Decomposing Shortwave Top-of-Atmosphere and Surface Radiative Flux Variations in Terms of Surface and Atmospheric Contributions

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ABSTRACT

A diagnostic tool for determining surface and atmospheric contributions to interannual variations in top-of-atmosphere (TOA) reflected shortwave (SW) and net downward SW surface radiative fluxes is introduced. The method requires only upward and downward radiative fluxes at the TOA and surface as input and therefore can readily be applied to both satellite-derived and model-generated radiative fluxes. Observations from the Clouds and the Earth's Radiant Energy System (CERES) Energy Balanced and Filled (EBAF) Edition 4.0 product show that 81% of the monthly variability in global mean reflected SW TOA flux anomalies is associated with atmospheric variations (mainly clouds), 6% is from surface variations, and 13% is from atmosphere–surface covariability. Over the Arctic Ocean, most of the variability in both reflected SW TOA flux and net downward SW surface flux anomalies is explained by variations in sea ice and cloud fraction alone ($r^2 = 0.94$). Compared to CERES, variability in two reanalyses—the ECMWF interim reanalysis (ERA-Interim) and NASA's Modern-Era Retrospective Analysis for Research and Applications, version 2 (MERRA-2)—show large differences in the regional distribution of variance for both the atmospheric and surface contributions to anomalies in net downward SW surface flux. For MERRA-2 the atmospheric contribution is 17% too large compared to CERES while ERA-Interim underestimates the variance by 15%. The difference is mainly due to how cloud variations are represented in the reanalyses. The overall surface contribution in both ERA-Interim and MERRA-2 is smaller than CERES EBAF by 15% for ERA-Interim and 58% for MERRA-2, highlighting limitations of the reanalyses in representing surface albedo variations and their influence on SW radiative fluxes.

1. Introduction

Because Earth's radiation budget (ERB) is such a fundamental part of the climate system, observations of ERB play a central role in global climate and numerical

weather prediction model development and evaluation. Most model evaluation analyses either compare climatological mean top-of-atmosphere (TOA) fluxes with observations in order to ensure the model adequately captures the mean state of the ERB, or they compare interannual variations in ERB in order to ensure the model correctly represents internal climate variability. Arguably, both are necessary in order to gain confidence that the model physics is representative of the real world and implemented properly. Model–observation climatology comparisons have been used to validate Earth

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system models (Park et al. 2014; Zhao et al. 2018), regional models (Lim et al. 2014), atmospheric reanalysis products (Zhang et al. 2016; Schmeisser et al. 2018), and specific model parameterizations, including ice cloud parameterizations (Baran et al. 2016; Eidhammer et al. 2017; Chern et al. 2016), cloud microphysics schemes (Gettelman et al. 2015), low cloud parameterizations (Cheng and Xu 2015; Qin et al. 2018; Song et al. 2018), mixed-phase cloud parameterizations (Furtado et al. 2016), parameterizations of deep convection (Boyle et al. 2015; Guo et al. 2014; Wang and Zhang 2016), and parameterizations of subgrid-scale cloud water content variability (Hill et al. 2015).

While fewer in number, comparisons between model and observed interannual variability in ERB have provided useful insight (Wong et al. 2006; Allan et al. 2014; Trenberth et al. 2014; Kolly and Huang 2018). In the shortwave (SW) region over dark surfaces (e.g., ocean), these model–observation comparisons test the model representation of cloud variations, particularly when the models are forced with observed sea surface temperature and sea ice boundary conditions [e.g., Atmospheric Model Intercomparison Project (AMIP) simulations]. However, over bright surfaces, the underlying reason for model–observed differences is often less obvious since the reflected SW TOA flux variations can either be due to cloud or surface (or both) variations. Unscrambling these requires using other observations that independently describe the cloud or surface conditions. In some cases, this may require that instrument simulators be used in the model in order to ensure the comparisons account for possible differences in how clouds are defined in models and observations.

This study uses a novel approach for comparing models and observations that enables separate atmospheric and surface contributions to reflected SW TOA and net downward SW surface radiative flux variations to be quantified. The methodology uses only upward and downward TOA and surface radiative fluxes, enabling its application to readily available satellite, reanalysis and climate model output without the need for introducing more complicated instrument simulators into the model. In section 2, we describe the methodology. This is followed by a summary of the data used in section 3, and a demonstration of the methodology as a tool for understanding ERB variability in observations and for reanalysis evaluation in section 4. Here we focus on comparisons between satellite observations and output from two reanalyses but note that the methodology can also be applied to output from any model providing TOA and surface radiative fluxes. A discussion is provided in section 5. Section 6 presents the summary and conclusions.

FIG. 1. Schematic diagram of single-layer reflecting system.
[Adapted from Stephens et al. (2015).]

2. Methodology

Stephens et al. (2015) describe a single-layer approximation for a reflecting system illuminated by a solar flux S consisting of a single scattering and absorbing atmospheric layer with reflectivity r and transmissivity t above a reflecting surface with albedo α (Fig. 1). The system reflectance and transmittance are given by $R = F_T^\uparrow/S$ and $T = F_S^\downarrow/S$, respectively, where $F_T^\uparrow$ is the upward flux at the TOA and $F_S^\downarrow$ is the downward flux at the surface, and the surface albedo is given by $\alpha = F_S^\uparrow/F_S^\downarrow$, where $F_S^\uparrow$ is the upward flux at the surface. The upward flux at the TOA is approximated by the sum of reflection by the atmosphere and multiple reflections between the surface and atmosphere (Fig. 1):

$$F_T^\uparrow = rS + tF_S^\uparrow. \quad (1)$$

Similarly, the downward flux at the surface is given by

$$F_S^\downarrow = tS + rF_S^\uparrow. \quad (2)$$

The system transmittance and reflectance can be expressed as

$$T = \frac{t}{1 - r\alpha} \quad (3)$$

$$R = r + \frac{t\alpha t}{1 - r\alpha} \quad (4)$$

and the intrinsic reflection and transmission of the atmosphere layer are given by

$$t = T \frac{1 - \alpha R}{1 - \alpha^2 T^2} \quad (5)$$

$$r = R - t\alpha T. \quad (6)$$

Equations (1)–(6) provide a means of separating the atmosphere-only and atmosphere–surface contributions

to radiation at the TOA and surface in terms of radiative fluxes derived from observations. This approach assumes r and t are independent of the direction of the radiation (i.e., radiation coming from the top and bottom of the atmosphere), which is a reasonable approximation for broadband radiation (Stephens et al. 2015). The formulation is used in Stephens et al. (2015) with CERES Energy Balanced and Filled (EBAF) TOA and surface fluxes (Loeb et al. 2009; Kato et al. 2013) to examine global, annually averaged atmosphere and atmosphere–surface contributions to hemispheric differences in TOA flux and the regional distribution of reflected flux at the TOA. They also use this approach to diagnose the annual cycle of globally averaged albedo under both clear and all-sky conditions. More recently, Sledd and L’Ecuyer (2019) use the methodology of Donohoe and Battisti (2011) to isolate atmospheric and surface contributions to mean TOA albedo over the Arctic.

Here we extend the Stephens et al. (2015) formulation to provide atmosphere-only and surface-only contributions to the variability in TOA and surface radiative fluxes. This requires an expression for the anomaly of the product of two variables. We define a deseasonalized monthly anomaly δx in variable x as the deviation from its monthly climatological mean $\delta x = x - \bar{x}$. Consider a second variable y with anomaly $\delta y = y - \bar{y}$. The anomaly in the product xy is

$$\delta(xy) = xy - \overline{xy}. \quad (7)$$

Since

$$xy = (\bar{x} + \delta x)(\bar{y} + \delta y) = \bar{x}\bar{y} + \bar{x}\delta y + \delta x\bar{y} + \delta x\delta y \quad (8)$$

and

$$\overline{xy} = \bar{x}\bar{y} + \overline{\delta x\delta y}, \quad (9)$$

it follows that

$$\delta(xy) = \bar{x}\delta y + \delta x\bar{y} + \delta x\delta y - \overline{\delta x\delta y}. \quad (10)$$

From Eq. (1), the anomaly in TOA flux can then be expanded as follows:

$$\begin{aligned} \delta F_T^\uparrow = & [\bar{r}\delta S + \delta r\bar{S} + \delta r\delta S - \overline{\delta r\delta S}] + [\bar{t}\delta F_S^\uparrow \\ & + \delta t\bar{F}_S^\uparrow + \delta t\delta F_S^\uparrow - \overline{\delta t\delta F_S^\uparrow}]. \end{aligned} \quad (11)$$

Similarly, from Eq. (2), the downward radiation at the surface is

$$\begin{aligned} \delta F_S^\downarrow = & [\bar{t}\delta S + \delta t\bar{S} + \delta t\delta S - \overline{\delta t\delta S}] + [\bar{r}\delta F_S^\uparrow \\ & + \delta r\bar{F}_S^\uparrow + \delta r\delta F_S^\uparrow - \overline{\delta r\delta F_S^\uparrow}]. \end{aligned} \quad (12)$$

The terms within the first set of square brackets on right-hand sides of Eqs. (11) and (12) are associated with variations in the atmosphere and sun and their covariability, and are independent of the surface. Terms within the second set of square brackets involve surface–atmosphere interactions. Note that $\delta F_S^\uparrow$ depends upon variations in both α and $F_S^\downarrow$. To determine the atmosphere-only (ATM) contribution to the anomaly, we first evaluate Eqs. (11) and (12) using a climatological surface albedo ($\bar{\alpha}$):

$$\begin{aligned} \delta F_T^\uparrow(\text{ATM}) = & [\bar{r}\delta S + \delta r\bar{S} + \delta r\delta S - \overline{\delta r\delta S}] \\ & + [\bar{t}\delta G_S^\uparrow + \delta t\bar{G}_S^\uparrow + \delta t\delta G_S^\uparrow \\ & - \overline{\delta t\delta G_S^\uparrow}] \end{aligned} \quad (13)$$

and

$$\begin{aligned} \delta F_S^\downarrow(\text{ATM}) = & [\bar{t}\delta S + \delta t\bar{S} + \delta t\delta S - \overline{\delta t\delta S}] \\ & + [\bar{r}\delta G_S^\uparrow + \delta r\bar{G}_S^\uparrow + \delta r\delta G_S^\uparrow - \overline{\delta r\delta G_S^\uparrow}], \end{aligned} \quad (14)$$

where $G_S^\uparrow$ is the same as $F_S^\uparrow$ but determined using $\bar{\alpha}$, which is calculated from the ratio of upward and downward climatological surface fluxes: $\bar{\alpha} = \bar{F}_S^\uparrow / \bar{F}_S^\downarrow$. To ensure $\delta G_S^\uparrow$ is only influenced by atmospheric variations, we compute $G_S^\uparrow$ as follows [based on Eq. (3)]:

$$G_S^\uparrow = \left(\frac{t}{1 - r\bar{\alpha}} \right) S \quad (15)$$

and $G_S^\uparrow$ from

$$G_S^\uparrow = \bar{\alpha} G_S^\downarrow. \quad (16)$$

The surface-only (SFC) contribution is then obtained from the differences between the observed $\delta F_T^\uparrow$ and $\delta F_S^\downarrow$ in Eqs. (11) and (12) and their respective counterparts in Eqs. (13) and (14). Accordingly, the surface contribution to TOA reflected and downward surface flux anomalies is expressed as follows:

$$\begin{aligned} \delta F_T^\uparrow(\text{SFC}) = & \bar{t}(\delta F_S^\uparrow - \delta G_S^\uparrow) + \delta t(\bar{F}_S^\uparrow - \bar{G}_S^\uparrow) \\ & + \delta t(\delta F_S^\uparrow - \delta G_S^\uparrow) - (\overline{\delta t\delta F_S^\uparrow} - \overline{\delta t\delta G_S^\uparrow}) \end{aligned} \quad (17)$$

and

$$\begin{aligned} \delta F_S^\downarrow(\text{SFC}) = & \bar{r}(\delta F_S^\uparrow - \delta G_S^\uparrow) + \delta r(\bar{F}_S^\uparrow - \bar{G}_S^\uparrow) \\ & + \delta r(\delta F_S^\uparrow - \delta G_S^\uparrow) - (\overline{\delta r\delta F_S^\uparrow} - \overline{\delta r\delta G_S^\uparrow}). \end{aligned} \quad (18)$$

find that EBAF and surface-observed mean downwelling radiative fluxes agree to within the measurement uncertainty of the ground observations and closely track one another at monthly and interannual time scales. However, at high latitudes the number of ground sites is limited. Since the EBAF SFC SW fluxes depend strongly upon cloud information from MODIS, it is useful to compare MODIS-based cloud information over the Arctic with active instruments such as *Cloud–Aerosol Lidar and Infrared Pathfinder Satellite Observations* (CALIPSO) data (Winker et al. 2010). Figures 3a–c compare MODIS and CALIPSO cloud fractions for daytime ocean north of 70°N during boreal summer months between 2006 and 2016. The MODIS results are produced using the CERES cloud mask applied to MODIS-Aqua (Q. Trepte et al. 2018, manuscript submitted to *IEEE Trans. Geosci. Remote Sens.*) for viewing zenith angles $<5^\circ$. The CALIPSO cloud fractions are determined from the CALIPSO Vertical Feature Mask (VFM) product using only cloud detections with the highest confidence according to the VFM quality assurance flag. In addition, clouds only detectable by CALIPSO at an 80-km horizontal resolution are omitted. As evident from Figs. 3a–c, the MODIS cloud fractions closely track those from CALIPSO from one year to the next, and the overall mean MODIS cloud fractions are within 0.01 of the CALIPSO mean values for each of the 3 months considered. In the absence of ground-based SW radiation measurements, this excellent agreement is encouraging and provides some validation of the inputs used in the EBAF-SFC calculations.

Other datasets used in this study include snow/ice maps from a combination of the National Snow and Ice Data Center (NSIDC) Near-Real Time Snow and Ice Extent (NISE) product (Brodzik and Stewart 2016), the National Environmental Satellite, Data, and Information Service (NESDIS) snow/ice map, and the CERES team's snow and sea ice fraction over the clear portions of CERES footprints (Minnis et al. 2008). The NISE product is based upon microwave imager data. It does not provide retrievals within 50 km of coastlines. The NESDIS product uses imager data to identify snow and sea ice near the coast. The CERES team's snow and sea ice fraction uses MODIS reflectances at 1.6 and 0.65 μm , and the brightness temperature differences between 3.7 and 11 μm (Minnis et al. 2008). Also considered are maps of monthly mean cloud fraction from EBAF-TOA Ed4.0 (Loeb et al. 2018a), which uses the CERES cloud mask (Q. Trepte et al. 2018, manuscript submitted to *IEEE Trans. Geosci. Remote Sens.*) applied to Terra-MODIS for March 2000–June 2002 and the average of Terra-MODIS and Aqua-MODIS from July 2002 onward.

FIG. 3. Daytime ocean cloud fraction for 70°–90°N from MODIS Aqua (CERES SSF edition 4) for viewing zenith angles $<5^\circ$ and CALIPSO VFM for high confidence clouds with clouds only detectable with 80-km resolution removed.

FIG. 4. Deseasonalized monthly anomalies in global mean (a) reflected SW TOA flux and (b) net downward SW surface flux (Total) together with the atmospheric (ATM) and surface (SFC) contributions to the total anomaly. Also shown are the SFC contribution to (c) reflected SW TOA flux and (d) net downward SW surface flux anomalies for equal-area latitude zones. Anomalies in each zone are normalized such that the sum of the zonal anomalies equals the global mean anomaly.

Monthly TOA reflected SW and surface radiative flux output from two reanalyses, MERRA-2 and ERA-Interim, are considered. MERRA-2 (Gelaro et al. 2017), produced by NASA's Global Modeling and Assimilation Office (GMAO), uses an atmospheric model comprising a finite-volume dynamical core with a resolution of $0.5^\circ \times 0.625^\circ$ and 72 vertical levels. MERRA-2 uses a 3DVAR algorithm based upon the gridpoint statistical interpolation (GSI) analysis system with 6-hourly updates. Clouds in MERRA-2 are generated internally and their radiative properties are parameterized. The radiative transfer code is based upon Chou and Suarez (1999) for SW radiation and Chou et al. (2001) for LW radiation. Sea surface temperature (SST) and sea ice concentration in MERRA-2 are based upon daily $1/4^\circ$ data from Reynolds et al. (2007) between 1982 and March 2006, and daily $1/20^\circ$ data from the Operational Sea Surface Temperature and Sea Ice Analysis (OSTIA) from Donlon et al. (2012) for April 2006 onward (Bosilovich et al. 2015). New glaciated land representation and seasonally varying sea ice albedo have been implemented in MERRA-2, leading to improved air temperatures and reduced biases in the net energy flux over these surfaces (Cullather et al. 2014).

ERA-Interim (Dee et al. 2011) is a global atmospheric reanalysis produced by the European Centre for Medium-Range Weather Forecasts (ECMWF). It consists of a global numerical weather prediction system called the Integrated Forecast System (IFS) coupled to a

4D-Var data assimilation system. The model uses a T255 grid and data are available at $0.75^\circ \times 0.75^\circ$ at 60 vertical levels. ERA-Interim uses a prognostic cloud scheme with the maximum-random overlap assumption. Radiative transfer calculations in ERA-Interim use Rapid Radiative Transfer Model (RRTM; Mlawer et al. 1997). SST and sea ice concentration are based upon various products for different time periods (see Table 1 in Dee et al. 2011). Over the period considered in this study, the main sources are from the NCEP Real-Time Global sea surface temperature (NCEP RTG) and OSTIA.

4. Results

a. Variability in global and zonal mean reflected SW TOA and net downward SW surface flux

Deseasonalized monthly anomalies in global mean reflected SW TOA flux and net downward SW surface flux are decomposed in terms ATM and SFC contributions in Figs. 4a and 4b, respectively. Substantial interannual variability is observed with standard deviations of 0.64 W m^{-2} for reflected SW TOA flux and 0.78 W m^{-2} for net downward SW surface flux. Anomalies after 2014 significantly exceed the 1σ values following a shift in the sign of the Pacific decadal oscillation index from negative to positive in spring 2014 and the major 2015/16 El Niño event (Loeb et al. 2018b). During this period there is a large decrease in reflected

SW TOA flux and an increase in net downward SW surface flux that is primarily associated with a decrease in marine stratus and stratocumulus in the eastern Pacific Ocean in response to warm sea surface temperatures (Loeb et al. 2018b). Early in 2017 the global anomalies reach -2 and 2.5 W m^{-2} for reflected SW TOA flux and net downward SW surface flux, respectively, but return to values close to the climatological means by 2018.

Most of the variability in the TOA and surface radiative fluxes is associated with ATM variations. Overall, the ATM contributes 81% of the variance in reflected SW TOA flux anomalies and the SFC contributes 6%. The remaining 13% is associated with covariability between ATM and SFC. For net downward SW surface flux, the breakdown is similar, with ATM accounting for 83%, SFC for 8%, and covariability for 9%. As will be discussed in sections 4b and 4d, the main source of ATM variations is due to cloud property variations. For the SFC contribution, variations are primarily due to variations in sea ice and snow cover in the mid- and high latitudes.

To quantify the SFC contribution in different latitude zones to the global mean anomaly, we calculate SFC anomalies for six equal-area latitude zones (42° – 90°N , 20° – 42°N , 0° – 20°N , 0° – 20°S , 20° – 42°S , and 42° – 90°S) and divide the anomalies in each zone by six, so that the sum of all six zones equals the global mean anomaly (Figs. 4c,d). The largest SFC contributions are from the 42° – 90°N and 42° – 90°S latitude zones. In the 42° – 90°N latitude zone, appreciable anomalies occur in June 2004, June 2012, April 2013, and May 2016, whereas for 42° – 90°S anomalies are pronounced in December 2007, November 2013, December 2014, November 2016, and December 2017. The largest impact of sea ice loss for 42° – 90°N occurs in June 2012, causing a reduction in global mean reflected SW TOA flux of 0.34 W m^{-2} and an increase in global mean net downward SW surface flux of 0.55 W m^{-2} . For 42° – 90°S the largest impact occurs in November 2016, causing the global mean reflected SW TOA flux to decrease by 0.33 W m^{-2} and the global mean net downward SW surface flux to increase 0.56 W m^{-2} . These anomalies are quite significant when viewed in the context of the overall variability in global mean reflected SW TOA flux and net downward SW surface flux anomalies, which have standard deviations of 0.64 and 0.78 W m^{-2} , respectively.

The ATM and SFC contributions to the variance in reflected SW TOA and net downward SW surface flux anomalies for each of the latitude zones is shown in Figs. 5a and 5b (note that here anomalies are not scaled by a factor of 6). As is the case for the global mean results in Figs. 4a and 4b, the ATM contribution

FIG. 5. ATM, SFC, and covariance contributions to the total variance in (a) reflected SW TOA and (b) net downward SW surface flux anomalies by latitude.

dominates in each latitude zone. We note that this is only true for the specific latitudes considered here. For net downward SW surface flux, the SFC contribution exceeds ATM for zonal means poleward of 70° (not shown). Not surprisingly, the greatest variability occurs in the tropics between 0° and 20°N , which contains the mean position of the intertropical convergence zone. There is larger variance in reflected SW TOA flux in the Southern Hemisphere subtropics (20° – 42°S) than in the Northern Hemisphere subtropics, whereas the variance in these two latitude zones is approximately the same for net downward SW surface flux. Poleward of 42° , reflected SW TOA and net downward SW surface flux contributions for ATM and SFC are larger in the Northern Hemisphere.

FIG. 6. Regional distribution of the variance in June anomalies of (a) the SFC contribution to net downward SW surface flux, (b) sea ice fraction, (c) the ATM contribution to reflected SW TOA flux (ATMr), and (d) cloud fraction.

b. Observed variations over the Arctic Ocean

Over the Arctic Ocean, the variance in the SFC contribution to net downward SW surface flux has a spatial distribution that closely resembles that of sea ice fraction (Figs. 6a,b). Maxima in these quantities occur over the Barents, Kara, Laptev, Chukchi, and Beaufort Seas, as well as the Hudson and Baffin Bays. These areas are well known to exhibit substantial year-to-year variability in sea ice fraction. The variance in the reflected ATM contribution to reflected SW TOA flux shows a similar spatial pattern to that for cloud fraction (Figs. 6c,d). The largest variance occurs over a large area between 120°W and 150°E and Hudson Bay.

The close correspondence in variability between the SFC contribution to net downward surface flux and sea ice fraction and the reflected ATM contribution to TOA flux and cloud fraction are also apparent when considering JJA anomalies averaged over the Arctic Ocean (Figs. 7a–d). The decreasing trend in sea ice fraction

during the CERES period (Fig. 7d) is associated with an increasing trend in net downward SW surface flux (Fig. 7b). A regression fit between anomalies in these two quantities produces a slope of -1.08 W m^{-2} per percent change in sea ice fraction, with an r^2 of 0.94 (Fig. 8d). In contrast, while there is no trend in the ATM contribution to net downward SW surface flux (Fig. 7b) and cloud fraction (Fig. 7c), year-to-year variations are well correlated (Fig. 8c), with an r^2 of 0.73. Similarly, anomalies in the SFC and ATM contributions to reflected SW TOA flux (Fig. 7a) closely track anomalies in sea ice and cloud fraction (Figs. 8a,b). Based upon these results, we find that most of the variability in both reflected SW TOA flux and net downward SW surface flux anomalies over the Arctic Ocean can be explained by variations in sea ice and cloud fraction, respectively. This is illustrated in Figs. 9a and 9b, which compare EBAF reflected SW TOA and net downward SW surface anomalies with those determined from combined SFC and ATM regression fits in Figs. 8a–d.

FIG. 7. Ocean-only 70°–90°N JJA anomalies in (a) reflected SW TOA flux and (b) SW net downward flux at the surface and corresponding atmospheric (ATM) and surface (SFC) contributions. Anomalies in (c) cloud fraction and (d) sea ice fraction.

c. Comparisons with reanalysis over the Arctic

Applying the approach described in section 2, we use TOA and surface radiative fluxes from ERA-Interim and MERRA-2 to calculate annual mean anomalies (defined for March–February) in ATM and SFC contributions to net downward SW surface flux. Figures 10a–c compare the results with CERES EBAF Ed4.0 for 70°–90°N over ocean. For the total net downward SW surface flux anomalies (Fig. 10a) and SFC contribution (Fig. 10c), MERRA-2 provides a better representation of interannual variations than ERA-Interim when compared to CERES EBAF. The correlation coefficient between MERRA-2 and CERES EBAF is 0.83 in Fig. 10a, while it is 0.70 for ERA-Interim. ERA-Interim underestimates the local maximum in net downward SW surface flux in 2007 and 2011, while MERRA-2 tracks the timing of maxima and minima observed in CERES EBAF, but underestimates the magnitude of the maxima in 2011 and 2016. Nevertheless, while interannual variations are reasonably well captured in MERRA-2, the overall upward trend in the SFC contribution is underestimated compared to CERES. The trend in the SFC contribution to net downward SW surface flux in CERES EBAF is $2.4 \pm 1.4 \text{ Wm}^{-2} \text{ decade}^{-1}$, compared to $0.84 \pm 0.7 \text{ Wm}^{-2} \text{ decade}^{-1}$ for MERRA-2 (trend uncertainty at 95% confidence level). In contrast, ERA-

Interim is in better agreement with CERES EBAF, with a trend of $2.1 \pm 0.7 \text{ Wm}^{-2} \text{ decade}^{-1}$.

It is interesting to note that while there is a large minimum in sea ice fraction in 2012 (Fig. 7d), the total net downward SW surface flux anomaly peaks in 2011. Even though the SFC contribution is largest in 2012, the combined effects of positive SFC and ATM contributions (Figs. 10a–c) mean that the total anomaly is larger in 2011 than 2012. MERRA-2 captures this, but underestimates the SFC contribution in 2011 (Fig. 10c). ERA-Interim misses it entirely, so that its local maximum in net downward SW surface flux occurs in 2012 instead of 2011.

The regional distribution of net downward SW surface flux anomalies and ATM and SFC contributions for JJA 2011 for CERES EBAF, ERA-Interim, and MERRA-2 are shown in Figs. 11a–i. In CERES EBAF, anomalies over the Arctic Ocean are generally greater than those in ERA-Interim and MERRA-2 (Figs. 11a,d,g). Local maxima in the SFC contribution over the Kara, Laptev, and Chukchi Seas are captured in all three cases (Figs. 11c,f,i), as are local minima along both coasts of Greenland in the Greenland Sea and Baffin Bay, but both ERA-Interim and MERRA-2 miss anomalies of $10\text{--}20 \text{ Wm}^{-2}$ over a large portion of the Arctic Ocean between 90° and 150°W. For the ATM contribution, anomaly patterns are similar but there are large differences in the magnitudes, and anomalies in the two

FIG. 8. Regression fits for (a),(c) reflected SW TOA and surface SW net downward ATM contributions against cloud fraction anomalies and (b),(d) reflected SW TOA and surface SW net downward SFC contributions against sea ice fraction anomalies for ocean-only 70° – 90° N JJA.

reanalyses cover a smaller area compared to CERES EBAF. The overall weaker anomalies in the reanalyses over the Arctic are presumably due to their underestimation of the interannual variation and long-term reduction of surface albedo in the Arctic (Cao et al. 2016).

d. Regional variance in net downward SW flux contributions

The regional distribution of variance in ATM and SFC contributions to monthly anomalies in net downward SW surface flux over the globe are shown in Figs. 12a–c for CERES EBAF, Figs. 12d–f for ERA-Interim, and Figs. 12g–i for MERRA-2. The dominant contribution to net downward SW surface flux is from variance in ATM over most of the globe except at high latitudes over sea ice, where the variance contribution from SFC variations dominates. For 70° – 90° N, the SFC contribution accounts for 75% of the overall variance, compared to 21% for ATM and 4% for atmosphere–surface covariance. The ATM contribution peaks in the

tropical Pacific and Indian Ocean regions, and is mainly associated with convective clouds. Overall, MERRA-2 overestimates the ATM variance compared to CERES EBAF, while ERA-Interim underestimates it. Globally, MERRA-2 ATM variance exceeds CERES EBAF by 17% and ERA-Interim variance is lower by 15%. In contrast, the overall SFC contribution in both ERA-Interim and MERRA-2 are smaller than CERES EBAF; ERA-Interim is lower than CERES EBAF by 15% and MERRA-2 by as much as 58% (a factor of 2.4). Over North America, there is good agreement in the SFC contribution between ERA-Interim and CERES, whereas MERRA-2 shows much weaker variance. Larger discrepancies are apparent over Asia, and neither ERA-Interim nor MERRA-2 shows any variability over Australia and southern Africa. This is likely because both use prescribed surface albedos. During the CERES period, substantial variations in surface properties occurred over Australia in response to the Millennium Drought (Loeb et al. 2017). Over sea ice off

FIG. 9. Comparison between anomalies in CERES observations and regression fits for (a) reflected SW TOA flux and (b) net downward SW surface flux for ocean-only and 70°–90°N (δf_c = anomalies in cloud fraction; δf_s = anomalies in sea ice fraction).

the coast of Antarctica, MERRA-2 SFC variance is substantially smaller than CERES EBAF and ERA-Interim.

For CERES EBAF, the covariance contribution from the correlation between ATM and SFC (Figs. 12c,f,i) shows a large negative contribution over sea ice off of Antarctica. There is also a negative covariance contribution in this area for ERA-Interim, but it is much weaker. For MERRA-2, the negative covariance contribution is almost entirely missing. Negative covariance contributions are also apparent in some portions of the Arctic Ocean (Fig. 12c), but are much weaker compared to those off the coast of Antarctica.

The negative covariance between ATM and SFC over sea ice off of Antarctica suggests that when positive (negative) anomalies in atmospheric transmittance (δt) occur, surface albedo increases (decreases). Closer inspection reveals that the negative covariance contribution is associated with a strong negative summertime correlation between cloud optical depth and sea ice fraction anomalies: when sea ice fraction is higher (lower) than average, cloud optical thickness is lower (higher) than average, resulting in a negative correlation

FIG. 10. Annual anomalies (defined for March–February) in net downward SW surface flux for 70°–90°N over ocean from EBAF 4.0, ERA-Interim, and MERRA-2: (a) Total, (b) ATM contribution, and (c) SFC contribution.

between ATM and SFC contributions to net downward SW flux. This inverse relationship between sea ice fraction and cloud optical depth anomalies is shown in Fig. 13 for a $1^\circ \times 1^\circ$ region centered at 63.5°S, 48°W during the months of December, January, and February (DJF) between March 2000 and October 2017. One could interpret this as a physical relationship, whereby thicker clouds are more likely to form over open water where there is a greater supply of moisture than over sea ice. However, we cannot discount the possibility of a systematic dependence in the cloud optical depth retrieval on sea ice fraction since cloud optical depth retrievals over open ocean rely on MODIS 0.63- μm radiances, whereas MODIS 1.24- μm radiances are used when sea ice fraction over an 8×8 pixel region exceeds 25% (since 1.24 μm is less sensitive to surface reflection contributions).

To investigate the possibility that the negative correlation between cloud optical depth and sea ice fraction is due to a cloud optical depth retrieval bias, we reprocessed a subset of 13 Januaries and Februaries between 2003 and 2018 using MODIS 1.24- μm radiances to retrieve cloud optical depths over both open

FIG. 11. Net downward SW surface flux anomalies for 70°–90°N for JJA 2011 for (a)–(c) EBAF Ed4.0, (d)–(f) ERA-Interim, and (g)–(i) MERRA-2, showing (left) Total, (middle) ATM, and (right) SFC.

water and sea ice. Considering the same $1^\circ \times 1^\circ$ region as in Fig. 13, we find that $1.24\text{-}\mu\text{m}$ liquid water cloud optical depth retrievals over open water are biased low by 10% on average compared to $0.63\text{-}\mu\text{m}$ retrievals. Accounting for this relative bias and the fraction of the grid box that

is open water, we estimate how cloud optical depth would vary with sea ice fraction if all months were processed using $1.24\text{-}\mu\text{m}$ radiances over both open water and sea ice. Figure 14 compares cloud optical depth retrievals against sea ice fraction using the standard

FIG. 12. Variance in net downward SW surface flux anomalies for (a)–(c) EBAF Ed4.0, (d)–(f) ERA-Interim, and (g)–(i) MERRA-2. (left),(middle) Variance in ATM and SFC components, respectively; (right) the contribution by the covariance between ATM and SFC. The number in the bottom-left corner of each figure corresponds to the global mean.

CERES algorithm (“standard algorithm”) and our estimate of $1.24\text{-}\mu\text{m}$ cloud optical depths (“ $1.24\text{-}\mu\text{m}$ only”). For the standard algorithm, the regression slope is $-0.057\%^{-1} \pm 0.02\%^{-1}$, while for the $1.24\text{-}\mu\text{m}$ only case it is reduced to $-0.037\%^{-1} \pm 0.02\%^{-1}$ (uncertainties at the 95% significance level). Thus, despite the cloud optical depth retrieval bias, these results suggest there is still a dependence in cloud optical depth on sea ice fraction, which leads to a negative covariance between ATM and SFC.

5. Discussion

Many previous studies have used satellite and ground measurements to evaluate clouds and radiation in reanalyses. These have mainly focused on how well the reanalyses represent the climatological mean state and seasonal cycle rather than interannual variations, which

is the focus of this study. Using cloud and radiation measurements at two surface sites in the Arctic, [Zib et al. \(2012\)](#) show that reanalyses tend to overestimate downwelling SW surface radiative fluxes, although ERA-Interim is amongst the best performers with a bias of only 5 W m^{-2} at these two locations. Similar conclusions are obtained in [Decker et al. \(2012\)](#) using flux tower measurements in other regions over land. Using ground observations at Barrow (now known as Utqiagvik), Alaska, [Walsh et al. \(2009\)](#) observed that the reanalyses tend to underestimate persistent low-level cloud cover during summertime, resulting in large biases in downward SW surface flux. Conversely, in their assessment of how reanalyses represent the relative contributions of the atmosphere and surface to the mean reflected SW TOA flux, [Sledd and L'Ecuyer \(2019\)](#) find that MERRA-2 and ERA-Interim overestimate the atmospheric contribution during summer and underestimate the surface contribution

FIG. 13. Scatterplot of monthly anomalies in cloud optical depth against anomalies in sea ice fraction for a $1^\circ \times 1^\circ$ region centered at 63°S , 48°W for all December, January, and February months.

year-round. Schmeisser et al. (2018) use CERES surface radiative fluxes from EBAF-SFC (Kato et al. 2018) and cloud fractions from the CERES SYN1deg product (Rutan et al. 2015) to evaluate radiation and clouds from five reanalyses over the northeast Pacific. They find that the reanalyses tend to underestimate cloud fraction by 8%–21%, resulting in an overestimation of downward SW flux at the surface that ranges from 4 to 21 W m^{-2} . It is noteworthy that the study of Schmeisser et al. (2018) is one of the few that evaluates interannual variations in clouds and radiation. They find that models that adequately reproduce the climatology in downward SW surface flux do not necessarily capture extreme anomalies very well. Thus, a model with low bias in the climatological mean is not guaranteed to provide an accurate representation of interannual variability. As a result, use of reanalyses to determine trends in clouds and radiation (e.g., over the Arctic) is highly uncertain. This is evident from the results in the present study (Fig. 10), which shows large trend differences in net downward SW surface flux over the Arctic Ocean between MERRA-2 and ERA-Interim.

Because the EBAF-SFC radiative fluxes are constrained by TOA fluxes (Kato et al. 2018), and because the CERES TOA SW fluxes have been shown to be stable to a few tenths of a watt per meter squared (Loeb et al. 2018a), we are confident that the anomalies and trends derived from CERES EBAF Ed4.0 provide a good representation of the actual interannual variability in the climate system. To verify this Kato et al. (2018) use surface measurements at a number of land and ocean buoy locations to compare against the CERES EBAF Ed4.0 values. They find that the EBAF surface

FIG. 14. Scatterplot of monthly liquid water cloud optical depth retrievals against sea ice fraction for the standard CERES cloud algorithm ("standard algorithm") and an estimate that uses $1.24\text{-}\mu\text{m}$ radiances to infer cloud optical depths over both open water and sea ice for the $1^\circ \times 1^\circ$ region in Fig. 13.

flux anomalies track closely with the measured values. In fact, the correlation coefficient between deseasonalized anomalies in EBAF-SFC and surface observations is 0.93 or greater.

6. Summary and conclusions

The CERES TOA and surface radiative fluxes are used to derive a diagnostic tool to determine surface and atmospheric contributions to reflected SW TOA flux and net surface downward SW flux variability. The method starts by using an existing framework that approximates the Earth–atmosphere system as a single layer reflecting and absorbing atmosphere above a reflecting surface. The framework provides a means of separating the atmosphere-only and atmosphere–surface contributions to radiation at the TOA and surface in terms of radiative fluxes derived from observations or models. In this study, we extend the framework to provide a set of equations enabling calculation of atmospheric (ATM) and surface (SFC) contributions to TOA and surface radiative flux anomalies. Over highly varying snow/sea ice and cloud conditions, the main contributions to TOA and surface flux variability are associated with atmospheric reflectance and transmittance, followed by surface albedo.

At global scale, ATM contributions to reflected SW TOA flux variability account for 81% of the total variability, while the SFC contributions account for 6%. The remaining 13% is associated with covariability between ATM and SFC. The breakdown is quite similar for net

downward SW surface flux (ATM: 83%; SFC: 8%; covariability: 9%). Not surprisingly, the surface contribution is primarily from middle to high latitudes and polar regions. Overall, the surface variations over the northern middle and high latitudes (42°–90°N) are 22% more variable relative to the same latitude range in the Southern Hemisphere. Years with appreciable summertime sea ice loss can have a large impact on global reflected SW TOA and net downward SW surface flux anomalies: anomalies in the SFC contribution at high latitudes can be as high as 0.34 W m^{-2} for reflected SW TOA flux and 0.56 W m^{-2} for net downward SW surface flux (by comparison, the typical standard deviations of global mean reflected SW TOA flux and net downward SW surface flux anomalies are 0.64 and 0.78 W m^{-2} , respectively).

Over the Arctic Ocean, spatial patterns of the variance in the SFC contribution to net downward SW surface flux are highly correlated with those in sea ice fraction. Similarly, the ATM contribution to reflected SW TOA flux closely resembles that for cloud fraction. When anomalies are averaged over the Arctic Ocean, most of the variability in both reflected SW TOA flux and net downward SW surface flux anomalies is explained by variations in sea ice and cloud fraction ($r^2 = 0.94$).

When the methodology introduced here is applied to ERA-Interim and MERRA-2 over the Arctic Ocean, MERRA-2 is found to provide a better representation of CERES-observed interannual variations in the total and SFC contribution to net downward SW surface flux anomalies than ERA-Interim. However, the overall upward trend in the SFC contribution in MERRA-2 is significantly smaller than ERA-Interim, which is in good agreement with CERES. During substantial summertime sea ice loss over the Arctic Ocean in 2011, regional anomalies in the SFC contribution to net downward SW surface flux in ERA-Interim and MERRA-2 are much smaller than CERES. In large portions of the Arctic Ocean, ERA-Interim and MERRA-2 miss anomalies of $10\text{--}20 \text{ W m}^{-2}$. Regional patterns in the ATM contribution differ markedly between ERA-Interim and MERRA-2, and both show large differences compared to CERES. At global scales, the regional distribution of variance in the ATM contribution to anomalies in net downward SW surface flux is overestimated in MERRA-2 by 17% compared to CERES while ERA-Interim underestimates the variance by 15%. The overall SFC contribution in both ERA-Interim and MERRA-2 is smaller than CERES EBAF: for ERA-Interim it is lower by 15% while for MERRA-2 it is lower by 58%.

In CERES EBAF, there is a marked negative covariance contribution resulting from a negative correlation

between the ATM and SFC contributions to net downward SW surface flux over sea ice off of Antarctica. There is also a negative covariance contribution in this region for ERA-Interim, but it is much weaker than CERES. For MERRA-2 it is totally absent. The negative covariance contribution in CERES is observed to be associated with a strong negative summertime correlation between cloud optical depth and sea ice fraction anomalies. When anomalies in sea ice fraction are higher (lower) than average, cloud optical depth anomalies tend to be lower (higher) than average. It is possible that this is a physical relationship in which thicker clouds are more likely to form over open water than sea ice in this region.

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