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Consequenses of Global Warming on Soil Processes in Arctic Regions[1]

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Summary: This paper reviews anticipated consequences of global change processes on soils in arctic regions. Permafrost is regarded as the main factor which controls many environmental processes, hence, not only changes in temperature, also changes in water contents, nutrients and related ecological factors are considered for effects on soil processes. It is evident for the description of soil processes that the individual variables act at very different time scales with long time lags and hystereses. The transitional stages from cold deserts to tundra and forest can be regarded as most important, and regions at margins of such systems can be identified as most effective for monitoring soil processes. Most actual ecological studies, however, suffer from their short term runs, and focus must be put into a combination between both, actual data sets and data archives. Such archives can be found in the time span of the Holocene, where drastical climatic changes occurred. The use of both, historical data and data from actual process studies can give better hints for forecasts and the interpretation of fossilised pictures. The search for fingerprints in soils is difficult, but is a valuable tool in order to establish direct relationships between climate change and soil processes.

Zusammenfassung: Der Zusammenhang zwischen Klimaänderungen und deren Folgen für arktische Böden wird anhand einer Literaturübersicht versucht darzustellen. Das Auftreten des Permafrostes kann als wesentlicher Faktor erkannt werden, der verschiedene umweltbezogene Prozesse steuert. Dazu zählen Änderungen der Temperatur, des Wasserstatus, der Nährstoffverhältnisse. Wichtig ist die Berücksichtigung der sehr unterschiedlichen Zeitskalen der einzelnen Prozesse bzw. deren Abbildungen. Randbereiche von kalten Wüsten, Tundren und Wäldern können als Orte stärkerer Veränderung angenommen werden und würden sich daher als Beobachtungsorte besonders gut eignen. Da die meisten ökologischen Untersuchungen aber in der Regel nur kurze Zeitintervalle (3–5 Jahre) überdecken, sind Datenarchive der Paläoökologie von besonderer Bedeutung. In der Kombination beider, Daten aus aktuellen ökologischen Programmen und Archiven aus erdgeschichtlichen Zeiträumen, z.B. dem Holozän, können dann Möglichkeiten für die Interpretation der Archive wie auch für verbesserte Voraussagen gewonnen werden. Die Suche nach Spuren von Klimaänderungen in Böden ist jedoch schwierig, dennoch lassen sich Methoden der Paläoökologie als wichtige Werkzeuge erkennen, um direkte Verbindungen zwischen Klimaänderungen und Bodenprozessen aufzuzeigen.

INTRODUCTION

Permafrost stretches beneath 20 % of the land masses (PÉWÉ 1982), and locally reaches down to several hundred meters. Today's permafrost distribution and structure is a product of past and present climate and terrain features. The depth of the active layer is a reflection of the dynamic equilibrium reached between hydrological and thermal properties of the soil and atmospheric conditions (HINZMAN et al. 1991).

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Such vast areas of permafrost, e.g., tundras cover about 1.1 x 107 km^2, are regarded to play a central role in the development of climate changes, especially when the big amounts of carbon stored in these environments are considered. About 25 % of the total world soil C-pool (= 180 x 10^{15} g) and 12 % of the global C-pool can be assumed to be located in the permafrost and the seasonally thawed active layer (OECHEL & VOURLITIS 1994). This carbon has the potential to become a main source of atmospheric CO_2.

Changes in the extent of permafrost regions consequently will also be mirrors of global climate change. Atmospheric and oceanic circulation patterns have consequences on changes in the extent of arid areas in the northern and southern hemisphere which can be related to climate changes in the northern hemisphere (BENSON et al. 1997, STOKES et al. 1997, STOCKER & SCHMITTNER 1997).

The anticipated mean global warming for the next century is about 0.1-0.3 °C per decade, but the effects in the polar regions can be expected to be stronger than in other regions (HOUGHTON et al. 1990). The continuous process of increasing temperature will result in thawing of permafrost and an increase in active layer thickness due to an increased heat flow into soil surface layers. The winter warming in Arctic regions will be 2 to 2.4 times above the global average, the summer warming only 0.5 to 0.7 times that of the global average (MAXWELL & BARRIE 1989). Possible feedbacks and side effects of such impacts, e.g., changes in albedo, precipitation, cloud formation, have recently been discussed in various meetings, books or journals. However, present-day data and models also indicate a decline of temperature in several areas of the north Atlantic (JONES 1990, WALSH 1991, CATTLE & CROSSLEY 1995).

Changes of earth's climate as anticipated from global warming models, however, are not new events in earth's history. As such, the climate change during the Allerød (12,000-11,000 BP) or the Preborial (10,300-9,200 BP) may be used as analogons for today's situation. From this and more recent climate changes, some extrapolations on environmental effects can be made. Changes have been described for distributions of plant and animal species by BIRKS (1990) and BOLIN et al. (1986) or for geomorphological processes by DERBYSHIRE (1976). Changes in landscapes and related environmental properties have to be considered in addition to regional and temporal variations (HULME et al. 1990, ROGERS & MOSLEY-THOMPSON 1995, MAXWELL

1997). The thawing process does not only effect the conditions of the biosphere but moreover the pedosphere and physical processes of soil behaviour. This will definitely influence the distribution of nutrients and further soil ecology which will have effects on remnants and fossilized structures.

Arctic soils can be regarded as archives for informations of paleoclimatic and paleoenvironmental records. Such proxies are especially well preserved in frozen soils, thus permafrost soils are well suited to carry informations about the last glaciation-deglaciation periods. Biological and non-biological proxy data of different kind can be found and analysed for studies and reconstructions on former climatological and environmental changes (HUIJZER & ISARIN 1997). Humic substances, soil layers, soil types and other features can be regarded as proxies. During this paper I will elaborate on tasks which seem relevant for the description of consequences of global warming for soil processes:

- monitoring of thawing of permafrost landscapes and changes in soil developments,
- ecological aspects of permafrost change and consequences for nutrient cycles from changing soil properties,
- reconstruction of Holocene events.

THAWING OF PERMAFROST LANDSCAPES AND CHANGES IN SOIL DEVELOPMENT

Distribution of permafrost and thermal behaviour

The distributions and dynamics of permafrost influence significantly soil physical, soil chemical and soil biological processes. Hence, the response of soils under permafrost to climate change is of special interest. Most sensitive areas are those at boundary regions where even small climatic variations may lead to the extent of frozen soils. Soil climatic conditions become instabile and force the retreat or advance of marginal permafrost areas. The energy exchange regime at the ground surface is a crucial factor for the extent and thickness of permafrost, the development of patterned ground and other periglacial features.

Thermal profiles of the permafrost across the north slope of Alaska indicate a temperature rise of 2-4 °C within the last century (OECHEL & VOURLITIS 1994). OSTERKAMP & ROMANOWSKY (1996) described changing permafrost conditions on a measured 11 years record. Change in the thermal regimes of the active layer were analysed for a six–years run (1987-1992) in Alaska and revealed a good relationship between active layer thickness and summer air temperatures (ZHANG et al. 1997). However, large scale data are rare and local data have to consider short term effects and microclimatic conditions (LABERGE & PAYETTE 1995). Thus, models of permafrost variability cope with broad limits of errors. Realistic forecasts are restricted to time spans below the scales of long-term Global Change scenarios. Although permafrost can be regarded as a heat sink, it also represents a main buffer for soil heat capacity. With permanently increasing temperatures, this buffer will gradually decrease and the present extent of permafrost inevitably will shift the southern limit northward (MAXWELL 1992).

There are several separate effects which have to be born in mind when discussing soil processes in relation to decreasing permafrost, i.e., changes from dry-cold to wet-warm or dry-warm. An important effect that vegetation has on permafrost is its role in shielding the soil surface from solar heat. Barren soils and rocks have significantly different reflectivity of solar radiation leading to different evaporation rates and thermal conductivity (BROWN 1970). This results in considerable differentiation of vertical profiles of temperature and humidity with respect to soil cover (ANGIEL 1990, BOUDREAU & ROUSE 1995). Widespread moss cover provides an insulation altering thermal conductivity from dry to wet moss cushions from 0.06-0.5 W m^{-1} K^{-1} (FAROUKI 1981). The latter figure is close to the thermal conductivity of water (0.56 W m^{-1} K^{-1}; solid soil: 0.3–0.4 W m^{-1} K^{-1}; frozen soil: ~ 1 W m^{-1} K^{-1}). The removal of plant cover by surface disturbance causes a degradation of the underlaying permafrost and thus lowers the permafrost table (NICHOLAS & HINKEL 1996). Another effective insulator is the snow cover which serves for permafrost stability (MACKAY & MACKAY 1974) and further influences directly the vegetation cover and soil development (BLUME & BÖLTER 1996). Such effects, however, are less pronounced when higher plants or shrubs are already growing.

Soil development under permafrost conditions

Soil development can be regarded as being slow in Arctic regions. This can be verified by reduced reactions to podzolization processes (TEDROW & CANTLON 1958, HINNERI 1974, COWELL et al. 1982). The formation of Ae-Bf or Ae-Bh horizons would require about 2000 years (PROTZ et al. 1984). Thawing, precipitation and elevated summer temperatures will increase the solubility of mineral elements and humic substances and their transport into deeper layers. Close interrelationships can be seen between soils, vegetation patterns and organic matter accumulation (KLIMOWICZ & UZIAK 1988, BLUME & BÖLTER 1996, BLUME et al. 1996, 1997).

Soil morphology shows distinct patterns of pedogenesis in cold environments. Qualitative field properties, e.g., depth of oxidation or silt coatings, can be used as good indicators for pedogenesis in arctic environments (FORMAN & MILLER 1984) although uncertainties remain due to the question whether, e.g., clay movement is a contemporary process or a relict from previous milder climates (BOCKHEIM & KOERNER 1997). BURNS (1980) observed maximum silt accumulations with dense herbaceous vegetation covers which had trapped eolian silt and which is subsequently illuviated down profile. A micromorphological study on tundra soils shows their evolutionary stages from cryogenic to biogenic factors (RUSANOWA 1996). These examples show that micromorphologic studies can be used to trace former environmental patterns. It should, however, be born in mind that parent material may be a more important factor in pedogenesis (KELLY & KING 1995). These processes, however, are slow and visible only along wide transects or chronosequence studies. As such, nitrogen and carbon increased significantly during the first few hundred years of pedogenesis on a 1200-yr-old series of moraines in the Yukon Territory (JACOBSON & BIRKS 1980). Later, the total nitrogen content decreased. Contents of carbon, clay and iron, pH, and other soil properties were used by WALKER et al. (1996) to estimate the ages of pingos in northern Alaska.

The most crucial effects in soil processes occur at temperatures around 0 °C, i.e., during freeze-thaw cycles. This temperature range thus becomes important for many chemical and biological processes. HINMAN (1970) observed that freeze-thaw cycles are followed by an increase in exchangable NH_4-N and decreased exchangable K. Soil aggregates of clay or silt, produced by freezing, swell and disperse in relation to electrolyte concentrations. Unfrozen water serves as solvent for inorganic and organic material, which lowers its freezing point. Water moves from warm to cold, i.e., from regions of low solute concentrations to high solute concentrations and from high–moisture zones to low-moisture zones (PERFECT et al. 1991). This creates movement of water along with gradients of nutrients and thermal conductivity (MARION 1995) and effects nutrient availability, cation exchange properties, soil weathering, and biological activity. Flowing water further poses a strong heat flux with pronounced effects on soil temperature and consequently on the depth of the active layer. Enhanced in soil water movement will lead to active and increased groundwater discharge and further depletions of nutrients. Hence, physically driven changes in soil processes will be a significant phenomenon during slow warming of permafrost regions.

Co-effects of relief and drainage, i.e., water saturation, have consequences on below-ground biological and chemical processes. All climate change models show increased wintertime soil moisture in the high northern latitudes but less soil moisture in summer in northern mid latitudes (UNEP/IUC 1997). The effects of relief are most evident in mountainous regions. Small-scale variations may cause similar situations elsewhere in the Arctic. Down-slope mass movements, such as solifluction, frost-heaving and creep can alter the ground surface significantly (RAPP 1970), which leads to an increased soil roughness. Changes in relief are prominent effects of active cryoturbation and thermokarst processes with influences on vegetation. Impermeable layers impede drainage which leads in a decline of aeration and impoverishments of nutrients by increased microbial activity. After warming, increased interactions between the chemical processes and physical constraints of soils (e.g., soil texture, porosity, heat transfer) influence the contaminant stability which has indirect effects on revegetation and carbon cycling. Progress in permafrost erosion can force the degradation of gas hydrates followed by releases of methane (MICHEL & EVERGREEN 1994). CHRISTENSEN (1993) showed an increase of methane release, holding this as a physically mediated and biologically related process in the active layer.

ECOLOGICAL ASPECTS OF PERMAFROST CHANGE

Shifts in ecotones and plant communities

Transitions between ecotones have great importance in monitoring changes of soil properties. LLOYD et al. (1994) examined the transition from tundra to steppe in central Alaska and the related changes between two ecological states. They speculated that this competition was forced by climatic change. Such regions may serve as models for observations of environmental change at appropriate time scales. Biologically, there is an old, winter-adapted flora in the arctic which is exposed to different kinds of stress. Shifts from tundra to boreal forest are as important as shifts from cold deserts to tundra environments. Migration of plants will be less vigorous than under conditions with less or no permafrost at all as frozen soil hampers deep root systems for long times. Hence, vegetation shifts have to be monitored carefully – in relation to soil properties and soil processes in areas of discontinuous permafrost and continuous permafrost. Plant functional types, such as dwarf shrubs, herbs, grasses, cryptogams, thus can be regarded as descriptors for climate effects which are mirrored in their palaeorecord. Their impact on soils is evident from effects on nutrient cycling, disturbance regime, or litter production (CHAPIN et al. 1996).

Changes in plant community composition will not only effect the thermal balance and the nutrient state of a soil, but also its structure. During a study on vegetation patterns in Alaska, GROSS et al. (1990) could show that the proportion of live aerial biomass (graminoids) increased with habitat wetness and depth of frozen soil, whereas shrub biomass decreased under these conditions. In addition, the lack of aerenchyma in dicotyledones limit their penetration of wet habitats. Consequently, this will effect the herbivores and the decomposer food web (FAJER et al. 1989, NORBY et al. 1986).

Effects of root systems

Root systems and their release of low-weight organic acids alter the fine structure of mineral particles and aggregates and produce higher internal surfaces. This further leads to changes in the exchange capacity of the soil as well as to changes in possible pores and niches for microorganisms (GEBAUER et al. 1996). Other effects may be caused by changes in C-allocation of plants or possible increased root growth. Buffers in carbon storage and nutrient reserves (BERENDSE & JONASSON 1992, JONASSON 1996) are important clues in the discussion of changing soil processes.

Such changes will occur at different time scales, depending on the responses of different plant covers. Graminoid or moss dominated communities will act on much shorter time scales than forested landscapes. The use of different soil layers for nutrient leads in different plasticities of communities and responses to environmenta changes (JONASSON et al. 1996, CALLAGHAN & JONASSON 1995). As the gain in photosynthesis is more pronounced in the C_3 pathway compared to the C_4 pathway (BILLINGS 1987), shifts in plant communities from prairies or tundras to shrubs and forests can be anticipated.

Due to the fact that shallow root systems are restricted to the active layer, tundra systems react much more sensitively to climatic changes than forests. Much more solar energy dissipates into evaporation of water from wetlands than from forest covered areas (LAFLEUR & ROUSSE 1995). On a short time scale (decades), thawed barren soils may increase the vegetation cover, which increases again insolation, and as a result, the permafrost table may move upwards again (SVOBODA 1994, NICHOLAS & HINKEL 1996). Trends in elevation of water contents in permafrost soils over a 30-yr period have been monitored by HINKEL et al. (1996). The transition state with strong local variability may last several years, and obviously will produce unpredictable situations for modellers. On longer times scales, however, non-linearities of model components may be smoothed out (OSTENDORF 1996).

As discussed by ALEXANDROVA (1988) plant succession series of semideserts or areas left by retreating glaciers are often shorter in comparison to low latitude zones. The vegetation is entirely determined by characteristics of pioneer species, i.e., their special life strategies (CALLAGHAN & EMANUELSON 1985). A study of recolonization of fresh moraines at Svalbard showed that after ten years a relatively large number of plants (14 species) appeared at many places. Thereafter, for the next 20 years, new colonization stagnated (PIROZNIKOV & GORNIAK 1992). The reason was seen in the soil structure which was dominated by sharp-edged solid aggregates and a continuous cryoturbation and solifluction. This prevented growth of roots and a subsequent accumulation of organics which otherwise would have served as a buffer for water and nutrients. The mechanical role of vegetation is an important clue in soil formation. Root elongation and the production of complex nets limit the superficial displacement of the soil and enhances its morphological expression (VLIET-LANOE 1993). These effects result in stabilization of soils and prevent solifluction and erosion by an increasing shearing resistance (VLIET-LANOE 1991).

Consequences from microbial activity

Freeze-thaw-cycles are resposible for bursts of microbial activity (SCHMIEL & CLEIN 1996, HASHIMOTO & NITTA 1997). Liberation of nutrients from weathering processes or lysis of cells and subsequent use of such „new“ material by bacteria has been discussed by SKOGLAND et al. (1988) and others. The inputs can also be followed by an increase in enzymatic activity (ROSS 1972). Such overshoot reactions, however, may be only short-lived, but may have trigger effects for other (bio-)chemical reactions. A depletion of the system of dissolved organic matter or fine changes in the C/N ratios affecting the remineralisation-immobilisation balance (MARION 1995). Unpredictable, however, remains the state of „readiness“ of organisms under snow covers for metabolic processes after snow melt (GALEN & STANTON 1995). Wetness allows the development of cyanobacteria and algae and local manuring from their slime production. Activity of microorganisms and microfauna allows the mulching and integration of organic matter into the mineral matrix. This increases biochemical weathering and results in stronger frost shattering (VLIET-LANOE 1993), and fine particles, amorphous clays and organic gels raise the frost susceptibility (VLIET-LANOE 1991).

The consequenses of increased carbon dioxide have received much attention with the aim to determine the so-called b-factor (BACASTOV & KEELING 1974), i.e., the extent to which an increase in CO_2 concentration increases the C storage in terrestrial ecosystems. Many plants acclimatise to elevated CO_2 concentrations relatively quickly and thus can serve as a buffer for CO_2. Species with rapid growth rates may be more responsive than slow growing species (HUNT et al. 1991). Data comparing arctic CO_2 fluxes in the 1970s to those measured twenty years later have shown a change from a CO_2-sink to source (OECHEL et al. 1995). This effect has been related to increased surface temperature. Many local aspects of tundra systems, however, have to be considered in order to formulate models or forecasts. JONASSON et al. (in prep.) resume that, in case of rising temperature, middle and high arctic regions will act as C-sinks, due to the expansion of the vegetation. Low arctic regions, however, will become C-sources, due to increased rates of decomposition. Measurements of CO_2-fluxes in a polygon tundra on Taymyr Peninsula revealed that the system acts as a sink for CO_2 during summer (BÖLTER & SOMMERKORN, unpubl.).

Nevertheless, the arctic system is a place of many microenvironmental conditions, and plants or microbes, which don't care much for macroscopic conditions, have to be reconsidered in the context of small-scale local microenvironmental conditions. The coupling or competition between aerobic and anaerobic processes still remains a challenge for tundra ecology. The sensitive transition between methanogens and methylotrophs, as well as the bypass of methane evaluation via plant stems, and the delicate change between these major processes affecting atmospheric impacts are extremely fine tuned and influences global models to a large extent. Such conditions, however, find their limit when the lowering of the water table results in the cessation of methane flux. Methane release is related to the below-ground and above-ground microbial communities and their activities. It can be well mediated by vegetation in wet tundra (GROSS et al. 1990), to soil moisture, temperature and the actual depth of the seasonally thawed active layer (VOURLITIS et al. 1993, CHRISTENSEN et al. 1997). OECHEL & VOURLITIS (1994) estimate that an increase in temperature from 2 to 12 °C would lead to a 6.7 fold higher methane efflux.

Consequences for the soil nutrient status

Along with the anthropogenic alteration of CO_2 there is a significant change in the global nitrogen cycle. Arctic soils can be regarded as deficient in nitrogen (SHAVER et al. 1986) which limits growth of several species. The amount of naturally fixed N_2 in terrestrial ecosystems can be assumed with about 100 x 10^{12} g N per year (SÖDERLUND & ROSSWALL 1982). This amount is outstripped by the industrial nitrogen fixation for fertilizers and during combustion. Much of this nitrogen enters the atmosphere via nitrous oxides or ammonia and is distributed by global atmospheric circulation patterns to remote areas. This creates a „bypass“ of N-support to arctic soil systems to its natural sources by cyanobacteria which are hitherto regarded as the main N-suppliers in arctic environments (HENRY & SVOBODA 1986); they may loose their prominent function in the food web.

Once a sufficient stock of nitrogen has been reached, further requirements are secured by a recycling of the system. The fertilisation of originally N-depleted areas will effect the plant growth and community composition; experimental studies on fertilization have given sufficient proof (e.g., CHRISTENSEN et al. 1997). However, microsite heterogeneity influences the balance between immobilization and remineralization at the process level on the one hand and at the organisms level on the other hand. This complicates assumptions made at system level for the influx and efflux of individual nitrogen compounds. It will effect the nutrient cycle by, e.g., increased decomposition rates and microbial turnover, and thus for all ecological processes at all levels in arctic systems (NEDELHOFFER et al. 1992, 1996).

Such scenery, with different time scales for mineralization, puts the arctic soil system in a wider context when the influence of nitrogen on the sequestration of carbon stocks is considered. Nitrogen mineralization was found not necessarily related to nitrogen flux because of the internal nitrogen cycle and high immobilization effects (MOORHEAD & REYNOLDS 1993). This becomes complicated as only a small fraction of the nitrogen is potentially mineralizable at prevailing cold temperatures (MARION & BLACK 1987). The liberation and redistribution seems to be of greater importance than only temperature shifts of 2-4 °C, as revealed by several experimental studies (e.g., JONASSON et al. 1996). Similarly, the sequestration of carbon contents is effected by temperature shifts. Studies in an Antarctic pergelic cryohemist showed that carbohydrates are less mineralized than under temperate climate conditions. Alkyl carbon units are enriched and incorporated into humic matter (BEYER et al. 1997).

Generally, changes in nutrient concentrations are generally much more effective on soil organisms than changes in temperature. The change in annual mean temperatures does not necessarily effect soil organisms to any great extent. If we consider temperature changes as effective driving forces, we have to look at events such as extreme long summers (or winters) which can effect the survival and breeding success of soil dwelling organisms. Their activity on bioturbation, dislocation of soil particles, and formation of tubes for soil aeration might be of more importance than cryoturbation processes. Today's observations thus should consider carefully climate „catastrophes“ which are smoothed out by long trend analyses. This also tackles the problem of analyses of qualitative data, and it should sharpen the eyes for new invadors, which can be regarded as preliminary indicators at small scales for large scale changes (e.g., SMITH 1994, GROBE et al. 1997).

RECONSTRUCTION OF HOLOCENE EVENTS – LEARNING FROM THE PAST

Holocene events

The reconstruction of former periglacial environments gives indications on the effects of global warming on soil environmental conditions (PÉWÉ 1983). Transitions to modern climate following the last glacial maximum was obviously punctuated by several rapid oscillations. Pollen analyses of Canadian soils reveal a rapid change to poplar and spruce at this time (RITCHIE 1989). This was also found in records from arctic ice cores (ALLEY et al. 1993) as well as in tropical corals (BECK et al. 1997, SMITH et al. 1997). HAHNE & MELLES (1997) further describe the transitions of vegetation pattern during Bolling, Older Dryas and Younger Dryas from sediment cores of lakes from Middle Siberia. Macrofossil and pollen records indicate shifts in treeline in British Columbia during several stages of the holocene (SPOONER et al. 1997). Isotope profiles indicate that in the time span from 18 ka to 8 ka BP mean temperatures in high latitudes could have changed by 5-7 °C very shortly, in the course of only a few decades (VAIKMÄE et al. 1995). Highest increases in temperature can be assumed for the time span between 11 ka to 8 ka. The active layer became thick enough for further plant colonization and caused the retreat of ice-rich permafrost which resulted in the proliferation of thermokarst (RAMPTON 1988).

The rapid climate warming at the end of the Younger Dryas with temperature shifts >5 °C within 50-100 years (JOHNSON et al. 1992, SMITH et al. 1997) caused various vegetation changes in response to climatic changes at that time. Some lags for growth responses become evident for some taxa, as shown by pollen analyses of lake sediments from atlantic Canada (MAYLE & CWYNAR 1995) and for changes in midge communities (WALKER et al. 1991). Shifts in the vegetation cover have also been monitored by pollen diagrams for central Siberia, where significant changes occurred at the boundaries from Younger Dryas to Boreal where shifts from non-tree pollen to tree pollen, mainly *Larix, Betula* and *Alnus,* were monitored (MELLES et al. 1996, HAHNE & MELLES 1997, HAHNE & MELLES in press).

Effects on soils can be anticipated especially from colonizations by *Alnus* sp., e.g., during the change from Preboreal to Boreal, ~9200–9000 yr BP in Canada. Their ability to form rhizomes have the potential of nitrogen fixation and thus change the nutrient stage of the soils. Other shifts in plant communities have been supposed by formations of peatlands which are indicative of increasing humidity. On the other hand, dry regions may have suffered from erosion and slopewash during the degradation of permafrost (WALTERS 1994). Different patterns of N-utilization by various functional groups of plants based on their mycorrhiza have been discussed by JONASSON & MICHELSEN (1996).

Transformations from tundra to forest-tundra took place during a period of approximately 150 years, about 5000-4000 years ago in Central Canada (MACDONALD et al. 1993), a time span which is just under consideration for today's climate models. The authors showed a response of the treeline vegetation to climate change during this period from pollen (mainly *Betula* and *Alnus*) in sediment cores and observations of buried cryic regosols and organic soils. ZOLTAI (1995) showed shifts in permafrost boundaries in northern Canada in relation to peat development. Hence, changes in soil biological processes in relation to growing root systems within the next century can be assumed for comparable regions. Storages of organic matter and changes in nutrient contents are due to new mycorrhizal systems, effects which can be attributed directly to decreases in permafrost.

Modern climate change

Data about plant growth after the little ice age about 400-500 years ago have been presented recently for *Cassiope tetragona* from deglaciated areas on Ellesmere Island (Havström et al. 1995). Based on these data, the number of leaves produced per year can directly be related to mean July temperature. More recent shifts in environmental conditions have been documented by changes in diatom assemblages at the end of the 18th century in an high-arctic ecosytem. These shifts were related to climate variables (Douglas et al. 1994). The anticipated warming of 1-3.5 °C over the next 100 years would shift current climate zones poleward by ca. 150-550 km (UNEP/IUC 1997).

Trends in temperature increase have been monitored since 1850 (Jones & Wigley 1990). The period between 1981 and 1991 was the warmest in the past 200 years in northern high latitudes, which resulted in a photosynthetic increase between 45 °N and 70 °N due to longer growing seasons (Myneni et al. 1997). The photosynthetic rate was found to have increased by 10 % whereas the increase in CO_2 was only 4 %. Consequently, photosynthesis won the race with CO_2 evolution from respiration and combustion. Temperature increase and new nutrients from previously frozen soils are discussed as reasons. Trends in warming of the 20th century are also evident from tree ring measurements from Mongolia (Jacoby et al. 1996).

Another data archive for climatic changes can be found in geographical patterns built from solifluction in mountainous regions. Increased solifluction rates are observed to increase in response to higher soil moisture during spring by snow melt (Harris 1993). Palynological investigations of solifluction lobes provide valuable palaeoclimatic informations (Burga 1993, Smith 1993). Comparisons with ^{14}C-dating of buried organic matter and differentiated soil horizons provide time scales for developments of soil patterns (Veit 1993, Ballantyne 1993). The alterations of phases of solifluction and soil development indicate climatic and ecologic varation (Nesje 1993). Onsets of solifluction appear to be directly related to tree line evaluation (Worsley 1993, Burga 1993). Stabilization due to plant growth prevails during warmer intervals as vegetation restricts the development of solifluction.

The analysis of buried organic matter as well as chemical descriptors of buried stratigraphic sections in permafrost regions provides nice tools to examine and understand former processes on long time scales. Palaeosols of these regions give hints for soil forming conditions under different climate regimes (Höfle & Ping 1996). However, a thorough differentiation into autochthonous and allochthonous material has to be taken into consideration, as this material can have severe impacts on the data interpretation (Vasil'chuk & Vasil'chuk 1997).

CONCLUSION

The contemplation of consequences of Global Warming in northern ecosystems shows a very complex structure. Although I have restrictedmainly to effects on permafrost, there are many other patterns which prevent a homogeneous picture of a greenhouse myself effect. The individual systems have to be regarded as indifferent in their reactions on a scenario of increasing temperature. Site and side effects, reactions of species, of local specifications have to be studied at their individual scales. Although the literature generally imposes the impression of a changing environment in the Arctic while temperature increases, there is no general guideline available for a broader forecast. Specific restrictions have to be considered for individual localities, species or processes.

The application of broad scenarios gained from overall models suffers from the difficulty to transpose their scales to those on which ecological research can be carried out. Although there is no lack in ideas for (short-time) experimental manipulation experiments, they – *vice versa* – suffer from extrapolations and validations to long time scales. Obviously, any experimental approach also has its disadvantages from individual manipulations. However, such experiments can provide frames in which way a system may react on temperature or moisture shifts, nutrient additions or introductions of plants or animals (e.g., Hillier et al. 1994, Harte & Shaw 1995, Chapin et al. 1995) or whether we are just at the moment of instabilities – „turning points“– where system change occurs.

This raises the question how soil scientists and soil biologists should arrange their data with other disciplines. The link to palaeoclimatic data is evident, but where is the match within those data sets? The introduction of results from recent studies into the time scale for the last 200-500 years seems reasonable, as this period may reveal the best highest resolution. Hence, short term effects could become visible and comparable to today's field observations. This time scale can also be used to display anthropogenic effects (Maslin & Berger 1997). Smith & Steenkamp (1990) have produced a scenario for ecological changes by climatic change for a subantarctic island with respect to different levels in ecology. Main effects of their model have to be attributed to side effects and feed–back mechanisms, such as faunal components which are stimulated by changes in soil moisture.

We also have to consider various man-made effects, e.g., CO_2 from burning fossile carbon resources. Recent estimates show that this effect may surpass all natural sources (DIE ZEIT 13.6.97) – effects which confuse all predictions from ecosystem models. Against this background, the discussions launched about the role of arctic systems as CO_2 source or sink may become of secondary nature. With the combustion of fuels, there is also linked an increase in dust particles. Such black particles from soot, when incorporated into snow, have the potential to decrease the reflectivity of the snowpack (Maxwell & Barrie 1989) and to force the melting process. Doubleday et al. (1995) have shown relationships between black particles and the occurrence of several diatoms in Arctic lake sediments. They conclude an interference between this effect and very recent climate change.

Similarly, changes in land use, mainly deforestation, are further facts with strong impacts on climatic effects (FELDMAN 1992). Deforestation can cause a cooling and thus maintain the altered climate regime (BONAN et al. 1992). The general warming of continents will have consequences in increases of growing-degree days of 30 to 40 percent along with a frost-free period lengthening by 20 to 40 percent. Shifts in treeline can be anticipated to reach about 200 to 300 km north, which further shifts the limits for agriculture and forestry (MAXWELL & BARRIE 1989). The warming may produce more favouroble conditions for landuse, but the actual state of soil nutrition will limit this factor, a fact which has been discussed also by SCHLESINGER (1990). Such shifts in treeline and landuse will be accompanied by human settlements with all their consequences of disturbances like pollution.

Many consequences are still behind a curtain of non-available informations. Some, indeed, could become available by more integrative research between more physically orientated research, ecological/ecophysiological research, and palaeoenvironmental research. In ecology, we still suffer from the lack of long term studies which would allow us to make realistic and data-based forecasts for future scenarios, although some long term projects (e.g. LTER, ITEX) have been launched recently. TILMAN (1989) showed by reviewing papers in ecology between 1977 and 1989 that less than 2 % of field studies lasted for 5 years or more. Hence, a main task remains in the differentiation between an environmental manifested trend and environmental noise.

The task of „looking forward" is necessary but arbitrariness remains to a great extent. This holds especially true when looking on other sources of global change rather than man's population increase and trace gas trends: CHARLES (1997) reviews papers in which he points to changes in ocean circulation and the impacts of effects on tropical oceans in this scenario. NICHOLIS (1997) could show that there is a positive trend in wheat yield in Australia due to recent climate trends.

I am aware of the problems that different scientific disciplines talk in different „languages" and operate at different scales of time and space; however, this should not hinder a brainstorming process for new ideas in research programs and – possibly – better extrapolations of actual data into future aspects. Also, we should make our research efforts better applicable to those people who are involved in political decisions for future plans. Research on climate change effects cannot be solved only by forecasting global parameters but needs more knowledge about possibilities and reliable values of anticipated scenarios: Not to hold predictions of environmental collapse but to learn more details about the ecological properties of earth's system in relation to man's impact.

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