

**Global and regional entropy production by radiation estimated from satellite
observations**

Seiji Kato¹ and Fred G. Rose²

¹ NASA Langley Research Center

² Science System & Applications Inc.

Corresponding author

Seiji Kato

seiji.kato@nasa.gov

Climate Science Branch NASA Langley Research Center

MS 420 Hampton, Virginia, 23681-2199

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24 **Abstract**

25 Vertical profiles of shortwave and longwave irradiances computed with satellite derived
26 cloud properties and temperature and humidity profiles from reanalysis are used to
27 estimate entropy production. Entropy production by shortwave is computed by the
28 absorbed irradiance within layers in the atmosphere and by the surface divided by their
29 temperatures. Similarly, entropy production by longwave is computed by emitted
30 irradiance to space from layers in the atmosphere and surface divided by their
31 temperatures. Global annual mean entropy productions by shortwave absorption and
32 longwave emission to space are, respectively, $0.852 \text{ Wm}^{-2}\text{K}^{-1}$ and $0.928 \text{ Wm}^{-2} \text{ K}^{-1}$. With a
33 steady state assumption, entropy production by irreversible processes within the Earth
34 system is estimated to be $0.076 \text{ Wm}^{-2}\text{K}^{-1}$ and by non-radiative irreversible processes to be
35 $0.049 \text{ Wm}^{-2}\text{K}^{-1}$. Both global annual mean entropy productions by shortwave absorption
36 and longwave emission to space increase with increasing shortwave absorption (i.e. with
37 decreasing the planetary albedo). The increase of entropy production by shortwave
38 absorption is, however, larger than the increase of entropy production by longwave
39 emission to space. The result implies that global annual mean entropy production by
40 irreversible processes decreases with increasing shortwave absorption. Input and output
41 temperatures derived by dividing the absorbed shortwave irradiance and emitted
42 longwave irradiance to space by respective entropy production are, respectively 282 K
43 and 259 K, which give the Carnot efficiency of the Earth system of 8.5%.

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1. Introduction

The hydrological cycle and dynamics in the earth system are driven by energy received from the Sun. Once averaged over a year and over the entire globe, about 71% of the solar irradiance is absorbed by Earth (e.g. Stephens et al. 2012). While dynamics in the atmosphere and ocean distributes energy absorbed by Earth, energy is converted into different forms. A part of shortwave irradiance absorbed by ocean and land is used to evaporate water vapor. Water vapor enters the atmosphere and heats the atmosphere when it condenses. Energy transport is also associated with heating and cooling by radiation, dynamics, and water vapor phase change, which in turn alter entropy of the earth system.

Entropy is produced by heating and cooling by irreversible processes. In addition, entropy is carried by radiation. Entropy produced by a blackbody is, therefore, the sum of entropy produced by radiative cooling and entropy carried by the blackbody radiation (Planck 1913). For the Earth system, entropy is imported by radiation emitted by the Sun. Heating due to absorption of solar radiation and scattering (Wu and Liu 2010) within the Earth system produce entropy. In addition to diabatic heating by irreversible processes occurring within the system, radiative cooling and heating by longwave radiation contribute to entropy production. Longwave radiation emitted to space then exports entropy. Earlier studies refer entropy produced by irreversible processes within the Earth system as material entropy (e.g Bannon 2015; Bannon and Lee 2017).

Global entropy budget was first estimated by Peixoto et al. (1991). Their estimate was primarily based on in situ surface and upper atmosphere data taken between 1963 and 1973 (Peixoto and Oort 1992). Stephens and O'Brien (1993) used satellites

broadband radiance measurements to estimate entropy production by longwave radiation emitted to space. Observational estimates of entropy production within the Earth system by absorption of shortwave radiation, however, have not been updated since the estimate by Peixoto et al. (1991) were made. Goody (2000) extended work by Peixoto et al. (1991) and estimated entropy produced by irreversible processes within the Earth system. Lucarini et al. (2011) used climate models and estimated entropy production by non-radiative irreversible processes. The analysis of thermodynamics of the climate system is also used to understand how general circulation changes under a warming climate (Kjellsson 2015; Laliberté et al. 2015).

In this study, we revise entropy budget estimated by earlier studies using satellite observations that have been taken since March 2000. We use cloud and aerosol properties derived from satellites to estimate entropy production by the absorption of shortwave irradiance and by the emission of longwave irradiance to space. Radiation also carries entropy but we primarily focus on entropy produced by heating and cooling within the atmosphere and at Earth's surface in this study.

Before we present entropy production estimates in Section 4, we use a 1D model to illustrate the relationship between entropy produced by radiation and entropy produced by irreversible processes within the system in Section 2. Section 3 explains data products used in this study and how entropy production is computed. Section 5 utilizes the results with the concept illustrated by the 1D model to discuss how entropy production changes when the energy input to Earth changes.

2. Conceptual 1D model

We use a 1D isothermal atmosphere model discussed by Bannon (2015) to describe the relationship between energy budget and entropy budget of the Earth climate system (**Figure 1**). The 1D model also helps to understand how entropy produced by shortwave absorption and longwave emission is related to irreversible processes within the Earth system. In addition, we extend the model to help relating entropy produced by non-radiative irreversible processes and entropy produced by heating and cooling by radiation within the Earth system.

We let the incoming solar irradiance at top-of-atmosphere (TOA) F_0 . The shortwave irradiance absorbed by the atmosphere is βF_0 where β is the absorptivity of the atmosphere. We assume that the atmosphere does not reflect shortwave irradiance but surface reflects shortwave irradiance αF_0 , where α is the reflectivity of the surface. The rest of the shortwave irradiance $(1-\alpha-\beta)F_0$ is absorbed by the surface. Excluding scattering by the atmosphere does not affect establishing the relationship between entropy produced by radiation and non-radiative processes. The surface is assumed to be black to longwave radiation and emits the longwave irradiance proportional to the fourth power of the temperature $F_{sfc} = \sigma T_{sfc}^4$, where T_{sfc} is the surface temperature. The emissivity of the atmosphere is ε so that the atmosphere emits the irradiance $F_{atm} = \varepsilon \sigma T_{atm}^4$, both downward and upward, where T_{atm} is the temperature of the isothermal atmosphere. The atmosphere absorbs the longwave irradiance εF_{sfc} emitted by the surface and the rest $(1-\varepsilon)F_{sfc}$ is transmitted to space. Energy is also transported by turbulence from the surface to the atmosphere at the rate of F_{tur} , which represents a non-radiative irreversible process in the model.

The energy balance at TOA, surface, and for the atmosphere is, respectively,

116
$$F_{TOA}^{net} = (1 - \alpha) F_0 - F_{atm} - (1 - \varepsilon) F_{sfc} \quad (1)$$

117
$$F_{sfc}^{net} = (1 - \alpha - \beta) F_0 + F_{atm} - F_{sfc} - F_{tur} \quad (2)$$

118 and

119
$$F_{atm}^{net} = \beta F_0 + \varepsilon F_{sfc} + F_{tur} - 2 F_{atm} \quad (3)$$

120 The relationship among these net energy fluxes is

121
$$F_{TOA}^{net} = F_{sfc}^{net} + F_{atm}^{net} \quad (4)$$

122 At the top of the atmosphere, the system receives entropy of radiation $J_{sun} = \frac{4}{3} \frac{F_0}{T_{sun}}$

123 by a blackbody radiation from the Sun. A 4/3 factor appears in the expression for the

124 entropy carried by radiation (denoted by J). Feistel (2011) shows that when entropy

125 exchange between two black bodies is considered, $(4/3)\sigma T^3$ is the only mathematical

126 expression for entropy carried by blackbody radiation to be consistent with the second

127 law of thermodynamics. In addition, Yourgrau et al. (1982, Section 3-9) show, using a

128 photon gas, that the 4/3 factor is necessary for the entropy carried by radiation. A simple

129 example treating both processes, entropy carried by radiation and entropy produced by

130 heating and cooling, is discussed in Section 102 of Planck (1913).

131 Reflecting the shortwave irradiance and converting direct to diffuse radiation

132 produce entropy J_{ref} (Wu and Liu 2010). Emission of the longwave irradiance to space

133 carries entropy, $J_{atm} = \frac{4}{3} \frac{F_{atm}}{T_{atm}}$ and $(1 - \varepsilon) J_{sfc} = (1 - \varepsilon) \frac{4}{3} \frac{F_{sfc}}{T_{sfc}}$. Therefore, entropy export to

134 space by radiation is

135
$$J_{TOA}^{net} = J_{ref} - J_{atm} - (1 - \varepsilon) J_{sfc} + (1 - \alpha) J_{sun} \quad (5)$$

136 Entropy produced by radiation exchange and heating at the surface is,

$$\dot{\Sigma}_{sfc} = \frac{F_{sfc}^{net}}{T_{sfc}} = \frac{(1-\alpha-\beta)F_0 + F_{atm} - F_{sfc} - F_{tur}}{T_{sfc}} \quad (6)$$

and entropy produced within the atmosphere is

$$\dot{\Sigma}_{atm} = \frac{F_{atm}^{net}}{T_{atm}} = \frac{\beta F_0 + \varepsilon F_{sfc} + F_{tur} - 2F_{atm}}{T_{atm}} \quad (7)$$

The sum of Eqs. (5), (6) and (7) is equal to the total entropy produced by the Earth

system $\dot{\Sigma}_{tot}$,

$$\dot{\Sigma}_{tot} = J_{TOA}^{net} + \dot{\Sigma}_{sfc} + \dot{\Sigma}_{atm}. \quad (8)$$

We can also express entropy production separated by individual heating and

cooling processes, absorbed shortwave irradiance $\dot{\Sigma}_{\downarrow}$, emitted longwave irradiance $\dot{\Sigma}_{emt}$,

and turbulent flux $\dot{\Sigma}_{tur}$,

$$\dot{\Sigma}_{\downarrow} = \frac{(1-\alpha-\beta)F_0}{T_{sfc}} + \frac{\beta F_0}{T_{atm}} \quad (9)$$

$$\dot{\Sigma}_{emt} = \frac{\varepsilon F_{sfc} - 2F_{atm}}{T_{atm}} + \frac{F_{atm} - F_{sfc}}{T_{sfc}} \quad (10)$$

$$\dot{\Sigma}_{tur} = \frac{F_{tur}}{T_{atm}} - \frac{F_{tur}}{T_{sfc}} \quad (11)$$

Similar to the sum of Eqs. (5), (6), and (7), the sum of Eqs. (9), (10), (11), and entropy

exported through TOA (i.e. Eq. 5) is the entropy produced by the Earth system

$$J_{TOA}^{net} + \dot{\Sigma} = J_{TOA}^{net} + \dot{\Sigma}_{\downarrow} + \dot{\Sigma}_{emt} + \dot{\Sigma}_{tur} = J_{TOA}^{net} + \frac{F_{sfc}^{net}}{T_{sfc}} + \frac{F_{atm}^{net}}{T_{atm}} \quad (12)$$

where F_{sfc}^{net} and F_{atm}^{net} are given by Eqs. (2) and (3) and J_{TOA}^{net} is given by Eq. (5). Equation

(12) states that the entropy produced by the system is the sum of the entropy carried by

radiation through the top boundary (TOA) and heating or cooling by net energy flux at

the surface and to the atmosphere. The last two terms on the right side of Eq. (12) can be written as

$$\frac{F_{sfc}^{net}}{T_{sfc}} + \frac{F_{atm}^{net}}{T_{atm}} = \left[\frac{(1-\alpha-\beta)F_0}{T_{sfc}} + \frac{\beta F_0}{T_{atm}} \right] - \dot{\mathcal{L}}$$

Terms in the first and second square brackets on the right side of Eq. (13) are, respectively, entropy produced by absorption of shortwave irradiance and by emission of longwave irradiance to space. Terms in the third bracket are entropy produced by radiation exchange between the surface and atmosphere and by the turbulent flux.

Using Eqs. (12) and (13), we can express the rate of entropy change of the system due to heating and cooling by radiation through TOA and by irreversible process within the system as

$$\frac{dS}{dt} = \dot{\Sigma} = \frac{Q_a}{T_a} - \frac{Q_e}{T_e} + \dot{\Sigma}_{irr} \quad (14)$$

Where

$$Q_a = (1-\alpha)F_0 \quad (15)$$

$$Q_e = (1-\varepsilon)F_{sfc} + F_{atm} \quad (16)$$

$$T_a = \frac{Q_a}{\frac{(1-\alpha-\beta)F_0}{T_{sfc}} + \frac{\beta F_0}{T_{atm}}} \quad (17)$$

$$T_e = \frac{Q_e}{\dot{\mathcal{L}}}$$

and

$$\dot{\Sigma}_{irr} = \frac{F_{atm} - \varepsilon F_{sfc}}{T_{sfc}} + \frac{\varepsilon F_{sfc} - F_{atm}}{T_{atm}} - \frac{F_{tur}}{T_{sfc}} + \frac{F_{tur}}{T_{atm}} \quad (19)$$

173 At a steady state when $F_{sfc}^{net} = F_{atm}^{net} = 0$, $\frac{dS}{dt} = 0$. Because of heating of ocean and land and
 174 melting ice and glacier, the global annual mean TOA net irradiance is estimated to be
 175 $0.71 \pm 0.1 \text{ Wm}^{-2}$ (Johnson et al. 2016; Loeb et al. 2018). When entropy production due to
 176 a positive net TOA irradiance can be handled by adding entropy storage term in Eq. (14)
 177 and subsumed into entropy production within the system (Bannon and Lee 2017), then, at
 178 a steady state, the difference of the entropy produced by longwave emission to space and
 179 shortwave absorption balances with entropy produced within the system is,

$$180 \quad \frac{Q_e}{T_e} - \frac{Q_a}{T_a} = \dot{\Sigma}_{irr} \quad (20)$$

181 Although $\frac{Q_e}{T_e} - \frac{Q_a}{T_a}$ equals $\dot{\Sigma}_{irr}$, vertical radiation exchange and energy flux by turbulence
 182 affect entropy production in the column, as indicated by Eq. (13), because temperature
 183 varies with height. This is different from energy balance described by Eq. (4) and makes

184 somewhat difficult to interpret regional $\frac{Q_e}{T_e} - \frac{Q_a}{T_a}$.

185 We further separate entropy produced by vertical exchange of longwave radiation
 186 and by non-radiative processes in $\dot{\Sigma}_{irr}$. Rewriting longwave terms in the third square
 187 bracket in the right side of Eq. (13) separating the entropy produced by net longwave
 188 surface and atmosphere irradiance and the entropy produced by longwave irradiance
 189 emitted to space

$$190 \quad \frac{F_{atm} - \epsilon F_{sfc}}{T_{sfc}} + \frac{\epsilon F_{sfc} - F_{atm}}{T_{atm}} = \left[\frac{F_{atm} - F_{sfc}}{T_{sfc}} + \frac{\epsilon F_{sfc} - 2F_{atm}}{T_{atm}} \right] - \dot{\Sigma}$$

The first term of the right side of Eq. (21) is the entropy produced by net longwave irradiance and the second term of the right side of Eq. (21) is the entropy produced by longwave irradiance emitted to space. When Eq. (21) is used to rewrite Eq. (14),

$$\frac{dS}{dt} = \left[\frac{(1-\alpha-\beta)F_0}{T_{sfc}} + \frac{\beta F_0}{T_{atm}} \right] + \left[\frac{F_{atm}-F_{sfc}}{T_{sfc}} + \frac{\varepsilon F_{sfc}-2F_{atm}}{T_{atm}} \right] - \frac{F_{tur}}{T_{sfc}} + \frac{F_{tur}}{T_{atm}} \quad (22)$$

At a steady state when $\frac{dS}{dt}=0$, entropy production by non-radiative irreversible process is,

$$\dot{S}_{tur} = - \left[\frac{F_{sfc,SW}^{net}}{T_{sfc}} + \frac{F_{atm,SW}^{net}}{T_{atm}} + \frac{F_{sfc,LW}^{net}}{T_{sfc}} + \frac{F_{atm,LW}^{net}}{T_{atm}} \right] \quad (23)$$

where $F_{sfc,SW}^{net}$ and $F_{atm,SW}^{net}$ are, respectively net shortwave irradiance at the surface and within the atmosphere and $F_{sfc,LW}^{net}$, and $F_{atm,LW}^{net}$ are, respectively, net longwave irradiance at the surface and within the atmosphere. Therefore, at the steady state, the sum of entropy production by shortwave absorption and longwave emission within the system and entropy production by non-radiative processes is zero. This expression is used by Lucarini et al. (2011) (their Eq. 10), Lucarini et al. (2014), and Lembo et al. (2019) in estimating entropy production by irreversible processes in climate models. In this study, as discussed in Section 3, we consider vertical temperature and radiative heating rate profiles and their correlation in an hourly temporal resolution and a $1^\circ \times 1^\circ$ spatial resolution in computing entropy production.

3. Data products

We use Clouds and the Earth's Radiant Energy System (CERES) Edition 4.1 SYN1deg-Month data product (Rutan et al. 2015; Kato et al. 2018, 2019) in this study. The

212 SYN1deg-Month product includes shortwave and longwave irradiance profiles computed
213 hourly for every equal-area grid. The size of the equal-area grids is $1^{\circ} \times 1^{\circ}$ between 45°N
214 and 45°S and increases toward poles (Doelling et al. 2013). Cloud and aerosol properties
215 are derived from Moderate Resolution Imaging Spectrometer (MODIS, collection 5
216 March 2000 through February 2016 and collection 6.1 from March 2016 onward for
217 clouds and collection 6.1 throughout for aerosols) on Terra and Aqua except for the
218 period from March 2000 through June 2002 when only MODIS on Terra is used. In
219 addition, cloud properties are also derived from geostationary satellites hourly over
220 regions between 60°N and 60°S to account for the diurnal cycle of clouds.

221 Temperature and humidity profiles used in producing SYN1deg-Month are taken
222 from the Goddard Earth Observing System Data Assimilation System reanalysis product
223 (Rienecker et al. 2008) version 5.4.1 (GEOS-5.4.1). Temperature and relative humidity
224 profiles are 6 hourly with a 1° spatial resolution. The temporal resolution of temperature
225 and humidity profiles is increased to hourly below 10 m from the surface. For surface
226 properties, we primarily use MODIS derived spectral surface albedos except over ocean.
227 Modeled spectral surface albedos are used for ocean (Jin et al. 2004) and other surface
228 types when the MODIS spectral albedo is not available. Surface spectral emissivities are
229 based on Wilber et al. (1999). Further details of inputs and algorithm to produce
230 SYN1deg-Month is given in Rutan et al. (2015) and Kato et al (2018, 2019).

231

232 **3.1. Computation of entropy production**

233 We compute shortwave and longwave upward and downward irradiances hourly at 35
234 levels in the atmosphere (plus cloud top and base heights, depending on retrieved heights)

for every equal area grid with a radiative transfer model (Fu and Liou 1993; Rose et al. 2006, 2013). For shortwave irradiance computations, a gamma distribution is used to express the distribution of hourly cloud optical thickness in an equal area grid separated by 4 different cloud types. We use two parameters to define a gamma distribution, the mean optical thickness and shape factor that is the mean optical thickness divided by the standard deviation of optical thicknesses. We infer the shape factor from linear and logarithmic means of cloud optical thickness (Wilks 1995). When the shape factor is less than 10, we use the algorithm that analytically integrates two-stream solutions of an integro-differential equation of radiative transfer over a gamma distribution (Barker 1996; Kato 2005). When the shape factor is greater than or equal to 10, we use a four-stream model (Liou et al. 1988). The four-stream model is also used for clear-sky irradiance computations. For longwave computations, we use a combination of two- and four-stream models (Fu et al. 1997).

The net irradiance at the i th level is defined as the downward irradiance minus upward irradiance, where the top-of-atmosphere is defined as level = 0 and the level number increases downward. The top layer between level = 0 and level = 1 is layer 1 and the layer number also increases downward. The absorbed irradiance by the i th layer $F_{i,abs}$ is the net irradiance at $i-1$ th level minus the net irradiance at i th level,

$$F_{i,x}^{net} = F_{i-1,x}^{net} - F_{i,x}^{net} = (F_{i-1,x}^{\downarrow} - F_{i-1,x}^{\uparrow}) - (F_{i,x}^{\downarrow} - F_{i,x}^{\uparrow}), \quad (24)$$

where upward and downward arrow indicate the direction of the irradiance and subscript x is either shortwave SW or longwave LW. The top-of-atmosphere shortwave + longwave net irradiance is therefore $F_0^{net} = F_{TOA,SW}^{net} + F_{TOA,LW}^{net} = F_{TOA}^{net}$.

The entropy production by absorption of shortwave irradiance is then

258
$$\dot{\Sigma}_{SW} = \sum_{i=1}^N \frac{F_{i,SW}^{kl}}{T_i} + \frac{F_{sfc,SW}^{net}}{T_{sfc}} \quad (25)$$

259 where T_i is the temperature of the i th layer, $F_{sfc,SW}^{net}$ is the net shortwave irradiance at the

260 surface and T_{sfc} is the surface skin temperature. Although entropy is produced by

261 scattering (Wu and Liu 2010), we ignore the production in our computations. To compute

262 the entropy production by longwave irradiance emitted to space, Stephens and O'Brien

263 (1993) divide the upward longwave irradiance at TOA by the effective emission

264 temperature. In this study, we utilize the two-stream source function technique discussed

265 in Toon et al. (1989) and Appendix A. Toon et al. (1989) replace the radiance in

266 computing the source function by the radiance computed with a two-stream

267 approximation. With this approximation, the source function can be written explicitly.

268 When the source function is specified, the azimuthally averaged radiance at a two-stream

269 angle can also be written explicitly. We compute the downward irradiance starting from

270 TOA moving downward toward the surface with this expression. We subsequently

271 compute the upward irradiance starting at the surface moving upward toward TOA with

272 the specified surface emissivity. We compute the entropy flux in the same way that the

273 irradiance is computed except that the source function is divided by the temperature of

274 the layer such that

275
$$\dot{\Sigma}_{LW} = \sum_{i=1}^N \frac{t_i F_{i,LW}^{\uparrow}}{T_i} + \frac{t_{sfc} F_{sfc,LW}^{\uparrow}}{T_{sfc}}, \quad (26)$$

276 where t_i is the transmission of the irradiance from i th level to TOA, $F_{i,LW}^{\uparrow}$ is the upward

277 longwave irradiance at the i th level and T_i is the temperature of the i th layer, and N is the

278 number of layers (~ 35). Once net irradiances are computed, entropy produced by

279 nonradiative irreversible processes is computed by

$$\dot{\Sigma}_{tur} = - \sum_{i=1}^N \dot{\epsilon}_i \dot{\epsilon}_i$$

4. Results

In this section, we discuss the result of global annual mean entropy production followed by spatial distribution of the production.

4.1. Global annual mean entropy production

Table 1 summarizes computed global mean shortwave and longwave irradiances averaged from July 2005 through June 2015. Because computed TOA irradiances from SYN1deg-Month are not constrained by observed ocean heating rates averaged over a decade by the method discussed in Loeb et al. (2018), the net TOA irradiance is 1.3 Wm^{-2} , which is larger than the net TOA irradiance derived from CERES observations of 0.71 Wm^{-2} (Johnson et al. 2016; Loeb et al. 2018). **Table 2** summarizes the annual global mean entropy production by shortwave and longwave irradiances computed from a 10-year period (July 2005 through June 2015), as well as the standard deviation of annual global mean values. One year or a longer time period is needed to the system to be close to a steady state (Lucarini et al. 2011). The entropy produced by absorption of shortwave irradiance is $0.853 \text{ Wm}^{-2}\text{K}^{-1}$ and entropy produced by emission of longwave to space is $-0.928 \text{ Wm}^{-2}\text{K}^{-1}$. These give the entropy production by irreversible processes of $0.076 \text{ Wm}^{-2}\text{K}^{-1}$. Entropy produced by longwave cooling in the atmosphere and surface is, respectively, $-0.729 \text{ Wm}^{-2}\text{K}^{-2}$ and $-0.173 \text{ Wm}^{-2}\text{K}^{-1}$. This gives the entropy produced by non-radiative irreversible process of $0.049 \text{ Wm}^{-2}\text{K}^{-1}$ ($-[0.853-0.729-0.173]$). The

difference between $0.076 \text{ Wm}^{-2}\text{K}^{-1}$ and $0.049 \text{ Wm}^{-2}\text{K}^{-1}$ is $0.027 \text{ Wm}^{-2}\text{K}^{-1}$, which is the entropy produced by longwave radiation exchange within the Earth system. Uncertainty in these estimates is difficult to quantify but it is estimated based on uncertainties in irradiances estimated in earlier studies. Relative uncertainties in the global annual mean TOA shortwave and longwave irradiances are, respectively, 1% and 0.75% (Loeb et al. 2009). Based on uncertainties given in Kato et al. (2018), relative uncertainties in the global annual mean surface net shortwave and longwave irradiances are, respectively, 3% and 10%. If we assume that the uncertainty in global mean temperature is less than 1K and ignore the correlation of errors in irradiances and temperatures, the uncertainty in the global annual mean entropy production estimate is about 10% (1σ). This uncertainty estimate also ignores the uncertainty in radiative heating rate and cloud vertical profiles discussed in Ham et al. (2017) and Kato et al. (2019).

For comparisons, entropy production estimated by Peixoto et al. (1991) is also included in **Table 2**. Our values are different from those estimated by Peixoto et al. (1991). The difference of entropy production by the longwave irradiance emitted to space is largely caused by partitioning the entropy production by the surface and atmosphere. However, entropy produced by shortwave absorption are significantly different, especially at the surface.

Non-radiative irreversible processes that increase entropy considered by Goody (2000) include turbulent enthalpy transport of $0.0024 \text{ Wm}^{-2}\text{K}^{-1}$, frictional dissipation of $0.0113 \text{ Wm}^{-2}\text{K}^{-1}$, and irreversible phase change, transport and precipitation of water of $0.0188 \text{ Wm}^{-2}\text{K}^{-1}$. Pauluis and Dias (2012) suggest that the atmospheric heating by the frictional dissipation of falling raindrops is nearly equivalent to the frictional dissipation

of turbulence in the atmosphere. We therefore replace Goody's estimate of entropy production by raindrops of $0.0013 \text{ mWm}^{-2}\text{K}^{-2}$ by $0.010 \text{ Wm}^{-2}\text{K}^{-1}$, which gives the frictional dissipation of $0.020 \text{ W}^{-2}\text{K}^{-1}$. Entropy produced by ocean estimated by Bannon and Najjar (2019) is $0.0016 \text{ Wm}^{-2}\text{K}^{-2}$. When these productions are added, entropy production by non-radiative irreversible processes is $0.043 \text{ Wm}^{-2}\text{K}^{-1}$. Our estimate of entropy produced by non-radiative irreversible process of $0.049 \text{ Wm}^{-2}\text{K}^{-1}$, agrees reasonably well to within 15% with the sum of the components. Also, our estimate agrees with the estimate of $0.055 \text{ Wm}^{-2}\text{K}^{-1}$ by Lucarini et al. (2011) using climate models, although their estimate is based on a simple expression [i.e. Eq. (23)] that ignores the vertical irradiance and temperature profiles in the atmosphere. The result of Lembo et al. (2019) who estimated entropy produced by non-radiative irreversible processes using CMIP5 models with the simple expression show a substantial spread among models and dependence to the method used for the estimate (i.e. direct and indirect methods). Nevertheless, our estimate is close to the mean value from 7 CMIP5 models estimated by the direct method of $0.045 \text{ Wm}^{-2}\text{K}^{-2}$ (Lembo et al. 2019).

As discussed in Stephens and O'Brien (1993), the sum of entropy production by the surface and atmosphere by longwave irradiances emitted to space is approximately equal to the longwave irradiance emitted to space divided by the effective temperature,

$$\frac{F_{TOA,LW}^\uparrow}{T_e} = \sigma^{1/4} \left(F_{TOA,LW}^\uparrow \right)^{3/4} \quad (28)$$

where the effective temperature is defined as

$$T_e = \left(\frac{F_{TOA,LW}^\uparrow}{\sigma} \right)^{1/4} \quad (29)$$

and σ is the Stefan-Boltzmann constant. The annual global mean TOA upward longwave irradiance is 240 Wm^{-2} , which gives the entropy production of $0.941 \text{ Wm}^{-2}\text{K}^{-1}$ by Eq. (28), while the global annual mean entropy production from SYN1deg-Month is $0.928 \text{ Wm}^{-2}\text{K}^{-1}$. The difference is presumably caused by the translucent atmosphere combined with decreasing temperature with height in the atmosphere. Longwave radiation emitted by the lower atmosphere transmits through the atmosphere in the wavelength regions where the atmosphere is not entirely opaque. Because of this difference, the effective emission temperature computed from dividing longwave irradiance emitted to space by longwave entropy production using SYN1deg-Month values is 259 K and larger than 255 K computed by Eq. (29) (**Table 3**). The result of Stephens and O'Brien (1993) also shows a smaller effective emission temperature when Eq. (29) is used. Therefore, the reason that our estimate shown in Table 2 is slightly larger than the estimate by Stephens and O'Brien (1993) who used Eq. (29) is probably due to calibration differences between ERBE and CERES instruments.

The effective absorption temperature of the shortwave irradiance computed with dividing the TOA net shortwave irradiance of 241 Wm^{-2} by entropy production by shortwave of $0.853 \text{ Wm}^{-2}\text{K}^{-1}$ is 282 K . While different definitions of the thermodynamic efficiency are proposed (e.g. Pauluis and Held 2002; Lucarini et al. 2011; Novak and Teilleux 2018), the efficiency that can be computed with these effective temperatures is the Carnot efficiency η of the Earth system defined by Bannon and Lee (2017). Based on our results, the Carnot efficiency is

$$\eta = \frac{T_a - T_e}{T_a} = 0.085 \quad (30)$$

4.2. Regional entropy production

Figure 2 shows regional annual mean entropy production by absorption of solar irradiance and by emission of longwave irradiance to space. Entropy produced by shortwave absorption is larger in the tropics and decreases toward poles. The spatial distribution of entropy produced by shortwave absorption resembles the spatial pattern of shortwave absorption, indicating that the spatial distribution is driven by shortwave absorption. The production is larger in the tropics and decreases toward poles. Absorption of shortwave irradiance produces more entropy at the surface than in the atmosphere. Similarly, entropy production by longwave emission is large over tropics and decreases toward poles. The effect of clouds is evident over western Pacific, central Africa, and Amazon. Similar to entropy production by shortwave, the spatial distribution is driven by the spatial distribution of emitted longwave irradiance to space. A larger contribution, however, comes from the atmosphere than from the surface. Once entropy production by shortwave absorption is subtracted from entropy production by longwave emission to space, the resulting entropy production increases toward poles (**Figure 3**). This is equal to entropy produced by irreversible processes (Eq. 20). The regional difference of entropy produced by longwave emission to space and shortwave absorption is the net effect of entropy produced by frictional dissipation, entropy produced by diabatic heating by water phase change associated with water vapor transport (i.e. evaporation and condensation temperature differs), frictional dissipation of falling raindrops, horizontal and vertical enthalpy transport by irreversible processes, and energy transport by radiation exchange. As indicated by Eq. (13) and unlike regional energy balance in the column, vertical

energy transports by irreversible processes and radiation exchange affect entropy production in the column.

Figure 4 shows the standard deviation of annual regional mean entropy produced by shortwave absorption and longwave emission to space. The variability of tropical convections causes large variability of both shortwave and longwave entropy production (**Figure 4** top). While both atmosphere and surface contribute the variability of longwave entropy production, the atmospheric contribution to the shortwave entropy production is smaller than the surface contribution. Clouds and water vapor variabilities are responsible for the variability of entropy production by longwave. The variability of clouds also causes the variability of entropy production by shortwave at the surface but water vapor and aerosol variabilities primarily drive the variability of shortwave entropy production in the atmosphere. Once entropy production by shortwave absorption is subtracted from entropy production by longwave emission to space, the variability due to tropical convection is small (**Figure 5**), which is analogous to a small net TOA irradiance over convective cloud regions due to partial cancellation of their shortwave and longwave radiative effects. As a result, large variabilities of the difference of entropy produced by shortwave absorption and longwave emission occur mostly over stratocumulus regions. Similar to **Figure 3**, the variability shown in **Figure 5** is the results of variability of frictional dissipation, radiation exchange, and heating and cooling by water phase change in the column.

Figure 6 shows the mean and standard deviation of annual mean entropy produced by non-radiative irreversible process, which is equal to entropy produced by longwave cooling in the atmosphere and surface minus entropy produced by shortwave

absorption in the atmosphere and surface (Eq. 23). The spatial pattern is similar to **Figures 3 and 5.**

5. Discussion

At a steady state, the sum of entropy produced by absorption of solar irradiance and emission of longwave irradiance to space and entropy produced by irreversible processes in the Earth system is zero. In addition, the 1D model suggests that entropy produced by the net shortwave and longwave irradiances in the atmosphere and at the surface and entropy produced by nonradiative irreversible process is zero at a steady state. The nonradiative irreversible processes include frictional dissipation of turbulence, diabatic heating by water phase change associated with water vapor transport (i.e. evaporation and condensation temperatures differ), and frictional dissipation of falling raindrops. Although the 1D model provides insight of how entropy is produced, the 1D model can be misleading when it is used to understand the sensitivity of entropy production to energy input to Earth. Once observations are combined, however, the 1D model can provide an insight of how entropy production changes when energy input to the earth system is changed. This point is demonstrated in the remaining of this section.

To understand the sensitivity of entropy production to energy input to Earth, we start with showing the relationship between global annual TOA emitted longwave irradiance and net shortwave irradiance anomalies derived from Edition 4.1 Energy Balanced and Filled (EBAF) product (Loeb et al. 2018) in **Figure 7**. The slope of the line derived by regressing net longwave irradiance anomalies versus shortwave irradiance anomalies is 0.39. The TOA emitted longwave irradiance, therefore, increases (decreases)

439 with increasing (decreasing) shortwave absorption. At an annual global scale, however,
 440 the magnitude of emitted longwave irradiance anomalies is generally smaller than the
 441 magnitude of shortwave absorption anomalies. Standard deviations of global annual
 442 anomalies of the absorbed shortwave irradiance and emitted longwave irradiance are,
 443 respectively, 0.40 Wm^{-2} and 0.25 Wm^{-2} . Similarly, when we plot anomalies of the annual
 444 global mean shortwave and longwave entropy production as a function of anomalies of
 445 the annual global mean TOA absorptivity a (TOA net shortwave irradiance divided by

446 F_0) (**Figure 8**), the slope is $0.948 \text{ Wm}^{-2}\text{K}^{-1}$ for $\frac{\partial}{\partial a} \left(\frac{F_{TOA,SW}^{net}}{T_a} \right)$ and $0.197 \text{ Wm}^{-2}\text{K}^{-1}$ for

447 $\frac{\partial}{\partial a} \left(\frac{F_{TOA,LW}^{\dagger}}{T_e} \right)$. Positive slopes indicate that both shortwave and longwave entropy

448 productions increase with shortwave absorption. Positive slopes also imply that when the
 449 shortwave absorption increases, relative increase of T_a or T_e is smaller than the relative
 450 increase of $F_{TOA,SW}^{net}$ or $F_{TOA,LW}^{\dagger}$. The increase of shortwave entropy production is,
 451 however, larger than the increase of longwave entropy production.

452 At a global annual scale, most energy is used for heating ocean so that interannual
 453 variability of TOA net irradiance agrees with the variability of ocean heating rate
 454 (Johnson et al. 2016). When absorbed shortwave irradiance is perturbed, shortwave
 455 anomalies are damped by ocean so that the magnitude of emitted longwave anomalies is
 456 smaller than the magnitude of absorbed shortwave anomalies. This smoothing or
 457 damping processes seem to be responsible for a negative

458 $\frac{\partial}{\partial a} \left(\frac{F_{TOA,LW}^\dagger}{T_e} \right) - \frac{\partial}{\partial a} \left(\frac{F_{TOA,SW}^{net}}{T_a} \right) = \frac{\partial \dot{\Sigma}_{irr}}{\partial a}$. A negative slope ($-0.694 \text{ Wm}^{-2}\text{K}^{-1}$) derived by a

459 linear regression applied to the $\dot{\Sigma}_{irr}$ and a relationship (**Figure 8**) is consistent with the

460 difference of the slopes of $\frac{\partial}{\partial a} \left(\frac{F_{TOA,SW}^{net}}{T_e} \right)$ and $\frac{\partial}{\partial a} \left(\frac{F_{TOA,LW}^\dagger}{T_e} \right)$. Although the slopes derived

461 from linear regression is not a partial derivative because other variables are not held

462 constant, we use the slope as proximity of the partial derivative and describe the

463 difference of entropy production sensitivity derived from observations and the 1D model

464 in the remaining of this section.

465 We take the derivative of entropy production with respect to shortwave

466 absorptivity,

467
$$\frac{\partial \dot{\Sigma}_{irr}}{\partial a} = \frac{F_{TOA,LW}^\dagger}{T_e} \left(\frac{1}{F_{TOA,LW}^\dagger} \frac{\partial F_{TOA,LW}^\dagger}{\partial a} - \frac{1}{T_e} \frac{\partial T_e}{\partial a} \right) - \frac{F_{TOA,SW}^{net}}{T_a} \left(\frac{1}{a} - \frac{1}{T_a} \frac{\partial T_a}{\partial a} \right) \quad (31)$$

468 and estimate the numerical value of the derivative. **Table 3** compares derivative values

469 derived from CERES data products using regression (second column) and theoretical

470 estimates using the 1D model (third column). The 1D model does not predict the change

471 of T_a with shortwave absorption so that observations are needed to estimate $\frac{\partial T_a}{\partial a}$. If we

472 set $\frac{\partial T_a}{\partial a} = 0$, the derivative $\frac{\partial \dot{\Sigma}_{irr}}{\partial a}$ from the 1D model is $-0.201 \text{ Wm}^{-2}\text{K}^{-1}$. The estimate of

473 $\frac{\partial \dot{\Sigma}_{irr}}{\partial a}$ by Bannon and Lee (2017) is $-0.236 \text{ Wm}^{-2}\text{K}^{-1}$. The reason for the difference

474 between our value of $-0.201 \text{ Wm}^{-2}\text{K}^{-1}$ and their value is that we use $T_a = 283 \text{ K}$ instead of

475 275 K . When we use the value of $\frac{\partial T_a}{\partial a}$ estimated from the CERES data products, the

476 derivative is $0.172 \text{ Wm}^{-2}\text{K}^{-1}$. The difference of the sensitivity derived from the CERES

477 data products and 1D model arises because of the difference of $\frac{\partial F_{TOA,LW}^\dagger}{\partial a}$ or $\frac{\partial F_{TOA,LW}^\dagger}{\partial F_{TOA,SW}^{net}}$.

478 The 1D model assumes that $\frac{\partial F_{TOA,LW}^\dagger}{\partial F_{TOA,SW}^{net}} = 1$ while the CERES observations indicate that the

479 slope of the relationship of annual global mean emitted longwave anomalies as a function

480 of annual global mean net shortwave anomalies is less than 1.

481 We test whether Eq. (31) can also be used to estimate the critical absorptivity to

482 make $\frac{\partial \dot{\Sigma}_{irr}}{\partial a} = 0$ with an assumption that all values are unaltered when the shortwave

483 absorption is altered. If we use $\frac{\partial}{\partial a} \left(\frac{F_{TOA,LW}^\dagger}{T_e} \right) = 0.197 \text{ Wm}^{-2}\text{K}^{-1}$ derived by directly fitting

484 $\frac{F_{TOA,LW}^\dagger}{T_e}$ versus a , and use $F_{TOA,SW}^{net}$, T_a , and $\frac{\partial T_a}{\partial a}$ derived from the CERES data products

485 (the second column of **Table 3**), $\frac{\partial \dot{\Sigma}_{irr}}{\partial a}$ is always negative for the physically possible

486 range of the absorptivity, $0 \leq a \leq 1$. Therefore, entropy production by irreversible

processes always decreases with increasing absorption of shortwave irradiance for the possible range of the planetary albedo.

6. Summary and conclusions

We used satellite observations of cloud and aerosol properties, temperature and humidity profiles from reanalysis to compute entropy production within the Earth system. We focused entropy production by heating and cooling by radiation that, at a steady state, balances with entropy production by irreversible processes within the system.

Entropy produced by longwave emission to space estimated in this study, $0.929 \text{ Wm}^{-2}\text{K}^{-1}$, is close to that of $0.925 \text{ Wm}^{-2}\text{K}^{-1}$ estimated by Peixoto et al. (1991). In addition, entropy carried by longwave irradiance emitted to space estimated in this study is in a close agreement with that estimated by Stephens and O'Brien (1993) using ERBE data. However, entropy produced by absorption of shortwave irradiance at the surface and by the atmosphere are significantly different from those estimated by Peixoto et al. (1991). With a steady state assumption, entropy production by irreversible processes within the Earth system is estimated to be $0.076 \text{ Wm}^{-2}\text{K}^{-1}$ and by non-radiative irreversible processes to be $0.049 \text{ Wm}^{-2}\text{K}^{-1}$.

Spatial distribution of entropy production is similar to the spatial distribution of net shortwave irradiance and longwave irradiance emitted to space. Larger standard deviation of annual mean of regional entropy productions by shortwave irradiance absorption and longwave irradiance emission to space is caused by the variability due to convections in the tropics. Once the entropy production by absorption of shortwave irradiance is subtracted from the entropy production by longwave irradiance emission to

space, the variability of the difference is larger over stratocumulus regions, resemble to the spatial distribution of the variability of net TOA irradiances.

Entropic absorption and emission temperatures derived from annual global mean TOA shortwave and longwave irradiance and entropy productions are, respectively, 282 K and 259 K. These entropic temperature gives the Carnot efficiency of the Earth system of 8.5%.

CERES observations suggest that, at a global annual scale, entropy produced by emission of longwave irradiance to space and by absorption of shortwave irradiance increases with increasing energy input to (absorptivity of) the Earth system. Increasing the entropy produced by the shortwave absorption is larger than increasing the entropy produced by the longwave emission, which makes entropy production by irreversible processes decreases when absorption of shortwave irradiance increases.

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Appendix A: Computation of material entropy production by longwave irradiance emitted to space

Following Toon et al. (1986) we use two approximations in addition to a two-stream approximation. First, the Planck function is represented by

$$B_n(\tau) = B_{0n} + B_{1n} \tau \quad (A1)$$

where B_{0n} is the Planck function evaluated at the temperature of the top of the n th computation layer with the optical thickness of τ_n and

$$B_{1n} = \frac{[B(T_{bn}) - B_{0n}]}{\tau_n} \quad (A2)$$

where $B(T_{bn})$ is the Planck function evaluated at the temperature of the bottom of the n th layer T_{bn} . Second, we use the radiance computed with a two-stream approximation instead of the true radiance in computing source function. With these approximations, the source function S_{ν} of longwave irradiances propagating upward and downward though the atmosphere are (Toon et al. 1989 Eqs. (53) and (54)), respectively,

$$S_{\nu}^+ = G \exp(-\lambda \tau_n) \exp(\lambda \tau) + H \exp(-\lambda \tau) + \alpha_1 + \alpha_2 \tau, \quad (A3)$$

and

$$S_{\nu}^- = J \exp(-\lambda \tau_n) \exp(\lambda \tau) + K \exp(-\lambda \tau) + \sigma_1 + \sigma_2 \tau, \quad (A4)$$

where G , H , J , K , α , and σ are defined in Table 3 of Toon et al. (1989) for the hemispheric mean two-stream approximation and superscripts $+$ and $-$ indicate upward (propagating from the surface to TOA) and downward (propagating from TOA to the surface). We also assume that the temperature changes linearly within the n th computational layer with the optical thickness of τ_n so that

$$T_n(\tau) = T_{0n} + T_{1n} \tau \quad (A5)$$

552 where

553
$$T_{1n} = \frac{T_{bn} - T_{0n}}{\tau} \quad (A6)$$

554 and T_{bn} and T_{0n} are the temperature at the bottom and top of the n th layer. The entropy

555 produced by the upward longwave irradiance by the n th layer is

556
$$\begin{aligned} \int_0^{\tau_n} \frac{S_{\nu t}^{\pm}}{T_{0n} + T_{1n}t} \exp(-t/\mu) \frac{dt}{\mu} &= \int_0^{\tau_n} \frac{S_{\nu t}^{\pm}}{T_{0n} \left(1 + \frac{T_{1n}}{T_{0n}}t\right)} \exp(-t/\mu) \frac{dt}{\mu} \\ &= \int_0^{\tau_n} \frac{S_{\nu t}^{\pm}}{T_{0n}} \exp(-t/\mu) \left(1 - \frac{T_{1n}}{T_{0n}}t + \frac{T_{1n}^2}{T_{0n}^2}t^2 - \dots\right) \frac{dt}{\mu} \end{aligned} \quad (A7)$$

557 Substituting Eqs. (A3) and (A4) in (A7) leads to

558
$$\begin{aligned} &\int_0^{\tau_n} \frac{S_{\nu t}^+}{T_{0n}} e^{-\frac{t}{\mu}} - \frac{S_{\nu t}^+}{T_{0n}} \frac{T_{1n}}{T_{0n}} t e^{-\frac{t}{\mu}} + \frac{S_{\nu t}^+}{T_{0n}} \left(\frac{T_{1n}}{T_{0n}}\right)^2 t^2 e^{-\frac{t}{\mu}} - \dots \frac{dt}{\mu} \\ &= A^+ \left(\frac{1}{T_{0n}} + \frac{T_{1n}}{T_{0n}^2} \frac{\mu}{\lambda\mu - 1} + \frac{T_{1n}^2}{T_{0n}^3} \frac{2\mu^2}{(\lambda\mu - 1)^2} + \sum \frac{T_{1n}^n}{T_{0n}^{n+1}} \frac{n!\mu^n}{(\lambda\mu - 1)^n} \right) \\ &+ B^+ \left(\frac{1}{T_{0n}} - \frac{T_{1n}}{T_{0n}^2} \frac{\mu}{\lambda\mu + 1} + \frac{T_{1n}^2}{T_{0n}^3} \frac{2\mu^2}{(\lambda\mu + 1)^2} + \sum (-1)^n \frac{T_{1n}^n}{T_{0n}^{n+1}} \frac{n!\mu^n}{(\lambda\mu + 1)^n} \right) \\ &+ C^+ \left(\frac{1}{T_{0n}} - \frac{T_{1n}}{T_{0n}^2} \mu + \frac{T_{1n}^2}{T_{0n}^3} 2\mu^2 + \sum (-1)^n \frac{T_{1n}^n}{T_{0n}^{n+1}} n!\mu^n \right) + D^+ \left(\frac{1}{T_{0n}} - 2 \frac{T_{1n}}{T_{0n}^2} \mu + \frac{T_{1n}^2}{T_{0n}^3} 6\mu^2 + \sum (-1)^n \frac{T_{1n}^n}{T_{0n}^{n+1}} (n+1)!\mu^n \right) \\ &+ \frac{G}{\lambda\mu - 1} \tau_n e^{-\frac{\tau_n}{\mu}} \left[-\frac{T_{1n}}{T_{0n}^2} + \frac{T_{1n}^2}{T_{0n}^3} \left(\tau_n + \frac{2\mu}{\lambda\mu - 1} \right) + \dots \right] - \frac{H}{\lambda\mu + 1} \tau_n e^{-\left(\lambda + \frac{1}{\mu}\right)\tau_n} \left[-\frac{T_{1n}}{T_{0n}^2} + \frac{T_{1n}^2}{T_{0n}^3} \left(\tau_n + \frac{2\mu}{\lambda\mu + 1} \right) + \dots \right] \\ &- \alpha_1 \tau_n e^{-\frac{\tau_n}{\mu}} \left[-\frac{T_{1n}}{T_{0n}^2} + \frac{T_{1n}^2}{T_{0n}^3} (\tau_n + 2\mu) + \dots \right] - \alpha_2 \tau_n^2 e^{-\frac{\tau_n}{\mu}} \left[-\frac{T_{1n}}{T_{0n}^2} + \frac{T_{1n}^2}{T_{0n}^3} (\tau_n + 3\mu) + \dots \right] \end{aligned} \quad (A8)$$

559 and

560

$$\begin{aligned}
& \int_0^{\tau_n} \frac{S_{\nu\tau}^-}{T_{0n}} e^{-\frac{t}{\mu}} - \frac{S_{\nu\tau}^+}{T_{0n}} \frac{T_{1n}}{T_{0n}} t e^{-\frac{t}{\mu}} + \frac{S_{\nu\tau}^-}{T_{0n}} \left(\frac{T_{1n}}{T_{0n}} \right)^2 t^2 e^{-\frac{t}{\mu}} - \dots \frac{dt}{\mu} \\
& = A \left(\frac{1}{T_{0n}} + \frac{T_{1n}}{T_{0n}^2} \frac{\mu}{\lambda\mu + 1} + \frac{T_{1n}^2}{T_{0n}^3} \frac{2\mu^2}{(\lambda\mu + 1)^2} + \sum \frac{T_{1n}^n}{T_{0n}^{n+1}} \frac{n!\mu^n}{(\lambda\mu + 1)^n} \right) \\
& + B \left(\frac{1}{T_{0n}} - \frac{T_{1n}}{T_{0n}^2} \frac{\mu}{\lambda\mu - 1} + \frac{T_{1n}^2}{T_{0n}^3} \frac{2\mu^2}{(\lambda\mu - 1)^2} + \sum (-1)^n \frac{T_{1n}^n}{T_{0n}^{n+1}} \frac{n!\mu^n}{(\lambda\mu - 1)^n} \right) \\
& + C \left(\frac{1}{T_{0n}} + \frac{T_{1n}}{T_{0n}^2} \mu + \frac{T_{1n}^2}{T_{0n}^3} 2\mu^2 + \sum (-1)^n \frac{T_{1n}^n}{T_{0n}^{n+1}} n!\mu^n \right) + D \left(\frac{1}{T_{0n}} + 2 \frac{T_{1n}}{T_{0n}^2} \mu + \frac{T_{1n}^2}{T_{0n}^3} 6\mu^2 + \sum (-1)^n \frac{T_{1n}^n}{T_{0n}^{n+1}} (n+1)!\mu^n \right) \\
& + \frac{J}{\lambda\mu + 1} \tau_n \left[-\frac{T_{1n}}{T_{0n}^2} + \frac{T_{1n}^2}{T_{0n}^3} \left(\tau_n + \frac{2\mu}{\lambda\mu + 1} \right) + \dots \right] - \frac{K}{\lambda\mu - 1} \tau_n e^{-\lambda\tau_n} \left[-\frac{T_{1n}}{T_{0n}^2} + \frac{T_{1n}^2}{T_{0n}^3} \left(\tau_n + \frac{2\mu}{\lambda\mu - 1} \right) + \dots \right] \\
& + \alpha_1 \tau_n \left[-\frac{T_{1n}}{T_{0n}^2} + \frac{T_{1n}^2}{T_{0n}^3} (\tau_n + 2\mu) + \dots \right] + \alpha_2 \tau_n \left[-\frac{T_{1n}}{T_{0n}^2} + \frac{T_{1n}^2}{T_{0n}^3} (\tau_n + 3\mu) + \dots \right]
\end{aligned}
\tag{A9}$$

where

$$A^+ = \frac{G}{\lambda\mu - 1} \left(e^{-\frac{\tau_n}{\mu}} - e^{-\tau_n \lambda} \right), \tag{A10}$$

$$B^+ = \frac{H}{\lambda\mu + 1} \left[1 - e^{-\left(\lambda + \frac{1}{\mu} \right) \tau_n} \right], \tag{A11}$$

$$C^+ = 1 - e^{-\frac{\tau_n}{\mu}}, \tag{A12}$$

$$D^+ = \mu - (\tau_n + \mu) e^{-\frac{\tau_n}{\mu}}. \tag{A13}$$

Expressions for an exponential assumption of the Planck function within a layer (Fu et al. 1997, eq 2.25) are,

$$\begin{aligned}
& \int_0^{\tau_n} \frac{S_{\nu\ell}^+}{T_{0n}} e^{-\frac{t}{\mu}} - \frac{S_{\nu\ell}^+}{T_{0n}} \frac{T_{1n}}{T_{0n}} t e^{-\frac{t}{\mu}} + \frac{S_{\nu\ell}^+}{T_{0n}} \left(\frac{T_{1n}}{T_{0n}} \right)^2 t^2 e^{-\frac{t}{\mu}} - \dots \frac{dt}{\mu} \\
& = A^+ \left(\frac{1}{T_{0n}} + \frac{T_{1n}}{T_{0n}^2} \frac{\mu}{\lambda\mu - 1} + \frac{T_{1n}^2}{T_{0n}^3} \frac{2\mu^2}{(\lambda\mu - 1)^2} + \sum \frac{T_{1n}^n}{T_{0n}^{n+1}} \frac{n!\mu^n}{(\lambda\mu - 1)^n} \right) \\
& + B^+ \left(\frac{1}{T_{0n}} - \frac{T_{1n}}{T_{0n}^2} \frac{\mu}{\lambda\mu + 1} + \frac{T_{1n}^2}{T_{0n}^3} \frac{2\mu^2}{(\lambda\mu + 1)^2} + \sum (-1)^n \frac{T_{1n}^n}{T_{0n}^{n+1}} \frac{n!\mu^n}{(\lambda\mu + 1)^n} \right) \\
& + C^+ \left(\frac{1}{T_{0n}} + \frac{T_{1n}}{T_{0n}^2} \frac{\mu}{\beta\mu - 1} - \frac{T_{1n}^2}{T_{0n}^3} \frac{2\mu^2}{(\beta\mu - 1)^2} - \sum (-1)^n \frac{T_{1n}^n}{T_{0n}^{n+1}} \frac{n!\mu^n}{(\beta\mu - 1)^n} \right) \\
& + \frac{G}{\lambda\mu - 1} \tau_n e^{-\frac{\tau_n}{\mu}} \left[-\frac{T_{1n}}{T_{0n}^2} + \frac{T_{1n}^2}{T_{0n}^3} \left(\tau_n + \frac{2\mu}{\lambda\mu - 1} \right) + \dots \right] - \frac{H}{\lambda\mu + 1} \tau_n e^{-\frac{(\lambda + \frac{1}{\mu})\tau_n}{\mu}} \left[-\frac{T_{1n}}{T_{0n}^2} + \frac{T_{1n}^2}{T_{0n}^3} \left(\tau_n + \frac{2\mu}{\lambda\mu + 1} \right) + \dots \right] \\
& + \frac{\varsigma B_0 e^{\beta\tau_n}}{\beta\mu - 1} \tau_n e^{-\frac{\tau_n}{\mu}} \left[-\frac{T_{1n}}{T_{0n}^2} + \frac{T_{1n}^2}{T_{0n}^3} \left(\tau_n + \frac{2\mu}{\beta\mu - 1} \right) + \dots \right]
\end{aligned}
\tag{A14}$$

and

$$\begin{aligned}
& \int_0^{\tau_n} \frac{S_{\nu\ell}^-}{T_{0n}} e^{-\frac{t}{\mu}} - \frac{S_{\nu\ell}^-}{T_{0n}} \frac{T_{1n}}{T_{0n}} t e^{-\frac{t}{\mu}} + \frac{S_{\nu\ell}^-}{T_{0n}} \left(\frac{T_{1n}}{T_{0n}} \right)^2 t^2 e^{-\frac{t}{\mu}} - \dots \frac{dt}{\mu} \\
& = A^- \left(\frac{1}{T_{0n}} + \frac{T_{1n}}{T_{0n}^2} \frac{\mu}{\lambda\mu + 1} + \frac{T_{1n}^2}{T_{0n}^3} \frac{2\mu^2}{(\lambda\mu + 1)^2} + \sum \frac{T_{1n}^n}{T_{0n}^{n+1}} \frac{n!\mu^n}{(\lambda\mu + 1)^n} \right) \\
& + B^- \left(\frac{1}{T_{0n}} - \frac{T_{1n}}{T_{0n}^2} \frac{\mu}{\lambda\mu - 1} + \frac{T_{1n}^2}{T_{0n}^3} \frac{2\mu^2}{(\lambda\mu - 1)^2} + \sum (-1)^n \frac{T_{1n}^n}{T_{0n}^{n+1}} \frac{n!\mu^n}{(\lambda\mu - 1)^n} \right) \\
& + C^- \left(\frac{1}{T_{0n}} + \frac{T_{1n}}{T_{0n}^2} \frac{\mu}{\beta\mu + 1} - \frac{T_{1n}^2}{T_{0n}^3} \frac{2\mu^2}{(\beta\mu + 1)^2} - \sum (-1)^n \frac{T_{1n}^n}{T_{0n}^{n+1}} \frac{n!\mu^n}{(\beta\mu + 1)^n} \right) \\
& + \frac{J}{\lambda\mu + 1} \tau_n \left[-\frac{T_{1n}}{T_{0n}^2} + \frac{T_{1n}^2}{T_{0n}^3} \left(\tau_n + \frac{2\mu}{\lambda\mu + 1} \right) + \dots \right] - \frac{K}{\lambda\mu - 1} \tau_n e^{-\lambda\tau_n} \left[-\frac{T_{1n}}{T_{0n}^2} + \frac{T_{1n}^2}{T_{0n}^3} \left(\tau_n + \frac{2\mu}{\lambda\mu - 1} \right) + \dots \right] \\
& + \frac{\eta B_0 e^{\beta\tau_n}}{\beta\mu + 1} \tau_n \left[-\frac{T_{1n}}{T_{0n}^2} + \frac{T_{1n}^2}{T_{0n}^3} \left(\tau_n + \frac{2\mu}{\beta\mu + 1} \right) + \dots \right]
\end{aligned}
\tag{A15}$$

We only keep terms with up to second power of T_{0n} in the denominator because high order terms are smaller in computing entropy production by Eqs. (A14) and (A15). These expressions compute entropy produced by longwave emission by the layer. These expressions therefore accurately compute entropy produced by longwave absorption and

579 emission by the layer. These expressions, however, do not explicitly compute the
580 transmission from each layer to space hence do not explicitly keep track of the emission
581 temperature of radiation emitted to space.
582

References

- Bannon, P. R., 2015: Entropy production and climate efficiency, *J. Atmos. Sci.* 72, DOI: 10.1175/JAS-D-14-0361.1.
- Bannon, P. R., and S. Lee, 2017: Toward quantifying the climate heat engine: Solar absorption and terrestrial emission temperatures and material entropy production, *J. Atmos. Sci.* DOI: 10.1175/JAS-D-16-0240.1.
- Barker, H. W., 1996: A parameterization for computing grid-averaged solar fluxes for inhomogeneous marine boundary layer clouds. Part I: Methodology and homogeneous biases, *J. Atmos. Sci.*, 53, 2289–2303, doi:10.1175/1520-0469(1996)053<2289:APFCGA>2.0.CO;2.
- Doelling, D. R., Loeb, N. G., Keyes, D. F., Nordeen, M. L., Morstad, D., Wielicki, B. A., Young, D. F., Sun, M., 2013: Geostationary enhanced temporal interpolation for CERES flux products. *J. Atmos. Oceanic Technol.*, DOI: 10.1175/JTECH-D-12-00136.1.
- Feistel, R., 2011: Entropy flux and entropy production of stationary black-body radiation, *J. Non-Equilib. Thermodyn.*, 36, 131-139, DOI: 10.1515/JNETDY.2011.009.
- Fu, Q., Liou, K. N., M. C. Cribb, T. P. Charlock, and A. Grossman, 1997: Multiple scattering parameterization in thermal infrared radiative transfer, *J. Atmos. Sci.*, 54, 2799-2812.
- Fu, Q., and K. N. Liou, 1993: Parameterization of the radiative properties of cirrus clouds. *J. Atmos. Sci.*, 50, 2008–2025, [https://doi.org/10.1175/1520-0469\(1993\)050<2008:POTRPO.2.0.CO;2](https://doi.org/10.1175/1520-0469(1993)050<2008:POTRPO.2.0.CO;2).
- Goody, R. 2000: Source and sinks of climate entropy, *Q. J. R. Meteorol. Soc.* 126, 1953-1970.
- Ham, S.-H., S. Kato, H. W. Barker, F. G. Rose, and S. Sun-Mack (2014), Effects of 3-D clouds on atmospheric transmission of solar radiation: Cloud type dependencies inferred from A-train satellite data, *J. Geophys. Res. Atmos.*, 119, 943–963, doi:10.1002/2013JD020683.
- Jin, Z., T. P. Charlock, W. L. Smith Jr., and K. Rutledge, 2004: A parameterization ocean surface albedo. *Geophys. Res. Lett.*, 31, L22301, doi:10.1029/2004GL021180.
- Johnson, G. C., J. M. Lyman, N. G. Loeb, 2016: Improving estimate of Earth's energy imbalance, *Nature Geoscience*, 6, 639-640.
- Kato, S., F. G. Rose, S.-H. Ham, D. A. Rutan, A. Radkevich, T. E. Caldwell, S. Sun-Mack, W. F. Miller, and Y. Chen, 2018: Radiative heating rates computed with clouds derived from satellite-based passive and active sensors and their effects on generation of available potential energy, submitted to *J. Geophysic. Res.*
- Kato, S., F. G. Rose, and T. P. Charlock, 2005: Computation of domain averaged irradiance using satellite derived cloud properties, *J. Atmos. Oceanic Technol.*, 22, 146–164, doi:10.1175/JTECH-1694.1.
- Kjellsson, J, 2015: Weakening of the global atmospheric circulation with global warming, *Clim Dyn*, 45:975-988, DOI: 10.1007/s00382-014-2337-8.
- Laliberté, F., J. Zika, L. Mudryk, P. J. Kushner, J. Kjellsson, K. Döös, 2015: Constrained work output of the moist atmospheric heat engine in a warming climate, *Science* 347 (6221), 540-543, DOI: 10.1126/science.1257103.

627 Lembo, V., F. Lunkeit, and V. Lucarini, 2019: TheDiaTo (v1.0) – a new diagnostic tool
 628 for water, energy, and entropy budget in climate models, *Geosci. Model Dev.*, 12,
 629 3805-3834, DOI: 10.5194/gmd-12-3805-2019.
 630 Liou, K.-N., Q. Fu, and T. P. Ackerman, 1988: A simple formulation of the delta-four-
 631 stream approximation for radiative transfer parameterizations, *J. Atmos. Sci.* 45,
 632 1940-1947.
 633 Loeb, N. G., D. R. Doelling, H. Wang, W. Su, C. Nguyen, J. G. Corbett, L. Liang, C.
 634 Mitrescu, F. Rose, and S. Kato, 2018: Clouds and the Earth Radiant Energy
 635 System (CERES) Energy Balanced and Filled (EBAF) Top-of-Atmosphere
 636 (TOA) Edition-4.0 data product, *J. Climate*, 31, doi: 10.1175/JCLI-D-17-0208.1.
 637 Lucarini, V., Fraedrich, K., Ragone, F., 2011: New results on the thermodynamic
 638 properties of the climate system, *J. Atmos. Sci.*, 68, DOI:
 639 10.1175/2011JAS3713.1.
 640 Pauluis, O., and J. Dias, 2012: Satellite estimates of precipitation-induced dissipation in
 641 the atmosphere, *Science*, 335, 10.1126/science.1215869.
 642 Peixoto, J. P., and A. H. Oort, 1992: *Physics of Climate*. American Institute of Physics,
 643 520 pp.
 644 Peixoto, J. P., A. H. Oort, M. De Almeida, and A. Tome, 1991: Entropy budget of the
 645 atmosphere, *J. Geophysic. Res.* 96, D6, 10981-10988.
 646 Planck, M., 1913: The theory of heat radiation, reprinted by the American Institute of
 647 Physics, the History of Modern Physics vol. 11.
 648 Rienecker, M. M., and Coauthors, 2008: The GOES-5 Data Assimilation System:
 649 Documentation of versions 5.0.1, 5.1.0, and 5.2.0. NASA Technical Report Series
 650 on Global Modeling and Data Assimilation, Vol. 27, NASA/TM-2008-105606, 97
 651 pp.
 652 Rose, F., D. A. Rutan, T. P. Charlock, G. L. Smith, and S. Kato, 2013: An algorithm for
 653 the constraining of radiative transfer calculations to CERES observed broadband
 654 top of atmosphere irradiance. *J. Atmos. Oceanic. Technol.*,
 655 DOI:10.11175/JTECH-D-12-00058.1.
 656 Rose, F., T. Charlock, Q. Fu, S. Kato, D. Rutan, and Z. Jin, 2006: CERES Proto-Edition
 657 3 radiative transfer: Tests and radiative closure over surface validation sites, paper
 658 presented at 12th Conference on Atmospheric Radiation, Am. Meteorol. Soc.,
 659 Madison, Wis., 10–14 July. [Available at <http://snowdog.larc.nasa.gov/cave/>.]
 660 Rutan, D. A., S. Kato, D. R. Doelling, F. G. Rose, L. T. Nguyen, T. E. Caldwell, and N.
 661 G. Loeb, 2015: CERES synoptic product: Methodology and validation of surface
 662 radiant flux, *J. Atmos. Ocean. Tech.*, 32, DOI: 10.1175/JTECH-D-14-00165.1.
 663 Stephens, G. L., and coauthors, 2012: An update on Earth's energy balance in light of the
 664 latest global observations. *Nat. Geosci.*, 5, 691–696,
 665 <https://doi.org/10.1038/ngeo1580>.
 666 Stephens, G. L., D. M. O'Brien, 1993: Entropy and climate. I: ERBE observations of the
 667 entropy production of the earth, *Q. J. R. Meteorol. Soc.*, 119, 121-152.
 668 Toon, O. B., C. P. McKay, and T. P. Ackerman, 1989: Rapid calculation of radiative
 669 heating rates and photodissociation rates in inhomogeneous multiple scattering
 670 atmospheres, *J. Geophys. Res.*, 94, D13, 16287-16301.

671 Wilber, A. C., Kratz, D. P., Gupta, S. K., 1999: Surface emissivity maps for use in
672 satellite retrievals of longwave radiation. NASA Tech. Memo. TP-1999-209362,
673 30 pp.
674 Wu, Wei, and Y. Liu, 2010: Radiation entropy flux and entropy production of the Earth
675 system, *Rev. Geophys.*, 48, 2008RG000275.
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Tables

Table 1: Global annual mean and standard deviation of TOA and surface net irradiances for July 2005 to June 2015 in Wm^{-2} .

	Irradiance (Wm^{-2})
Top-of-atmosphere	
NET Shortwave	241 (0.21)
NET Longwave	-240 (0.18)
Surface	
NET Shortwave	164 (0.28)
NET Longwave	-53.0 (0.33)

Table 2: Global annual mean radiative entropy production in $\text{mWm}^{-2}\text{K}^{-1}$ and standard deviation of annual means computed from a 10-year period

Entropy production process	This study (SYN1deg-Month, Ed4.1 [#])	Peixoto et al. (1991)	Stephens and O'Brien (1993)	Goody (2000), Pauluis and Dias (2012), and Bannon and Najjar (2009)
TOA incoming Shortwave	54.6*	55.1*		
TOA outgoing Longwave	1238* (0.8)		1230*	
Atmosphere outgoing Longwave at TOA	-826 (1.3)	-854		
Surface outgoing Longwave at TOA	-103 (1.0)	-71		
Atmosphere absorbed Shortwave	304 (0.7)	258		
Surface absorbed Shortwave	549 (1.3)	-29.5		
Atmosphere net Longwave	729 (1.6)			
Surface net Longwave	173 (1.2)			
Irreversible processes	76 ^{\$} (1.1)	697 ^{\$}		
Irreversible nonradiative processes	49 (1.2)			43 [@]

* Entropy flux carried by radiation (i.e. including a 4/3 factor).
global mean of July 2006 through June 2015.
\$ -(Atm absorbed SW + Sfc absorbed SW - Atm outgoing LW - Sfc outgoing LW).
@ Sum of Itemized values explained in Section 5.

Table 3: Global annual mean values that appear in Eq. (2) derived from SYN1deg-Month and the ideal energy balanced model

	Global annual mean	
	Ed4 SYN EBAF-TOA	Ideal energy balanced model
Downward shortwave irradiance F_0 (Wm ⁻²)	340	340
Net shortwave irradiance F_{sw} (Wm-2)	241	241
Emitted longwave irradiance F_{lw} (Wm-2)	240	241
Shortwave absorptivity a	0.708	0.709
Effective absorption temperature T_a (K)	282 ($=F_{sw}/S_a$)	283 [@]
Effective emission temperature T_e (K)	259 ($=F_{lw}/S_e$)	255 ($=F_{sw}/\sigma$) ^(1/4)
$\partial F_{LW}/\partial a$ (Wm-2)	125 [*]	340 ($=F_0$)
$\partial T_a/\partial a$ (K)	85 [*]	0
$\partial T_e/\partial a$ (K)	80 [*]	90 [$=T_e/(4a)$]
$\frac{F_{TOA,LW}^\dagger}{T_e} \left(\frac{1}{F_{TOA,LW}^\dagger} \frac{\partial F_{TOA,LW}^\dagger}{\partial a} - \frac{1}{T_e} \frac{\partial T_e}{\partial a} \right)$ (Wm-2K-1)	0.197 ⁺	1.0
$\frac{F_{TOA,SW}^{net}}{T_a} \left(\frac{1}{a} - \frac{1}{T_a} \frac{\partial T_a}{\partial a} \right)$ (Wm-2K-1)	0.948 ^{\$}	1.201 [#] (0.828 ^{\$})
$\partial (F_{TOA,LW}^\dagger/T_e)/\partial a$ (Wm-2K-1)	0.197 [*]	
$\partial (F_{TOA,SW}^{net}/T_s)/\partial a$ (Wm-2K-1)	0.888 [*]	

$\partial \bar{\Sigma}_{irr} / \partial a$ (Wm ⁻² K ⁻¹)	-0.694 [*]	-0.201 [#] (0.172 ^{&})
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712 * derived from linear regression

713 @ derived from Ed4 SYN

714 # computed with setting $\partial T_a / \partial a = 0$ K, i.e. $= F_0 / T_s$.

715 \$ computed with setting $T_a / \partial a = 59$ K.

716 + computed with $\partial F_{LW} / \partial a = 132$ Wm⁻² and $\partial T_e / \partial a = 71$ K

717 & 1.0 – 0.828 Wm⁻²K⁻¹.

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Figure caption list

Figure 1: One-dimensional isothermal atmosphere model. F_0 is solar irradiance at top-of-atmosphere, β is the absorptivity of the atmosphere, α is the reflectivity of the surface, T_{sfc} is the surface temperature, T_{atm} is the temperature of the atmosphere, ε is the emissivity of the atmosphere, and F_{tur} is the energy flux by turbulence.

Figure 2: Annual mean entropy produced by absorption of shortwave irradiance (left) and by emission of longwave irradiance to space (right) in $\text{mWm}^{-2}\text{K}^{-1}$. Middle plots are absorption by and emission from the atmosphere and bottom plots are absorption by and emission from the surface. Top plots are the sum of middle and bottom plots. Eighteen years of data from March 2000 to February 2018 are used.

Figure 3: Annual mean of entropy production by longwave irradiance emission to space minus entropy production by absorption of shortwave irradiance (Eq. 20). Eighteen years of data from March 2000 to February 2018 are used.

Figure 4: Standard deviation of annual mean entropy produced by absorption of shortwave irradiance (left) and by emission of longwave irradiance to space (right) in $\text{mWm}^{-2}\text{K}^{-1}$. Top plots are absorption by and emission from the atmosphere and surface. Middle plots are absorption by and emission from the atmosphere and bottom plots are absorption by and emission from the surface. Eighteen years of data from March 2000 to February 2018 are used.

Figure 5: Standard deviation of annual mean entropy produced by longwave irradiance emission to space minus annual mean entropy produced by shortwave irradiance absorption. Eighteen years of data from March 2000 to February 2018 are used.

Figure 6: Left) Annual mean of entropy production by non-radiative irreversible processes. Right) Standard deviation of annual mean entropy produced by non-radiative irreversible processes. Eighteen years of data from March 2000 to February 2018 are used.

Figure 7: Global annual anomalies of top-of-atmosphere (TOA) emitted longwave irradiance versus global annual anomalies of TOA shortwave net irradiance. Eighteen years of data from March 2000 to February 2018 are used. The solid line is linear regression line with the slope of 0.39. The slope of the dashed line is 1. R and m are, respectively, correlation coefficient and slope.

Figure 8: Annual global anomalies of entropy production by (left) longwave irradiance emission to space, (middle) shortwave irradiance absorption, and (right) longwave irradiance emission minus shortwave absorption as a function of annual global anomalies of absorptivity (i.e. absorbed shortwave irradiance divided by top-of-atmosphere downward shortwave irradiance). Eighteen years of data from March 2000 to February 2018 are used. R and m are, respectively, correlation coefficient and slope.

Figures

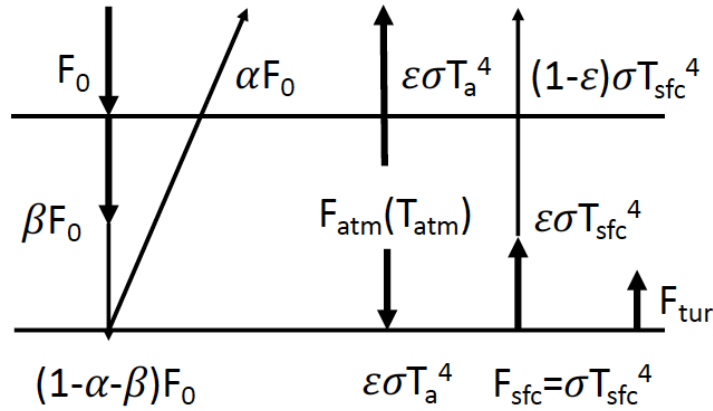


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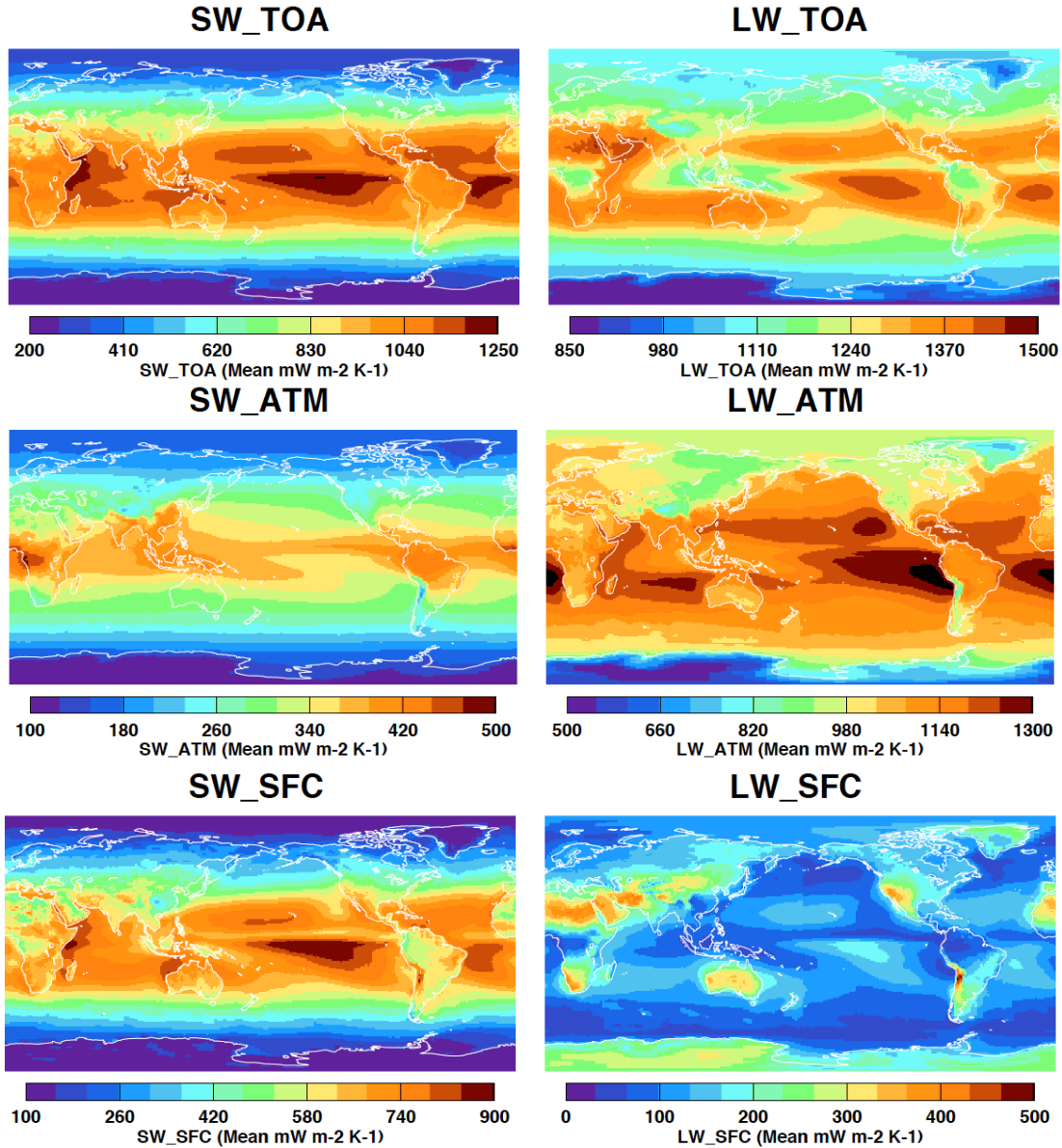


Figure 2: Annual mean entropy produced by absorption of shortwave irradiance (left) and emission of longwave irradiance to space (right) in $\text{mWm}^{-2}\text{K}^{-1}$. Middle plots are absorption by and by emission from the atmosphere and bottom plots are absorption by and emission from the surface. Top plots are the sum of middle and bottom plots. Eighteen years of data from March 2000 to February 2018 are used.

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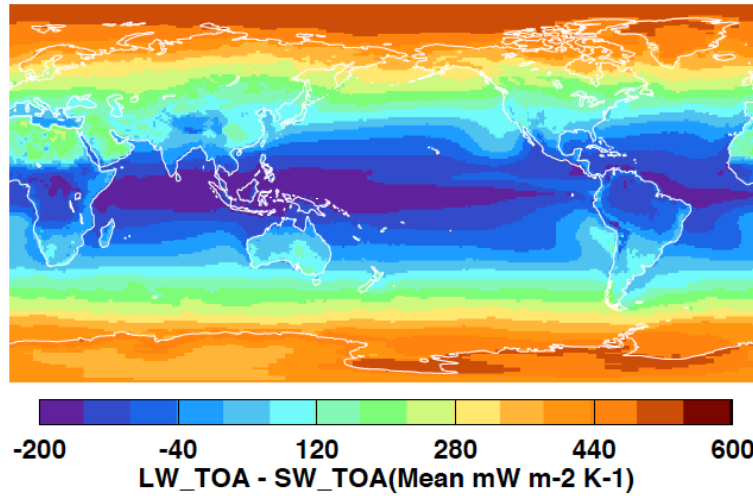


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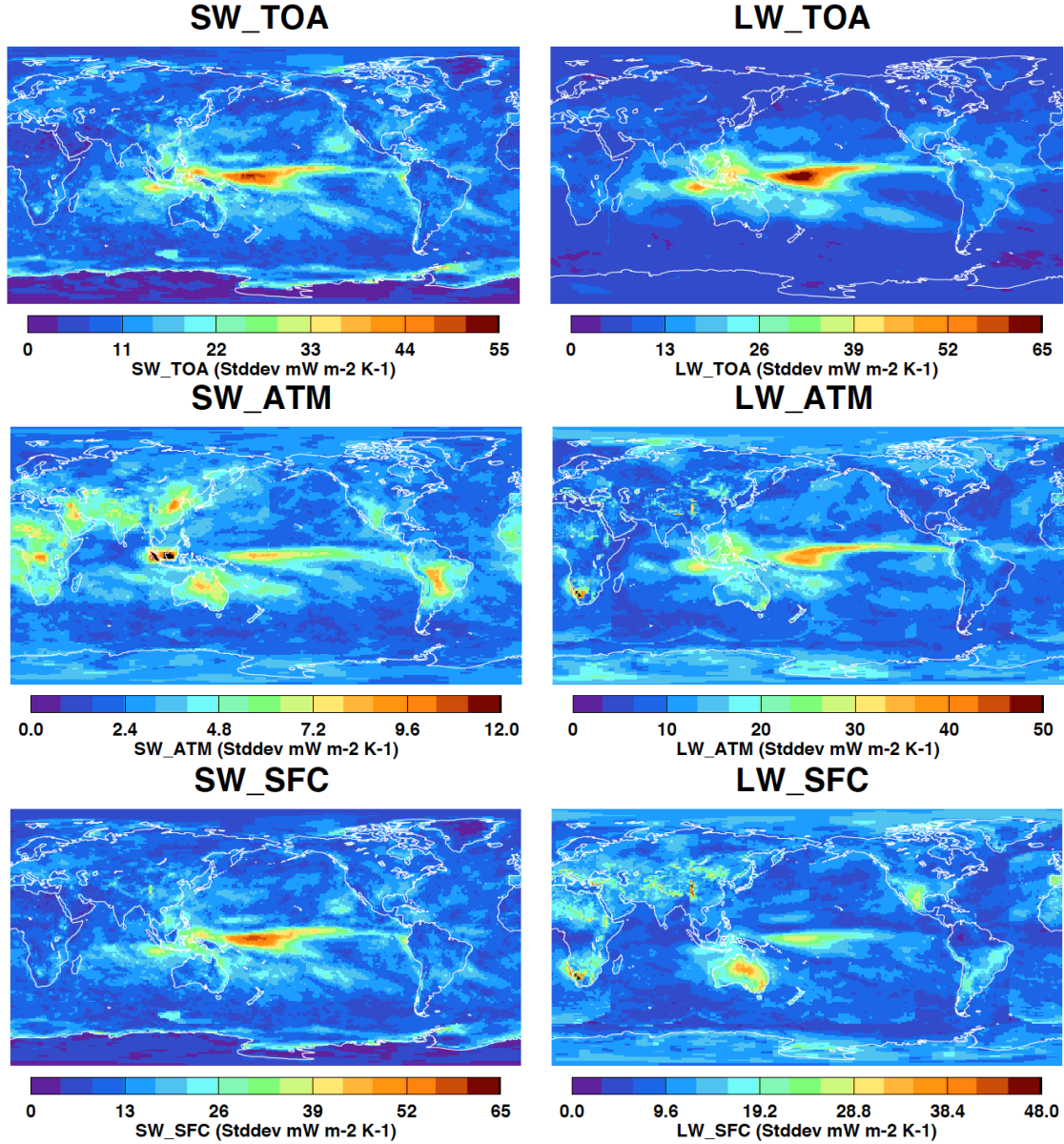


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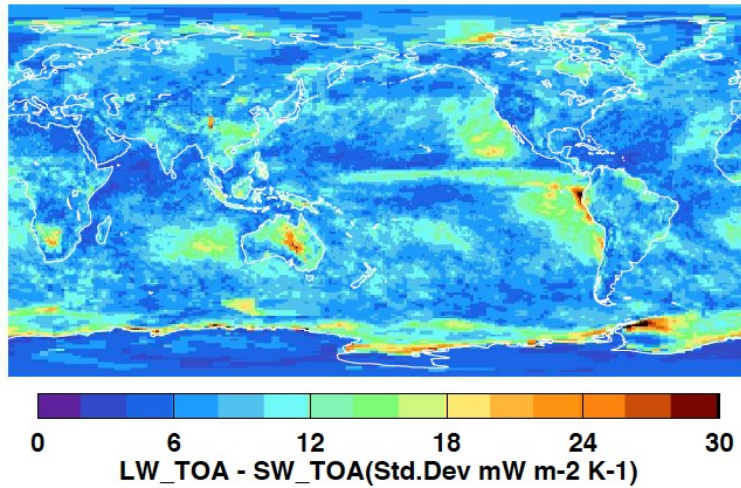


Figure 5: Standard deviation of annual mean entropy produced by longwave irradiance emission to space minus annual mean entropy produced by shortwave irradiance absorption. Eighteen years of data from March 2000 to February 2018 are used.

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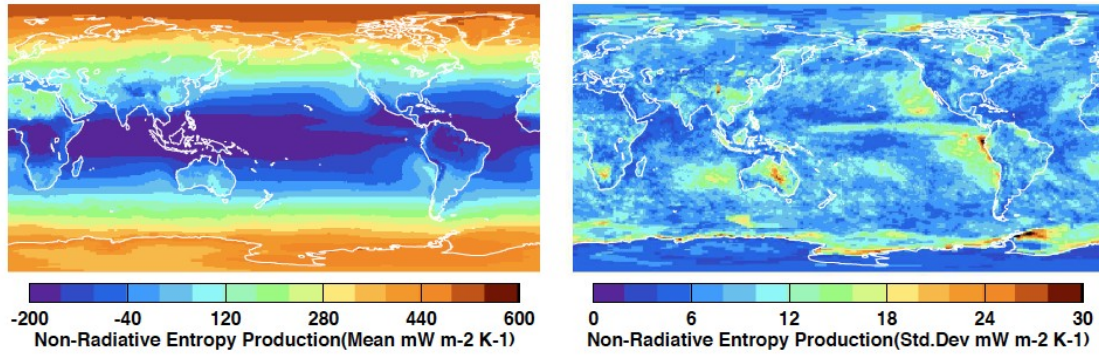


Figure 6: Left) Annual mean of entropy production by non-radiative irreversible processes. Right) Standard deviation of annual mean entropy produced by non-radiative irreversible processes. Eighteen years of data from March 2000 to February 2018 are used.

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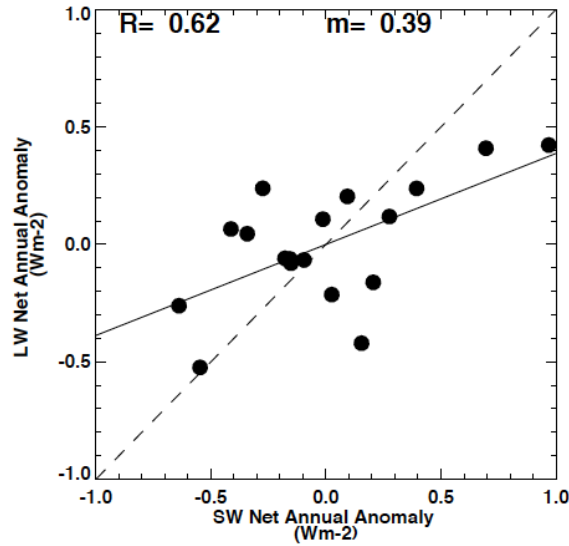


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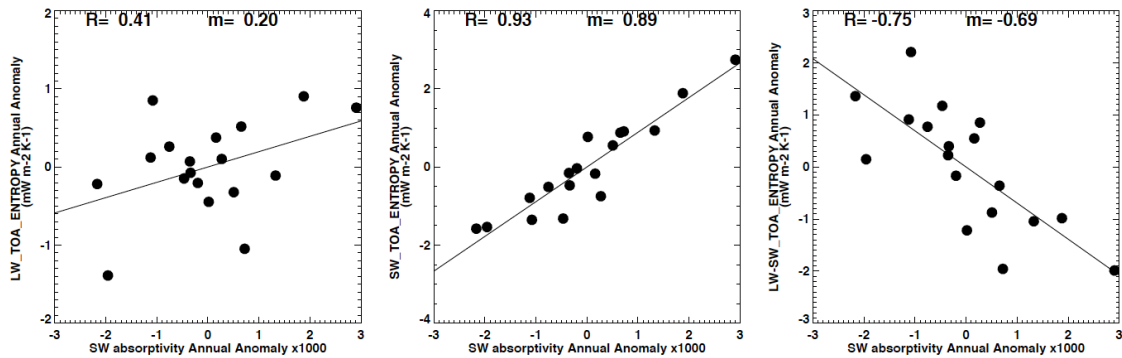


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