limnic; – = absent; r = rare; c = common; a = abundant).

Sample no.	No. of couplets	*Betula nana* (1)	*Dryas octopetala* (1)	*Salix* spp. (1)	*Arctostaphylos alpina* (1)	*Juniperus communis* (1)	*Alnus* cf. *glutinosa* (1–2)	*Potentilla nivalis* tp (1)	*Melandrium angustifolium* (1)	Poaceae (1–2)	*Carex* spp. (1–2)	*Ranunculus* sect. *Batrachium* (3)	Bryopsida (1–3)
2	51–100	–	–	–	–	–	–	–	–	–	–	–	–
3	101–145	–	r	–	–	–	–	–	–	–	–	–	r–
13	595–638	r	r	–	–	–	–	–	–	r	–	–	r
14	640–689	c	r	–	–	–	–	–	–	–	–	–	–
15	691–737	c	–	–	–	–	–	–	–	–	–	–	r
16	741–789	c	r	–	–	–	r	–	–	r	r	–	r
17	794–840	r	r	–	–	–	–	–	r	–	–	–	r
18	845–910	c	c	–	–	–	–	–	–	r	–	–	r
19	975–1036	r	r	–	–	–	–	–	–	–	r	–	r
20	1039–1086	r	r	–	r	–	–	–	–	–	–	–	–
21	1092–1135	r	–	r	–	–	–	–	–	–	–	–	r
22	1150 ± 10	r	–	–	–	–	–	–	–	–	–	–	–
23	1150–1190	r	r	–	–	–	–	–	–	–	–	–	–
24	1230–1291	r	r	r	–	–	–	–	–	–	–	–	r
25	1292–1298	c	–	–	–	–	–	–	–	–	–	–	r
26	1299–1318	r	r	–	–	–	–	–	–	–	–	–	r
27	1319–1350	r	r	r	–	–	–	–	–	–	–	–	–
28	1397–1445	r	r	–	r	–	–	–	–	–	–	–	r
29	1447–1498	r	r	r	–	–	–	–	–	–	–	–	r
30	1507–1554	r	–	–	–	r	–	–	–	r	r	–	–
31	1557–1598	c	r	–	–	–	–	–	–	–	–	–	–
32	1599–1629	a	a	r	–	–	–	–	–	–	c	–	a
	Organic lens												
33	1630–1691	c	r	–	–	–	–	–	–	–	–	–	r
34	1692–1696	r	r	–	–	–	–	–	–	r	–	r	–
	Clay layer												
35	1697–1748	r	r	–	–	–	–	–	–	–	–	–	r
36	1748/1749	c	c	r	–	–	–	–	–	r	–	r	r
	Organic lens												
37	1749–1792	c	r	r	–	–	–	–	–	r	r	–	r
38	1793–1830	c	r	r	–	–	–	–	–	–	–	–	r
39	1831–1842	c	c	r	r	–	–	–	–	r	r	–	c
40	1843–1892	c	c	–	–	–	–	r	–	–	r	–	r
41	1893–1932	c	c	r	–	–	–	–	–	r	r	–	c
42	Tectonised clay	c	c	r	r	–	–	r	–	–	–	–	r

11 590 ± 100 and 10 205 ± 150 ^{14}C years BP, while four dates show ages of between 9475 ± 105 and 7245 ± 115 ^{14}C years BP (Table 3a). These latter measurements are clearly younger when compared with the main trend (Fig. 3a) and may be explained by their contamination with modern carbon during preparation (Wohlfarth *et al*. 1998). Samples from the lowermost part of the sequence (older than couplet 1700) give fairly constant radiocarbon ages of between 10 845 ± 95 and 11 365 ± 95 years BP and agree very well with each other within two standard errors (Fig. 3a). The same holds true for the macrofossils derived from the organic lens (samples P-32, P-32a), which show ages of 10 515 ± 95 and 10 590 ± 100 ^{14}C years BP. A slightly larger discrepancy can be seen between samples P-27 (10 935 ± 340), P-22a (10 205 ± 150), P-22b (11 005 ± 175) and P-14 (10 595 ± 90), i.e. between couplets 1350–640 (Table 3a, Fig. 3a), although these ages still overlap within the error limits. In the case of samples P-22a and P-22b, the ‘true’ age may be represented by sample P-22a, where *Dryas octopetala* leaves were dated, which are less resistant to reworking than the twigs dated in sample P-22b. If this holds true, sample P14 has to be regarded as too old, i.e. it may have contained reworked plant material.

Age assignment

The lowermost part of the sequence (couplets older than 1697) is characterized by ages ranging from 11 365 ± 95 to 10 975 ± 95 ^{14}C years BP. Between couplets 1697 and 1599 the ^{14}C ages drop to 10 590 ± 100 and 10 515 ± 95 BP. The next younger dates, obtained between couplets 1350 and 640, indicate no clear age trend (Table 3a, Fig. 2a). However, if we exclude samples P-22b and P-14 with the argument that reworked plant material was dated, sample P-22a may indicate gradually decreasing ages (Fig. 3a).

Radiocarbon ages around 11 400–11 000 ^{14}C years BP are characteristic of the later part of the Alleröd, while the drop in ages from 11 000 to about 10 600 ^{14}C years BP marks the Alleröd/Younger Dryas transition in high-resolution dated terrestrial lake sediment sequences (Goslar *et al*. 1995, 1999; Björck *et al*. 1996). Based on the assumption that the atmospheric ^{14}C/^{12}C content was the same worldwide, we may use the rapid decrease in ages to correlate our data set to other high-resolution records (Goslar *et al*. 1999). Such an approach allows us to evaluate whether the couplets are real annual laminations, i.e. varves.

In Fig. 3b we compare the Pudozh ^{14}C/lamination record with the ^{14}C-dated high-resolution record of

Distichium sp. (1)	*Ditrichum* sp. (1)	*Hylocom-nium* sp. (1)	*Scorpidium scorpioides* (2–3)	*Drepano-cladus* sp. (2–3)	*Cenococcum geophilum* (1–2)	Coleoptera (1–3)	Formicidae (3)	Trichoptera (1–3)	Nematocera (1)	Chironomi-dae (1–3)	*Daphnia pulex* tp. (3)	Rodentia (1)	Radiocarbon age (yr BP)
–	–	–	–	–	r	–	–	–	–	–	–	–	
–	–	–	–	–	–	–	–	–	–	–	–	–	
r	–	–	r	r	r	r	–	–	–	–	–	–	
–	–	–	–	–	r	–	–	–	–	–	–	–	10,595 ± 90
–	–	–	–	–	–	–	–	–	–		–	–	
r	–	–	–	r	r	–	–	–	–	–	–	–	
–	–	–	–	–	r	–	–	–	–	–	r	–	
r	–	–	–	r	–	–	–	r	–	–	–	–	
–	–	–	–	–	r	r	–	–	–	–	–	–	9,085 ± 195
–	–	–	–	–	–	–	–	–	–	–	–	–	7,715 ± 170
–	–	–	–	–	–	–	–	–	–	–	–	–	
–	–	–	–	–	–	–	–	–	–	–	–	–	11,005 ± 175
													10,205 ± 150
–	–	–	–	–	–	–	–	–	–	–	–	–	
–	–	–	–	–	–	–	–	–	–	–	–	–	7,245 ± 115
–	r	–	–	r	–	–	–	–	–	r	–	–	
–	–	–	–	–	r	–	–	–	–	–	–	r	
–	–	–	–	–	–	–	–	–	–	–	–	–	10,935 ± 340
–	r	–	–	–	–	–	–	–	–	–	–	–	
–	r	–	–	–	–	–	–	–	–	–	–	–	
–	–	–	–	–	–	–	–	–	–	–	–	–	
–	–	–	–	–	r	–	–	–	–	–	r	–	
r	r	r	–	–	–	r	r	–	r	–	–	r	10,590 ± 100
													10,515 ± 95
r	–	–	–	–	r	r	–	r	–	–	–	–	
–	–	–	–	–	–	–	–	–	–	–	–	–	
–	–	–	–	–	–	–	–	–	–	–	–	–	11,070 ± 110
–	–	–	–	–	–	–	–	–	r	–	–	–	11,070 ± 95
–	–	–	–	–	–	r	–	–	–	–	–	–	10,845 ± 95
r	r	–	–	–	–	–	–	–	–	–	–	–	
–	–	–	–	–	r	–	–	r	r	r	–	r	11,365 ± 95
–	–	–	–	–	–	–	–	–	–	–	–	–	10,975 ± 95
–	–	–	–	–	–	–	–	–	–	–	–	–	11,030 ± 110
–	–	–	–	–	–	r	–	r	–	–	–	–	9,475 ± 105
													11,000 ± 125

Lake Gosciaz in Poland (Goslar *et al*. 1995; Goslar & Madry 1998), and with the ^{14}C-dated part of the Swedish varve chronology (Goslar *et al*. 1999). For this comparison, we assume that each couplet (fine sand/silt and clay lamina) represents 1 year and that the drop in radiocarbon ages occurred at the same time as in the two high-resolution records, i.e. at 12 650 $\pm$ 60 calendar years BP (Goslar & Madry 1998; Goslar *et al*. 1999). The best fit between the Pudozh data set and the combined Lake Gosciaz/Swedish varve record can be achieved if the local varve year/couplet no. 1700 is set equal to 12,650 calendar years BP (Fig. 3b). The match shows that the AMS ^{14}C dates from Pudozh, including the diverging dates between the local varve years/couplets 1350–640, range well within the AMS ^{14}C/calendar-year curve and indicates that each couplet very likely represents 1 year. A higher or lower sedimentation rate for each individual lamina would cause the radiocarbon dates to be much more centred or wider spread. If we then adopt the best estimate of *c*. 12 650 and *c*. 11 500 calendar years BP for the Alleröd/Younger Dryas and Younger Dryas/Preboreal boundaries (Björck *et al*. 1996; Goslar & Madry 1998), couplets no. 1700 and 550 can be set equal to 12 650 and 11 500 calendar years BP, respectively. The whole 1933-year-long sequence would then encompass the time period between *c*. 12 933 and 11 000 calendar years BP or, late Alleröd to early Preboreal.

Discussion

The combined AMS ^{14}C and varve chronology for Pudozh shows that the deposition of these sediments occurred between approximately 12 900 and 11 000 calendar years BP, or between late Allerød and early Preboreal. The consistency of the radiocarbon ages in the bottom part of the sequence shows that the sedimentation started several thousand years later than previously estimated based on palaeomagnetic correlations (Bakhmutov & Zagniy 1990; Ekman & Iljin 1991; Bakhmutov *et al*. 1994). If the age assignment of the different ice-marginal deposits is correct (Ekman & Iljin 1991), the sediments at Pudozh cannot have been deposited in connection with the receding ice margin during the Vepsovo/Krestets, Luga or Neva stages.

Four AMS ^{14}C measurements on terrestrial macrofossils from a varved clay sequence at Nizhnee Mjagrozero, which is situated in the northern part of Lake Onega (Saarnisto *et al*. in prep.) (Fig. 1), gave ages of between 11 600 and 11 300 ^{14}C years BP (Table 3b). These radiocarbon dates, which cover a varve-year

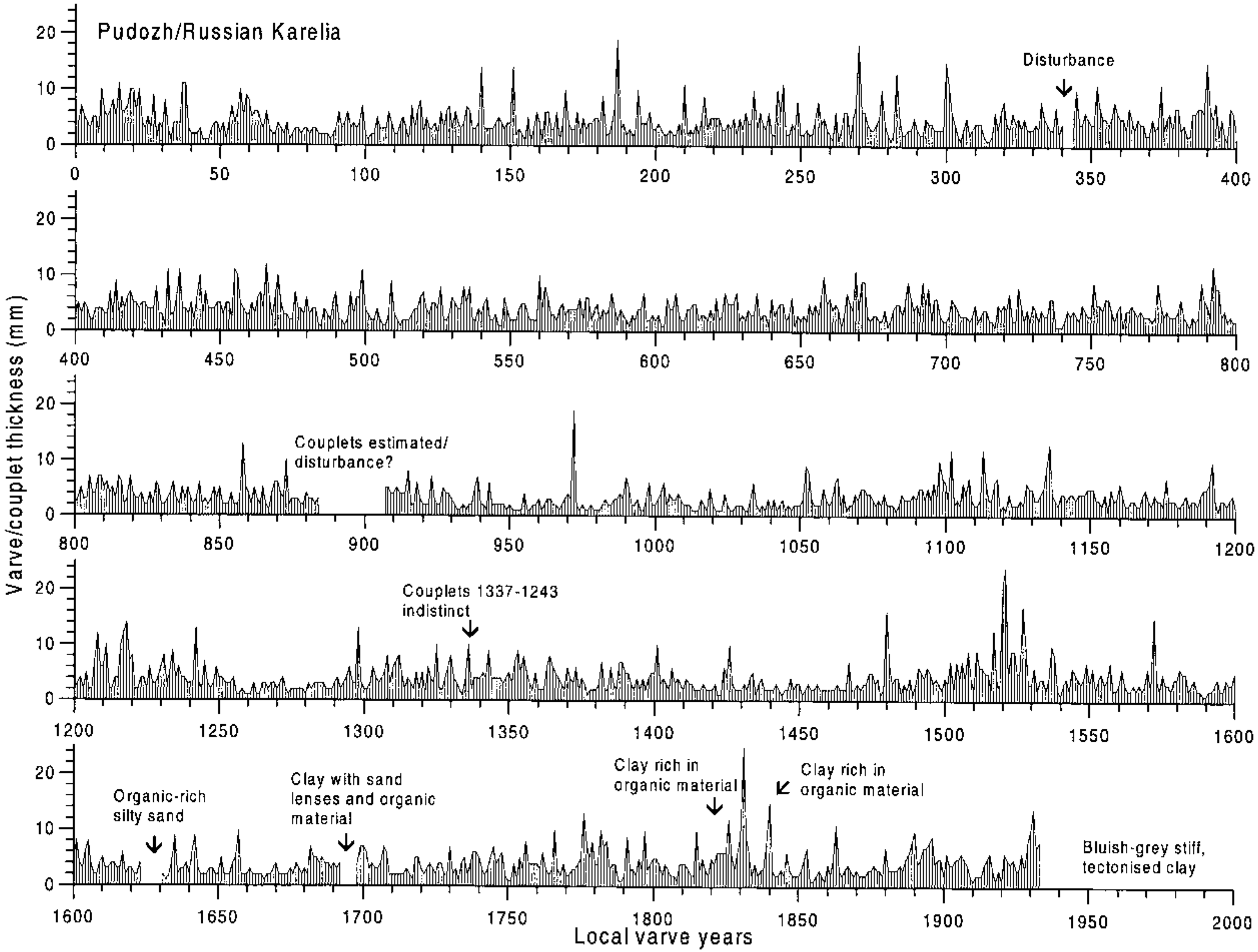

Fig. 2. Varve-thickness diagram from Pudozh. The disturbed sections and indistinct laminations are indicated.

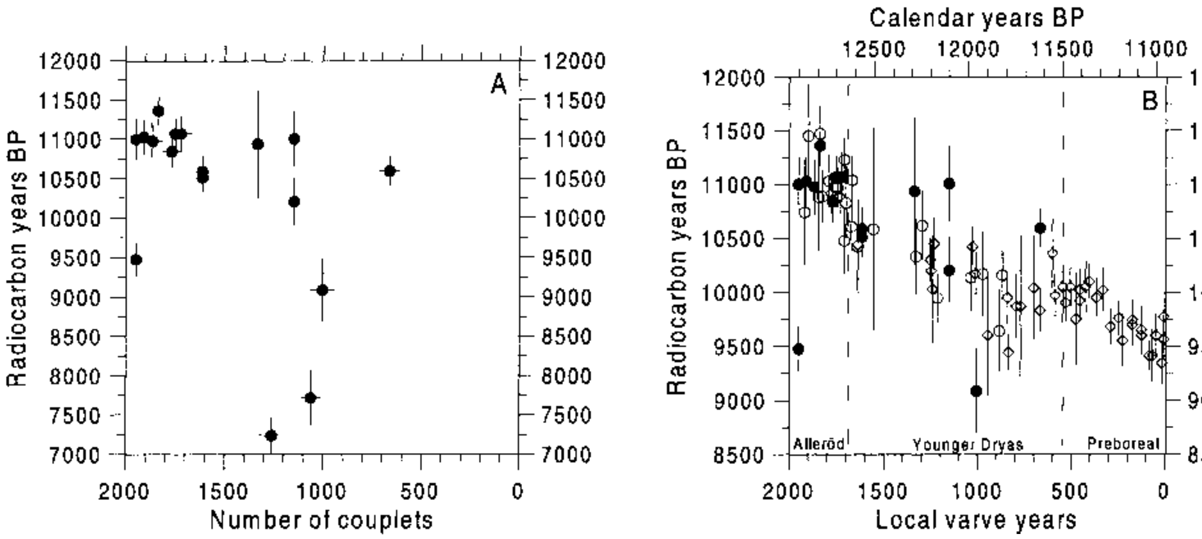

Fig. 3. □A. AMS ^{14}C dates on terrestrial plant macrofossils (filled circles) from the sequence Pudozh displayed with double standard error □B. compared to the AMS ^{14}C dated, annually laminated sequence from Gosciaz in Poland (open diamonds) (Goslar *et al.* 1995) and to the Swedish varve chronology (open circles) (Goslar *et al.* 1999). All dates are displayed with double standard error.

interval of 110 years, are clearly older than those obtained from the Pudozh sequence and show that the northern part of Lake Onega had already become ice-free several hundred years before varve deposition started at Pudozh.

Compared to glaciolacustrine varved clays, where often only the proximal/bottom deposits show sandy summer layers, the 1933-year-long sequence at Pudozh is characterized by silty sandy summer layers (Table 1). The intercalated organic rich horizons with well-preserved terrestrial macrofossils in the lower part of the sequence show that the macrofossils cannot have been transported very far and that the deposition of the sediments must have occurred rapidly, close to the contemporaneous shore and in fairly shallow water. The rare occurrence of limnic plants and animal remains in the macrofossils assemblage could be explained by a low transparency of the water column, as indicated by the high minerogenic content of the sediments. The distinct silty sand/clay laminations point to seasonal variations in sediment input, where the coarser fraction is deposited during the melting season, e.g. from dead ice remaining in the catchment, and the fine fraction after the melting peak and/or during cold months, when the water body is ice covered. The intercalated organic layers in the lower part of the sequence, which were rich in terrestrial macrofossils, may be interpreted as deriving from unstable soils close to the shore.

Estimates for the lake level of Lake Onega in the Vodla River/Pudozh area during the deglaciation and the early Holocene vary according to different authors. Devyatova (1986) places the Younger Dryas and Preboreal lake levels at around 57 m and 45–50 m, respectively. Saarnisto *et al.*'s (1995) shoreline diagram, on the other hand, implies a lake level of *c.* 33 m a.s.l., already between *c.* 10 100 and 9400 ^{14}C years BP. Following these and other studies (Zemljakov 1936; Kakkuri 1993), the direction of the isobases of the present-day land uplift and the tilt axes of Lake Onega run along a NNE–SSW line, from the mouth of the River Vodla to the outlet of the River Svir. This implies that the Vodla River/Pudozh area, relative to the other parts of Lake Onega, did not experience any major isostatic changes. The studied section in Pudozh extends between approximately 45 and 37 m a.s.l.. Following our adapted chronology, laminated sediments were deposited between *c.* 12 900 and 11 000 calendar years BP. Furthermore, based on an estimated sedimentation rate of *c.* 4–4.5 mm/laminae/year, the uppermost non-laminated 134 cm of the sequence may have been deposited during a time interval of approximately 300 years. This means that sediment deposition ceased completely at around 10 700 calendar years BP, when the water level fell below 45 m. If we adopt Saarnisto *et al.*'s (1995) lake level estimates (*c.* 33 m between 10 100 and 9400 ^{14}C years BP) for the Vodla River/Pudozh area, the laminated sediment sequence could not have been deposited in the Lake Onega basin. However, following Devyatova's (1986) lake level estimates, the area west and northwest of Pudozh may have been part of a large estuary of the Vodla River during the Younger Dryas and early Preboreal.

The vegetation around the site during late Allerød until early Preboreal was characterized by open, treeless dwarf shrub heaths with shade-intolerant species, such as *Betula nana, Dryas octopetala, Arctostaphylos alpina, Melandrium angustiflorum* and also *Distichium* sp. and *Ditrichum* sp. Some of these species lived on

Table 3a. AMS ^{14}C dates from the laminated sequence at Pudozh (L = leaves and leaf fragments; N = nutlet; C = catskin scale; T = twig; F = flower; S = seed; UIP = unidentified terrestrial plant fragments; * = sample not pretreated; # = inferred value).

Laboratory no. (Ua-)	Sample no.	Macrofossil content	Couplet	Radiocarbon age (yr BP)	$\delta^{13}C$ (‰ PDB)
12392	P-14	640–689	*Dryas octopetala* (L), *Betula nana* (N, C, L)	10 595 ± 90	−29.00
11587	P-19	975–1936	*Betula nana* (N; C; L), UIP	9085 ± 195	−28.74
11588	P-20	1039–1086	*Dryas octopetala* (L), *Betula nana* (N, C), UIP	7715 ± 170	−27.80
12388	P-22b	1150 ± 10	*Alnus* sp. (T)	11 005 ± 175	−29.50
	P-22a	1150 ± 10	*Dryas octopetala* (L)	10 205 ± 150	
12393	P-24	1230–1291	*Salix* sp. (T)	7245 ± 115	−29.00
12822	P-27	1319–1350	*Dryas octopetala* (L), *Betula nana* (N, C, L), *Salix* sp. (L), UIP	10 935 ± 340	−28.55
12394	P-32a	1599–1629 Organic lens	*Dryas octopetala* (L)	10 590 ± 100	−26.69
11501	P-32	1599–1629 Organic lens	*Dryas octopetala* (L), *Betula nana* (L)	10 515 ± 95*	
11589	P-35	1697–1748	*Betula nana* (N, C, L), *Salix* sp. (L)	11 070 ± 110	−29.19
12389	P-36	1748–1749 Organic lens	*Dryas octopetala* (L, F), *Betula nana* (N, C, L), *Salix* sp. (L)	11 070 ± 95	−28.90
12390	P-37	1749–1792	*Dryas octopetala* (L), *Betula nana* (N, L) *Salix* sp. (L, F)	10 845 ± 95	−29.09
12391	P-39	1831–1842	*Dryas octopetala* (L), *Betula nana* (L), *Salix* sp. (L)	11 365 ± 95	−29.15
12823	P-40	1843–1892	*Dryas octopetala* (L), *Betula nana* (C, S, L) ?*Salix* (L)	10 975 ± 95	−28#
11590	P-41	1893–1932	*Dryas octopetala* (L), *Betula nana* (N, C, L), *Salix* sp. (L)	11 030 ± 110	−28.15
11536	P-42b	Tectonized clay	*Dryas octopetala* (L)	9475 ± 105	
12825	P-42b, c	Tectonized clay	*Betula nana* (C, N), *Salix* sp. (L)	11 000 ± 125	−27.87

Table 3b. AMS ^{14}C dates from Nizhnee Mjagrozero (L = leaves and leaf fragments; C = catskin scale; F = flower).

	Sample no.	Macrofossil content	Radiocarbon age (yr BP)	$\delta^{13}C$ (‰ PDB)
10966	NMO 9–11	*Salix* spp. (C, L), *Dryas octopetala* (L)	11,530 ± 145	−29.20
10967	NMO 13	*Salix* spp. (F, L), *Dryas octopetala* (L)	11,505 ± 90	−29.63
10968	NMO 17	*Salix/Betula* (L), *Dryas octopetala* (L)	11,325 ± 95	−29.67
10969	NMO 18–19	*Salix/Betula* (L), *Dryas octopetala* (L)	11,570 ± 310	−29.97

well-drained soils, while sedges, grasses and bryophytes grew on poorly drained soils.

A few macrofossil finds of *Alnus* cf. *glutinosa* in late Alleröd sediments show that this species was an early immigrant to this part of Europe. In the studied material, two nutlets of *Alnus* were found, which have been tentatively referred to *A. glutinosa*. Further west in Europe, *Alnus glutinosa* arrived in early Holocene, but based on pollen stratigraphical work it has been suggested that *Alnus* arrived in Russian Karelia already in Allerød (Elina & Filimonova 1987, 1996). Most of the pollens in these pollen stratigraphic investigations are referred to as *Alnus incana*, but some were also identified to *A. glutinosa* and *A. fruticosa*. The finds of *Juniperus communis* in the later part of Younger Dryas are in accordance with several previous macrofossil records from northwestern Europe and show that juniper immigrated early.

No detailed pollen stratigraphic investigations covering the later part of the Weichselian and the early Holocene have so far been published from the area east of Lake Onega. The closest records, which are from the area west and northwest of Lake Onega (Elina & Filimonova 1996), suggest that the vegetation during the Allerød was dominated by scattered forests of *Betula pubescens*, with admixtures of *Alnus incana* and *Picea*. The tundra vegetation was composed of *Betula nana* and periglacial complexes included *Artemisia* and Chenopodiaceae. During Younger Dryas, the vegetation changed into a tundra environment with *Betula nana, Lycopodium dubium* and *Diphasiastrum alpinum* (Elina & Filimonova 1996). The Preboreal is characterized by north-taiga forests with *Betula pubescens*, a tundra landscape with *Betula nana*, *Salix*, *Lycopodium dubium* and *Diphasiastrum tristachium.*

In contrast to pollen stratigraphical studies (Elina & Filimonova 1996), the macrofossil results from Pudozh indicate that the vegetation was essentially unchanged from Allerød into early Preboreal. This finding is in line with that of Bondestam *et al* (1994), among others, who found that arctic steppe-tundra and tundra prevailed in Finnish Karelia from late Younger Dryas until mid-Preboreal, but here macroscopic remains of *Betula nana* were rare in Younger Dryas sediments. The different picture indicated by our results compared to the pollen stratigraphical studies can be explained by the presence of redeposited or long-distance-transported pollen, which hampers the interpretation of pollen assemblages. The proportion of redeposited or long-distance-transported pollen is greater in late-glacial sediments deposited during a time with unstable soils and a more or less tree-less vegetation, and studies of plant macrofossils are an indispensable complementary tool for the reconstruction of the past vegetation (e.g. Birks 1993; Dinter & Birks 1996).

Little can be said about the climatic conditions on the basis of the macrofossil results. At the present time, *Dryas octopetala* is rarely found south of the arctic tree-line, but it thrives well in fairly warm climates if it gets enough light. It appears to be the competition for light rather than high temperature that determines the southern limit of this species. The most warmth-demanding plants represented are *Juniperus communis* and *Alnus*, but both have geographical ranges extending north of the arctic tree-line. *Juniperus communis* indicates a mean summer temperature above *c.* 7°C, and *Alnus* indicates a mean July temperature above *c.* 8°C, when the modern geographical ranges of these taxa are compared with temperature maps.

The occurrence of organic-rich horizons during late Allerød/earliest Younger Dryas (between couplets 1600 and 1900) points to unstable soils and increased erosion in the surrounding catchment, which may be related to the melting of stagnant ice and indirectly to warmer temperatures before 12 550 calendar years BP. However, no change in laminae thickness can be observed coinciding with the cessation of catchment soil erosion. A distinct decrease in thickness is only visible between couplets 1500 and 1400 corresponding to *c.* 12 350–12 450 calendar years BP and between couplets 1100 and 900 or *c.* 11 850–11 950 calendar years BP. From couplet 800 upwards, a two-step increase in laminae thickness can be observed, which coincides approximately with 11 750 and 11 450 calendar years BP. Based on the age model developed for the sequence, this two-step increase may be paralleled with the last *c.* 250 years of Younger Dryas and with the beginning of the Holocene, respectively.

Conclusions

AMS ^{14}C and macrofossil analyses have been performed on a 1933-year-long laminated sequence from Pudozh in eastern Russian Karelia. The AMS ^{14}C dates show that deposition of these sediments started during the late Allerød, i.e. when the northern part of Lake Onega had already become ice-free. The site cannot therefore indicate the position of the retreating active ice margin.

The comparison with other AMS ^{14}C dated, high-resolution sequences indicates that the laminations are very likely true varves and that the whole sequence spans between *c.* 12 900 and 11 000 calendar years BP, i.e. late Allerød until early Preboreal.

The silty sand and clay laminae, which can be observed throughout the whole sequence, do not resemble typical glaciolacustrine varved clays. Their deposition may instead be due to the seasonal melting of dead ice in the catchment and sediment transport into a shallow basin. Since published estimates for water level changes in Lake Onega are controversial, it cannot be determined whether the basin in which these varves were deposited was connected to Lake Onega.

Based on the macrofossil analysis, we can conclude that the local vegetation, which consisted of open, tree-less dwarf shrub heaths with shade-intolerant species, was essentially unchanged from late Allerød to early Preboreal. Increased soil erosion during late Allerød/early Younger Dryas is indicated by organic-rich horizons in the lower part of the sequence.

Our investigations, which have provided the first AMS ^{14}C dates for a late glacial sequence in Russian Karelia, show the importance of an independent and well-established chronology before attempting any type of correlation. They also emphasize the need for macrofossil studies as a necessary complementary tool for the reconstruction of the past vegetation.

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