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Estimation of different forest-related contributions to the radiative balance using observations in southern Finland

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The effect of boreal forest on the Earth's energy budget was estimated in field conditions in Hyytiälä, southern Finland. The indirect aerosol effect due to new particle formation, the effect of forest carbon sequestration and the effect of forestation- or deforestation-related albedo change were investigated. The young, fast-growing boreal forest was estimated to be a relatively significant carbon sink and an important aerosol source, and thus the effect on the Earth's energy budget was seen to be negative. The estimated values (reported as annually averaged energy budget perturbations per square meter of forest) were –36 W m^{-2} due to the carbon sink, +10 W m^{-2} due to the albedo effect and between –5 and –14 W m^{-2} due to the aerosol effect. However, the values are uncertain.

Introduction

The connection between forests and climate is important, complicated and far from fully understood. Recently, forests and activities concerning them have been the subject of intense political discussion due to their potential for climate change mitigation through carbon dioxide sequestration. However, focusing only on the carbon balance — as the Kyoto Protocol to the

UN Framework Convention on Climate Change does — disregards several other potentially important effects of forests on the global energy budget. Forests may contribute significantly to balances of other greenhouse gases such as methane or N_2O, and activities such as aforestation and deforestation also have a direct physical impact on the radiation energy balance through changes in the surface albedo. According to Betts (2000), this impact may, in boreal areas, be comparable in magnitude but opposite in sign to the cooling effect caused by carbon sequestration. Furthermore, biogenic aerosol emissions from forests may exert an additional cooling effect that is also significant as compared with the sequestration and surface albedo-change effects.

Forests emit a large variety of volatile organic vapours (e.g. Jansson *et al*. 2001). Several of the less volatile oxidation products of these gases may participate in the formation of aerosols. The effect of biogenic aerosol emissions from forests on climate is poorly understood, and only a few quantitative estimates have been made.

Estimating the contribution of biogenic aerosol emissions to the radiative balance is much more difficult than estimating the radiative forcing due to forest carbon sequestration or forestation-related surface albedo change. There are two main reasons for this. First, the precise role of biogenic emissions in aerosol formation is unknown, making it impossible to accurately determine the amount or type of aerosol particles formed specifically due to the presence of a forest. Second, while albedo-related radiative forcings are caused by changes occurring in a defined area, and carbon-budget changes can be assumed to affect the entire atmosphere relatively evenly, biogenic aerosols can be transported far from their site of production, but are too short-lived to spread through the entire atmosphere (e.g. Boucher *et al*. 2001).

The vast difference in atmospheric lifetimes makes it also difficult to accurately compare carbon dioxide- and aerosol-related radiation energy budget perturbations. Aerosols have lifetimes measured in days or weeks, while carbon dioxide emissions or sinks exert their influence on the atmosphere over several decades.

In general, aerosol emissions can affect the global radiative balance in three ways. First, aerosols can alter directly the radiative properties of the atmosphere by reflecting and/or absorbing both sunlight and infrared radiation. This "direct effect" can thus cause either cooling or warming. Second, aerosols can contribute to the formation of clouds and act as cloud condensation nuclei (CCN) within existing clouds, thus increasing the net cloud albedo. This is often referred to as the "cloud-albedo" or "first indirect" effect, and is generally cooling in nature. Third, an increase in concentration (but a decrease in size) of CCN caused by aerosol emissions can increase average cloud lifetimes and liquid water contents, thus decreasing precipitation. This cloud-lifetime effect, often called the "second indirect effect", is very difficult to quantify but probably, though not certainly, exerts a cooling effect. (e.g. Boucher *et al*. 2001).

Estimating the direct climatic effects of biogenic aerosols would require more detailed knowledge of their chemical and physical composition than is currently available. The cloud-lifetime effect, likewise, can probably not be estimated using a "bottom-up" approach, but requires "top-down" calculations in the form of advanced climate models. The scope of this study is thus limited to estimating the atmospheric cooling associated with the first indirect effect. Since biogenic aerosols are not very strongly absorbing, and since cloud long-wave albedo changes associated with increased CCN concentrations are usually minimal, any positive contributions to the radiative balance are very probably insignificant. Results from studies concerning other aerosols (Rotstayn 1999, Rotstayn and Penner 2001) indicate that an inclusion of the direct and cloud-lifetime effects would probably increase the strength of the total cooling, but probably by no more than a factor of two.

In this study, we use the term "radiative forcing" to denote the annually averaged perturbation in the Earth's radiation energy budget caused by some activity, before any temperatures or states of the atmosphere-surface system are allowed to reach equilibrium. Our definition of the radiative forcing differs somewhat from the International Panel on Climate Change's (IPCC) definition (Boucher *et al*. 2001), generally used in Global Circulation Model (GCM) simulations, which requires stratospheric temperatures

to reach equilibrium before final forcing values are calculated. In bottom-up, order-of-magnitude approaches, such as that employed in this study, such a definition is clearly impractical. Also, most climate studies usually refer to global radiative forcing values, while this study focuses on comparing the different climate effects of small areas of forest. The results obtained have the same dimension as global radiative forcings; W m^{-2}. However, this should be understood to mean watts per square meter of forest. It is tempting to call the value obtained a "local" radiative forcing, but this is not appropriate: a "local forcing" usually refers to a perturbation that an activity in a certain area would have on the radiative energy budget of the atmospheric column above that area if no other parts of the atmosphere were affected. The calculations presented in this study are, on the contrary, based upon the assumption that aerosol emissions from a unit area of forest are spread over a much larger area, in order for the changes in CCN concentrations to be relatively small.

The spatial scale of our forcing estimates is thus intermediate in nature: we assume the cause to be "local", but the effects to be "global" — we are estimating the total, global difference in energy absorbed by the climate system due to the emissions (or sequestration) of one square meter of forest. The results are expressed as W m^{-2} instead of, say, J m^{-2} $year^{-1}$, only for reasons of convenience. The main purpose of this study is to compare the magnitude of three different climate effects of boreal forests with each other — comparisons with other types of "forcing" estimates should be made only with utmost care. For example, the assumption that the CCN concentration changes caused by aerosol emissions are small makes it impossible to accurately extrapolate global radiative forcing estimates from the data obtained, as the changes in CCN concentrations caused by emissions from large forest areas are likely to be large. One possible approach is to divide the forcing values by the Earth's surface area to obtain a differential — a measure of the rate of change of the global radiative forcing as a function of forest area. However, even the differential approach is valid only if forests are compared with land types that produce no aerosols themselves.

The intermediate nature of our forcing concept is partially demonstrated by the fact that our forcing values lie in between the "global" and "local" values presented in other studies — global aerosol forcing estimates usually have magnitudes on the order of –1 W m^{-2} (e.g. Boucher *et al*. 2001), while local forcing estimates can be as large as –56 W m^{-2} (Redemann *et al*. 2000). It should, however, be noted that these local forcing estimates deal with the direct rather than the cloud-albedo effect, and with anthropogenic rather than biogenic aerosols.

In this study we will estimate the effect of boreal forests on the Earth's energy budget. We will investigate the effect of boreal forest aerosol emissions, carbon sequestration and forestation-related albedo change based on data collected from the SMEAR II site at Hyytiälä, Finland.

Experimental methods

Continuous measurements of ultrafine aerosol particle concentrations, their vertical net flux and relevant background data (local meteorology, micrometeorology, vertical profiles of inorganic gases) were made at the SMEAR II station and have been described in detail by Kulmala *et al*. (2001). The local formation rate of particles and the fate of nascent particles have been identified.

Site description

The aerosol observation started at the SMEAR II station (Station for Measuring Forest Ecosystem-Atmosphere Relations) in Hyytiälä, southern Finland (61°51´N, 24°17´E, 181 m above sea level) in January 1996, and carbon flux measurements in April 1996. The station represents boreal coniferous forests, which cover 8% of the earth's land surface and store about 10% of the total carbon of the terrestrial ecosystem. The biggest city near the SMEAR II station is Tampere, which is about 60 km from the measurement site, and has about 200 000 inhabitants.

The SMEAR II station is located in a homogeneous Scots pine stand (*Pinus sylvestris* L.), sown in 1962 next to the Hyytiälä forest station in southern Finland. Like 29% of the forests

in southern Finland, the Scots pine forest in Hyytiälä is of medium site quality, *Vaccinium* type according to the Cajander site class system (Cajander 1909), and has a typical growth rate of 8 m^3 ha^{-1} yr^{-1}. The forest is 40 years old, halfway through the rotation time for this site type, which is about 80 years. The dominating species in 56% of the forest area in southern Finland is the Scots pine. The forest has been managed along the usual silvicultural guidelines (Peltola 2001). The height of the dominant stand is 13.3 m and the all-sided needle area is 7 m^2 m^{-2}. The tree biomass is 68 t ha^{-1} (above-, and below-ground) (Ilvesniemi and Liu 2001). The homogeneous fetch in the prevailing wind direction (230°) is 250 m (Vesala *et al*. 1998). The annual mean temperature in 1961–1990 was +2.9 °C and the annual mean precipitation was 709 mm.

Aerosol events

Continuous measurements of the submicron aerosol number size distribution have been performed every 10 minutes at the SMEAR II station since January 1996. The aerosol size distributions were measured with a system consisting of two Differential Mobility Particle Sizers (DMPS) at 2 m above ground. The DMPS spectrums of the selected aerosol formation events were analyzed to obtain the start and end times of the events as well as the condensation growth rate of new particles. This growth rate was then used to calculate the time for the new particles to reach a given CCN threshold diameter. A 30-minute-average concentration of particles of a size greater than the threshold diameter was then computed at the start of the event and at the time the new particles had reached the threshold diameter. This time was calculated from the growth rate, assuming that the growth rate was constant during the time period. The difference of these concentrations was taken as the number of CCN formed. In practice only the events where the air mass stayed unchanged enough during the growth were considered. If the growth was observable only for a few hours, or the new particle mode was badly obscured by pre-existing aerosols, the event was classified to be too uncertain to be used in calculations.

Fluxes

The eddy covariance measurements were carried out at the heights of 23.3 and 46.0 m. The measurement system consists of an ultrasonic fast-response anemometer (Gill Solent 1012R) and a fast-response gas analyser (Li-Cor LI-6262). The storage flux (the accumulation of CO_2) was estimated by means of gas gradient measurements with URAS 4 (Hartmann & Braun) infrared analyzers in sequence. Vesala *et al*. (1998) have described the measuring system and the site in more detail, and the post-processing procedure is presented by Rannik (1998).

Theory

Estimation of biogenic secondary aerosol contribution to the radiative balance

In this section we first estimate the number of biogenic aerosols capable of CCN activation produced at the measurement site, and then calculate the annually averaged radiation energy budget perturbation caused by changes in CCN concentrations.

The production of new particles capable of CCN activation during nucleation events was estimated from data obtained during the years 1996, 1997, 1998, 1999, 2000 and 2001 at the SMEAR II station in Hyytiälä, as described in the previous section. Radiative forcing values were calculated for particles of a threshold diameter 40, 60, and 80 nm.

It was assumed that all particles with a diameter greater than the threshold value were capable of forming CCN. This assumption is reasonable at least for the higher threshold values, as the minimum radius, r, (in nanometres) required for CCN activation is (Hobbs 1993):

$$r = 15.3\varepsilon^{-0.31}S^{-2/3}, \quad (1)$$

where ε is the soluble fraction of the particle and S is the supersaturation in percent. For 40 nm diameter particles at the Hyytiälä site, soluble fractions were usually higher than 0.2 (Hämeri *et al*. 2001). The supersaturation required for the

CCN activation was then around 1.4% for 40-nm particles and 0.50% for 80-nm particles. The latter values, at least, are a level often reached during a cloud formation (Hobbs 1993). The error caused by the fraction of particles greater than the threshold value that do not activate due to lower soluble fractions is offset to some extent by the activation of smaller particles with higher soluble fractions. It is unclear to which extent the formation and/or growth of the new particles was caused by biogenic emissions, so the values obtained must be treated as maximum-limit estimates. Based on Kulmala *et al*. (2001), aerosol production was assumed to occur within a 1000-m-high boundary layer. The number of recently formed aerosols bigger than a certain threshold size over one square meter of forest in each nucleation event, N_1, was thus directly obtained from the measured concentration change (1 particle (cm^{-3}) = 10^9 particles (m^{-2})). In order to simplify the following calculations, the emitting forest area was fixed as one square meter and N_1 was given as a pure number.

The produced aerosols were assumed to spread out over a volume with a height (h) of 1000 m and a sufficiently large base area (A, with dimensions of m^{-2}) for the resulting change in CCN concentration to be small as compared with the original concentration (in practice, A = 100 m^2 or greater was sufficient). The assumption that the height of the volume was equal to the height of the formation layer means, in effect, that the vertical spreading of aerosol particles was disregarded. The change in CCN concentration (ΔN, with dimensions of particles (m^{-3})) in this volume was then:

$$\Delta N = \frac{N_1}{Ah}. \tag{2}$$

The cloud albedo change caused by a change in CCN concentrations can be estimated by using the linear form of Twomey's cloud sensitivity equation (Hobbs 1993, Twomey 1977):

$$\frac{\partial R_c}{\partial N} \approx \frac{\Delta R_c}{\Delta N} = \frac{R_c(1-R_c)}{3N}, \tag{3}$$

from which we obtain:

$$\Delta R_c = \frac{R_c(1-R_c)N_1}{3NAh}. \tag{4}$$

Here R_c is the original cloud albedo (assumed to vary between 0.4 and 0.6) and N is the original CCN concentration (in particles (m^{-3})). Based on data from Heymsfield (1993), N was assumed to vary between 100 cm^{-3} for thin maritime clouds and 600 cm^{-3} for thicker continental clouds. Using the observed aerosol concentrations in Hyytiälä during the non-event times these estimations seem to be realistic. Cloud albedos were naturally assumed to correlate with N; thicker clouds were assigned higher albedo values. Originally a differential expression, Eq. 3 is valid if the changes are sufficiently small, i.e. $\Delta N << N$ and $\Delta R_c << R_c$.

Next, the net albedo change caused by the biogenic CCN was calculated. In order to do this accurately, the surface albedo R_s had to be estimated. Accounting for multiple reflections between the surface and clouds, the following expression was obtained for the total albedo R_{tot} (Sagan *et al*. 1979) and the total albedo difference ΔR_{tot}:

$$\begin{aligned} R_{tot} &= R_c + \frac{R_s(1-R_c)^2}{1-R_cR_s} \Rightarrow \Delta R_{tot} \\ &= \Delta R_c + R_s\left[\frac{\left[1-(R_c+\Delta R_c)\right]^2}{1-(R_c+\Delta R_c)R_s} - \frac{(1-R_c)^2}{1-R_cR_s}\right]. \end{aligned} \tag{5}$$

The net difference in reflected solar energy, ΔE (in Joules) caused by the cloud albedo change (which in turn is caused by the aerosol emissions from one square meter of forest) is simply:

$$\Delta E = ScA\Delta R_{tot}t \tag{6}$$

where t is the average total time an aerosol particle spends in a cloud before being removed from the atmosphere by precipitation, c is a dimensionless correction factor and S is the daily average insolation at the top of the cloud layer.

According to Pruppacher and Jaenicke (1995), the weighted global average time an air parcel spends in a cloud is 3.03 h, and the average number of evaporation-condensation cycles undergone by atmospheric water vapor before precipitation is 11. We assume, to a first approximation, that, these residence times also apply to aerosol particles, and calculate the value of t to be approximately 33 h. This approach also

avoids the problem of having to estimate average cloud coverages, as only the "effective" aerosol lifetime is used.

S was estimated for each event based on the day of the year and the insolation equations given by Hartmann (1994). The latitude was assumed to be 61.5. The values obtained for *S* are top-of-the-atmosphere insolation values. The correction factor *c*, which was varied between 0.5 and 0.7, attempts to account for the effects of clouds above the volume into which the aerosols are transported. Note that ΔE is in fact independent of the spread area *A* because $\Delta R_c \propto 1/A$. The average energy budget perturbation caused by each event (*F*) is then simply

$$F = \Delta E/(189\ 216\ 000\ \text{s}), \qquad (7)$$

where ΔE has been divided by the number of seconds in a year, multiplied by the number of years in the study. The total energy budget perturbation caused by aerosol emissions from the forest area is the sum of the individual "event" perturbations. This value represents the annually averaged perturbation on the Earth's energy budget caused by the aerosol emissions from one square meter of forest. (The dimension of *F* is Watts, the m^{-2}-dependence is added to reflect the fact that the perturbation has been calculated per square meter of emitting forest.)

Estimation of radiative forcing due to carbon sequestration

Based on the work by Myhre *et al*. (1998), Betts (2000) presented two equations that relate the global radiative forcing, F_{global}, to a terrestrial carbon stock change ΔC, taking into account the nonlinearity of the concentration-forcing relation and also the airborne fraction. Combining Betts' two equations, and using the approximation $\ln(1 + x) \approx x$ when $x << 1$, we obtain:

$$\begin{aligned} F_{\text{global}} &= 5.35\ \text{W m}^{-2} \times \ln\,(1 + \Delta C/2C_0) \\ &\approx 5.35\ \text{W m}^{-2} \times \Delta C/2C_0 \\ &\quad (\text{when } \Delta C/2C_0 << 1), \end{aligned} \qquad (8)$$

where C_0 is the atmosphere's current carbon content, about 730 Pg. (The factor of 2 in the denominator represents the airborne fraction, i.e. the fact that only around 50% of carbon emissions remain in the atmosphere.) Based on Myhre *et al*. (1998), the error margin of the equation was assumed to be ±1%. It will be seen later that this is insignificant as compared with other error sources. When ΔC is replaced by the measured annual carbon balance value $\Delta C/A$ (in g m^{-2} given as carbon), we obtain a value for the global radiative forcing caused by forest carbon sequestration per square meter of forest (with dimensions of W m^{-4}; watts per square meter per square meter of forest). In order to determine the total perturbation to the Earth's energy budget caused by this sequestration (the "radiative forcing per square meter of forest" as defined in this study), we must multiply the result by the Earth's surface area (5.1×10^{14} m^2), and obtain:

$$\begin{aligned} F = \Delta C \times 0.00187\ &\text{W m}^{-2}\ \text{g}^{-1} \\ &(\Delta C/2C_0 << 1). \end{aligned} \qquad (9)$$

In order to compare the radiative forcing caused by carbon dioxide sequestration with the forcing caused by aerosol emissions, we must also account for the different time-scales. Equation 7 represents the annual mean radiation energy budget perturbation caused by the aerosol emissions. Equation 9, on the other hand, represents the radiative forcing the carbon dioxide sequestered by the forest in one year would have exerted for as long as it remained in the atmosphere. In order to compare the results of the two equations with each other, the value obtained from Eq. 9 should thus be multiplied by the atmospheric lifetime of carbon dioxide — a variable that is not precisely known or even defined. Based on Boucher *et al*. (2001), we used a maximum value of 200 years, a minimum value of 5 years and a best-guess value of 100 years. Whether or not future energy budget perturbations should, for political, economical or environmental reasons, be discounted in comparison with current perturbations, is outside the scope of this study.

Disregarding the slight future decrease in the radiative forcing per carbon dioxide molecule, and the fact that the IPCC definition of radiative forcing assumes that stratospheric temperatures reach equilibrium, the value obtained by multiplying the result of Eq. 9 by the atmospheric

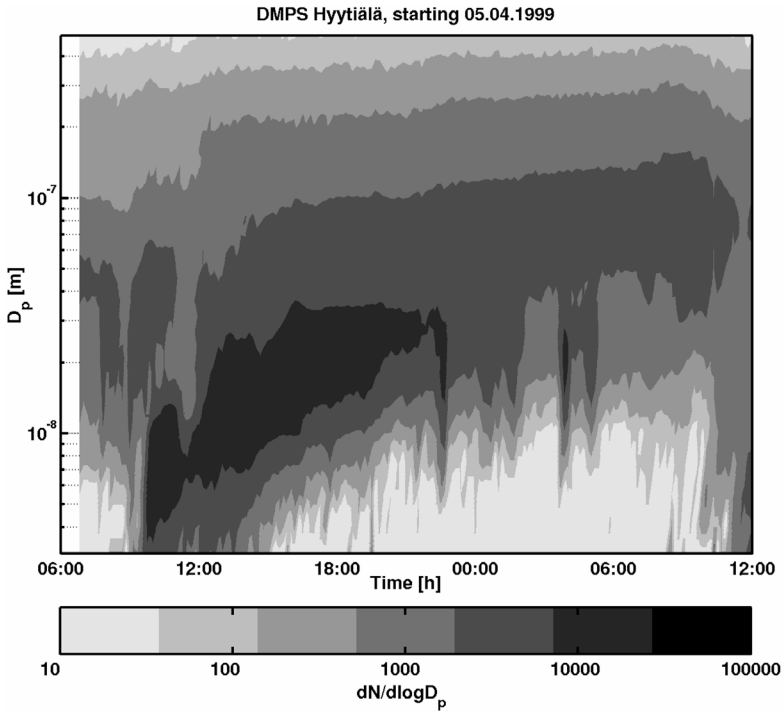

Fig. 1. An aerosol formation event and subsequent growth at Hyytiälä (5–6 Apr. 1999).

lifetime of carbon dioxide can be compared with the aerosol-forcing estimates obtained in the previous section. It should, however, be noted that Betts (2000) referred to aboveground carbon stocks, while the fluxes measured at the SMEAR II station represent the net ecosystem exchange, which includes both aboveground and soil carbon stock changes.

Results and discussion

Formation and growth of atmospheric aerosols

The aerosol number size distribution measurements show clearly detectable aerosol particle formation events at approximately 50 days per year (*see* Mäkelä *et al*. 2000). The typical weather conditions during particle formation events are sunny and half cloudy with wind coming from west to north (Kulmala *et al*. 2001). The most typical time for these events is March–April. Subsequent to the new particle formation, significant particle growth is usually observed. Almost 20% of the events continue sufficiently long to produce particles with diameters larger than 80 nm. Particles of this size can become effective cloud condensation nuclei (CCN), especially if they are hygroscopic.

A typical example of a clearly detectable aerosol particle formation event with subsequent growth of long duration is presented in Fig. 1. The growth can clearly be seen. The aerosol number concentration of particles with diameters greater than 40, 60, 80 and 100 nm is presented in Fig. 2. The event affects, after a delay, all number concentrations. When air masses are not changing, it is justified to use 40, 60 or 80 nm as the CCN-threshold diameter. However, for the threshold of 40 nm, the risk of the influence of different air masses is minimized.

Carbon fluxes

Based on the eddy covariance measurements of CO_2, the net ecosystem exchange (NEE) is obtained for each half-hour period. However, due to the calibration and maintenance of the instruments and the lack of turbulence, which especially concerns stable low-windy summer nights, the data coverage is not complete. The filling in of the gaps in the flux record was carried

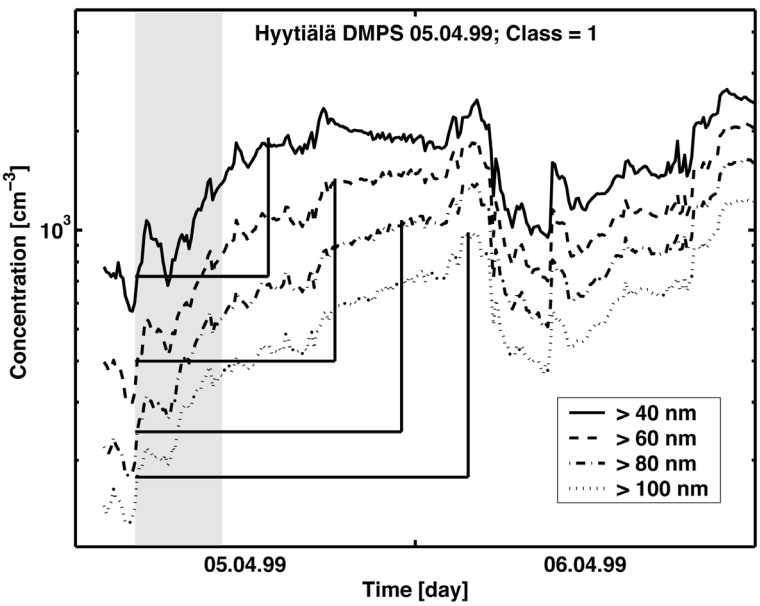

Fig. 2. Threshold number concentration of particles with diameters greater than 40 nm, 60 nm, 80 nm, and 100 nm measured at Hyytiälä (5–6 Apr.1999).

out using regressions fitted by means of air and soil temperature and photosynthetic photon flux density (PPFD) data (*see* Markkanen *et al*. 2001, Suni *et al*. 2003). In winter, the forest is a small source of carbon resulting from the soil respiration, and during growing seasons the forest acts as carbon sink. For the period 1997–2001, the annual carbon balances varied from –240 to –160 g m^{-2} given as carbon, where the negative sign indicates the net sink. The average carbon balance value, ΔC, was –194 g m^{-2}, with a standard deviation of 38 g m^{-2}, both given as carbon.

Radiative forcing

Unless otherwise stated, all forcing values should be interpreted to mean watts per square meter of forest.

Radiative forcing due to aerosol emissions

Table 1 presents the values of the parameters used to obtain maximum, minimum and "best-guess" estimates for the radiative forcing per square meter of forest caused by aerosol emissions. Table 2 presents the estimated values of radiative forcing per square meter of forest for three different CCN-activation thresholds and three different parameter sets. The smaller threshold diameter we use the bigger is the estimated value for radiative forcing. The best-guess forcing decreases from –13.8 to –5.4 W m^{-2}

Table 1. Parameters used to obtain minimum, maximum and "best-guess" estimates.

	Minimum case	Maximum case	"Best guess"
Correction factor (c)	0.5	0.7	0.6
Cloud albedo (R_c)	0.7	0.4	0.5
Surface albedo (R_s)	0.4	0.1	0.3200–0.1083*
CCN density (N)	600 cm^{-3}	100 cm^{-3}	200 cm^{-3}
Effective lifetime (t)	33 hours	33 hours	33 hours

* the surface albedo range in the "best guess" case was based on average monthly albedoes measured in Hyytiälä. The albedo was calculated as the ratio of reflected global radiation and global radiation. The measurements were made using two Redemann TP3 pyranometers [Astrodata, Estonia] operating in the 300–4800 nm wavelength range.

when changing the threshold diameter from 40 to 80 nm. On the other hand, for the threshold diameter of 40 nm, the estimated minimum forcing (–2.6 W m^{-2}) is almost 30 W m^{-2} larger in magnitude than the estimated maximum forcing.

Radiative forcing due to carbon sequestration and albedo change

By inserting the carbon balance value of –194 ± 38 g m^{-2} given as carbon into Eq. 9, and by taking into account the 1% error margin of the equation itself, we obtain a value of –0.36 ± 0.07 W m^{-2} for the radiative forcing per square meter of forest caused by carbon sequestration at the SMEAR II station. Multiplying this value by the assumed CO_2 lifetime of 5–200 years leads to a range from –1.45 to –86 W m^{-2}, with –36 W m^{-2} being the best-guess value. This is the theoretical mean energy budget perturbation that the annual CO_2 sequestration would exert, if all the perturbation were to occur within one year.

Using a rather extreme GCM-method, Betts (2000) calculated the increase in radiative forcing caused by boreal forestation-related albedo change to be from +3 to +20 W m^{-2}. This value was originally reported as carbon stock equivalents, and was converted into a radiative forcing per square meter of forest using Eq. 8. By using the "bottom-up" method employed in the previous sections, the forcing caused by the albedo change associated with cropland-to-forest conversion in southern Finland was estimated to be on the order of magnitude of +10 W m^{-2}, in good agreement with Betts's estimate. It should be noted that while the sequestration-related forcing corresponds to the annual carbon-stock change, the albedo-related forcing corresponds to the difference between open fields and forest with 100% canopy coverage. As cropland-to-forest conversion does not happen within one year, the forcing values cannot be directly compared. On the other hand, while sequestration ceases or at least slows down after the forest reaches maturity, albedo changes are permanent in the sense that forests absorb more light than cropland for as long as they are allowed to remain standing. Whether or not, and at which rate, forests continue to emit biogenic aerosols after reaching maturity, is a matter beyond the scope of this study.

Global forests

By using a combination of remote sensing and national inventories, the FAO Forest Resource Assessment 2000 report (FAO 2001) arrived at a total global forest area of 3869 ± 116 million ha, of which 187 million ha consists of forest plantations. The forest area measurement is rather sensitive to the parameters used. In boreal areas, the crown cover parameter (10%) used by FAO can be criticised, as it leads to the inclusion of areas "most non-specialists would consider to be tundra" (Matthews 2001). Whether or not these tundra-like "forest" areas emit biogenic aerosols, or sequester significant amounts of carbon, is beyond the scope of this study. On the other hand, Canadian forest resources may have been underestimated in the FRA 2000 due to national inventories that define only economically productive wooded areas as "forest" (Matthews 2001).

The subdivision of the total forest area by type is presented in Table 3. Explicit error margins for forest areas by type are not given, but can be assumed to be somewhat less than 3% for boreal and temperate forests. The total area of boreal forests presented in Table 3 (1748 million ha) corresponds to about 3.4% of the Earth's surface area. Multiplying the aerosol-related forcing values by 0.034 gives an extremely rough estimate for

Table 2. Total biogenic aerosol forcing estimates per square meter of boreal forest.

CCN activation threshold	Minimum case	Maximum case	"Best guess"
40 nm	–2.6 W m^{-2}	–31.7 W m^{-2}	–13.8 W m^{-2}
60 nm	–1.6 W m^{-2}	–19.5 W m^{-2}	– 8.4 W m^{-2}
80 nm	–1.0 W m^{-2}	–12.3 W m^{-2}	– 5.4 W m^{-2}

the annually averaged global radiative forcing caused by boreal aerosol emissions: between –0.03 W m^{-2} and –1.1 W m^{-2}.

The Forest Resource Assessment report estimated, in general terms, that 50% of the land area of the world was covered by forests in prehistoric times as opposed to 30% today. The World Resource Institute Pilot Assessment of Global Ecosystems (Matthews *et al.* 2000) report is more careful, and estimates that "one fifth to one half of the world's forest cover has been converted to other uses since pre-agricultural times." If the difference between forest and non-forest aerosol emissions are significant for both boreal and non-boreal forest types, our maximum-case values would seem to indicate that anthropogenic deforestation may, through changes in aerosol emissions and cloud albedo values, have influenced the climate to an extent comparable to all anthropogenic greenhouse gas emissions.

Conclusions

Processes within boreal forests were seen to have significant impacts on the radiation energy budget. Biogenic aerosol formation causes, by increasing the cloud albedo, an annual-average radiative cooling of –1.0 to –31.7 W m^{-2}, with a best-guess value of –8.4 W m^{-2} (watts per square meter of forest). Comparable values are between –1.45 and –86 W m^{-2}, with a best-guess value of –36 W m^{-2}, for the cooling caused by forest carbon sequestration and around +10 W m^{-2} for the warming caused by the surface albedo change following cropland-to-forest conversion.

As a very rough estimate, the annually averaged global radiative forcing caused by boreal aerosol emissions was calculated to be between –0.03 and –1.1 W m^{-2}. If the upper limiting value is applicable also to non-boreal forests, the aerosol-related climate effects of deforestation may be globally significant.

Table 3. Forest areas by type, FRA 2000.

Tropical	47%	1818×10^6 ha
Boreal	33%	1748×10^6 ha
Temperate	11%	426×10^6 ha
Subtropical	9%	348×10^6 ha

The tentative global forcing values presented above might, however, be unrealistically large in at least three different ways. First, the use of the linear form of Twomey's equation is not permissible when the change in CCN concentrations is not small as compared with the original CCN concentration. The albedo change — and thus the radiative forcing — per an additional CCN particle is greatest when the number of added CCN is small. Presumably, a large part of the CCN present in the atmosphere in regions close to boreal forests is produced by the forests in the first place. Also, new-particle formation may well occur simultaneously over large areas of the forest, causing large changes in CCN concentrations. In multiplying the forcing per square meter of forest with the global boreal forest area fraction, we have thus used Twomey's equation incorrectly, and exaggerated the global effect of boreal aerosol emissions. Second, the assumption that vertical aerosol fluxes can be disregarded is optimistic. If, for example, vertical fluxes were to spread the newly formed particles into a volume of height 2000 m, instead of 1000 m, our forcing value would be halved. On the other hand, new particle formation might also occur above the 1000 m, increasing the number of aerosols formed per unit area. Third, in reporting the above forcing values as "climate effects of boreal forests", the implicit assumption was made that other land types emit no aerosols. If, hypothetically, other land types were to emit similar numbers of aerosols as boreal forests, the real aerosol-forcing effect of the forest would be zero, in the sense that no aerosol-related climate changes would be observed if the forest area changed.

In order to accurately estimate the regional radiative forcing caused by emissions from a large forest area, we would first need to know the vertical CCN concentration profile and the corresponding cloud albedo values for different cloud layers, both in the absence of a forest. Next, we would need to measure the (three-dimensional, time-dependent) flux of CCN-forming aerosols from the forest area into the region in question, and calculate the new cloud albedo values by solving the differential form of Twomey's equation. With the help of (season-dependent) surface albedo and cloudiness values we could then obtain

the net change in albedo, and thus obtain the perturbation in the Earth's energy budget. Finally, we would need to calculate the sum of the regional radiative forcings for every region affected by aerosol emissions from the forest area. Dividing this by the Earth's surface area would then yield a global radiative forcing estimate for the aerosol emissions from this forest area.

In the absence of detailed three-dimensional aerosol flux data, as well as CCN and cloud-albedo profiles, the only reliable conclusion that can be drawn from the data obtained from the SMEAR II station is that the total global energy budget perturbation — per unit area of forest — caused by biogenic aerosol emissions is probably smaller than that caused by carbon sequestration. However, it has the same order of magnitude.

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