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Ecosystems and climate interactions in the boreal zone of northern Eurasia

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Abstract
The climate system and terrestrial ecosystems interact as they change. In northern Eurasia these interactions are especially strong, span all spatial and timescales, and thus have become the subject of an international program: the Northern Eurasia Earth Science Partnership Initiative (NEESPI). Without trying to cover all areas of these interactions, this paper introduces three examples of the principal micrometeorological, mesometeorological and subcontinental feedbacks that control climate–terrestrial ecosystem interactions in the boreal zone of northern Eurasia. Positive and negative feedbacks of forest paludification, of windthrow, and of climate-forced displacement of vegetation zones are presented. Moreover the interplay of different scale feedbacks, the multi-faceted nature of ecosystems–climate interactions and their potential to affect the global Earth system are shown. It is concluded that, without a synergetic modeling approach that integrates all major feedbacks and relationships between terrestrial ecosystems and climate, reliable projections of environmental change in northern Eurasia are impossible, which will also bring into question the accuracy of global change projections.

Keywords: northern Eurasia, biogeophysical and biogeochemical feedbacks, boreal forest and bog, windthrow

1. Introduction

Climate and terrestrial ecosystems interact to enhance and/or to moderate change, making change nonlinear and possibly transitional. The present situation requires new scientific approaches as the equilibrium has been disrupted and a mounting body of evidence shows changes in the states of both the ecosystem and climate, with human impact/reactions contributing to the swiftest of these changes (ACIA 2005, IPCC 2001). Contemporary climatic change in northern Eurasia is among the largest in the world, is projected to remain so, and may affect the global climate system (NEESPI 2004). Ecosystems here are vulnerable to external forcing, especially along their boundaries (in transient zones) and, when affected, may exercise important controls on the global Earth system. This situation raises the stakes in our quest for understanding of multifaceted processes that control natural interactions

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(feedbacks) and forced impacts and systems' responses in northern Eurasia and makes the need for this understanding urgent. Below, the problem is introduced (section 2) and three examples of how the regional ecosystems in northern Eurasia interact with the climate on different spatial scales are presented (section 3). Then it is concluded that a synergetic approach with a greater attention to these interactions should be employed in projections of global Earth system change (section 4).

2. Interaction of climate and ecosystems in northern Eurasia

A combination of factors, conditions and links makes it very difficult to answer the question about the final sign and magnitude of the terrestrial ecosystems–climate interactions. These interactions can be interpreted in terms of natural biogeochemical and biogeophysical feedbacks (Claussen 2004). The biogeochemical feedbacks are associated with changes of terrestrial biomass, soil chemical properties and microbiology and, thus, with changes of the chemical composition of the atmosphere. Effects of vegetation and soil changes on the surface energy and water cycles are named biogeophysical feedbacks. These feedbacks directly affect surface and near-surface energy, water and momentum fluxes via changes in surface albedo, roughness, moisture availability for evapotranspiration, etc. *Hydrology–vegetation feedbacks* constitute a special subclass because water deficit controls the vegetation growth, and these feedbacks may determine chains of specific biogeochemical and biogeophysical processes, depending on regional climate and ecosystem type. In the context of global climate change, the main attention should be focused on the most vulnerable ecosystems and on 'hot' positive feedbacks which, when initiated, may cause nonlinear run-away processes in the climatic system and the biosphere.

Vegetation is one of the most variable components of each terrestrial ecosystem, except deserts. There are general effects of vegetation on the albedo (usually causing a decrease compared to bare soil, especially in the presence of snow cover) and on surface roughness (usually causing an increase). Vegetation may generate micro- and mesoscale effects of advection and turbulence due to spatial heterogeneity. It enhances regional precipitation and evaporation (Rauner 1972, Pielke 2001). It controls the land structure, preventing erosion, as well as affecting smoothing of the near-surface temperature gradients. Vegetation affects the surface energy balance, controls evaporation, runoff, soil moisture, snowmelt, and the partitioning between sensible and latent heat losses. The direct effects of changing land cover and spatial mosaic then manifest themselves in temperature, the hydrological cycle and atmospheric circulation, thus extending the impacts beyond the region where vegetation is changed (Baldocchi *et al* 2000). These changes in turn may feed back to the vegetation. On a global scale, biophysical land–atmosphere couplings due to (a) interactions between vegetation and snow (Sturm *et al* 2001), (b) desertification processes (Zolotokrylin 2003), (c) interactions between vegetation and bare soil or between different vegetation types (Chase *et al* 2001) and (d) variation of sensible and latent heat fluxes (Chapin *et al* 2000) are the primary paths of interaction between the land surface and the atmosphere in northern Eurasia.

3. Examples of ecosystem–climate feedbacks of different special scales

The previous section represents a brief overview of the problem under study. A more extensive overview is presented in chapter 3.5 of the NEESPI Science Plan (NEESPI 2004). Below we present examples of three spatial scales of ecosystem–climate feedbacks in northern Eurasia: micrometeorological, mesometeorological and subcontinental (near-global) using observational evidence and bioclimatological modeling.

3.1. Example 1. The paludification process in the European taiga

The issue of paludification in forests within the boreal zone beyond the permafrost zone has been widely discussed in the scientific literature (Crawford *et al* 2003, Lavoie *et al* 2005). During the last 10–15 years in northwestern European Russia, increased precipitation (Groisman *et al* 2005) and weak surface drainage across plain watersheds with heavy clay soils blocked by moraines have resulted in surplus water in the upper soil layers and an increasing area of bogged forests. In the European taiga, this surplus water is an indication of the paludification process. Three years of eddy covariance, energy budget and water table measurements were conducted over an ombrotrophic peat bog and a *Sphagnum–Vaccinium myrtillus* spruce forest in the southern taiga in the Tver region, 55°N, 33°E (Kurbatova *et al* 2002). Analysis of these measurements for contrasting weather conditions (warm and dry versus cool and wet) shows that the main differences in the radiation balance structures of the two ecosystems are caused by differences in surface albedo (table 1). Albedo by definition is the ratio of the mean upward (from the surface) radiant energy flux in a given spectral band to the mean downward (to the surface) radiant energy flux in the same spectral band (e.g. Gravenhorst *et al* 1999). Throughout this paper, the shortwave albedo (0.3–3 μm) is considered. During the growing season, the albedo of bog vegetation with lower density and canopy depth is approximately twice that of spruce forest. Differences between longwave radiation balances of the spruce forest and bog are small. Correspondingly, the total effect is that the net radiation of the spruce forested land is higher than that of the bog, with relative differences of about 30% under cool and wet conditions and up to 70–90% under dry and warm ones (table 1). Thus, the paludification process substantially changes the radiation balance structure, causing surface cooling. The current temperature changes across the entire boreal zone of Eurasia are characterized by a strong warming (GCAG 2006). Thus, it could be assumed that, over the European taiga, the warming might be stronger if the paludification process described above did not provide mitigation, in this case, a negative biogeophysical feedback. The hydrology–vegetation feedbacks of paludification are difficult to estimate directly because of the nonlinearity of the

Table 1. Daily mean and standard deviations of the ratios between the radiation balance components of an upper bog and a *Sphagnum–Vaccinium myrtillus* spruce forest. Based on approximately two thousand 30 min-averaged eddy-covariance measurements in the southern taiga in the Tver region site (55°N, 33°E) during the two contrasting growing seasons of 1998 and 1999. B and B_L are net and longwave radiation balances, respectively; A is the surface albedo. Cool and wet conditions were selected in year 1998 and warm and dry conditions in year 1999. Mean daily temperatures, T, water vapor pressure deficit, VPD, and their standard deviations for these two periods are also presented.

Weather type	T (°C)	VPD (hPa)	B_{forest}/B_{bog}	$B_{L,forest}/B_{L,bog}$	A_{forest}/A_{bog}
Cool, wet	10.06 ± 3.16	1.47 ± 1.12	1.28 ± 0.77	1.017 ± 0.028	0.49 ± 0.03
Warm, dry	13.89 ± 0.96	8.86 ± 3.15	1.89 ± 2.18	0.995 ± 0.0096	0.44 ± 0.04

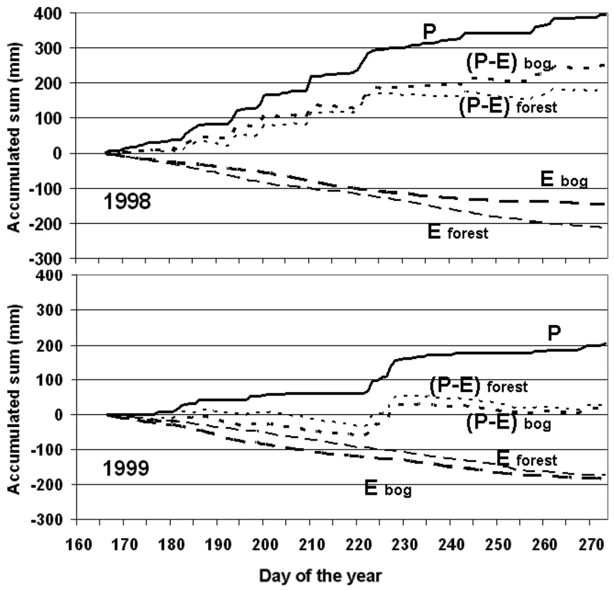

Figure 1. Examples of positive and negative water balances (P–E) of a wet spruce forest and an ombrotrophic peat bog in the Central Forest Reserve (Fedorovskoe, the Tver region, 55°N, 33°E). P and E are the accumulated precipitation and evapotranspiration sums for a common period of observations during a wet (1998) and a dry (1999) year, respectively. For a dry growing season, the total evapotranspiration from the bog is higher than from the spruce forest.

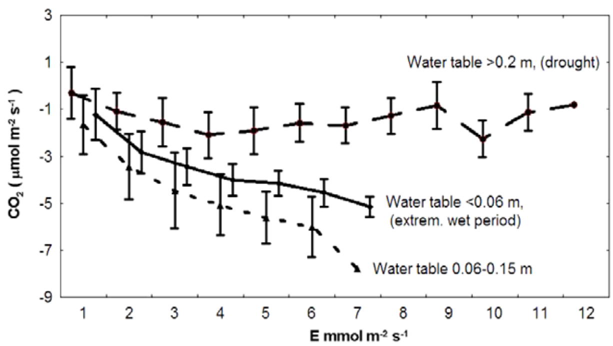

Figure 2. Water use efficiency (amount of carbon dioxide uptake per amount of water use) changes depending on the bog water table level. Evapotranspiration rates (E) change slightly under different water table levels but under both drought (to a greater extent) and excessive water, CO_2 assimilation decreases. Under drought conditions (with low water table), CO_2 assimilation is of low efficiency (0–2 μmol m^{-2} s^{-1}) with stomata closed. Under favorable conditions (with water table 0.06–0.15 m), the CO_2 assimilation increases as evapotranspiration increases with stomata open. Under very wet conditions (with water table less than 0.06 m), CO_2 assimilation decreases compared to favorable conditions but remains greater than under drought. Results based on the suite of meteorological, hydrological and turbulent flux observations in the Central Forest Reserve (Fedorovskoe, the Tver region) during three growing seasons (1998–2000).

relationship between the water budget changes and the net ecosystem CO_2 exchange in forest and bog ecosystems. In southern taiga ecosystems, a positive seasonal water budget, when precipitation, P, exceeds evaporation, E, is typical (figure 1). But, in anomalous dry vegetation seasons (e.g. the bottom panel in figure 1) negative values of P–E, causing a lower water table, affect the ecosystem production processes (figure 2). This figure also shows that the CO_2 exchange of the bog is directly related to evapotranspiration rates and bog water table level, and that the efficiency of the water use by the bog vegetation can be suppressed in both prolonged dry and prolonged wet weather conditions (as defined by the water table position). In the ecosystems studied, the rate of carbon uptake is 63 gC m^{-2} for the wet growing season. On an annual basis, the taiga–bog ecosystems are a CO_2 source to the atmosphere and at geological timescales the carbon pool of terrestrial boreal ecosystems increases due to the peat accumulation. Methane represents another greenhouse gas emitted by bogs and contributing to the total greenhouse effect in the atmosphere. At the Fedorovskoe site, there were no systematic studies of the methane emission dynamics. However, it was shown that for boreal ecosystems of northern Eurasia the pattern of methane emission in the context of climate warming is opposite to the one of carbon dioxide (Velichko *et al* 1998). It was also shown that methane released from thawing permafrost (another important factor of global warming across half of northern Eurasia) generates a strong positive feedback to global warming (Walter *et al* 2006). Another consequence of paludification is that, at the first stage of the process, the rooting systems of trees weaken due to poorer aeration and thus may reduce the critical value of wind speed leading to a windthrow (cf the next example).

3.2. Example 2. Windthrow feedback on the mesoscale

One of the important effects of the present as well as projected climatic changes is the increased frequency of severe extratropical storm events like storms Lothar in 1999 and Kyrill in 2007 (Leckebusch *et al* 2007). These storms may result in widespread damage within forest ecosystems. Wind damage of trees, including uprooting of trees (windthrow) and tree trunk breakage, trigger a number of direct and indirect positive and negative feedback mechanisms in a climate–forest system. Each event of windthrow or break requires a combination of several environmental factors. However, Flesch and Wilson (1999) showed that occurrence of damage events correlates well with tree sway statistics (which in turn are linked to turbulence characteristics of air flow like turbulent kinetic energy—TKE) and with the velocity of air flow pressing on vegetation at its exposed edge (which defines the wind load as $F \approx \int_0^h aU^2\,dz$, where a is the affected plant area (m^2), U—wind speed ($m\,s^{-1}$) and h is the height of an average tree in a stand (m)). In the present study, a three-dimensional atmospheric boundary-layer model, SCADIS, which accounts for flow dynamics within a plant canopy (Sogachev *et al* 2002, Sogachev and Panferov 2006) is used to assess the consequences of a windthrow. A series of model experiments shows that, once a windthrow gap occurs, it results in approximately a twofold increase of TKE and threefold of wind load on the downwind gap edges compared to undisturbed forest (figure 3). If another storm with velocities exceeding the critical one (e.g. 10 $m\,s^{-1}$ for a gap size $\approx$10 h) occurs within the next 3–4 years when the natural regeneration has not had time to show significant growth, a 'secondary' windthrow may occur and the gap size will increase further with an increment depending on wind speed and site conditions. The method was tested at an existing spruce forest in Solling, Germany where a clear cut of approximately 160 m $\times$ 160 m ($\approx$5 h $\times$ 5 h) was made in 2003. Numerical experiments carried out with SCADIS allowed mapping of the areas of high wind load and turbulence around the clear cut. Directly after the cutting in 2003 the first winter storm in 2004 coming from the west resulted in a windthrow approximately 20 m wide on the downwind SE edge of the forest within the mapped area. The critical wind velocity for the 'secondary' windthrow under the west wind is predicted to be 7 $m\,s^{-1}$ in the crown space. These critical values were exceeded during the storm Kyrill, three years after the 'primary' windthrow. The west wind of Kyrill resulted in a secondary windthrow at the downwind edge which concurred with the model predictions. The anticipated higher frequency of severe storms, therefore, will increase the probability of 'secondary' and subsequent windthrow events, providing a positive biogeophysical feedback to wind forcing causing the damaged area to grow. According to Knohl *et al* (2002) windthrow would result in $\sim$164 mmol CO_2 m^{-2} day^{-1} efflux from the boreal forest ecosystem, adding a positive biogeochemical feedback to a changing climate. Some modeling studies (Betts 2000) have suggested climate cooling caused by changes in surface albedo (cf examples 1 and 3) due to large-scale loss of boreal forest cover might produce a negative feedback to global warming. Although the windthrown areas could hardly be described as 'large scale', even in the vast boreal forests of Canada or Russia, subsequent factors like insect infestations could lead to a considerable increase of damaged areas (cf Ravn 1985).

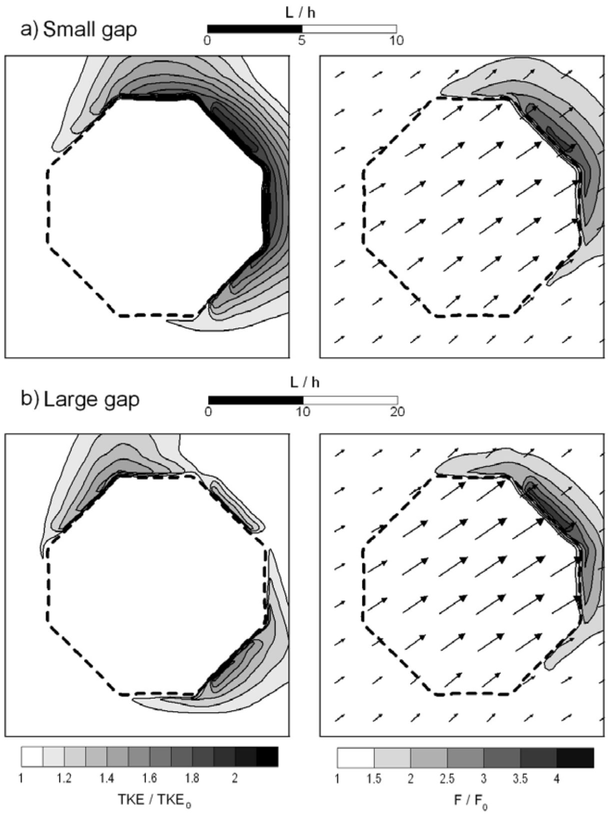

Figure 3. Distribution of turbulent kinetic energy (TKE) (left panels) and wind load, F (right panels) over windthrow gaps of different size, L: (a) small ($L/h \approx 12$) and (b) large ($L/h \approx 22$), where h = mean tree height. The TKE and F are normalized on their values above the undisturbed forest: TKE_0 and F_0, respectively. Arrows in the right panels denote wind velocity and direction at the height of maximal plant area density ($\approx 2/3h$). Dashed lines show the borders of vegetation. Note that TKE maxima are displaced depending on gap size providing self-growth of the windthrow gap crosswise to the main flow.

3.3. *Example 3. Role of the biosphere–climate systems interaction in the projections of the future change in northern Eurasia*

Among the major strategies of plants to survive is their ability to adapt to the environment by means of tolerating possible disturbances/extremes. Environmental factors which are, to a large extent, governed by the climate control the spatial distribution of ecosystems and their composition. Basic climate requirements for an ecosystem's survival are typical plant requirements, which involve a range of optimal temperature and moisture regimes. When external forcing and/or feedbacks move a particular ecosystem close to the climatic bounds of its survival, it is in danger. If and when one of these limits is crossed, the ecosystem starts degrading, and a process of its replacement by a new ecosystem accelerates. These principles are the basis for the bioclimatic modeling to assess the land cover changes for the decade 2090 associated with the climate change scenario HadCM2GGa1 based on a scenario of a 1% yr^{-1} increase of greenhouse

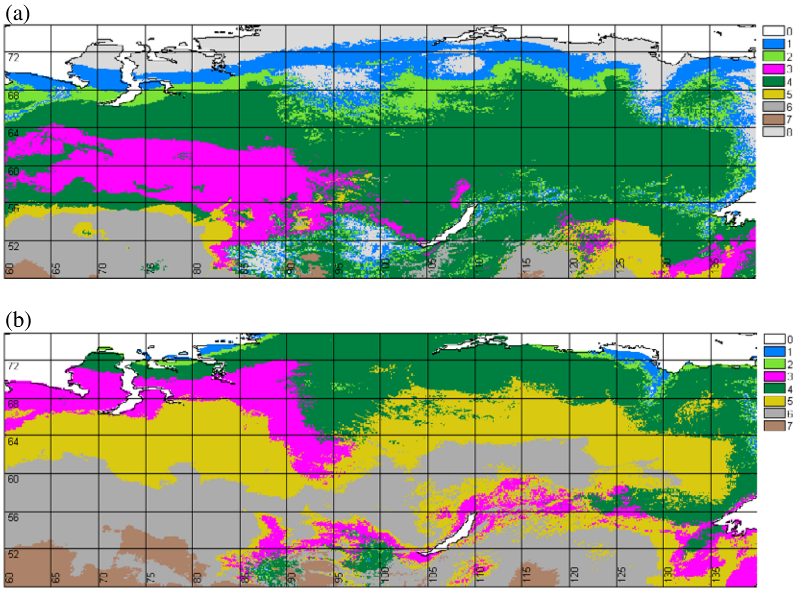

Figure 4. Vegetation distribution over Siberia modeled by coupling the Siberian bioclimatic model (Tchebakova *et al* 2003) with the current climate (a) and that of 2090 (b) evaluated using the climate change scenario HadCM3GGa1 (Gordon *et al* 2000). Water (0), tundra (1), forest–tundra (2), darkleaf taiga (3) and lightleaf taiga (4), forest–steppe (5), steppe (6), semi-desert (7) and polar desert (8).

gases and the Hadley Centre global climate model calculations (Gordon *et al* 2000). Three climatic variables—growing degree days (GDD_5, the sum of daily temperatures above 5 °C), negative degree days (the sum of daily temperatures below 0 °C) and annual moisture index (the ratio between GDD_5 and annual precipitation)—were used as input to the Siberian bioclimatic model (Tchebakova *et al* 2003) to generate a pattern of ecosystem distribution corresponding to the current (figure 4(a)) and 2090 climate (figure 4(b)). These three climatic indices for 2090 are derived from the HadCM2GGa1 output. Figure 4 shows sweeping changes in land cover with the warming anticipated in the last decade of the 21st century: (1) the tundra and forest–tundra zones (currently ~1/3 of the Siberian area) practically disappear; (2) the taiga zones (currently about 2/3 of Siberia) move northward and are reduced to ~40% of the present area; (3) the steppe, forest–steppe, semi-desert and desert areas (practically absent now) are projected to occupy up to 45% (forest–steppe) and up to 15% (steppe, desert and semi-desert) of the area. Summer albedo change by 2090 was calculated as the difference between albedos ascribed to each pixel from Budyko (1974) according to vegetation types predicted for the current and 2090 climates. 'Summer' in Siberia is defined as a period with no snow cover which varies from 2–3 months in the north to 5–7 months in the south (Reference books 1965–1970). Summer albedo would increase over 72% of the area in the southern and middle latitudes in Siberia due to the forest retreat. In the northern latitudes and highlands, tundra would be replaced by forest with decreased albedo in 28% of the territory. By 2090, vegetation change in Siberia, over the territory between 60°E–140°E and 50°N–75°N, would cause a 1.2% albedo increase (figure 5, table 2). Even if we ignore a shorter snow cover period under the predicted warmer climate, these calculations suggest that, considering summer albedo change only, shortwave radiation balance, B_S, in summer would increase by $\sim$30 MJ m^{-2} in 1/3 of the area in the north and would decrease by $\sim$80 MJ m^{-2} in 2/3 of the area in the south. Summer net radiation (B), estimated as $B = 0.83B_S - 0.05$ (Pivovarova 1977), would decrease by about 66 MJ m^{-2} (or about 567×10^{12} MJ for the entire area with positive albedo change, table 2) and thus result in cooling of 2/3 of the area in southern and central Siberia. This pattern of land cover change would increase B by 25 MJ m^{-2} (or about 82×10^{12} MJ for the entire area with negative albedo change, table 2) and thus warm 1/3 of Siberia in the north, enhancing an even greater warming than predicted in the high latitudes. The southern borderline of taiga in Siberia is shaped by forest fires which rapidly promote equilibrium between the vegetation and the climate. In this region, conditions favorable to fire have already been unusually frequent during the past two decades (Groisman *et al* 2007, Soja *et al* 2007). Fire and the melting of permafrost are considered to be the principal mechanisms that facilitate vegetation changes across Siberian landscapes (Polikarpov *et al* 1998). Dramatic changes predicted in the area along the present southern boundary of the forest zone in Siberia (figure 4) are closely connected to the state of permafrost warming. Gradual thawing of the permafrost might impact the natural sequence of disturbances such as forest fires and cause a decline of forest extent and its replacement by steppes in well-drained territories or by bogs in poorly drained territories (Romanovsky *et al* 2007, Soja *et al* 2007). The projection described above and its potential interplay with the permafrost and forest fire dynamics clearly indicate that, for a reliable regional and probably global pattern of climate and ecosystem changes, simultaneous interactive model simulations should be conducted instead of a sequential approach.

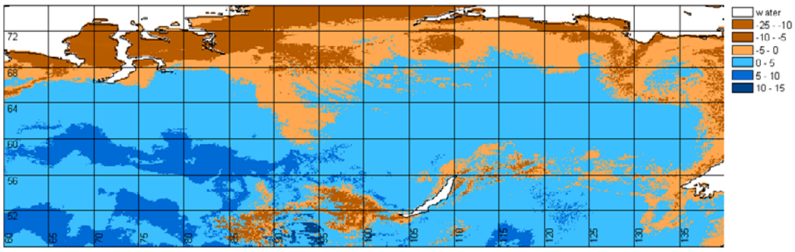

Figure 5. Albedo change in Siberia by 2090 (%) caused by changes in vegetation cover in a warming climate (shown in figure 4).

Table 2. Summer albedo and net radiation balance changes resulting from the estimates of land cover change for 2090 shown in figure 4 according to the HadCM2GGa1 climate change scenario compared to the present climate conditions.

Area (km^2)	Area (%)	Albedo change, Δ (%)	Weighted albedo change (%)	Average albedo change (%)	Net radiation change (MJ)
83 921.3	0.70	<-10	-8.8	-1.2	82×10^{12}
1057 811.9	8.90	-10–5	-66.7		
2157 001.2	18.14	-5–0	-45.3		
7164 174.4	60.24	0–5	150.6	$+2.4$	-567×10^{12}
1379 838.8	11.60	5–10	87.0		
50 442.1	0.42	>10	5.3		
		Mean	$\Delta 1.2$	$\Delta 1.2$	-485×10^{12}

4. Discussion and conclusions

In section 3, three examples of climate–terrestrial ecosystems interactions across three different spatial scales are presented. They cannot pretend to cover all possible interactions within northern Eurasia but already show the scales of interplay and cross-connections. Paludification affects the strength of the tree root system and (in the eastern half of northern Eurasia) the insulation of the permafrost. This in its turn (in addition to the CO_2 and CH_4 uptake/intake) affects land cover by feedback to the forest windthrow and permafrost thaw processes. Climate change forcing that causes an increase in windthrow produces changes of surface albedo leading to surface cooling but may also generate an additional CO_2 release to the atmosphere and thus contribute to further warming. The climate change forcing due to the increase of greenhouse gases in the atmosphere causing large scale changes in land cover in northern Eurasia may generate a suite of positive and negative feedbacks to the surface energy budget that would feedback both to the forcing itself and to the greenhouse gases' emissions.

The biosphere's resilience to external impacts is a key issue of modern ecology and geography, and its vegetation component is among the most labile (e.g. compared to geomorphology, soil, etc). Currently, field studies (alone and/or in combination with regional modeling, e.g. examples 1 and 2) allow off-line studying of the ecosystem–climate feedbacks only on micro- and mesometeorological spatial scales and on relatively short timescales. For larger spatial and temporal scales, we rely upon modeling. We know that a large-scale change in land cover (example 3) would generate additional regional forcing (actually, biogeophysical feedbacks) and thus compromise the general circulation model (GCM) run assumptions. Furthermore, changes in biomass, soil and wetlands carbon, and permafrost thawing (that inevitably must accompany such changes) would generate additional and substantial forcings (biogeophysical *and* biogeochemical feedbacks) on both the GCM forcing *and* the emission scenario itself. For several reasons (NEESPI 2004), the amplitudes of the changes in northern Eurasia in all components of the global Earth system have been and are anticipated to be among the greatest in the world. Thus, for their reliable projections a synergic approach that properly accounts for interactions with terrestrial ecosystems is a must.

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