

Thesis for the degree of Philosophiae Doctor

On biosphere-atmosphere interactions in a changing climate

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List of papers

Paper I

Rydsaa, J.H., Stordal, F., Gerosa, G., Finco, A., Hodnebrog, Ø., 2016. Evaluating stomatal ozone fluxes in WRF-Chem: Comparing ozone uptake in Mediterranean ecosystems. *Atmospheric Environment* 143, 237-248.

Paper II

Rydsaa, J.H., Stordal, F., Tallaksen, L.M., 2015. Sensitivity of the regional European boreal climate to changes in surface properties resulting from structural vegetation perturbations. *Biogeosciences* 12, 3071-3087.

Paper III

Rydsaa, J.H., Stordal, F., Bryn, A., Tallaksen, L.M., Effects of shrub cover increase on the near surface atmosphere in northern Fennoscandia, *Submitted to Biogeosciences, bg-2016-373*

1 Introduction

Vegetation has received more attention as a key element in the climate system in recent years. This is owing to the growing evidence of its important influence on land-atmosphere interactions. Vegetation controls surface properties such as albedo, rate of evapotranspiration and surface roughness. Through these, it determines the fraction of incident radiation that is absorbed by the surface and the partitioning of excess energy into fluxes of heat and moisture that is returned to the atmosphere. Vegetation interacts with the atmosphere through the hydrological cycle and bio-chemical cycles, e.g. by emitting trace gases and volatile organic compounds. These in turn act to alter the atmospheric composition. Changes to the vegetation cover result in alterations in all of these properties and processes, and have the ability to substantially alter weather and climate on local as well as global scales.

Anthropogenic influence on the vegetation cover, whether direct, through deforestation and land use change, or indirect, by influence of pollution and global warming, results in alterations of biosphere-climate interactions. The interactions result in feedback loops that may re-inforce or dampen any initial forcing. Depending on the type and location of the vegetation, the net effect of these feedback loops will vary (Bonan, 2008). Increased knowledge of these feedbacks to climate may, in addition to improve climate change projections, motivate and guide mitigation strategies and control measures of anthropogenic influence on the biosphere. It is of great scientific interest to increase our knowledge of the full reach of our influence on the climate system through direct and indirect alterations to the terrestrial biosphere.

One important anthropogenic influence on vegetation is through air pollution. The air pollutant that causes the most damage to vegetation worldwide is near surface ozone, which in addition to being an important GHG is a toxic oxidant that can cause damage to human as well as plant tissue (The Royal Society, 2008). The adverse effects on vegetation were first recognized in the 1950's, and estimation of damage by means of measurements, modelling and model development, today comprise a broad field of research. Adverse effects include reduced biomass production, which affect crop yield and forestry causing severe economic loss (Van Dingenen et al., 2009; Yue and Unger, 2014). It also affects biodiversity due to cross species differences in ozone sensitivity, and reduces carbon sequestration, thereby increasing atmospheric carbon (Sitch et al., 2007; Pacifico et al., 2015).



Figure 1: One healthy and one damaged leaf as a result of ozone exposure.

In order to estimate current and future damage to vegetation, one needs to take into account not only the emissions of precursor gases, but also the meteorological factors that influence the ozone photochemical production and transport, in addition to the range of meteorological, chemical and biological conditions that influence the uptake in the exposed vegetation. To make accurate estimates of current and future ozone uptake and corresponding potential damage, advanced modelling tools are required and are under constant development. One aim of this thesis is to address the suitability of one such tool for estimating fluxes of ozone to vegetation.

The global vegetation cover is changing in response to human activities also through the effect of global warming. The impact of increased temperatures and CO₂ levels on ecosystems is observed around the globe, and one of the most sensitive areas is in the northern high latitudes (Bonan, 2008). The Arctic is warming at about twice the rate as the rest of the globe (IPCC, 2007), and this is causing what is known as the Arctic greening. This refers to the observed increase in high latitude biomass and northward expansion of boreal trees and shrubs in high latitude ecosystems over the past decades. The enhanced warming in high latitudes is in part caused by local feedback mechanisms (Serreze and Barry, 2011), and evidence gathered over the past decades show that the Arctic greening contributes to these feedbacks. However, there are still large uncertainties with regard to both the sensitivity of the high latitudes ecosystems to climate change, and the strength of the corresponding feedbacks, in both present and future climate. In this thesis, we aim to contribute to the increased understanding of the biophysical feedbacks accompanying particular changes in the high latitudes ecosystems.

1.1 Motivation and research objectives

The biophysical properties of vegetation in many ways represent the most direct link between the biosphere and the atmosphere. Focus in the research conducted as part of this thesis is on the biophysical aspects of biosphere-atmosphere interactions; from leaf level details, to biophysical effects on the atmosphere resulting from changes on regional ecosystem level.

Numerical modelling is one of the most versatile tools available to us to quantify and investigate the complex interactions of the biosphere and atmosphere, on both leaf level and regional to global scale. In order to make confident estimates of the atmospheric effects of changes in the terrestrial ecosystems, a precise description of vegetation properties, behavior and interactions with the environment is important.

On leaf level, Paper I in this thesis investigates the modelling of one important process; the parameterization of stomatal openings on the leaf surfaces, which act to control the release of water vapor and entrance of CO₂ and other trace gases. In this study, the stomatal fluxes are validated in the context of one direct and important application of such parameterizations; the estimation of uptake of near surface ozone in vegetation, with the aim to estimate accurate doses and corresponding damage to vegetation.

In Papers II and III in this thesis, focus is shifted from leaf level to regional scale. On regional scale, a precise estimate of the realistic response of the vegetation or ecosystems to environmental forcing, is important for the confidence we can place on the estimates of the resulting feedbacks to the atmosphere. Focus in these studies is placed on particularly the atmospheric response to semi-realistic estimates of vegetation change in response to climate forcing. For this purpose, we have focused on one of the most sensitive regions on the planet with respect to ecological response to climate forcing; the boreal and Arctic biomes. It has been suggested that these are the biomes with the highest potential to feed back to global climate, resulting mainly from biophysical alterations of the land surface. Furthermore, the work conducted in this part of the thesis was part of an increased focus within our department, which in time developed into the LATICE project, focused on land-atmosphere interactions in cold climates.

Objectives

The development of modelling tools for use in estimating current and future ozone induced damage to various types of vegetation is an important aspect of developing emission control strategies and mitigation planning. In addition to developing new modelling tools, improving and adjusting existing models may prove an efficient strategy. The first study in this thesis represents a validation study with the aim to quantify and conclude on the accuracy of the biophysical representation of leaf level stomatal conductance in a well-known and widely used parameterization scheme. In this study, we aim to conclude on the suitability of the modelling system in contributing to one important issue reliant on this description; the estimation of vegetation damage induced by ozone uptake. Although not the main focus of the study, the results are easily transferable and relevant to other sides of the modelling of biosphere-atmosphere interactions influenced by stomatal conductance; such as evapotranspiration and CO₂ uptake. More precisely our motivation is summarized in research objective 1.

1. Can the Weather Research and Forecasting model (WRF) coupled with chemistry (WRF-Chem), with the Wesely dry deposition scheme, be utilized in estimating ozone-fluxes in vegetation?
 - a) How accurate are the modelled fluxes compared to measurements?
 - b) Are some vegetation types better represented by the model?
 - c) How do known weaknesses of the current model parameterization influence the results?
 - d) Can the model parameterization be improved, if so how?

In the next two studies (Papers II and III), the focus shifts from leaf level biophysics to biophysical properties of vegetation at the ecosystem level. Effects on the atmosphere related to the changes of those biophysical properties corresponding to large scale changes in the terrestrial ecosystem induced by current and future climate change are targeted. In this part of the thesis we aim to apply moderate and realistic changes to the high latitude vegetation cover on a finer temporal and spatial scale as compared to previous studies, and investigate the resulting feedback processes and effects on the overlying atmosphere. More specifically, the research objectives that motivated our investigations on this matter were;

2. How will the biophysical feedbacks, resulting from observed and expected high latitude vegetation changes affect the high latitude atmosphere?
 - a) How will the biophysical changes resulting from northward migrating boreal tree cover influence the surface fluxes of heat and moisture?
 - b) Will changes to the high latitude biosphere include a positive or a negative feedback to an initial warming?
 - c) How will the various vegetation changes in the boreal zone differ in their influence on the near surface atmosphere?
 - d) Will slow migration of large tree species dominate results compared to less drastic, yet faster vegetation changes, such as shrub expansion?
 - e) How will the height and stature of the vegetation influence its effect on the overlying atmosphere?
 - f) In which seasons are the influences of increased vegetation cover more important?
 - g) Are the effects of increased vegetation cover on the overlying atmosphere sensitive to environmental variables such as temperature and snow cover?
 - h) Can we expect an enhanced feedback in the future as compared to the present, and what will the effect of future vegetation expansion have on the atmosphere?

In the following, a brief presentation of the scientific background for of this work is presented in Chapter 2, and the tools and methodology applied is presented in Chapter 3. In Chapter 4 the main findings corresponding to each of the research papers are presented and discussed, along with some ideas for future work in Chapter 5. Finally, the scientific papers are attached, containing further details on methods and results, in Chapter 6.

2 Scientific background

In the following a brief presentation of the scientific background on the subjects of this thesis is given for the purpose of providing a context for the work presented here. The main focus is on recent research on each subject, and on uncertainties still existing within the field, and briefly on how the studies presented in this thesis aim to contribute to limit these uncertainties.

2.1 Near surface ozone and effects on vegetation

Tropospheric ozone is a secondary pollutant, and the main source is photochemical reactions in the near surface atmosphere. The formation rates are dependent on the meteorological conditions and abundance of precursor gases such as NO_x (NO and NO_2), carbon monoxide (CO), methane (CH_4), and non-methane volatile organic compounds (VOCs). In remote areas, and for the hemispheric background concentrations, reservoir gases such as HNO_3 and PAN (peroxyacetylnitrate) are also important in ozone formation, as illustrated in Fig. 2.

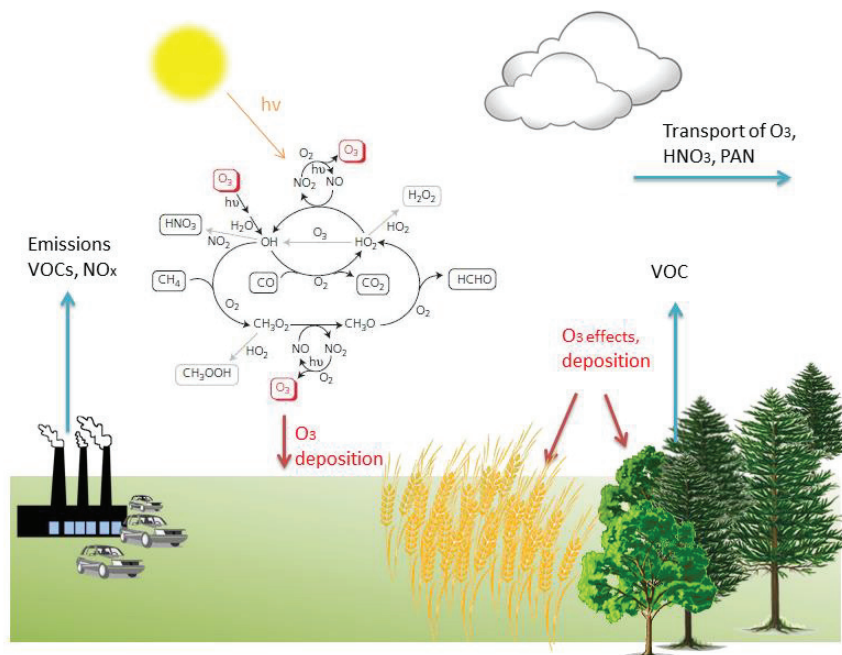


Figure 2: Illustration of the dominant features of tropospheric ozone production (in a high NO_x environment) and loss mechanisms. (Chemical cycle inset adapted from The Royal Society (2008)).

The adverse effects of ozone exposure to vegetation were first recognized in the 1950s and is today well documented (The Royal Society, 2008). Since the 1950's, the northern hemisphere (NH) free troposphere mixing ratios of ozone have more than doubled (IPCC, 2013, Chapter 2.2.2.3), and field experiments across the world have documented that today's background concentrations of ozone are high enough that damage to ozone sensitive vegetation can be expected (Hollaway et al., 2012). Legislations limiting precursor gas emissions may cause the risk of ozone damage to vegetation in Europe to decline in coming decades (Klingberg et al., 2014). However, increased future emissions of precursor gasses related to urbanization and industrialization in many rapidly developing countries, combined with climatic conditions promoting ozone formation, give reason to expect further increase in regional near surface ozone mixing ratios, and NH background concentrations over the coming decades (Ashmore, 2005;Hollaway et al., 2012).

Documented adverse effects on vegetation resulting from ozone exposure include reduced biomass production, early senescence, reduced biodiversity due to cross species differences in ozone sensitivity, and visible injuries reducing economic value (Hayes et al., 2007;Mills et al., 2011a). In some areas, this may have economic effects due to decreases in crop yield and forestry, and in extreme cases affect food security. According to Van Dingenen et al. (2009) the global economic loss due to ozone damage to four common crops; maize, wheat, rice and soybean, was \$ 14-26 billion in the year 2000, with 40% of the damage occurring in India and China. Yue and Unger (2014) estimated an average decrease of 4-8% in gross primary production of crops in the Eastern United States over the decade 1998-2007, with even higher reduction of 11-17% in certain east coast "hot-spot" regions.

Another important effect of large scale ozone damage in coming years is the reduction by ozone on plant photosynthetic rate, which may influence the land carbon sink through decreased CO₂ uptake. This will yield an increase in atmospheric CO₂ leading to an indirect effect on the atmospheric radiative budget, that could potentially double that of the radiative forcing by atmospheric ozone itself (Sitch et al., 2007;Pacifico et al., 2015).

The most traditional means of estimating damage to vegetation is through concentration based indices, which were based on the assumption that the higher concentration of ozone in the ambient air, the higher is the risk of damage to the vegetation. Accumulated ozone concentration

per day-lit hour over the threshold value of 40 ppb (AOT40) is one such index which has been widely used in air pollution regulations in Europe. However, it has become increasingly clear over the past decades, that the risk of damage is more accurately described using a flux based index, linking the risk of damage to the actual uptake of ozone into the vegetation (CLRTAP, 2015). Today, flux based indexes like the POD_Y (phytotoxic O_3 dose over a flux threshold value Y) is regarded more appropriate, and recommended for risk assessment. The POD_Y is based on the stomatal ozone flux exceeding a species specific flux threshold value Y integrated over time, which yields a phytotoxic absorbed dose. This dose is interpreted by empirical species specific dose-response relationships (CLRTAP, 2015; Mills et al., 2011b; Pleijel et al., 2007), which will give an estimate of the actual damage to the plant.

The ozone flux from the atmosphere to the surface can be estimated by using a resistance analogy, based on Ohm's law for electrical currents in a circuit. The procedure is based on placing a number of resistances in series or in parallel to account for the meteorological, chemical and biological processes acting to control the flux of ozone. The flux from the ambient air to the surface is then calculated by multiplying the canopy height ozone concentration with the surface conductance which is given as the reciprocal of the surface resistance;

$$F = g \times C$$

where F is the flux, g is the surface conductance and C is the canopy height ozone concentration.

The ozone flux into the leaves is through the stomata, which is regulated by a number of environmental and biological factors, such as the temperature, light, water availability through the soil, and the evaporative demand of the atmosphere. In addition, the phenology and ozone damage itself can influence the stomatal openings. A common procedure of calculating the stomatal conductance (or conversely, the resistance) while accounting for the various conditions regulating the stomatal opening, is by applying a multiplicative approach based on Jarvis (1976). This approach is recommended in the CLRTAP (2015) mapping manual, and applied in the DO_3SE model (Emberson et al., 2000). Other approaches are also available and used in other models. There are several models in use, and more are constantly being developed.

Our contribution to this development is presented in Paper I of this thesis. In this paper, we aim to test the skill of the WRF-Chem model (Grell et al., 2005), and the Wesely dry deposition

scheme (Wesely, 1989), in estimating ozone stomatal fluxes, by comparing modelled estimates to various types of vegetation. The details regarding the approach taken by this model in calculating stomatal ozone flux is presented in Chapter 3.3, and in Paper I, Section 2. A summary of our findings is presented in Chapter 4.1 in this thesis.

2.2 Changes in the high latitude ecosystems

The polar amplification of global warming is in part due to a cascade of local feedback mechanisms, such as sea ice decline and snow cover decrease, that act to increase the initial greenhouse gas forcing (Serreze and Barry, 2011). The warming at high latitudes over past decades has led to changes in the boreal and Arctic vegetation cover, and the term “Arctic greening” refers to the observed increase in high latitude “greenness”, commonly related to increased photosynthetic activity and biomass (Xu et al., 2013). Although greening is observed across the circumpolar Arctic and boreal regions, the rate of greening in response to the warming trends is far from homogenous across the area, or ecosystems. The ecological response also varies in speed across vegetation species and ecosystems (Corlett and Westcott, 2013). In the context of current climate change, some vegetation shifts, happening on the order of centuries, may be considered slow, while some are achieved within decades, and as such may be considered fast.

2.2.1 Tree line migration; slow changes

One observed change to the high latitude ecosystems in response to increased temperatures, is the migration of tree lines towards higher latitudes and altitudes. The substitution of tundra and shrub covered areas by boreal forest to the north of the previous boreal edges and the replacement of boreal forest by grasses or temperate trees along the southern edges have been observed across the boreal biome (Soja et al., 2007; Hofgaard et al., 2013; de Wit et al., 2014; Tommervik et al., 2009; Chapin et al., 2005).

Observations confirm anticipated patterns based on reconstructions from paleoclimate data from previous periods of high latitude temperature increase (Davis and Shaw, 2001), and modelled estimates of vegetation response to climate warming (Jeong et al., 2014; Jeong et al., 2011; Strengers et al., 2010; Wolf et al., 2008). Although tree line migration represents a widely documented feature of recent climate response across the boreal forest, there are considerable time lags associated with this type of vegetation shift, in the sense that the climatic conditions are

presently shifting northward faster than the migration speed of the forest ecosystems. The migration of tree lines may as such be labeled a “slow” vegetation shift in relation to climate change.

2.2.2 Shrub expansion; fast changes

In contrast to the slow migration of forest ecosystems, the increase of shrub vegetation cover in the Arctic biomes is considered a much faster response in high latitudes ecosystems. Repeat photography (Sturm et al., 2001b; Hofgaard et al., 2013), satellite studies (Bhatt et al., 2010) and vegetation surveys (Elmendorf et al., 2012; Myers-Smith et al., 2015b) have provided evidence of large increases in the extent of shrubs across the Arctic over just the last few decades. They document that vast areas of previously tundra covered areas have been converted to shrub cover, in addition to increase in shrub height, density and abundance.

Based on a synthesis of field data from Arctic Alaska, Chapin et al. (2005) estimated that since the 1950's, the cover of tall shrubs in northern Alaska has increased by 1.2 % per decade, increasing the total cover from 14 to 20 %. They estimate that a temperature increase of 1-2 K is generally effective in triggering shrub growth within a decade. Based on tundra vegetation surveys of 158 plant communities spread across 46 sites around the globe, Elmendorf et al. (2012) found biome-wide evidence for increased canopy height, litter and abundance of low and tall shrubs with increased summer warming. They found that shrub growth was particularly sensitive to summer warming in locations that were relatively warmer to begin with. They emphasize a dependence of shrub expansion not only on summer warming, but also on soil moisture and presence of permafrost. These findings are supported by Myers-Smith et al. (2015a), who analyzed circumpolar dendroecological data from 37 Arctic and alpine sites in order to estimate the climate sensitivity of shrub growth. They found that sensitivity of growth to warming was higher at sites with higher moisture, and for taller shrubs, growing at the upper ranges of their latitudinal or altitudinal limits, and in areas where permafrost was thawing.

Sturm et al. (2001a) suggest that part of why shrub cover increase is so rapid, is because of shrub-snow interactions triggering positive feedback mechanisms promoting further shrub expansion. In addition to enhanced atmospheric warming, they found that the insulating effect of thicker snow cover in shrub covered areas leads to better insulation of the winter time soil beneath the shrubs, enhancing nutrient supply to further shrub growth.

By investigating dendroecological data from sites across the Arctic, in combination with future climate scenarios, Pearson et al. (2013) predict that about half of the present tundra covered regions could be covered by shrubs by 2050. They conclude that the atmospheric and soil warming found in modelling studies may be substantially underestimated, due to underrepresented shrub expansion, and that the resulting warming should be expected much sooner than estimated. This illustrates the uncertainties related to future estimates of shrub expansion, and associated feedback effects.

Shrub expansion represents a less drastic change in surface properties as compared to forest migration, and thus more subtle influence on the overlying atmosphere. However, the vast extent and timespan of just decades as compared to centuries, makes this vegetation change a key factor in estimating present and future climate response in the high latitudes.

The biophysical effects on the regional atmosphere resulting from various shifts in high latitude ecosystems are the focus of objective 2 in this thesis, and discussed further in Papers II and III.

2.3 High latitude vegetation changes and climate; biophysical feedbacks

Increase in vegetation cover in high latitudes leads to both biochemical and biophysical feedbacks. Extensive evidence gathered over the past decades has established that changes in high latitude ecosystems are part of the feedback mechanisms amplifying high latitude warming (e.g. Sturm et al., 2001a; Bonfils et al., 2012; de Wit et al., 2014). These changes may have the largest direct impacts on the local scale. However, local scale feedbacks may propagate to regional and continental scales through cross-scale links, and possibly lead to critical transitions in the large scale climate (Rietkerk et al., 2011).

2.3.1 The albedo effect

The albedo varies across vegetation types, and is generally higher for short and low vegetation, as compared to taller, more complex canopy vegetation. This is because less short wave radiation is reflected back to the atmosphere from a taller and structurally more complex canopy.

The largest impact of this effect is related to increase of high latitude (evergreen) vegetation where snow cover is present in parts of the year, as snow completely covers low vegetation, while taller vegetation protrudes the snow cover, thereby decreasing the otherwise high winter surface albedo. Through the decrease in surface albedo during the snow covered season, the

boreal evergreen forest has been estimated to have the greatest effect of all biomes on global warming (Bonan, 2008). By investigating the climate benefits of afforestation mitigation strategies, Arora and Montenegro (2011) also found that in high latitudes, the warming effect of decreased surface albedo related to increased boreal forest cover, dominated the cooling effect of increased carbon sequestration, supporting similar findings of Betts et al. (2007).

Compared to evergreen forests, shrub expansion into tundra ecosystems generally leads to a weaker and more seasonally dependent albedo effect. However; Sturm et al. (2005) observed that the presence of tall shrubs protruding the snow cover during winter acted to lower the winter time surface albedo by 30%, which was about two thirds of what might be expected from a tundra to forest change. The decrease in albedo caused melting to start several weeks earlier. They estimated an increase in the absorption of solar radiation by 69-75%, depending on latitude.

While the feedback effects of increased boreal forest cover have been studied in a number of global and regional modelling studies (e.g. Liess et al., 2012; Snyder and Liess, 2014; Arora and Montenegro, 2011), increased shrub cover has so far been examined in only a few modelling studies. A common approach has been applying a 20% increase in shrub land cover north of 60 °N , and investigating the response on the soil and atmosphere (Bonfils et al., 2012; Lawrence and Swenson, 2011; Swann et al., 2010). These studies found that increased shrub cover leads to a mean annual warming with a peak in the spring melting season, mainly due to lowering of surface albedo, and effects on snow melt.

2.3.2 The evapotranspiration (ET) effect

The effect of increased ET associated with higher leaf area index (LAI) and increase in complex canopies was investigated by Swann et al. (2010). Focusing on tall, deciduous vegetation, they found that increased ET caused by an increase of deciduous vegetation species in the Arctic, reinforced by sea-ice interactions, strengthened the greenhouse effect of the overlying atmosphere leading to a temperature increase of the same magnitude as the one caused by the albedo effect.

As different types of vegetation have varying water use efficiency, photosynthetic rate and capability to draw water from the soils and transpire it into the atmosphere, the amount of water retained in the ground and canopy in response to climate forcing will vary. In addition, the total

surface of the canopy will determine the amount of rainwater that is intercepted and how much can evaporate from the canopy and surface beneath it. The combined effect will give the rate of evapotranspiration from the surface, and determine the amount of water vapor that is returned to the atmosphere through latent heat fluxes. This could in turn affect regional or continental scale evapotranspiration-precipitation feedbacks (Thompson et al., 2004; Beringer et al., 2005; Eugster et al., 2000), as further discussed in Section 4 and in Paper III.

Changes in ET will affect the partitioning of excess energy into sensible and latent energy fluxes (the relationship called the Bowen ratio). Beringer et al. (2005) measured in general warmer and drier fluxes along a transition zone going from Arctic tundra to boreal forest, confirming findings by Thompson et al. (2004), who also found increased sensible to latent heat flux ratio associated with taller and more complex canopies. The change in this ratio resulting from shifts in the extent of the boreal forest was also investigated in Paper II in this thesis.

In addition to the albedo and the ET effect, observational and modelling studies have evaluated a range of other effects of increased high latitude vegetation cover, such as increased snow cover (Sturm et al., 2005), effects on soil moisture and temperature affecting permafrost thaw (Blok et al., 2010; Lawrence and Swenson, 2011), and dynamic influence on the strength and position on the polar jet stream (Liess et al., 2012). Further investigation of the sensitivity of the atmosphere to the biophysical changes related to specific shifts in the high latitude vegetation is the focus of Papers II and III in this thesis, and a summary of the findings is presented in Chapters 4.2 and 4.3, respectively.

The relative importance of each feedback loop will vary with the type and location of the vegetation. Bonan (2008) argues that at low latitudes, where broadleaved tree species dominate in a humid and warm climate, enhanced tree cover will have a net cooling effect, due to the dominating effect of increased evapotranspiration (ET) and decreased warming by enhanced CO₂ sequestration over other feedbacks like warming due to decreased albedo. In contrast, the main effect of increased cover of evergreen needle leaved trees at high latitudes would be enhanced warming due to the strong effects of lower albedo, especially by masking of the snow covered ground in winter and spring. The net effect of various feedback mechanisms may thus vary across vegetation types, as well as location, as illustrated in Fig. 3.

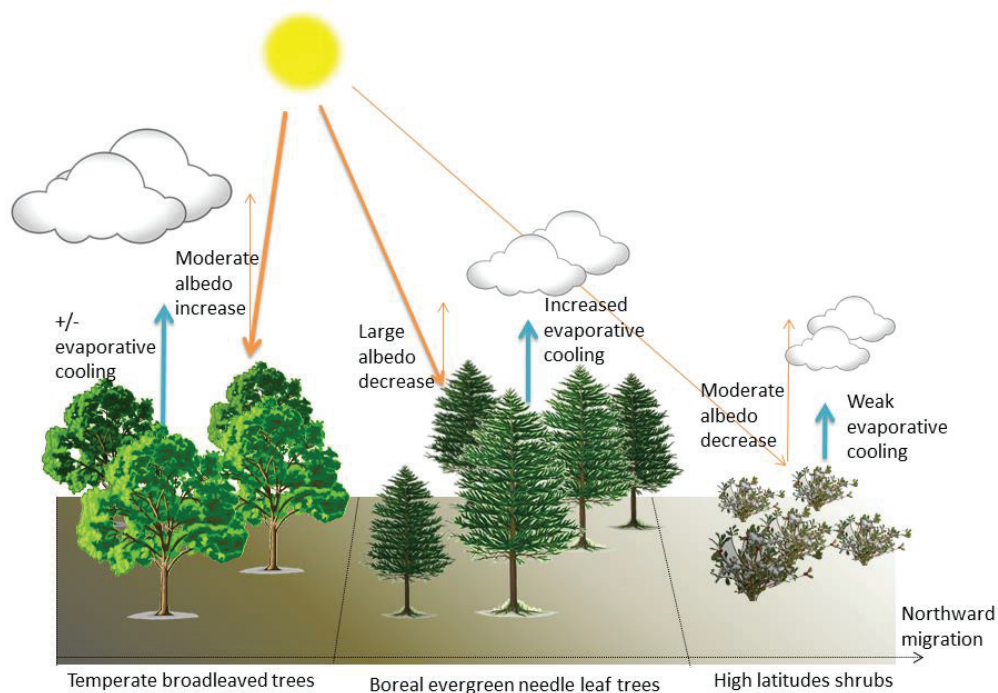


Figure 3: Simplified illustration of the relative importance of various feedback effects resulting from migration of ecosystems towards higher latitudes (or altitudes) in response to warming.

The biochemical feedback loops are mainly related to the carbon sequestration associated with more biomass, in addition to the enhanced permafrost thaw, leading to release of carbon and methane stored in the soil in permafrost regions. In high latitude ecosystems however, the biophysical properties are regarded the most influential with respect to global mean temperatures and local feedback effects (Bonan, 2008; Pearson et al., 2013).

2.4 Approaches and uncertainties in modelling vegetation change

The increased recognition of vegetation as an important factor in the climate system has resulted in development of more advanced treatments of vegetation in numerical weather and climate models. Improved representation of vegetation properties and dynamics, and corresponding influence on climate, is regarded one of the great improvements in the CMIP5 suite of earth system models as compared to its predecessor (IPCC, 2013, Chapter 9.4.4).

As highlighted by Mckenney et al. (2007), a major unknown in estimating the future feedbacks of vegetation re-distribution, is the extent to which species are able to disperse into their new suitable habitats, e.g. to follow their “climate envelope”. There are large differences in migration rates across ecosystems, and additionally, migration rates are influenced by disturbances and geographical features. For example, Myers-Smith et al. (2011) presented a synthesis of findings relating the sensitivity of high latitude shrub expansion to various environmental factors. While warming and lengthening of the snow free season promotes greening, they highlight the more complex influence of altered disturbance regimes resulting from herbivory, permafrost thaw, tundra fires and anthropogenic activity. This complexity adds to the uncertainties and spread in modelled estimates of future high latitude vegetation distribution.

In this section, a brief presentation of some methods and tools used in modelling vegetation changes are presented. Without going into detail about the complex and varied representation of vegetation and vegetation dynamics that is included in current earth system and vegetation models, some general methods in representing changes to the terrestrial vegetation in modelling are presented, along with some current uncertainties within modelling of vegetation-climate feedbacks that are of particular relevance to this thesis. Also, some perspectives of future challenges within this research field are highlighted. Special emphasis is on high latitude vegetation changes, as this is of particular interest in the thesis.

In order for models to accurately quantify the feedbacks to climate resulting from alterations in terrestrial ecosystems, all aspects of such changes should be accounted for; re-distribution of plant species, alterations in ecosystem composition and diversity, alterations in canopy structure, disturbances (by e.g. herbivores, fires or insects), allocation and usage of carbon, nitrogen and other nutrients, and corresponding changes in water use efficiency. The perfect vegetation model would account for all of these aspects of vegetation changes. However; until such a model comes into existence, we make use of a variety of modeling tools which each accounts for some fraction of the full picture; so that put together, science is gradually reaching new levels of understanding of the complex mechanisms that comprise biosphere-atmosphere feedbacks.

Vegetation models have often been separated into two main (yet often overlapping) groups; statistical models and process based models. The first group is based on establishing empirically based relationships between different climatic and/or environmental factors and species or

ecosystems. The second aims to explicitly resolve the various physiological, biophysical and biochemical processes and interactions that influence the distribution, growth and interactions of separate species or ecosystems.

Bonan (2016) provides a historical overview of the main features in the development of the process based models employed as part of present day earth system models; from the first generation models developed in the 1960s, which used bulk transfer equations and prescribed descriptions of surface albedo, roughness and soil water without describing the influence of vegetation at all, to today's third generation models. In addition to being able to account for the hydrological cycle and the explicit effects of vegetation and soil on the energy and water fluxes, third generation models are characterized by the direct link between photosynthesis and stomatal conductance, and were developed with a main motivation related to the investigation of the climate effect of reduced stomatal conductance in response to increased atmospheric CO₂.

The development and application of increasingly complex products from satellite data has also contributed to an increasing level of accuracy in vegetation models over the past decades. The fraction of Absorbed Photosynthetically Active Radiation (fPAR) is one much used satellite product, which is the ratio of vegetation-absorbed to incident PAR. It is derived from spectral reflectance data as a measure of vegetation greenness (Prentice et al., 2007). Seasonal cycles from such data are often used to either validate or prescribe vegetation phenology, and applied in constructing maps of vegetation composition by defining broad groups of plant functional types (PFTs). In addition, datasets with seasonally varying LAI and vegetated fraction are employed in models. However; as pointed out by Bonan (2016), the distinction between the two is not very clear, and the concept of fractional vegetation cover is not always consistent between models and satellite products.

In vegetation models, the different vegetation types are commonly grouped together according to their characteristics and biophysical properties in relation to atmospheric interactions, in what is referred to as PFTs. Each PFT is assigned with parameters and functions describing their biogeophysical properties and other aspects of their interaction with the environment. The distributions of vegetation types in the model domain may be prescribed according to aforementioned satellite based datasets and remain the same throughout the simulation. Alternatively, the distribution may be updated throughout the simulation in response to the environmental

forcing. This is the case in dynamic vegetation models (DVM). The redistribution is then based on climatic conditions, carbon concentration and available nutrients to the simulated ecosystem, in addition to elaborate routines describing plant specific establishment, competition, disturbances and mortality. The vegetation dynamics and interaction with the atmosphere is described with an ever increasingly level of detail, and varies greatly among models. This contributes to increase the spread among climate model projections especially when applied to future scenarios (IPCC, 2013, Chapter 9.4.4; Friedlingstein et al., 2013), and particularly in high latitudes. This is in part owing to the inter-model differences related to nutrient availability (IPCC, 2013, Chapter 9.4.4) and varied representation of snow-vegetation interactions and effects on albedo (Lorantý et al., 2014). These demonstrated impacts on the climate response in high latitudes predicted by DVMs highlight the importance of also considering other approaches to estimate the atmospheric effects of potential shifts in the Arctic and boreal vegetation distribution.

Manual re-distribution of vegetation cover in model simulations has often been applied in studies where the atmospheric effects of particular vegetation changes were of main interest. Manual re-distribution of vegetation properties may be applied to general vegetation features such as the greenness or vegetation density in an area, or to specific vegetation types, typically represented by dominant PFTs. The re-distribution is based on certain criteria reflecting the aim of the study, and is often linked to specific environmental or climate characteristics derived from observations or empirical data (as in the case of Paper III in this thesis). Alternatively, alterations are applied to represent observed trends in vegetation in an area, such as migration of boreal forests, and shrub expansion in Arctic tundra (Bonfils et al., 2012; Swann et al., 2010; Lawrence and Swenson, 2011; Snyder and Liess, 2014; Liess et al., 2012, and as illustrated in Paper II). A similar approach was taken when historical, prescribed land use change and deforestation patterns were applied in several CMIP5 climate simulations. This is regarded a great innovation in ESMs since the predecessor CMIP, despite introducing an additional spread in climate model projections, which is partly caused by counteracting effects of evapotranspiration and albedo changes (IPCC, 2013, Chapter 9.4.4).

Based on historical and/or current observations, simplified future projections for an area may also be derived. In this thesis, Papers II and III also follow such an approach, by applying

simplified perturbations to the current boreal and Arctic vegetation cover, represented by dominant PFTs, in order to investigate the atmospheric effects resulting from specific changes in the high latitude ecosystem. A more accurate description of the methodology and parameterization of vegetation properties in the model applied in the work included here is given in Chapter 3.2, and in the model and methodology sections of Papers II and III.

An “intermediate” approach has sometimes been applied, with the aim to apply continuously updated environmental drivers for vegetation change, without running a dynamical model. Alo and Wang (2010) rather applied a cyclic update of the vegetation cover based on a year-to-year climate development in response to annually updated vegetation distribution.

There are, however, still considerable uncertainties associated with modelling vegetation and changes to vegetation cover that are common to all types of the mentioned models and methods. The representation of vegetation types by use of PFTs is by far the most dominant method across vegetation modelling tools. Wullschleger et al. (2014) argue that there is mounting evidence that uncertainties in vegetation modelling in part arises from the incorrect or incomplete classifications of PFTs and their parameterization, particularly at high latitudes. Prentice et al. (2007) highlighted that there is no international standard or consistent, global high resolution vegetation map available to use for validation across vegetation models. They also argue that current PFT representation is simplistic and parameter values are neither agreed on, or are particularly well founded. To our knowledge, various satellite-based global and regional vegetation maps still have considerable differences among them. Procedures applied in generating such maps vary among products, and Wullschleger et al. (2014) also highlight that the number of representative PFTs differs among models and land use datasets. They particularly recommend four major aspects of future development regarding the use of PFTs in climate modelling; a review of the PFT classification in relation to integrating remote sensing products and validation, amended inclusion of appropriate PFTs in regional models, database compilation with emphasis on trait identification with regard to both above and below ground properties, and data development for cross-model validation. Several have argued that as the definition of PFTs is based on a discretization of what is a continuous variation in plant traits defining their interactions among each other and with climate, new techniques in describing plant variations in modelling is required (Verheijen et al., 2013; Wang et al., 2012; Wullschleger et al., 2014).

Uncertainties in the results arising from parameter values related to PFTs are also discussed in Papers I-III in this thesis.

3 Research tools and methodology

The main research tool applied in the studies presented in this thesis, are numerical models. A regional weather or climate model is used to dynamically downscale results from a global weather or climate model, for the purpose of providing results on finer spatial resolutions. This has advantages when regional and local scale processes are important to resolve (Rummukainen, 2010; Giorgi and Mearns, 1991).

Although important on a global scale, all research questions addressed here are related to local and regional scale processes, and for that reason we chose to apply regional models to resolve them. Simulating land surface processes are dependent on the highly heterogeneous features of the land surface, such as topography, land-water boundaries and vegetation cover which is the main focus here. Simulations on so-called convection-permitting resolutions are beneficial for processes such as soil-moisture-precipitation feedbacks and local wind patterns (Rummukainen et al., 2015). Particularly simulations conducted for Papers I and III in this thesis were important to run on fine, convection permitting spatial scales, and required running a non-hydrostatic model. For the purpose of Paper I, chemical processes were also required on a fine temporal and spatial scale.

The measurements used in Paper I were gathered by co-authors in that paper, Giacomo Gerosa and Angelo Finco and are described in Paper I. For the chemical initial and boundary conditions, data from the Oslo CTM3 model was used, which was run by co-author Øivind Hodnebrog.

3.1 Atmospheric model

For the purpose of all three studies, we applied the WRF modeling system (Skamarock et al., 2008). WRF is a community model, with users registered around the world. The code is maintained at the Mesoscale and Microscale Meteorology (MMM) Division at the National Center for Atmospheric Research (NCAR), who also provide support for the user community.

WRF is a widely used and highly adjustable state-of-the-art numerical weather prediction system with a wide variety of applications on scales ranging from local regions with a resolution of a few hundred meters, to global simulations with much coarser resolution. It can be run in a non-hydrostatic or hydrostatic mode, and with a wide range of physical parameterization schemes.

The setup may as such be adjusted to simulate case-specific short-term local weather events as in Paper I, or decadal long climate simulations, as in Paper II.

The setup in Paper I was based on review of literature and NCAR guidelines for chemical simulations, in addition to various sensitivity tests (not shown). Particularly, we have tested the effect of spatial and vertical resolution, domain sizes and chemical boundary conditions with focus on simulated ozone concentrations, anthropogenic and biogenic emissions related to the diurnal cycle of near surface ozone concentrations, and nudging parameters and PBL schemes with the aim to accurately resolve the land sea-breeze.

The applied setup in Papers II and III was based partly on the Polar WRF setup and validation studies (however not equal to), and partly on the NCAR recommendations of physical parameterizations for cold climates (shorter and longer simulations). The Polar WRF version has been evaluated in a series of experiments (Hines and Bromwich, 2008; Hines et al., 2011; Wilson et al., 2011, 2012). Main findings of a benchmark study of the forecast hydrological cycle of Polar WRF, a one year long forecast covering the entire Arctic region, include a tendency of excessive summer precipitation linked to overestimated mid-latitude annual precipitation, and excessively warm and moist atmospheric boundary layer (Wilson et al., 2012). In addition, too low cloud fraction, underestimated downwelling LW, and overestimated SW radiation were found. These findings are invoked to explain a slight warm bias in the 2 m temperature north of 60 °N (Wilson et al., 2011; Hines et al., 2011). In these studies however, the biases are judged sufficiently small to conclude that the modelling system accurately predicts the annual cycle of the 2 m temperature. The findings of excessively warm and moist boundary layer in WRF simulations were supported by findings by Mölders and Kramm (2010), studying a short-time winter inversion event using WRF in two different setups, and they highlight the strong impact of the choice of physical parameterizations on the results. Although according to findings by Jin et al. (2010), excess precipitation in the WRF model seemed detached from land surface processes, and remained overestimated across four different land surface schemes.

Although domains and model versions vary between our experiments and the ones analyzed in these studies, the findings of these validation studies are kept in mind while interpreting the results in our papers.

For atmospheric initial and boundary conditions, we used the ERA Interim 6-hour reanalysis for all three studies. A brief presentation of the most important parts of the model in relation to the work presented here is described below.

3.2 Land Surface model

For all three studies, we have run the atmospheric model coupled to the Noah Land Surface Model (LSM) (Tewari et al., 2004). It is a widely used LSM currently coupled to several regional and global modelling systems. The land surface model computes the surface fluxes of energy and water, and controls the energy and water budget of the surface and soils. The parameterization of vegetation plays a key part in these processes, as illustrated in Fig. 4.

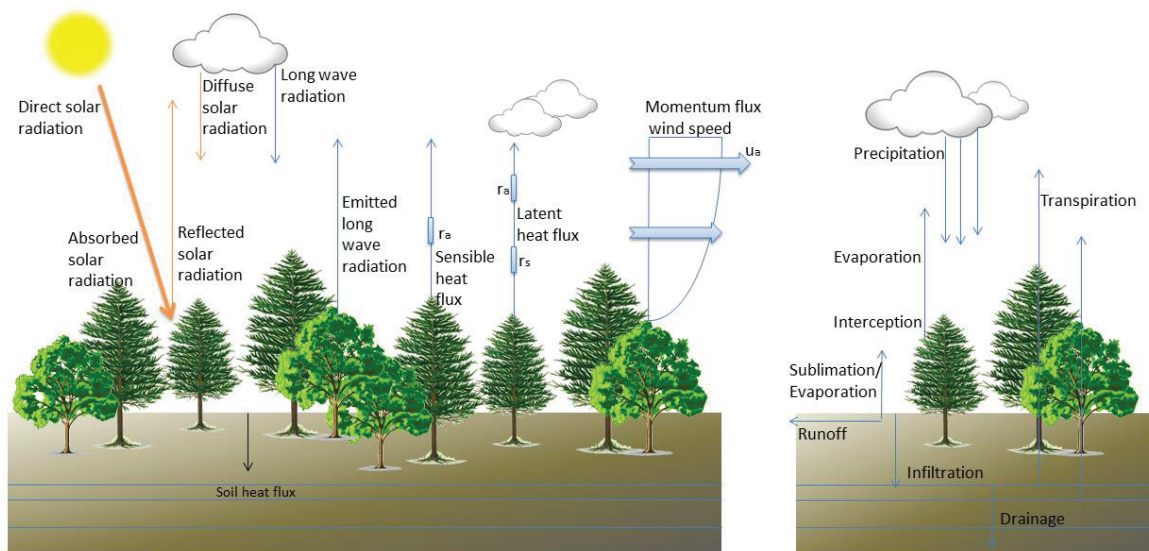


Figure 4: Surface processes handled by the land surface model. Surface energy budget (left) and surface water budget (right).

In our simulations, the soil is divided into four layers of varying thickness, summing up to a total of 2 m. The top surface layer is a combined soil, vegetation and snow layer, which properties are dependent on soil and vegetation category. The partitioning of energy between sensible and latent heat fluxes is related to the vegetation dependent exchange coefficients for heat and moisture. Besides meteorological conditions such as wind speed and atmospheric stability, they

are dependent on the vegetation's stomatal resistance, greenness fraction, roughness length, rooting depth and LAI.

3.2.1 Noah-UA

For the purpose of addressing the research questions in Paper III, the model was run with the Noah-UA land surface model, which is the Noah LSM, with parameterization for snow-vegetation interactions added by Wang et al. (2010) at the University of Arizona. The added parameterizations include consideration of the vegetation shading effect on snow sublimation and snowmelt, under-canopy resistance, improvements to the ground heat flux computation when snow is deep, and revision of the momentum roughness length computation when snow is present.

3.2.2 Treatment of vegetation in Noah

In the Noah LSM, the vegetation cover is represented by plant functional types (PFTs), which are interpolated to the simulation domain from satellite based datasets. There are several such datasets with varying resolution and vegetation categories provided with the standard model package download. In the simulations, the dominant PFT for each grid cell is used for calculating interactions with the atmosphere. The dominant vegetation category in a given grid cell determines a range of biophysical parameters related to its interaction with the atmosphere. These parameters include the height and density of the canopy, the number of soil layers available to the plants' roots, minimum canopy resistance, snow depth water equivalent required for total snow cover, and ranges for maximum and minimum values for LAI, emissivity and surface roughness length. The surface albedo and green vegetation fraction is either prescribed per vegetation category, or interpolated from monthly datasets.

In the model the dominant categories in each grid cell are fixed throughout the simulation, as there is no parameterization for competition, survival and establishment.

3.2.3 Land cover data

For land use data in Paper II, the MODIS IGBP modified 21 class land surface data was used. This dataset is available with the standard WRF package download. The dataset is based on the original 1 km resolution MODIS IGBP vegetation map (Friedl et al., 2010), but excludes permanent wetland. It has three tundra classes and lakes added by the Land Team at EMC/NCEP. To represent high latitude and altitude ecosystems more accurately, the vegetation category of

open shrub land was replaced by various tundra vegetation classes north of 60 degrees latitude in the modified MODIS dataset.

For the land cover in Paper III, we used the newly available 20 class MODIS 15 sec resolution dataset (Broxton et al., 2014). In this dataset, most of the shrub and tundra covered part of our study area is covered by the dominant vegetation category of “open shrubland”, consisting of low shrubs of <0.5m height. As we would like to also study the effects of shrub height on the atmospheric response, we used this category as a basis to make adjustments in order to distinguish shrubs of various heights, in accordance with their temperature based potential habitats. More about the details of this approach is found in Paper III, Section 2.

3.3 Chemistry model

In WRF-Chem (Grell et al., 2005), a chemistry module is completely embedded in WRF, allowing it to simulate the coupling between chemistry and meteorology. The chemistry package consists of the following main components; a dry deposition scheme, anthropogenic emissions, biogenic emission, gas-phase chemical mechanisms, photolysis schemes and aerosol schemes. Tracer transport is computed by the meteorological part of the model. It is employed for e.g. forecasting for field campaigns, testing in relation to air pollution abatement strategies, and assimilation of satellite and in-situ chemical measurements. For our purpose the WRF-Chem package is used in Paper I to estimate the ozone distribution and deposition in order to compare estimated stomatal fluxes into vegetation with measurements gathered at various field campaigns in different Mediterranean ecosystems.

3.3.1 Chemical input data

The chemical initial and boundary conditions are gathered from the chemistry transport model OsloCTM3 (Søvde et al., 2012). For the anthropogenic emissions we use the TNO MACC II gridded anthropogenic emission database (Kuenen et al., 2014), which is a gridded emission inventory covering UNECE-Europe for the years 2003-2009. It contains European emissions of air pollutions (CH₄, CO, NH₃, NMVOC, NO_x, PM₁₀, PM_{2.5} and SO₂). It is furthermore divided per source sector, and the emissions from area sources have been distributed in a sector-specific way, while the point source emissions keep their particular coordinates (Denier van der Gon et al., 2010a; Denier van der Gon et al., 2010b). Hodnebrog et al. (2011) demonstrated that the resolution of the emission inventory greatly influences accuracy in the modelling of ozone

distribution. Based on these findings, and for the purpose of this study, we were given access to the TNO MACC II high resolution dataset (HAC Denier van der Gon, pers. comm.), which covers Europe with a resolution of $\sim 7 \text{ km} \times 7 \text{ km}$. Examples of the NO emissions as re-gridded to the inner domains are shown in Fig. 5.

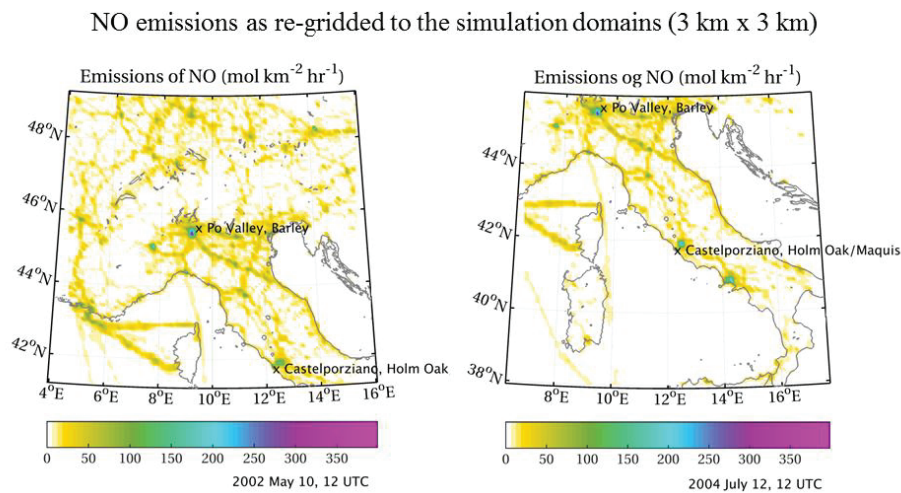


Figure 5. NO emissions at two points in time as re-gridded from the TNO MACC II emissions inventory resolution to the WRF-Chem inner domains simulated in Paper I, with resolution 3km x3km. The sites of measurements are indicated.

3.3.2 Calculation of ozone fluxes using WRF-Chem

The dry deposition of gaseous species and aerosols in WRF-Chem is based on the work of Wesely (1989). The Wesely scheme uses the “big-leaf” approach, where the surface vegetation is treated as one big canopy. To classify the plant properties, the dry deposition scheme uses the same vegetation categories as the rest of the model. In the dry deposition scheme these categories are grouped together to 11 categories, by giving them the same value for parameters that are only land use and season dependent, using five seasonal categories that distinguish the main stages of plant phenology through the year.

The deposition velocity is calculated utilizing a resistance analogy. The deposition velocity reflects the surface conductance, and is given as the reciprocal of the overall surface resistance. The total surface deposition velocity v_d is given as the sum of the reciprocal of three resistances placed in series, and can be expressed as

$$v_d \equiv (r_a + r_b + r_c)^{-1} \quad (3.1)$$

where r_a represents the aerodynamic resistance of the turbulent air between a specified height above the canopy and the surface. r_b gives the resistance of the quasi-laminar sub layer adjacent to the deposition surface, and r_c represents the bulk surface resistance. The surface resistance represents the plant canopy, twigs and stem, and the soil. It is computed as the sum of seven minor resistances, placed in four parallel pathways to account for the various processes controlling the flux of gas particles deposited at the surface. This resistance network introduced by Wesely (1989) is illustrated in Fig. 6.

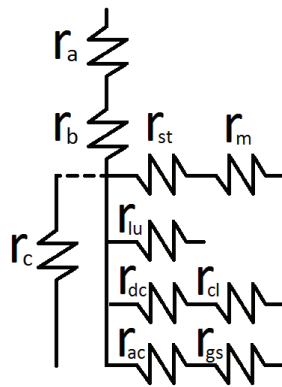


Figure 6. Schematic diagram of the pathway resistances used in the Wesely module. r_a is the aerodynamic resistance, r_b is the quasi-laminar sublayer resistance, r_c is the bulk surface resistance. The latter consists of the stomatal resistance (r_{st}), mesophyll resistance (r_m), cuticular resistance of the upper canopy (r_{lu}), cuticular resistance of lower canopy (r_{cl}), resistance of buoyant convection (r_{dc}), resistance of gas transfer through the canopy (r_{ac}) and the soil and ground resistance (r_{gs}). (Modified from Wesely (1989))

The main entry of gaseous species into the leaf is through the stomata, which is parameterized as the pathway regulated by the stomatal resistance and the mesophyll resistance (marked as r_{st} and r_m in Fig. 4). In the Wesely expression for stomatal resistance the surface temperature and the solar radiation is accounted for by the following expression:

$$r_{st} = r_i(1 + (200(G + 0.1)^{-1})^2) \left(400(T_s(40 - T_s))^{-1} \right) \quad (3.2)$$

where G is the solar irradiation in $W m^{-2}$, and T_s is the surface air temperature between 0 and 40°C. Outside this temperature range the resistance is set to a large value, assuming that the stomatal transfer has stopped. The parameter r_i represent the minimum stomatal resistance for water vapor and is vegetation type and season dependent. After passing through the stomata, the

gas encounters the mesophyll resistance, which for simplicity is set to zero for ozone, assuming a near instant oxidation of sub-stomatal mesophyll once it has entered the vegetation. The combined mesophyll and stomatal resistance for gaseous species is calculated by scaling the stomatal resistance by the ratio of molecular diffusivity of water vapor, to that of the gas of interest.

The non-stomatal pathways are grouped together according to the resistance network depicted in Fig. 3. The cuticular resistances are only dependent on vegetation type and season, however they are increased whenever the canopy is wet due to rain or dew. The buoyancy resistance is dependent on the short wave radiation and solar zenith angle, giving it a diurnal variation. The soil and in-canopy air transfer resistances are solely dependent on the Wesely vegetation type and season. A comprehensive description on the parameterization of the single resistances can be found in Wesely (1989).

In order to estimate the flux of ozone into the vegetation, the bottom layer ozone concentration is scaled according to the ratio of the canopy height deposition velocity to that of the grid cell average velocity, in order to get the canopy height concentration. This concentration is then multiplied with the combined stomatal and mesophyll conductance, given as the reciprocal of the resistance, to get the flux:

$$F_{st} = g_{st} * C_c \quad (3.3)$$

where F_{st} is the stomatal flux of ozone, g_{st} is the stomatal conductance, and C_c is the canopy-height ozone concentration, and assuming that the ozone concentration inside the leaf is zero. These fluxes are used for comparison to measured fluxes in Paper I of this thesis.

4 Presentation of findings

In this section, the research corresponding to each of the scientific papers included in thesis is presented in relation to the research objectives as defined in Section 1.1. For each study, a brief explanation of why and how our research was conducted is presented, along with a discussion aimed to relate our findings to similar studies in the field. More details of the methods and results are described in each of the scientific papers.

4.1 Evaluation of ozone fluxes from the Wesely scheme in WRF-Chem

The first research objective is the focus of Paper I. This study investigates the skill of the widely used community model WRF-Chem equipped with the Wesely deposition scheme in estimating the flux of ozone into vegetation. This is a highly adjustable model which we believe could prove to be a versatile and powerful tool in estimating ozone damage to vegetation in both present and future climates.

However; some weaknesses regarding the surface conductance parameterization in the Wesely dry deposition scheme have been found in previous studies that were aimed at evaluating the dry deposition fluxes produced by this scheme. Limitations related particularly to the surface resistance have been highlighted. More specifically, the scheme lacks an explicit account for the plants' water stress through parameterizations of effects of vapor pressure deficit or soil moisture deficit (Hardacre et al., 2015; Fowler et al., 2009). This relates particularly to the stomatal resistance parameterization, which compared to other parameterizations, as e.g. in the DO₃SE model (Emberson et al., 2000), is simplified and only dependent on the ambient temperature and available sunlight, in addition to vegetation type and season (as shown in Eq. 3.2).

Building on the findings by previous validation studies, we aimed to give an indication of the ability of this parameterization to estimate stomatal ozone fluxes and potentially harmful ozone doses. Furthermore, emphasis was placed on how the known weaknesses in the current parameterization might influence these results. By comparing estimated fluxes for three different vegetation types to flux measurements gathered from field campaigns in three different typical Mediterranean ecosystems; a holm oak forest, a maquis ecosystem and a barley field, under varying meteorological conditions, this study identifies strengths and weaknesses of the current parameterization, with particular emphasis on the influence of water availability.

Our results show that the WRF-Chem system generally underestimated both the ozone mixing ratios (by 5%) and temperatures (by 2%) as averaged across all measurement periods, and overestimated the stomatal fluxes by 58% as compared to the measurements (Fig. 7)

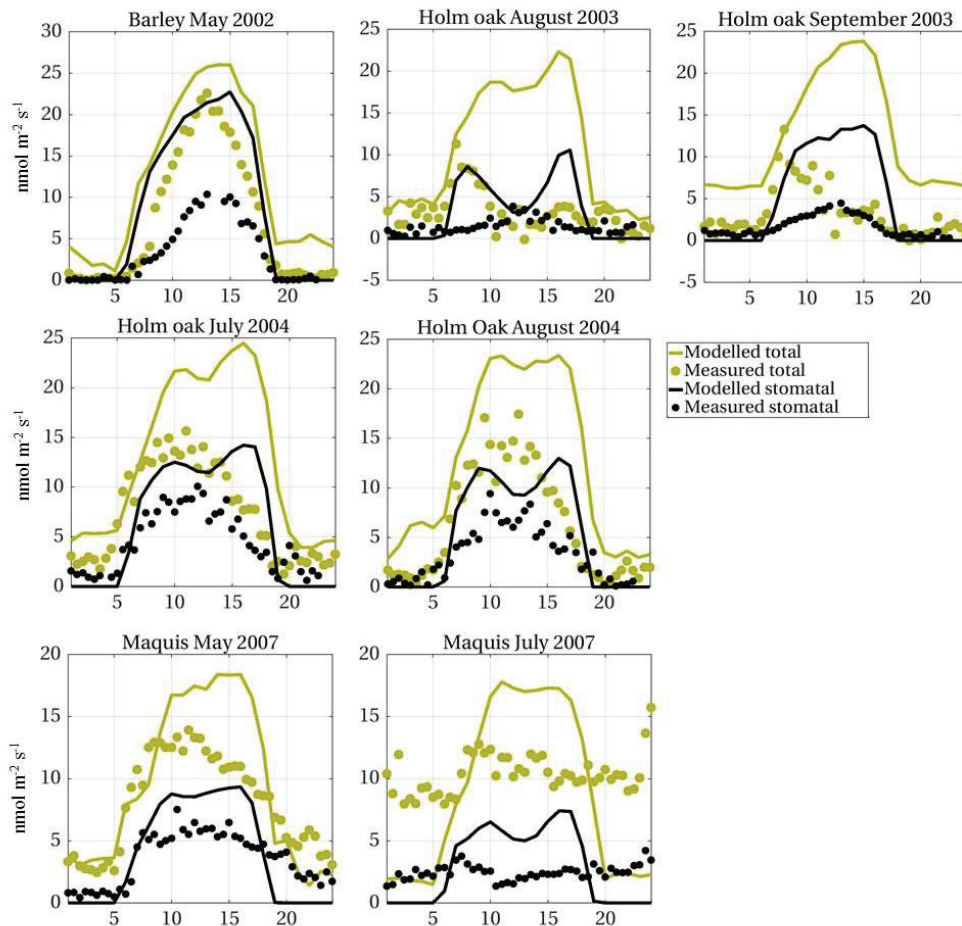


Figure 7. Mean diurnal modelled (lines) and measured (dots) stomatal (black) and total (green) surface fluxes for each measurement period. Figure from Rydsaa et al.,(2016)

The reasons for modelled biases of stomatal fluxes varied somewhat between vegetation types (Objective 1b). For the holm oak and maquis ecosystems, there was a clear mismatch in diurnal pattern for the stomatal conductance and stomatal flux for all measurement periods, which was caused by the temperature response function within the Wesely scheme. This function assumes an optimal temperature for stomatal opening at about 20 °C, which seems low for these particular types of vegetation. In contrast, the optimal temperature for Mediterranean woody species as

suggested in CLRTAP (2015) is 23°C, which would decrease the observed bias produced by the Wesely scheme, in which the same optimal temperature is used for all vegetation types.

For the barley field, the overall daytime stomatal conductance level was too high, causing a severely overestimated stomatal flux of ozone. The results point to the Wesely scheme's value for the minimum resistance. Based on these results, the value representing the category of "Agricultural land" is likely too low in the case of this particular crop. Again comparing to the CLRTAP (2015) suggestion for the generic crop value for Mediterranean areas, the Wesely scheme's value is significantly lower. Our findings support those of Wu et al. (2011), who also concluded that large uncertainties were associated with the minimum stomatal resistance values in the Wesely scheme in WRF-Chem.

The Wesely scheme's parameterization for stomatal conductance imposes a requirement of daylight on stomatal opening (Eq. 3.2). However; unlike the modelled estimates, the measurements do indicate that nighttime values for stomatal conductance are not zero. As such, the low modelled nighttime stomatal flux fails to reproduce particularly the woody vegetation's ability to open their stomata during nighttime. This supports results by Mereu et al. (2009), who found high nighttime stomatal uptake of ozone in Mediterranean woody species, especially in drought situations. Caird et al. (2007) suggests that the parameterization of zero nighttime stomatal conductance is inaccurate for most plants.

With regard to Objective 1c, this study confirms that the previously known weaknesses of the surface resistance parameterization, particularly with respect to the lack of explicitly accounting for the plant's water availability, result in overestimated daytime ozone stomatal fluxes particularly in warm and dry periods. Although such dependencies are included in most stomatal conductance parametrizations (Damour et al., 2010), the Wesely scheme only takes into account the water status of the plants in an indirect way, through the dependence on season and vegetation type (Wesely, 1989). For the purpose of estimating absorbed and potentially damaging doses of ozone to vegetation, such dependencies are of particular importance. In drought and water stress situations, while the vegetation closes the stomata to prevent water loss, it prevents ozone from entering the plant and thereby lowers the risk of ozone-induced damage. For this reason, estimating potential risk of ozone induced damage to vegetation in (semi) arid

areas such as in Mediterranean ecosystems, is particularly sensitive to such parameterizations (e.g. CLRTAP, 2015;Gerosa et al., 2004;Gerosa et al., 2009a).

To illustrate the effect of a simple improvement of the model performance, we tested including critical VPD as suggested in Pleijel et al. (2007). This acts to remove the afternoon peak values in estimated stomatal fluxes, and the overall mean fluxes, and bias for the period is reduced.

Based on results presented in Paper I, we conclude that the limited environmental conditions accounted for in the Wesely conductance calculations do cause specific and substantial biases compared to measurements, particularly in drought periods. The dependence on daylight for stomatal openings seems inaccurate and causes systematic biases for the vegetation types investigated in Paper I. For the purpose of estimating ozone doses in vegetation, the stomatal resistance parameterization should be improved by extending it to include an explicit and time varying dependence on the water availability of the vegetation in question, in addition to a revision of the temperature response function and night time stomatal conductance in woody vegetation (Objective 1d).

The Wesely deposition scheme is a widely known parameterization which is employed in a variety of chemical transport models, and as such a well validated scheme. To our knowledge however, Paper I is the first study to explicitly test the stomatal component of the estimated surface fluxes resulting from this scheme. Our findings could form the basis of further development efforts that might lead to the employment of this modelling tool in future estimation of ozone induced damage to vegetation.

4.2 Sensitivity of the high latitudes atmosphere to biophysical effects of vegetation changes

The research questions in Objective 2 were investigated in Papers II and III. More specifically; Paper II focuses on the biophysical effects on the atmosphere resulting from three specific vegetation shifts in the boreal vegetation, and is mainly aimed at Objective 2 a-d, whereas Paper III focuses on the effects of increased shrub and tree cover specifically, corresponding to research Objectives 2 e-g. Following is a short presentation of each study, along with a discussion placing our findings in context with related research.

The sensitivity of the atmosphere to the biophysical effects of northward migrating boreal tree lines (Objective 2a) was part of the focus of Paper II. For this purpose we chose to aim our investigation at the boreal European vegetation cover. As mentioned in Chapter 2, a trend of northward migration of the boreal European forest has been confirmed in observational studies (e.g. Soja et al., 2007; Hofgaard et al., 2013). For this study, we applied WRF V3.5.1, coupled to the Noah land surface model. For the vegetation cover, the MODIS modified IGBP vegetation data set was chosen. The atmospheric response was investigated over a 10 year simulation period. The 10 year period was chosen as a compromise between sufficient length to represent a robust climate signal and computational cost. More details about the experimental setup and the simulations can be found in Section 2 of Paper II.

To limit uncertainties related to the dynamical response of the biosphere to climate changes, several previous studies have applied manual rather than dynamical vegetation changes in studies where the main objective was to study the resulting effects on the atmosphere. To reduce the complexity related to the ecosystem migration in our investigation, we chose a similar approach, by applying simplified perturbations to the vegetation cover based on observed and anticipated trends in forest migration in the area. The aim was to represent dominating vegetation shifts while keeping simplicity in the setup, enabling us to investigate the atmospheric response corresponding to each specific shift in vegetation type. We aimed to apply changes that were moderate enough to be achieved within a 100 year timespan, assuming a mean migration rate based on cross-species observational studies, resulting in a shift of up to 108 km in the northward direction. The perturbations to the vegetation cover applied in Paper II are illustrated in Fig. 8.

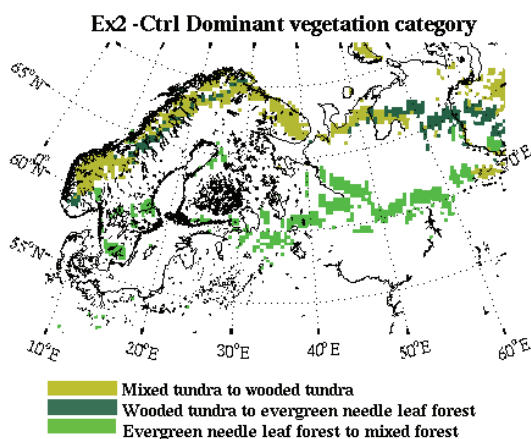


Figure 8: Changes to the MODIS IGBP land cover applied in experiments, as compared to the control simulation. The colors indicate areas where the tree lines have been moved northward, and where shrub cover has been expanded. (Figure from Rydsaa et al., 2015)

To achieve a simplified, yet realistic experimental setup, we further limited the vegetation shifts to only apply to neighboring grid cells covered by certain dominating vegetation; the evergreen forest was only allowed to “migrate” into grid cells covered by wooded tundra, which represents tundra already partly covered by deciduous trees and shrubs. Along the southern edges of the boreal forest, the mixed forest, comprising both evergreen and deciduous trees, was allowed to “take over” for pure evergreen needle leaved forest. The category of mixed tundra was converted to wooded tundra, representing a slight increase in shrub and tree cover. Although simplified, the vegetation shifts in these experiments represented more moderate and perhaps more realistic perturbations given the 100 year timeframe, as compared to other similar studies as in the following examples.

Other studies applying dynamical vegetation shifts resulting from future climate projections, like a doubling of atmospheric CO₂, have resulted in enhanced greening and vegetation shifts on a substantially larger scale as compared to our study (e.g. Jeong et al., 2011; Jeong et al., 2014). Using the CCSM3 model with a DVGM on a resolution of 2.8° x 2.8°, their vegetation changes included higher LAI across all six PFT categories covering the northern high latitudes, more vegetation in the permafrost region and northward migration of the boreal forest. They do however highlight uncertainties related to these vegetation shifts, particularly related to known model biases.

With a motivation to amend for such uncertainties inherent in DGVMs, Snyder and Liess (2014) based their vegetation changes on an assumed theoretical timeframe of 100-500 years. They used the CAM3-CLM3 model and applied a 2.8° northward shift in the circumpolar boreal evergreen forest (represented by a one grid cell shift, and as such the smallest possible shift within that model setup). This led to the replacement of both grass, shrubs, and bare ground by evergreen forest in their simulations, with the aim to examine not only the regional, but also global atmospheric responses.

In a study more similar to ours, Liess et al. (2011), using the WRF model on a circumpolar domain covering the entire Arctic (on a 30 km x 30 km resolution), for the month of July, applied a shift in which vegetation in all grid cells covered by the PFT “open shrubland” was

replaced by evergreen (and deciduous) needle leaf forest in the circumpolar area north of 50 and 60°N (depending on the longitude). This however, represented a larger areal extent, and larger change in all biophysical properties (like LAI, albedo and roughness length), as compared to changes applied in our study. They also highlighted some unrealistic features of their vegetation shift, namely the lack of a simultaneous increase in shrub cover, and movement of the southern border of their boreal forest. Both of these features have been included in our simulations in Paper II.

In contrast to the above mentioned studies, which all were carried out on circumpolar large domains, our study in Paper II was conducted on a limited sub-region of the boreal and Arctic biome, on higher horizontal resolution. As the vegetation shifts in our study represents a smaller fraction of the total study area, and the vegetation types replacing each other in our experiments are more structurally similar (as reflected in their biophysical properties, see Table 2, Paper II), comparatively less drastic effects on the atmosphere following our perturbations may be expected. However, as highlighted by Beringer et al., (2005), their observations suggested substantial non-linearity in the atmospheric response to changes in the structural properties of vegetation, and they showed that moderate structural and biophysical changes may cause significant feedbacks to the atmosphere, and is therefore of interest for further investigation.

Emphasis in the investigation in Paper II was placed on the effects on land-atmosphere fluxes of heat and moisture (Objective 2a), along with other near-surface atmospheric variables. The northward migrating evergreen forest led to increased latent heat fluxes with an annual average of 4.4 W m^{-2} , and with a peak of 18 W m^{-2} in July (Fig. 10). The large increase in summer LH in

our

Annual average differences in surface variables as result of northward migration of boreal forest and shrub expansion, EX2-Ctrl

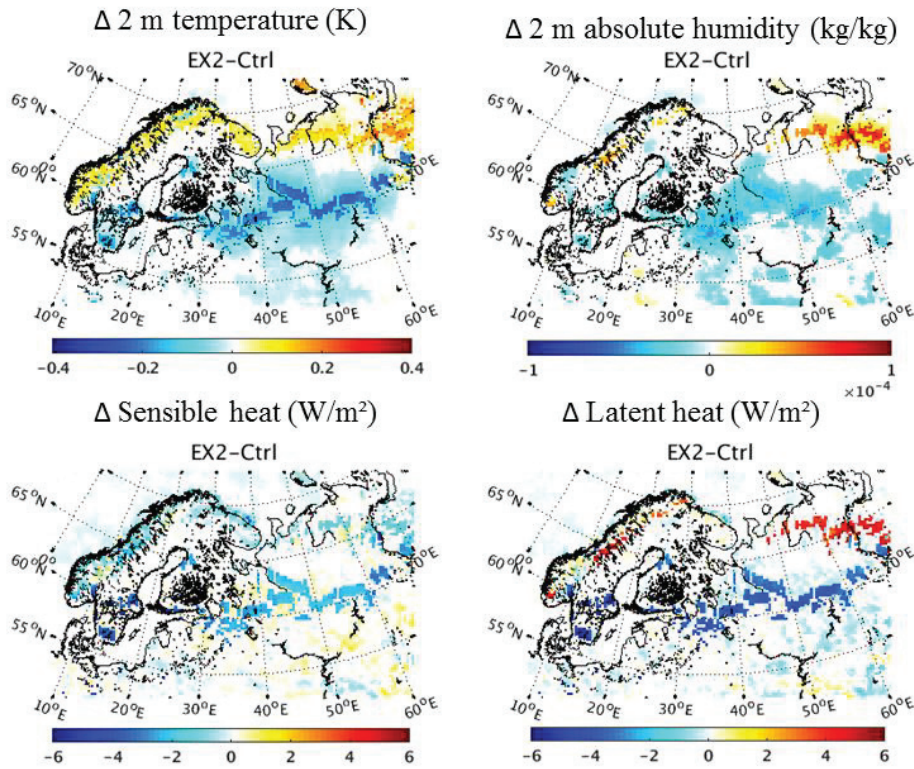


Figure 9: Result of northward migration of boreal forest and shrub expansion given as a 10 year average in 2 m temperature (upper left panel) and moisture (upper right panel), and surface fluxes of heat and moisture (lower left and right panel ,respectively) as compared to the control simulation. (Only showing significant results at the 95% confidence level). Figure modified from Rydsaa et al., (2015).

simulations is caused largely by the simulated increase in LAI (and corresponding increase in inception and canopy evaporation). This led to enhanced atmospheric water content in our simulations (Fig. 9, upper right panel), increased cloud cover (as evident from the radiation fluxes, Fig. 10) and increased summer precipitation (+ 3.4 %). The increased LW and lowered incoming SW radiation in the summer season for this vegetation change is shown in Fig. 10 (thin, stippled line).

The increase in LH, cloud cover and rainfall enhances the partitioning of increased absorbed radiation (due to decreased albedo) into latent rather than sensible heat flux, and thereby decreases heating of the overlying atmosphere. This results in a lowering of summer SH, with a mean reduction of the July sensible heat flux by 11.2 W m^{-2} in areas with northward expanding evergreen forest, despite the counteracting effect of a slight increase in surface roughness associated with this vegetation shift (which leads to an increase during the other seasons). In areas with shrub expansion (conversion from mixed tundra to wooded tundra), the largest monthly mean effect was seen in June, with an average reduction of sensible heat of 1.3 W m^{-2} .

This decrease in summer SH stands in contrast to results of other studies simulating enhanced vegetation cover at high latitudes (Liess et al., 2011; Snyder and Liess, 2014; Jeong et al., 2014). It is largely explained by the comparatively moderate vegetation shift and corresponding small effects on structural canopy properties that affect the SH flux, combined with the strong increase in LH, leading to an ET effect (here acting to reduce surface incoming SW radiation) dominating over the albedo effect (a positive feedback) during the summer months for this particular vegetation change.

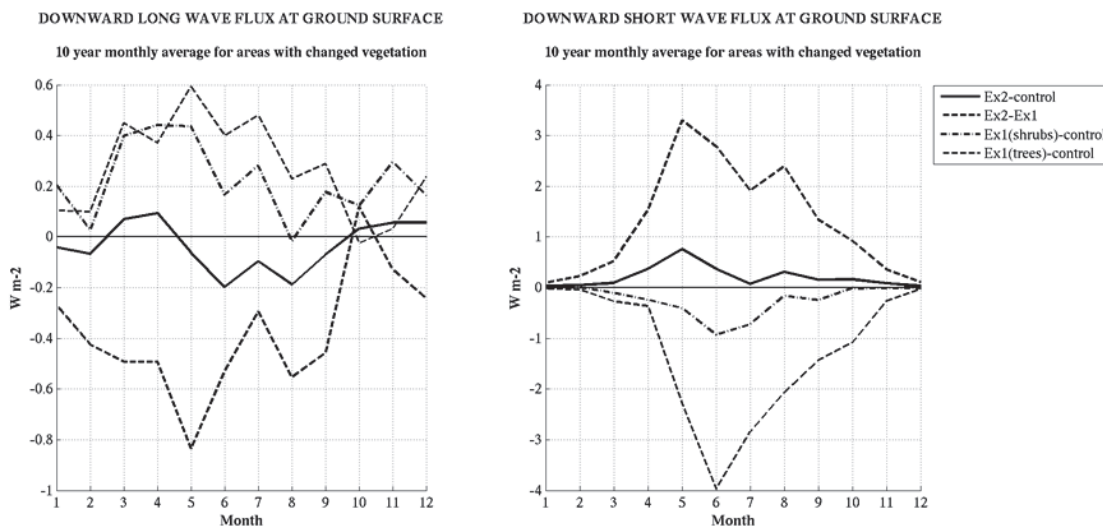


Figure 10. Monthly 10-year average changes in LW (left) and SW (right) downwelling in each area with changed vegetation. Average of all vegetation changes shown as the thick black line (Ex2-Control).

The flux repartitioning with reduction in summer SH in our simulations led to a modest annual 2 m temperature increase of 0.12 K in areas with evergreen forest expansion (Objective 2b). Our

simulations show a peak temperature increase of 0.45 K in autumn, when the albedo decrease was larger and sensible heat increase was largest. This annual mean and seasonal response was largely related to the applied vegetation changes, where evergreen trees were replacing wooded tundra. The largest albedo response was seen in autumn when the deciduous trees shed their leaves. This led to increases in absorbed SW at the surface and increased SH, warming the lower atmosphere.

The atmospheric response resulting from the separate vegetation changes along the northern and southern edges of the boreal forest largely pointed in different directions (Objective 2c); the high northward shrub and evergreen forest expansion lead to a decreasing albedo and enhanced near surface temperatures by 0.12 K, and a deeper and wetter planetary boundary layer. The replacement of evergreen forest by more broadleaved species along the southern border of the boreal forest led to lower LAI, higher albedo and lower surface fluxes. This resulted in decreased heating of the boundary layer and lower near surface temperatures and humidity. This vegetation shift as such led to an annually averaged negative feedback of about 0.2 K (Fig. 9). However, there were large seasonal differences in atmospheric response for the individual vegetation shifts.

These results support findings by Jeong et al. (2011), who found that vegetation changes resulting from a doubling of atmospheric CO₂ over the northern high latitudes led to an reinforced increase in daytime near surface temperatures in high latitudes (~0.9 K), but limited the initial increase in temperature in the mid-latitudes temperate forest regions by ~0.4 K.

Beringer et al. (2005) examined the potential influence of structural vegetation changes by measuring surface energy exchanges along tundra to forest transition zone in Alaska during the summer. Beringer et al. (2005) reported small differences in latent energy flux between the sites, and a decrease in ground evaporation was compensated by an increase in ET moving from tundra towards more complex canopies. Sensible heat flux was increased going from tundra to more complex canopy vegetation, with both shrubs and trees enhancing the sensible heat flux to the atmosphere.

Our summer increase in LH is slightly higher than the mean July value of 11.5 W m⁻² found in Liess et al. (2011), who similarly to us found that expanding the needle leaved boreal forest led to an increase in LH rather than SH (+8.3 W m⁻²). The larger increase in SH as compared to our

results relates to larger changes in surface roughness length following their vegetation perturbations, and resulted in an overall annual warming of 0.63 K. Similar to our study, they found that the increased ET and LH led to a more humid PBL and increased cloud cover, leading to reduced SW downward flux, and increased LW. They also found areas within their study domain where these effects dominated over the albedo decrease, and led to cooling rather than warming. Their study did not include an analysis of precipitation.

This partly contrasts findings in a follow-up study by Snyder and Liess (2014), who found a slight decrease in summer LH following a northward shift of boreal forest, and an annual increase of only 1 W m^{-2} . They explained their findings by the counteracting effects of increased ET ($+3.2 \text{ W m}^{-2}$) and decreasing ground evaporation (-2.2 W m^{-2}), and little or no increase in near surface atmospheric humidity or precipitation. Furthermore they found annual mean increases in sensible heat ($+2.7 \text{ W m}^{-2}$) as response to lowered albedo and increased net radiative forcing, leading to an annual warming of 0.3 K (JJA temperature increase of 0.4 K, and a peak in spring of 0.7 K). They found little or no increase in near surface humidity and precipitation, and a decrease in low cloud cover. They emphasize the role of coarse resolution in their simulations and argue that compared to the study by Liess and Snyder (2011), less fine scale surface processes can be captured on such a coarse resolution. They also highlight that the static representation of vegetation represents a limitation as it does not capture important processes related to vegetation dynamics.

Jeong et al. (2014) found an increase in summer LH of 4.35 W m^{-2} . They also found an increase in summer SH ($+4.12 \text{ W m}^{-2}$) leading to a peak increase in JJA near surface temperatures of 1.95 K, although the main albedo effect was seen in spring. The increase in surface fluxes was caused by an increase in net radiation due to increased SW radiation in spring and summer (5.49 W m^{-2} in MAM, 13.19 W m^{-2} in JJA), despite a decrease in LW caused by reduced low cloud fraction (-6.98% in JJA). The decrease in cloud cover was linked to a decrease in PBL moisture despite the relatively large increase in LH, which they explain by the relatively large summer temperature increase and corresponding increase in saturation vapor pressure in the $2\times\text{CO}_2$ environment forcing the simulation. No analysis of the effect on precipitation was included in their study.

As discussed further in Paper II, the spring effect on the albedo and surface temperature in our simulations may be underestimated, due to the limited treatment in this model version of the snow-masking effect of trees when the ground is covered by snow. Although the snow masking by the wooded deciduous vegetation on the tundra might be substantial (Sturm et al., 2001a), the effect of evergreen trees in spring most likely would exceed this, given a more elaborate parametrization. As such, the results for spring in these simulations might be underestimated with regard to the springtime albedo effect. This uncertainty in the results for the spring season in Paper II is investigated further in Paper III, by applying a more advanced parameterization of snow-vegetation effects.

The areas with simulated shrub expansion (conversion from mixed to wooded tundra), experienced a small increase in temperatures throughout the year (Objective 2c) and increased PBL height (Fig. 9). The high north vegetation shifts as such contribute as positive feedbacks to the theoretical initial warming, as expected based on the findings by other studies (Bonfils et al., 2012; Lawrence and Swenson, 2011). Although the effects on surface fluxes of heat and moisture resulting from this vegetation shift are weak as compared to the ones related to the boreal forest migration (Objective 2d), the results for the near surface temperature are interesting. The effect of increased shrub cover is further investigated in Paper III.

Based on a broad analysis of available observational data from Arctic and boreal ecosystems, Eugster et al. (2000) found a general increase in latent heat flux in areas with deciduous trees replacing evergreen needle leaf forest, due to lower canopy conductance in conifer trees. A similar, yet less drastic replacement was made along the southern border of the boreal forest in our experiment, where we applied a replacement of evergreen needle leaf forest by mixed forest (representing a mix of coniferous and deciduous trees). This shift in vegetation, however, caused a modelled year-round reduction in latent heat. This result is closely related to the model parameterization related to this vegetation shift, which implies a decrease in LAI, but no increase in stomatal conductance, and it thereby acts to decrease the latent heat release to the atmosphere. The relatively high stomatal conductance in evergreen needle leaved trees (as compared to the other vegetation types) in our simulations also supports the significant increase in latent heat flux in summer months in areas with evergreen forest expansion.

Out of the mentioned model studies, not all include an analysis of clouds cover or precipitation, however, the only two reporting enhanced cloud cover and rain in response to increased evergreen needle leaved forest are the Liess and Snyder (2011) study and ours. Based on previous validation studies of the WRF modelling system in high latitudes, a persistent bias have been an overestimation of PBL moisture and especially summer convective precipitation (Wilson et al., 2011, 2012). This was supported by findings by Mölders and Kramm (2010), who highlighted the strong dependence on the choice of physical parameterizations on these results. However, according to findings by Jin et al. (2010), excess precipitation in the WRF model seemed detached from land surface processes, and remained overestimated across tests of four different land surface schemes. They did, however, not include an analysis of the dependency of land cover or land cover parameter values, which remain the same across land surface schemes. Based on our results, a further investigation related to cloud cover and precipitation in the WRF model in relation to vegetation parameters would be an interesting topic for further investigation. Also, here we have chosen to perturb only the vegetation type in each area. The greenness fraction is not altered, which influences the evapotranspiration and thereby available energy for sensible heat, as demonstrated by Hong et al. (2009). This choice in perturbation seemed more appropriate for the aim of this study. However, it might have influenced the results for the surface fluxes.

We consider the vegetation shifts applied in Paper II as more moderate and, although simplified, in areal extent more realistic as compared to the ones applied in other studies with similar aim. Although stronger effects on near surface temperature and surface fluxes as compared to our results are to be expected in more extreme scenarios, it is also clear that the simulated atmospheric response is also strongly dependent on the model dependent vegetation properties, and that dominating feedback effects vary among model studies. Particularly the rate of increase in ET and LH, and the fate of excess moisture in the atmosphere seem to vary greatly among models. Further investigation into the mechanisms behind these differing results would be an interesting topic for further research.

4.3 Further investigation of the effects of increased shrub cover

The effects of increased shrub cover were largely dominated by the northward migration of boreal forest in Paper II. A further investigation of the biophysical effects related to this widely

observed vegetation change was the focus of Paper III. In this study we have again applied WRF (V3.7.1), but this time coupled with the Noah-UA land surface model, which has a more elaborate treatment of snow-vegetation interactions as compared to Noah.

For the first experiment we adjusted the existing shrub cover by applying a cover with varying shrub heights according to present day summer temperature limits, based on well documented temperature-vegetation relationships for the region. This led to an increase in low alpine shrubs (height 2 m) and in sub-alpine and boreal tall shrubs and deciduous small leaved trees (height > 5 m and 10 m, respectively) (Fig. 1, Paper III). To evaluate the sensitivity of the atmospheric response to inter-annual variations, simulations were conducted for a warm and a cold spring and summer season. The mean atmospheric response across these seasons represents a broad range in atmospheric conditions. To evaluate the sensitivity to a potential further expansion in shrub and tree cover related to a 1 K increase in mean summer temperature, we conducted additional simulations for each season, applying a new vegetation cover shifted according to new temperature limits. This led to the expansion of sub-alpine and boreal shrubs and low trees across the domain (Fig. 2, Paper III).

Other modelling studies have suggested that the effect of increased Arctic shrub cover is largest in spring due to the strong albedo decrease resulting from the influence on the melting season (Bonfils et al., 2012; Lawrence and Swenson, 2011; Miller and Smith, 2012). This was confirmed in our results (Objective 2f). The relatively subtle increase in shrub and tree cover led to increase in near surface temperatures across both the spring and summer seasons, with the largest effect seen during the melting season in mid-May, with a peak in 2 m temperature increase of 0.8 K. Furthermore, the warming was larger in the warm spring season than in the cold, during which the temperature increase as averaged over the total area with vegetation change reached 1 K (Fig. 11 and 12). Although these numbers represent comparatively moderate temperature responses, these results support previous modelling findings of peak warming anomalies following shrub cover increase during the spring melting season (Bonfils et al., 2012; Lawrence and Swenson, 2011).

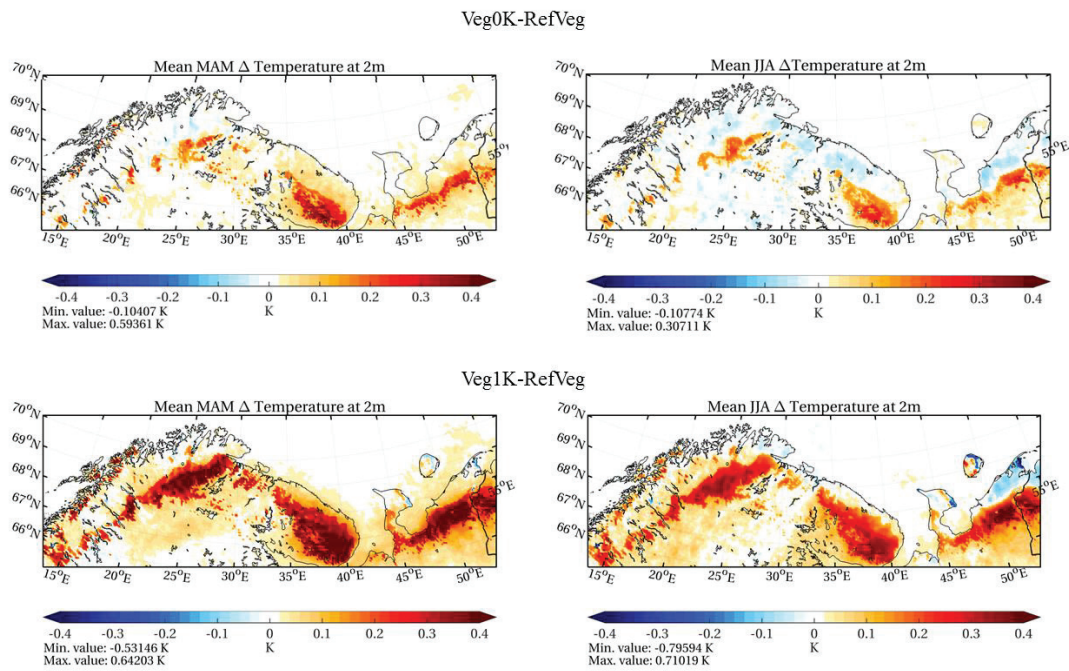


Figure 11. Effects of shrub cover increase on the seasonal mean near surface temperature. Effects of shrub cover increase on the spring season (top left panel) and summer season (top right panel), and effects of a vegetation shift corresponding to a 1 K increase in summer temperatures (bottom row), spring season (left) and summer season (right). Only showing significant results at the 95% confidence level according to a Mann-Whitney U-test of equal median. Figure modified from Paper III.

Swann et al. (2010), using the CAM3-CLM3 on a $2.8^\circ \times 2.8^\circ$ resolution found an annual temperature increase of 0.2 K resulting from mainly increased ET following a 20% increase in deciduous broadleaved tree cover in previously bare ground areas north of 60°N (representing an area about 1.75 times the size of Alaska). In their simulations the strongly enhanced ET caused an enhanced greenhouse effect through increase in atmospheric water vapor, reinforced by sea-ice interactions. They found that the warming caused by increased greenhouse effect was as strong as the one caused by decreased surface albedo. They furthermore found the largest effects on surface fluxes in summer, with a July increase in LH of 4.16 W m^{-2} , and a comparatively weak increase in SH, of 1 W m^{-2} .

By applying a uniform $\sim 20\%$ increase in shrub cover with height 0.5 m in areas with bare ground north of 60°N , Bonfils et al. (2012) found a peak warming of nearly 2 K in the June melting season (using the CAM4-CLM4 model). By increasing the height of the total shrub cover to 2 m, the result for the peak warming increased to 4.75 K in May. Similarly, Lawrence and Swenson (2011) found a peak warming resulting from shrub cover increase in May-June of $1.5\text{-}2 \pm 1 \text{ K}$.

This contrasts findings by Jeong et al. (2014), who found the largest temperature response to increased vegetation greenness in the summer season (+1.95 K), and a greater warming response also in the autumn season (0.98 K). Despite more drastic vegetation alterations as compared to our study, they find a similar spring season warming of 0.9 K.

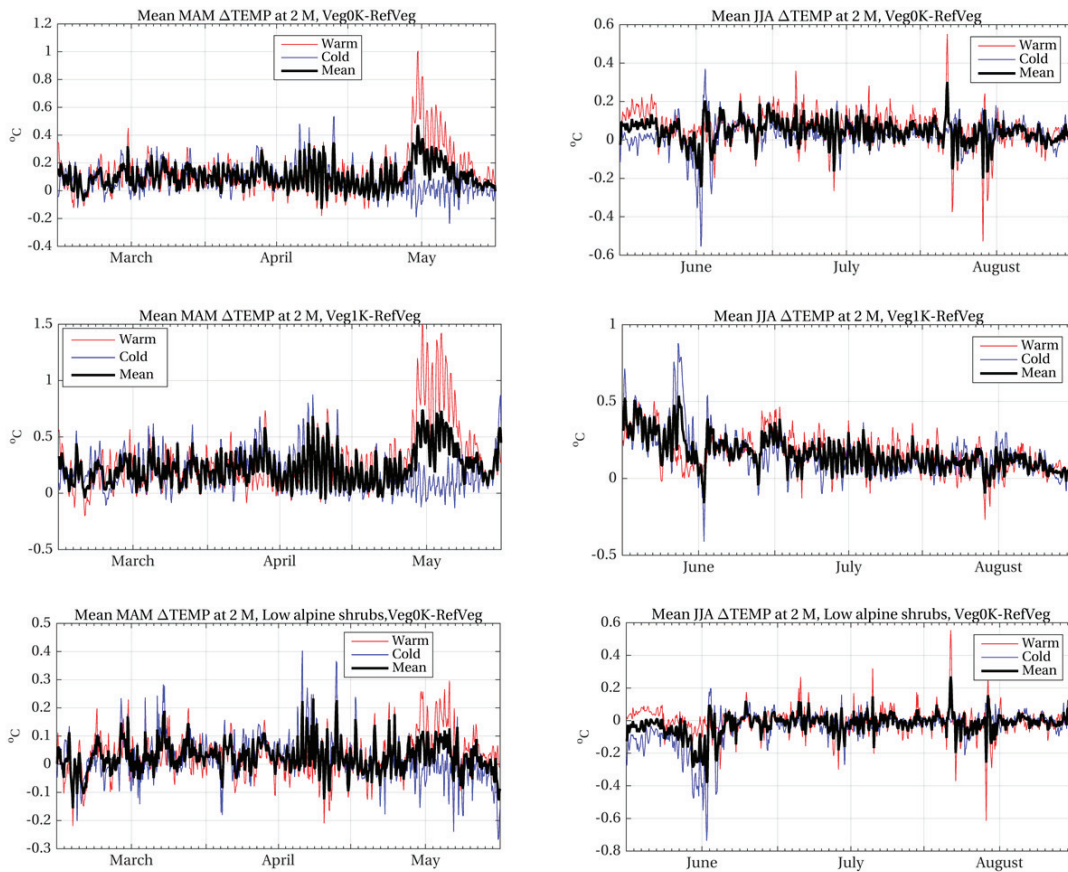


Figure 12. Time evolution of temperature anomalies following increased shrub and tree cover for the spring season (left column) and summer season (right column) as averaged over all areas with vegetation change, and averaged over only areas with low Alpine shrub increase (bottom row). Red lines represent the warm season feedback, the blue lines represent cold season feedbacks, and black lines represent mean values, which form the basis for the mapped values. Note that scales differ.

The spring temperature response in our simulations are largely attributed to the albedo feedback of enhanced snow melt due to shrub and tree increase, leading to an increase in SW absorbed at the surface, despite a weaker SW downwelling attributed to increased cloud cover (Fig. 13 and 14). As in Paper II, the increased net SW leads to enhanced LH more than SH, a deeper and wetter boundary layer, and increased cloud cover and precipitation (Fig. 14).

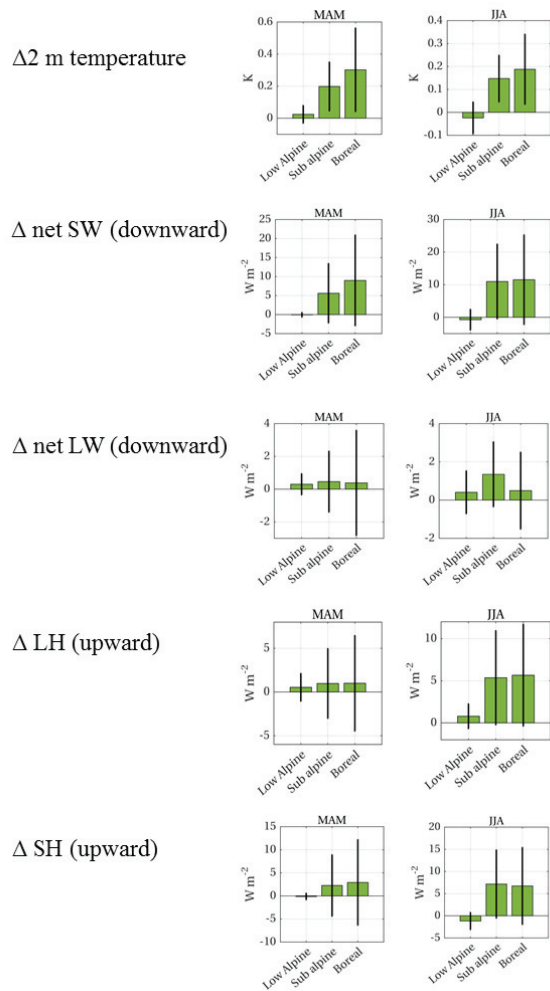


Figure 13. Mean response as averaged over the separate areas with vegetation changes (black lines indicate one σ range about the mean). Note that scales differ among panels. Figure modified from Paper III.

warm the overlying atmosphere (Bonfils et al., 2012). The peak effects on temperature, latent and sensible heat fluxes in their simulations were increased and shifted from June to May in the case with taller shrubs, according to the maximum reduction of albedo related to the earlier emergence of canopy from beneath the snow in spring, and a stronger effect on snowmelt in the tall shrub case.

The high sensitivity to shrub height, and corresponding increase in LAI was confirmed in our findings in Paper III. We found that the atmospheric warming, increase in surface fluxes, and

Summer warming was not as pronounced in our simulations, partly due to enhanced early summer snow cover and increased surface albedo in areas with low alpine shrubs (Fig. 11, bottom right panel and Fig 12, upper right panel), leading to a reduction in net SW radiation absorbed by the surface and correspondingly lowered SH and temperature decrease. The resulting mean warming response across all areas with vegetation changes was a comparatively moderate increase of 0.05 K, concealing the fact that the warming in areas with taller vegetation was almost as pronounced as during the spring season. The effects of increased shrub cover on the near surface temperature are shown in Fig. 12.

One of the objectives of this paper was to investigate the sensitivity to shrub and tree height on the atmospheric response (Objective 2e), in addition to expansion into tundra areas. Other modelling studies found that taller shrubs have a greater ability to

albedo decrease were larger in areas with increase in taller vegetation (Fig. 13). The atmospheric response in areas with expansion of low alpine shrubs (lower vegetation) were more sensitive to environmental conditions such as snow cover and mean temperature.

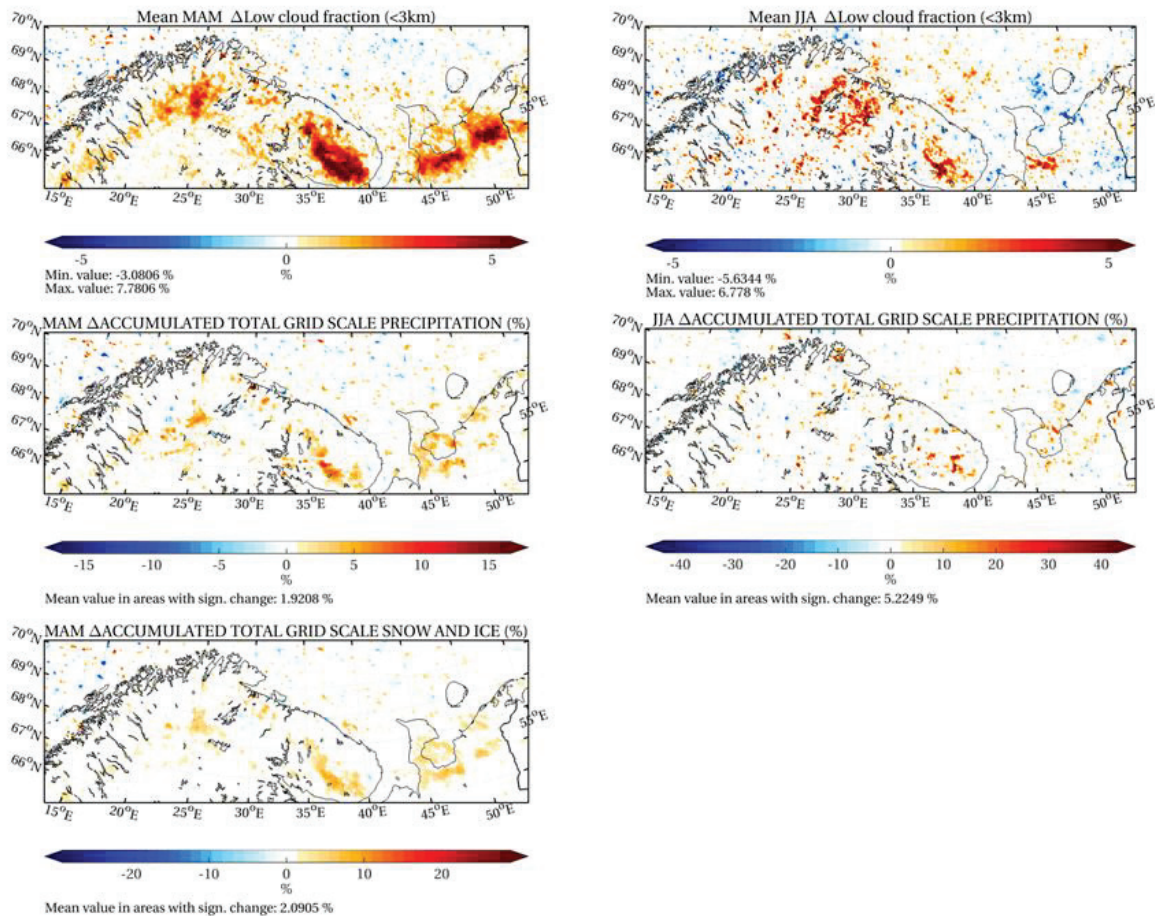


Figure 14. Mean seasonal effects on low level cloud cover fraction (top row), relative change in accumulated seasonal precipitation (middle row) and spring season snow and ice precipitation (bottom panel). Only showing significant changes at the 95% confidence level. For precipitation, significance tests are conducted on daily values of accumulated precipitation, rather than three-hourly values. Mean over spring seasons in left column, and summer season in right column. Note that scales differ among panels. Figure from Paper III.

Similar to the findings of Bonfils et al. (2012), a strong increase in LH in our simulations (fourth row, Fig. 13) led to increased cloud cover and precipitation in both spring and summer seasons (Fig. 14). The enhanced atmospheric moisture increased LW (fifth row, Fig. 13) to the surface

and reduced the albedo effect through reduced SW downwelling, although the net SW was increased in areas with taller vegetation due to strongly decreased surface albedo.

Like Swann et al. (2010), Bonfils et al., (2011) also found large increase in LH ($> 5 \text{ W m}^{-2}$ and $> 12 \text{ W m}^{-2}$ in low and tall shrub cases, respectively) acting to enhance warming through the greenhouse effect, particularly in the case with taller shrubs. In addition to increased downward net LW radiation ($5\text{-}10 \text{ W m}^{-2}$), they similarly to our results found that the increased atmospheric water content resulted in a slight increase in summer season rain, particularly in the tall shrub case.

This contrasts the findings of Jeong et al. (2014), who experienced that despite a strong increase in LH ($+4.35 \text{ W m}^{-2}$), the atmospheric relative humidity (RH) decreased, resulting in a decrease in summer low cloud cover, as mentioned in the discussion of Paper II. Similarly, Swann et al., (2011) found an increase in LH (1.39 W m^{-2} in spring, 4.16 W m^{-2} in summer) rather than SH (1.30 W m^{-2} in spring, 1 W m^{-2} in summer). However, they found that the low cloud fraction was reduced, despite increased atmospheric water vapor leading to an enhanced greenhouse effect.

The vegetation distribution applied according to a theoretical 1 K shift in summer temperatures may serve as a simplified proxy for a future scenario with regard to vegetation re-distribution, providing the results are interpreted with care (Objective 2h). We found that the larger feedback to the summer season warming (Fig. 10, bottom row and Fig. 11, second row) was related mainly to taller vegetation, and stronger in the future scenario. The spring response to increased shrub and tree cover was enhanced mainly in the warm spring season in both the present and future scenario. Based on these results, we might expect that in a warmer climate, the positive feedback effect of shrub expansion would strengthen in both spring and summer seasons.

The northward migration of taller trees and the sub-alpine ecotone in the future scenario enhanced warming in both seasons, and in some areas summer warming even more than spring. As the summer temperatures are highlighted here as one of the main environmental drivers of shrub expansion, we can conclude that the main feedback to shrub and tree growth in these areas are subject to a time lag corresponding to that of establishment of taller shrubs and sub alpine trees in tundra areas, as concluded by de Wit et al. (2014) and coherent with the results of Bonfils et al. (2012).

As in Paper II, our simulations in Paper III represented more subtle vegetation changes as compared to those applied in other modelling studies with similar aim. The vegetation changes in Paper III were applied according to well established climate-vegetation relationships in the area of interest, and aimed at representing the shrub and low tree cover in accordance with present day temperature limits. In addition a future projection, based on a theoretical summer temperature increase of 1 K, was simulated. In both simulations the shrub height increase was limited to areas already covered with lower shrubs, and trees were only allowed to expand into areas with pre-existing shrub cover. As anomalies from studies with similar aim (Swann et al., 2010; Lawrence and Swenson, 2011; Bonfils et al., 2012; Jeong et al., 2014; Jeong et al., 2011) represented effects of increased shrub and tree cover as compared to bare ground (Swann et al., 2010; Bonfils et al., 2012) or Arctic grasslands (Lawrence and Swenson 2011), or according to the dynamic vegetation response resulting from a doubling of CO₂ and corresponding heating (Jeong et al., 2014; Jeong et al., 2011), the atmospheric effects resulting from our vegetation perturbations should be expected to be less extreme. However, comparing results in Paper III to the other modelling studies and to the findings in Paper II, the near surface warming associated with our more drastic vegetation shifts were often of the same magnitude, particularly in the peak melting season.

Compared to other studies with similar aim, our study is conducted on, to our knowledge, state-of-the-art high resolution (so-called “convection-permitting” at 5.4 km x 5.4 km). As highlighted in Snyder and Liess (2014), one might expect that the modelled effects on the atmosphere caused by fine scale changes in the surface properties are better captured when resolved on higher resolution. However, it is hard to estimate to which degree this aspect has influenced our results in comparison with other modelling studies.

Based on the above comparison of modelling studies it is clear that there is a qualitatively consistent temperature response in atmospheric temperature across models and vegetation perturbations in response to increased high latitude vegetation. However considerable differences arise in the response related to the surface fluxes of heat and moisture, and particularly to the fate of additional atmospheric moisture. In our simulations strong increases in LH following the vegetation perturbations in both Paper II and Paper III lead to enhanced cloud cover and precipitation, particularly in summer. This is also the case in Bonfils et al. (2012). However, the

opposite effect on cloud cover is seen in other studies (Swann et al., 2010;Jeong et al., 2014). This relates not only to the nature of the applied changes to vegetation cover, but also to the parameterization of vegetation characteristics, and model dependent calculations of cloud microphysics and convective dynamics. In addition, the mentioned difference in spatial resolutions among the modelling studies might have influenced these results.

5 Discussion and suggestions for future research

The work presented as part of this thesis has contributed to identify several interesting topics for further research. Some of these are emphasized in the following.

Based on the findings of Paper I, we agree with the conclusions of previous validation studies in that the Wesely dry deposition scheme in WRF-Chem should be revised to better account for environmental conditions such as water availability. In addition, we suggest that some further investigation on the suitability of the temperature response function and vegetation-specific values for the minimum stomatal resistance parameters could act to improve estimation of the stomatal resistance used in the gaseous deposition. For the purpose of applying WRF-Chem in estimating potential ozone induced damage to vegetation, the stomatal conduction parameterization could be improved according to other well tested models, such as the DO₃SE model. Further testing and validation towards measurements in various ecosystems would be required in order to establish a suitable setup for applying this model in dose estimates. If the stomatal dose simulated by the model can be improved, we believe the WRF-Chem model could be a powerful and versatile tool in estimating risk of ozone induced damage to vegetation under present and future climate conditions.

The findings related to the partitioning of excess energy into latent and sensible heat fluxes in response to increased vegetation cover in Papers II and III were interesting, and to some degree they differ as compared to the results of several observational and modelling studies. It is also worth noting that this aspect of land atmosphere interactions differ notably among other modelling studies, as it is closely linked to the parameterization of vegetation properties and vegetation changes in models. Based on our findings of substantial increase in LH due to vegetation increase in high latitudes, it would be interesting to further investigate this aspect in relation to previously known biases in the WRF model, regarding excess atmospheric moisture and convective precipitation (particularly in summer) (Wilson et al., 2012; Mölders and Kramm, 2010). Special focus could be on vegetation types and particularly the corresponding parameter values.

Also, our results show that the contrasting effects of albedo decrease and increased cloud cover due to enhanced ET impact on net short wave radiation and corresponding warming. In our Paper

III the net response varies among vegetation shifts, and furthermore we have seen that this effect also varies greatly among modelling studies. This is an aspect of increased vegetation cover that is difficult to measure in situ, and modelled effects are highly sensitive to model parameterizations of vegetation properties (such as LAI, rooting depth, stomatal conductance and albedo). However, it is of great importance in estimating atmospheric effects of high latitude vegetation changes. It would be interesting to investigate this aspect further in sensitivity analyses and modelling experiments. As highlighted by Wullschläger et al., (2014) and in Bonan (2016), further investigations and development of PFT parameter values and validation datasets regarding plant functions and properties are necessary to increase accuracy in biosphere-atmosphere modelling, and would greatly enhance confidence in all aspects of modelled results.

Paper III in this thesis is conducted on a comparatively fine resolution as compared to other modelling studies focused on the atmospheric effects of increased shrub cover. To our knowledge, there are no other studies on equally fine resolution, and limited numbers of studies with similar aim, and consequentially, relating our results to other studies has proven somewhat challenging. More fine scale simulations of vegetation changes and corresponding feedbacks to the atmosphere are needed in order to increase understanding of these processes, both in the present and the future climate.

The term ecological “tipping point” can refer to the level of vegetation response, where the atmospheric warming resulting from increased shrub and tree cover feedbacks enhances the further growth to such a degree that the response becomes nonlinear in relation to the initial warming (Brook et al., 2013). Given the strong impact of the northward migrating sub-alpine ecotone on the summer temperature shown in Paper III in this thesis, we find the possibility for such a future tipping point in the circumpolar tundra area probable. This could be investigated further by e.g. model sensitivity experiments, fine scale vegetation distribution modelling and possibly observational studies.

Substantial uncertainties regarding the future response of high latitude ecosystems to climate change contribute to large spread in climate model projections of future climate (IPCC, 2013, Chapter 9; Friedlingstein et al., 2013; Piao et al., 2013). This is in part due to limitations in the representation of high latitude ecosystems in climate models (Wullschleger et al., 2014). To

improve understanding of cold climate vegetation-atmosphere feedback processes increased efforts in model development and observational studies for validation should be prioritized.

References

- Alo, C. A., and Wang, G. L.: Role of dynamic vegetation in regional climate predictions over western Africa, *Clim Dynam*, 35, 907-922, DOI 10.1007/s00382-010-0744-z, 2010.
- Arora, V. K., and Montenegro, A.: Small temperature benefits provided by realistic afforestation efforts, *Nat Geosci*, 4, 514-518, Doi 10.1038/Ngeo1182, 2011.
- Ashmore, M. R.: Assessing the future global impacts of ozone on vegetation, *Plant Cell Environ.*, 28, 949-964, 2005.
- Beringer, J., Chapin Iii, F. S., Thompson, C. C., and McGuire, A. D.: Surface energy exchanges along a tundra-forest transition and feedbacks to climate, *Agr Forest Meteorol*, 131, 143-161, <http://dx.doi.org/10.1016/j.agrformet.2005.05.006>, 2005.
- Betts, R. A., Falloon, P. D., Goldewijk, K. K., and Ramankutty, N.: Biogeophysical effects of land use on climate: Model simulations of radiative forcing and large-scale temperature change, *Agr Forest Meteorol*, 142, 216-233, DOI 10.1016/j.agrformet.2006.08.021, 2007.
- Bhatt, U. S., Walker, D. A., Raynolds, M. K., Comiso, J. C., Epstein, H. E., Jia, G., Gens, R., Pinzon, J. E., Tucker, C. J., Tweedie, C. E., and Webber, P. J.: Circumpolar Arctic Tundra Vegetation Change Is Linked to Sea Ice Decline, *Earth Interact*, 14, 8, 2010.
- Blok, D., Heijmans, M. M. P. D., Schaepman-Strub, G., Kononov, A. V., Maximov, T. C., and Berendse, F.: Shrub expansion may reduce summer permafrost thaw in Siberian tundra, *Glob. Change Biol.*, 16, 1296-1305, 10.1111/j.1365-2486.2009.02110.x, 2010.
- Bonan, G. B.: Forests and climate change: Forcings, feedbacks, and the climate benefits of forests, *Science*, 320, 1444-1449, 10.1126/science.1155121, 2008.
- Bonan, G. B.: Terrestrial Ecosystem and Earth System Models, in: *Ecological climatology : Concepts and applications*, 3rd ed., edited by: Bonan, G. B., Cambridge University Press, Cambridge, 2016.

Bonfils, C. J. W., Phillips, T. J., Lawrence, D. M., Cameron-Smith, P., Riley, W. J., and Subin, Z. M.: On the influence of shrub height and expansion on northern high latitude climate, *Environ Res Lett*, 7, Artn 015503

10.1088/1748-9326/7/1/015503, 2012.

Brook, B. W., Ellis, E. C., Perring, M. P., Mackay, A. W., and Blomqvist, L.: Does the terrestrial biosphere have planetary tipping points?, *Trends Ecol Evol*, 28, 396-401,

10.1016/j.tree.2013.01.016, 2013.

Broxton, P. D., Zeng, X. B., Sulla-Menashe, D., and Troch, P. A.: A Global Land Cover Climatology Using MODIS Data, *J Appl Meteorol Clim*, 53, 1593-1605, 10.1175/Jamc-D-13-0270.1, 2014.

Caird, M. A., Richards, J. H., and Donovan, L. A.: Nighttime stomatal conductance and transpiration in C-3 and C-4 plants, *Plant Physiol*, 143, 4-10, 10.1104/pp.106.092940, 2007.

Chapin, F. S., Sturm, M., Serreze, M. C., McFadden, J. P., Key, J. R., Lloyd, A. H., McGuire, A. D., Rupp, T. S., Lynch, A. H., Schimel, J. P., Beringer, J., Chapman, W. L., Epstein, H. E., Euskirchen, E. S., Hinzman, L. D., Jia, G., Ping, C. L., Tape, K. D., Thompson, C. D. C., Walker, D. A., and Welker, J. M.: Role of land-surface changes in Arctic summer warming, *Science*, 310, 657-660, DOI 10.1126/science.1117368, 2005.

Mapping Critical Levels for Vegetation, Chapter III of Manual on methodologies and criteria for modelling and mapping critical loads and levels and air pollution effects, risks and trends:

www.icpmapping.org, access: 19.02, 2015.

Corlett, R. T., and Westcott, D. A.: Will plant movements keep up with climate change?, *Trends Ecol Evol*, 28, 482-488, 10.1016/j.tree.2013.04.003, 2013.

Damour, G., Simonneau, T., Cochard, H., and Urban, L.: An overview of models of stomatal conductance at the leaf level, *Plant Cell Environ.*, 33, 1419-1438, 10.1111/j.1365-3040.2010.02181.x, 2010.

Davis, M. B., and Shaw, R. G.: Range Shifts and Adaptive Responses to Quaternary Climate Change, *Science*, 292, 673-679, 10.1126/science.292.5517.673, 2001.

de Wit, H. A., Bryn, A., Hofgaard, A., Karstensen, J., Kvalevag, M. M., and Peters, G. P.: Climate warming feedback from mountain birch forest expansion: reduced albedo dominates carbon uptake, *Glob. Change Biol.*, 20, 2344-2355, 10.1111/gcb.12483, 2014.

Denier van der Gon, A. Visschedijk, H. van der Brugh, and Droge, R.: A high resolution European emission data base for the year 2005, A contribution to UBA-Projekt PAREST: Particle Reduction Strategies, Utrecht, 2010a.

Denier van der Gon, H. A. Visschedijk, J. Kuenen, H. van der Brugh, R. Droge, and Schaap, M.: High resolution European emission grids for anthropogenic sources for the years 2003-2007, *Geophysical Research Abstracts*, 12, 2010b.

Elmendorf, S. C., Henry, G. H. R., Hollister, R. D., Bjork, R. G., Boulanger-Lapointe, N., Cooper, E. J., Cornelissen, J. H. C., Day, T. A., Dorrepaal, E., Elumeeva, T. G., Gill, M., Gould, W. A., Harte, J., Hik, D. S., Hofgaard, A., Johnson, D. R., Johnstone, J. F., Jonsdottir, I. S., Jorgenson, J. C., Klanderud, K., Klein, J. A., Koh, S., Kudo, G., Lara, M., Levesque, E., Magnusson, B., May, J. L., Mercado-Diaz, J. A., Michelsen, A., Molau, U., Myers-Smith, I. H., Oberbauer, S. F., Onipchenko, V. G., Rixen, C., Schmidt, N. M., Shaver, G. R., Spasojevic, M. J., Porhallsdottir, P. E., Tolvanen, A., Troxler, T., Tweedie, C. E., Villareal, S., Wahren, C. H., Walker, X., Webber, P. J., Welker, J. M., and Wipf, S.: Plot-scale evidence of tundra vegetation change and links to recent summer warming, *Nat Clim Change*, 2, 453-457, 10.1038/Nclimate1465, 2012.

Emberson, L. D., Ashmore, M. R., Simpson, D., Tuovinen, J. P., and Cambridge, H. M.: Modelling and mapping ozone deposition in Europe, 6th International Conference on Acidic Deposition, Tsukuba, Japan, 2000, ISI:000172011900067, 577-582,

Eugster, W., Rouse, W. R., Pielke, R. A., McFadden, J. P., Baldocchi, D. D., Kittel, T. G. F., Chapin, F. S., Liston, G. E., Vidale, P. L., Vaganov, E., and Chambers, S.: Land-atmosphere energy exchange in Arctic tundra and boreal forest: available data and feedbacks to climate, *Glob. Change Biol.*, 6, 84-115, DOI 10.1046/j.1365-2486.2000.06015.x, 2000.

Fowler, D., Pilegaard, K., Sutton, M. A., Ambus, P., Raivonen, M., Duyzer, J., Simpson, D., Fagerli, H., Fuzzi, S., Schjoerring, J. K., Granier, C., Neftel, A., Isaksen, I. S. A., Laj, P., Maione,

M., Monks, P. S., Burkhardt, J., Daemmgen, U., Neiryneck, J., Personne, E., Wichink-Kruit, R., Butterbach-Bahl, K., Flechard, C., Tuovinen, J. P., Coyle, M., Gerosa, G., Loubet, B., Altimir, N., Gruenhage, L., Ammann, C., Cieslik, S., Paoletti, E., Mikkelsen, T. N., Ro-Poulsen, H., Cellier, P., Cape, J. N., Horvath, L., Loreto, F., Niinemets, U., Palmer, P. I., Rinne, J., Misztal, P., Nemitz, E., Nilsson, D., Pryor, S., Gallagher, M. W., Vesala, T., Skiba, U., Brüeggemann, N., Zechmeister-Boltenstern, S., Williams, J., O'Dowd, C., Facchini, M. C., de Leeuw, G., Flossman, A., Chaumerliac, N., and Erisman, J. W.: Atmospheric composition change: Ecosystems-Atmosphere interactions, *Atmospheric Environment*, 43, 5193-5267, 10.1016/j.atmosenv.2009.07.068, 2009.

Friedl, M. A., Sulla-Menashe, D., Tan, B., Schneider, A., Ramankutty, N., Sibley, A., and Huang, X.: MODIS Collection 5 global land cover: Algorithm refinements and characterization of new datasets, *Remote Sens Environ*, 114, 168-182, <http://dx.doi.org/10.1016/j.rse.2009.08.016>, 2010.

Friedlingstein, P., Meinshausen, M., Arora, V. K., Jones, C. D., Anav, A., Liddicoat, S. K., and Knutti, R.: Uncertainties in CMIP5 Climate Projections due to Carbon Cycle Feedbacks, *J Climate*, 27, 511-526, 10.1175/JCLI-D-12-00579.1, 2013.

Gerosa, G., Marzuoli, R., Cieslik, S., and Ballarin-Denti, A.: Stomatal ozone fluxes over a barley field in Italy. "Effective exposure" as a possible link between exposure- and flux-based approaches, *Atmospheric Environment*, 38, 2421-2432, DOI 10.1016/j.atmosenv.2003.12.040, 2004.

Gerosa, G., Finco, A., Mereu, S., Marzuoli, R., and Ballarin-Denti, A.: Interactions among vegetation and ozone, water and nitrogen fluxes in a coastal Mediterranean maquis ecosystem, *Biogeosciences*, 6, 1783-1798, 2009a.

Giorgi, F., and Mearns, L. O.: Approaches to the Simulation of Regional Climate Change - a Review, *Rev Geophys*, 29, 191-216, Doi 10.1029/90rg02636, 1991.

Grell, G. A., Peckham, S. E., Schmitz, R., McKeen, S. A., Frost, G., Skamarock, W. C., and Eder, B.: Fully coupled "online" chemistry within the WRF model, *Atmospheric Environment*, 39, 6957-6975, 10.1016/j.atmosenv.2005.04.027, 2005.

Hardacre, C., Wild, O., and Emberson, L.: An evaluation of ozone dry deposition in global scale chemistry climate models, *Atmos Chem Phys*, 15, 6419-6436, 10.5194/acp-15-6419-2015, 2015.

Hayes, F., Mills, G., Harmens, H., Norris, D., and Vegetation, P. C. C. f. t. I.: Evidence of Widespread Ozone Damage to Vegetation in Europe (1990-2006), Centre for Ecology and Hydrology Bangor, UK, 2007.

Hines, K. M., and Bromwich, D. H.: Development and testing of Polar Weather Research and Forecasting (WRF) Model. Part I: Greenland ice sheet meteorology, *Mon. Weather Rev.*, 136, 1971-1989, Doi 10.1175/2007mwr2112.1, 2008.

Hines, K. M., Bromwich, D. H., Bai, L. S., Barlage, M., and Slater, A. G.: Development and Testing of Polar WRF. Part III: Arctic Land, *J Climate*, 24, 26-48, Doi 10.1175/2010jcli3460.1, 2011.

Hodnebrog, O., Stordal, F., and Berntsen, T. K.: Does the resolution of megacity emissions impact large scale ozone?, *Atmospheric Environment*, 45, 6852-6862, 2011.

Hofgaard, A., Tømmervik, H., Rees, G., and Hanssen, F.: Latitudinal forest advance in northernmost Norway since the early 20th century, *J Biogeogr*, 40, 938-949, 10.1111/jbi.12053, 2013.

Hollaway, M. J., Arnold, S. R., Challinor, A. J., and Emberson, L. D.: Intercontinental trans-boundary contributions to ozone-induced crop yield losses in the Northern Hemisphere, *Biogeosciences*, 9, 271-292, 10.5194/bg-9-271-2012, 2012.

Hong, S. B., Lakshmi, V., Small, E. E., Chen, F., Tewari, M., and Manning, K. W.: Effects of vegetation and soil moisture on the simulated land surface processes from the coupled WRF/Noah model, *Journal of Geophysical Research-Atmospheres*, 114, Artn D18118 Doi 10.1029/2008jd011249, 2009.

IPCC: Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change, Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 1535 pp., 2013.

Jarvis, P. G.: Interpretation of variations in leaf water potential and stomatal conductance found in canopies in field., *Philos. Trans. R. Soc. Lond. Ser. B-Biol. Sci.*, 273, 593-610, 1976.

Jeong, J. H., Kug, J. S., Linderholm, H. W., Chen, D. L., Kim, B. M., and Jun, S. Y.: Intensified Arctic warming under greenhouse warming by vegetation-atmosphere-sea ice interaction, *Environ Res Lett*, 9, Artn 094007

Doi 10.1088/1748-9326/9/9/094007, 2014.

Jeong, S. J., Ho, C. H., Park, T. W., Kim, J., and Levis, S.: Impact of vegetation feedback on the temperature and its diurnal range over the Northern Hemisphere during summer in a 2 x CO₂ climate, *Clim Dynam*, 37, 821-833, DOI 10.1007/s00382-010-0827-x, 2011.

Jin, J., Miller, N. L., and Schlegel, N.: Sensitivity Study of Four Land Surface Schemes in the WRF Model, *Advances in Meteorology*, 2010, 11 pp., 10.1155/2010/167436, 2010.

Klingberg, J., Engardt, M., Karlsson, P. E., Langner, J., and Pleijel, H.: Declining ozone exposure of European vegetation under climate change and reduced precursor emissions, *Biogeosciences*, 11, 5269-5283, 10.5194/bg-11-5269-2014, 2014.

Kuenen, J. J. P., Visschedijk, A. J. H., Jozwicka, M., and van der Gon, H.: TNO-MACC_II emission inventory; a multi-year (2003-2009) consistent high-resolution European emission inventory for air quality modelling, *Atmos Chem Phys*, 14, 10963-10976, 10.5194/acp-14-10963-2014, 2014.

Lawrence, D. M., and Swenson, S. C.: Permafrost response to increasing Arctic shrub abundance depends on the relative influence of shrubs on local soil cooling versus large-scale climate warming, *Environ Res Lett*, 6, Artn 045504

10.1088/1748-9326/6/4/045504, 2011.

Liess, S., Snyder, P. K., and Harding, K. J.: The effects of boreal expansion on the summer Arctic frontal zone, *Clim Dyn*, 10.1007/s00382-011-1064-7, 2011.

Liess, S., Snyder, P. K., and Harding, K. J.: The effects of boreal forest expansion on the summer Arctic frontal zone, *Clim Dynam*, 38, 1805-1827, DOI 10.1007/s00382-011-1064-7, 2012.

Lorantý, M. M., Berner, L. T., Goetz, S. J., Jin, Y. F., and Randerson, J. T.: Vegetation controls on northern high latitude snow-albedo feedback: observations and CMIP5 model simulations, *Glob. Change Biol.*, 20, 594-606, Doi 10.1111/Gcb.12391, 2014.

Mckenney, D. W., Pedlar, J. H., Lawrence, K., Campbell, K., and Hutchinson, M. F.: Potential impacts of climate change on the distribution of North American trees, *Bioscience*, 57, 939-948, Doi 10.1641/B571106, 2007.

Mereu, S., Gerosa, G., Finco, A., Fusaro, L., Muys, B., and Manes, F.: Improved sapflow methodology reveals considerable night-time ozone uptake by Mediterranean species, *Biogeosciences*, 6, 3151-3162, 2009.

Miller, P. A., and Smith, B.: Modelling Tundra Vegetation Response to Recent Arctic Warming, *Ambio*, 41, 281-291, 10.1007/s13280-012-0306-1, 2012.

Mills, G., Hayes, F., Simpson, D., Emberson, L., Norris, D., Harmens, H., and Búker, P.: Evidence of widespread effects of ozone on crops and (semi-)natural vegetation in Europe (1990-2006) in relation to AOT40-and flux-based risk maps, *Glob. Change Biol.*, 17, 592-613, DOI 10.1111/j.1365-2486.2010.02217.x, 2011a.

Mills, G., Pleijel, H., Braun, S., Búker, P., Bermejo, V., Calvo, E., Danielsson, H., Emberson, L., Fernandez, I. G., Grunhage, L., Harmens, H., Hayes, F., Karlsson, P. E., and Simpson, D.: New stomatal flux-based critical levels for ozone effects on vegetation, *Atmospheric Environment*, 45, 5064-5068, DOI 10.1016/j.atmosenv.2011.06.009, 2011b.

Myers-Smith, I. H., Forbes, B. C., Wilmking, M., Hallinger, M., Lantz, T., Blok, D., Tape, K. D., Macias-Fauria, M., Sass-Klaassen, U., Levesque, E., Boudreau, S., Ropars, P., Hermanutz, L., Trant, A., Collier, L. S., Weijers, S., Rozema, J., Rayback, S. A., Schmidt, N. M., Schaepman-Strub, G., Wipf, S., Rixen, C., Menard, C. B., Venn, S., Goetz, S., Andreu-Hayles, L., Elmendorf, S., Ravolainen, V., Welker, J., Grogan, P., Epstein, H. E., and Hik, D. S.: Shrub expansion in tundra ecosystems: dynamics, impacts and research priorities, *Environ Res Lett*, 6, Artn 045509 10.1088/1748-9326/6/4/045509, 2011.

Myers-Smith, I. H., Elmendorf, S. C., Beck, P. S. A., Wilmsking, M., Hallinger, M., Blok, D., Tape, K. D., Rayback, S. A., Macias-Fauria, M., Forbes, B. C., Speed, J. D. M., Boulanger-Lapointe, N., Rixen, C., Levesque, E., Schmidt, N. M., Baittinger, C., Trant, A. J., Hermanutz, L., Collier, L. S., Dawes, M. A., Lantz, T. C., Weijers, S., Jorgensen, R. H., Buchwal, A., Buras, A., Naito, A. T., Ravolainen, V., Schaepman-Strub, G., Wheeler, J. A., Wipf, S., Guay, K. C., Hik, D. S., and Vellend, M.: Climate sensitivity of shrub growth across the tundra biome (vol 5, pg 887, 2015), *Nat Clim Change*, 5, 2015a.

Myers-Smith, I. H., Elmendorf, S. C., Beck, P. S. A., Wilmsking, M., Hallinger, M., Blok, D., Tape, K. D., Rayback, S. A., Macias-Fauria, M., Forbes, B. C., Speed, J. D. M., Boulanger-Lapointe, N., Rixen, C., Levesque, E., Schmidt, N. M., Baittinger, C., Trant, A. J., Hermanutz, L., Collier, L. S., Dawes, M. A., Lantz, T. C., Weijers, S., Jorgensen, R. H., Buchwal, A., Buras, A., Naito, A. T., Ravolainen, V., Schaepman-Strub, G., Wheeler, J. A., Wipf, S., Guay, K. C., Hik, D. S., and Vellend, M.: Climate sensitivity of shrub growth across the tundra biome, *Nat Clim Change*, 5, 887-+, [10.1038/Nclimate2697](https://doi.org/10.1038/Nclimate2697), 2015b.

Mölders, N., and Kramm, G.: A case study on wintertime inversions in Interior Alaska with WRF, *Atmos Res*, 95, 314-332, <http://dx.doi.org/10.1016/j.atmosres.2009.06.002>, 2010.

Pacifico, F., Folberth, G. A., Sitch, S., Haywood, J. M., Rizzo, L. V., Malavelle, F. F., and Artaxo, P.: Biomass burning related ozone damage on vegetation over the Amazon forest: a model sensitivity study, *Atmos Chem Phys*, 15, 2791-2804, [10.5194/acp-15-2791-2015](https://doi.org/10.5194/acp-15-2791-2015), 2015.

Pearson, R. G., Phillips, S. J., Lorant, M. M., Beck, P. S. A., Damoulas, T., Knight, S. J., and Goetz, S. J.: Shifts in Arctic vegetation and associated feedbacks under climate change, *Nat Clim Change*, 3, 673-677, [10.1038/Nclimate1858](https://doi.org/10.1038/Nclimate1858), 2013.

Piao, S. L., Sitch, S., Ciais, P., Friedlingstein, P., Peylin, P., Wang, X. H., Ahlstrom, A., Anav, A., Canadell, J. G., Cong, N., Huntingford, C., Jung, M., Levis, S., Levy, P. E., Li, J. S., Lin, X., Lomas, M. R., Lu, M., Luo, Y. Q., Ma, Y. C., Myneni, R. B., Poulter, B., Sun, Z. Z., Wang, T., Viovy, N., Zaehle, S., and Zeng, N.: Evaluation of terrestrial carbon cycle models for their response to climate variability and to CO₂ trends, *Glob. Change Biol.*, 19, 2117-2132, [10.1111/gcb.12187](https://doi.org/10.1111/gcb.12187), 2013.

Pleijel, H., Danielsson, H., Emberson, L., Ashmore, M. R., and Mills, G.: Ozone risk assessment for agricultural crops in Europe: Further development of stomatal flux and flux-response relationships for European wheat and potato, *Atmospheric Environment*, 41, 3022-3040, 2007.

Prentice, I. C., Bondeau, A., Cramer, W., Harrison, S. P., Hickler, T., Lucht, W., Sitch, S., Smith, B., and Sykes, M. T.: Dynamic Global Vegetation Modeling: Quantifying Terrestrial Ecosystem Responses to Large-Scale Environmental Change, in: *Terrestrial Ecosystems in a Changing World*, edited by: Canadell, J. G., Pataki, D. E., and Pitelka, L. F., Springer Berlin Heidelberg, Berlin, Heidelberg, 175-192, 2007.

Rietkerk, M., Brovkin, V., van Bodegom, P. M., Claussen, M., Dekker, S. C., Dijkstra, H. A., Goryachkin, S. V., Kabat, P., van Nes, E. H., Neutel, A. M., Nicholson, S. E., Nobre, C., Petoukhov, V., Provenzale, A., Scheffer, M., and Seneviratne, S. I.: Local ecosystem feedbacks and critical transitions in the climate, *Ecol Complex*, 8, 223-228, DOI 10.1016/j.ecocom.2011.03.001, 2011.

Rummukainen, M.: State-of-the-art with regional climate models, *Wires Clim Change*, 1, 82-96, 10.1002/wcc.8, 2010.

Rummukainen, M., Rockel, B., Barring, L., Christensen, J. H., and Reckermann, M.: Twenty-First-Century Challenges in Regional Climate Modeling, *B Am Meteorol Soc*, 96, Es135-Es138, 10.1175/Bams-D-14-00214.1, 2015.

Serreze, M. C., and Barry, R. G.: Processes and impacts of Arctic amplification: A research synthesis, *Global Planet Change*, 77, 85-96, DOI 10.1016/j.gloplacha.2011.03.004, 2011.

Sitch, S., Cox, P. M., Collins, W. J., and Huntingford, C.: Indirect radiative forcing of climate change through ozone effects on the land-carbon sink, *Nature*, 448, 791-U794, Doi 10.1038/Nature06059, 2007.

Skamarock, W. C., Klemp, J. B., Dudhia, J., Gill, D. O., Barker, D. M., Duda, M. G., Huang, X.-Y., Wang, W., and Powers, J. G.: A Description of the Advanced Research WRF Version 3, National Center for Atmospheric Research, Boulder, Colorado, USA, 2008.

Snyder, P. K., and Liess, S.: The simulated atmospheric response to expansion of the Arctic boreal forest biome, *Clim Dynam*, 42, 487-503, DOI 10.1007/s00382-013-1746-4, 2014.

Soja, A. J., Tchebakova, N. M., French, N. H. F., Flannigan, M. D., Shugart, H. H., Stocks, B. J., Sukhinin, A. I., Parfenova, E. I., Chapin, F. S., and Stackhouse, P. W.: Climate-induced boreal forest change: Predictions versus current observations, *Global Planet Change*, 56, 274-296, DOI 10.1016/j.gloplacha.2006.07.028, 2007.

Strengers, B. J., Muller, C., Schaeffer, M., Haarsma, R. J., Severijns, C., Gerten, D., Schaphoff, S., van den Houdt, R., and Oostenrijk, R.: Assessing 20th century climate-vegetation feedbacks of land-use change and natural vegetation dynamics in a fully coupled vegetation-climate model, *Int J Climatol*, 30, 2055-2065, 10.1002/joc.2132, 2010.

Sturm, M., McFadden, J. P., Liston, G. E., Chapin, F. S., Racine, C. H., and Holmgren, J.: Snow-shrub interactions in Arctic tundra: A hypothesis with climatic implications, *J Climate*, 14, 336-344, Doi 10.1175/1520-0442(2001)014<0336:Ssiat>2.0.Co;2, 2001a.

Sturm, M., Racine, C., and Tape, K.: Climate change - Increasing shrub abundance in the Arctic, *Nature*, 411, 546-547, Doi 10.1038/35079180, 2001b.

Sturm, M., Douglas, T., Racine, C., and Liston, G. E.: Changing snow and shrub conditions affect albedo with global implications, *J Geophys Res-Bioge*, 110, Artn G01004
10.1029/2005jg000013, 2005.

Swann, A. L., Fung, I. Y., Levis, S., Bonan, G. B., and Doney, S. C.: Changes in Arctic vegetation amplify high-latitude warming through the greenhouse effect, *Proceedings of the National Academy of Sciences*, 107, 1295-1300, 10.1073/pnas.0913846107, 2010.

Søvde, O. A., Prather, M. J., Isaksen, I. S. A., Berntsen, T. K., Stordal, F., Zhu, X., Holmes, C. D., and Hsu, J.: The chemical transport model Oslo CTM3, *Geosci Model Dev*, 5, 1441-1469, 10.5194/gmd-5-1441-2012, 2012.

Tewari, M., Chen, F., Wang, W., Dudhia, J., LeMone, M. A., Mitchell, K., Ek, M., Gayno, G., Wegiel, J., and Cuenca, R. H.: Implementation and verification of the unified NOAA land

surface model in the WRF model, 20th conference on weather analysis and forecasting/16th conference on numerical weather prediction, 2004.

The Royal Society: Ground-level ozone in the 21st century: future trends, impacts and policy implications, Royal Society Policy Document 15/08, Commissioned report, The Royal Society, London, 2008.

Thompson, C., Beringer, J., Chapin, F. S., and McGuire, A. D.: Structural complexity and land-surface energy exchange along a gradient from arctic tundra to boreal forest, *J Veg Sci*, 15, 397-406, DOI 10.1111/j.1654-1103.2004.tb02277.x, 2004.

Tommervik, H., Johansen, B., Riseth, J. A., Karlsen, S. R., Solberg, B., and Hogda, K. A.: Above ground biomass changes in the mountain birch forests and mountain heaths of Finnmarksvidda, northern Norway, in the period 1957-2006, *Forest Ecol Manag*, 257, 244-257, 10.1016/j.foreco.2008.08.038, 2009.

Van Dingenen, R., Dentener, F. J., Raes, F., Krol, M. C., Emberson, L., and Cofala, J.: The global impact of ozone on agricultural crop yields under current and future air quality legislation, *Atmospheric Environment*, 43, 604-618, 10.1016/j.atmosenv.2008.10.033, 2009.

Verheijen, L. M., Brovkin, V., Aerts, R., Bonisch, G., Cornelissen, J. H. C., Kattge, J., Reich, P. B., Wright, I. J., and van Bodegom, P. M.: Impacts of trait variation through observed trait-climate relationships on performance of an Earth system model: a conceptual analysis, *Biogeosciences*, 10, 5497-5515, 10.5194/bg-10-5497-2013, 2013.

Wang, Y. P., Lu, X. J., Wright, I. J., Dai, Y. J., Rayner, P. J., and Reich, P. B.: Correlations among leaf traits provide a significant constraint on the estimate of global gross primary production, *Geophys Res Lett*, 39, Artn L19405
10.1029/2012gl053461, 2012.

Wang, Z., Zeng, X., and Decker, M.: Improving snow processes in the Noah land model, *Journal of Geophysical Research: Atmospheres*, 115, D20108, 10.1029/2009JD013761, 2010.

Wesely, M. L.: Parameterization of surface resistances to gaseous dry deposition in regional-scale numerical-models, *Atmospheric Environment*, 23, 1293-1304, 1989.

Wilson, A. B., Bromwich, D. H., and Hines, K. M.: Evaluation of Polar WRF forecasts on the Arctic System Reanalysis domain: Surface and upper air analysis, *Journal of Geophysical Research-Atmospheres*, 116, Artn D11112

Doi 10.1029/2010jd015013, 2011.

Wilson, A. B., Bromwich, D. H., and Hines, K. M.: Evaluation of Polar WRF forecasts on the Arctic System Reanalysis Domain: 2. Atmospheric hydrologic cycle, *Journal of Geophysical Research-Atmospheres*, 117, Artn D04107

Doi 10.1029/2011jd016765, 2012.

Wolf, A., Callaghan, T. V., and Larson, K.: Future changes in vegetation and ecosystem function of the Barents Region, *Climatic Change*, 87, 51-73, 10.1007/s10584-007-9342-4, 2008.

Wu, Z., Wang, X., Chen, F., Turnipseed, A. A., Guenther, A. B., Niyogi, D., Charusombat, U., Xia, B., William Munger, J., and Alapaty, K.: Evaluating the calculated dry deposition velocities of reactive nitrogen oxides and ozone from two community models over a temperate deciduous forest, *Atmospheric Environment*, 45, 2663-2674, <http://dx.doi.org/10.1016/j.atmosenv.2011.02.063>, 2011.

Wullschleger, S. D., Epstein, H. E., Box, E. O., Euskirchen, E. S., Goswami, S., Iversen, C. M., Kattge, J., Norby, R. J., van Bodegom, P. M., and Xu, X. F.: Plant functional types in Earth system models: past experiences and future directions for application of dynamic vegetation models in high-latitude ecosystems, *Annals of Botany*, 114, 1-16, 10.1093/aob/mcu077, 2014.

Xu, L., Myneni, R. B., Chapin, F. S., Callaghan, T. V., Pinzon, J. E., Tucker, C. J., Zhu, Z., Bi, J., Ciais, P., Tommervik, H., Euskirchen, E. S., Forbes, B. C., Piao, S. L., Anderson, B. T., Ganguly, S., Nemani, R. R., Goetz, S. J., Beck, P. S. A., Bunn, A. G., Cao, C., and Stroeve, J. C.: Temperature and vegetation seasonality diminishment over northern lands, *Nat Clim Change*, 3, 581-586, 10.1038/Nclimate1836, 2013.

Yue, X., and Unger, N.: Ozone vegetation damage effects on gross primary productivity in the United States, *Atmos Chem Phys*, 14, 9137-9153, 10.5194/acp-14-9137-2014, 2014.

6 Scientific papers

