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Climate warming feedback from mountain birch forest expansion: reduced albedo dominates carbon uptake

Running head: "Expanding mountain forest warms climate"

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Abstract

Expanding high elevation and high latitude forest has contrasting climate feedbacks through carbon sequestration (cooling) and reduced surface reflectance (warming), which are yet poorly quantified. Here, we present an empirically-based projection of mountain birch forest expansion in south-central Norway under climate change and absence of land use. Climate effects of carbon sequestration and albedo change are compared using four emission metrics. Forest expansion was modeled for a projected 2.6 ^{o}C increase of summer temperature in 2100, with associated reduced snow cover. We find that the current (year 2000) forest line of the region is circa 100 m lower than its climatic potential due to land use history. In the future scenarios, forest cover increased from 12 to 27% between 2000 and 2100, resulting in a 59% increase in biomass carbon storage and an albedo change from 0.46 to 0.30. Forest expansion in 2100 was behind its climatic potential, forest migration rates being the primary limiting factor. In 2100, the warming caused by lower albedo from expanding forest was 10 to 17 times stronger than the cooling effect from carbon sequestration for all emission metrics considered. Reduced snow cover further exacerbated the net warming feedback. The warming effect is considerably stronger than previously reported for boreal forest cover, because of the typically low biomass density in mountain forests and the large changes in albedo of snow-covered tundra areas. The positive climate feedback of high latitude and high elevation expanding mountain forests with seasonal snow cover exceeds those of afforestation at lower elevation, and calls for further attention of both modelers and empiricists. The inclusion and upscaling of these climate feedbacks from mountain forests into global models is warranted to assess the potential global impacts.

Introduction

Climate feedbacks from land cover change are complex (Bonan, 2008). Carbon uptake in forests contributes significantly to the land carbon sink (Schimel, 1995) and hence moderates climate warming, while increased forest cover may offset the potential cooling effect of carbon sequestration by decreasing surface albedo (Betts, 2000; Bala *et al.*, 2007). Especially in northern regions with persistent snow cover during an extensive part of the year, decreased reflectance can be significant and dominate changes in the local energy balance (Jackson *et al.*, 2008). Tree-less boreal heaths and shrub vegetation, and alpine tundra are generally completely snow-covered during winter, whereas the canopy of trees frequently lack a stable reflecting snow layer (Holtmeier & Broll, 2005). A better understanding of the trade-off between carbon sequestration and albedo change in boreal regions is necessary for predictions of climate change (Pearson *et al.*, 2013) and for quantification of implications of afforestation policies as a means to combat climate change (Canadell & Raupach, 2008).

Global, hemispheric and regional carbon budgets indicate an important land sink at higher latitudes, usually attributed to forest (Pan *et al.*, 2011). High altitude forest expansion – when disregarding anthropogenic drivers – is driven primarily by temperature (Körner & Paulsen, 2004). However, current altitudinal forest distribution in Eurasian mountainous areas has been lowered by past land use, e.g. summer farming, domestic and semi-domestic livestock grazing, heath burning, logging, and fodder collection (Bryn & Daugstad, 2001; Gehrig-Fasel *et al.*, 2007), and may in many places be maintained by a continued high grazing pressure (Hofgaard, 1997; Gehrig-Fasel *et al.*, 2007; Bryn, 2008; Aune *et al.*, 2011). Consequently, reduced land use intensity in combination with climate warming may lead to a significant increase in forest cover in mountainous regions while sustained land use would hamper a temperature-driven forest expansion. In projections of changing forest cover from process-based models, bioclimatic control of forest expansion dominates limitations from land use (Smith *et al.*, 2001; Hickler *et al.*, 2012). Expansion rate and growth of high-altitude forests is, in addition to being driven by temperature and land use, controlled by factors restricting tree recruitment, such as seed production and dispersal, soil conditions, and resistance of resident alpine vegetation to invasion (Cairns & Moen, 2004; Dullinger *et al.*, 2004; Speed *et al.*, 2010; Walker *et al.*, 2012; Milbau *et al.*, 2013).

The importance of the surface albedo effect for periodically snow-covered boreal regions is larger than in temperate and tropical regions, where reduced cloudiness from deforestation is the main biophysical factor changing absorption of solar radiation (Bala *et al.*, 2007) though increased evapotranspiration could have an opposite effect (Swann *et al.*, 2010). Climate feedbacks of afforestation (Betts, 2000) or deforestation (Bala *et al.*, 2007) on a global and biome scale have been assessed as idealized scenarios where land cover changes were radical and occurred instantaneously or during a very short period. Others have used coupled earth-system models to evaluate the sensitivity of climate to changes in surface reflectance and carbon pools at the global scale (Claussen *et al.*, 2001; Sitch *et al.*, 2005). The literature focusing on smaller scales is usually focused on biofuels areas (Bright *et al.*, 2011; Loarie *et al.*, 2011) or forest disturbances (O'Halloran *et al.*, 2012). We are not aware of climate feedback studies specifically regarding mountain areas, although expanding mountain forest and advancing treelines are a common phenomenon and expected to become more dominant under climate warming (Harsch *et al.*, 2009). There is a strong need for better predictions of shifts in future vegetation in northern regions and their associated feedbacks (Pearson *et al.*, 2013).

In the last decades, field observations and satellite images have indicated higher terrestrial productivity of deciduous broadleaf forests and tundra at higher latitudes (de Jong *et al.*, 2012) and expansion of forests in the tundra (Kharuk *et al.*, 2010; Van Bogaert *et al.*, 2011; Rannow, 2013). Forest-tundra ecotones have typically lower biomass than productive forests at lower elevations and latitudes (Betts, 2000; Bala *et al.*, 2007), which suggests that the climate feedbacks from changing forest cover in these areas cannot be uncritically extrapolated to zones with expanding forests at higher altitudes or latitudes.

The Dovrefjell-Rondane mountain region in south-central Norway is an area where birch forest is currently expanding at the cost of tundra vegetation (Dalen & Hofgaard, 2005; Hofgaard *et al.*, 2010). Similar to most of mountainous Norway (Rössler *et al.*, 2008; Potthoff, 2009; Bryn *et al.*, 2013) and other European mountains (Gehrig-Fasel *et al.*, 2007), vegetation in the study region has been strongly impacted by previous summer farming (Bryn & Daugstad, 2001). Sheep grazing remains a driver of vegetation change in Norway (Speed *et al.*, 2010), although its importance is expected to be reduced in the future (Lundberg, 2011; Wehn *et al.*, 2012). Here, we use the Dovrefjell-Rondane region as a case study to calculate climate feedbacks of advancing mountain forest. The objectives of our study are i) to quantify changes in land cover, biomass carbon pools and albedo under climate change and absence of grazing pressure; ii) to compare the net climate feedback of carbon uptake and albedo change using common emission metrics and to iii) evaluate the additional climate feedback of changed snow cover under climate change. We employ an empirical, bottom-up approach which is complementary to earth-system models focusing on large-scale climate feedbacks from vegetation changes.

Materials and Methods

Study area

The Dovrefjell-Rondane mountain area in south-central Norway around 62^{o} North (Figure 1) covers approximately 3790 km^2 and ranges in elevation from 243 to 2186 m. The topography varies from gently undulating mountain plateaus to steep valleys, while the dominating acidic Precambrian bedrock is partly covered by thin and discontinuous layers of glacial till. The alpine-boreal vegetation of the study area reflects climatic and altitudinal gradients - from coastal humid to inland continental climate and from the valley bottoms to the mountain plateaus (Bakkestuen *et al.*, 2008). Boulder fields dominate at high elevation, interspersed with snow-bed vegetation and mid-alpine dry heath and meadow vegetation. Dwarf shrubs and lichen heaths are common at low-alpine plateaus. The forest line, here defined by trees more than 2.5 m tall within patches larger than one hectare, is constituted by mountain birch (*Betula pubescens* ssp. *tortuosa*) (Aas & Faarlund, 2000). Forest is defined as area with less than 25 m spacing between each tree, and > 25% tree crown cover. Boreal forest types, bogs and fens, and cultivated land and pastures dominate at the lower sub-alpine elevations (Table 1).

Vegetation transitions under land use and climate change

Vegetation cover was compiled into a seamless map, using results from mapping projects performed between 1980 and 1997 by the Norwegian Institute of Land Inventory. This actual vegetation (AV) map includes 30 vegetation classes and 9 other land cover types (for a full list, see supporting information S1). The classification was based on a combination of species composition, indicator species and vegetation physiognomy. The seamless map was converted into raster format of 10 m resolution and joined with a digital elevation model (DEM). The mapped birch forest line was updated using aerial orthophotos from 2010. GIS-operations were run in ArcMap (version 10.0).

Based on the vegetation distribution of the AV map, the distribution of the potential natural vegetation (PNV) – a hypothetical vegetation state where the biophysical effects of land use have been removed (Somodi *et al.*, 2012) - was assessed under current and future climate. The current PNV was established by locating grid cell points for the upper potential climatic forest line (FL_{upc}) in the AV map. To identify the grid cells with maximum altitude of the local FL_{upc}, a circular neighborhood analysis (focal statistics, an ArcMap function) was implemented, using a search radius of 6.25 km^2 for all grid cells, resulting in the local position of FL_{upc}. Drivers of local differences in FL_{upc} such as aspect, slope, vegetation type and geology were tested (ANOVA). Significant relations between forest vegetation type (a lower FL_{upc} for birch forest of lower productivity) and aspect (north-facing slopes had a lower FL_{upc}) were used to extrapolate FL_{upc} from the maximum altitude grid cells to the surrounding cells within the search radius, resulting in local differences in FL_{upc}. Then, open vegetation types below the local FL_{upc} with a potential for forest growth (e.g. deforested by former land use), were reset to the most probable forest type using empirical studies (Bryn, 2008; Bryn & Hemsing, 2012; Dalen & Hofgaard, 2005) which resulted in the PNV. Further details on GIS operations are found in Bryn *et al.* (2013).

The future PNV was established by determining the FL_{upc} for selected climate scenarios (see below), which then provided the limit for maximum forest expansion. The elevation of FL_{upc} under the present climate was increased based on the projected change in summer T, using a lapse rate of 0.7°C per 100 m altitude.

The transition from open vegetation types to birch forest types was based on vegetation studies from the study area (Dalen & Hofgaard, 2005; Bryn, 2008; Bryn & Hemsing, 2012) (Table 2) and assumed to be influenced by climate only while land use drivers were absent. The changes followed a rule-based envelope method described in detail by Hemsing and Bryn (2012), where vegetation type polygons were split up in layers (envelopes), based on aspect and altitude. Vegetation change was not allowed on land cover types that were not expected to support forest regrowth because of lacking suitable soil (Holtmeier, 2009) or management, e.g. bogs and mires, existing forest, agricultural areas, built-up areas, surface waters, boulder fields and bare rock (Table 1; Table 2). Because wind is a strong determinant of vegetation development above the forest line (through direct wind damage, drought, reduced snow insulation and mechanical damage from wind-blown ice crystals (Holtmeier & Broll, 2010)), forest growth was also not allowed on wind-exposed mountain ridges.

Forest growth was modeled with an iterative model, where forest only advances pixel-wise (25x25 m) from established forest cover. In each iteration step, possible forest expansion is determined based on distance to current forest, the location of FL_{upc} and the restrictions for vegetation change described above. Climatic forest expansion was modeled until the forest expansion reached a steady state, at a rate of 1 m altitudinal advance per year. This rate was an average of estimates from relevant continental climatic regions without interference from grazing, i.e. larch in Siberia (1.5 ± 0.9 m yr^{-1}, continental (Kharuk *et al.*, 2010)), spruce in Alaska (0.8 – 1.0 m yr^{-1}, slightly continental (Lloyd & Fastie, 2002)) and birch and pine in Kola Peninsula (0.6 m yr^{-1}, slightly continental (Mathisen *et al.*, 2013)).

Forest expansion was modeled until all vegetation cover below the FL_{upc} suitable for forest growth had become forest. The ultimate limiting factor for further forest expansion in mountain regions is presence of soil (Holtmeier, 2009), implying that further forest advancement demands soil development, starting with the formation of an organic-rich layer enhancing seedling survival. Soil development involves vegetation development interacting with the geological substrate, climate, topography and time which requires centuries in cold climates on hard Precambrian rocks and boulder fields (Van Breemen & Buurman, 2002), and is beyond the scope of this study.

Scenarios

The regionally best correlated climate variable with the climatically controlled forest lines is the summer tritherm (i.e. mean temperature of June, July, and August) (Dahl, 1998; Körner & Paulsen, 2004; Bryn, 2008). The change in the summer tritherm for a reference period 1961-1990 and scenario period 2071-2100 was calculated using four climate scenarios. The climate scenarios were dynamically downscaled scenarios to the study area from Global Circulation Models (Hadley Centre (HAC): HadAm3H and Max Planck-Institute (MPI): ECHAM4) using two IPCC emission scenarios (A2 and B2) (IPCC, 2007). HAC-A2 and HAC-B2 were downscaled to the study area with a spatial resolution of 55x55 km while MPI-B2 was downscaled to both 55x55 km and 25x25km (Engen-Skaugen *et al.*, 2008) (Table 3).

The regional climate models did not provide predictions of snow cover change. Therefore, snow cover and snow depth for the reference and scenario period in the study region was taken from a CMIP5 simulation (Taylor *et al.*, 2012) using the CESM1.0 global climate model (http://www.cesm.ucar.edu/models/cesm1.0/) following the Representative Concentration Pathway 8.5 (RCP8.5) (Meinshausen *et al.*, 2011). The RCP8.5 is the high end scenario of current climate projections and leads to the strongest radiative forcing in 2100 (Meinshausen *et al.*, 2011), and is comparable with the A2 emission scenario (IPCC, 2007). Monthly simulated snow depth was compared with measured snow depth at four meteorological stations (Fokkstugu, Brekkom, Venabu, Nerskogen) in and near the study region, all of which were located between 770 and 930 m a.s.l.. Simulated snow depth was on average 33% higher than measured snow depth between October and May (30 year averages; 63 cm and 53 cm, respectively; data not shown).

Land cover, biomass carbon stock (see below) and albedo were estimated for years 2000 (AV2000) and 2100 (PNV2100), and for a hypothetical steady state between climate and vegetation when the birch forest had reached its full expansion under climate change (PNV_{stst}). Changes that occurred below and above the FL_{upc} in the AV map were attributed to land use change and climate change, respectively.

Carbon sequestration

Site-specific data on forest biomass pools and growth in the study area are scarce. Therefore, our estimates of birch forest growth were based on a growth curve of unmanaged stands of birch (*Betula pubescens*) in Finland under present climate and climate warming, simulated using a gap-type simulation model (Karjalainen, 1996). Here, forest growth was described as a function of stand age with slow growth at young age, a maximum growth rate at 50 years and a maximum biomass stock at 100 years, after which the biomass is reduced to 70% of maximum biomass and reaches a steady state at a stand age of 150-200 years (Figure 2a). The relation between biomass and stand age was substantiated by inventory data from 2000 to 2004 for birch forest stands from the Oppland County located at least 900 m above sea level (Norwegian Forest Inventory (NFI), data not shown), where most of the birch forest in the study area is located. The NFI data showed that maximum standing volume at 80-120 years stand age was between 25 to 28 m^3 ha^{-1}, or approximately 10.5 to 11.7 Mg C ha^{-1} total (above- and belowground) tree biomass after converting standing volume to biomass using biomass expansion factors for birch (Lehtonen *et al.*, 2004). Standing volume in old birch forest in the Oppland County (>160 years stand age) was between 60 and 70% of maximum standing volume.

We calibrated forest growth curves for the two dominating vegetation types (4a and 4b, Table 2) in the study area. Based on biomass stocks reported for these vegetation types in unmanaged mountain birch in cold, subalpine climate in Northern Sweden, maximum aboveground biomass for lichen-and heath forest was averaged from 4.6 (Bylund & Nordell, 2001) and 5.5 (Starr *et al.*, 1998) Mg C ha^{-1}, and in bilberry and meadow forest from 13.8 (Dahlberg, 2001) and 10.6 Mg C ha^{-1} (Starr *et al.*, 1998). Aboveground biomass was converted to total biomass using a ratio of 3:2 for above- to belowground biomass based on allometric equations for birch (Lehtonen *et al.*, 2004; de Wit *et al.*, 2006) (Table 2). Karjalainen (1996) simulated a 20% increase in biomass in *Betula pubescens* given a 4.3 ^{o}C increase in temperature. The ratio of biomass increase per ^{o}C temperature was used to create forest growth curves under a linear temperature increase of 2.6 ^{o}C from 2000 to 2100 (Figure 2a).

Alpine heath and alpine meadow were attributed a total biomass of 1 and 3 Mg C ha^{-1}, respectively, based on an overview of biomass stocks in a model study of vegetation changes in the Barents region (Wolf *et al.*, 2008) (Table 2). We assumed that absence of grazing lead to an immediate establishment and growth of birch (Hofgaard *et al.*, 2010; Speed *et al.*, 2010).

Changes in carbon stocks were calculated for each pixel in annual timesteps and summarized for different vegetation types.

Surface albedo and radiative forcing

Effects of changing surface reflectance on atmospheric radiative forcing (in Watt m^{-2}) related to changes in vegetation and snow cover were calculated using the DISORT radiative transfer model (Stamnes *et al.*, 1988). Here, daily energy fluxes at the top of the atmosphere are calculated in a single column (1x1 degree). The model is forced with 3-hourly meteorological data (obtained from the year 2004 from the European Centre for Medium-Range Weather Forecasts (ECMWF)). For the snow-free season, fixed surface albedo values in two spectral broad bands (visible and near infrared) for the vegetation classes in Table 2 were derived from

MODIS satellite data for 60° to 70° NB (Zhou *et al.*, 2003; Gao *et al.*, 2005). Albedo (α) for snow-covered surfaces for the same vegetation classes was calculated in the DISORT model using the method described by Betts (2000):

$$\alpha = \alpha_0 + (\alpha_D - \alpha_0)\,(1-e^{-0.2S})$$

where α_0 is the local albedo under snow-free conditions, S is the snow mass in kg m^{-2}, and α_D represents the influence of vegetation type and surface temperature when there is snow cover. α_D receives a maximum value under deep snow cover (based on MODIS satellite data as described above (Gao *et al.*, 2005); Table 2), and is reduced as a function of surface temperature.

Thus, albedo for snow-covered surfaces in a vegetation class varies while the albedo in a vegetation class is fixed in the snow-free season. The model uses present-day concentrations of aerosols which were kept constant during the study period. All changes in radiative forcing were thus driven by changes in surface properties of the study region. Albedo changes after vegetation type transitions were assumed to occur linearly during 20 years, while snow cover changed linearly between 2000 and 2100. To be able to separate the effect from land cover change and climate change, the surface albedo change and radiative forcing was also calculated assuming no change in snow cover. Radiative forcing was calculated for 10-year intervals.

Emission metrics

To compare climate effects of different forcing agents such as carbon sequestration and albedo, a range of emissions metrics have been developed (Fuglestvedt *et al.*, 2003). Typically, emission metrics are used to compare the climate effects of different greenhouse (GHG) gases which vary in radiative efficiency (radiative effect in the atmosphere per unit mass) and lifetime that complicate a direct comparison. Choices are required to make consistent comparisons leading to equally valid, but alternative methods of comparison (Peters *et al.*, 2011a). The most common emissions metrics is Radiative Forcing (RF), particularly when expressed in its integrated form as the Global Warming Potential (GWP), used in the Kyoto Protocol (Aamaas *et al.*, 2013). Another metric is the Global Temperature Potential (GTP) (Shine *et al.*, 2005; Shine *et al.*, 2007), which meets some criticized points of the GWP (see for example discussions in Shine (2009) and Peters *et al.* (2011a)). An objective determination of the "best" emissions metric is impossible, as all require choices that depend on the way climate policy is framed and interpreted (Fuglestvedt *et al.*, 2003; Tol *et al.*, 2012; Aamaas *et al.*, 2013).

The emission metrics RF has been used for evaluation of other agents which cause a radiative forcing, for instance by the IPCC (e.g., Figure 2.20 in Solomon *et al.* (2007)). RF, expressed in terms of carbon, has been used to compare albedo and carbon sequestration from land use change (Betts, 2000; Rotenberg & Yakir, 2010), which has received some critique (Pielke *et al.*, 2002; Davin *et al.*, 2007). The main issue is that land use change leads to changes in the local energy balance, which may not lead to radiative imbalances at the top-of-the-atmosphere, where RF is defined. Nevertheless, the RF is still a useful concept to apply to give a first order comparison of the RF from carbon with that of land use change (Betts, 2000), acknowledging that more detailed climate modeling will refine these results (e.g. Bala *et al.* (2007)).

We compared carbon sequestration with changes in albedo using a variety of emission metrics, i.e. Radiative Forcing (RF), integrated Radiative Forcing (iRF), mean global Temperature change (T), and integrated mean global Temperature change (iT) (Peters *et al.*, 2011a). We choose a variety of metrics, to ensure our results are robust to different value choices. We consider the metrics in absolute form, and not normalized to CO_2, though this is simply a choice of presentation and this choice does not affect the results. The sources and sinks of CO_2 were first converted into RF using standard radiative efficiencies and an impulse response function for CO_2 (Forster *et al.*, 2007; Aamaas *et al.*, 2013). The RF for the change in albedo resulting from vegetation cover change is estimated using a radiative transfer model (see previous section). To be able to compare this with the RF from CO_2 we convert the local RF into a global average RF using the ratio of the areas (A); $RF_{global} = RF_{local}*A_{local}/A_{global}$. This step simply renormalizes the RF from the local to global level. Thus, the metric values represent the global responses due to the changes in the local study area (e.g., global temperature response due to change in the Dovrefjell-Rondane mountain area) and this is a standard application of emission metrics (Fuglestvedt *et al.*, 2003; Aamaas *et al.*, 2013).

Since the changes in carbon uptake and albedo change are a function of time, we estimated the metrics using convolutions with the metrics for a pulse emission (Peters *et al.*, 2011b). The Impulse Response Function (IRF) used for the convolution includes the uptake of CO_2 by the land and ocean sinks (Joos *et al.*, 2013). This overcomes the weakness in other studies (e.g. (Betts, 2000; Rotenberg & Yakir, 2010)) which assume that CO_2 has a constant airborne fraction. The RF gives the RF due to sequestration and albedo change in previous time periods, and using an IRF for the temperature response, this is further distributed in time due to the thermal inertia of the ocean (Peters *et al.*, 2011b). The parameters for the metrics were based on previous work (Aamaas *et al.*, 2013).

Results

Mountain birch forest expansion

Under increased summer temperatures (+2.6^{o}C, HAC-A2 scenario) and in the absence of grazing pressure, the actual vegetation in 2000 (AV2000) was transformed into the potential natural vegetation (PNV) as a function of time (Figure 1, Figure 2b, Table 3, Table 4). The main vegetation transitions were from alpine heath and snowbed vegetation to birch forest. The potential endpoint of change under higher summer temperature was the PNV_{stst} situation, a theoretical steady state that is presented to illustrate the time lag between vegetation cover at its climatic potential and vegetation change limited in its expansion by biotic and abiotic factors. The altitude of the upper potential climatic forest line (FL_{upc}) (Table 4) extended upwards from on average 956 m (AV2000) to 1028 m (PNV), to reach 1067 m after 100 years (PNV2100) which was 100 m below the potential forest line in the PNV_{stst}.

Birch forest expansion was largest initially because of the large area available for forest expansion at lower elevations (Figure 1, Figure 2). Open areas below the FL_{upc} started to fill in with birch forest, simultaneous with expansion of birch forest at the upper climatic forest line in 2000. The vegetation cover that is approximately similar to the potential natural vegetation, unaffected by former land use (PNV), was reached after about 50 years. The transition from actual vegetation to PNV resulted in a 9% increase in birch forest area, while the transition to PNV2100 and PNV_{stst} gave an additional 6% and 20% of forest, respectively (Table 4). New birch forest below the FL_{upc} in the current climate was attributed to absence of grazing pressure, while new forest cover above this line was attributed to climate warming. Thus, 65% of vegetation change by 2100 was a legacy of former land use while 35% related to forest expansion under a warmer climate.

The results of vegetation change were similar for all four scenarios (Table 3). The limiting factor for forest advancement is the altitudinal advance rate, which lags behind predicted temperature changes because of abiotic and biotic interactions. In the hypothetical steady state situation, all vegetation with a potential for forest growth has become forest under a +2.6^{o}C temperature increase. The ultimate limiting factor for further forest expansion in the study region is presence of soil, which is derived from properties of the vegetation categories. The simulated forest expansion using the four temperature scenarios resulted therefore in the same land cover in the hypothetical steady state situation.

Carbon sequestration and albedo change

The dynamics of the vegetation cover change until 2100 (Figure 2b) was the main driver for albedo change (Figure 3a) and change in biomass carbon stock or biomass carbon flux (Figure 2c) with the largest changes taking place in the first 100 years. Conceptually, the biomass carbon flux is found by combining biomass growth per ha (the derivative of Figure 2a) with the area change (Figure 2b). In practice, this calculation is done at the grid level.

The area changes became gradually smaller as the forest limit became higher due to smaller areas available for forest expansion. The carbon flux was small initially due to the form of the biomass growth curve (Figure 2a) and peaked in the latter half of the 21st century because of combined large changes in forest expansion and the maximum forest growth rate at around 50 years. After this, the rate of carbon uptake slowly declined because of lower forest growth rates and smaller changes in forest area. Towards the end of the period, there was loss of carbon from biomass (Figure 2c, Figure 3b) because of aging forest vegetation as shown by the forest growth curve. The period beyond 100 years serves as an illustration of the delay between actual forest development and its climatic potential. Forest development after 2100 is difficult to predict as factors driving birch forest cover and carbon accumulation become progressively uncertain after 2100.

The effect of reduced snow cover on albedo exceeded the effect of birch forest expansion on albedo (Figure 3a). Albedo decreased from 0.459 to 0.421 for the AV2000-PNV2100 transition (only vegetation change), and to 0.303 for the same transition when reduced snow cover was included (Table 3, Table 4). The change in albedo resulted in climate warming for all metrics, and this effect was enhanced when we included the effect of reduced snow cover (Figure 3a, c-f). The metrics for carbon were based on the change in biomass carbon flux (Figure 2c). During the entire vegetation development, the effect of albedo dominated the carbon effect for all metrics. The integrated metrics have higher values because they accumulated the effect of earlier times. The ratio of albedo to carbon uptake effect in 2100 was respectively 10 (RF), 15 (iRF), 11 (T) and 17 (iT) (Figure 3c-f). Considering both vegetation and snow cover change, the ratio of albedo effect to carbon uptake in 2100 for the given metrics rose to 49, 65, 51, and 70 respectively. The instantaneous metrics, RF and T, were similar in magnitude, as are the integrated metrics iRF and iT. The integrated metrics gave slightly higher ratios, indicating that in these metrics the albedo effect is even more dominant.

Discussion

Climate feedback from mountain birch expansion

Mountain birch forest expansion due to climate warming and reduced grazing pressure in the Dovre-Rondane region in south-central Norway resulted in a net warming climate feedback, where climate warming driven by a lower albedo strongly dominated climate cooling from carbon uptake in biomass. Our empirically-based results from a high-latitude case study provides insights on climate feedbacks from vegetation changes at a spatial scale and landscape heterogeneity that is usually too detailed to be incorporated in global models (i.e. (Betts, 2000; Claussen *et al.*, 2001; Sitch *et al.*, 2005; Bala *et al.*, 2007). Climate feedbacks may occur at scales smaller than grid cells commonly used in global change models, a point which was made earlier by comparing temperatures observations at forested and cleared plots which showed local cooling effects from deforestation (Lee *et al.*, 2011).

Even when adjusting for carbon densities of different forest types, our study indicates a higher feedback from expanding high latitude mountain forest than reported earlier (Betts, 2000; Claussen *et al.*, 2001; Chapin *et al.*, 2005; Sitch *et al.*, 2005; Bala *et al.*, 2007). In fact, the feedback specifically calculated for Nordic countries was cooling (Betts, 2000). These studies used biomass carbon densities for Nordic or boreal forest of 120 Mg C ha^{-1} (Betts, 2000), 55 Mg C ha^{-1} (Claussen *et al.*, 2001) and 50 Mg C ha^{-1} (Bala *et al.*, 2007). Biomass densities in Norwegian and Swedish forest are lower, i.e. 39 Mg C ha^{-1} (de Wit *et al.*, 2006) and 35 Mg C ha^{-1} (Ågren *et al.*, 2007), while mountain reported birch forest biomass densities vary between 8 and 17 Mg C ha^{-1} (Starr *et al.*, 1998; Bylund & Nordell, 2001). Using the mean biomass density for Sweden and Norway instead of the original biomass densities results in an increase in the ratio of albedo to carbon feedback for Nordic forest from 0.5 (cooling) (Betts, 2000) to 1.6 (warming), and from 2 to 3 for northern boreal forest (Claussen *et al.*, 2001). Similarly, by using the mountain birch biomass density of 13 Mg C ha^{-1} from our study, the ratio of albedo to carbon feedback becomes 5 (Betts, 2000) and 9 (Claussen *et al.*, 2001), closer to the range ratios of in emission metrics that we found, i.e. 10 to 17. More differentiation of biomass pools related to vegetation type would improve model simulations of climate feedbacks of land cover change.

Rates of vegetation change

The steady-state vegetation cover from our model indicated that the vegetation state in the study region in 2100 was several hundred years behind its climatic potential. The rates of vegetation change we employ are result in much smaller vegetation change than the instantaneous land cover changes in many large-scale modeling studies (Betts, 2000; Claussen *et al.*, 2001; Moen *et al.*, 2004; Sitch *et al.*, 2005; Bala *et al.*, 2007; Wramneby *et al.*, 2010). Empirical studies of forest encroachment rates in boreal regions (Gehrig-Fasel *et al.*, 2007; Van Bogaert *et al.*, 2011; Hofgaard *et al.*, 2013), supported by experimental manipulations showing strong nutrient and herbivory limitation of birch growth in alpine and tundra vegetation (Olofsson *et al.*, 2009; Hofgaard *et al.*, 2010; Speed *et al.*, 2010; Zamin & Grogan, 2012), could be used to validate modeling of vegetation dynamics to a larger extent than is done today.

Other potential climate feedbacks from mountain forest expansion

Our analysis did not include changes in soil carbon, implying that the soil carbon stock does not change with forest cover and temperature within the time frame of this study. So far, the evidence for combined effects of temperature and afforestation effects on soil carbon storage appears to be inconclusive. Tundra soils can support similar carbon densities as forests (Sjögersten & Wookey, 2009). Temperature increases and increased forest cover may promote soil organic decomposition, the latter through increasing winter soil temperature from insulation with thicker snow cover while increased litter inputs would promote soil carbon storage (Knorr *et al.*, 2005; Kammer *et al.*, 2009). Environmental controls of soil organic matter in the forest-tundra ecotone clearly deserve more attention.

Another potential climate feedback from expanding forest that we did not include is enhanced transpiration and increased water vapor, which is thought to be of more relevance in tropical than in boreal regions (Bala *et al.*, 2007). However, Swann *et al.* (2010) and Bonfils *et al.* (2012) showed that expanding deciduous forest in the arctic increased evapotranspiration and atmospheric water vapor, resulting in a warming feedback that was stronger than the effect from albedo only. Such processes could be relevant in our study area and would further strengthen our estimated warming feedback from expanding birch forest.

Albedo effect of reduced snow cover

Reduction of snow cover in our study area had a considerably larger effect on radiative forcing than changes in forest cover, contrary to a study of feedbacks from reduced snow cover and expanding forest in Arctic Alaska, where changes in vegetation cover were expected to exacerbate the warming effect of reduced snow cover with a factor of 2 to 7 (Chapin *et al.*, 2005). The prediction of reduced snow cover in our study area is consistent with a set of high-resolution regional climate models that all simulated decreases in snow cover in Northern Europe under climate warming, despite increases in precipitation (Raisanen & Eklund, 2012). However, simulated snow depth was 33% higher than local measurements of snow depth, possibly indicating that the predicted decline of snow cover under climate change was overestimated. Additionally, small-scale variations in snow cover – associated with variations in albedo - related to topography and wind is a known driver of plant growth and diversity in the Dovre-Rondane mountain area (Evju *et al.*, 2012). Simulations of radiative forcing of reduced snow cover need validation with empirical methods for more reliable quantification.

Relevance to mountain forest expansion elsewhere

The climate warming feedback from expanding birch forest from our high-latitude case study is relevant in other regions where deciduous tree species dominate the forest line. In particular at high latitudes, the most wide-spread tree species are in forest lines are mountain birch (the species in our study area) and *Larix* species, where mountain birch dominates in the Nordic countries, north-western Russia and the Ural mountain range, Iceland and Scotland (Walter & Breckle, 1989). *Larix* species dominates in large parts of Siberia and Canada (Walter & Breckle, 1989). These areas overlap with regions where currently a 'greening' trend is observed (Zhou *et al.*, 2001; Piao *et al.*, 2006), although to what extent forest expansion of deciduous species contribute is not known. Birch and *Larix* species also occur world as co-dominants or local dominants in evergreen-dominated forest-tundra zones in North America, Central Europe, and Eastern and Central Asia (Walter & Breckle, 1989). Common for the high latitude deciduous forests are relatively low biomass pools compared with forest at lower elevations and altitudes (Olson *et al.*, 1985), seasonal snow cover, and an expectation of climate warming driving both vegetation and snow cover changes (ACIA, 2005).

Forest expansion as legacy of former land-use

Circa two-third of the change in vegetation cover in our study was associated with forest regrowth at altitudes below the estimated natural forest line, related to former land-use. Lowered forest lines from human activities are common also in other mountain regions in Europe (Gehrig-Fasel *et al.*, 2007; Batllori & Gutierrez, 2008; Bryn, 2008). In mountain regions with little history of land use (e.g. in North America and Russia (Roberts *et al.*, 2006;

Dyukarev *et al.*, 2011)), forest expansion could be more limited than we found here. Including land-use as a driver of past and current vegetation patterns in large-scale vegetation modeling would improve credibility of predictions of land cover under global change.

Synthesis

Our empirically-based results of climate feedbacks of expanding mountain forests indicate a stronger warming feedback than previously found for boreal forests, primarily because of low carbon storage in high-altitude forests. Afforestation of mountain regions, specifically in high latitudes, must be considered as a counter-effective tool for climate mitigation. Climate feedbacks from expanding mountain forests present a challenge for accurate representation in global climate models, because the process occurs at a scale and heterogeneity that may be difficult to capture by the size of grid cells employed in global models. The positive climate feedback of high latitude and high elevation expanding mountain forests with seasonal snow cover exceeds those of afforestation at lower elevation and latitudes, and calls for further attention of both modelers and empiricists.

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List of Figures

Figure 1 Study area Dovrefjell-Rondane mountain region in Norway. Vegetation state under four scenarios is shown for a part of the study area (right-hand panels). Colours depict area covered birch forest under actual vegetation in 2000 (AV2000), under potential natural vegetation under the present climate (PNV), under PNV in 2100 under climate warming (PNV2100) and when in a steady state with regard to climate warming (PNVstst).

Figure 2 Forest growth, area change and change in forest carbon stock for 2000 (Actual Vegetation, year 0) until 2500 (birch forest cover has reached its climatic potential, Potential Natural Vegetation at steady state). Panel a: Forest growth curves (above- and belowground biomass, in Mg C ha^{-1}) versus stand age for lichen and heath birch forest (A) and bilberry and meadow forest (B) under present climate (solid) and climate change (dotted) line. Panel b: Annual vegetation cover in the study area in 10^5 ha for conifers, birch forest, alpine heath, snow bed vegetation and other land cover (not supporting birch forest growth); Panel c) Annual change in birch forest biomass expressed as biomass carbon flux in Gg C yr^{-1}.

Figure 3 Changes in of albedo, biomass carbon stock and in emission metrics in annual time steps between 2000 (Actual Vegetation, year 0) and 2100 (Potential Natural Vegetation, year 100). Panel a), albedo; b) biomass (above- and belowground) carbon stock in Gg C; c) global radiative forcing in W m^{-2}; d) integrated global radiative forcing (RF) in W m^{-2}; e) global temperature change in K; f) integrated temperature change in K. The metrics cover carbon sequestration (solid line), albedo with constant snow (dashed line), and albedo with snow cover change (dotted line).

Table 1 Land cover characteristics of actual vegetation of Dovre-Rondane mountain area

Land cover type	% coverage
Alpine heath	42.3
Boulder fields and bare rock	21.4
Birch forest	12.0
Snowbed vegetation	7.8
Bogs and mires	4.9
Alpine meadow	4.5
Coniferous forest	2.5
Water	2.2
Agricultural areas	1.7
Other	0.5
sum	100

Table 2 Maximum biomass (above- and belowground) and albedo of vegetation types, and potential transitions of vegetation types (VT) under climate warming.

Vegetation type (VT)	VT code	Productivi-ty class[a]	MODIS land cover type[b]	Maximum biomass Mg C ha^{-1}	Maximum albedo[c] under snowcover	snow-free albedo[c]	Potential new VT
Snowbed, moss	1a	L	G	0	0.831/0.569	0.049/0.248	1a
Snowbed, sedge/grass	1b	M	G	0	0.831/0.569	0.049/0.248	4b
Alpine heath, dry grass	2b	L	OSL	1	0.828/0.569	0.045/0.216	4a
Alpine heath, lichen	2c	M	OSL	1	0.828/0.569	0.045/0.216	4a
Alpine heath, dwarf shrub	2e	H	OSL	1	0.828/0.569	0.045/0.216	4b
Alpine meadow	3	M-H	OSL	3	0.828/0.569	0.045/0.216	4c
Billberry birch forest	4a	L	DBF	17.4	0.360/0.300	0.030/0.250	4a
Lichen and heather birch forest	4b	M	DBF	7.4	0.360/0.300	0.030/0.250	4b
Meadow birch forest	4c	H	DBF	17.4	0.360/0.300	0.030/0.250	4c

a. low (L), medium (M) or high (H).

b. Grassland (G), open shrubland (OSL) or deciduous broadleaf forest (DBF).

c. Visible/near infra red light range

Table 3 Projected forest line, vegetation cover and biomass stocks (above- and belowground biomass) for all climate scenarios in 2100 and at steady state

Global climate model	Emission scenario; downscaling grid (in km^2)	Δ summer T (°C)	Potential altitudinal increase forest line (m) at steady state	Birch forest cover in 2100 (%)	Alpine heath cover in 2100 (%)	Biomass stocks (2100/steady state) (Mg C ha^{-1})
HadAm3H	A2 (55x55)	2.6	371	29.2	32.5	2.51/4.29
HadAm3H	B2 (55x55)	1.8	257	29.2	32.5	2.51/3.73
ECHAM4	B2 (55x55)	3.3	471	29.2	32.5	2.51/4.32
ECHAM4	B2 (25x25)	3.5	500	29.2	32.5	2.51/4.32

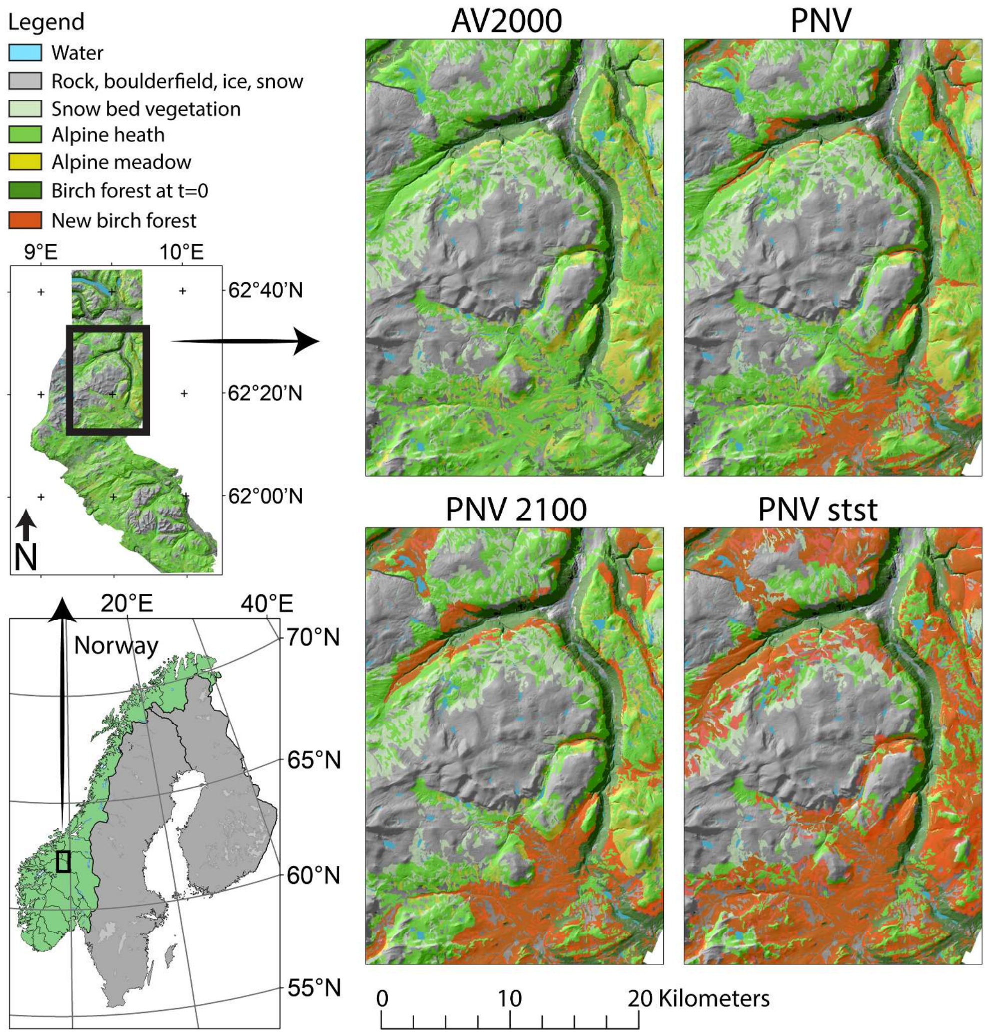
Table 4 Key characteristics of scenarios and resulting changes in forest line, land cover, biomass stock and albedo for Dovrefjell-Rondane mountain area. AV, actual vegetation. PNV, potential natural vegetation. 2000 and 2100 refer to years. Stst, steady state.

	Climate scenario		Forest line elevation[c]	Land cover		Biomass stock	Albedo
Vegetation	ΔT^a (°C)	Δ snow depth[b] (cm)	mean (std) (m)	Birch forest (%)	Alpine heath and snowbed (%)	(Mg C ha^{-1})	(unitless)
AV2000	0	108	956 (24)	12	55	1.58	0.459
PNV	0	108	1028 (125)	21	46	1.81	0.429
PNV2100	2.6	108	1067 (134)	27	40	2.51	0.421
	2.6	63	1067 (134)	27	40	2.51	0.303
PNVstst	2.6	108	1167 (171)	47	20	4.29	0.366
	2.6	63	1167 (171)	47	20	4.29	0.264

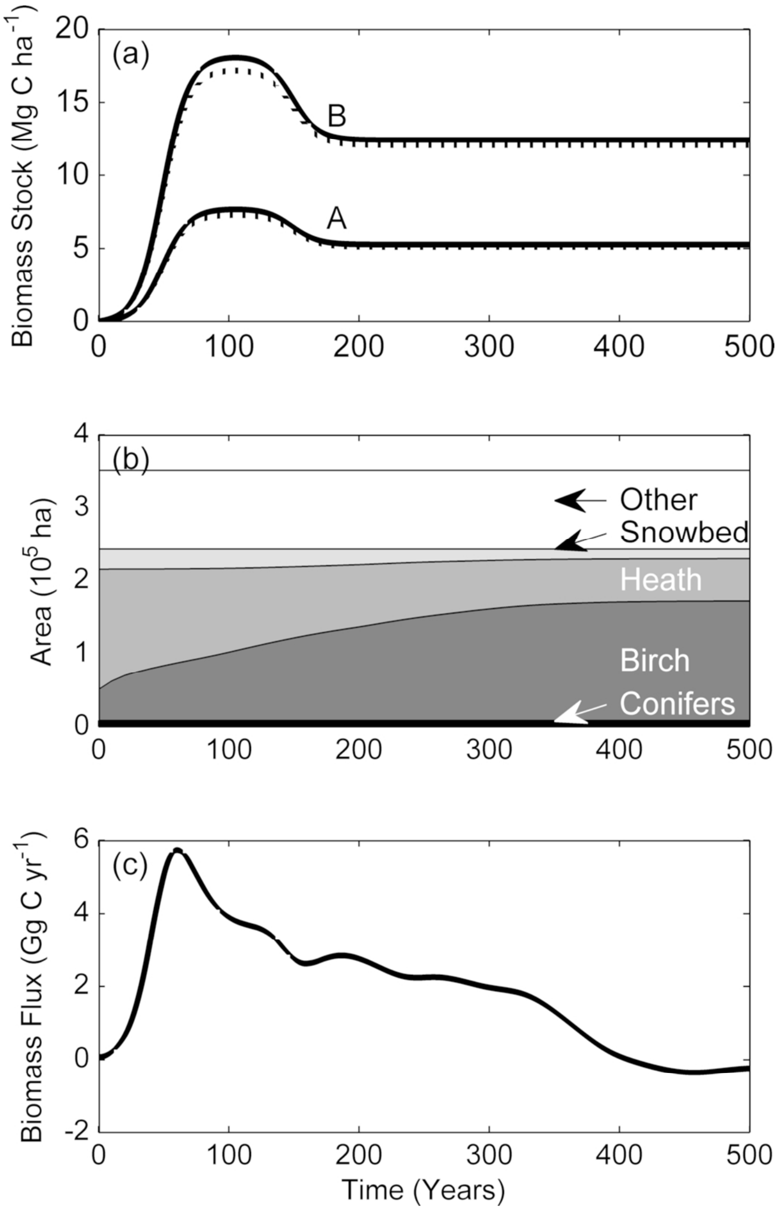
[a] Delta summer (June-August) temperature between control (1961-1990) and projection period (2071-2100).

[b] Mean snow depth in April between control (1961-1990) and projection period (2071-2100).

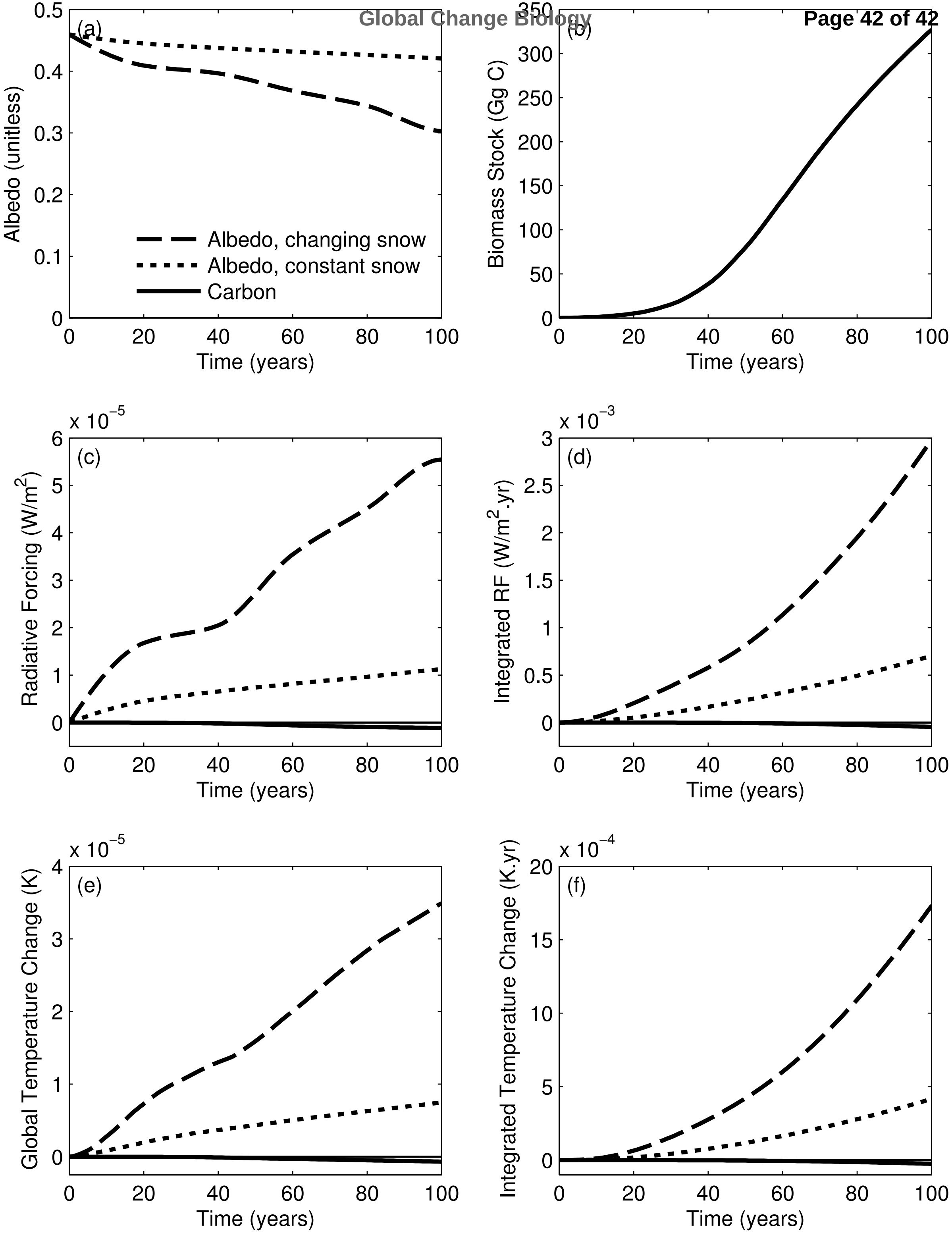
[c] Forest line restricted by non-climatic factors (boulder fields, water). The standard deviation represents natural variations in the forest line.



Study area Dovrefjell-Rondane mountain region in Norway. Vegetation state under four scenarios is shown for a part of the study area (right-hand panels). Colours depict area covered birch forest under actual vegetation in 2000 (AV2000), under potential natural vegetation under the present climate (PNV), under PNV in 2100 under climate warming (PNV2100) and when in a steady state with regard to climate warming (PNVstst).
238x248mm (300 x 300 DPI)



Forest growth, area change and change in forest carbon stock for 2000 (Actual Vegetation, year 0) until 2500 (birch forest cover has reached its climatic potential, Potential Natural Vegetation at steady state). Panel a: Forest growth curves (above- and belowground biomass, in Mg C ha-1) versus stand age for lichen and heath birch forest (A) and bilberry and meadow forest (B) under present climate (solid) and climate change (dotted) line. Panel b: Annual vegetation cover in the study area in 105 ha for conifers, birch forest, alpine heath, snow bed vegetation and other land cover (not supporting birch forest growth); Panel c) Annual change in birch forest biomass expressed as biomass carbon flux in Gg C yr-1.


(a)
Albedo (unitless)
Time (years)
Albedo, changing snow
Albedo, constant snow
Carbon
(b)
Biomass Stock (Gg C)
Time (years)
(c)
x 10^{-5}
Radiative Forcing (W/m^2)
Time (years)
(d)
x 10^{-3}
Integrated RF (W/m^2.yr)
Time (years)
(e)
x 10^{-5}
Global Temperature Change (K)
Time (years)
(f)
x 10^{-4}
Integrated Temperature Change (K.yr)
Time (years)