Observing and Modelling the Surface Radiative Budget and Cloud Radiative Forcing at the Cabauw Experimental Site for Atmospheric Research (CESAR), the Netherlands, 2009–17[Ⓔ]

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ABSTRACT

A dataset of 9 years in duration (2009–17) of clouds and radiation was obtained at the Cabauw Experimental Site for Atmospheric Research (CESAR) in the Netherlands. Cloud radiative forcings (CRF) were derived from the dataset and related to cloud cover and temperature. Also, the data were compared with RCM output. Results indicate that there is a seasonal cycle (i.e., winter, spring, summer, and autumn) in longwave (CRF-LW: 48.3, 34.4, 30.8, and 38.7 W m^{−2}) and shortwave (CRF-SW: −23.6, −60.9, −67.8, and −32.9 W m^{−2}) forcings at CESAR. Total CRF is positive in winter and negative in summer. The RCM has a cold bias with respect to the observations, but the model CRF-LW corresponds well to the observed CRF-LW as a result of compensating errors in the difference function that makes up the CRF-LW. The absolute value of model CRF-SW is smaller than the observed CRF-SW in summer, mostly because of albedo differences. The majority of clouds from above 2 km are present at the same time as low clouds, so the higher clouds have only a small impact on CRF whereas low clouds dominate their values. CRF-LW is a function of fractional cloudiness. CRF-SW is also a function of fractional cloudiness, if the values are normalized by the cosine of solar zenith angle. Expressions for CRF-LW and CRF-SW were derived as functions of temperature, fractional cloudiness, and solar zenith angle, indicating that CRF is the largest when fractional cloudiness is the highest but is also large for low temperature and high sun angle.

1. Introduction

The latest IPCC report (IPCC 2014) reaffirms that confidence in the representation of processes involving clouds remains low. As the evolution of clouds is primarily dependent on regional and local processes it is hard to determine globally the strength and the sign of cloud feedback to the climate system. This is probably the reason why the range of possible values of the

climate sensitivity of numerical prediction models [i.e., the amount by which Earth will warm when the concentration of carbon dioxide (CO₂) is doubled] is high, ranging from less than 2°C at the low end to more than 5°C at the high end (Stephens 2005). Much effort has been spent on directly testing cloud parameterizations in climate models, but the credibility of the representation of clouds can also be evaluated from an assessment of the impact of clouds on the radiation budget of the atmosphere.

A global perspective on clouds can be derived from satellite datasets such as the Earth Radiation Budget Experiment (ERBE) and the Clouds and the Earth's Radiant Energy System (CERES). Both datasets provide top-of-atmosphere (TOA) radiative fluxes that can be used to evaluate the model cloud representation or to determine the impact of clouds on the radiation balance of Earth. An often used method to determine cloud impact is to measure cloud forcing, which is defined as the difference between the net all-sky and the net clear-sky radiant fluxes (Charlock and Ramanathan 1985).

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However, TOA fluxes only provide a limited view on clouds as the impact of clouds on climate is largely governed by the manner in which they redistribute heating and cooling within the atmosphere. Such redistribution cannot be gleaned from a TOA perspective alone so that atmospheric and surface radiative budget studies and datasets are needed to complete our understanding of clouds. For example, the recent availability of active remote sensing data from the *CloudSat* and *CALIPSO* missions (Mace et al. 2009) has been instrumental in adding precision to the vertical distribution of clouds, and thus will be a key in improving the accuracy of atmospheric heating and cooling profiles due to the presence of clouds. For an atmosphere in radiative and convective equilibrium the loss of radiative energy from the atmosphere to the surface is balanced by the input into the atmosphere of sensible and latent heat, so that the surface radiation budget is closely coupled to the hydrological cycle. Even though satellite datasets have in the past been used to obtain surface radiative budgets (L'Ecuyer and Stephens 2003, and many others) it is preferable to use surface datasets as they are less dependent upon assumptions about the state of the atmosphere between the satellite sensor and the surface level at which the value of radiant fluxes are desired. They also can serve as validation material for the satellite datasets which by virtue of their much wider global coverage remain an attractive tool in understanding clouds.

In the last two decades, several comprehensive datasets of cloud radiative effects were collected in a wide variety of climate regimes. Much attention has been given to the impact of clouds on the Arctic ice/snow surface (Intrieri et al. 2002; Wang and Key 2003; Shupe and Intrieri 2004; Miller et al. 2015; Bennartz et al. 2013; Walsh et al. 2009) as cloud forcing there is dominated by longwave emission from low clouds thought to induce melt-off and thus affecting the mass balance of ice sheets. Radiative budget climatologies of the Arctic Barrow, Alaska (Dong and Mace 2003; Dong et al. 2010), and the midlatitude Southern Great Plains (Dong et al. 2006) Atmospheric Radiation Measurement (ARM) sites have offered a comprehensive view of the seasonal and yearly variability of cloud forcing. Budget/forcing studies of the tropics are provided by May et al. (2012) for the monsoon region at the ARM site of Darwin, Australia; by McFarlane et al. (2013) for the ARM tropical western Pacific sites; for the eastern Pacific Ocean region by Cronin et al. (2006), who used mooring arrays during the Eastern Pacific Investigation of Climate Studies (EPIC) underneath the stratus decks that are observed; by Ghate et al. (2009), who collected five years of single buoy data for the Pacific Ocean; and

by Fairall et al. (2008), who collected data from the Tropical Atmosphere Ocean (TAO) buoy lines. For southern Europe, Mateos et al. (2013) and Salgueiro et al. (2014) provide multiyear shortwave radiative cloud forcing data; Marty et al. (2002) studied the altitude dependence of cloud forcing in the Alps in Switzerland. The ARM Mobile Facility was deployed at Shouxian, China, for several months in 2008 to collect cloud forcing and surface radiation data (Qiu et al. 2013) and in the western Sahel region of Africa during the Radiative Atmospheric Divergence using ARM Mobile Facility, GERB, and AMMA Stations (RADAGAST) experiment (Miller et al. 2012; GERB is the Geostationary Earth Radiation Budget; AMMA is the African Monsoon Multidisciplinary Analysis). Further oceanic studies of cloud effects on the radiative balance are provided by ship cruises in the Atlantic (Kalisch and Macke 2012) and by a ship-based experiment over the Southern Ocean—namely, the Clouds, Aerosols, Precipitation, Radiation, and Atmospheric Composition over the Southern Ocean (CAPRICORN) experiment (Protat et al. 2017).

Even though these studies were taken in different climate regimes some tentative conclusions may be drawn from them. 1) In the tropics at low latitudes, negative shortwave cloud forcing dominates, whereas in the Arctic at high latitudes the positive longwave cloud forcing is the most important. 2) The midlatitudes represent a transition region where the net forcing averaged over a year might be positive or negative depending on the distance to either pole or equator or by the type of clouds that are prevalent on the location at hand. 3) Cloud forcing is dependent on cloud fraction as expected, but not as a linear combination of a clear-sky and a cloudy sky component; in general a more complex functional dependence is required, which has implications for the parameterization of clouds in weather and climate models. 4) The few comparisons with model output yield scattered results, indicating the difficulty of models to adequately represent cloud forcing.

Despite the availability of studies with cloud forcing datasets there are only a few comparisons of cloud forcing data with output from models. Most studies are confined to comparisons of data with reanalysis results using the European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA) or the National Centers for Environmental Prediction reanalysis projects (Cronin et al. 2006; Fairall et al. 2008). Despite the fact that in such studies the single point output of models is tightly constraint by data sources from the immediate surroundings only ambiguous conclusions about the quality of the reanalysis models could be derived. Results are no better when regional or global

climate models (RCMs and GCMs, respectively) are used. Miller et al. (2012) used two GCMs from phase 3 of the Coupled Model Intercomparison Project (CMIP3) to intercompare cloud forcing data and model output and found that both models underestimated the longwave and shortwave cloud forcing. Kalisch and Macke (2012) used a single column version of the ECMWF model developed at the Max Planck Institute in Hamburg (ECHAM) and found significant difference between data and model output under convective conditions. Last, Protat et al. (2017) compared data with regional weather forecasts over the Southern Ocean using the Australian Community Climate and Earth System Simulator (ACCESS) and found considerable scatter in the comparisons and an underestimation of low cloud cover impacting cloud forcing calculations.

In conclusion, the summary here can be viewed as a motivation to obtain additional cloud and radiation observational and model data of high quality and long duration. Thus, one purpose of this paper is to produce such a dataset of surface radiative fluxes and to study the impact of clouds on these fluxes. A second purpose is to compare these data with output from an RCM. The output of this paper will also provide the satellite community with validation datasets to improve their derived surface radiation products.

To this end we analyze radiation and cloud data at the Cabauw research site in the Netherlands for the 9-yr period 2009–17 and compare them with output from an RCM. Data are first condensed to daily averages and then processed to monthly and eventually yearly time series. Surface cloud radiative forcing is calculated and the relationship between cloud forcing and other relevant parameters such as cloud fraction, water vapor column, temperature, and solar zenith angle is derived. An analysis of errors is performed on the output so that the future user of these data will be able to gauge the accuracy of their own datasets vis-à-vis the data at hand. Section 2 describes the data sources and analysis procedure. Although the radiation data are, as we will show, of the highest quality, a considerable and non-trivial effort is still required to identify periods with cloud-free conditions from which the baseline clear-sky fluxes can be derived necessary to produce time series of cloud forcing. The procedure to do so yields a “virtual” time series of clear-sky radiation for periods when the sky is partly cloudy or completely overcast. Such virtual time series have larger errors than those associated with direct radiation flux output, which is treated in section 3. In section 4 the Regional Atmospheric Climate Model (RACMO) will be described. Section 5 shows the results, and section 6 provides a discussion and conclusions.

2. Data sources and analysis procedures

a. CESAR

The Cabauw Experimental Site for Atmospheric Research (CESAR) is situated at 51.978°N, 4.938°E in a rural grassland region at a distance of 22 km southwest of the city of Utrecht (<https://www.cesar-observatory.nl>). It was established in 1972 and in the first 20 years of its existence was used to study exchange processes between Earth’s surface and the atmosphere. Central to the site is a tower of 213-m altitude that is instrumented at the 20-, 40-, 80-, 140-, and 200-m levels with thermodynamic, turbulence, and wind probes. Since the mid-1990s the site has also been used to study the interaction of clouds, aerosols, and radiation using in situ and remote sensing instruments and to monitor trace gases such as CO₂, methane, and radon. A suite of instruments has been installed for the (remote) detection of hydrometeors and aerosols, such as a microwave radiometer, several cloud radars, a cloud lidar, a Raman lidar, an all-sky camera, an infrared scanning radiometer, and filter radiometers. In 2005 a Baseline Surface Radiation Network site was placed several hundreds of meters south of the tower, and shortwave and longwave radiometers were mounted at the top of the tower to observe the surface-reflected shortwave radiative flux and the surface-emitted longwave radiative flux. An overview of the Cabauw site is given by Monna and Bosveld (2013).

Although KNMI is the site manager of CESAR, a consortium of Dutch universities and scientific and technological research institutes jointly runs the scientific program. Each of the institutions of the CESAR-consortium runs their own scientific equipment, and CESAR provides a platform for collaboration in the field of atmospheric sciences. Over the years, international research groups have temporarily located their instruments at CESAR, and the site has contributed to European field experiments [such as the European Integrated Project on Aerosol Cloud Climate and Air Quality Interactions (EUCAARI; see Kulmala et al. 2011) and many others].

b. Cloud forcing

The net surface radiative flux R_{net} is defined as the difference between the incoming and outgoing sum of shortwave (SW) and longwave (LW) radiation, where the downward radiation is assigned as positive:

$$R_{\text{net}} = (LW_{\downarrow} - LW_{\uparrow}) + (SW_{\downarrow} - SW_{\uparrow}). \quad (1)$$

The net cloud radiative forcing (CRF) is then defined as the difference between the net surface radiative flux

under all-sky conditions and the net radiative flux under clear-sky conditions:

$$\text{CRF} = R_{\text{net,all-sky}} - R_{\text{net,clear-sky}}. \quad (2)$$

In this paper, fluxes are obtained at 10-min intervals and processed thereafter.

c. Clear-sky radiation

Proper calculation of CRF involves assigning values to the LW and SW downward radiation for clear skies. Thus it is necessary to compute a virtual downward clear-sky flux for periods when the sky is partly cloudy or overcast. Any method to do so will have to interpolate between cloud-free periods. As a baseline for such interpolation schemes individual 10-min data points are to be marked as clear or cloudy. The simplest way to determine whether a data point is clear or cloudy is to invoke the remote sensing instruments that detect hydrometeors. At CESAR there are a number of instruments to detect clouds, and their quality was extensively evaluated by Boers et al. (2010). The best instrument to detect clear and cloudy skies is the Nubiscope (NUB; an infrared scanning radiometer) with the total sky imager (TSI) as a close second best. Less suitable are the cloud lidar and cloud radar. Although either of the last two instruments are capable of detecting hydrometeors they are not accurate at the very high time resolutions that are used in this paper as they are pointed in the nadir direction and will thus not detect off-nadir clouds. Off-nadir clouds have a high impact on the radiation measured by the hemispheric radiometers, so it was decided not to use the lidar and cloud radar in the determination of fractional cloudiness at the 10-min intervals to indicate whether the radiometers were exposed to a clear sky. A third remote sensing instrument, a microwave radiometer, is not usable for this purpose either as it only detects liquid water clouds, whereas in the Netherlands there is a sizable portion of clouds present in the ice phase throughout all seasons. However, information about clouds can be obtained by examining the shortwave and longwave radiometer observations.

1) DOWNWELLING SHORTWAVE RADIATION

Long and Ackerman (2000) developed a method to isolate daytime clear-sky periods while making exclusive use of broadband radiometers. They later polished the method to estimate fractional cloudiness as well (Long et al. 2006). As they exclusively rely on radiation instruments, they have limited information to obtain their desired parameters and thus their method has a certain ad hoc character. Nevertheless, some aspects of their method are useful to supplement the NUB and TSI data and we

will use them here. For cloud detection priority was given to NUB. However NUB was absent for repair/calibration for several months throughout the 9-yr period at which time TSI took over the cloud detection. Together, the two instruments were able to cover the entire 9-yr period for the determination of daytime clear skies. Supplemental information about daytime cloud cover can be obtained by examining the functions $\text{SW}_{\downarrow}/\mu_0$, where μ_0 is the cosine of solar zenith angle, and $\text{std}(\text{SW}_{\downarrow,\text{diffuse}})/(\text{SW}_{\downarrow,\text{diffuse}})$. Here the indication “std” reflects the 10-min-time-scale standard deviation of the parameter within the parentheses. Long and Ackerman (2000) found that $\text{SW}_{\downarrow}/\mu_0$ should fall within a limited interval to be assigned as clear, with its variation within prescribed limits being entirely due to varying aerosol content. The expression $\text{std}(\text{SW}_{\downarrow,\text{diffuse}})/(\text{SW}_{\downarrow,\text{diffuse}})$ defines the fraction of standard deviation of the diffuse downward directed shortwave radiation to the total diffuse downward directed shortwave radiation within a 10-min period. For a clear sky it should fall below a prescribed limit that is a function of solar zenith angle. In practice, data points were assigned as clear-sky data points when NUB or TSI assigned a point to be clear while at the same time the radiation data adhered to the limits $1100 < \text{SW}_{\downarrow}/\mu_0 < 1300$ and $\text{std}(\text{SW}_{\downarrow,\text{diffuse}})/\text{SW}_{\downarrow,\text{diffuse}} < 0.8\mu_0^{-0.6}$. These limits are not the same as those used by Long and Ackerman (2000) but are optimized values obtained after examining the entire 9-yr dataset.

Per individual year, all clear-sky data points were then collected and fitted with the function

$$\mu_0 \ln(\text{SW}_{\downarrow}/\mu_0) = A + B\mu_0. \quad (3)$$

The validity of this procedure was explained by Boers et al. (2017). The factors A and B determine the mean zenith angle variation of the clear-sky solar radiation for the entire year. Next, individual days with more than eight clear-sky points with $\mu_0 > 0.10$ were selected, and for each day the values of A_i and B_i were determined using Eq. (3). The A_i and B_i values were quality checked by comparing them with the yearly values of A and B . Differences of up to 15% with the yearly averaged values were possible, but no action was undertaken to correct them because the fits were deemed of good quality. For the remaining days (which constitute the majority of days within a year) the values of A_i and B_i were simply linearly interpolated between days for which acceptable values of A_i and B_i could be generated. In this way a daily clear-sky shortwave virtual radiative flux time series was created.

2) UPWELLING SHORTWAVE RADIATION

For the reflected clear-sky virtual flux $\text{SW}_{\uparrow,\text{clear}}$, the albedo (Alb) was calculated using the Beljaars and Bosveld (1997) formula:

$$\text{Alb} = 0.33 - 0.13\mu_0. \quad (4)$$

Data from (almost) clear days were collected, and Eq. (4) was compared with the measured reflected flux, which led to the conclusion that Eq. (4) still provides an excellent parameterization of the albedo.

3) DOWNWELLING LONGWAVE RADIATION

We suppose that the downward longwave radiation $\text{LW}_\downarrow$ under clear sky can be expressed as

$$\text{LW}_{\downarrow, \text{clear-sky}} = \varepsilon \sigma T^4. \quad (5)$$

Here ε is the effective emissivity of the atmosphere above the level at which radiation is measured, σ is the Stefan–Boltzmann constant, and T is the temperature at which the longwave radiation is measured. To estimate the clear-sky radiation in the absence of clear-sky periods is a well-trodden path. Li et al. (2017) evaluated no less than 15 different parameterizations to estimate the effective emissivity ε , and the evaluated methods in their paper do not even count additional ad hoc measures such as those developed by Dürr and Philipona (2004) and Long and Turner (2008). The latter two computed clear skies from longwave radiometer data that depend on the short-term variability of the longwave radiation under clear skies. The essence of the methods evaluated by Li et al. (2017) is that ε is a function of the temperature at the observation level and of the vertically integrated water vapor (IWV). So the parameterization schemes were constructed by calculating radiative fluxes for a wide variety of temperature–humidity profiles with different IWV and observation-level temperatures. It is thus in principle possible to choose any of the listed parameterizations and compute $\text{LW}_{\downarrow, \text{clear-sky}}$ for our dataset. However, we decided to develop our own ad hoc method because of the availability of clear-sky data points as identified by NUB and TSI. This method is explained in the appendix and relies on the calculation of the clear-sky effective emissivity under cloudy conditions by interpolating effective emissivity values inferred from downward longwave radiation measured at clear-sky periods.

4) UPWELLING LONGWAVE RADIATION

Assuming a surface emissivity of unity, the upwelling longwave radiation under clear-sky conditions is considered to be equal to the measured upwelling longwave radiation under all-sky conditions:

$$\text{LW}_{\uparrow, \text{clear-sky}} = \text{LW}_{\uparrow, \text{all-sky}}. \quad (6)$$

This approach conforms to the manner in which cloud forcing calculations are performed in RACMO.

d. Cloudnet

To probe the vertical distribution of clouds we use the combination of radar and lidar observations by applying the Cloudnet procedure. Cloudnet was originally designed as a program to compare the cloud data from several current numerical weather forecast models with ground-based remote sensing observations (Illingworth et al. 2007). To this end, a detection algorithm was constructed that ingested the data thermodynamic profiles from a numerical weather prediction (NWP) model together with data from ground-based radar, lidar, and microwave radiometers to automatically assign cloud heights, cloud boundaries, and the phase of hydrometeors. Cabauw was one of the three stations for which this algorithm was constructed and evaluated (the others were Chilbolton, United Kingdom, and Palaiseau, France). We apply the algorithm here to process all 9-yr data. The cloud distributions that are thus obtained have a vertical resolution of 90 m from 250 m above the surface to an altitude of 12 km and have a time resolution of approximately 15 s. These data are averaged in the Cloudnet processing to a time resolution of 1 h and projected on the model vertical grid for ease of assessment of the model output. The 1-h Cloudnet data are used in this study.

3. Uncertainties

All instruments are calibrated to absolute standards at the radiation calibration facility in Davos, Switzerland, at intervals of five years. Uncertainties were investigated by Wang et al. (2009) and comprise calibration uncertainties and operational uncertainties, the latter of which are larger than the former. From Wang et al. (2009) and Shi and Long (2002) we estimate the operational accuracy of individual 10-min averages to be on the order of $10\text{--}14 \text{ W m}^{-2}$. Taking 14 W m^{-2} as the upper limit in uncertainty, then the daily average will incur an uncertainty of a little more than 1 W m^{-2} . Here it is assumed that errors are statistically independent. Although calibration to absolute standards probably did remove most bias, intercomparison of different sensors revealed that residual biases remained at a level of 0.1%, or 1 W m^{-2} . Such bias cannot be removed by averaging over a month (i.e., the minimum time period that will be shown in this paper), and therefore monthly values of measured radiation incur a precision error of 1 W m^{-2} . Combining the accuracy and precision of the radiation observations, the total uncertainty of monthly radiation values is 2 W m^{-2} .

Different standards should be employed for the virtual clear-sky fluxes that were calculated under conditions of a sky that was partly cloudy or overcast. The calculation of these fluxes involves assumptions about the state of the atmosphere and assessment of the validity of interpolation schemes and depends on how much of the sky was clear, in which case observed fluxes could be used rather than calculated virtual fluxes. It is hard to come up with precise estimates, but an indication of potential errors is given when different ways to compute the monthly averages are compared. For the shortwave fluxes, the virtual clear-sky flux can be calculated using the yearly mean value of the coefficients A and B of Eq. (3), or using their 10-min values. Then the difference between the fluxes based on these calculations yields a quantity that can serve as the potential uncertainty. In this way we find typical uncertainties of 7 W m^{-2} . This means that the upward clear-sky flux has an uncertainty of 2 W m^{-2} (considering that the shortwave albedo is in the range of 0.2–0.3). Similarly, for the longwave fluxes an indication of errors is given when our estimates of downwelling clear-sky fluxes using the 2- or 80-m tower emissivity [see Eqs. (6) and (7)] are intercompared, or when the fluxes that are based on the tower estimates are compared with those using standard parameterizations as listed in Li et al. (2017) such as the Prata (1996) formulation. For the clear-sky downwelling longwave flux, monthly uncertainties on the basis of the difference between the fluxes based on the 2- or the 80-m emissivity are 4 W m^{-2} . When comparing the tower-based fluxes with the parameterization from Prata (1996) there is a monthly uncertainty of 8 W m^{-2} . Therefore we put the uncertainty in calculating the monthly clear-sky flux at 6 W m^{-2} . For the upwelling clear-sky fluxes, the combined operational uncertainty and bias applies (2 W m^{-2}).

Given these estimates, the uncertainty in the monthly net radiative balance can be obtained using Eq. (1). All forcing and flux components are additive so that the total uncertainty in radiative forcing is the root of the sum of the squares of the individual component uncertainties. For the all-sky radiative flux balance, the total uncertainty is $(2^2 + 2^2 + 2^2 + 2^2)^{1/2} = 4 \text{ W m}^{-2}$ with the uncertainties equally balanced between the longwave and the shortwave components. For the clear-sky radiative flux balance, it is $(7^2 + 2^2 + 6^2 + 2^2)^{1/2} = 10 \text{ W m}^{-2}$ with the uncertainty in shortwave balance being slightly larger than that of the longwave balance. This means that the monthly total cloud forcing [Eq. (2)] has an uncertainty of $(10^2 + 4^2)^{1/2} = 11 \text{ W m}^{-2}$, with most of this uncertainty being attributable to the uncertainty in estimating the clear-sky fluxes, as would be expected. The component shortwave and longwave cloud

forcings thus have uncertainties of $(2^2 + 2^2 + 7^2 + 2^2)^{1/2} = 8 \text{ W m}^{-2}$ and $(2^2 + 2^2 + 6^2 + 2^2)^{1/2} = 7 \text{ W m}^{-2}$, respectively.

4. RACMO

The hydrostatic RCM KNMI-RACMO (hereinafter RACMO) is originally built from the parameterization package of physical processes employed in the ECMWF physics merged with the dynamical kernel of the High Resolution Limited Area Model (HIRLAM; Undén et al. 2002) for NWP. RACMO2.3 applied in this study is based on ECMWF cycle 33r1 (ECMWF-IFS 2009). In RACMO2.3, both the shortwave and longwave radiative schemes utilize versions of the Rapid Radiative Transfer Model (RRTM): Clough et al. (2005) for shortwave radiation and Mlawer et al. (1997) for longwave radiation, and the independent column approximation (ICA; Morcrette et al. 2008) is applied to account for the effect of clouds on radiation at the subgrid scale. Other physics components include a turbulent kinetic energy (TKE)-driven eddy-diffusivity mass-flux scheme (Siebesma et al. 2007; Lenderink and Holtslag, 2004; Baas et al. 2008) for mixing and cloud processes in the boundary layer, a scheme for deep convection [originated by Tiedtke (1989)], a prognostic cloud scheme (Tiedtke 1993; Tompkins et al. 2007), and the “HTESSEL” land surface/soil scheme (Balsamo et al. 2009). Apart from the use of the shortwave component of RRTM and the application of ICA, the model formulation of RACMO2.3 is very similar to its predecessor (van Meijgaard et al. 2012).

The model time series used in this study are taken from a continuous climate-mode simulation with RACMO (1979–2017) forced by atmospheric fields from ERA-Interim (Dee et al. 2011) at the lateral boundaries. Prescribed sea surface temperature results from a slab ocean model (Attema and Lenderink 2014) are merged with information from ERA-Interim. The land surface and soil state, initialized from ERA-Interim at the start of the climate simulation (0000 UTC 1 January 1979), evolves freely as part of the climate-mode simulation. The RACMO modeling domain comprises western Europe, the horizontal resolution is 12 km, the vertical mesh counts 40 model levels, and the time step size is typically 300 s. To facilitate the comparison with observations taken at Cabauw, (sub-) hourly time series of (near-) surface and atmospheric model state variables and fluxes have been archived for the grid cell column encompassing Cabauw. Parameters relevant to this paper include cloud fraction profile and total cloud cover derived from the maximum-random overlap assumption, as well as all-sky and clear-sky upward and

downward shortwave and longwave radiative fluxes at the surface.

5. Results

a. The general climatic cloud conditions at Cabauw

The Netherlands are located at the edge of the European continent at a latitude at which westerly winds dominate. This means that the area is exposed to the passage of frontal systems and rainbands originating from the Atlantic Ocean and the North Sea in all months of the year. Cabauw is situated in relatively open flat, rural grassland (polder landscape) where most land is surrounded by dykes and the groundwater table is tightly controlled by water management systems. This means that there is sufficient moisture available year-round to supply agriculture. Also, during summer months the surface boundary layer is often coupled to the low clouds because the top of the boundary layer will exceed lifting condensation level. As the westerlies drift to lower latitudes in winter, more frequent frontal passages will increase cloud fraction relative to the summer months. However, because of the absence of significant sunlight the boundary layer is then often stable and uncoupled from even the lowest clouds. We refer to Boers et al. (2010) for a more extensive discussion of cloudiness conditions at Cabauw.

b. The height distribution and temporal variation of cloudiness

Focusing first on the measured and modeled cloud fractions, Fig. 1a shows three time series of monthly cloud fractions, with a month-by-month comparison in Fig. 1b and the average monthly mean in Fig. 1c. The Nubiscope recorded smaller cloud fractions than Cloudnet. The difference of typically 10% is mostly caused by the sensitivity of the cloud radar to high cirrus clouds. For NUB, in the infrared part of the spectrum high clouds are difficult to discern against a cold (water vapor background), in particular if water vapor content in the lower atmosphere is relatively high. However, all three systems showed the same time variation. There is a seasonal cycle in cloudiness with high cloud amounts in winter and low cloud amounts in summer. In the mid-latitude sea climate of CESAR this is mainly related to the seasonal variation of the difference in ocean water temperature and land air temperature. All three systems show this cycle (Fig. 1c). RACMO cloud amount is higher than that recorded by the Nubiscope but lower than that recorded by the Cloudnet.

Cloudnet records higher cloud fractions than are modeled by RACMO in the atmosphere above 1.5 and

FIG. 1. Observed (Nubiscope and Cloudnet) and modeled (RACMO) cloud fraction: (a) time series, (b) month-by-month comparisons of modeled cloud fraction with observations, and (c) 9-yr monthly means. Error bars represent the standard deviation of the monthly mean and not the calibration/operational uncertainty of the instruments.

below 8 km (see Fig. 2). Both profiles record high cloud fraction near altitudes of 1 km caused by daytime convection and stratus clouds and near 9–10 km caused by cirrus clouds. At the lowest level for which Cloudnet observations are available (250 m) RACMO records slightly higher cloud fractions than Cloudnet. To understand this difference we compared RACMO temperature and integrated water vapor path with observations. Figure S1a in the online supplemental material shows that RACMO has a small low bias in temperature (less than 0.5 K) in summer increasing to 0.5–1.0 K in winter. However, at the same time the integrated water vapor path is not biased (Fig. S1b). As most water vapor is located in the lowest 1–2 km of the atmosphere it is likely that the excess of clouds close to the surface in RACMO is caused by excess condensation in a relatively moist atmosphere. The cold bias of RACMO is a long standing issue, the cause of which is unclear. As an addition to Fig. 2, averaged profiles of cloud fraction for the four seasons are shown in Fig. S2 of the

FIG. 2. Mean vertical profiles of observed (Cloudnet) and modeled (RACMO) cloud fraction. The error bars represent the standard deviation over the nine yearly mean profiles.

online supplemental material. In winter cloudiness is higher than in summer at all levels up to the tropopause, but the seasonal variation in vertical structure of the clouds is relatively small. Also, even though there are some differences between the observations and the RACMO output, the latter follows the former remarkably well. A tentative conclusion from these comparisons is that the lower cloud fractions (as seen from the surface; Fig. 1a) of RACMO when compared with that of Cloudnet is at least partly caused by the frequent absence of modeled RACMO clouds at intermediary levels (2–8 km) in comparison with the Cloudnet clouds. This result is similar to that found by Illingworth et al. (2007). Below (section 5c) we will explore the relationship between the vertical cloud structure and cloud forcing in more detail.

c. Cloud radiative forcing: its composition and time variation

Figures 3a–c show a comparison of total cloud forcing and the shortwave and longwave CRF components between observations and model. There is a seasonal cycle in all three time series. CRF-LW is positive with a

FIG. 3. Observed and modeled total cloud forcing and longwave and shortwave cloud forcing components: (a) time series, (b) month-by-month comparison of observed and modeled cloud forcing and cloud forcing components, and (c) 9-yr averaged yearly cycle.

minimum in summer and a maximum in winter. CRF-SW is negative with the largest negative values in summer and the smallest values in winter. The resulting total CRF is negative in summer and positive in winter. The time series in Fig. 3a and the comparison in Fig. 3b indicate that the more negative the value for CRF-SW, the larger the departure is between the observed and modeled total CRF. Figure 3c indicates that this coincides with large negative values of CRF-SW from May throughout October. From January through to April, the CRF-SW of RACMO is larger than the observations. RACMO CRF-LW values on the average almost perfectly match the observations.

Figures 4a–c indicate that the model output and observations of the difference function of the all-sky and clear-sky longwave fluxes are well matched. However, the breakdown into the component fluxes (Fig. 5) indicates that in the winter months the RACMO fluxes are smaller than the observations by about 10 W m^{-2} . However, because this difference applies for all fluxes, they compensate each other so that in the end they do not affect the cloud forcing results. This is an

FIG. 4. As in Fig. 3, but for longwave cloud forcing and all-sky and clear-sky cloud forcing components.

interesting result because the low bias for the RACMO longwave flux components in the winter months is probably due to the RACMO low bias in temperature in winter. Therefore, the comparison of cloud forcing components between model and observations will not reveal such a structural issue if the individual forcing components that make up difference functions are all biased by the same amount and occur at the same times during a year.

Figures 6a and 6b show that RACMO yields high values of the clear-sky shortwave flux-difference function in comparison with the observations. This causes the RACMO cloud forcing to be more negative than the observations, in particular in spring, summer, and autumn. Figure 6c shows the flux components and the intercomparison for individual months. There are two issues that play a role here. First, RACMO takes its aerosol concentration from the CMIP5 simulations, which are biased low to the typical conditions in the Netherlands. This means that the incoming SW radiation is biased high in particular in summer. Second, a comparatively low SW albedo of 0.18 is assigned to the model grid cell of Cabauw. Observations show that the observed SW albedo is variable with sun angle, but ranges from 0.23 to 0.27. The lower model albedo

FIG. 5. Observed and modeled longwave flux components: (a) time series, (b) month-by-month comparison of observed and modeled flux components, and (c) 9-yr monthly averages.

produces smaller upward shortwave fluxes in both the model all-sky and the model clear-sky fluxes (Fig. 7). These fluxes do not compensate each other in the difference calculations that make up the shortwave cloud forcing components. Therefore a comparatively large negative shortwave cloud forcing remains for the RACMO results.

d. Cloud radiative forcing, fractional cloudiness, temperature, and sun angle

CRF-SW and CRF-LW are functions of cloud fraction (Dong et al. 2006; Miller et al. 2015; Dong et al. 2010; Shupe and Intrieri 2004). However, sun angle (for CRF-SW) and temperature (for CRF-LW) are also important. If sun angle is low, CRF-SW must be small, no matter how large the fractional cloudiness may be. So, a seasonal and latitudinal variation in CRF-SW is expected. Longwave radiation is a function of the IR-radiant temperature of clouds and of the temperature of the ambient water vapor laden air. Hence CRF-LW is a function of temperature and thus has a seasonal and latitudinal variation as well. Below we find relationships between CRF, fractional cloudiness, temperature, and sun angle.

FIG. 6. As in Fig. 3, but for shortwave cloud forcing and all-sky and clear-sky cloud forcing components.

1) CRF-LW, FRACTIONAL CLOUDINESS, AND TEMPERATURE

Figure 8 shows that there is a relationship between the fractional cloudiness and CRF-LW. Although a straight line can in principle be drawn through the data (broken line), the physics of the relationship dictates that the line ought to go through the origin, so an additional power-law relationship (solid line) was drawn through the data. This provides a better fit.

If we consider the longwave component of Eqs. (1) and (2) then

$$\text{CRF}_{\text{LW}} = (\text{LW}_{\downarrow \text{all-sky}} - \text{LW}_{\uparrow \text{all-sky}}) - (\text{LW}_{\downarrow \text{clear-sky}} - \text{LW}_{\uparrow \text{clear-sky}}). \quad (7)$$

As shown in section 2, the all-sky and clear-sky longwave upward fluxes are considered the same, so they cancel in Eq. (7). Therefore the solid line represents the fit

$$\text{CRF}_{\text{LW}} = (\text{LW}_{\downarrow \text{all-sky}} - \text{LW}_{\downarrow \text{clear-sky}}) = ac^b, \quad (8)$$

with $a = 68 \text{ W m}^{-2}$ and $b = 1.43$. For fractional cloudiness $c = 1$, the all-sky downward flux is equal to the flux under a solid cloud deck. Hence

FIG. 7. As in Fig. 5, but for shortwave flux components.

$$a = (\text{LW}_{\downarrow \text{cloud-base}} - \text{LW}_{\downarrow \text{clear-sky}}). \quad (9)$$

Thus, averaged over a year, there is a difference of 68 W m^{-2} between the clear-sky flux and the downwelling longwave flux underneath a solid cloud deck. It is presumed that the factor a is likely a function of temperature as both the cloud base and the clear-sky

FIG. 8. Observed longwave cloud forcing as a function of fractional cloudiness. The dotted line is the straight-line fit through the data. The solid line represents the function $\text{CF}_{\text{LW}} = ac^b$.

FIG. 9. Various longwave fluxes as functions of surface temperature.

components in Eq. (9) are functions of temperature. The temperature dependence of the longwave fluxes is shown in Fig. 9 where fluxes represent monthly averages. There appears to be a gradual transition of the overcast/all-sky fluxes from winter (cold) to summer (warm) temperature. In winter, the cloud-base temperature is close to the surface temperature so that the flux from an overcast sky is almost that of the surface. With increasing temperature toward summer the cloud-base flux becomes closer to the clear-sky flux.

The overcast and the clear-sky flux can be related to the surface temperature T_{surf} by power-law dependences on temperature

$$\text{LW}_{\downarrow\text{overcast}} = a_1 T_{\text{surf}}^{b_1}, \quad (10)$$

with $a_1 = 1.0396 \times 10^{-6}$ and $b_1 = 3.4713$. Similarly,

$$\text{LW}_{\downarrow\text{clear-sky}} = a_2 T_{\text{surf}}^{b_2}, \quad (11)$$

with $a_2 = 1.2382 \times 10^{-11}$ and $b_2 = 5.4472$. (For the calculation use temperature in kelvins.) With high temperatures the downward flux under a cloudy sky resembles that of a clear sky because the sky temperature is a function not only of the land surface temperature but also of water vapor column, which is increasing with temperature. Inserting Eqs. (10) and (11) in Eq. (9) we obtain from Eq. (8)

$$\text{CRF}_{\text{LW}} = c^b (a_1 T_{\text{surf}}^{b_1} - a_2 T_{\text{surf}}^{b_2}). \quad (12)$$

Because Cabauw is occasionally exposed to Arctic temperatures in winter and to tropical temperatures in summer, it is not unreasonable to suggest that Eq. (12) also mimics a latitudinal transition of this function from the pole toward the equator. Figure 8 then indicates that for tropical temperature CRF-LW becomes small. This

FIG. 10. The parameterized cloud radiative forcing components: (a) the longwave radiative cloud forcing as a function of cloud fraction (horizontal axis) and surface temperature (vertical axis) and (b) the shortwave radiative cloud forcing as a function of cloud fraction (horizontal axis) and cosine of solar zenith angle (cmu; vertical axis).

point has been made earlier by Stephens et al. (2004) and others, but with Eq. (12) we suggest here a functional form for the vanishing of CRF-LW in the tropics. Equation (12) is plotted in Fig. 10a as a function of cloud fraction and temperature. (See the next section for Fig. 10b.) For cold temperatures the CRF-LW approaches a value of just over 70 W m^{-2} . This suggests that there is an upper limit to the longwave cloud forcing in the Arctic. Observations of CRF-LW from the Arctic region do indeed indicate that CRF-LW does not exceed $\sim 80\text{--}85 \text{ W m}^{-2}$.

2) CRF-SW, FRACTIONAL CLOUDINESS, AND SUN ANGLE

For CRF-SW no straightforward relationship with fractional cloudiness emerges (Fig. 11a). In the summer when the sun is high in the sky (May, June, and July) there is an almost straight-line dependence, but this relationship falls apart in the transition spring and autumnal period. It then emerges again in the winter

FIG. 11. (a) CRF-SW and (b) normalized CRF-SW (normalized by the cosine of solar zenith angle) as a function of fractional cloudiness.

months. It is clear that the changing sun angle in the spring and autumn impacts changes in CRF-SW more than the fractional cloudiness. To eliminate the effects of changing solar zenith angle, the shortwave radiation was normalized by the cosine of solar zenith angle for every 10-min data point. Next, the flux differencing, CRF-SW calculation, and averaging were repeated. The monthly averaged normalized fluxes were then plotted as a function of fractional cloudiness (Fig. 11b). Again a functional relationship emerges with data from all seasons collapsing onto the same line. So, the seasonal scatter in data points in Fig. 11a is mostly due to the seasonal changes in solar zenith angle. For monthly averages we therefore suggest a simple parameterization of the shortwave radiative cloud forcing as follows:

$$\text{CRF}_{\text{SW}} = a_3 \overline{\mu_0} c^{b_3}. \quad (13)$$

Here $a_3 = -331.1$ and $b_3 = 1.39$. The overbar represents the monthly average of the cosine of solar zenith angle.

In Fig. S3 of the online supplemental material it is shown that for RACMO relationships that are almost identical to those from the observations could be

derived for the component cloud forcings and as a function of fractional cloudiness. RACMO longwave fluxes closely follow the same relationship with temperatures as found from the data (online supplemental Fig. S4). Equation (13) was plotted in Fig. 10b as a function of cloud fraction and mean monthly averaged cosine of solar zenith angle ($\overline{\mu_0}$). For a given temperature and sun angle Fig. 10 suggests that the shortwave cloud forcing largely changes in exactly the opposite direction along the cloud fraction axis to that of the longwave cloud forcing (see also Stephens et al. 2012a,b).

e. Comparison with other datasets

To compare our results with other data we selected studies for which averages were calculated over at least a year (although there are other studies available with data from shorter time spans). Also, all data were taken at single points, over either land or sea, and for the shortwave radiation they were averaged over all values of the solar zenith angle. The available nine studies with the addition of our own results are listed in Table 1. Data comprise seasonal averages. We adhere to the Northern Hemispheric seasonal indications. Therefore winter corresponds to December–February, spring is March–May, summer is June–August, and autumn is September–November. The studies are grouped in three climatic regions, namely Arctic, midlatitude, and (sub) tropical. Our results are comparable to those of the Great Plains (Dong et al. 2006) and Spain (Salgueiro et al. 2014), although the latter only shows shortwave results. In autumn and winter, the three studies show similar results for CRF-SW. The CRF-SW of Salgueiro et al. (2014) is smaller than either of the other two. For CRF-LW our results are larger than those from the Great Plains. A plausible reason is that the fractional cloudiness over the Great Plains is lower than at Cabauw by 0.1 or more. Large values of CRF-LW (40–50 W m^{-2}) can be found in the subtropics (Ghate et al. 2009; Miller et al. 2012). In the tropics CRF-LW becomes smaller than 20 W m^{-2} . It is surprising to find that large negative values of CRF-SW (up to -80 W m^{-2}) can be found in the Arctic region (Dong et al. 2010; Intrieri et al. 2002), although in the winter values are small.

Table 2 shows an inventory of values of the parameter a from Eq. (9). This parameter constitutes the average difference of the cloud base downwelling longwave radiative flux and the downwelling clear-sky longwave radiative flux (averages are taken over all seasons). Except for the Miller et al. (2015) study from Greenland, it appears that this value does not appear to be much dependent on the climatic regime. A value of 60–70 W m^{-2} seems appropriate.

TABLE 1. Summary of cloud forcing data from the literature, including those from this paper. All data are averages over at least one year and are taken at a single point, over either land or sea. The units of the numbers quoted are watts per meter squared. The “~” indicates data for which no tabular information was available in the paper. In that case the numbers were inferred from the available graphs and are thus approximate.

Reference	Where	How long	Winter LW	Spring LW	Summer LW	Autumn LW	Winter SW	Spring SW	Summer SW	Autumn SW	Winter total	Spring total	Summer total	Autumn total
Dong et al. 2010.	Barrow, AK	10 yr: 1998–2008	9.6	16.5	53.8	42.8	0.3	−7.2	−80.0	−17.7	9.9	9.2	−26.2	25.1
Miller et al. 2015	Summit, Greenland	1.5 yr: 2011–13	~40	~35	~50	~40	0	~−10	~−15	~−5	~40	~25	~35	~35
Intrieri et al. 2002	Arctic (Canada–United States) SHEBA	1 yr: 1997–98	~30	~40	~50	~30	0	~−30	~−60	~−50	~30	~30	~−4	~30
Midlatitudes														
This paper														
Dong et al. 2006	Cabauw, Netherlands	9 yr: 2009–17	48.3	34.4	30.8	38.7	−23.6	−60.9	−67.8	−32.9	24.7	−26.5	−37.0	5.8
Salgueiro et al. 2014	Southern Great Plains	6 yr: 1997–2002	27.1	26.4	12.3	19.8	−34.0	−61.3	−37.1	−33.4	−6.9	−34.9	−24.8	−13.6
	Evora, Spain	8 yr: 2003–10					−26.6	−45.5	−18.8	−28.4				
Subtropics and tropics														
Fairall et al. 2008	Equatorial Pacific between 8°S and 12°N	4 yr: 1999–2002		11.8		28.6		−49.7		−71.6		−37.9		−43.0
Ghate et al. 2009	Southeastern Pacific: 20°S, 85°W	5 yr: 2001–05	~45	~40	~50	~52	~−110	~−70	~−75	~−120	~−65	~−30	~−25	−68.0
Miller et al. 2012	West African Sahel	2006	~35	~30	~40	~40	~−20	~−40	~−80	~−40	~15	~−10	~−40	~0
McFarlane et al. 2013	Nauru, Pacific (totals only)	11 yr: 1999–2010		LW: 12.3				SW: −65.0						
	Manus, Papua New Guinea (totals only)	11 yr: 1999–2010		LW: 15.1				SW: −94.7						
	Darwin, Australia (totals only)	9 yr: 2002–10		LW: 12.4				SW: −59.6						

TABLE 2. Summary of the value of a [Eq. (9)], which is the average difference between the cloud base longwave downwelling radiation and the clear-sky longwave downwelling radiation.

Reference	Where	W m^{-2}
Dong et al. 2010	Barrow	80
Miller et al. 2015	Summit	30
Shupe and Intrieri 2004	SHEBA	60
This paper	Cabauw	68
Dong et al. 2006	Southern Great Plains	62
Fairall et al. 2008	Equatorial Pacific between 8°S and 12°N	60
Kalisch and Macke 2012	Parameterized	50–70

6. Discussion and conclusions

In this study we constructed a 9-yr time series of cloud radiative forcing and analyzed seasonal changes and its relationship to cloud and temperature parameters. The data were also compared with output from an RCM. CRF-LW is positive, ranging from 50 W m^{-2} in winter to 25 W m^{-2} in summer. CRF-SW is always negative, ranging from -10 W m^{-2} in winter to -90 W m^{-2} in summer. Consequently, total CRF has a large seasonal amplitude with small positive values in winter and large negative values in summer. The seasonal amplitudes can be well explained by the latitude of Cabauw. There, the large seasonal change in sun angle is responsible for the strong seasonal variation in CRF-SW. In fact in winter CRF-SW is so low that the total CRF becomes positive as CRF-LW is always positive. The range of CRF values computed for Cabauw corresponds well to results from other temperate climate regimes such as the climate on the Southern Great Plains (Dong et al. 2006). The observations of the vertical distribution of clouds indicate that only few high clouds exist by themselves. The majority of high clouds occur in combination with low clouds so that their influence on CRF is small. Also, higher clouds have a smaller absolute value of CRF than lower clouds because low-level water vapor will swamp the CRF-LW signal of high clouds, and the higher transparencies of cirrus clouds will decrease the CRF-SW signal relative to that of low clouds. CRF-LW is dependent on fractional cloudiness and temperature and a functional relationships between these parameters was derived. CRF-SW is a function of solar zenith angle which partly shields its dependence on fractional cloudiness. However, when the CRF-SW is normalized by the cosine of the solar zenith angle its dependence on fractional cloudiness reveals itself.

The analysis in this paper focuses on explaining monthly CRF variations from seasonal variations in the four radiation components and total cloud cover. It is clear that some variability remains unexplained, especially for the SW forcing. In the midlatitude sea climate of CESAR it

was found that the vertical distribution of cloud fraction does not vary much over the seasons. Thus its contribution to seasonal CRF variability is expected to be much smaller than the contribution of total cloud fraction. Other factors may contribute to the variability in CRF like the diurnal variation in summertime convective clouds and variations in cloud type.

The observations were compared with the output from the KNMI-RACMO RCM. The model has a small cold bias with respect to the observations, in particular in winter. This means that most LW fluxes are biased low with respect to the observations in wintertime. Nevertheless, it was found that the modeled CRF-LW corresponds very well to the observations. The reason is that cloud forcing is a function that is the difference between two terms and constant biases in both terms will not affect the difference itself. RACMO employs a comparatively low surface albedo for the grid cell at Cabauw and a comparatively high clear-sky shortwave radiation, the latter of which is due to a prescribed low aerosol content. The result is a small negative bias in CRF-SW in summer that increases to larger negative values in winter (up to 10 W m^{-2}).

RACMO cloud fraction is larger than that of NUB but smaller than that of Cloudnet. Despite its cold bias with respect to the observations, RACMO integrated water vapor corresponds very well to the observations. The result is that RACMO cloud fraction at low altitudes in the troposphere is equal to that of the Cloudnet observations as is shown in Fig. 2. However, at higher altitudes in the atmosphere RACMO cloud cover is lower than the Cloudnet observations. Even so, for RACMO the same conclusion applies as for the observations: Only a few percent of high clouds occur by themselves; the majority of high clouds occur in combination with low clouds. So, the conclusion that the observed CRF predominantly originates from low clouds also applies to the RACMO CRF.

The time period over which data at Cabauw were taken was 9 years. This is not enough to constitute a climatological relevant time span. However, because meteorological observations at the Cabauw research site are to continue for at least another 10 years because of the recent start of the Ruysdael Observatory (<http://www.ruysdael-observatory.nl>), it will be possible to continue these observations into the foreseeable future so that trends eventually may be obtained.

APPENDIX

Determining the Virtual Clear-Sky Longwave Radiation

It is in principle possible to apply a straight interpolation of the longwave fluxes between clear-sky periods.

However, when the time period between clear-sky periods is long (i.e., longer than 24 h) a straight longwave flux interpolation is not able to account for changes in temperature on time scales shorter than 24 h. Therefore we identified all clear-sky points on the basis of NUB or TSI. For all of the clear-sky points the effective emissivity ε_{eff} was calculated from the actually observed values of temperature at 2 m and downward longwave fluxes:

$$\varepsilon_{\text{eff}} = \frac{\text{LW}_{\downarrow, \text{clear-sky}}}{\sigma T_{2\text{m}}^4}. \quad (\text{A1})$$

The effective emissivity was linearly interpolated between the clear-sky points to obtain the emissivity values at all other data points. The downward flux under clear sky could then be evaluated for all remaining data points (i.e., using the actually observed temperatures). Here we make use of the fact that the time variation of ε_{eff} is dominated by the relatively slow variations in IWV. However, some more rapid variation in emissivity due to variations in the 2-m temperature was present as well. The root cause is that the 2-m temperature has a strong diurnal cycle. In other words, it is likely that the 2-m temperature is not the “right” temperature because it sets off spurious short-term temporal variations in effective emissivity. Even though the 2-m temperature is used to account for this extra variation in the emissivity, its influence was deemed too strong as evidenced by unacceptably large fluctuations in the downward flux under conditions with low fractional cloudiness. Therefore we investigated the application of temperature values at the higher tower levels (i.e., at 10, 20, 40, 80, 140, and 200 m), that is, replacing the subscript “2m” in Eq. (A1) by other levels. It was found that the 80-m level provided the optimum level to account for the temperature variations in the effective emissivity.

Therefore, we repeated the emissivity interpolation procedure using the temperature of the 80-m tower level to compute the virtual clear-sky downwelling longwave radiation at the remaining (partly) cloudy points. This procedure is empirical but is based on a careful examination of the data files. It uses supplemental tower data that are normally not available elsewhere. However, the fact that the 80-m level is—in our findings—a better level to use when calculating the emissivity is possibly one reason for any lingering variability in more traditional surface (i.e., 2 m) datasets that are used to fit a parameterization constant of the type of Eq. (A1).

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