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DOES SOIL CHANGE CAUSE VEGETATION CHANGE OR VICE VERSA? A TEMPORAL PERSPECTIVE FROM HUNGARY

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Abstract. The long-term relationship between major climatic change, vegetation change, and soil development is complex and poorly understood. In northeastern Hungary, for example, geochemical and pollen studies from a lake sedimentary sequence indicate that in the early postglacial, vegetation changed from a coniferous to deciduous forest, and soils from a podzol to brown earth. But which changed first? Did climatic warming result in a transformation from one soil type to another, which in turn resulted in a change in forest composition, or did the vegetation change first and subsequently alter the soil? How long did these soil transformation processes take? And what mechanisms were involved in the development of a brown-earth soil from a podzol? This paper presents the results of a study addressing some of these questions using palaeoecological analyses of a sedimentary sequence from lake Kis-Mohos Tó in northeastern Hungary. A proposed model for the process by which a podzol becomes transformed into a brown earth is presented, and possible triggering mechanisms are discussed. Results suggest that in northeastern Hungary the postglacial increase in deciduous populations was not consequent on soil type; rather, deciduous trees increased on podzolic soils, and this increase was one of the triggering mechanisms responsible for the development of brown-earth soils.

Key words: brown earth; climatic change; geochemistry; permafrost; podzol; pollen; soils; wildfires.

Introduction

Modelling of the long-term relationship between climate and vegetation change has received much attention over the past few decades (cf. Wright et al. 1993). In comparison there have been relatively few models developed to describe the long-term relationship between climate change, vegetation change, and soil development. Those that have been developed (Iversen 1958, van der Hammen et al. 1971, Birks 1986, Pennington 1986), suggest that soil development was probably a limiting factor to the postglacial increase in temperate trees in northern Europe. Estimates of a time delay up to 1500 yr before temperate trees could colonize have been attributed to soil maturation processes (Pennington 1986). However, these models mainly concentrate on northern Europe where ice covered much of the landscape during the full glacial and the development of soils would have involved the colonization and primary succession of bare rock (cf. Matthews and Whittaker 1987, Crouch 1993). But what about regions that were not covered with ice during the full glacial? Much of central and southern Europe remained unglaciated and although the predominate vegetation was steppe-like communities including grasses, *Artemisia,* and Chenopodiaceae, there were also a number of coniferous and deciduous trees in refugia (Bennett et al. 1991, Willis 1992, 1994, 1996). Although soils in these regions would have been cold (in some cases semipermanently frozen) and nutrient poor, it is likely that podzols covered many areas of central and southern Europe during the full glacial as in the modern boreal forest. The impact of postglacial climatic warming on soil development of these regions is therefore significantly different. For example, during the late glacial, vegetation in Hungary was dominated by coniferous forest (Willis et al. 1995). Early in the postglacial a rapid change to deciduous woodland occurred. Climatic change was the ultimate driving force behind this change, but how was this related to vegetation change and changes in the soils? Present-day coniferous forests in central Europe produce an acidic litter and are usually situated upon acidic podzols typically in cool humid regions; mixed deciduous woodland produce a calcium-rich litter and are situated upon brown-earth soils (Russell 1961, Wild 1993). Did the change in climate therefore result in a change from one soil type to another, which in turn resulted in a change in forest composition or did the vegetation change first and subsequently alter the soil? How long did these soil formation processes take? And what mechanisms were involved in the development of a brown-earth soil from a podzol?

This study attempts to address some of these questions through the palaeoecological analyses of a sedimentary sequence from Kis-Mohos Tó in northeastern Hungary. Palaeoecological techniques of pollen analysis and geochemistry on an 8.79-m sedimentary se-

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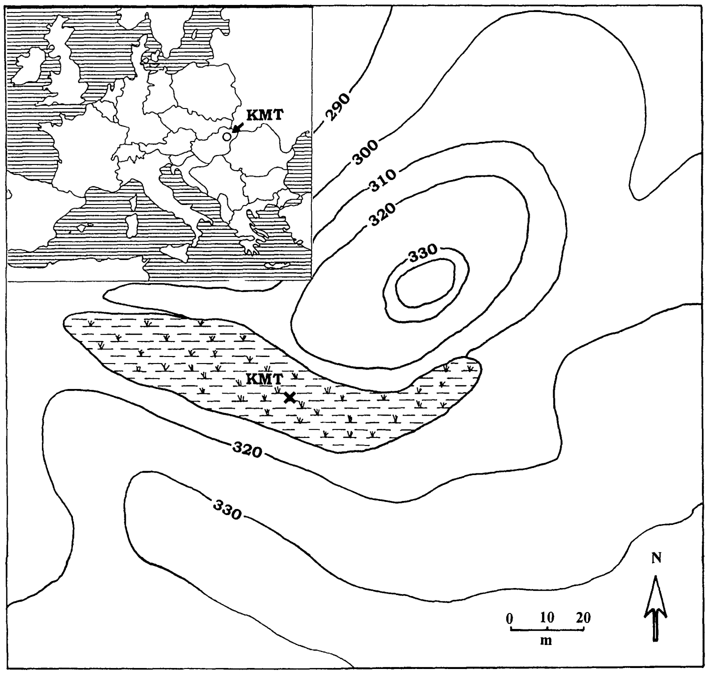

FIG. 1. Topographic map of the Kelemer region, northeastern Hungary (10-m contours) showing the location of Kis-Mohos Tó (KMT) and the core locality (x). Inset map indicates location of KMT in Hungary.

quence from Kis-Mohos Tó provide a detailed record of changes occurring on the slopes around the lake basin since the late glacial. Concentrating only on the sediments covering the time period from ≈14 600–8000 cal. yr BP (prior to anthropogenic activity), results from this study suggest that it is possible to detect the processes and triggering mechanisms involved in the transition from podzol to brown-earth soil with postglacial climatic warming. We present a model for this mechanism and discuss the relationship of changes in the soils to climate and vegetational change.

METHODOLOGY

Study area

The study area is in the Kelemer region of northeastern Hungary. This region, in stark contrast to the Hungarian plain (≈60 km southwest), is composed of undulating hills and valleys formed in alluvial sands and gravels that were deposited during the Tertiary period. Within one of these valleys (at an altitude of 310 m above sea level) there is a small *Sphagnum* peat bog (≈60 m diameter), Kis-Mohos Tó (20°24′30″ E, 48°24′40″ N), from which an 8.86-m sedimentary sequence was collected.

The Kis-Mohos Tó peat bog (Fig. 1) has no inflowing or outflowing streams and is probably hydrologically maintained by the continental climate of the region, which provides relatively high amounts of precipitation (650–700 mm/yr) with a maximum in June and a minimum in February and a mean annual temperature of 9.5°C (January: −3°C; July: 20°C) (Kakas 1960). The microclimate in the valley is cooler (summer: 16°C, winter: −6°C) and moister (P. Sumegi, *personal communication,* 1995). Slopes surrounding the bog are covered with an acidic nonpodzol brown forest soil upon which is situated a mixed deciduous woodland dominated by *Quercus cerris* and *Carpinus betulus.*

Analyses

Two continuous, undisturbed sedimentary sequences (8.86 m) were obtained from the Kis-Mohos Tó basin using a 5 cm diameter modified Livingstone piston corer (Wright 1967) and the following analyses were carried out.

1) The main lithostratigraphic features of the sedimentary sequence were identified through macroscopic investigation and the measurement of percentage of inorganic and organic material at 4-cm intervals throughout the core by loss-on-ignition (Aaby 1986). Analyses of the surface textures of the quartz sand grains in the section of the profile between 730 and 686 cm was carried out using scanning electron microscopy following the technique of Krinsley and Doornkamp (1973).

2) The geochemical elements in the sediment were measured in 4-cm sections throughout the core using a modified technique of the bulk sediment digestion of Bengtsson and Enell (1986) (M. Braun, *personal communication,* 1996). Acid-soluble concentrations of Al, K, Mg, Li, Cr, Sr, Ba, Fe, Mn, P, and Ca were determined using ICPEAS (inductively coupled plasma emission atomic spectrometry). Elemental concentrations of extant leaf litter and soils from the basin were also measured using the same technique whereby fresh

TABLE 1. Radiocarbon age determinations for the Kis-Mohos Tó sedimentary sequence. Calibrated ages were calculated using the calibration program CALIB (Stuiver and Reimer 1993).

Code number	Depth (cm)	Mean (cm)	Age (^{14}C yr BP)	Age (^{14}C cal. yr BP)
AA-11984	316	361	1345 + 50	1281
deb-3338	372–384	378	2148 ± 57	2130
deb-3298	388–400	394	2994 ± 55	3190
deb-3329	410–417	409	2945 ± 53	3080
deb-3339	437–457	447	3574 ± 45	3849
deb-3347	470–486	478	3598 ± 49	3883
AA-12993	484	484	4270 ± 75	4840
deb-3301	520–530	525	4751 ± 62	5540
deb-3300	550–570	560	6250 ± 73	7170
deb-3296	600–616	608	7379 ± 82	8130
deb-3324	658–674	666	7685 ± 51	8140
AA-11986	675	675	8020 ± 100	8950
AA-11987	868	868	12 495 ± 95	14 640

litter and brown-earth samples (0.5 g) were ground, dried, and then digested in the same manner as for the sediments.

3) Samples for analysis of the pollen contained within the sediment were collected using a 1-cm^3 volumetric subsampler. The sampling interval was 8 cm throughout the core with a finer interval of 2 cm for the section between 686 and 710 cm. Samples were processed for pollen (Berglund and Ralska-Jasiewiczowa 1986) with exotic pollen added to each sample in order to determine the concentration of pollen (Stockmarr 1971). A minimum count of 300 grains (all pollen plus spores) per sample was made in order to ensure a statistically significant sample size (Maher 1972) and charcoal abundance was determined using Clark's (1982) point count method.

4) Radiocarbon dating of the sequence was obtained by both bulk and AMS (Accelerator Mass Spectrometry) analyses. Nine bulk samples of sediment were analyzed for radiocarbon ages at the Nuclear Research Centre of the Hungarian Academy of Sciences, Debrecen, Hungary and four samples of plant macrofossils were analyzed for AMS dates at the NERC (National Environment Research Council) radiocarbon dating facility at East Kilbride, Glasgow. In order to allow comparison with the COHMAP (Co-operative Holocene Mapping Project) computer-simulated climatic data (Kutzbach and Guetter 1986, Kutzbach et al. 1993; P. Behling, *personal communication,* 1994), the dates were calibrated using the calibration program CALIB (Stuiver and Reimer 1993) and are labelled cal. yr BP.

RESULTS

Spatial representation

In this study there are two spatial perspectives to consider: the spatial representation of the pollen and that of the chemical elements. Previous work has demonstrated that small lakes (<5 ha) contain pollen predominantly from a local source area (Janssen 1966, Jacobson and Bradshaw 1981, Jackson 1990, 1994). The small size (<1 ha) of the Kis-Mohos Tó basin, which was an openwater pond from 14 600 to 2000 cal. yr BP, suggests that the majority of the pollen rain entering this basin will be from a local and extralocal source and that the pollen record will represent changes that were occurring in the vegetation on the slopes surrounding the basin and immediate vicinity (up to ≈1000 m).

Chemical elements in the lake sediments are derived from a number of different sources but represent two main components: elements from the atmosphere and surrounding soils carried into the basin by overland flow, throughflow, and precipitation, and elements formed as a result of chemical processes within the lake sediment. With the exception of the component entering the lake from the atmosphere (which is thought to be low in comparison to the other elements), the spatial representation of the chemical data is therefore local and restricted to the watershed (Mackereth 1966, Bormann and Likens 1979, Engstrom and Hansen 1984, Engstrom and Wright 1984, Engstrom and Swain 1986).

Temporal representation

Results from the 13 radiocarbon dates (four AMS and nine bulk dates) (Table 1, Fig. 2) indicate an almost linear relationship of sediment deposition with time. The basal sediments extend back to 14 640 cal. yr BP and linear interpolation suggests that 1 cm of sediment was deposited in the basin during ≈20 yr. Although it was not possible to obtain dates for the period between ≈14 600 and 9000 cal. yr BP, which includes the late glacial/postglacial transition when a change in sedimentation rate might have occurred, the available dates suggest that the accumulation rate has been uniform (Fig.2).

Vegetational changes from 14 600 to 8000 cal. yr BP

During the late glacial and up until 9500 cal. yr BP a coniferous forest-steppe of *Pinus, Picea, Larix,* and *Betula* surrounded the Kis-Mohos Tó basin (Fig. 3).

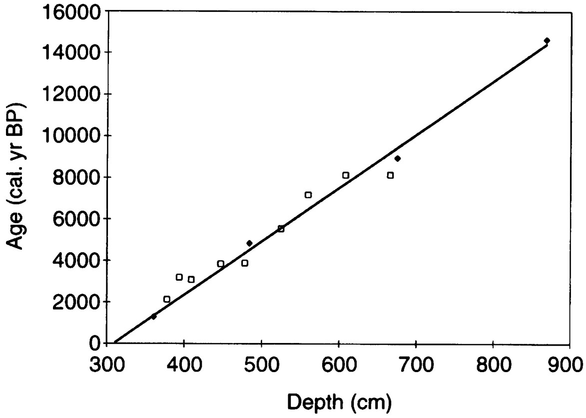

FIG. 2. Calibrated radiocarbon age (cal. yr BP, using program CALIB; Stuiver and Reimer 1993) determinations plotted against depth. The open squares indicate bulk dates; the black squares indicate AMS (Accelerator Mass Spectrometry) dates. The dates are fitted with a linear regression.

Pollen from the trees accounted for >60% of the total pollen with the remaining percentage composed of steppe elements such as grasses, Chenopodiaceae, Cyperaceae, and *Artemisia.* These percentage values are similar to levels recorded in pollen rain from the present-day western European boreal forest-steppe (west of 100° E)(Peterson 1983) and suggest an open coniferous forest with patches of steppe-like communities.

From 11 000 to 10 000 cal. yr BP the dominance of *Larix* and *Betula* in the forest increased with a decline in *Picea* and the steppe elements, suggesting a more closed canopy. *Larix* is notoriously underrepresented with mean values of <1% recorded in forests where *Larix* is dominant now (Peterson 1983). Values of up to 10% *Larix* in the early postglacial woodland around Kis-Mohos Tó therefore suggest that it was a significant component of the early postglacial forest.

Between 9600 and 9500 cal. yr BP, a dramatic change occurred in the forest. There was rapid decline (≈80 yr) in all dominant components of the forest (*Pinus, Larix, Betula*) and an increase in mixed deciduous woodland. This woodland was composed of *Quercus, Corylus, Tilia, Fraxinus ornus, Fraxinus excelsior, Ulmus, Alnus,* and *Carpinus orientalis* and accounted for >90% of the total pollen. At this transition an increase in the charcoal concentration suggests that burning was occurring (Fig. 3). An increase in the Filicales type is probably associated with this burning.

The transition between two very different woodland types took ≈100 yr to complete. Mixed deciduous woodland then persisted around Kis-Mohos Tó until anthropogenic activity affected the woodland at ≈8000 cal. yr BP.

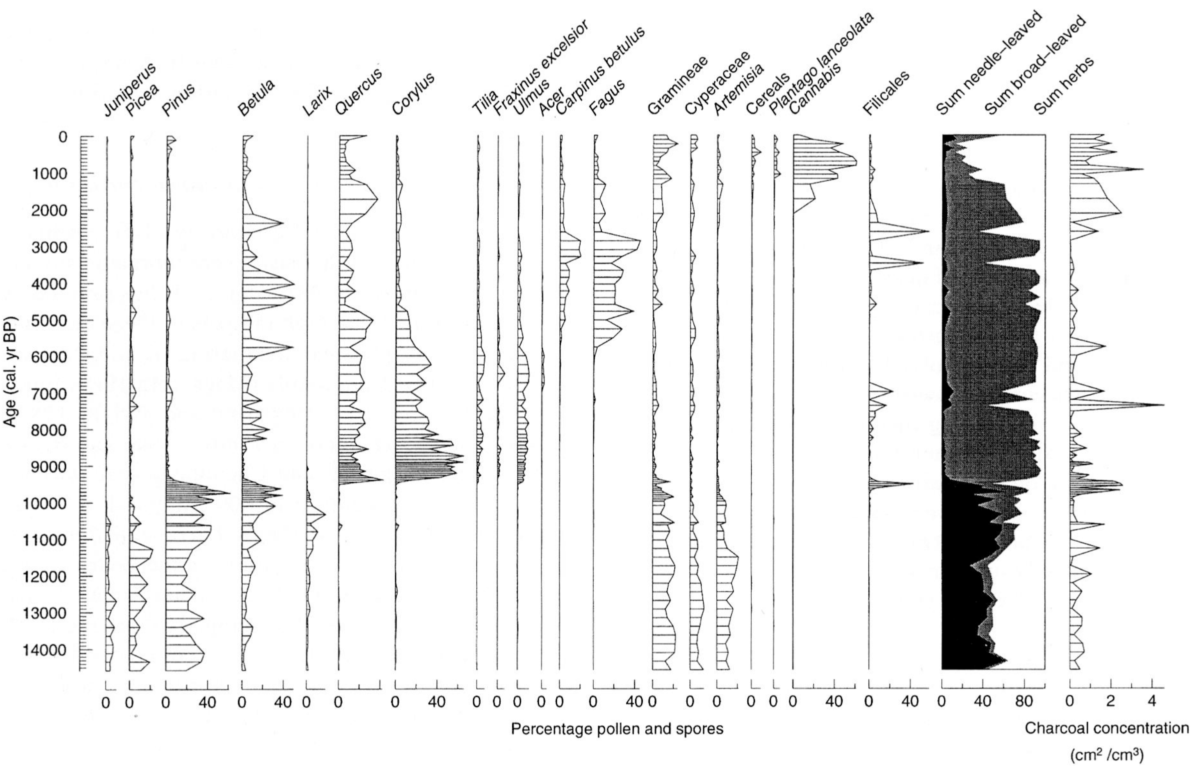

FIG. 3. Percentage pollen and spore diagram of selected taxa from Kis-Mohos Tó plotted against calibrated radiocarbon age. Calculation of the pollen data used a pollen sum of total land pollen excluding indeterminates. Also presented are curves of charcoal concentration (cm^2/cm^3) and a summary diagram of percentage sum needle-leaved, broad-leaved, and herbs.

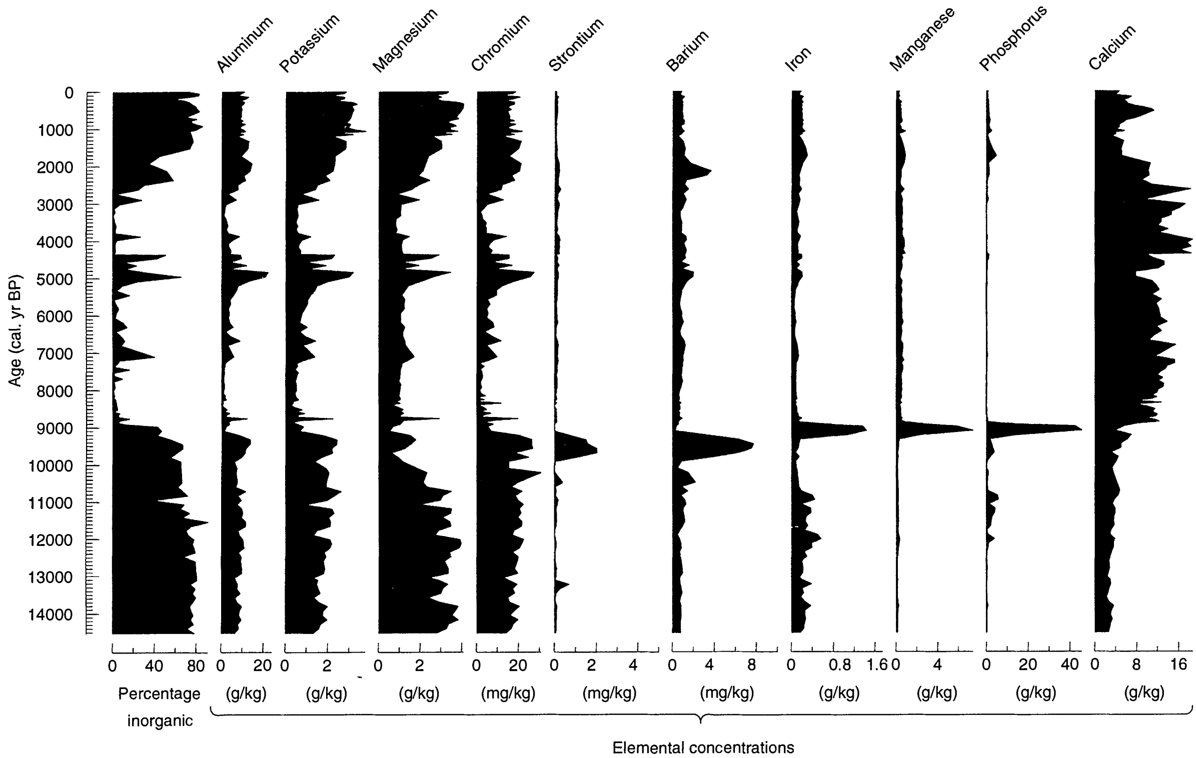

FIG. 4. Physical characteristics (%) and elemental concentrations (g/kg, mg/kg) from the Kis-Mohos Tó sediment plotted against calibrated radiocarbon age (cal. yr BP).

Geochemical changes from 14 600 to 8000 cal. yr BP

The geochemical evidence from Kis-Mohos Tó suggests that four distinctive stages can be detected in the sediment between ≈14 600 and 8000 cal. yr BP (Fig. 4). The first stage from ≈14 600 to 9800 cal. yr BP resulted in the influx of inorganic silicate minerals including K, Al, Mg, Li, and Cr. SEM (scanning electron microscopy) analyses of quartz grains from this section of the core clearly indicate features of frost shattering (Gausses and Coudé-Latridou 1987).

The second stage involved a large increase in Sr and Ba, which occurred between ≈9800 and 9200 cal. yr BP and a corresponding increase in charcoal concentration (Fig. 5). No further indication of frost shattering was found on quartz grains in this section of the core. Following this was a third stage when an influx of Fe, Mn, and P occurred between ≈9200 and 9000 cal. yr BP, and fourthly an increase in Ca and percentage organic from ≈8800 cal. yr BP.

DISCUSSION

Results from this study suggest that between 14 600 and 8000 cal. yr BP a series of distinct processes occurred in and around the Kis-Mohos Tó basin, which caused changes both in the vegetation and elemental composition of the lake sediment (Fig. 5). We infer that the changes in the elemental composition reflect the processes involved in changing from a podzol to acidic brown-earth soil. The processes involved and their relationship to changes in the vegetation and climate are discussed below.

Process of change: podzol to brown-earth soil

During the late glacial the majority of geochemical elements deposited in the basin were inorganic silicate minerals such as Al, K, Mg, and Cr. It has previously been demonstrated that high influxes of these elements are indicative of physical weathering of silicate minerals from nutrient-poor soils (Mackereth 1966, Engstrom and Hansen 1984, Engstrom and Wright 1984). A coniferous forest-steppe of *Pinus, Picea,* and *Betula* with grasses, *Artemisia,* and Chenopodiaceae was situated upon this soil until ≈11 000 cal. yr BP followed by a coniferous forest of *Pinus, Larix,* and *Betula.* A combination of the acidic nature of the bedrock, the coniferous trees, and the cool late glacial climate would have ensured that a podzol developed in this region. It is also probable that a litter layer accumulated from at least ≈11 000 cal. yr BP when the forest became denser and contained a predominance of deciduous trees (*Larix* and *Betula*). Litter breakdown would probably have been slow due to the low biological activity associated with cooler temperatures (Killham 1994). We infer that the first process involved with the change from podzol

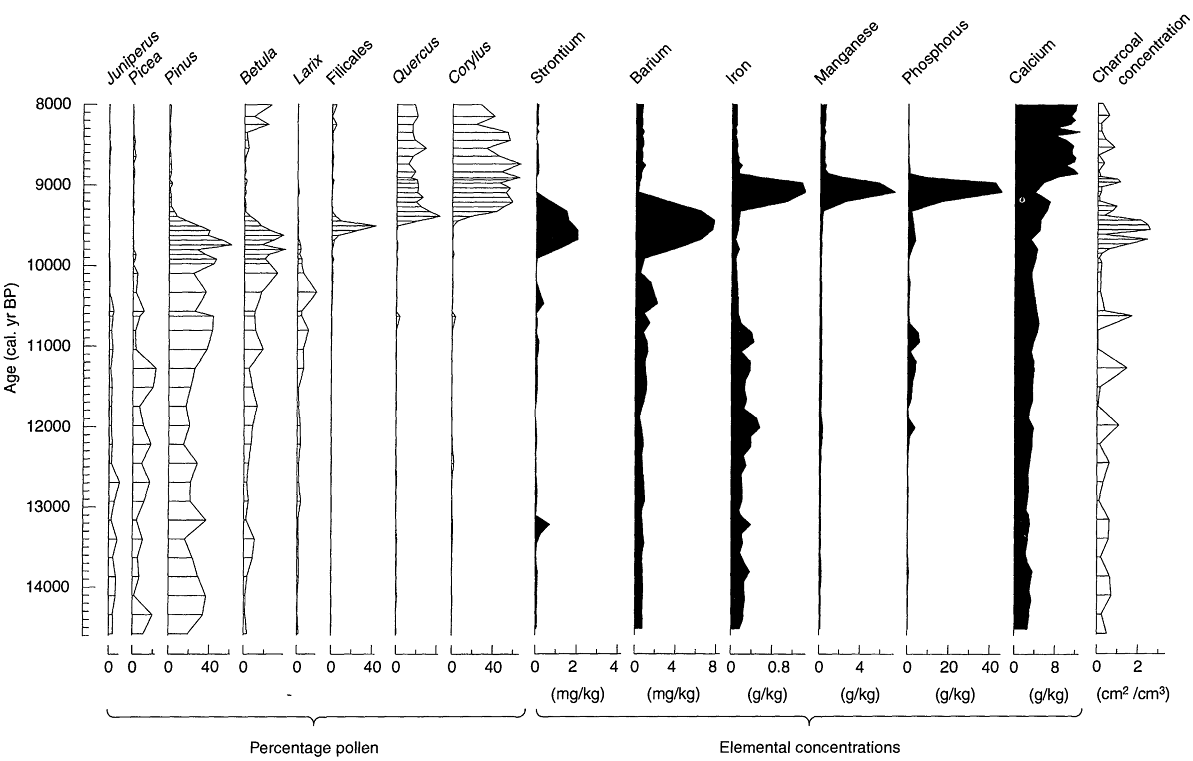

FIG. 5. Pollen (%), charcoal (cm^2/cm^3) and geochemical (g/kg, mg/kg) results of selected taxa/elements for the period 14 600–8000 cal. yr BP.

to brown-earth soil was the breakdown of this litter layer.

1. *Breakdown of litter layer.*—Between 9800 and 9200 cal. yr BP large amounts of Sr and Ba were transported into the basin (Fig.5). Sr and Ba are rarely measured in geochemical records from lakes in temperate regions, although some study has been made of Sr concentrations of lakes in arid environments (Kinsman 1969, Chivas et al. 1986, Gasse and Fontes 1989). Measurements from the Kelemer region suggest both elements occur in the bedrock upon which the basin is situated (P. Sumegi, *personal communication,* 1995). However, the curves of Sr and Ba increase independently of the other inorganic elements (Fig. 4), suggesting that the processes bringing Sr and Ba into the basin are different. Studies of Sr and Ba in vegetation has suggested that that these two elements are preferentially taken up from the soil instead of Ca by certain species (Bowen and Dymond 1955) where they are concentrated in leaves, branches, and shoots. Studies also suggest that this uptake varies greatly between vegetation type (Bowen and Dymond 1955) and that the more acidic the soil solution, the greater the amount of Sr and Ba available to the plant.

TABLE 2. Results from the measurements of Sr and Ba concentrations (mg/kg) of various leaf litter types (freshly ground samples) and brown-earth soil (sample from A horizon) from the slopes around Kis-Mohos Tó basin.

Litter type	Ba (mg/kg)	Sr (mg/kg)	Ca (mg/kg)
Picea abies	300.33	49.67	14 268.3
Pinus nigra	9.06	9.22	6791.6
Carpinus/Quercus	80.12	49.82	15 969.8
Betula	83.25	18.02	10 126.2
Brown-earth soil	97.8	13.84	1361.3

Preliminary analyses carried out as part of this study (Table 2) indicate that all arboreal litter found today around the Kis-Mohos Tó basin contains Sr and Ba and that the highest concentration was in *Picea abies* needles. During the late glacial the slow rate of decomposition would have resulted in the formation of a thick litter layer highly concentrated in chemical elements including Sr and Ba (Dryness et al. 1986). As the litter layer became broken down, Sr and Ba were released into the lake as throughflow and overland flow. This was the first stage in the transition from a podzol to a brown-earth soil.

2. *Release of Fe and Mn from the podzol.*—Immediately following the peak in Sr and Ba there was an increase in Fe and Mn between ≈9200 and 9000 cal. yr BP (Fig. 5). Recent studies on podzols in the northeastern United States have indicated that Fe has maximum net accumulations in the lower B horizons and Mn occurs at greater depths in C horizons (Jersak et al. 1995). Many studies have been made on the mech-

anisms resulting in the release of these elements into lake sediments (Mackereth 1966, Engstrom and Hansen 1984, Engstrom and Wright 1984). Three mechanisms for the release of these elements into lake sediments that have been suggested by other studies include: (a) Physical weathering and intense geologic erosion. (b) Chemical weathering as a result of changing redox conditions in the soils. In this process it is suggested that Fe and Mn from weathered rocks in the catchment are retained as hydrated oxides in the soils but become mobile under waterlogging conditions (Mackereth 1966, Engstrom and Wright 1984, Killham 1994). (c) Chemical weathering as a result of changing soil acidity. It has been demonstrated that organic soils that are rich in humic acid such as beneath coniferous vegetation increase the mobility of Fe and Mn by forming water-soluble complexes (Muir et al. 1964, Lundstrom 1993). In the last two mechanisms Fe and Mn would be transported into the lake by throughflow.

The Fe and Mn curves do not show a close similarity to percentage inorganic curve or elements (Al, K, Mg, Li, Cr) taken to be indicative of physical weathering. It is therefore probable that the peaks in Fe and Mn between ≈9200 and 9000 cal. yr BP are either as a result of reducing conditions caused by waterlogging of the soils or an increase in organic acids in soils due to the breakdown of the coniferous litter layer. Removal of the Fe and Mn layers from the podzol would have promoted the movement of moisture and organisms throughout the soil profile (Killman 1994).

The similarity of the P curve to the Fe and Mn curves (Fig. 4) suggests that there is a close relationship between these three elements. The relationship of P to other elements in the soil is highly complex. However, the close association of P with the Fe curve suggests that it is occurring as a sorbed component of amorphous iron oxide (Russell 1961, Wild 1993). Under acidic or neutral conditions, for example, phosphate ions become strongly held by hydrated oxides of Fe as surface films (Wild 1993). In reducing conditions, iron-chelating agents will remove the film of phosphate and this will be transported into the basin with the Fe and Mn.

3. *Buildup of Ca.*—We suggest that the third and final stage of change from a podzol to a brown earth involved the addition of Ca from ≈8800 cal. yr BP (Fig.5). Brown-earth soils often contain a higher content of exchangable Ca than podzols, although the soils still remain typically acidic (Russell 1961). As mentioned previously, there is low Ca in the bedrock of this region. However, deciduous trees are particularly efficient at the uptake of Ca, which becomes concentrated in the plant and enters the nutrient cycle via litter and organic decomposition (Packham et al. 1992). Ca is also very easily leached from soils (Packham et al. 1992). Thus the increase from ≈8800 cal. yr BP is thought to represent the gradual buildup of Ca through decidous litter decomposition and nutrient cycling resulting in the development of a brown-earth soil.

In summary, we suggest that three distinct processes occurred in the transition from a podzol to a brown-earth soil: the breakdown of the litter layer; the removal of the iron and manganese layers in the soil; and a buildup of calcium. The triggering mechanisms for these processes and their relationship to changes in the vegetation will be discussed below.

Triggering mechanisms

1. *Climatic change.*—During the late glacial much of central and southeastern Europe remained unglaciated (Denton and Hughes 1971) but cold (Hertelendi et al. 1992, Willis 1994, Willis et al. 1995) with January temperatures as low as −20°C (Kordos 1977, 1987, Kutzbach and Guetter 1986, Kutzbach et al. 1993). Climatic modelling for the grid squares covering central Europe and palaeoclimatic reconstructions (Kordos 1977, Kutzbach and Guetter 1986, Kutzbach et al. 1993, Willis et al. 1995; P. Behling, *personal communication,* 1994) suggest that postglacial climatic warming started as early as 12 000 cal. yr BP and that by 9000 cal. yr BP, July temperatures were as much as 5°C warmer than present day. Throughout much of southern Europe the lateglacial/postglacial transition to temperate vegetation started at ≈11 200 cal. yr BP (Willis 1994). But at Kis-Mohos Tó and other sites in eastern Europe (Huttunen et al. 1992, Willis 1994) the transition to deciduous vegetation did not occur until up to 2000 yr later. The results from Kis-Mohos Tó also suggest that the soils did not change until 10 000 cal. yr BP. It is therefore suggested that a triggering mechanism other than postglacial increase in temperature was responsible for the transitions seen from ≈10 000 cal. yr BP.

There is much evidence from both palaeoenvironmental reconstructions and computer-simulated climatic modelling for central Europe that along with an increase in temperature, the climate became drier in the early postglacial in central Europe (Kordos 1977, 1987, Kutzbach and Guetter 1986, Kutzbach et al. 1993, Willis et al. 1995; P. Behling, *personal communication,* 1994). This decrease in precipitation is possibly reflected in the pollen record by the decrease in *Picea* and *Larix* from ≈10 000 cal. yr BP (Fig.3). For example, in the present-day boreal forest of the western Soviet Union, *Picea abies* and *Larix sibirica* have better root growth on cooler and wetter soils than *Pinus sylvestris* (Korotaev 1987). However, it can be suggested that it was the indirect effect of a decrease in precipitation that was the most important triggering mechanism in that it appears to have increased the frequency of wildfires.

2. *Increase in wildfires.*—Evidence from the charcoal record suggests that between 9800 and 9200 cal. yr BP there was an increase in both the intensity and regularity of fires. The Filicales record (Fig. 5) coincides with the high charcoal concentrations and possibly represents the expansion of ferns with the increase

of light from the opened canopy. Furthermore, the charcoal peak between 9800 and 9200 cal. yr BP corresponds closely to the peak in Sr and Ba and suggests that the breakdown of the litter layer was associated with an increase in wildfires. Evidence from this study also suggests that following the litter layer breakdown there was a release of Fe and Mn layers from the podzol and a rapid transition from coniferous to deciduous forest (Fig. 5).

Two hypotheses can be presented as to why the Fe and Mn layers were released once the litter layer was broken down. The first is that the podzol was underlain with permafrost. Geomorphological evidence, for example, suggests that extensive permafrost extended across central Europe (Bell and Walker 1992) and that northern Hungary was in a zone of discontinous permafrost (Maruszczak 1987, Pécsi 1987). Periglacial features have been identified in loess sequences south of the Kelemer region dating back to 11 500 cal. yr BP (Pécsi 1987, Szekel 1987) and SEM analyses of quartz grains within the Kis-Mohos Tó sediment indicate that frost shattering associated with cold conditions and possibly permafrost persisted in this region much later until at least 10 000 cal. yr BP. Studies of the present-day distribution of permafrost emphasize that it is not air temperature that determines permafrost depth, but rather the energy exchange (vegetation composition, precipitation, wind, snow cover, litter layer, etc.) acting in the vicinity of the ground surface (Williams and Smith 1989). As the forest cover increased from ≈11 000 cal. yr BP, therefore, it is possible that a thick developing litter layer prevented any further thaw of the permafrost by reducing the energy exchange. Studies of the present-day boreal forest indicate that the most important effect of burning is the reduction of the litter layer, thus reducing its insulative effect (Brown 1983, Dryness et al. 1986, Bonan 1992). It is only a regular removal of the litter layer that will result in a significant thaw of the permafrost (Dryness and Norum 1983). Thus with increased regularity of fire the litter layer would be gradually reduced and permafrost thaw increased. Waterlogging is a common process associated with permafrost melting. As the ice melts, the frozen layer beneath impedes drainage and causes waterlogging (Viereck et al. 1983). This would create redox conditions in the lowest levels of the soils and result in the mobilization of Fe and Mn.

The second hypothesis does not involve permafrost but an increase in organic acids due to the breakdown of the coniferous litter layer. This increase would have resulted in the formation of organically bound complexes and altered Fe and Mn into more mobile forms (Muir et al. 1964, Lundstrom 1993). Also, higher soil temperatures following fires are thought to increase organic matter decomposition and nutrient mineralization (MacLean et al. 1983, Dryness et al. 1986). Thus the wildfires brought about an increase in the nutrient status of the soils and removed the Fe and Mn layers.

Although both hypotheses are plausible, we suggest that waterlogging resulting from permafrost melt is more likely since the peak in Fe and Mn occurs after the litter layer breakdown and not during it, suggesting that a critical threshold was passed. We propose that if the release of Fe and Mn is related solely to the formation of organically bound complexes from increased acidity, then the release in Fe and Mn would have coincided with the Sr and Ba increase.

3. *Increase in soil temperature.*—We infer that the third mechanism closely associated with the litter layer breakdown was an increase in soil temperature. Although there can never be direct evidence for an increase in soil temperatures, there is evidence from the sediments that the soil temperatures in the Kis-Mohos Tó basin were still low enough at 10 000 cal. yr BP for frost shattering to occur and that it ceased after ≈9600 cal. yr BP. There is also the question of the change from coniferous to deciduous woodland. This occurred at ≈9800 cal. yr BP immediately after burning and the breakdown of the litter layer. The inference is therefore made that the transition from coniferous to deciduous is closely associated with burning and the breakdown of the litter layer. Present-day studies in boreal forest ecosystems suggest that the effect of burning upon the coniferous vegetation would not in itself have been detrimental (Payette 1992). In addition, the soils, if anything, would have been becoming more acidic, again to the advantage of coniferous vegetation. Therefore another explanation must be sought.

We suggest that a reduction in the litter layer resulted in soil temperature increase because of greater exposure to solar energy. This gave deciduous woodland the competitive edge and a rapid increase occurred as soon as the soils were warm enough.

4. Change in vegetation (coniferous to deciduous).—The final stage in the transition from podzol to brown earth involved the addition of Ca to the soils from ≈8800 cal. yr BP. We propose that this mechanism was driven by the change in vegetation. As mentioned above, as soon as the coniferous litter layer was broken down, deciduous trees became established on the slopes around Kis-Mohos Tó. Deciduous trees are particularly good at the uptake of Ca from the bedrock and this enters the nutrient cycle through litter decomposition and leaching. Addition of Ca following the release of the Fe and Mn layers resulted in the development of an acidic brown-earth soil.

Relative rates of change

The radiocarbon dating of the early postglacial must be considered tentative because there are no radiocarbon dates between 14 640 and 8950 cal. yr BP. However, whatever uncertainity there may be about the exact timing of the transitions, the rate of soil change relative to vegetation change is unambiguous. Soil development from podzol to brown earth is both a con-

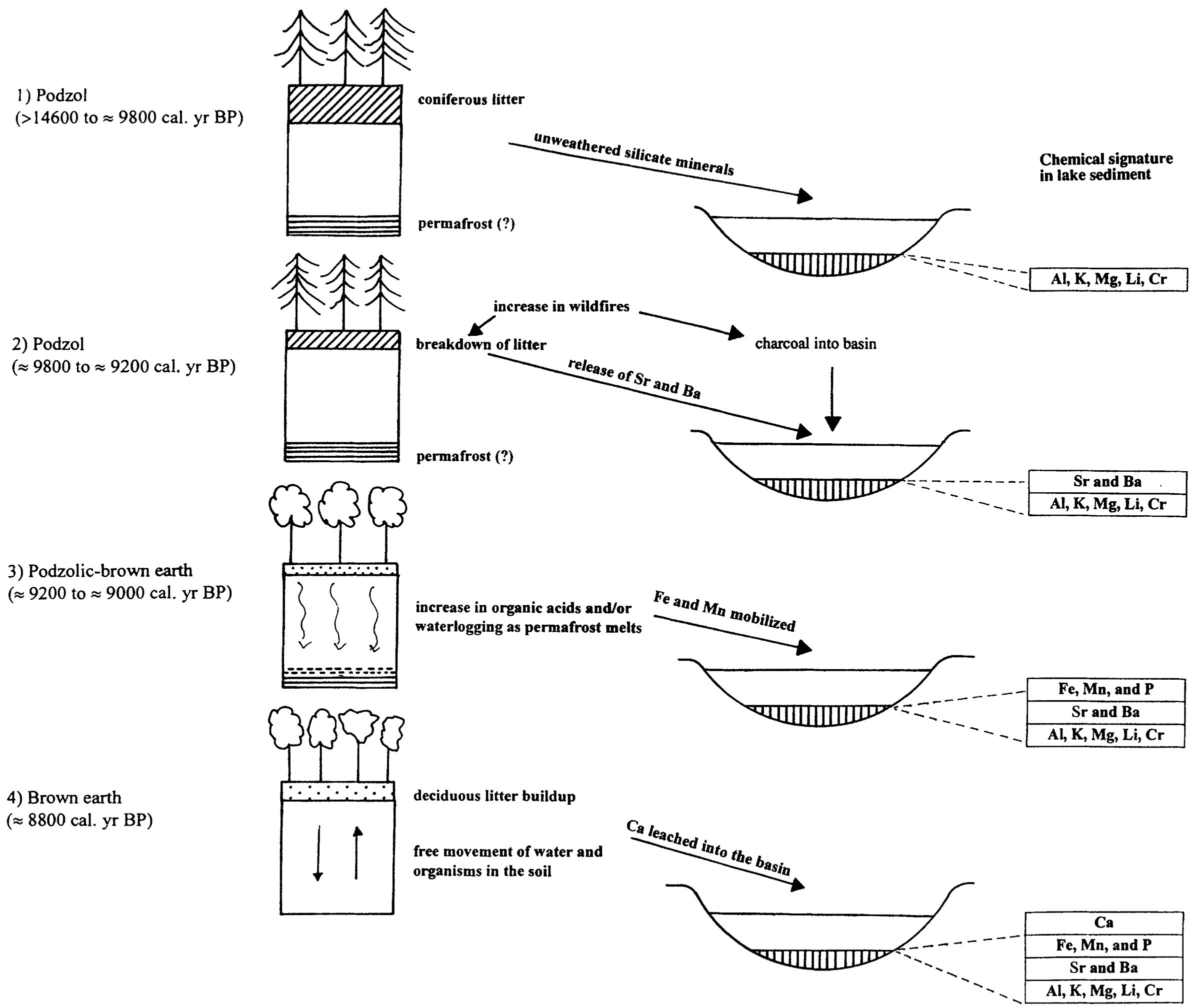

FIG. 6. Schematic diagram to illustrate the main processes (1–4) occurring in the transition from a podzol to a brown earth at Kis-Mohos Tó.

sequence of and slower than vegetation change from coniferous to deciduous woodland.

The three stages involved in the transition from podzol to brown-earth soil thus took ≈1000 yr to complete (Figs. 2 and 5). The first stage, involving the breakdown of the litter layer occurred between 9800 and 9200 cal. yr BP; the second stage involving the mobilization of Fe and Mn took ≈200 yr to complete and occurred between 9200 and 9000 cal. yr BP; and the third stage involving the addition of Ca to the soil profile occurred from ≈8800 cal. yr BP. The rate of the transition from coniferous to deciduous vegetation was more rapid. A change from forest dominated by *Pinus, Larix,* and *Betula* to a mixed-deciduous forest of *Quercus, Corylus, Tilia, Fraxinus excelsior, Ulmus, Carpinus orientalis,* and *Alnus* occurred in ≈100 yr between 9500 and 9400 cal. yr BP.

Conclusions

We conclude that it is possible to infer from this study the process and mechanism by which a podzol changes to a brown-earth soil following postglacial climatic warming. A schematic diagram of the proposed model is presented in Fig. 6. The process involved: (1) the breakdown of the coniferous litter layer (releasing Sr and Ba into the lake) gradually reducing the insulative effect of the litter layer; (2) the release of Fe and Mn layers from within the podzol either as a result of redox conditions from waterlogging (caused by permafrost melt), or an increase in organic acids (from the litter layer breakdown), thus transforming Fe and Mn into mobile forms through organically bound complexes; and (3) the increase of Ca from deciduous litter buildup and nutrient cycling within the soil. Triggering mechanisms included climatic change, an increase in frequency of wildfires, and a vegetation change from coniferous to deciduous woodland, inititating the buildup of Ca.

Models of early postglacial woodland succession (Iversen 1958, van der Hammen et al. 1971, Birks 1986, Pennington 1986) suggest that deciduous tree populations only increased after brown-earth soil development had taken place. However, there is strong evidence in this study for the inference that the increase in deciduous tree populations was not consequent on soil type (podzolic or brown earth); rather, deciduous tree populations increased on podzolic soils and this increase was one of the triggering mechanisms responsible for the development of brown-earth soils.

Results further suggest that with the increase in wildfires and a reduction of the litter layer, a critical threshold was passed where deciduous forest rapidly increased (within 100 yr). We infer that this threshold was soil temperature, which increased as a result of litter layer reduction and increased solar input. The timing of the deciduous forest increase and the decline of coniferous forest were so rapid that it almost suggests that woodland replacement following one clearance by fire caused this dramatic transition.

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