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SPATIO-TEMPORAL DEVELOPMENT OF THE SOOMAA MIRE SYSTEM IN SW ESTONIA

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Abstract. The spatio-temporal development pattern of a mire system and major factors affecting this process were studied. The Soomaa mire system (25 100 ha) in south-western Estonia is one of the largest and best preserved mire systems in Estonia. Using the bulk densities of peat layers and radiocarbon dates average accumulation rates and increments of different peat types were calculated. On the basis of the obtained and already existing stratigraphical data the probable development of the mire system was reconstructed. Paludification of the Soomaa region started about 10 000 BP and ombrotrophication 7000–6000 BP. The expansion rate of the mire system was rather stable during the whole Holocene and no distinct relationship between the climate and the expansion of mires was detected. The results suggest that the expansion of mires was mostly an autogenous process, while ombrotrophication might have been favoured also by climate changes. The spatial dynamics of paludification was mainly determined by local factors – topographical and hydrological characteristics of the area.

Key words: mire system, expansion of mires, paludification, ombrotrophication, Estonia.

INTRODUCTION

Mires are accumulative ecosystems that are characterized by peat deposition. This property gives mires an ability to change their relief, water regime, and trophicity, but also certain autonomy in their development. Bogs are described as externally controlled systems, which reflect climatic changes (Granlund, 1932), but also as systems capable of strong autoregulation. The lateral expansion caused by runoff water collection and thus intensive paludification in lagg areas is considered here as an autogenous process. Also local topographical and hydrological factors are assumed to play an important role in mire development. Recent studies (Mäkilä, 1997) suggest that autogenous factors may interact with local site factors and climate.

Under excessively moist climatic conditions, which predominate in the northern temperate zone, extensive mire systems may form in the presence of favourable geomorphological features of the subsoil and a favourable geological structure of the Quaternary cover (Kiryushkin, 1980). The mire systems may, in turn, start to influence the general hydrological regime of the region. However, to detect the impact of developing mire systems on the environment, we should know the regularities of their formation and growth.

The aim of the present study was to reconstruct the probable development process of one of the largest and best-preserved mire systems in Estonia, i.e. to establish the spatio-temporal dynamics of paludification and succession of mire types. In particular, the goal was to determine the role of climatic, local, and autogenous factors in the development of mires, i.e. to find out how climate in the Holocene, local (topography, edaphic properties of subsoil, hydrological regime of the region, etc.), and autogenous (tendency of mire massifs to grow and expand because of discharge waters) factors have affected the expansion of mires and formation of mire systems.

STUDY AREA

The mire system studied is located on the territory of Viljandi and Pärnu counties in SW Estonia (Fig. 1) and belongs to Soomaa National Park (36 700 ha). The study area consists of the main part of the mire system (except Riisa mire). Soomaa, which belongs to the catchment of the Pärnu River, is part of the Pärnu Lowland in Lower Estonia (Kink, 1996). It borders on a slope of the Sakala Upland in the east and south-east. According to the geobotanical subdivision of Estonia, the study area belongs to the Intermediate Estonian region of bogs and swamp forests (Laasimer, 1965) and, according to the classification of mire districts, to the district of large bogs in south-west Estonia (Masing & Allikvee, 1988). Soomaa includes large mires but also transit rivers (Halliste, Raudna, and Lemmjõgi), floodplains, and forests between the mires. The region represents a plain dissected by shallow river beds and, in places, by steep-sloped mire massifs rising above the surrounding area. The predominating height of the study area is 20–30 m a.s.l. The bog massifs rise up to 9 m above the surrounding area. The mineral surface is inclined to the west, north-west, and north (Kink, 1996).

Ice retreated from the Soomaa area with the retreat of a glacier from the belt of marginal formations of the Sakala Stage before the Pandivere Stage, around 12 500–13 000 BP (Kalm et al., 1994). During the retreat of the ice a local ice-dammed lake formed between the glacier and the Sakala Upland, and for several centuries the whole study area was covered with water. Around 11 200–10 600 BP this area was part of the Baltic Ice Lake. During that period sediment accumulation levelled markedly the Late-Glacial relief of Soomaa.

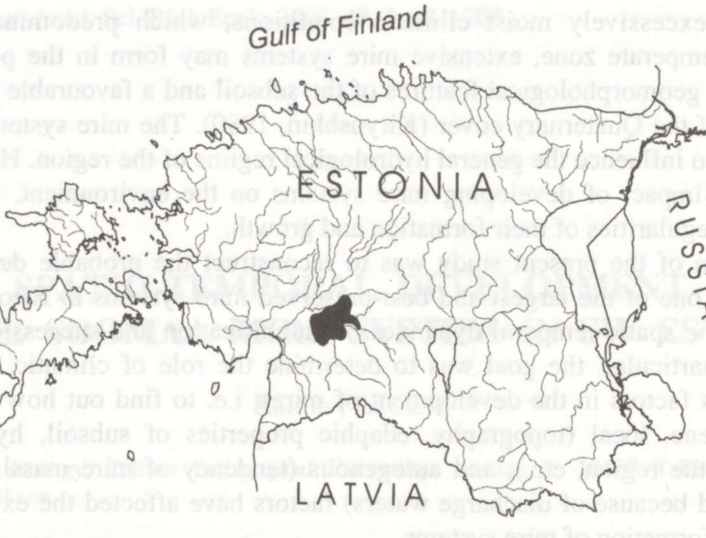

Fig. 1. Location of the Soomaa mire system.

After the regression of the Baltic Ice Lake, a network of rivers developed and the formation of vegetation and soil started. In depressions lakes or other shallow water bodies preserved (Kalm et al., 1994).

In the Quaternary cover glaciolimnic fine-grained sandy-loamy sediments predominate (sand, sandy loam, loam, clay), also till occurs in places (Kalm et al., 1994). At the present time most of the area is covered with peat. The thickness of the peat layer is up to 11 m at Soomaa. The rivers (Halliste, Raudna, Lemmjõgi) have a small gradient and the river beds are shallow. This causes certain slowing of the currents in the rivers that originate from the Sakala Upland, an increase in their water level, and periodic floods in the area.

As Soomaa is geobotanically situated in Intermediate Estonia, which is the transitional area between West and East Estonia, also the bogs have a transitional character there. However, judging from the morphology of massifs, the bogs of the study area are of West Estonian type (plateau-like massifs); their plant cover shows also some features similar with East Estonian bogs (Allilender, 1990). Of the mire area (25 060 ha), 73% is represented by bog communities, 21% by mixotrophic mire communities, and 6% by fen communities. Widespread are extensive treeless and wooded hollow-ridge and pool-ridge complexes, also bog forests, dwarf shrub bog, and grassy bog occur often. At the margins of the massifs transitional mire areas, fens, and swamp forests are predominant.

Soomaa comprises four big mires: Kuresoo, Kikepera, Öördi, and Valgeraba.

Kuresoo mire (11 000 ha), which takes up the northern and north-eastern parts of the region, is the largest mire at Soomaa. It is situated on a plain rising towards the north-east; there is practically no depression under the mire. Bog communities spread on 9580 ha and mixotrophic mire communities on 1520 ha (Veber, 1988). Kuresoo mire consists of 10 massifs (massif = hydrologically independent bog cupola) with areas of 300–1300 ha (on average 750 ha) and 1.6–6 km in cross-section (Loopmann, 1974). Water runoff from such a large mire area takes place through several mire streams and outflows between the massifs and mire funnels in the middle of the mire which are connected to overgrown mire streams. The surfaces of the massifs are mainly flat and their slopes are steep. The marginal Karuskose slope of the bog lying in the south-western margin of Kuresoo reaches up to 8 m. The average thickness of the peat deposit is 3.2 m (Loopmann, 1974); in bog areas the thickness of peat is mostly 3–6 m, in the south-western part up to 8.6 m.

Kikepera mire (6900 ha) is situated in the western and south-western part of Soomaa. Bog, mixotrophic mire, and fen communities occur on 4420, 1740, and 740 ha, respectively (Veber, 1988). The mire comprises seven massifs of West Estonian type with steep slopes and flat plateaus. The northern part of Kikepera is split up by the Jõhve Brook, which starts from the middle of the mire. In the northern part lies Kikepera residual lake (2 ha). The southern part of Kikepera, which consists of two massifs, is connected to the northern part by a fen and mixotrophic mire area, where the peat layer is thin. The greatest thickness of the peat layer is 8.2 m (Veber, 1988).

Öördi mire (4910 ha) lies in the south-eastern part of Soomaa. Of the mire area, bog communities take up 2250 ha, mixotrophic mire communities 1890 ha, and fen communities 770 ha (Veber, 1988). Öördi mire consists of three massifs. In its eastern part Öördi residual lake, 5 ha in area, is located. Three mire streams originate from the contact area between the massifs. The peat layer is thickest (9.5 m) in the middle of the mire (Veber, 1988).

Valgeraba mire (2150 ha) lies in the eastern part of Soomaa. Most of its territory is covered by bog communities (2050 ha); mixotrophic mire takes up 100 ha (Veber, 1988). Valgeraba consists of two massifs, which are separated by a mesotrophic mire area. The western massif is flat. The eastern massif is a convex bog of eccentric shape and with steep slopes. The height of the eastern massif with respect to the mineral land lying to the north is up to 10 m and it has the thickest peat deposit (11 m) at Soomaa.

MATERIAL AND METHODS

To establish the thickness and stratigraphy of the peat deposits of mires, and to determine the relief and composition of the mire bottom, materials of the Hydrometeorological Survey of the Estonian SSR (1958) were used: these included maps of the topography of the mineral subsoil and of the composition and thickness of the peat deposits (scale 1:50 000), and sections of peat deposits (1:25 000 × 1:100) involving 15 profiles with a total length of about 110 km, compiled from the data of 137 stratigraphically studied points and 168 survey points (Archive of the Institute of Meteorology and Hydrology).

Also technical plans of peat deposits at a scale of 1:25 000 for Kuresoo and Öördi mires, and at a scale of 1:10 000 for Valgeraba mire, which contained information from 45 stratigraphically studied points and 291 survey points provided by the Geological Survey of Estonia (1982–83), were used.

The topography and hydrography of the study area were described using topographic maps (1:10 000, 1:25 000, and 1:50 000), and water runoff schemes (1:25 000 and 1:50 000), compiled by the Hydrometeorological Survey (Archive of the Institute of Meteorology and Hydrology). Data on the water regime of the rivers were obtained from hydrological yearbooks.

In each mire detailed investigation of the peat deposit, including ^{14}C datings, was performed at one site, in Valgeraba mire on both massifs (Fig. 2). The mire areas with the thickest peat deposition, i.e. probably the oldest mire areas, were selected as study sites. A Belarus peat corer (peat auger measuring 75 mm × 500 mm) was used. At each study site the layers (with a thickness of at least 2 cm) were distinguished according to the peat colour and texture. The depth, thickness, botanic composition (peat type), and decomposition degree (in %) of peat were detected macroscopically (according to Largina, 1977) for every visually distinguishable layer. All the described layers were sampled (422 samples altogether) to establish their moisture content.

Beds marking the beginning of paludification, transition into the ombrotrophic stage, etc. were sampled for ^{14}C dating (11 samples). ^{14}C datings were performed in the radiocarbon laboratory of the Institute of Geology at Tallinn Technical University. Dendrological corrections were calculated for datings using the program "Cal 15" (Table 1). The average values of the calibrated data ranges were used in calculations.

To determine the moisture content, the samples were dried at 60 °C until they acquired a constant weight. On the basis of the moisture content and decomposition degree, the relevent bulk density was found from special empirical tables. According to peat bulk density and calibrated radiocarbon dates, the mean peat increment rates and dry matter accumulation rates were calculated separately for bog and fen peats of each mire (Table 2). Assuming that peat accumulation was constant in time in the periods between radiocarbon datings, the increment of bog and fen peats was determined for each mire on the basis of the degree of decomposition. Using the obtained average peat increment rates, the ages of the peat layers were calculated (see Ilomets, 1984) for each stratigraphically studied point. The calculation of ages started from the surface of the peat deposit, i.e. from zero age. The ages of sublayers between the layers with a different degree of decomposition and the time of the start of peat accumulation and ombrotrophication were calculated. Also, the depths corresponding to the ages of 2000, 4000, 6000, and 8000 years were calculated and the age isolines of peat deposits were drawn on the stratigraphical profiles (Fig. 3). The contact points

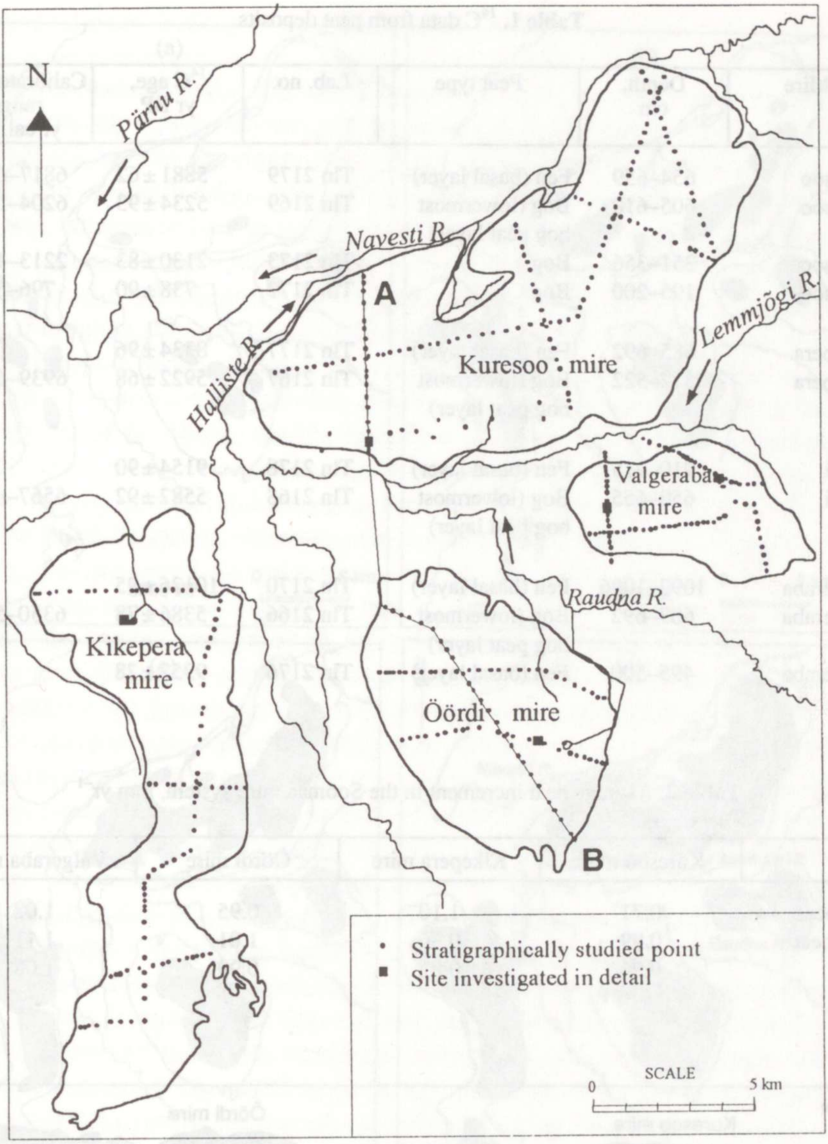

Fig. 2. Soomaa mire system.

of the isolines with the mineral subsoil coinciding with ancient mire borders were marked on the maps. The areas between the profiles were interpolated according to the depth of the peat. The proportions of bog and fen areas were measured from the stratigraphical profiles obtained (Fig. 4). The expansion rates were calculated on the basis of the model where mires were transformed into areas of circular shape.

Table 1. ^{14}C data from peat deposits

Mire	Depth, cm	Peat type	Lab. no.	^{14}C age, yr BP	Calibrated date range, yr cal. BP
Kuresoo	654–659	Fen (basal layer)	Tln 2179	5881 ± 62	6817–6621
Kuresoo	605–610	Bog (lowermost bog peat layer)	Tln 2169	5234 ± 93	6204–5856
Kuresoo	351–356	Bog	Tln 2173	2130 ± 83	2213–1923
Kuresoo	195–200	Bog	Tln 2172	738 ± 90	796–536
Kikepera	685–692	Fen (basal layer)	Tln 2177	8334 ± 96	
Kikepera	517–522	Bog (lowermost bog peat layer)	Tln 2167	5922 ± 68	6939–6633
Öördi	910–915	Fen (basal layer)	Tln 2178	9154 ± 90	
Öördi	650–655	Bog (lowermost bog peat layer)	Tln 2168	5582 ± 92	6567–6267
Valgeraba	1092–1096	Fen (basal layer)	Tln 2170	10126 ± 95	
Valgeraba	687–692	Bog (lowermost bog peat layer)	Tln 2166	5386 ± 78	6350–5974
Valgeraba	495–500	Fen (basal layer)	Tln 2176	9952 ± 28	

Table 2. Average peat increment in the Soomaa mire system, mm yr^{-1}

	Kuresoo mire	Kikepera mire	Öördi mire	Valgeraba mire
Fen peat	0.71	1.10	0.95	1.02
Bog peat	0.99	0.76	1.01	1.11
Total	0.96	0.82	0.99	1.08

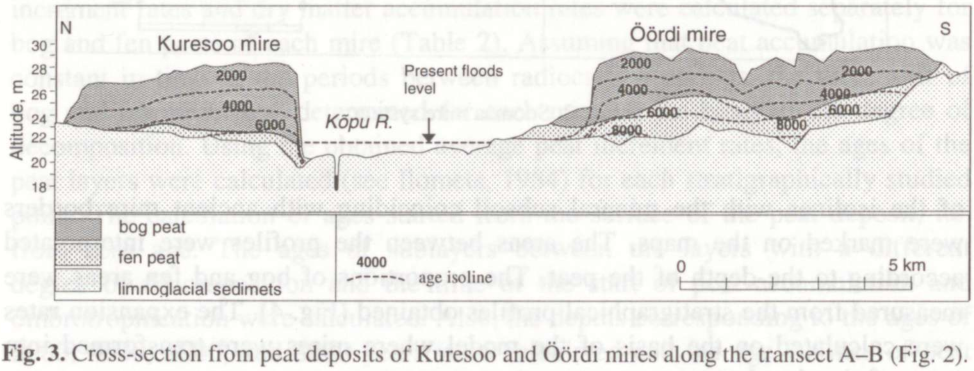

Fig. 3. Cross-section from peat deposits of Kuresoo and Öördi mires along the transect A–B (Fig. 2).

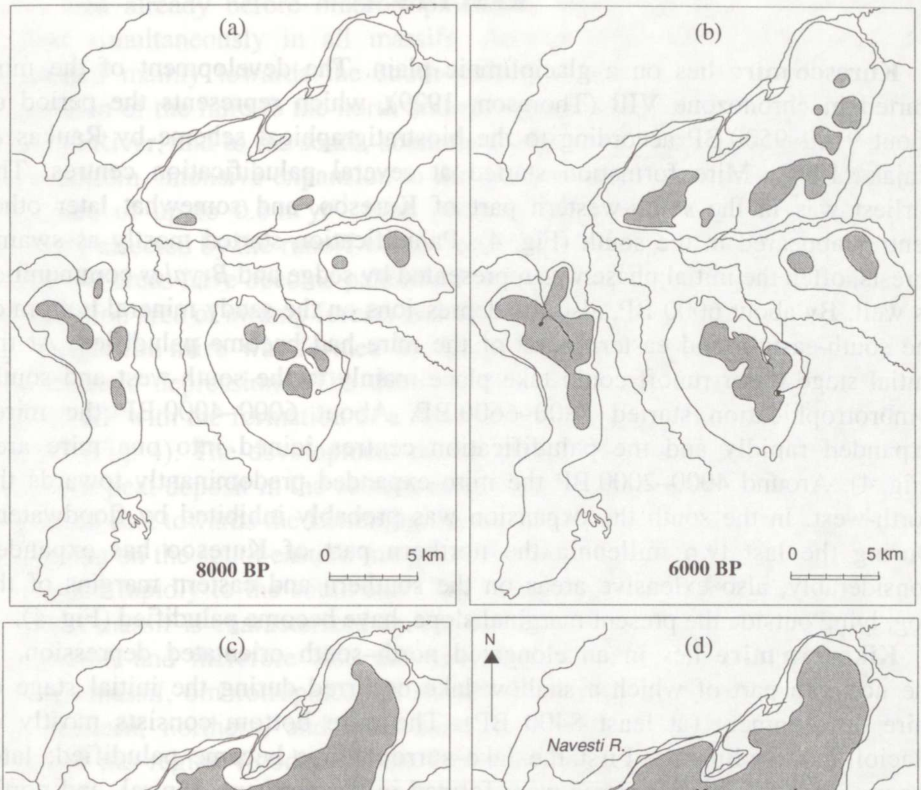

Fig. 4. Soomaa mire system 8000 BP (a), 6000 BP (b), 4000 BP (c), and 2000 BP (d).

RESULTS

Kuresoo mire lies on a glaciolimnic plain. The development of the mire started in chronozone VIII (Thomson, 1929), which represents the period of about 9100–9500 BP according to the biostratigraphical scheme by Raukas & Kajak (1997). Mire formation started at several paludification centres. The earliest was in the south-western part of Kuresoo, and somewhat later other centres appeared in the south (Fig. 4). Paludification started mostly as swamp forests, often the initial phase was represented by sedge and *Bryales* communities as well. By about 6000 BP, also the depressions on the sandy mineral bottom of the south-eastern and eastern parts of the mire had become paludified. At the initial stage water runoff could take place mainly to the south-west and south. Ombrotrophication started 7800–6600 BP. About 6000–4000 BP the mires expanded rapidly and the paludification centres joined into one mire area (Fig. 4). Around 4000–2000 BP the mire expanded predominantly towards the north-west. In the south the expansion was probably inhibited by floodwaters. During the last two millennia the northern part of Kuresoo has expanded considerably, also extensive areas on the southern and eastern margins of the bog, lying outside the present marginal slope, have become paludified (Fig. 4).

Kikepera mire lies in an elongated north–south orientated depression, in the northern part of which a shallow lake occurred during the initial stage of mire development (at least 8400 BP). The mire bottom consists mostly of glaciolimnic sediments. First the lake surroundings became paludified; later separate paludification centres were formed in the northern, central, and north-western parts of the present mire (Fig. 4). Mire development started with the formation of swamp forests, in isolated depressions *Bryales* fen developed. Runoff took place predominantly to the north along the Jõhve Brook. The centres widened rapidly on the flat mineral surface and joined up. Ombrotrophication started about 6800 BP, and 6000–4000 BP the mire expanded rather rapidly (up to 0.5 m yr^{-1}) south- and westwards, towards the rising mineral subsoil (Fig. 4). In the east the mire reached the present marginal slope, but its further expansion was hampered by floodwater from the Halliste River. At the end of this period paludification started in several centres in the southern part of Kikepera mire. Between 4000 and 2000 BP expansion was greatest in the southern part of the mire (0.3–0.5 m yr^{-1}). During the last 2000 years the southern and central areas of the mire have expanded markedly and they have become connected.

The mineral bottom of **Öördi mire** is represented by a glaciolimnic plain with rising southern and south-eastern margins. In the eastern part of the mire depression a shallow lake existed. Mire formation started about 9200 BP with the paludification of the area north-west of the lake. About 200–500 years later paludification centres appeared in the north-eastern, northern, and north-western parts of the mire (Fig. 4). The initial phase of the mire was dominated by *Bryales* fen. Paludification centres widened rapidly 8000–6000 BP and joined into one mire area already before ombrotrophication, which started *c.* 6300–6600 BP, almost simultaneously in all massifs. Around 6000–4000 BP the mire area expanded mainly towards the south-west at a rate of up to 0.4 m yr^{-1}. The expansion of the mire to the north and north-east was hampered by floods of the Raudna River, and to the south, south-east, and east by the rising gradient of the mire bottom. Intensive expansion to the north-west, west, south-west, and south at a rate of up to 0.6 m yr^{-1} took place *c.* 4000–2000 BP. This process was strongly affected by the runoff waters from the massifs. During the last millennia extensive areas have become paludified outside the present marginal slope in the distribution area of swamp forests and wooded fen communities.

Valgeraba mire was formed by the paludification of two separate shallow glaciolimnic depressions. The development of the mire started about 10 300–10 100 BP with the formation of a *Bryales* fen in the northern part of the eastern massif (Fig. 4). The development of the western massif began somewhat later. Soon the peat deposit in the eastern massif started to inhibit the water flow from the south-east towards the Lemmjõgi River. Thus the water level rose in the areas bordering on the south-eastern margin of the massif and the paludification centre expanded rapidly to the south-east. The initial phase in the development of the western massif is characterized by the distribution of swamp forest, where peat increment, and therefore also the widening of the centre, was slow. On the eastern massif, ombrotrophication started about 6300–6100 BP. By *c.* 6000 BP the western, northern, and north-eastern margins of the eastern massif had reached the region flooded by the Lemmjõgi River and the expansion of the massif in these directions had stopped. Around 6000–4000 BP the eastern massif widened rapidly (0.6 m yr^{-1}) southwards; also the western massif expanded markedly. At the end of this period, after a fire (charcoal-rich layer), ombrotrophication started on the western massif. Around 4000–2000 BP the western massif widened rather rapidly (0.3–0.5 m yr^{-1}) to the east and south-east. The process was favoured by an intensive vertical growth of the bog massif, which led to an increasing outflow from the massif to the surrounding mineral land. During the recent millennia the area between the two massifs has become paludified, extensive expansion has taken place also at the eastern margin of the eastern massif and at the northern margin of the western massif.

DISCUSSION

In Estonia intensive paludification started 8500–8000 BP (Allikvee & Ilomets, 1995). In the context of the general paludification of Estonia, the Soomaa region became paludified relatively early, indicating extremely favourable local conditions for mire formation: glaciolimnic sediments with low water permeability and a low surface gradient. Considering the dynamics of paludification on the mire system level, we summarized the parameters for individual mires and thus obtained the averaged values. The expansion of the mire system in time has been relatively regular, without any interruptions or abrupt accelerations (Fig. 5). The rate of paludification has rather constantly increased from 2.24 ha yr^{-1} (10 000–8000 BP) to 4.06 ha yr^{-1} (recent millennia). At the same time the lateral growth (m yr^{-1}) of mires has generally decreased. The expansion of mires (Fig. 6) was greatest (0.63 m yr^{-1}) at the intial phase of mire development (10 000–8000 BP), slowest 6000–4000 BP (0.22 m yr^{-1}), and it has been rather stable during the last 4000 years (0.34 m yr^{-1}). In spite of a very slow lateral expansion rate 6000–4000 BP, the area of the mire system increased markedly because of the great number of massifs (mire areas). It may be concluded that a relatively

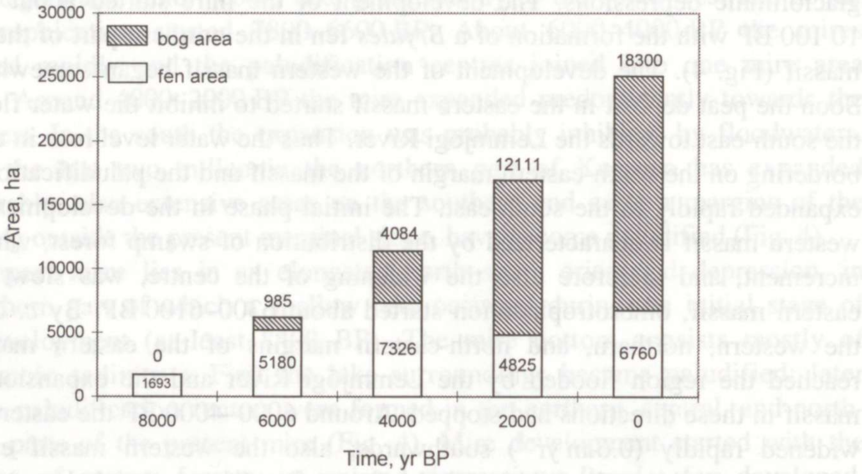

Fig. 5. Expansion and ombrotrophication of the Soomaa mire system.

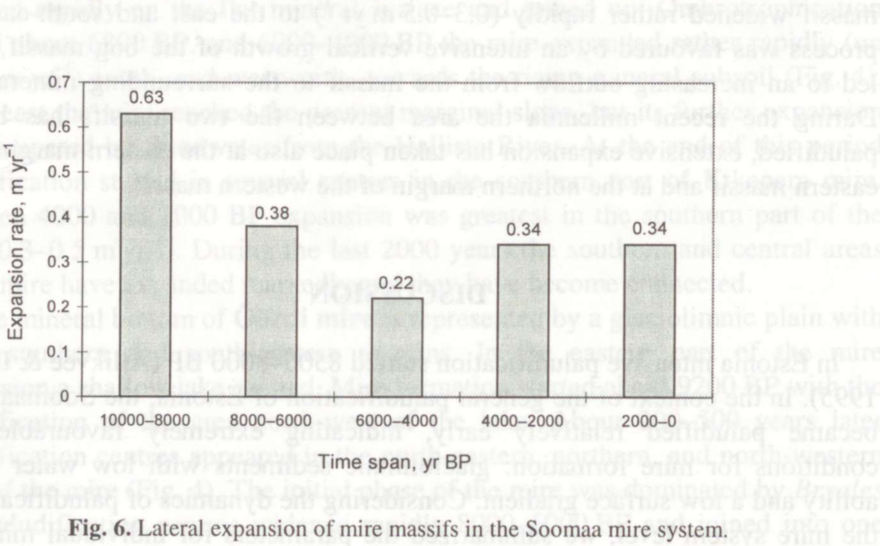

Fig. 6. Lateral expansion of mire massifs in the Soomaa mire system.

rapid expansion of mires took place in recent millennia. This conclusion, however, may be biased by possible underestimation of the age of fen and mesotrophic mire areas (particularly of swamp forests), distributed at the marginal parts of the massifs.

Paludification dynamics

The spatial dynamics of paludification shows somewhat more distinct trends compared with temporal one. All paludification centres formed in flat areas or shallow depressions, at a sufficient distance from running waters. The expansion of the mire system was more intensive in the direction of the margins of the area, against the rise in the mineral subsurface relief. The expansion was slower towards the rivers, and often the mire margin remained at the same line for several millennia (Fig. 4). Such a spatial course of paludification may undoubtedly be explained by local geomorphological and hydrological conditions. The location of the paludification centres at a moderate distance from flowing waters and their slow expansion towards the rivers were probably caused by periodic floods with river water, which inhibited peat accumulation. Peculiarities of the water regime may also account for the formation of steep bog slopes: intensive floods had stopped the expansion of mires for several millennia; however, the height of the mire area continued increasing. Because of permanent peat accumulation, the site that became paludified first rose higher above the surrounding area and was released earlier from the impact of floods that inhibited peat deposition. Therefore, ombrotrophication took place earlier on these sites and they became the centres of bog massifs. Thus, the peculiarities of the topography, water regime, and composition of the mineral subsoil have greatly determined the location of the present bog massifs. The expansion of mires towards the rise in the subsoil relief is connected with the damming up of the water by massifs. The mire areas, which had formed on plains, hampered the flow of the water from the highest areas towards the rivers. This caused a rise in water level and paludification of the areas bordering on the massifs and having a surface gradient decreasing towards the massif. The expansion of mires depended on the subsoil gradient and increasing height of massifs – against the rising subsoil the massifs could expand only as the peat deposit became thicker and the height of the massifs increased. Thus, for instance, during their development the eastern massif of Valgeraba mire and the southern part of Kikepera mire have risen 7–8 m against the inclination of the subsoil. This was possible only thanks to a rather rapid peat increment. Considering that mires expanded intensively during the entire Holocene in spite of the reduced area suitable for paludification, we may suggest a strong autogenous control over the paludification process, i.e. the expansion of massifs, supported by the waters flowing off from these massifs, seems to be a largely autogenous process.

Vegetation development

The development of mires has most often (38% of cases) been initiated by tree–reed communities (swamp forests). This refers to nutrient-rich conditions, which could have been caused by the peculiarities in the local water regime (periodic floods). Often (27% of cases) the development started as a *Bryales* fen. In this way the shallow closed depressions, now forming the centres of many older massifs, became paludified. In 16% of the cases mire development began as a *Carex* fen.

Because of peat increment the areas paludified earlier turned into topologically higher regions. Therefore these areas also underwent ombrotrophication earlier and in the course of later development became centres of bog massifs. Ombrotrophication of the oldest parts of the four mires studied took place almost simultaneously, 6000–7000 BP, which according to Allikvee & Ilomets (1995) coincides with the first period of bog formation in Estonia (7000–6500 BP). Bog formation could be related to warm and relatively dry beginning of the Atlantic climatic period, which caused a lowering in the water table and therefore supported the invasion of *Sphagnum* mosses. However, there are numerous mire areas that reached the ombrotrophication phase later at different times, which shows that ombrotrophication is a result of mire development, i.e. it is determined by local development processes. The long-lasting fen phase in some mire areas (4000 years on the eastern massif of Valgeraba mire, 2700 years in the central part of Öördi mire) may be explained by local conditions – ombrotrophication was inhibited by the flow of nutrient-rich waters down from higher mineral lands or by floodwater.

Often a charcoal-rich peat layer, indicative of a fire on the mire, precedes in peat deposits the bed marking ombrotrophication. A fire changes strongly hydrological and edaphic conditions, and supports the invasion of *Sphagnum* mosses and accelerates ombrotrophication processes (Tolonen, 1987). Thus, the fires that have occurred in the study area may be considered a factor accelerating ombrotrophication. In some cases also paludification of mineral land is connected with fires.

CONCLUSIONS

Paludification started on the study area very early (10 300–10 100 BP), immediately after the regression of the ice-dammed lake, which indicates favourable conditions for paludification. The formation of the extensive Soomaa mire system has been strongly influenced by local factors, particularly by the topographic and hydrological peculiarities of the site. These factors also determined the spatial dynamics of paludification, the location of mire massifs, and the formation of mire slopes. The relatively even increase in the area of mires in time has been caused by autogenous development of massifs. The transition of mire areas into the ombrotrophic bog stage is first of all a result of autogenous development of the mire, though also some relations to climate were established.

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REFERENCES

Allikvee, H. & Ilomets, M. 1995. Sood ja nende areng. In *Eesti. Loodus* (Raukas, A.,comp.), pp. 327–347, Valgus, Eesti Entsüklopeediakirjastus, Tallinn.

Allilender, K. 1990. Pärnu madaliku idaosa rabade (Kuresoo, Valgeraba, Kikepera, Öördi raba) areng ja taimkate. Diploma paper, University of Tartu (manuscript in the library of the Institute of Geography, University of Tartu).

Granlund, E., 1932. De svenska högmossarnas geologi. Deras bindningsbetingelser, utvecklingshistoria och utbredning jämte sambandet mellan högmossbildning ach försumpning. *Sveriges Geologiska Undersökning. Ser. C. Avhandlingar ach uppsatser. Årsbok 26*, 1. Norstedt & Söner, Stockholm.

Ilomets, M. 1984. On the cyclical nature of the development of bogs. In *Estonia. Nature, Man, Economy*, pp. 68–77. Academy of Sciences of the Estonian SSR, Estonian Geographical Society, Tallinn.

Kalm, V., Kink, H., Andresmaa, E., Orru, M. & Ainsaar, L. 1994. Soomaa inventeerimine. Aruanne (manuscript at the Institute of Geology, TU, Tartu).

Kink, H. 1996. *Eesti kaitsealad – geoloogia ja vesi*. Estonian Acad. Publ., Tallinn.

Kiryushkin, V. N. 1980. *Formation and Development of Mire Systems*. Nauka, Leningrad (in Russian).

Laasimer, L. 1965. *Eesti NSV taimkate*. Valgus, Tallinn.

Largina, I. F. 1977. *Peatlands and Their Investigation*. Nedra, Moskva (in Russian).

Loopmann, A. 1974. Kuresoo. *Eesti Loodus*, 5, 272–278.

Masing, V. & Allikvee, H. 1988. Eesti soode valdkonnad. In *Eesti sood* (Valk, U., ed.), pp. 247–275. Valgus, Tallinn.

Mäkilä, M. 1997. Holocene lateral expansion, peat growth and carbon accumulation on Haukkasuo, a raised bog in southeastern Finland. *Boreas*, **26**, 1–14.

Raukas, A. & Kajak, K. 1997. Quaternary cover. In *Geology and Mineral Resources of Estonia.* (Raukas, A. & Teedumäe, A., eds.), pp. 125–136. Estonian Acad. Publ., Tallinn.

Thomson, P. 1929. Die regionale Entwicklungsgeschichte der Wäldes Estlands. *Acta comment. Univ. Tartuensis. A* XVII, Dorpat.

Tolonen, K. 1987. Natural history of raised bogs and forest vegetation in the Lammi area, southern Finland studied by stratigraphical methods. *Ann. Acad. Sci. Fenn. Ser. A 3. Geol.–Geogr.*, 144.

Veber, K. 1988. Eesti soode loend. In *Eesti sood* (Valk, U., ed.), pp. 287–302. Valgus, Tallinn.

SOOMAA SOOSTIKU ARENGU AJALIS-RUUMILINE DÜNAAMIKA JA SEDA MÕJUTAVAD TEGURID

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On rekonstrueeritud Edela-Eestis paikneva Soomaa soostiku tõenäoline arengukäik. Selleks koguti soode turbalasundeist 422 proovi kuivaine sisalduse määramiseks, 11 proovi dateeriti ^{14}C-meetodil ning kasutati ka varasemaid andmeid soode turbalasundite ehituse kohta. ^{14}C-dateeringute ja turbakihtide kuivainesisalduse baasil arvutati keskmine turba akumulatsioonikiirus ja juurdekasv. Saadud näitajate ja varasemate stratigraafiliste andmete põhjal leiti lasundite vanus kõigis uuritud punktides (137) ning taastati soode laienemise tõenäoline ajalis-ruumiline käik. Piirkond vabanes Balti jääpaisjärve vete alt 10 600 a.t., soostumine algas 10 300–10 100 a.t. ja rabafaasi jõudsid esimesed sooalad 7000–6000 a.t. Ulatusliku soostiku kujunemine oli põhjustatud eeskätt piirkonna lokaalsetest teguritest, nagu tasasest pinnamoest ja vettpidavast pinnakattest, mis pärssisid vee äravoolu ning lõid soodsad tingimused soostumiseks. Soostumise ruumilise dünaamika määrasid samuti lokaalsed tegurid. Soode laienemise kiirus on olnud suhteliselt ühtlane läbi holotseeni. See võib olla põhjustatud rabalaamade laienemisprotsessi autogeensest iseloomust, s.t. laamadele omasest tendentsist laieneda kõrguskasvu ja laamalt valguvate vete toel. Kuigi madalsoo jõudmine rabafaasi on soo arengu loomulik tulemus, viitab uuritud nelja soo rabastumise võrdlemisi sünkroonne algus võimalikule seosele kliimaga.