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Aerosol-radiation feedback loop based on satellite data

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The climate feedback is a response of the climate system to a perturbation through a number of mechanisms. Perturbations can be due to natural factors, like volcanic activity or changes in solar activity, or anthropogenic such as emissions of long-lived greenhouse gases and aerosol particles.

Atmospheric aerosols affect the Earth’s radiation budget. The aerosols impact radiation directly by scattering and absorbing incoming solar radiation and indirectly by changing cloud properties via formation of cloud condensation nuclei.

Here, the aerosol radiation feedback loop associated to the continental biosphere-aerosol-cloud-climate (COBACC) feedback loop, is suggested. This negative feedback loop connects increasing atmospheric CO₂ concentration, rising temperatures, the formation of aerosol particles due to the emission of biogenic volatile organic compounds, changes in ratio of diffuse to global radiation in the clear sky condition, and changes in the plant gross primary production.

In this study, in-situ atmospheric measurement data in Hyytiälä station, as well as satellite atmospheric measurement data (CERES (Clouds and the Earth’s Radiant Energy System) and MODIS (Moderate Resolution Imaging Spectroradiometer instrument)) around Hyytiälä station and a small area in the western plain of Siberia for clear sky conditions in June and July around noon, were used.

Three methods for detecting clear-sky conditions were considered: brightness parameter, global radiation smoothing and lastly MODIS cloud mask method. Here, MODIS cloud mask method was selected as the most suitable method due to availability of data and global coverage.

This study proved partly the existence of the aerosol radiation feedback loop by finding positive correlation between some of the components of the feedback loop, such as condensation sink(CS) concentration, rising temperatures, the formation of aerosol particles due to the emission of biogenic volatile organic compounds, changes in ratio of diffuse to global radiation in the clear sky condition, and changes in the plant gross primary production.

In addition, the impact of relative humidity on the relation between R and temperature was investigated. It was found that it is important to take into account the swelling effect in order to investigate the relation between R and temperature. In contrast, solar zenith angle does not have an impact on the relation during study period (June – July).
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List of Abbreviations

$R_d$  Diffuse Radiation
$R_g$  Global Radiation
R      Ratio of diffuse radiation to global radiation
CS     Condensation Sink
RH     Relative Humidity
1. Introduction

1.1 Background

1.1.1 Climate system

Climate change has been considered an environmental issue over the past century. It is essential to investigate the causes of changes in the Earth climate system since it can be a threat for the livelihood of humans and all kinds of species on the Earth. Climate change has many impacts on the Earth such as drought, rising sea level, storms, melting of glaciers, etc. All these impacts can make significant problems for humankind and other species on the Earth.

IPCC (2013) define climate change as any change in climate over time, whether due to natural variability or as a result of human activity. Climate change can be caused by changes in the atmospheric abundance of greenhouse gases and aerosols, in solar radiation and in land surface properties which cause an imbalance in the radiative energy of the climate system. During these changes in the climate called radiative forcing, the amount of incoming solar radiative energy is not equal to the amount of outgoing radiative energy (Boucher, 2015). Radiative forcing is measured at the top of atmosphere with unit of $Wm^{-2}$. Positive radiative forcing refers to the estimation of the gain of energy, which leads to the warming of the system. On the other hand, negative radiative forcing is an estimation of the loss of energy that causes cooling of the system.
Several studies have revealed that an increase of greenhouse gas emissions such as carbon dioxide and methane enhance greenhouse effect, which causes warming effect, and high atmospheric aerosol concentration can have a cooling effect on Earth’s climate (Kulmala et al., 2001; IPCC, 2013; Kulmala et al., 2014b). However, due to a lack of information about temporal distribution of aerosol and clouds, there is uncertainty associated with the influence of aerosol particles on the climate (Houghton, 1996; Kulmala et al., 2001; Kulmala et al., 2014b).

Climate feedback is a response of the climate system to a perturbation through a number of mechanisms (Boucher, 2015). A positive feedback loop occurs when the original perturbation is amplified, whereas a negative feedback loop happens when the original disturbance is dampened. The term climate feedback loop refers to the process in which the initial perturbation in climate-related quantities, such as greenhouse gases emissions from human activities, can be amplified or dampened through changes in the components of the climate feedback loop (Kulmala et al., 2014b; Boucher, 2015). Components of the climate system are surface (biosphere), atmosphere, ocean (hydrosphere) and cryosphere. There is a close interaction between these components of the climate system. Therefore a change in one of them can impact other components via complicated linked processes, feedbacks, and interactions (Kulmala et al., 2013).

### 1.1.2 Aerosols

Atmospheric aerosol particles refer to solid or liquid particles suspended in the air (Hinds, 2012). Atmospheric aerosols affect climate by impacting on the Earth’s radiation budget. They impact radiation directly by scattering and absorbing incoming solar radiation and indirectly by changing cloud properties via formation of cloud condensation nuclei (CCN). Furthermore, aerosol particles impact air quality by decreasing visibility and putting human health in risk (Stieb, Judek, and Burnett, 2002; Akimoto, 2003; Qi et al.,
All these impacts depend on the size of aerosol particles. The size of the aerosol particle diameter varies from some nanometres to hundreds of micrometres. Aerosols are divided into different modes based on their diameter sizes (dp), namely nucleation mode (dp < 25 nm), Aitken mode (25 nm < dp < 90 nm), accumulation mode (90 nm < dp < 1 µm), and coarse mode (dp > 1 µm) (Sundström, Nousiainen, and Petäjä, 2009). The Aitken and accumulation modes are also called fine mode with diameter sizes from a hundred nanometers to less than 1 µm (Boucher, 2015). There is evidence that particulate matter, especially fine mode particles cause adverse human health effects (Dockery et al., 1993; Kim, Kabir, and Kabir, 2015).

Particles can impact Earth’s radiations budget, in other words they can scatter solar radiation when their size is bigger than 100 nm in diameter or some of them like black carbon can absorb solar radiation and transform it into heat. (Boucher, 2015; Sundström et al., 2015a).

Under certain super saturation conditions, aerosol particles can form cloud droplets through condensation of water vapour on their surface, which depends on the chemical properties and size of the aerosol particles (Andreae and Rosenfeld, 2008). When particles are bigger than 50 nm in diameter, they can act as cloud condensation nuclei (CCN). An increase in aerosol concentration generally causes concentration of CCN to rise as well as concentration of cloud droplets, which leads to an increase of cloud albedo. This is called the aerosol indirect effect.

Another way to classify aerosol particles is based on the source of their origin. Aerosol can originate from natural sources (like emission from the vegetation, volcanoes and ocean), and anthropogenic sources (like industrial emissions and fossil fuel combustion) (Boucher, 2015).

Aerosols can also be classified according to their formation process into two groups: primary aerosols and secondary aerosols (Tomasi, Fuzzi,
and Kokhanovsky, 2017). Primary aerosols are particles emitted directly to atmosphere (Boucher, 2015). Primary aerosol can be emitted from natural source such as emission of sea spray, release of mineral dust, emission of biomass burning, and volcanic dust and anthropogenic sources like fuel combustion, transport activities, and industrial process (Tomasi, Fuzzi, and Kokhanovsky, 2017). Secondary aerosols are particles formed via gas to particle conversion in which condensable vapours lead either to growth of pre-existing particles by condensation processes or nucleation of new particles (Tomasi, Fuzzi, and Kokhanovsky, 2017). The latter one can be called new particle formation (NPF) (Weber et al., 1997).

Volatile organic compounds play an important role in the formation of secondary organic aerosol (SOA), which are formed through gas to particle conversion of organic compounds (Ehn et al., 2014). Volatile organic compounds are emitted to atmosphere from anthropogenic and biogenic sources (Guenther et al., 1995; Atkinson, 2000; Guenther et al., 2000; Ehn et al., 2014). Terrestrial vegetation is the main source of biogenic volatile organic compounds (BVOC) like isoprenes, monoterpenes, sesquiterpenes, as well as oxygen-containing compounds, such as alcohols or ketones (Hoffmann et al., 1997; Ehn et al., 2014; Boucher, 2015). There are different kind of BVOCs depending on plant species (Niinemets et al., 2010). After the BVOCs are oxidized by e.g. OH, O$_3$, and NO$_3$, the resulting low volatility compounds can participate in aerosol formation (SOA formation) and can condense on existing aerosol particles (Schallhart et al., 2016).

As mentioned above, secondary aerosols can be formed due to nucleation during the gas to particle conversion process. Nucleation of new particles or new particle formation (NPF) processes is the formation of nanometer-size clusters, which further grow via condensation of available vapours to reach detectable sizes (Kulmala and Kerminen, 2008; Lushnikov, Zagaynov, and Lyubovtseva, 2010). However, all newly formed particles do not grow to
larger sizes. The growth of newly formed particles depends on the competition between the processes of condensational growth of clusters and scavenging by pre-existing particles (Riipinen et al., 2011; Kulmala et al., 2014a; Kulmala et al., 2017).

1.1.3 Radiation

The climate system is in equilibrium when the amount of solar radiation absorbed by the system and terrestrial radiation emitted to space is equal. Incoming shortwave radiation comes from the sun and includes short wavelength such as ultraviolet, visible, and near-infrared (Boucher, 2015). Terrestrial radiation is also called longwave as it includes longer wavelength (infrared), and it is emitted from the earth’s surface and atmosphere (Boucher, 2015). The cooling of the Earth occurs when emitted terrestrial radiation escapes to space and it exceeds the incoming radiation. The process by which aerosol particles interact with solar radiation and deflect solar radiation into different direction anisotropically is called scattering (Boucher, 2015). There are a number of components that affect the amount of solar radiation scattered by aerosol particles, namely solar zenith angle, altitude, the surface properties, relative humidity and clouds.

The term solar zenith angle refers to the angle between solar beam and perpendicular line on horizontal surface (Fatemi and Kuh, 2013). Changes in solar zenith angle affect solar radiation, which reaches surface. Solar zenith angle is larger during winter noon time than summer noon time. Summer solstice is when solar zenith angle is smallest and solar radiation, which reaches surface is highest during a year. When the solar zenith angle is large, sunlight has to pass through a thicker layer of the atmosphere before it reaches the surface, therefore less radiation arrives to the surface due to scattering and absorption of radiation.

It has been shown that many aerosol components are hygroscopic,
and aerosols take up water and grow to bigger sizes (swelling effect) as relative humidity (RH) increases (Cheng et al., 2008). Consequently, optical properties of aerosol particles are affected by relative humidity (Zieger et al., 2013). The influence of RH on aerosol scattering depends on chemical composition and size of aerosol particle since these two variables determine scattering property and the ability of taking up water of aerosol particles.

In absence of clouds, aerosol particles can scatter solar radiation more effectively. Aerosol particles can still scatter solar radiation when there are thin clouds in atmosphere or they are above clouds (Boucher, 2015). However, aerosols such as biomass burning generated carbonaceous particles located above clouds can effectively absorb solar radiation and may have a warming effect (Torres, Jethva, and Bhartia, 2012; Jethva, Torrres, and Ahn, 2016).

Henning, Motta, and Mugnier (2013) explain global solar radiation includes two components, namely direct solar irradiance and diffuse solar irradiance. Direct solar irradiance refers to the amount of solar radiation arriving to an area unit of surface. This component can be expressed as normal direct solar irradiance (or normal beam) and horizontal solar irradiance (or horizontal beam). This classification is based on the geometrical relation between the sun, earth and the surface. Henning, Motta, and Mugnier (2013) define normal direct solar irradiance as the amount of received solar radiation by a surface, which is perpendicular to the solar beam whereas horizontal solar irradiance is the amount of received solar radiation by a surface, which is horizontal to the ground. Henning, Motta, and Mugnier (2013) define diffuse solar irradiance as the amount of solar radiation received by the area unit of surface not directly from sun. This component arrives to the surface from all possible directions except the solar beam direction. Diffuse solar irradiance is produced when incoming solar radiation is either scattered by atmospheric components or reflected by any surface surrounding the receiving surface. The first case is called diffuse sky solar irradiance and the latter one is called diffuse ground solar irradiance. Henning, Motta, and Mugnier
(2013) state most of the measured data of diffuse solar irradiance is diffuse sky solar irradiance received by a horizontal surface. Global solar irradiance is defined by Henning, Motta, and Mugnier (2013) as the sum of the diffuse solar irradiance and direct solar irradiance:

\[ G_T = G_{\text{diff}} + G_{\text{dir}} \]

where \( G_T \) is global solar irradiance, \( G_{\text{diff}} \) is diffuse solar irradiance, and \( G_{\text{dir}} \) is direct solar irradiance.

### 1.1.4 Feedback mechanisms

There are a number of feedback loops in the climate system. Here we focus on two feedback loops associated with continental biosphere-aerosol-cloud-climate (COBACC) feedback (Figure 1.1). These two feedback loops are related to forest-atmosphere climate interactions proposed by Kulmala et al. (2004) and Kulmala et al. (2013) where the influence of \( \text{CO}_2 \) and aerosol particles on climate were investigated.

COBACC feedback mechanism is initiated by the increase in \( \text{CO}_2 \) (Kulmala et al., 2013). The T-loop (upper loop in Figure 1.1) starts from a rise in temperature when \( \text{CO}_2 \) concentration increases. Hence, terrestrial vegetation emits more BVOC to atmosphere, which leads to enhanced secondary aerosol formation and an increase in CCN concentration. As a result, T-loop is closed with the decrease of temperature due to higher cloud albedo through higher cloud droplet number concentrations (CDNC) and the indirect effect of aerosol on climate. The T-loop counts as a negative feedback (Paasonen et al., 2013).
The difference between the net ecosystem exchange of $CO_2$ (NEE) and total ecosystem respiration (TER) present the plant gross primary production (GPP). Photosynthesis measured by GPP happens during the growing season and in sunlight in the boreal forest zone (Hari and Mäkelä, 2003). The term photosynthesis refers to the process in which plants and certain
organisms take in CO$_2$ and using solar radiation and water to produce oxygen and energy-rich organic compounds. Kulmala et al. (2004) and Kulmala et al. (2013) state there is a negative feedback between increase of CO$_2$ concentration and plant growth. They explain that increasing CO$_2$ concentration enhances photosynthesis, which in turn limits the increase of CO$_2$ concentration.

The GPP-loop (lower loop in Figure 1.1) illustrates a BVOC and SOA concentrations increase due to enhanced GPP. Condensation sink (CS) is used in this feedback loop as a measure of SOA concentration (Kulmala et al., 2014b). As a result of higher SOA concentration, scattering from the surface of aerosols increases and consequently diffuse radiation rises. Many studies have shown that plants use diffuse radiation more efficiently than direct radiation (Cohan et al., 2002; Mercado et al., 2009; Li et al., 2014). Diffuse radiation is distributed more uniformly in the canopy due to entering the canopy from different directions. This causes leaves to absorb light evenly and fewer leaves will be light-saturated. Hence, the leaves use photons efficiently for photosynthesis. Increase of diffuse radiation leads to plants becoming more photosynthetically active. Therefore, increase of the ratio between diffuse radiation and global radiation (R) due to rise of diffuse radiation will increase GPP. Hence, an increase of GPP closes the positive GPP-loop.

The relation between all the steps of the COBACC feedback loop have been studied in prior literature. For instance, it has been shown there is a temperature dependency for BVOC emissions, which have an important role in the formation of SOA (Kulmala et al., 2004). It was shown by Kulmala et al. (2013), there is a positive correlation between monoterpane concentrations, which are known precursors for condensing organic vapours, and CS in Hyytiälä station. Their findings can explain there is possibly a positive correlation between SOA and CS. Furthermore, Kulmala et al. (2013) found positive correlation between R and CS as well as R and GPP in Hyytiälä station.
1.2 Satellite-based atmospheric research

Utilizing satellite data to study the climate system is essential because they provide comprehensive global observations. Although in-situ measurements are more accurate about aerosol physical, optical and chemical properties as well as local concentrations, they are limited to a small area (Sundström et al., 2012).

Satellite data observe spatial distribution of aerosol particles with sufficient resolution on global scale as well as to distinguish fine particles from coarse particles (Kaufman, Tanré, and Boucher, 2002; Sundström et al., 2012). This latter ability is useful to separate natural aerosol particles from anthropogenic particles (Kaufman, Tanré, and Boucher, 2002). The benefit of this ability is to gain information about global aerosol budget for assessing contribution of anthropogenic aerosol particles in the aerosol budget and to the aerosol radiative forcing of the climate (Kaufman et al., 1997). Furthermore, due to the short lifetime of aerosol particles and their strong spatial variation, utilizing satellite data to estimate global aerosol budget trends is beneficial.

Satellite observations have been used in several studies related to the climate aerosol interactions. For example, Boucher and Tanré (2000) estimated the global mean shortwave radiative flux change due to aerosols from the POLDER (POLarization and Directionality of the Earth’s Reflectance) satellite retrievals, (Sundström et al., 2015b) determined the aerosol direct radiative effect by using CERES (Clouds and the Earth’s Radiant Energy System) and MODIS (Moderate Resolution Imaging Spectroradiometer instrument) over China, and Kaufman et al. (2005) used MODIS measurements to estimate the anthropogenic component of the aerosol optical thickness over oceans.
Satellite observations can also be useful for cloud screening and providing information about clouds. This information is important for studies related to aerosols to detect cloud-free conditions. Many research papers have used this ability of satellite instruments to detect cloud-free conditions including Martins et al. (2002), Bellouin et al. (2003), and Kaufman et al. (2005).

1.3 Aims

This thesis analyses the aerosol radiation feedback loop (Figure 1.2) associated with the lower part of the COBACC feedback loop using satellite data. Whilst many steps remain the same, some are different. The aerosol-radiation feedback loop starts with linking CO$_2$ concentrations to temperature, whereas the lower part of the COBACC feedback loop starts with connecting CO$_2$ concentrations to GPP and BVOCs production. Here, I show that the lower part of the COBACC feedback loop still exists, even without the initials steps taking place.

The hypothesis in this study is that increasing CO$_2$ causes a rise of temperature, which leads to enhanced SOA (secondary organic aerosol) formation due to BVOC oxidation products. Then, increase of SOA formation causes number of aerosol particles to increase, leading to a larger surface area, which scatters radiation more efficiently. As a result, the ratio of diffuse radiation to global radiation will increase due to enhanced scattering by aerosol particles. Increase of diffuse radiation leads to plants becoming more photosynthetically active. Hence, increase of photosynthesis closes the loop via negative feedback to the carbon sink.

In order to confirm the hypothesis of the connection between the components of this feedback loop utilizing satellite data, I started by investigating the relation between temperature and aerosol surface area, as well
Next, I analysed the relation between temperature and R. It was possible to temporarily ignore the relation between different components in the feedback loop, as they have been investigated and quantified in prior literature using in-situ data (More details at the end of section 1.1.4). Following the investigation of the relation between R and temperature, dependency of this relation from solar zenith angle and relative humidity was studied.
2. Materials and Methods

2.1 Observation sites

Here, two boreal forest environments were studied with ground-based and satellite-based data, specifically SMEAR (Station for Measurement of Ecosystem-Atmosphere Relations) II measurement station located in Hyytiälä, southern Finland (Hari et al., 2013) (61°51-62°5 N, 23°5-24°5 E, UTC+2) and a region in Siberia (61°51-62°5 N, 100°5-101°5 E, UTC+7). The term boreal forest refers to coniferous forests located in the northern hemisphere region (Hari and Kulmala, 2008). Boreal forests are an important source of different BVOCs specially monoterpenes (Spracklen, Bonn, and Carslaw, 2008).

The ground-based data that I used in our analysis was obtained from SMEAR II measurement station (called here Hyytiälä station). A boreal forest surrounds this station, and major pollution sources are located far away. Long-term measurements of aerosol particle number size distribution, aerosol concentrations, and other aerosol physical and chemical properties are available in this station as well as meteorological trace gas quantities and ecosystem measures.

The satellite data was used for both Hyytiälä station and west Siberian plain.
2.2 Materials

2.2.1 In-situ data

I used several quantities obtained from ground-based measurements including, global radiation, diffuse radiation, temperature, aerosol size distribution to calculate condensation sink (CS), brightness parameter, and relative humidity (RH).

To find clear sky conditions, I used brightness parameter (P) data calculated according to Kulmala et al. (2010). This parameter refers to the ratio of the integral of observed global radiation ($R_{\text{obs}}$) to the integral of the theoretical radiation ($R_{\text{theor}}$) during daytime radiation (from sunrise to 16 in afternoon) as follows

$$P = \frac{\int_{\text{sunrise}}^{16} R_{\text{obs}}}{\int_{\text{sunrise}}^{16} R_{\text{theor}}}$$

Kulmala et al. (2013) state that calculation of theoretical radiation is based on the elevation angle of the sun and altitude of the measurement site. This variable presents the maximum amount of solar radiation.

Diffuse radiation (0.30 – 4.8 µm) as well as global Radiation (0.30 – 4.8 µm) measured above canopy (18 m above ground) by Reeman TP-3 pyranometer and Middleton SK08 pyranometer, respectively, were used here. The global radiation and diffuse radiation data were available from January 1998 to December 2017 and March 2000 to March 2010, respectively. The ambient temperature data used here measured at above canopy (16.8 m above ground) by Pt100 sensor. This variable is available from January 1998 to December 2017.

Condensation sink was used as this variable is connected to total surface area of aerosol particles (Kulmala et al., 2013; Kulmala et al., 2014b). CS describes the rate at which condensable vapours are removed from the
atmosphere due to their condensation on aerosol particles (Kulmala et al., 2014b). The CS was defined by Kulmala et al., 2012 as:

\[ CS = 4\Pi D \sum \beta_m(d_p) d_p N(d_p) \]

where D is the diffusion coefficient of the condensing vapor, the \( d_p \) is diameter of a particle, and \( N(dp) \) is the particle number concentration in the size \( dp \), and \( \beta_m \) is transitional correction factor. CS was calculated from particle number size distributions from DMPS (Differential Mobility Particle Sizer) according to (Kulmala et al., 2012) and it is available from March 1996 to April 2016.

The relative humidity (RH) which refers to the ratio between the partial pressure of water vapour and saturation pressure of water vapour was also used here. RH is available from January 2000 to December 2017.

All ground-based data are available with a 1-min time resolution, but CS and RH are available with a 10-min time resolution and 5-min resolution, respectively.

### 2.2.2 Satellite-based data

I used satellite data including cloud mask from MODIS (Moderate Resolution Imaging Spectroradiometer instrument), radiation from CERES (Clouds and the Earth’s Radiant Energy System) instrument.

#### 2.2.2.1 MODIS

The MODIS is an instrument developed for the Earth Observing System (EOS) on board of Terra and Aqua satellites (Ackerman et al., 2008). This instrument provides measurement of radiance at 36 wavelength bands including visible and infrared bands (0.4-14.5 micro meter) (Ackerman et al., 2008). Cloud mask data from MODIS is a global level 2 product generated
at 1Km and 250m spatial resolutions (Ackerman et al., 1998). MODIS radiance measurement includes two visible bands at 250 m spatial resolution (bands 1 and 2), five more visible bands at 500 m resolution (bands 3 to 7), and the remaining 29 visible and infrared bands at 1000 m resolution (Li et al., 2003). The radiances that were used in MODIS cloud mask algorithm to identify whether the observed view of the Earth surface is obstructed by thick aerosols, clouds, or cloud shadows are bands 1 and 2 (620–670 nm and 841–876 nm), bands 5 and 6 (1230–1250 nm and 1628–1652 nm), bands 18-21 (931–941 nm, 915–965 nm, 3.660–3.840 \( \mu m \), 3.929–3.989 \( \mu m \)), bands 26 and 27 (1.360–1.390 \( \mu m \) and 6.535–6.895 \( \mu m \)), band 29 (8.400–8.700 \( \mu m \)), bands 31 and 32 (0.780–11.280 \( \mu m \) and 11.770–12.270 \( \mu m \)), and band 35 (13.785–14.085 \( \mu m \))(Ackerman et al., 1998; Li et al., 2003).

Clouds can be identified by higher reflectance and lower temperature than surface of Earth. MODIS cloud mask algorithm is based on the cloud-surface contrast and it applies several visible and infrared threshold tests to determine clear pixels (Ackerman et al., 1998). MODIS cloud mask have 48 bits of output for each field of view which provide information on the individual spectra test (Team et al., 1997; Frey et al., 2008). The bits 1 and 2 give combined results from all individual tests by defining clear-sky confidence level of each pixel as cloudy, probably clear, confidence clear, and high confidence clear (Team et al., 1997; Frey et al., 2008).

### 2.2.2.2 CERES

CERES refers to instruments on board of Terra, Aqua, and Suomi National Polar-orbiting Partnership (NPP) platforms. Radiance measurements in the shortwave (0.3–5.0 \( \mu m \)), infrared (8.0–12.0 \( \mu m \)) and total (8.0–200 \( \mu m \)) broadband channels is provided by CERES. In this study, the SYN1deg Ed4A (edition 4A) data, which are level 3 products of CERES processing, was used. According to Data Quality Summary (2017), SYN1deg Ed4A provides monthly,
monthly 1-hourly, daily, 3 Hourly, and hourly averaged observed top-of-atmosphere (TOA) fluxes, computed TOA fluxes and fluxes at surface and different pressure levels of 70, 200, 500, and 850 hpa. These products are available in 1x1 degree spatial resolution and information observed by Terra and Aqua CERES was combined. Computed fluxes in SYN1deg Ed4A are produced by using Langley Fu-Liou radiative transfer model along with MODIS and geostationary satellite cloud properties. SYN1deg Ed4A data are produced in four different cases, namely all-sky, clear-sky, pristine, and cloudy sky. For pristine sky condition, aerosols are removed from clear-sky computations. For cloudy sky flux cases, all-sky computation is used.

Here, I used 1 hourly clear-sky adjusted surface fluxes (Wielicki et al., 1996). According to information from CERES website, the term clear-sky refers to conditions in which the probability of clear sky at CERES footprint (20 Km nominal resolution) is 99%. The parameters of the SYN1deg Ed4A products that were used in this study included geolocation (latitude, longitude), downward clear-sky shortwave (SW) surface flux, and surface shortwave diffuse flux.

2.2.3 Re-analysis data

In this study, I used 2meter temperature and dew point temperature from ERA (European Centre for Medium-Range Weather Forecasts (ECMWF) Reanalysis)-Interim, which refers to the global atmospheric re-analysis data generated by the ECMWF ((Dee et al., 2011). This product provides a 3-hourly surface parameter including surface temperature with 80 Km horizontal resolution. The ERA-Interim is a product of global observation assimilation, for example satellite remote sensing, in situ, radio sounding, with a prior information about global atmospheric state generated by a global forecast (Dee et al., 2011; Balsamo et al., 2015). Two analyses are performed at 00:00 and 12:00 UTC to generate this product (Balsamo et al., 2015).
Apart from using re-analysis temperature data directly to study relation between R and temperature, re-analysis temperature and dew point temperature data were used to calculate RH in Siberia as follows (White, 2004),

\[
\text{RH} = 100\% \times \frac{es(T_d)}{es(T)}
\]

where \( es \) is saturation water vapour pressure expressed with Teten’s formula

\[
es(T) = a_1 \exp\left(a_3 \frac{T-T_0}{T-a_4}\right)
\]

\( a_1, a_3, \) and \( a_4 \) for saturation over water are 611.21 Pa, 17.502, and 32.19 K, respectively.

### 2.3 Methods

To begin this study, it was needed to find a method to detect clear sky conditions. The reason that I focused on clear sky conditions is to diminish the effect of clouds on radiation fluxes. The presence of clouds impacts radiation fluxes due to reflection of solar radiation to space by clouds (Stephens and Greenwald, 1991; Kulmala et al., 2014b). Furthermore, in this study it was important to find conditions in which, all parts of the sky be clear without any isolated clouds due to short-period enhancement of solar global irradiance. This phenomenon happens during partly cloudy sky when the sun is not obscured. Previous studies proposed multiple theories for the physical mechanism of this phenomenon. For example Piacentini et al., 2011 explained that enhancement occurs when direct radiation is reflected at the edge of cumulus clouds and global irradiance is enhanced.

To find clear sky conditions, I examined different methods, namely brightness parameter (P), global radiation smoothing, and MODIS cloud mask.
In the first method, I used brightness parameter (P) data calculated for Hyytiälä station. For detecting clear sky conditions, I selected time periods with 90% of data having P value greater than 0.8 between 10AM to 14PM. The reason this time window was selected is that radiation is higher during this time of day. Another benefit of investigating in this time window is that this method could easily be compared with cloud mask method. The overpass time of satellite during day time over Hyytiälä is about this time.

The second method used to detect clear sky condition was global radiation smoothing. In this method, I determined days as cloudy if absolute differences between global radiation data and a smoothed global radiation curve is too large. The threshold value for cloudy days used here is when the difference is greater than 15 W/m². For this method, I used ground-based data from Hyytiälä station. I examined the data per day and discarded the day as cloudy if the threshold value was exceeded during any part of the day.

The last method I used here was utilizing cloud mask product from MODIS. In this study, I considered just pixels with high confidence clear sky and I used cloud mask data from Terra (MYD35L2). Here, clear sky conditions were detected for an area of 1x1 degree. For this purpose, the 1Km pixels of cloud mask products were combined to obtain 1x1 degree area. Once I detected clear sky conditions for the overpass time of satellite during day time over Hyytiälä station (around 10 AM to 14 PM), I selected data points for the hour around 12AM to study the highest period of radiation during the day.

A major problem with considering the above-mentioned criteria in the cloud mask method was the low amount of data points due to low probability of a 1x1 degree area to be all clear during overpass time of satellite. The solution was to collect days for which 90% of a 1x1 degree area had clear sky conditions during over pass time of satellite. This means that 10% of the area could be cloudy or partly cloudy.
Once clear sky conditions were detected, to confirm the hypothesis of the connection between the components of aerosol-radiation feedback loop, first the relation between temperature and condensation sink (CS), as well as the relation between condensation sink and ratio of diffuse radiation to global radiation (R) with in-situ data was investigated for Hyytiälä station. Then, the relation between in-situ temperature and R was studied for Hyytiälä station.

In this study, in-situ global radiation less than 500 W/m$^2$ was discarded to assure cloudy sky conditions are not involved. As Figures 3.1 (A) to (D) show global radiation is usually around 700 to 800 W/m$^2$ during clear days in June and July around noon. In order to prevent instrumental error, I discarded ground-based R greater than 0.8 because diffuse radiation during clear days can not have such a high value to affect R by such an extent.

Next, the relation between the above-mentioned components of the feedback loop with satellite-based and re-analysis data was investigated to study whether the feedback loop can be applied globally. Finally, more investigation was done to study impacts of daily median temperature, solar zenith angle and relative humidity on the relation between R and temperature.

The relation of R with the past 24 hours median temperature was investigated. Here, 24 hours is the median temperature of 24 hours data prior to the observation temperature. The reason the past 24 hours median temperature was used here was that aerosol particle number size distribution are not immediately affected by temperature changes.

In order to quantify the swelling effect on the relation between R and temperature in Hyytiälä station, parametrization was used in this study. First, to investigate impact of RH on R, the ratio of satellite-based R to the fitted line equation ($f(T_1)$) from satellite-based R vs re-analysis temperature as
a function of RH was plotted. Next, the same was done with temperature instead of RH. The ratio of satellite-based R to the fitted line equation (f(RH1)) from plot of satellite-based R/ f(T1) vs RH was depicted against temperature. Same method was applied for Siberia.
3. Results and discussion

3.1 Comparison of clear-sky detection methods

Three methods were compared in this study, including brightness parameter (P), global radiation smoothing, and MODIS cloud mask.

Table 3.1 shows the number of clear days detected by these methods during June and July from 2000-2009. Here, clear days mean clear sky conditions detected during 10AM to 14 PM for P method and MODIS cloud mask method and a whole day as clear for global radiation smoothing. MODIS cloud mask was not available for the year 2004. In addition, Table 3.1 presents the number of clear days which all three methods detected. It is apparent from this table that MODIS cloud mask detects more clear days than the two other methods. Furthermore, there are more common clear days between the MODIS cloud mask method and the two other methods than common days between the brightness parameter method and the global radiation smoothing method.
Table 3.1: An overview of number of clear days detected by three different clear day detection methods during June and July from 2000-2009 in Hyytiälä station. N indicates number of clear days detected by each method. Common clear days between methods are indicated in the respective row and column.

<table>
<thead>
<tr>
<th>Method</th>
<th>N</th>
<th>Global radiation smoothing</th>
<th>Brightness parameter</th>
</tr>
</thead>
<tbody>
<tr>
<td>MODIS cloud mask</td>
<td>123</td>
<td>22</td>
<td>35</td>
</tr>
<tr>
<td>Brightness parameter</td>
<td>45</td>
<td>15</td>
<td></td>
</tr>
<tr>
<td>Global radiation smoothing</td>
<td>79</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Figure 3.1(A) shows global radiation and diffuse radiation in an example day for which only P method detects clear sky conditions but other methods identify it as cloudy sky condition. Figure 3.1(B) shows global radiation and diffuse radiation in an example day for which only global radiation smoothing method detects clear sky conditions but other methods identify it as cloudy sky conditions. Figure 3.1(C) and 3.1(D) show in-situ and satellite-based global radiation and diffuse radiation, respectively, for the same example day for which only cloud mask method detects clear sky conditions. Comparison of these figures shows that only the cloud mask method using satellite-based radiation data (Figure 3.1(D)) finds clear sky conditions.

A possible explanation for the smoother radiation curves with satellite-based data is that spatial resolution is for a 1x1 degree wider area for Ceres data, thus it is averaged data.

Figure 3.1(E) shows an example day that was detected as a clear sky condition with all the above-mentioned methods. As can be seen from this figure the common day detected by all three methods looks clear.
FIGURE 3.1: In-situ global radiation \( (R_g) \) and diffuse radiation \( (R_d) \) for example days for which only (A) brightness parameter method (B) global radiation smoothing method (C) cloud mask method detected clear sky conditions. (D) Satellite-based \( R_g \) and \( R_d \) for an example day for which only cloud mask method detected clear sky conditions. (E) In-situ \( R_g \) and in-situ \( R_d \) for an example day for which all methods detected as a clear sky conditions.
The brightness parameter method has two challenges. First, there is no certain threshold to define clear sky conditions with high confidence. The brightness parameter method has been used in different studies with different thresholds for clear sky conditions (Kulmala et al., 2013; Kulmala et al., 2014b). This study used the same approach, but with a different threshold. In this study, days were counted as clear days when 90% of data of a day between 10 AM to 14 PM have P value greater than 0.8.

The global radiation smoothing method has its own challenges because there is no certain threshold to detect clear days. This method allows for subjective choices. Another challenge of using this method is that there are no previous studies to confirm the accuracy of this threshold.

A possible explanation for the difference in the number of detected clear days by the satellite-based method and ground-based methods can be due to the different sizes of the area that was considered. As mentioned before, for cloud mask data, a region of 1x1 degree was considered for detecting clear sky conditions, which is a large area in comparison to the area measured by the instrument in Hyytiälä station. Furthermore, for cloud mask data, the clear sky conditions was considered if 90% of this 1x1 degree region is clear. It is possible this method finds clear sky conditions when there are some isolated clouds above Hyytiälä station. In addition, if there is small cloud cover over the station, in-situ measurements are more affected with a larger calculated ratio of diffuse radiation to global radiation (R) than the satellite derived R.

Another minor cause for the difference in the numbers of detected clear day by the satellite-based method and ground-based methods can be due to errors in the surface measurements. These are caused by calibration and cosine response errors (Long and Ackerman, 2000).

MODIS cloud mask was selected as the best method for this study due to detecting clear sky conditions globally, and data is readily available.
Furthermore, MODIS cloud mask with using satellite-based radiation identify a wider area of 1x1 degree, as a result, radiation measurements are less affected by enhanced scattering by surrounding clouds.

### 3.2 Viability of aerosol-radiation feedback loop

In this section, I will discuss the relation between components of the aerosol-radiation feedback loop with in-situ data for Hyytiälä station. These components include temperature and condensation sink (CS), condensation sink and ratio of diffuse radiation to global radiation (R), temperature and R, which are presented in panel A of Figures 3.2 to 3.4. The panel B of Figures 3.2 to 3.4 will be discussed in section 3.3 to compare in-situ results with satellite-based results.

As can be seen in Figure 3.2 (A), there is a positive strong correlation between CS and in-situ temperature. CS also correlates strongly with in-situ R (Figure 3.3 (A)). As discussed in the section 1.1.4, the rise of temperature leads to increase aerosol surface area through rise of BVOCs and consequent formation of secondary organic aerosol (SOA). As CS is linked to total surface area of aerosol particles, rise of temperature increases CS. The positive correlation between CS and R supports the hypothesis where an increase in aerosol surface area, increases diffuse radiation.
Figure 3.2: Condensation sink (CS) as a function of (A) in-situ temperature and (B) reanalysis temperature for a 1x1 degree region around Hyytiälä station during day time (closest to 12PM) in June and July from 2000 to 2015.
As Figure 3.4 (A) shows, there is a moderate positive correlation between R and temperature for ground-based data.
Figure 3.4: Comparison of (A) in-situ and (B) satellite ratio of diffuse radiation ($R_d$) to global radiation ($R_g$) as a function of temperature for a 1x1 degree region around Hyytiälä station during day time (closest to 12PM) in June and July from 2000 to 2009. The solid line shows the least-squares exponential fit to measurements points.

Taken together these results indicates that the existence of the aerosol-radiation feedback loop is confirmed for some of the main steps, namely temperature and CS, CS and R, R and temperature. The relation between GPP and R, the last step of aerosol-radiation feedback loop was investigated in previous studies like Kulmala et al. (2013).
3.3 Comparison of in-situ and satellite-based data

In order to confirm whether satellite-based data can be utilized for the aerosol-radiation feedback loop, I compared my obtained results for in-situ data with satellite-based data.

The results of the correlational analysis between ground-based data and satellite-based data are shown in Table 3.2 and in the Appendix, Figures A.1 to A.3. As can be seen from Table 3.2, there is a moderate positive correlation between ground-based and satellite-based diffuse radiation from 2000 to 2009. In addition, Table 3.2 shows a strong positive correlation between ground-based and satellite-based global radiation from 2000 to 2016 and between ground-based temperature and re-analysis temperature from 2000 to 2017. The results, as shown in Table 3.2, indicate that satellite-based global radiation and re-analysis temperature can be used interchangeably with ground-based global radiation and temperature. Due to the low correlation in June and July, care should be taken when using diffuse radiation from satellite-based. The differences are probably due to the fact satellite based diffuse radiation is a calculated function for the surface derived from multiple measured variables, whereas in-situ diffuse radiation is directly from an instrument on the surface.

As can be seen in Figures 3.2(A) and 3.2(B), the in-situ temperature behaves similarly as re-analysis temperature when it is plotted against CS. Furthermore in-situ radiation and satellite-based radiation exhibit a similar pattern against CS ( Figures 3.3(A) and 3.3(B) ).

Table 3.3 shows an overview of the results for the relation between ratio of diffuse radiation to global radiation (R) and temperature with different data sources. As can be seen from Table 3.3, there is a positive correlation
### Table 3.2: Linear correlational analysis of relation between ground-based data and satellite-based data in Hyytiälä station.

Here, $R_d$, $R_g$, and $R$ represent diffuse radiation, global radiation, and ratio of diffuse radiation to global radiation, respectively.

<table>
<thead>
<tr>
<th>Data source</th>
<th>$r$ (Whole year)</th>
<th>$P$ (Whole year)</th>
<th>$r$ (June and July)</th>
<th>$P$ (June and July)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ground-based $R_d$ vs satellite-based $R_d$ (2000-2009)</td>
<td>0.50</td>
<td>$6.31 \times 10^{-15}$</td>
<td>0.35</td>
<td>$3.72 \times 10^{-3}$</td>
</tr>
<tr>
<td>Ground-based $R_g$ vs satellite-based $R_g$ (2000-2016)</td>
<td>0.91</td>
<td>$3.3 \times 10^{-97}$</td>
<td>0.62</td>
<td>$2.94 \times 10^{-14}$</td>
</tr>
<tr>
<td>Ground-based temperature vs re-analysis temperature (2000-2017)</td>
<td>0.96</td>
<td>$1.83 \times 10^{-74}$</td>
<td>0.96</td>
<td>$1.83 \times 10^{-74}$</td>
</tr>
<tr>
<td>Ground-based $R$ vs satellite-based $R$ (2000-2009)</td>
<td>0.46</td>
<td>$2.19 \times 10^{-9}$</td>
<td>0.47</td>
<td>$8.72 \times 10^{-5}$</td>
</tr>
</tbody>
</table>
between R and temperature with all the types of data used. These results corroborate our earlier finding in Table 3.2, which indicates satellite-based radiation and re-analysis temperature compare well with ground-based data.

The comparison between ground-based and satellite-based R against temperature in Hyytiälä station in June and July from 2000 to 2009 is illustrated in Figures 3.4(A) and 3.4(B). It is apparent from these figures that R and temperature are correlated slightly better with satellite-based data than ground-based data. These results are likely to be related to differences between ground-based and satellite-based radiation, because Table 3.2 shows that there is a strong correlation between ground-based and reanalysis temperature.

The differences between radiation from separate data sources may be explained by the discrepancy between spatial resolution of ground-based and satellite-based radiation. Satellite based radiation is for a wider area (1x1 degree) whereas ground based radiation is measured at the surface at Hyytiälä station. As cloud mask data is also for a wider area, satellite based-radiation is more practical here. Furthermore, aerosol size distribution measured at the station is more connected to satellite-based radiation over a wider area of 1x1 degree.

These comparisons show that the relationship is similar in both ground-based and satellite-based data and validates the method to utilise satellite-based data.
TABLE 3.3: Linear correlational analysis of relation between ratio of global radiation to diffuse radiation (R) and temperature with ground-based data and satellite-based data during June and July in a 1x1 degree region around Hyytiälä station and Siberia. N indicates number of data points.

<table>
<thead>
<tr>
<th></th>
<th>Date</th>
<th>r</th>
<th>P</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>In-situ R vs In-situ temperature (Hyytiälä station)</td>
<td>2000-2009</td>
<td>0.34</td>
<td>$6.04 \times 10^{-3}$</td>
<td>63</td>
</tr>
<tr>
<td>In-situ R vs re-analysis temperature (Hyytiälä station)</td>
<td>2000-2009</td>
<td>0.27</td>
<td>$3.27 \times 10^{-2}$</td>
<td>63</td>
</tr>
<tr>
<td>Ceres R vs in-situ temperature (Hyytiälä station)</td>
<td>2000-2016</td>
<td>0.45</td>
<td>$2.07 \times 10^{-7}$</td>
<td>126</td>
</tr>
<tr>
<td>Ceres R vs re-analysis temperature (Hyytiälä station)</td>
<td>2000-2016</td>
<td>0.43</td>
<td>$6.64 \times 10^{-7}$</td>
<td>126</td>
</tr>
</tbody>
</table>

3.4 Aerosol-radiation feedback loop in west Siberia

To investigate whether the aerosol-radiation feedback loop can be expanded for the wider boreal region, I studied the west Siberian plain in addition to Hyytiälä station.

Figures 3.5(A) and 3.5(B) show ratio of satellite-based global radiation to diffuse radiation (R) as a function of re-analysis temperature in Hyytiälä station and the west Siberian plain, respectively, in June and July from 2000-2016. There is a positive correlation between R and temperature in both boreal forest regions. However, this correlation is weaker in the west Siberian plain than Hyytiälä station but there is a similar trend in both regions.

Comparison of correlational statistics between R and temperature with linear and exponential fitting in Hyytiälä station and the west part of Siberia was studied. It was found that the exponential correlation coefficient
is almost the same as the linear one, though the exponential correlation is slightly stronger.

Taken together, these results and the results of relation between CS and temperature, which is exponential, and CS and R, which is linear, indicate that it is reasonable to assume R have an exponential connection to temperature.
FIGURE 3.5: Satellite ratio of diffuse radiation \((R_d)\) to global radiation \((R_g)\) as a function of re-analysis temperature for a 1x1 degree region around (A) Hyytiälä station and (B) west Siberian plain during day time (closest to 12PM) in June and July from 2000 to 2016. The solid line shows the least-squares exponential fit to measurements points.
3.4.1 Impacts of daily median temperature, solar zenith angle and relative humidity

The comparison of the current temperature on aerosol particles and the past 24 hours median temperature on aerosol particles indicates there is no significant difference between correlation coefficient of R against current temperature and the past 24 hours median temperature. Furthermore, as can be seen from Figure 3.6, which is coloured with the difference of days from summer solstice (21st June), changes in solar zenith angle and path length of arriving radiation to the surface does not affect diffuse radiation and consequently the relation between temperature and R. In this figure, negative values represent days before 21st June and positive values are days after 21st June.

![Figure 3.6: Ratio of satellite-based diffuse radiation (Rd) to global radiation (Rg), as a function of re-analysis temperature for a 1x1 degree region around Hyytiälä station during day time (closest to 12PM) in June and July from 2000 to 2016. The solid line shows the least-squares exponential fit to measurements points. Colour scale indicates difference of days from 21st June.](image-url)
Figure 3.7 shows satellite-based $R$ as a function of re-analysis temperature, coloured by RH from 2000-2016 in June and July around noon in Hyytiälä station. As can be seen from Figure 3.7, the swelling effect can impact the relation between $R$ and temperature as the high values of RH correspond to higher $R$ values.

![Figure 3.7: Ratio of satellite-based diffuse radiation ($R_d$) to global radiation ($R_g$), as a function of re-analysis temperature for a 1x1 degree region around Hyytiälä station during day time (closest to 12PM) in June and July from 2000 to 2016. The solid line shows the least-squares exponential fit to measurements points. Colour scale indicates relative humidity.](image)

Figure 3.8a shows the high values of RH are mostly located above the fitted line and the low values of RH are placed under the fitted line. As can be seen from Figure 3.8b normalizing $R$ with RH improved the correlation coefficient between $R$ and temperature in comparison with previous results in Figure 3.5. Same result was obtain for Siberia (Table 3.4).
Figure 3.8: The ratio between satellite-based R (ratio of diffuse radiation to global radiation) and the fitted line equation (f(T1)) from satellite-based R vs re-analysis temperature (from Figure 3.5) as a function of relative humidity (RH) (A), (B) the ratio of satellite-based R to the fitted line equation (f(RH1)) from plot of satellite-based R/ f(T1) vs RH as function of temperature in Hyytiälä station.
Figure 3.9: The ratio of satellite-based R (ratio of diffuse radiation to global radiation) to the fitted line equation (fit(RH2)) from plot of satellite-based R/ f(T2) vs RH as function of temperature in (A) Hyytiälä station and (B) west Siberia (Hyytiälä parameterization was used with Siberia data).

Table 3.4 shows the statistical analysis of the fitted line parametrizations. There is a strong correlation between R and the product of parametrization (fit(RH).fit(T)) in both Hyytiälä and Siberia data, which confirms that the parameterization works well here. In addition, there is a stronger correlation in Siberian data with the Hyytiälä parameterization than the Siberia
A possible explanation for this might be that to obtain the Hyytiälä parametrization, in-situ RH was used, unlike the Siberia parameterization where RH was calculated from re-analysis data.

Table 3.4: Correlation coefficient of fitted line parametrizations in Hyytiälä and Siberia for clear sky conditions at noon. Where \( f(T1) \) and \( f(RH1) \) refers to first step of parameterization and \( f(T2) \) and \( f(RH2) \) refers to second step of parameterization, using iteration. This table shows exponential correlational analysis of relation between ratio of satellite-based \( R \) (ratio of diffuse radiation to global radiation) to \( f(T) \) against RH, as well as ratio of satellite-based \( R \) to \( f(RH) \) against re-analysis temperature. In addition, linear correlational analysis between satellite-based \( R \) and product of parametrization is shown here.

<table>
<thead>
<tr>
<th>Location of data</th>
<th>r</th>
</tr>
</thead>
<tbody>
<tr>
<td>Satellite-based ( R/f(T1) ) vs. RH</td>
<td>0.51</td>
</tr>
<tr>
<td>Satellite-based ( R/f(RH1) ) vs. re-analysis temperature</td>
<td>Hyytiälä</td>
</tr>
<tr>
<td>Satellite-based R vs. ( f(RH1)f(T1) )</td>
<td>Hyytiälä</td>
</tr>
<tr>
<td>Satellite-based ( R/f(T2) ) vs. RH</td>
<td>Hyytiälä</td>
</tr>
<tr>
<td>Satellite-based R vs. ( f(RH1)f(T2) )</td>
<td>Hyytiälä</td>
</tr>
<tr>
<td>Satellite-based R vs. ( f(RH2)f(T2) )</td>
<td>Hyytiälä</td>
</tr>
<tr>
<td>Satellite-based ( R/f(T1) ) vs. RH</td>
<td>Siberia</td>
</tr>
<tr>
<td>Satellite-based ( R/f(RH1) ) vs. re-analysis temperature</td>
<td>Siberia</td>
</tr>
<tr>
<td>Satellite-based R vs. ( f(RH1)f(T1) )</td>
<td>Siberia</td>
</tr>
<tr>
<td>Satellite-based ( R/f(T2) ) vs. RH</td>
<td>Siberia</td>
</tr>
<tr>
<td>Satellite-based ( R/f(RH2) ) vs. re-analysis temperature</td>
<td>Siberia</td>
</tr>
<tr>
<td>Satellite-based R vs. ( f(RH2)f(T2) )</td>
<td>Siberia</td>
</tr>
<tr>
<td>Satellite-based ( R/f(RH2) ) vs. re-analysis temperature</td>
<td>Siberia with Hyytiälä parametrization</td>
</tr>
<tr>
<td>Satellite-based R vs. ( f(RH2)f(T2) )</td>
<td>Siberia with Hyytiälä parametrization</td>
</tr>
</tbody>
</table>
4. Conclusions

The main goal of this study was to analyse the aerosol-radiation feedback loop associated with the COBACC feedback loop using satellite data. This feedback loop connects increasing atmospheric CO$_2$ concentration, rising temperatures, the formation of aerosol particles due to the emission of BVOCs, the change of ratio of diffuse to global radiation in the clear sky condition, and changes in GPP. The aerosol-radiation feedback loop is hypothesized to be negative for the CO$_2$ concentration increase. The aerosol-radiation feedback loop was investigated here by analysing in-situ atmospheric measurement data as well as satellite atmospheric data over Hyytiälä station and a small area in the western plain of Siberia for clear sky conditions in June and July around noon. Here, a region of 1x1 degree in both Hyytiälä and the west part of Siberia was considered.

The first aim of the present research was to investigate the existence of the aerosol-radiation feedback loop with utilizing in-situ data. For this purpose the relations between the components of aerosol-radiation feedback loop were studied, namely temperature and condensation sink (CS), CS and ratio of diffuse radiation to global radiation (R), and R and temperature.

The second aim of this study was to investigate whether it is possible to study the aerosol-radiation feedback loop globally with using satellite data. To achieve this purpose, satellite data was used to investigate whether the same results can be obtained as results from in-situ data. Furthermore, in
this investigation, the impacts of solar zenith angle and RH on the relationship between R and temperature were studied.

One of the main findings to emerge from this study is that the aerosol-radiation feedback loop was confirmed due to finding positive correlation between different components of the aerosol-radiation feedback loop by using in-situ data, namely CS and temperature, CS and R, and R and temperature in this study, whereas the connection between R and GPP has been shown in previous studies (for example, Kulmala et al. (2013)).

The second finding was that there is positive correlation between satellite-based data and in-situ data in Hyytiälä station, namely diffuse radiation, global radiation, and temperature. Furthermore, the investigation of the relation between CS versus temperature or R, as well as R versus temperature show better results with satellite data than in-situ data. Taken together, these findings confirmed that satellite-based data compares well with ground-based data.

The results showed that there are very similar results for the relation between R and temperature in both Hyytiälä and the west plain of Siberia. This major finding suggests that the aerosol-radiation feedback loop can be expanded globally for boreal forests by using satellite data.

The research has found that solar zenith angle has no influence on the relation between R and temperature during study period (June – July). Whereas the investigation of the impact of RH on the relation between R and temperature showed that it is important to take into account the swelling effect in order to inspect this phenomenon.

Further research should focus on determining the relation between R and temperature by considering air-mass trajectories. It would be useful to investigate more about the properties of aerosol particles and divide them to clean and polluted origins.
Considerably more work will need to be done to quantify all the steps of aerosol-radiation feedback loop such as the evaluation of the plant gross primary production (GPP), which measures photosynthesis, and ratio of diffuse radiation to global radiation step with satellite data. This step represents a change in R due to an increase in diffuse radiation possibly leading to plants becoming more photosynthetically active and likely increasing GPP.
A. Appendix

Figure A.1: (A) Satellite-based global radiation ($R_g$) as function of in-situ $R_g$. (B) Satellite-based diffuse radiation ($R_d$) as function of in-situ $R_d$ in June and July from 2000 to 2016 and 2000 to 2009, respectively, in Hyytiälä station
FIGURE A.2: Satellite-based ratio of diffuse radiation to global radiation \( \frac{R_d}{R_g} \) as function of in-situ \( \frac{R_d}{R_g} \) in June and July from 2000 to 2009 in Hyytiälä station.

FIGURE A.3: Re-analysis temperature as function of in-situ temperature in June and July from 2000 to 2009 in Hyytiälä station.
Figure A.4: Satellite ratio of diffuse radiation ($R_d$) to global radiation ($R_g$) as a function of the past 24 hours median re-analysis temperature for a 1x1 degree region around Hyytiälä station (A) and west Siberian plain (B) during day time (closest to 12PM) in June and July from 2000 to 2016. The solid line shows the least-squares exponential fit to measurements points.
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