

1 **The seasonal cycle of atmospheric heating and temperature**

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ABSTRACT

The seasonal cycle of the heating of the atmosphere is divided into a component due to direct solar absorption in the atmosphere, and a component due to the flux of energy from the surface to the atmosphere via latent, sensible, and radiative heat fluxes. Both observations and coupled climate models are analyzed. The vast majority of the seasonal heating of the Northern extratropics (78% in the observations and 67% in the model average) is due to atmospheric shortwave absorption. In the southern extratropics, the seasonal heating of the atmosphere is entirely due to atmospheric shortwave absorption in both the observations and the models, and the surface heat flux opposes the seasonal heating of the atmosphere. The seasonal cycle of atmospheric temperature is surface amplified in the northern extratropics and nearly barotropic in the southern hemisphere; in both cases, the vertical profile of temperature reflects the source of the seasonal heating.

In the northern extratropics, the seasonal cycle of atmospheric heating over land differs markedly from that over the ocean. Over the land, the surface energy fluxes compliment the driving absorbed shortwave flux; over the ocean they oppose the absorbed shortwave flux. This gives rise to large seasonal differences in the temperature of the atmosphere over land and ocean. Downgradient temperature advection by the mean westerly winds damps the seasonal cycle of heating of the atmosphere over the land and amplifies it over the ocean. The seasonal cycle in the zonal energy transport is 4.1 PW.

Finally, we examine the change in the seasonal cycle of atmospheric heating in 11 CMIP3 models due to a doubling of atmospheric carbon dioxide from pre-industrial concentrations. We find the seasonal heating of the troposphere is everywhere enhanced by increased short-wave absorption by water vapor; it is reduced where sea ice has been replaced by ocean which

27 increases the effective heat storage reservoir of the climate system and thereby reduces the
28 seasonal magnitude of energy fluxes between the surface and the atmosphere. As a result, the
29 seasonal amplitude of temperature increases in the upper troposphere (where atmospheric
30 shortwave absorption increases) and decreases at the surface (where the ice melts).

31 1. Introduction

32 Averaged annually and globally, the atmosphere receives approximately two thirds of its
33 energy input from upward energy fluxes from the surface (longwave, sensible, and latent
34 heat fluxes) and the remaining one third from direct atmospheric absorption of shortwave
35 radiation (Kiehl and Trenberth 1997; Trenberth et al. 2009; Trenberth and Stepaniak 2004).
36 This result follows from the fact that (A) the atmosphere is more transparent than absorb-
37 ing in the shortwave bands resulting in more shortwave radiation absorbed at the surface
38 than within the atmosphere itself (Gupta et al. 1999), and (B) the surface is in energetic
39 equilibrium (provided that energy is not accumulating in the system) such that the net
40 shortwave radiation absorbed at the surface is balanced by an upward energy flux toward
41 the atmosphere (Dines 1917). As a result, in the annual average, the atmosphere is heated
42 from below (by surface fluxes) rather from above (from atmospheric shortwave absorption).
43 Energy is primarily redistributed vertically from the input region at the surface to the region
44 of net radiative cooling aloft by convection (Held et al. 1993).

45 The seasonal input of energy into the atmosphere has received less attention in the liter-
46 ature and is not subject to the same constraints imposed on the annual average. Specifically,
47 the oceans can store large quantities of energy in the annual cycle (Fasullo and Trenberth
48 2008a) and there is therefore no requirement that the seasonal variations in net shortwave
49 absorption at the surface be balanced by an upward energy flux toward the atmosphere.
50 Consequently, although the atmosphere is more shortwave transparent than shortwave ab-
51 sorbing during all seasons, there is no *a priori* requirement that the atmosphere be heated
52 from below rather than above in the annual cycle. The relative contributions of atmospheric

53 shortwave absorption and surface heating to the seasonal heating of the atmosphere is an
54 unresolved issue in climate dynamics and it is the focus of this study.

55 The seasonal flow of energy in the climate system has been thoroughly documented
56 by Trenberth and Stepaniak (2004) and Fasullo and Trenberth (2008b). There, it was
57 demonstrated that the large seasonal variations in shortwave radiation at the top of the
58 atmosphere (TOA) were primarily balanced by an energy flux into the ocean. In this regard,
59 the seasonal input of energy into the the atmospheric column is the residual of two large
60 terms: the net shortwave flux at the TOA and the net energy flux through the surface.
61 To better elucidate the seasonal heating of the atmosphere, we take the unconventional
62 approach of dividing the surface energy flux into solar and non-solar components. This
63 choice is motivated by the fact that the solar flux through the surface is an exchange of
64 energy between the Sun and the surface whereas the non-solar surface energy flux represents
65 an energy exchange between the surface and the atmosphere that (potentially) serves to
66 heat the atmosphere seasonally. Our framework shows that the vast majority of the seasonal
67 heating of the atmosphere is due to atmospheric absorption of shortwave radiation as opposed
68 to seasonal variations in the upward energy flux from the surface to the atmosphere.

69 The division of seasonal atmospheric heating into upward surface fluxes and shortwave
70 atmospheric absorption has implications for the vertical structure of the seasonal temper-
71 ature response, the hydrological cycle, the temporal phasing of the seasonal cycle and, the
72 change in seasonality due to global warming. Heating the air column from below destabi-
73 lizes the air column often triggering convection and a vertical temperature profile at the
74 adiabatic lapse rate (Manabe and Wetherald 1967). In contrast, heating the atmosphere at
75 an upper level stabilizes the air column and results in a temperature response that mimics

76 the radiative heating profile (Fels 1985). We demonstrate that, throughout most for the do-
77 main, the annual cycle of temperature has a vertical profile that reflects the distribution of
78 shortwave atmospheric heating. The partitioning of atmospheric heating into surface fluxes
79 and atmospheric absorption is also useful for understanding the strength of the hydrological
80 cycle which is intimately connected to the upward surface fluxes (Takahashi 2009).

81 The phase of the seasonal cycle of temperature within the atmosphere is also dictated
82 by the heating source. For example, the upward energy fluxes from the surface to the
83 atmosphere lag the insolation (especially over the ocean) because the surface must first heat
84 up before it can flux energy to the atmosphere. In contrast, shortwave absorption in the
85 atmosphere is phase locked to the insolation. Therefore, an atmosphere that is heated by
86 shortwave absorption will have a phase lead in the seasonal cycle of temperature relative to
87 an atmosphere that is heated by surface fluxes.

88 Changes in the seasonal heating of the atmosphere due to increasing CO₂ concentrations
89 will have a direct impact on the seasonal cycle of atmospheric temperatures. The source
90 of the seasonal heating of the atmosphere is anticipated to change with global warming
91 as a consequence of (1) reduced sea ice extent leading to a larger effective surface heat
92 capacity (Dwyer et al. 2012) and a smaller seasonal cycle of surface heat fluxes upward to
93 the atmosphere, and (2) the moistening of the atmosphere (Held and Soden 2006) leading
94 to an enhanced seasonal cycle shortwave atmospheric absorption because water vapor is a
95 strong shortwave absorber (Arking 1996) and the largest increases in shortwave absorption
96 occur in the summer (when insolation is the greatest). Predicting how the seasonal cycle of
97 atmospheric temperature will respond to global warming hinges critically on understanding
98 how the seasonal heating of the atmosphere will change.

99 In this paper, we analyze the seasonal heating of the atmosphere in observations and in
 100 an ensemble of state of the art coupled climate models. We use observations and models in
 101 conjunction because the surface heat fluxes are poorly constrained in the observations and
 102 the similarities of the results in the observations and models demonstrate that the conclusions
 103 we reach are a consequence of the fundamental physics in both nature and the models, and
 104 not due to the uncertainty in the observational fluxes. This paper is organized as follows.
 105 In Section 2 we describe the data sets and models used and the basic method of analysis we
 106 will use throughout this study. In Section 3 we partition the zonal average seasonal heating
 107 of the atmosphere into shortwave atmospheric absorption and upward surface heat fluxes.
 108 We also analyze the spatial structure of the seasonal amplitude of atmospheric temperature.
 109 In Section 4, we trace the seasonal flow of energy through the climate system. We then
 110 analyze the seasonal cycle of energy fluxes averaged over the extratropical regions of each
 111 hemisphere and quantify seasonal energy fluxes between the ocean domain and the land
 112 domain. In Section 5 we analyze the change in the seasonal cycle due to a doubling of CO₂
 113 in an ensemble of coupled climate models. A summary and discussion follows in Section 6.

114 2. Methods and datasets

115 a. Methods

116 The vertically integrated atmospheric energy budget is expressed as :

$$117 \frac{1}{g} \int_0^{P_s} \frac{\partial (c_p T + Lq)}{\partial t} dP = SWABS + SHF - OLR - \frac{1}{g} \int_0^{P_s} \left(\vec{U} \cdot \nabla E + \tilde{E} \nabla \cdot \vec{U} \right) dP \quad (1)$$

118 where P is the pressure (P_S is the surface pressure), E is moist static energy, OLR is the
 119 outgoing longwave radiation at the top of the atmosphere (TOA), and the term on the far
 120 right is the atmospheric energy flux divergence in advective form; \vec{U} is the horizontal velocity,
 121 and g is the acceleration of gravity. The symbol (\sim) represents the departure from the vertical
 122 average and the integration represents the mass integral over the atmospheric column. The
 123 advective form of the vertically integrated energy flux divergence is derived and discussed in
 124 the Appendix. $SWABS$ is the shortwave absorption within the atmospheric column defined
 125 as:

$$126 \quad SWABS = SW \downarrow_{TOA} - SW \uparrow_{TOA} + SW \uparrow_{SURF} - SW \downarrow_{SURF}, \quad (2)$$

127 and represents the direct heating of the atmosphere by the sun. SHF is the upward flux of
 128 energy from the surface to the atmosphere and is composed of sensible heat fluxes ($SENS$),
 129 latent heat fluxes (LH) and longwave fluxes (LW) from the surface and the atmosphere:

$$130 \quad SHF = SENS \uparrow_{SURF} + LH \uparrow_{SURF} + LW \uparrow_{SURF} - LW \downarrow_{SURF}. \quad (3)$$

131 We emphasize that SHF is defined as the energy exchange between the surface and the
 132 atmosphere and does not include the shortwave flux through the surface because the net
 133 shortwave flux at the surface represents an exchange of energy between the sun and the
 134 surface; it does not directly enter the atmospheric energy budget. A schematic of the energy
 135 exchange between the sun, atmosphere and, the surface is presented in Figure 1. Conceptu-
 136 ally, the atmospheric energy tendency on the left hand side of Equation 1 is the difference
 137 between the atmospheric heating (by both surface fluxes – SHF – and by direct solar ab-

138 sorption within the atmosphere – *SWABS*) and the losses of energy from the atmospheric
139 column (by the emission of outgoing longwave radiation and the atmospheric energy flux
140 divergence).

141 We wish to analyze the role of the energy fluxes in amplifying/dissipating the seasonal
142 cycle of temperature in the atmosphere. The magnitude of the seasonal cycle in temperature
143 is quantified as the amplitude of the seasonal harmonic of temperature. The seasonal ampli-
144 tude of the energy fluxes in Equation 1 is defined as the amplitude of the seasonal harmonic
145 of the energy flux in phase with the solar insolation; this definition accounts for both the sea-
146 sonal magnitude and phase of the energy fluxes with positive values amplifying the seasonal
147 cycle of temperature in the atmosphere and negative values reducing the seasonal amplitude
148 of temperature. We note that the conclusions reached in the this manuscript do not depend
149 on the choice of phase used to define the seasonal cycle. The same qualitative conclusions are
150 reached if we define the seasonal amplitude using the phase of the atmospheric temperature
151 or the total atmospheric heating ($SWABS + SHF$).

152 *b. Data sets and model output used*

153 1) OBSERVATIONAL DATA

154 The longwave and shortwave radiative fluxes at the TOA and the shortwave fluxes at
155 the surface are from the Clouds and Earth’s Radiant Energy System (CERES) experiment
156 (Wielicki et al. 1996). We use the long term climatologies of the CERES TOA fluxes from
157 Fasullo and Trenberth (2008a) that are corrected for missing data and global average energy
158 imbalances. The surface shortwave radiation is taken from the CERES “AVG” fields that

159 are derived by assimilating the satellite observations into a radiative transfer model to infer
160 the surface radiative fluxes (Rutan et al. 2001). All calculations are performed separately for
161 each of the four CERES instruments (FM1 and FM2 on Terra from 2000 -2005 and FM3 and
162 FM4 on AQUA from 2002 – 2005). We then average the results over the four instruments
163 to compose monthly averaged climatologies over the observation period.

164 The atmospheric heat flux divergences are calculated using the velocity, temperature,
165 specific humidity and geopotential fields from the ERA interim analysis. We use the 6 hourly
166 instantaneous fields with a horizontal resolution of 1.5° and 37 vertical levels to calculate the
167 atmospheric moist static energy fluxes using the advective form of the energy flux equations
168 (Trenberth and Smith 2008) as discussed in the Appendix. This method satisfies the mass
169 budget by construction and allows us to accurately calculate the energy flux divergences
170 without explicitly balancing the mass budget with a barotropic wind correction. We note
171 that, the calculated heat flux divergences are in close agreement with similar calculations by
172 Fasullo and Trenberth (2008b) and the conclusions reached in this study do not depend on
173 the dataset and methodology used to calculate the atmospheric energy fluxes. We calculate
174 the vertical integral of the atmospheric energy tendency as follows: (1) the temperature
175 and specific humidity tendency at each level is calculated as the centered finite difference of
176 the monthly mean fields and (2) the mass integral is calculated as the weighted sum of the
177 tendencies at each level multiplied by c_P and L respectively.

178 The *SHF* is calculated as the residual of the other terms in Equation 1, similar to
179 Trenberth (1997).

180 2) MODEL OUTPUT

181 We use model output from the World Climate Research Programme’s (WCRP) Cou-
182 pled Model Intercomparison Project phase 3 (CMIP3) multi-model database (Meehl et al.
183 2007): a suite of standardized coupled simulations from 25 global climate models that
184 were included in the International Panel on Climate Change’s Fourth Assessment Report
185 (<https://esgcat.llnl.gov:8443/index.jsp>). We use the pre-industrial (PI) simulations in which
186 greenhouse gas concentrations, aerosols, and solar forcing are fixed at 1850 levels and the
187 models are run forward for 400 years. We calculate model climatologies from the last 20
188 years of the PI simulations. The 16 coupled models that provided all the output fields that
189 are required for the analysis presented in this study are listed in Table 1.

190 *SWABS* and *SHF* are calculated directly from the radiative and turbulent fluxes at the
191 TOA and surface using Equations 2 and 3. The atmospheric column integrated energy
192 tendency is calculated from the finite difference of the monthly mean vertical integral of
193 the moist static energy. The atmospheric energy flux divergence is then calculated as the
194 residual of Equation 1. We note that the method we use for calculating the *SHF* differs
195 markedly between the models (where the surface energy fluxes are standard model output)
196 and the observations (where surface energy fluxes are scarce and are diagnosed as a residual
197 in this study).

3. Zonal average seasonal cycle of atmospheric heating

Figure 2 shows the observed seasonal variations of the zonally averaged *SWABS* and *SHF* with the annual average at each latitude removed. The seasonal cycle of *SWABS* is in phase with the solar insolation and has a seasonal amplitude of order 60 W m^{-2} in the extratropics. In the global and annual average, 21% of the incident shortwave radiation at the TOA is absorbed in the atmosphere (while 49% is absorbed at the surface and 30% is reflected back to space). The spatio-temporal structure of *SWABS* is predominantly ($R^2 = 0.96$) due to the spatio-temporal distribution of insolation; the spatial and seasonal variations in the shortwave absorptivity of the atmosphere make a very small contribution to the spatio-temporal distribution of *SWABS* (i.e. *SWABS* is well approximated by assuming a spatial and temporal invariant fraction of the insolation is absorbed within the atmosphere). We find that, using the isotropic shortwave model of Donohoe and Battisti (2011), approximately 92% of *SWABS* (in the global average) is absorbed on the downward pass from the TOA to the surface and the enhancement of *SWABS* due to reflection of the Earth's surface is minimal. We note that Kato et al. (2011) recently demonstrated that CERES surface shortwave fluxes have uncertainties of order 10 W m^{-2} associated with uncertainties in the cloud and aerosol fields assimilated into the radiation model used to derive the fields. Projecting these errors onto the seasonal cycle of *SWABS* requires knowledge of the spatio-temporal structure of those uncertainties that are unknown and beyond the scope of this work. If the errors in CERES surface shortwave fluxes are zonally uniform and project perfectly onto the annual cycle (worse case scenario) the seasonal anomalies in *SWABS* derived here have uncertainties of order 20%. If the errors are random in space and time,

220 the errors in the seasonal anomalies in *SWABS* are less than 1%.

221 The seasonal variations of *SHF* are substantially smaller than the seasonal variations in
222 *SWABS*. Over the southern ocean (between 30°S and 70°S) the seasonal variation in *SHF*
223 oppose the seasonal heating of the atmosphere. In contrast, over the latitudes that have
224 a substantial land fraction (between 45°N and 70°N and poleward of 70°S) the seasonal
225 variations in *SHF* are in phase with the insolation. We understand these results as follows.
226 The land surface is nearly in energetic equilibrium in the annual cycle due to the small
227 heat capacity of the land surface (Fasullo and Trenberth 2008b), and so in these regions
228 the seasonal variations in shortwave radiation at the surface are balanced by upwards *SHF*
229 fluxes to the atmosphere. In contrast, the large heat capacity of the ocean allows the seasonal
230 variations in shortwave radiation at the surface to be stored within the ocean mixed layer
231 and the seasonal variations in surface shortwave radiation are not fluxed to the atmosphere.
232 In fact, the ocean stores more energy seasonally than it absorbs directly from the sun (by as
233 much as 30% in the latitude band of the Southern ocean) due to a net flux of energy from
234 the atmosphere to the ocean (*SHF*) during the warm season.

235 We quantify the contribution of *SWABS* and *SHF* to the seasonal heating of the atmo-
236 sphere as the amplitude of the annual Fourier harmonic in phase with the local insolation
237 (see Section 2a for a discussion). This definition takes into account both the amplitude and
238 phase of the annual cycle of energy fluxes with positive flux amplitudes amplifying the sea-
239 sonal heating of the atmosphere and negative flux amplitudes reducing the seasonal heating
240 of the atmosphere. We point the reader toward Figure 6 as a demonstration of how the total
241 heating of the atmosphere is nearly in phase with the insolation and note that the same
242 qualitative conclusions found here hold if we define the amplitude of the seasonal cycle from

243 the phase of the total heating or the phase of the column averaged atmospheric temperature.
244 At all latitudes, the seasonal amplitude of *SWABS* is positive (*SWABS* is phase locked to
245 the insolation) and exceeds that of *SHF* (solid lines in lower left panel of Figure 3). The
246 seasonal amplitude of *SHF* is negative in the latitudes where ocean is prevalent and positive
247 in the latitudes where land is prevalent. This result coincides with the seasonal phasing of
248 *SHF* relative to the insolation noted over the same regions in Figure 2. We show in the
249 bottom right panel of Figure 3 the fraction of atmospheric heating due to *SWABS*, defined
250 as $|SWABS| / (|SWABS| + H(|SHF|))$ where $||$ brackets denote seasonal amplitudes and
251 H is the Heaviside function. *SWABS* accounts for the vast majority of the seasonal at-
252 mospheric heating at all latitudes and all of the seasonal heating of the atmosphere in all
253 latitude bands where ocean is prevalent.

254 The dominance of *SWABS* (relative to *SHF*) in the seasonal heating of the atmosphere
255 is a stark contrast to the annual average atmospheric heating (top panels of Figure 3) where
256 heating by *SHF* exceeds that by *SWABS* at all latitudes. In the global and annual average,
257 the atmospheric heating is due to approximately two parts *SHF* and one part *SWABS* (see
258 the top right panel of Figure 3). Conceptually, this result follows from the fact that, although
259 the atmosphere is more transparent than absorbing resulting in more shortwave radiation
260 reaching the surface than is absorbed within the atmosphere on all timescales, the annual
261 average surface energy budget requires that the surface shortwave flux be balanced by *SHF*
262 to the atmosphere. On shorter timescales, such as the seasonal cycle, no such balance is
263 required: a significant fraction of the shortwave flux to the surface can be stored in the
264 surface layer on shorter timescales.

265 *SWABS* and *SHF* from CMIP3 PI models are co-plotted with the observations in Figure

266 3 where the shading represents $\pm 1\sigma$ about the CMIP3 ensemble average. The observations
267 and the models are in excellent agreement in all regions and seasons. The only significant
268 difference between the models and observations is the annual the average *SHF* in the Arctic
269 that is biased low in the models. Walsh et al. (2002) previously demonstrated that the
270 downwelling surface fluxes are lower in the models than the observations in this region due
271 to more clouds and optically thicker clouds than are observed and we believe this is the root
272 cause of the bias. We emphasize that *SHF* is calculated as a residual from Equation 1 in the
273 observations and directly from Equation 3 in the models; the correspondence of the relative
274 contributions of *SWABS* and *SHF* to the seasonal and annual average atmospheric heating
275 suggests that our conclusions are a consequence of fundamental physics in Nature and in the
276 models and not due to methodology of our calculations or the observational field used here.

277 The source (*SHF* versus *SWABS*) of the seasonal heating of the atmosphere manifests
278 itself in the spatial structure of the seasonal amplitude of temperature: observations are
279 shown in Figure 4. *SHF*s primarily heat the lower troposphere whereas atmospheric heating
280 by *SWABS* is nearly barotropic throughout the troposphere as can be seen in the left panel
281 of Figure 4 which shows the vertical distribution of the seasonal amplitude of *SWABS* aver-
282 aged poleward of 40° from a GFDL2.1 simulation of the pre-industrial climate¹. The nearly
283 barotropic profile of shortwave absorption in the troposphere is consistent with the profile of
284 water vapor absorption (Chou and Lee 1996) whereas the isolated maximum in the strato-
285 sphere is due to ozone. In the latitude bands in which land is prevalent (poleward of 45°N and
286 70°S), the seasonal amplitude of *SHF* is positive (see Figure 3) and the seasonal amplitude
287 of temperature is surface amplified. In the latitude bands where ocean is prevalent (30°N

¹The shortwave atmospheric heating is not readily available in the CMIP3 archive.

288 to 40°N and 30°S to 70°S) and *SWABS* dominates the seasonal heating and the seasonal
289 amplitude of temperature is nearly barotropic in the troposphere. The seasonal amplitude of
290 temperature in the CMIP3 models (not shown) have a qualitatively similar structure in the
291 latitude-level plane as in the observations. The seasonal amplitude of temperature therefore
292 reflects the spatial structure of the atmospheric heating, suggesting that the seasonal heating
293 of the atmosphere is not well mixed through the atmospheric column (i.e. via convection).
294 In summary, the seasonal heating of the atmosphere is predominantly due to *SWABS* and
295 the vertical structure of the atmospheric response (the seasonal amplitude of temperature)
296 reflects the dominant source of heating.

297 **4. The seasonal cycle of energy fluxes**

298 The source of the seasonal heating of the atmosphere was discussed in the previous
299 section. We now ask: how does the atmosphere balance the energy input from *SWABS* and
300 *SHF* over the seasonal cycle? We start by looking at the zonal average seasonal energy
301 balance. We then analyze the seasonal energy balance averaged over the extratropics in each
302 hemisphere (subsection a) and the contribution of atmospheric energy transport between
303 land and ocean regions to the seasonal cycle energy budget (subsection b). Finally, we
304 demonstrate that the the source of the seasonal heating has implications for the vertical
305 structure of the seasonal temperature response within the different regions (subsection c).

306 The seasonal amplitude (defined again as the seasonal amplitude of the annual Fourier
307 harmonic in phase with the insolation) of all the atmospheric energy fluxes in Equation 1
308 are shown in Figure 5 for both the observations (solid lines) and the CMIP3 models (shad-

309 ing). The models and observations are in excellent agreement and the bulk structure of
310 the seasonal amplitude at different latitudes is robust across the suite of CMIP3 models.
311 With the exception of the tropics, meridional heat transport, OLR, and the loss of energy
312 to atmospheric storage (the negative atmospheric energy tendency) all have negative sea-
313 sonal amplitudes and thus act to damp the seasonal input of energy into the atmosphere².
314 In general the seasonal heating of the atmosphere (by *SWABS* plus *SHF*) is balanced by
315 (listed in order of decreasing importance): (A) reduced (meridional) heat transport conver-
316 gence (MHT), (B) enhanced OLR and, (C) atmospheric energy storage. As the atmosphere
317 accumulates energy seasonally and temperature increases (term C) it exports energy dy-
318 namically to adjacent regions (term A) and radiatively to space (term B) and we can think
319 of these three terms of the response of the atmosphere to seasonal heating. Energetic con-
320 straints require that the combined response be equal in magnitude to the combined heating
321 by *SHF* and *SWABS* and Figure 5 shows that the atmospheric response is largest in the
322 regions where *SHF* amplifies the seasonal cycle. The relative magnitudes of the response
323 terms (OLR vs. meridional heat transport convergence vs. tendency) have been discussed
324 by Donohoe (2011) and Donohoe and Battisti (2012) where it was argued that meridional
325 heat transport convergence is the most efficient mechanism for the atmosphere to export
326 energy, followed by OLR and energy storage.

327 The seasonal amplitude of both OLR and meridional heat transport convergence in the

²On the equatorward side of heat transport maximum (between 25°N and 40°N) the meridional heat transport divergence is in phase with the seasonal insolation and the heat transport amplifies the seasonal cycle. This effect is non-local; more energy is exported to the high latitudes in the cold season leading to a cooling of the subtropical atmosphere in the cold season.

328 southern extratropics is muted (with the exception of Antarctica) compared to that in the
329 northern extratropics. This result follows from the fact that both *SWABS* and *SHF* heat
330 the atmosphere in the northern extratropics where as *SHF* reduces the seasonal heating of
331 the atmosphere in the southern extratropics. The total seasonal input of energy to the atmo-
332 sphere is reduced in the southern hemisphere compared with the northern hemisphere and
333 thus the atmospheric response (OLR, heat transport, and energy tendency) is reduced which
334 coincides with the nearly seasonal invariance of storm activity in the southern hemisphere
335 (Trenberth 1991; Hoskins and Hodges 2005).

336 *a. Seasonal energy fluxes averaged over the extratropics*

337 The seasonal cycle of energy fluxes (with the annual average removed) averaged over the
338 extratropics of each hemisphere (poleward of 42°) is shown in the top panels of Figure 6.
339 The observations (solid lines) and CMIP3 ensemble (shading) are in excellent agreement in
340 both the seasonal amplitude of the energy fluxes and the phasing of each term. *SWABS* is in
341 phase with the insolation and has similar seasonal amplitudes in the two hemispheres. In the
342 southern extratropics, *SHF* is out of phase with the insolation; the seasonal heating of the
343 atmosphere is accomplished entirely by *SWABS* and a portion of the seasonal atmospheric
344 heating by *SWABS* is transferred to the ocean via *SHF*. Therefore, the seasonal storage of
345 the energy in the ocean *exceeds* the seasonal variations in shortwave radiation at the surface.
346 In contrast, *SHF* is in phase with the insolation in the NH extratropics. As a consequence,
347 the seasonal amplitude of both OLR and (meridional) heat transport convergence in the
348 northern extratropics is enhanced relative to the seasonal cycle in the southern extratropics

349 as is the seasonal cycle of atmospheric temperature. The atmospheric energy tendency leads
 350 the insolation in both hemispheres. In the southern extratropics the phase lead is 54 days in
 351 the observations and 51 ± 5 days in the CMIP3 PI ensemble (ensemble average and standard
 352 deviation). In the northern extratropics, the energy tendency leads the insolation by 62 days
 353 in the observations and 61 ± 4 days in the CMIP3 ensemble. Stated otherwise, the column
 354 average atmospheric temperature— which is in quadrature phase with the energy tendency—
 355 lags the insolation by approximately 30 days in the northern extratropics and 40 days in the
 356 southern extratropics or approximately one tenth of the annual forcing period. This phase
 357 lag is consistent with a system that is sinusoidally forced and has a linear damping (due to
 358 OLR and MHT energy export) that is approximately an order of magnitude larger than the
 359 heat capacity times the angular frequency of seasonal forcing³.

360 The contribution of the various energy fluxes to the seasonal heating of the extratropical
 361 atmosphere in each hemisphere is summarized in the bottom panels of Figure 6. The left
 362 column shows the seasonal amplitude of the fluxes that heat the atmosphere seasonally (have
 363 positive seasonal amplitudes), the middle column shows the fluxes that damp the seasonal
 364 cycle (have negative seasonal amplitudes) and, the right column shows the atmospheric
 365 energy tendency. By construction, the sum of the heating terms (height of the left column)
 366 is balanced by the sum of the middle and right column. The key difference between the
 367 two hemispheres is that *SHF* serves as a heating term in the northern extratropics and as
 368 a damping term in the southern extratropics. As a result, the seasonal cycle of atmospheric

³The temperature response (T) of a system that is forced at angular frequency f satisfies the equation
 $C \frac{dT}{dt} = -\lambda T + e^{ift}$ where C is the heat capacity and λ is the linear damping (OLR and heat transport
 convergence). The phase lag of the temperature response (relative to the forcing) is $\text{atan}(\frac{fC}{\lambda})$.

369 energy, MHT, and OLR is larger in the northern hemisphere than that in the southern
370 hemisphere.

371 *b. The contrast in seasonal atmospheric energy fluxes between the land and ocean domains*

372 The contrast of the seasonal phasing of SHF in the northern and southern extratropics
373 is best understood by subdividing the northern extratropics into land and ocean domains
374 (Figure 7). We also divide the observed atmospheric heat transport divergence into merid-
375 ional and zonal components. We note that, in the zonal averages that were presented above,
376 the zonal heat transport divergence is zero (by the divergence theorem) and that the zonal
377 heat transport approximates the exchange of energy between the ocean and land domains
378 in the northern extratropics where the coastlines are primarily orientated North to South.
379 Over the land domain, the seasonal amplitude of the SHF is larger than that of $SWABS$ and
380 is in phase with the insolation (upper right panel of Figure 7). We understand this result as
381 follows. First, the heat capacity of the land surface is very small resulting in a surface that
382 is nearly in energetic equilibrium with the seasonal variation in surface shortwave radiation.
383 Hence, the upward SHF s are in phase with the insolation. Second, the atmosphere is more
384 transparent than absorbing for all seasons resulting in a seasonal amplitude of downwelling
385 solar fluxes at the surface that exceeds $SWABS$. Therefore, the seasonal heating of the atmo-
386 sphere over the land domain is dominated by surface energy fluxes as opposed to $SWABS$;
387 this is shown in the lower right panel of Figure 7 and is very much akin to the annual average
388 energy balance.

389 The phase of SHF over the land results in a large seasonal flux of energy to the atmosphere

390 that must be balanced by meridional and zonal energy exports, OLR, and atmospheric
391 storage. Zonal energy fluxes (the dashed black lines in the upper panels of Figure 7) are
392 the dominant mechanism of energy export. The zonal export of energy from the land to the
393 ocean in the summer (and vice versa in the winter) is primarily accomplished by advection
394 of the land-ocean temperature contrast by the time averaged atmospheric flow (not shown).
395 This result agrees with the conclusion of Donohoe (2011) that zonal heat export is the most
396 efficient energy export process for the extratropical atmosphere over the land and ocean
397 domains. (Seasonal variations in MHT also contribute to energy export but the difference in
398 the seasonal cycle of MHT over the land and the ocean domain is minimal; see the bottom
399 panel of Figure 7). The zonal heat export into the ocean domain is equal and opposite
400 that to the land domain and thus tends to amplify the seasonal cycle of the atmospheric
401 temperature and energy fluxes over the ocean domain. This dynamical import of energy
402 to the atmosphere above the ocean domain during the warm season is balanced primarily
403 by energy export to the ocean via *SHF*. We emphasize that the seasonal energy storage in
404 the northern ocean exceeds the seasonal variations in absorbed shortwave radiation at the
405 surface which is a consequence of the zonal atmospheric heat import that ultimately is derived
406 from shortwave heating of the land surface. As a hypothetical illustrative example, if the
407 zonal flow of the atmosphere suddenly ceased in the middle of the summer, the atmosphere
408 over the oceans would start cooling because the seasonal heating by *SWABS* is completely
409 removed by *SHF* (c.f. the height of the red and blue bars in the lower left panel of Figure
410 7). Similarly, in the winter, the ocean provides a source of heating (via *SHF*) that is nearly
411 identical in magnitude to the atmospheric heating by the summer sun (via *SWABS*). The
412 energy flux from the ocean to the atmosphere during the winter attenuates the seasonal

413 cycle of atmospheric temperatures over the land via the zonal atmospheric energy import.
414 The portion of shortwave radiation incident on the land surface during the summer that gets
415 stored in the ocean is returned to the land domain and warms the atmosphere above the land
416 (relative to the purely radiative case) in the winter. The zonal atmospheric energy transport
417 between the ocean and the land in the northern extratropics has a seasonal amplitude of 4.1
418 PW and is of comparable magnitude to the annual mean meridional heat transport in the
419 atmosphere (Fasullo and Trenberth 2008a).

420 *c. The seasonal temperature response by region*

421 The seasonal input of energy into the atmosphere differs markedly between the ocean
422 domain where the input is entirely by *SWABS* with a nearly vertically invariant heating
423 profile throughout the troposphere (see left panel of Figure 4) and the land domain where
424 *SHF* makes a substantial contribution to the lower atmosphere only. The source of seasonal
425 heating is clearly reflected in the vertical structure of the seasonal amplitude of temperature
426 averaged over the land and ocean domains of the northern hemisphere, shown in Figure 8.
427 Over the northern land domain, the seasonal amplitude of temperature is surface amplified
428 (reflecting the role of *SHF*) whereas over the northern ocean domain the seasonal amplitude
429 is nearly barotropic to the tropopause (consistent with the profile of *SWABS*). Averaged
430 over the whole of the northern extratropics, the seasonal amplitude of temperature is slightly
431 surface amplified. The seasonal cycle of temperature averaged over the southern extratropics
432 is nearly barotropic, consistent with the vertical heating profile of *SWABS* only over the
433 southern ocean. The similarity of the vertical profile of seasonal heating and the seasonal

434 temperature response in each region suggests that the troposphere is not well mixed (by
435 vertical turbulent energy fluxes) in the annual cycle; heating at a given vertical level results
436 in a response localized in the vertical. The input of seasonal energy at the surface over
437 land and its subsequent removal at the surface over the ocean (see Subsection b) begs
438 the question: at what vertical level does the zonal heat transport occur and how does the
439 vertical structure of the temperature response reflect the vertical structure of the (zonal)
440 heat transport? Further investigation is under way.

441 **5. The response of the seasonal cycle of the atmosphere** 442 **to CO₂ doubling**

443 We now analyze the impact of the doubling of carbon dioxide on the seasonal heating
444 of the atmosphere by *SWABS* and *SHF* and on the seasonal cycle of temperature. We
445 have two expectations: First, as the globally averaged temperature increases the atmosphere
446 will moisten (Held and Soden 2006) and the percent of incident shortwave insolation that
447 is absorbed in the atmosphere will increase because water vapor is a strong absorber of
448 shortwave radiation (Arking 1996; Chou and Lee 1996). The increase in *SWABS* will be
449 greatest in the summer when the insolation is strongest, resulting in an increase in the
450 amplitude of *SWABS*. Second, the melting of sea ice in the high-latitudes will expose ocean
451 that was previously insulated from seasonal heat uptake. This is akin to replacing land with
452 ocean and will result in a reduction of the seasonal amplitude of *SHF* and thus cause less
453 net seasonal heating of the lower troposphere where ice melts.

454 *a. Model runs used*

455 We analyze output from the 1% CO₂ increase to doubling experiments in the CMIP3
456 archive (Meehl et al. 2007). The initial conditions for each model come from the equilibrated
457 pre-industrial (PI) or in some cases (CCSM, MRI, and ECHAM) the present day (PD)
458 simulations. Atmospheric CO₂ is increased by 1% per year until CO₂ has doubled relative to
459 the PI concentration at 70 years. The simulations are then run forward for an additional 150
460 years with carbon dioxide fixed at twice the PI concentration. We average the model output
461 over the last 20 years of these simulations (years 201-220 after CO₂ has started to ramp
462 up) and compare the climatological fields to their counterparts in that model’s PI (or PD)
463 simulations. The 11 models that provided the necessary output fields used in this section
464 are indicated with a ‘yes’ in the last column of Table 1. Hereafter, we will refer to these runs
465 as the 2XCO₂ runs.

466 *b. Changes in the seasonal heating of the atmosphere due to CO₂ doubling*

467 The top panel of Figure 9 shows the CMIP3 ensemble average difference in the seasonal
468 amplitude of atmospheric heating by *SWABS* and *SHF* between the 2XCO₂ and the PI (or
469 PD) runs. The ensemble average seasonal amplitude of *SWABS* increases by of order 2 W
470 m⁻² in the extratropics due to CO₂ doubling. This change is very robust across models in
471 the extratropics (1 σ about the ensemble average is given by the red shaded error in Figure
472 9 and is positive throughout the extratropics). Averaged over all the models, the fraction
473 of incident shortwave radiation that is absorbed in the atmosphere increases by 0.8% in the
474 annual and global average, from 22.4% in the PI simulations to 23.2% in the 2XCO₂ runs.

475 The change in the seasonal amplitude of *SWABS*, in turn, is consistent with the seasonal
476 amplitude of the insolation at each latitude multiplied by the (0.8%) global and annual
477 average increase in the atmospheric absorptivity (not shown). We use the atmospheric
478 radiative kernels of Previdi (2010) to diagnose the contribution of water vapor shortwave
479 absorption to the change in the seasonal amplitude of *SWABS*. The product of the kernel
480 and the water vapor change due to CO₂ doubling in each CMIP3 model gives the change
481 in atmospheric shortwave heating due to the change in water vapor. The ensemble average
482 change in the seasonal amplitude of that quantity is shown by the dashed red line in Figure
483 9. Water vapor changes account for almost all of the change in the seasonal amplitude of
484 *SWABS* with the exception of the high latitude of the southern ocean where we suspect
485 changes in ozone may also contribute. We note that the change in *SWABS* in the CMIP
486 models due to CO₂ doubling are almost entirely in the clear sky radiative fields (not shown)
487 suggesting that clouds play a minimal role in the *SWABS* changes.

488 To examine the change in the vertical structure of the seasonal heating by *SWABS* due
489 to CO₂ doubling, we show in the left panel of Figure 10 the change (relative to the PI sim-
490 ulation) in the seasonal amplitude of atmospheric shortwave heating averaged poleward of
491 42° in the GFDL 2.1 simulation⁴. The enhanced *SWABS* heating in a moister atmosphere
492 (i.e. in the 2XCO₂ world) is primarily in the upper troposphere where the fractional changes

⁴We did not analyze the change in the vertical distribution of shortwave radiative heating in the other CMIP3 models because these fields are not available in the CMIP3 archive. Given the robust nature of the humidity response to increasing CO₂ however, we anticipate that the change in the vertical structure of shortwave atmospheric heating in the other CMIP3 models will be similar to that shown in the left panel of Figure 10.

493 in water vapor due to CO₂ doubling are largest (not shown). Although the absolute change
494 in specific humidity is smaller in the upper troposphere than in the lower troposphere, the
495 downwelling radiation in the lower troposphere is depleted at the frequencies of shortwave
496 water vapor absorption relative to downwelling radiation in the upper troposphere. As a
497 result, the relatively small changes in specific humidity in the upper troposphere have a
498 disproportionately large local heating effect on the heating aloft but a relatively small im-
499 pact on the column integrated *SWABS*. Integrated over the atmospheric column, shortwave
500 absorption is enhanced by 2Wm⁻² due to more absorption in the wings of water vapor ab-
501 sorption bands in the moister atmospheric column. As a consequence, the seasonal heating
502 of the atmosphere by *SWABS* is enhanced in a warmer world and more so in the upper
503 troposphere than in the lower troposphere.

504 The most pronounced (and robust across models) change in the seasonal heating of the
505 atmosphere due to CO₂ doubling is the reduced seasonal amplitude of *SHF* poleward of
506 60° in both hemispheres (the solid blue line in Figure 9). The vast majority of this change
507 occurs within the sub-domain where sea ice melts, as shown in Figure 9. This result can
508 be understood as follows. Sea ice insulates the ocean from the exchange of energy with
509 the atmosphere (Serreze et al. 2007) so the effective surface heat capacity of a region with
510 extensive sea ice is much smaller than that of an open ocean (i.e. the heat capacity of
511 the ocean mixed layer). As a consequence, the contributions to the seasonal heating of the
512 atmosphere above regions that are covered with sea ice is similar to that above land regions:
513 seasonal variations in *SHF* to the atmosphere amplify the seasonal heating due to *SWABS*
514 (as in the upper right hand panel of Figure 7). Melting the sea ice exposes the atmosphere
515 to the higher heat capacity of the open ocean so seasonal variations in shortwave radiation

516 at the surface over these regions are now balanced by ocean heat storage as opposed to the
517 upward energy fluxes (SHF). Hence, the seasonal flow of energy is now from the atmosphere
518 to the ocean during the summer (as in the upper left panel of Figure 7). Thus the seasonal
519 amplitude of SHF decreases as the ice melts and more of the atmosphere is exposed to the
520 (high heat capacity) ocean mixed layer as was demonstrated by (Dwyer et al. 2012).

521 The seasonal amplitude of SHF increases between 45° and 60° in both hemispheres due
522 to CO_2 doubling. In the northern hemisphere, this is largely due to an increase (relative to
523 the PI simulation) in the seasonal amplitude of downwelling shortwave radiation at the land
524 surface (not shown) resulting in larger seasonal variations in SHF , as required by the surface
525 energy budget given the small heat capacity of the land surface. This process accounts for
526 the vast majority of the increases in the seasonal amplitude of SHF in the midlatitudes of the
527 northern hemisphere (not shown) but it does not explain the increased seasonal amplitude
528 of SHF in the midlatitudes of the southern hemisphere. The cause of the enhanced seasonal
529 amplitude of SHF between $50^\circ S$ and $60^\circ S$ is under further investigation.

530 The change in the seasonal amplitude of all terms in the atmospheric energy budget is
531 shown in the lower panel of Figure 9. Between 40° and 50° in each hemisphere, CO_2 doubling
532 results in more net ($SWABS + SHF$) seasonal heating of the atmosphere. To balance the
533 energy budget, the atmosphere exports more energy meridionally to adjacent regions during
534 the summer while a small amount of the enhanced heating is stored in the column. In
535 contrast, in the high latitude regions where the change in surface energy fluxes leads to
536 less seasonal heating of the atmosphere, the atmosphere imports less energy from the low
537 latitudes during the winter. The atmospheric energy heat transport adjusts to transport
538 energy from the regions with enhanced seasonal heating to regions with reduced seasonal

539 heating suggesting that dynamical as opposed to local radiative feedbacks play a primary
540 role in bringing the system into seasonal energy balance. Local radiative feedbacks do play
541 a role in bringing the Arctic into seasonal balance; the reduced seasonal heating there is
542 partially balanced by a reduced seasonal amplitude of energy loss by OLR.

543 *c. Changes in the seasonal amplitude of temperature due to CO₂ doubling*

544 The spatial structure of the change in the seasonal amplitude of temperature due to
545 CO₂ doubling reflects the change in atmospheric heating induced by a moistening of the
546 atmosphere and the melting of sea ice in the high latitudes. As shown in section 5b, doubling
547 CO₂ causes a robust increase in the seasonal shortwave heating in the upper troposphere
548 throughout the extratropics and a robust decrease in the seasonal heating of the atmosphere
549 by surface fluxes poleward of 60° in both hemispheres. These changes in the seasonal heating
550 of the atmosphere have a clear and robust imprint on the change in the seasonal amplitude of
551 temperature relative to the PI simulations (right panel of Figure 10). The seasonal amplitude
552 of temperature decreases in the lower atmosphere where the seasonal amplitude of *SHF* is
553 reduced in a 2XCO₂ world. In contrast, the seasonal amplitude of temperature is enhanced
554 in the upper troposphere of the extratropics in a 2XCO₂ world where the seasonal amplitude
555 of shortwave heating is enhanced. The vertical profile of the change in the amplitude of
556 the seasonal cycle of temperature matches that of the shortwave absorption. The vertical
557 structure of the seasonal temperature response to CO₂ doubling is robust across models, as
558 assessed by a one sample t-test of the change due to CO₂ doubling in each model at the 99%
559 confidence interval (the regions enclosed by the red and blue dashed contours in Figure 10

560 are significantly different from zero).

561 At 30° in each hemisphere, the seasonal amplitude of the near surface temperature is
562 enhanced in the 2XCO₂ simulations despite the reduction in the seasonal amplitude of *SHF*.
563 This behavior is a consequence of a deepening of the subtropical boundary layer in the
564 warmer planet and the climatological phasing of *SHF* in this region which opposes the
565 dominant solar heating (see Figure 5). The surface damping of the seasonal cycle by *SHF*
566 over a deeper layer results in an enhanced seasonal cycle of temperature at the surface in
567 the 2XCO₂ world. The cause of the deepened boundary layer is beyond the scope of this
568 work but we speculate that a reduction of subsidence associated with the weakening (Tanaka
569 et al. 2005) and widening (Seager and Coauthors 2007) of the Hadley circulation is the root
570 cause.

571 **6. Summary and discussion**

572 The seasonal cycle of atmospheric temperature has large socioeconomic and ecological
573 impacts. Both the amplitude and phase of the seasonal cycle are projected to change due to
574 global warming (Mann and Park 1996) and trends in the seasonal cycle have been observed
575 over the last century (Stine et al. 2009; Thomson 1995). Understanding the source of the
576 seasonal heating of the atmosphere is critical to understanding the projected change in the
577 seasonal cycle of temperature in the atmosphere.

578 The seasonal heating of the atmosphere differs markedly from the annual average atmo-
579 spheric heating. While the annual average heating is dominated by upward energy fluxes
580 from the surface, the vast majority of the seasonal heating is due to shortwave absorption

581 within the the atmosphere that is nearly vertically invariant throughout the troposphere.
582 The annual average surface energy budget requires that the net shortwave flux at the sur-
583 face be balanced by an upward energy flux to the atmosphere. The same constraint does
584 not apply to the annual cycle where shortwave surface heating can be balanced by surface
585 energy storage. Thus, although the atmosphere is more shortwave transparent than short-
586 wave absorbing, our results show that the seasonal heating of the atmosphere is dominated
587 by shortwave atmospheric heating because the shortwave absorption is considerable and the
588 insolation that is transmitted to the surface primarily goes into storage (especially over the
589 ocean). In fact, across most of the planet, the atmosphere is seasonally heated by directly
590 absorbing energy from the sun (by *SWABS*) during the summer and subsequently fluxes a
591 portion of this energy to the ocean. In contrast to the annual average, over the seasonal cycle
592 the atmosphere is heated from above and is cooled slightly from below (the global average
593 seasonal amplitude of *SHF* is slightly negative).

594 The limited heat capacity of the land surface requires that seasonal variations in surface
595 solar radiation over the land domain are primarily balanced by upward energy fluxes to the
596 atmosphere so that the heating of the atmosphere over the seasonal cycle is primarily by
597 upward surface energy fluxes (*SHF*) and secondarily by atmospheric shortwave absorption
598 (*SWABS*), as can be seen in the lower right panel of Figure 7. In the midlatitudes of the
599 NH, the gross differences in the seasonal atmospheric heating over the land domain and the
600 ocean domain forces a seasonally varying zonal energy exchange between the land and ocean
601 domain of 4.1 PW which is comparable in magnitude to the annually averaged atmospheric
602 meridional heat transport in each hemisphere. The vertical structure of the seasonal ampli-
603 tude of atmospheric temperature clearly reflects the different contribution of *SWABS* and

604 *SHF* to the net seasonal heating over the land and ocean domains. Where ocean is promi-
605 nent and seasonal heating by *SWABS* is dominant, the seasonal amplitude of temperature
606 is nearly barotropic throughout the troposphere and coincides with the vertical structure of
607 *SWABS*. In contrast, over the land domain the seasonal amplitude of temperature is surface
608 amplified, reflecting the contribution of upward energy fluxes from the surface (*SHF*) to the
609 seasonal heating.

610 The observed energy fluxes documented in this manuscript were calculated from the TOA
611 radiative fluxes and the atmospheric reanalysis. Both have substantial errors. However,
612 we have found similar results using alternative sets of reanalysis and methodologies for
613 calculating the atmospheric heat transport (Fasullo and Trenberth 2008b). The largest
614 uncertainty in our analysis is the TOA and surface shortwave fluxes which have errors of
615 order 10 W m^{-2} Kato et al. (2011). Nonetheless, the qualitative conclusions reached here
616 are robust beyond the observational error; we know the atmosphere is a significant absorber
617 in the shortwave and this leads to substantial seasonal heating by *SWABS*.

618 The change in the seasonal heating of the atmosphere due to CO_2 doubling is a con-
619 sequence of two different physical processes that are robust across the CMIP3 ensemble
620 (Figure 9) and have a clear physical interpretation: First, enhanced CO_2 causes a moisten-
621 ing of the atmosphere which, in turn, causes more shortwave absorption in the troposphere—
622 particularly in the upper troposphere (see the left panel of Figure 10). These effects are
623 most pronounced in the summer, when the insolation is the greatest, leading to an enhanced
624 seasonal cycle of heating in a warmer world. Second, enhanced CO_2 causes a reduction in
625 the area covered by sea ice, which results in more of the seasonal variations in solar inso-
626 lation being transmitted to the (large heat capacity) ocean. Thus the seasonal heating of

627 the atmosphere by upward surface energy fluxes (*SHF*) is reduced in the high latitudes in a
628 $2XCO_2$ world. The change in the seasonal heating of the atmosphere due to CO_2 doubling
629 has a clear imprint on the seasonal amplitude of atmospheric temperature; the seasonal cycle
630 of temperature increases in the upper troposphere of the extratropics (where the seasonal
631 amplitude of *SWABS* increases) and decreases at the surface in the polar regions (where the
632 seasonal amplitude of *SHF* decreases) due to CO_2 doubling (Figure 10). As a consequence,
633 the atmospheric column in a $2XCO_2$ world is stabilized in the summer and destabilized in
634 the winter.

635 In our study, we have formulated the atmospheric energy budget in terms of the shortwave
636 energy absorbed within the atmosphere and the net (non-solar) exchange of energy between
637 the surface and the atmosphere. Our approach differs from the traditional approach of
638 Fasullo and Trenberth (2008a) that views the atmospheric energy budget in terms of the
639 difference between the total energy flux into the top of the atmosphere and the surface.
640 The traditional viewpoint emphasizes the near seasonal balance of insolation at the TOA
641 and the energy flux (solar included) to the surface; by and large, the oceans are seasonally
642 heated by the sun. The traditional approach is less useful for understanding the source
643 of the seasonal heating of the atmosphere (where the seasonal heating of the atmosphere
644 is the residual of the TOA and surface fluxes). Our formulation illuminates the relative
645 importance of atmospheric shortwave absorption and surface energy fluxes for the seasonal
646 cycle of temperature in the troposphere.

647 Our work demonstrates that the atmospheric response to heating is localized in the ver-
648 tical and further suggests that the net radiative forcing at the tropopause (i.e. the Solomon
649 et al. 2007, definition of radiative forcing) is not a useful concept on short timescales be-

650 cause it fails to distinguish between energy absorbed within the atmospheric column and
651 energy absorbed at the surface. The vertical structure of atmospheric heating within the
652 troposphere is irrelevant provided the surface layer is in energetic equilibrium and the tropo-
653 sphere is well mixed in the vertical. Our results demonstrate that neither of these conditions
654 are satisfied in either the climatological or perturbed ($2XCO_2$) seasonal cycles and the atmo-
655 spheric temperature response depends critically on the vertical distribution of the heating.
656 This work begs the question: on what timescales and regimes is the radiative forcing at
657 the tropopause a useful concept and when is the response of the system contingent on the
658 vertical structure of the atmospheric forcing? We hope to explore the impact of the vertical
659 structure of atmospheric forcing on the atmospheric temperature response across a myriad
660 of spatio-temporal scales in future work.

661

662

663 **Derivation of the atmospheric energy budget equation**

664 The vertically integrated atmospheric energy budget equation is developed, starting from
 665 the the dry energy and moisture equations of Trenberth and Smith (2008) at a given vertical
 666 level (their Equations 3a and 5). Multiplying the moisture equation by L , adding it to the
 667 dry energy equation and vertically integrating over pressure levels from the TOA to the
 668 surface (P_S) gives:

$$\begin{aligned} \frac{1}{g} \int_0^{P_S} \frac{\partial}{\partial t} (c_P T + K + Lq) dP + \frac{1}{g} \int_0^{P_S} \vec{U} \cdot \nabla (E + K) dP + \\ \frac{1}{g} \int_0^{P_S} \omega \frac{\partial}{\partial P} (E + K) dP = SWABS + SHF - OLR, \end{aligned} \quad (A1)$$

669 where E is the moist static energy, K is the kinetic energy, ω is the pressure velocity and the
 670 vertically integrated diabatic heating (Q1-QF) have been reorganized into the terms used in
 671 this study. The first term is the energy tendency in the atmospheric column. The second
 672 term is the horizontal advection of energy. The third term represents the vertical advection
 673 of energy and is primarily associated with subsidence warming in regions of net descent and
 674 adiabatic cooling in regions of net ascent. The right hand side of the equation is the net
 675 heating of the atmospheric by radiative and diabatic processes.

676 We can rewrite the vertical advection term (third term on the right) by first integrating

677 by parts and then invoking the Boussinesq approximation:

$$\begin{aligned}
& \frac{1}{g} \int_0^{P_S} \omega \frac{\partial}{\partial P} (E + K) dP \\
&= \frac{1}{g} [(E + K) \omega] \Big|_0^{P_S} - \frac{1}{g} \int_0^{P_S} \left([E + K] \frac{\partial \omega}{\partial P} \right) dP \quad (\text{A2}) \\
&= \frac{1}{g} (E_S + K_S) \frac{\partial P_S}{\partial t} + \int_0^{P_S} (E + K) \nabla \cdot \vec{U} dP .
\end{aligned}$$

678 where E_S and K_S are the moist static and kinetic energy at the surface. If we subdivide the
679 E into the vertical average, $[E]$, and an anomaly from the vertical average, \tilde{E} we can make
680 one additional simplification and gain insight into the physical interpretation of equation
681 A2:

$$\begin{aligned}
& \frac{1}{g} \int_0^{P_S} \omega \frac{\partial}{\partial P} (E + K) dP \\
&= \frac{1}{g} (E_S + K_S) \frac{\partial P_S}{\partial t} + \frac{[E + K]}{g} \int_0^{P_S} \nabla \cdot \vec{U} dP + \frac{1}{g} \int_0^{P_S} (\tilde{E} + \tilde{K}) \nabla \cdot \vec{U} dP . \quad (\text{A3})
\end{aligned}$$

682 If the mass of the atmospheric column is conserved, both the surface pressure tendency and
683 the vertical integral of the divergence will be zero and only the third term on the right of
684 equation A3 remains. This term says that energy is input into the column (the energy flux
685 divergence on the LHS is negative) when there is convergence at levels of relatively high E
686 ($\tilde{E} > 0$) and divergence at levels of relatively low E ($\tilde{E} < 0$). Since E increases with height
687 in the atmosphere (with the exception of the boundary layer), this statement says that the
688 column gains energy when there is convergence aloft and divergence at the surface as is the
689 case on the poleward side of the thermally direct Hadley cell (the subtropics) where the
690 vertical structure of horizontal divergence forces large scale subsidence warming. If the mass
691 of the column is not conserved, then mass balance requires that

$$\frac{1}{g} \int_0^{P_s} \nabla \cdot \vec{U} dP = -\frac{1}{g} \frac{\partial P_s}{\partial t}. \quad (\text{A4})$$

692 (In equation A4, we have ignored the mass source associated with evaporation minus pre-
 693 cipitation which is two orders of magnitude smaller than the other terms (Trenberth 1997)).
 694 Substituting Equation A4 into equation A3 results in the near cancellation of the first two
 695 terms; the two terms differ by the energy contrast between the surface and the column
 696 average energy. Using *tilde* to represent the anomaly from the vertical average gives

$$\begin{aligned} & \frac{1}{g} \int_0^{P_s} \omega \frac{\partial}{\partial P} (E + K) dP \\ & = \frac{1}{g} \left(\tilde{E}_s + \tilde{K}_s \right) \frac{\partial P_s}{\partial t} + \frac{1}{g} \int_0^{P_s} \left(\tilde{E} + \tilde{K} \right) \nabla \cdot \vec{U} dP \quad . \quad (\text{A5}) \end{aligned}$$

697 The magnitude of $\left(\tilde{E}_s + \tilde{K}_s \right)$ in the first term is comparable to that of the magnitude
 698 of \tilde{E} in the integrand of the second term. \tilde{E} increases with pressure and crosses zero in
 699 the mid-troposphere. $\nabla \cdot \vec{U}$ also has a simple vertical structure with divergence aloft and
 700 convergence at the surface (e.g. in the upper and lower branches of the Hadley cell) or vice
 701 versa. Therefore, the magnitude of the second term in equation A5 is of order the product
 702 of the average magnitude of $\nabla \cdot \vec{U}$ and \tilde{E} . Integrated over the column as a whole, the layers
 703 of convergence and divergence nearly balance each other out; the average magnitude of the
 704 divergence exceeds the column average divergence by two orders of magnitude (not shown).
 705 Because $\frac{\partial P_s}{\partial t}$ is equal to the column average divergence (by equation A4), the first term on
 706 the RHS in equation A5 is approximately two orders of magnitude smaller than the second
 707 term, and we thus neglect the first term in this study. The impact of ignoring this term on
 708 the calculations performed in this manuscript is shown in Figure 11 and is small.

709 In our calculations, we also ignore kinetic energy which is two orders of magnitude smaller
710 than the moist static energy. We assess the magnitude of the kinetic energy's contribution
711 to the seasonal amplitude of energy fluxes in the following manner: (1) the climatological
712 energy flux divergence and total column energy tendency of Trenberth and Stepaniak (2003)
713 are used as a starting point, (2) the kinetic energy contribution is subtracted from the total
714 energy flux divergence and tendency (3) the seasonal amplitude of these fields is calculated
715 as the annual harmonic of the zonal mean energy flux in phase with the insolation. The
716 seasonal amplitude of these fields including and excluding the kinetic energy is shown in
717 Figure 12. We emphasize that the inclusion of kinetic energy makes a small contribution to
718 the calculations presented in this manuscript. Excluding kinetic energy from the equation,
719 the atmospheric energy budget equation becomes

$$\frac{1}{g} \int_0^{P_s} \frac{\partial (c_P T + LQ)}{\partial t} dP = SWABS + SHF - OLR - \frac{1}{g} \int_0^{P_s} \left(\vec{U} \cdot \nabla E + \tilde{E} \nabla \cdot \vec{U} \right) dP,$$

720 (A6)

721 which is equation 1 of this paper.

722 We emphasize that the differences between the flux form methodology of calculating
723 energy fluxes used by Fasullo and Trenberth (2008a) and Trenberth and Stepaniak (2003)
724 and the advective form of the energy fluxes used here (compare Figures 12 and 5) are
725 very small and do not affect the conclusions found in this manuscript. The advantages of
726 this form of the equation over the more commonly used flux form (Trenberth and Stepaniak
727 2003) are that (A) the decomposition of the heat flux divergence into advective and divergent
728 components lends insight into the processes that contribute to the accumulation of energy

729 in the column, (B) the heat flux calculations can be done without explicitly balancing the
730 mass budget with a barotropic wind correction and (C) the energy budget is invariant to the
731 zero point energy.

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738

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817 **List of Tables**

818 1 Models used in this study and their resolution. The horizontal resolution refers
819 to the latitudinal and longitudinal grid-spacing or the spectral truncation.
820 The vertical resolution is the number of vertical levels. The last column
821 indicates if the model is included in the analysis of the 2XCO₂ runs in Section
822 5. 43

Abbreviation	Full Name	Horizontal Resolution	Vertical Resolution	2XCO ₂ run
BCCR-BCM2.0	Bjerknes Centre for Climate Research, University of Bergen, Norway	T63	L31	Yes
CCCMA-CGCM3.1	Canadian Centre for Climate Modeling and Analysis, Canada	T47	L31	Yes
CNRM-CM3	Meteo-France/Centre National de Recherches Meteorologique, France	T63	L45	Yes
CSIRO-MK3.0	Australian Commonwealth Scientific and Research Organization (CSIRO), Australia	T63	L18	Yes
GFDL-CM2.0	NOAA/Geophysical Fluid Dynamics Laboratory, USA	2.0° X 2.5°	L24	No
GFDL-CM2.1	NOAA/Geophysical Fluid Dynamics Laboratory, USA	2.0° X 2.5°	L24	Yes
IAP-FGOALS	National Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics (LASG), China	T42	L26	No
MPI-ECHAM5	Max Planck Institute for Meteorology, Germany	T63	L31	No
INM-CM3.0	Institute for Numerical Mathematics, Russia	4° X 5°	L21	Yes
IPSL-CM4.0	Institute Pierre Simon Laplace, France	2.5° X 3.75°	L19	Yes
Micro3.2 (Medres)	National Institute for Environmental Studies, and Frontier Research Center for Global Change, Japan	T42	L20	No
Micro3.2 (Hires)	National Institute for Environmental Studies, and Frontier Research Center for Global Change, Japan	T106	L56	No
MRI-CGCM2.3.2a	Meteorological Research Institute, Japan	T42	L30	Yes
NCAR-CCSM3.0	National Center for Atmospheric Research, USA	T85	L26	Yes
UKMO-HADCM3	Hadley Centre for Climate Prediction and Research/Met Office, UK	2.5° X 3.8°	L19	Yes
MIUB-ECHOG	University of Bonn, Germany	T30	L19	Yes

TABLE 1. Models used in this study and their resolution. The horizontal resolution refers to the latitudinal and longitudinal grid-spacing or the spectral truncation. The vertical resolution is the number of vertical levels. The last column indicates if the model is included in the analysis of the 2XCO₂ runs in Section 5.

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910 kinetic energy; dashed lines do not include kinetic energy. All calculations
911 were done from Fasullo and Trenberth's (2008b) study.

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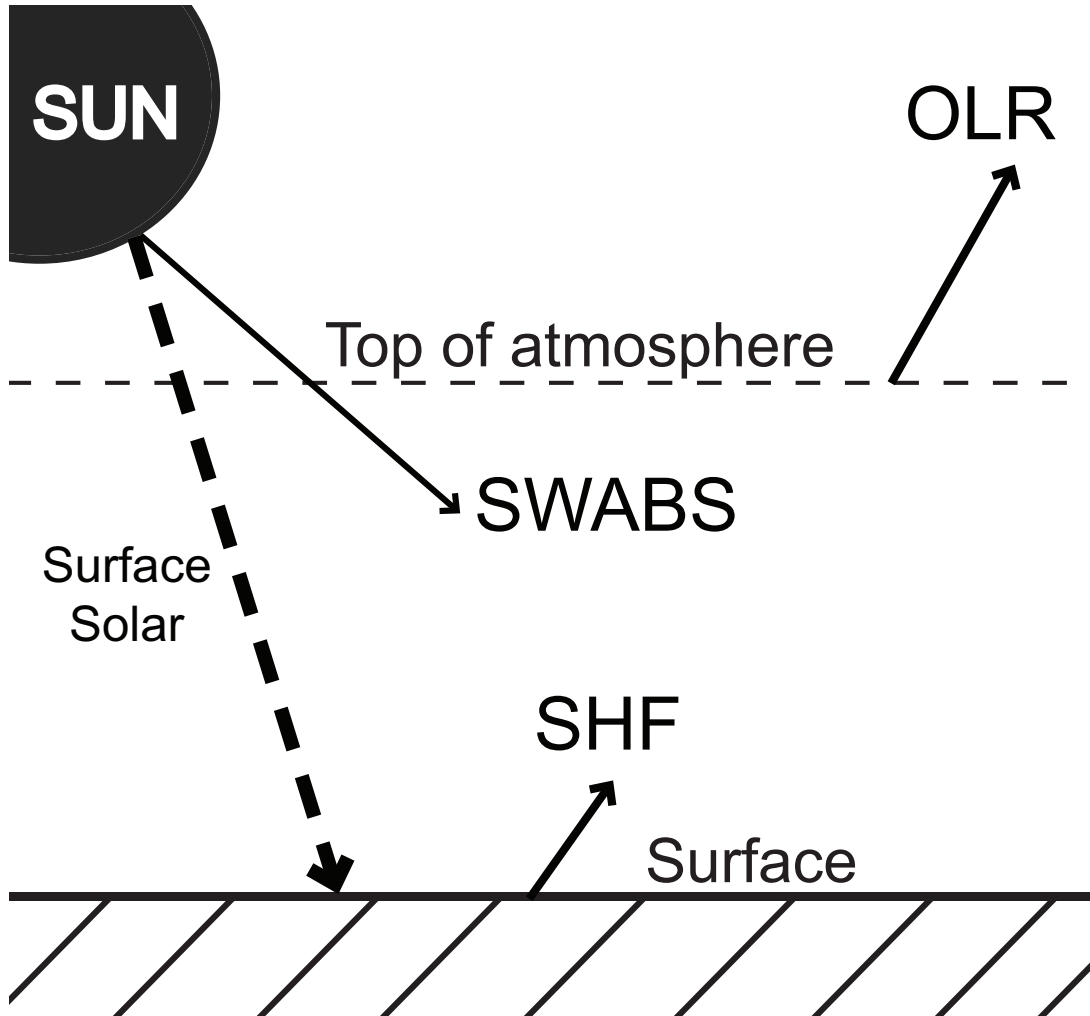


FIG. 1. Schematic of the energy exchanges between the sun, the atmosphere, and the surface. *SWABS* is the solar insolation absorbed within the atmosphere. *SHF* is the net upward energy flux from the surface to the atmosphere. *OLR* is the outgoing longwave radiation at the top of the atmosphere. The surface solar flux (dashed line) is the solar flux to the surface and does not enter the atmospheric energy budget because this radiation passes through the atmosphere.

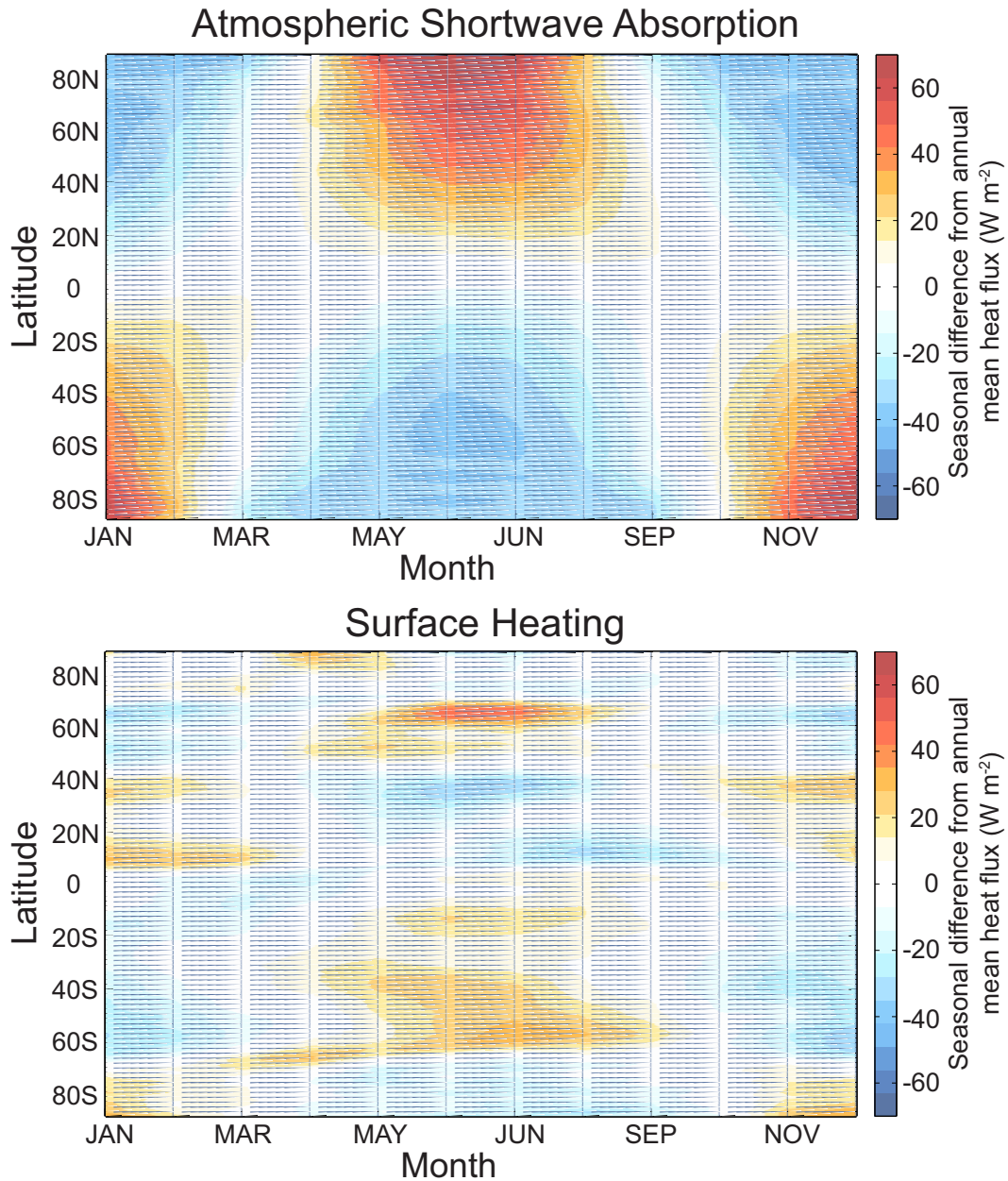


FIG. 2. Observed zonal mean seasonal cycle of atmospheric heating by atmospheric solar absorption (*SWABS* – top panel) and by upward surface heat fluxes (*SHF* – bottom panel) in W m^{-2} . The annual average at each latitude has been removed. The atmospheric solar absorption is calculated from the CERES data at the TOA and surface and the surface heating is calculated from the residual of the terms in Equation 1 as discussed in the text.

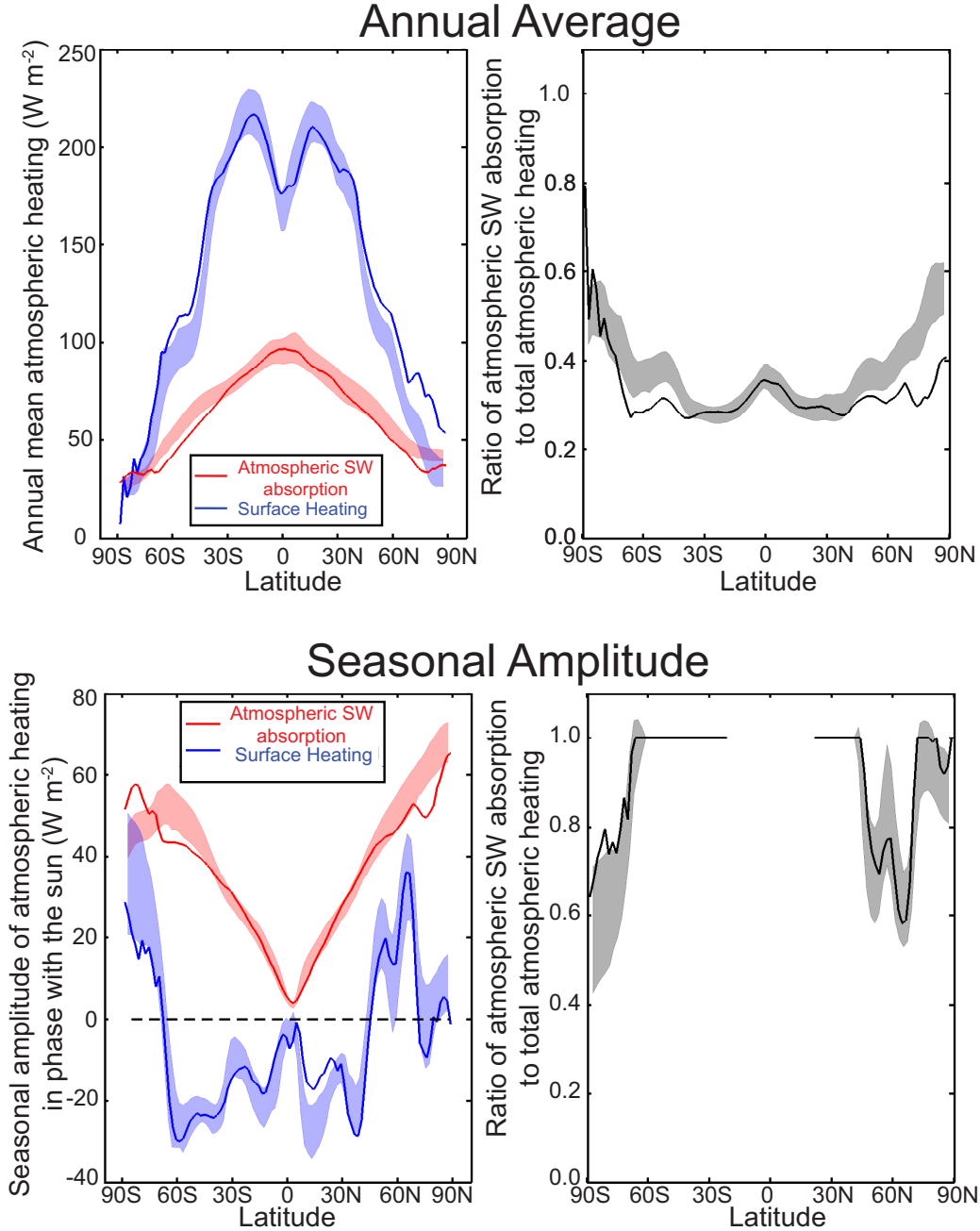


FIG. 3. Zonal mean heating of the atmosphere in the annual average (top panel) and in the seasonal cycle (bottom panel). The heating is divided into atmospheric shortwave absorption (SWABS, red) and upward surface fluxes (blue). The right panels show the fractional contribution of SWABS to the total heating (defined as $SWABS / (SWABS + H(SHF))$) where H is the Heaviside function and the tropics are excluded from the seasonal calculation. The seasonal amplitude is defined throughout as the amplitude of the Fourier harmonic in phase with the sun. In each figure, the solid line is the observations and the shading is $\pm 1\sigma$ about the ensemble mean pre-industrial simulations from the CMIP3 models.

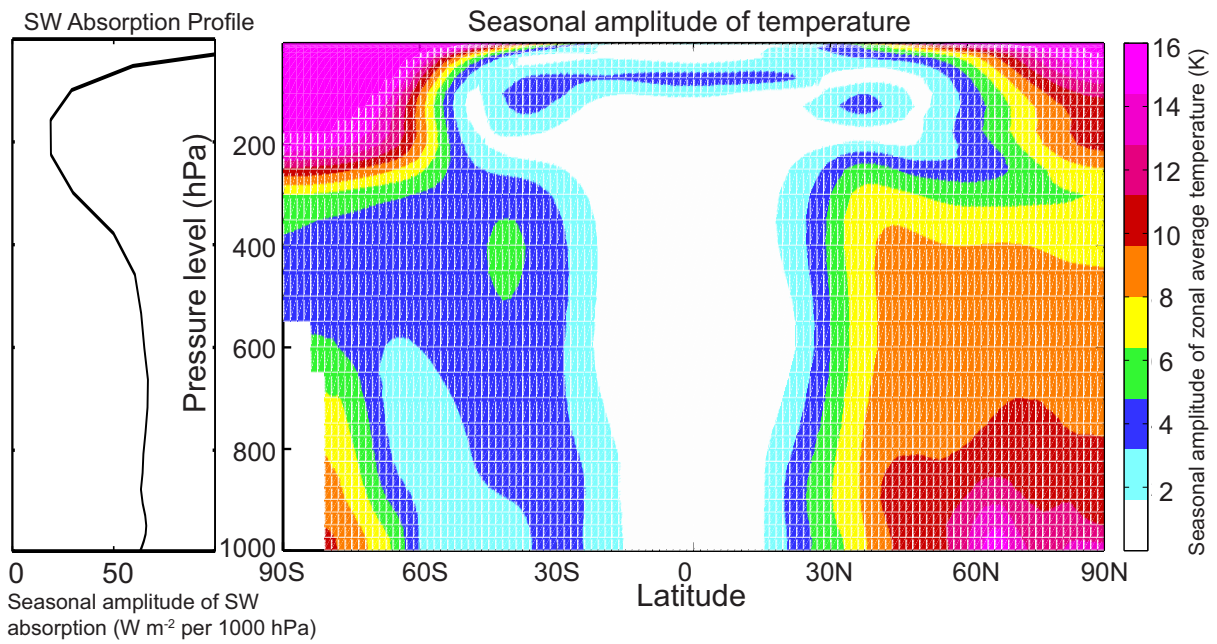


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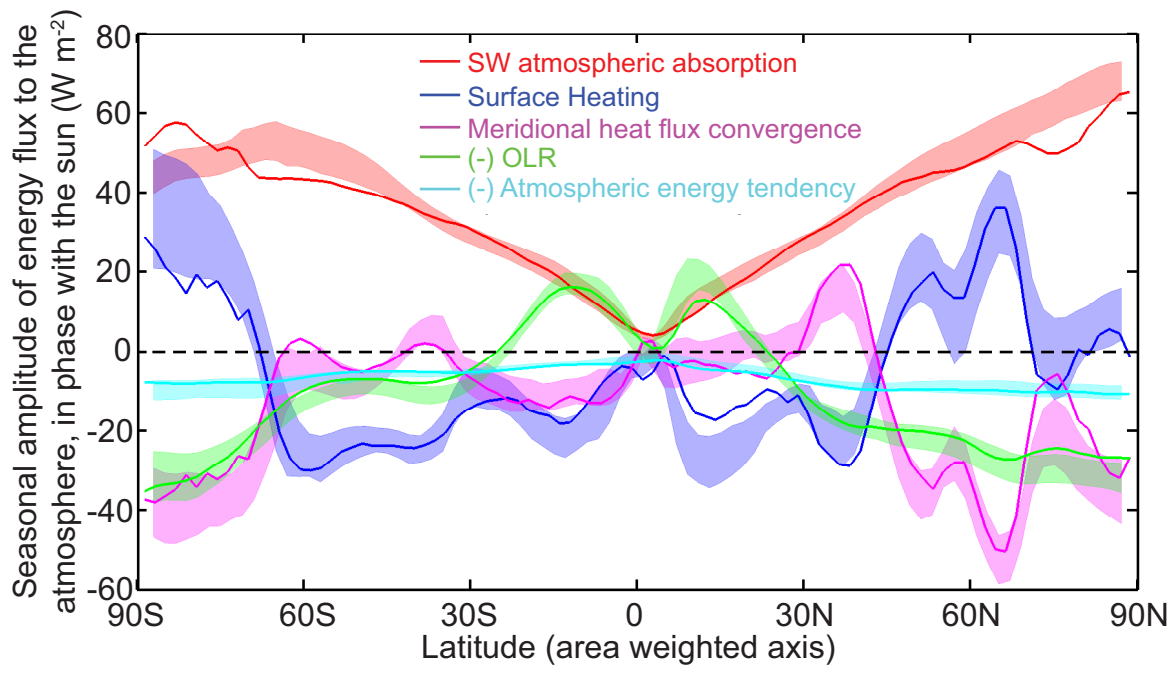


FIG. 5. The seasonal amplitude of atmospheric energy fluxes in phase with the sun (positive fluxes amplify the seasonal cycle, negative fluxes reduce the seasonal cycle). Solid lines are observations and shaded regions represent $\pm 1\sigma$ about the ensemble mean pre-industrial simulations from the CMIP3 models.

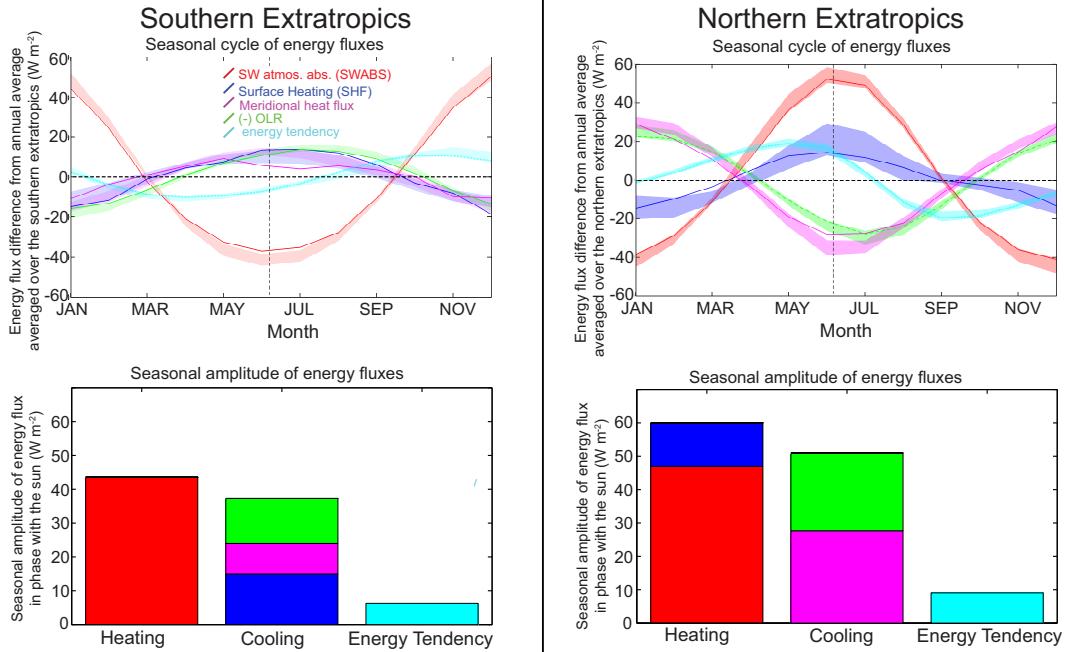


FIG. 6. (top panel) The seasonal cycle of atmospheric energy fluxes (in W m^{-2}) averaged over the extratropics – defined as poleward of 42° – in the southern hemisphere (left panel) and the northern hemisphere (right panel). The observations are shown by the solid lines and the shaded region represents $\pm 1\sigma$ about the CMIP3 PI ensemble average. The dashed vertical lines represent the winter solstice in the SH plot and summer solstice in the NH plot. The annual average of each term has been removed. (bottom panel) The seasonal amplitude of the atmospheric energy fluxes in phase with the seasonal cycle of solar insolation averaged over the extratropics (the left panel is the southern extratropics and the right panel is the northern extratropics). The terms that amplify the seasonal cycle in temperature (heating) are in the first column. The seasonal energy loss terms (cooling) are in the second column. The third column is the energy stored in the atmospheric column (energy tendency). The individual terms are color coded in the legend in the upper left panel and explained in the text.

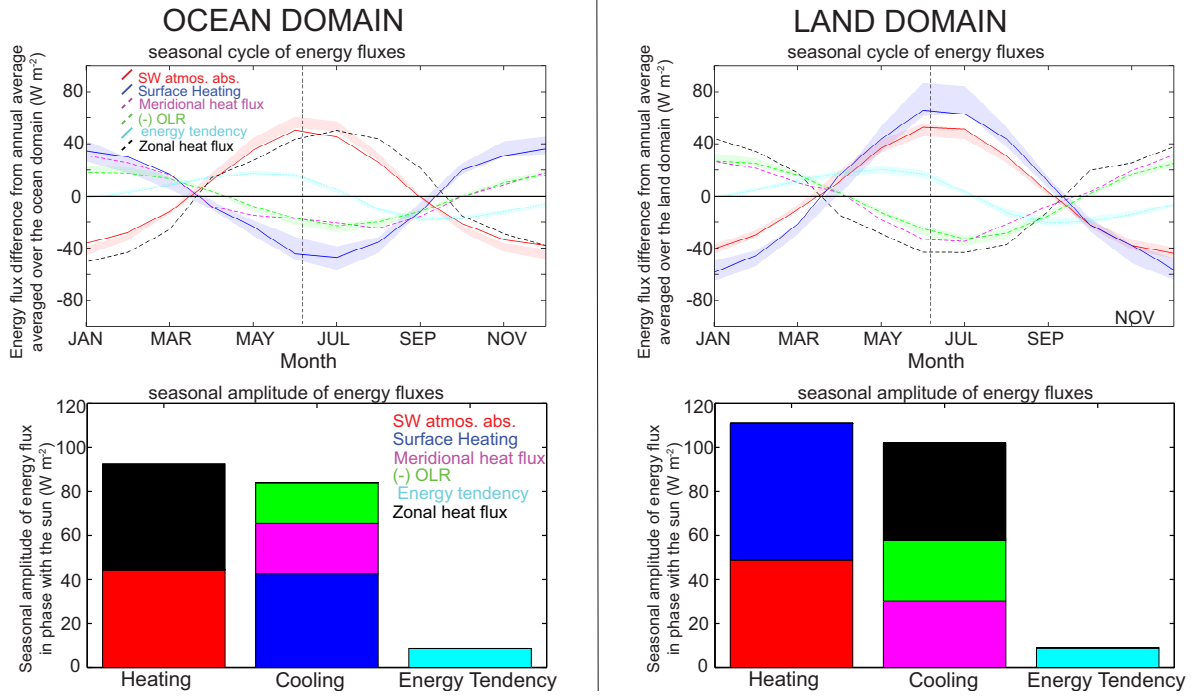


FIG. 7. (top panels) The seasonal cycle of energy fluxes averaged over the atmosphere in the NH extratropical ocean domain (left panel) and land domain (right panel). Observations are given by solid lines and the shading represents 1 of the CMIP3 pre-industrial ensemble. The atmospheric heat fluxes are decomposed into zonal and meridional components in the observations. The vertical dashed line represents the summer solstice. (bottom panels) The seasonal amplitude of energy fluxes (in phase with the sun) averaged over the ocean/land domains. The amplifying fluxes are on the left and the damping (i.e, out of phase fluxes) are in the middle (colors are described in the legend).

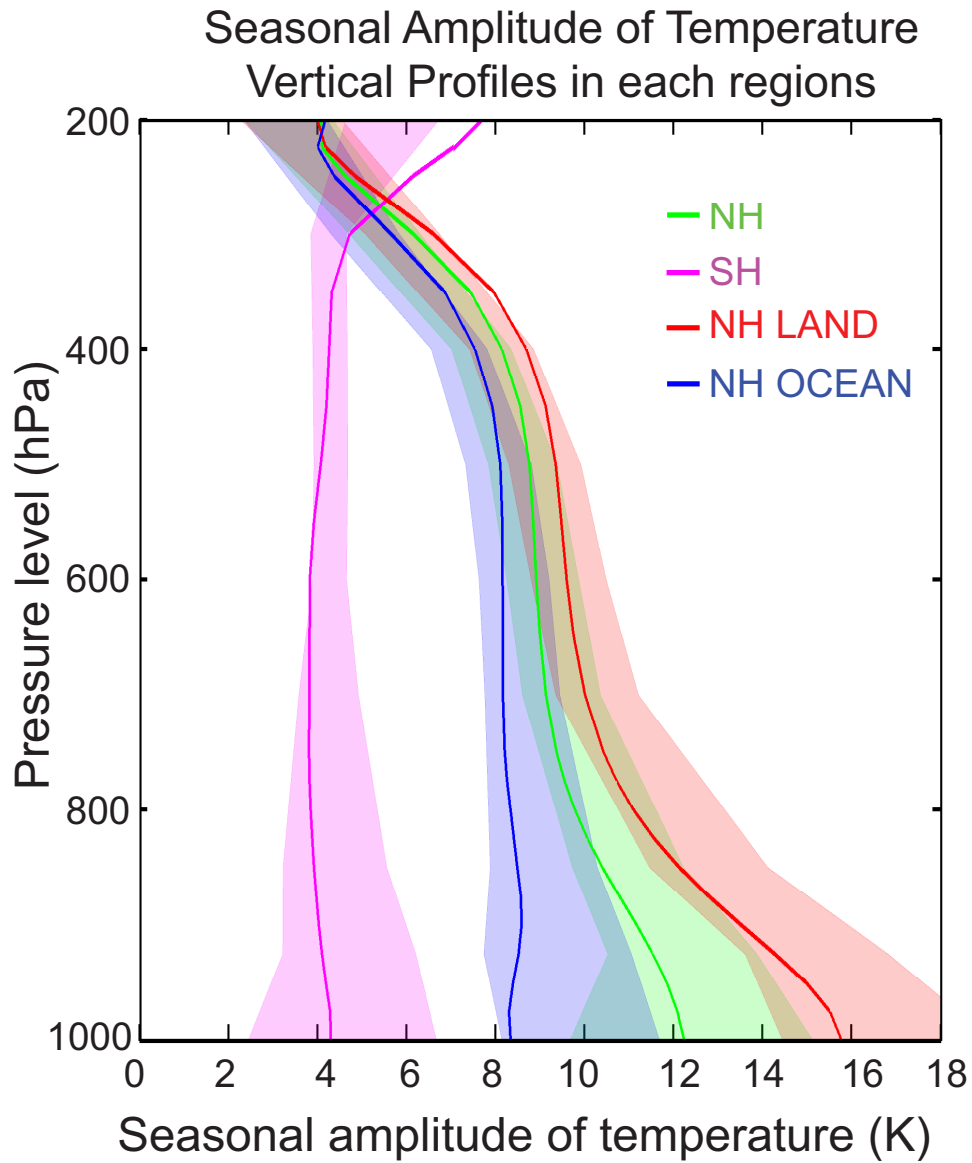


FIG. 8. The seasonal amplitude of temperature averaged over the extratropics (poleward of 42°) in each hemisphere (in K). The northern extratropics is further decomposed into ocean and land domains. The observations are given by the solid line and the shading represents $\pm 1\sigma$ about the ensemble mean pre-industrial simulations from the CMIP3 models.

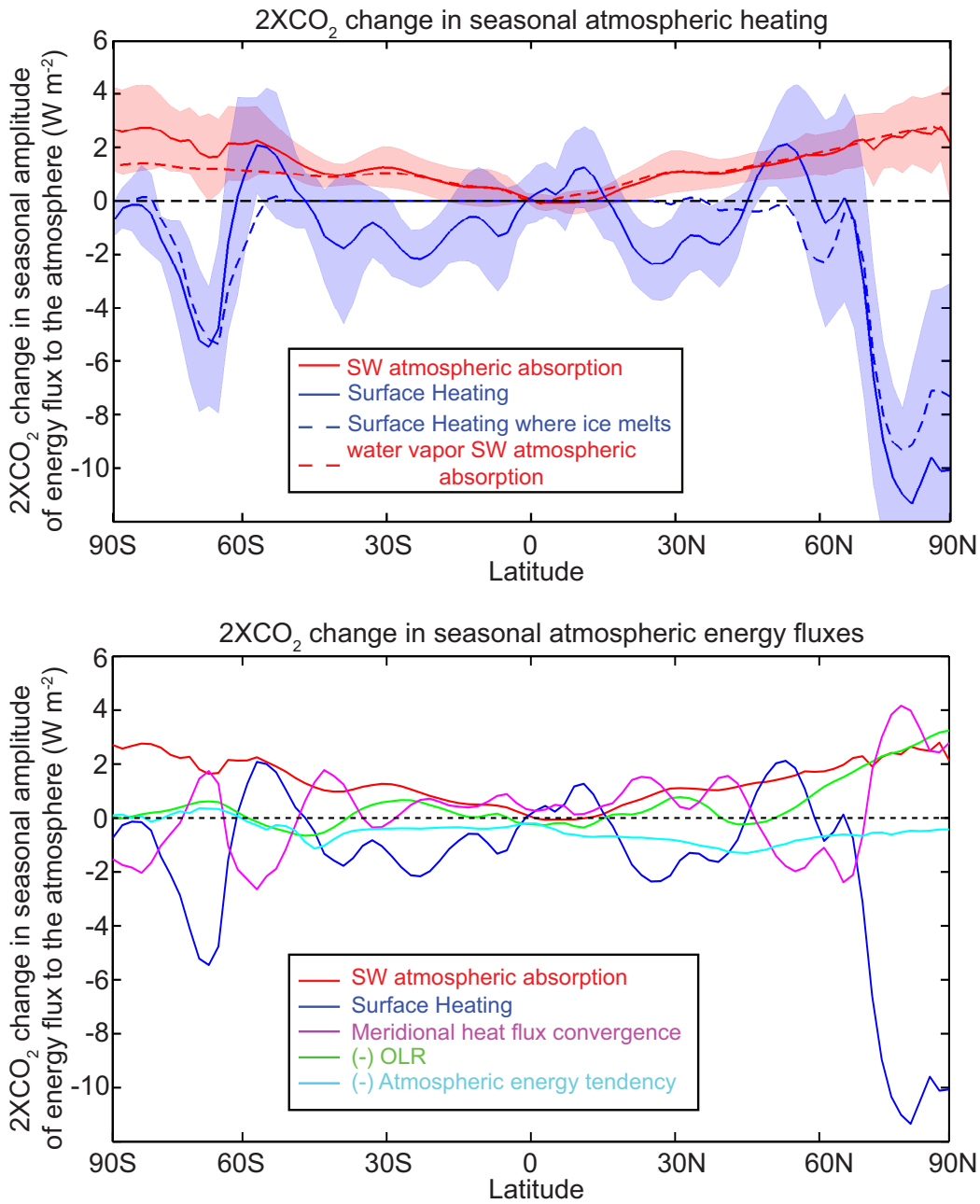


FIG. 9. (Top Panel) The change in the seasonal amplitude of atmospheric heating in the CMIP3 CO₂ doubling experiment. The solid red line is the ensemble average change in *SWABS*, the shaded red area is $\pm 1\sigma$ of the model response and, the dotted red line is the change due to water vapor changes as diagnosed from the water vapor shortwave kernel. The blue line is the change in the seasonal amplitude of *SHF* (with shading $\pm 1\sigma$) and the dotted blue line is the change within the sub portion of the latitude band where the sea ice fraction decreases by more than 10% relative to the PI simulation. (Bottom Panel) The ensemble average change in the seasonal amplitude of all terms in the atmospheric energy budget.

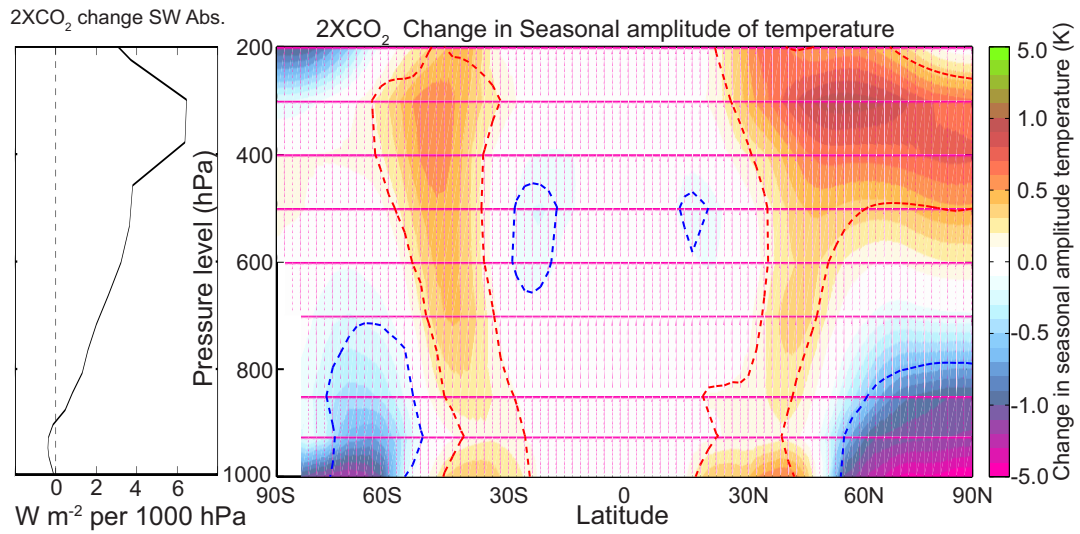


FIG. 10. (Right Panel) Zonal and ensemble average change in the seasonal amplitude of temperature in the CMIP3 CO₂ doubling experiments. The contours show the regions of significant change as assessed by a one sample t-test at the 99% confidence interval. (Left Panel) The vertical profile of the change in the seasonal amplitude of shortwave radiative heating in the GFDL 2.1 CO₂ doubling experiment expressed as the change in column integrated *SWABS* in W m⁻² that would result if that heating rate were vertically invariant over the entire column.

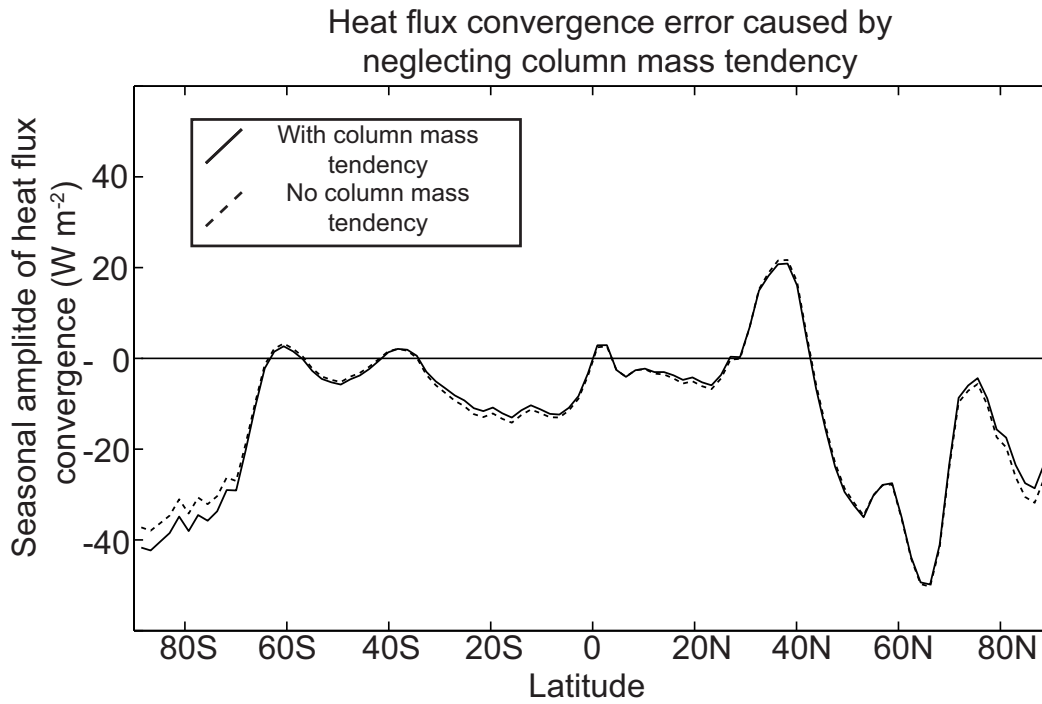


FIG. 11. Effect of neglecting the mass column tendency on the energy flux calculations presented in this manuscript. The solid line is the seasonal amplitude of the zonal average energy flux convergence using all terms and the dashed line is the same calculation neglecting the column mass tendency (first term in Equation A5).

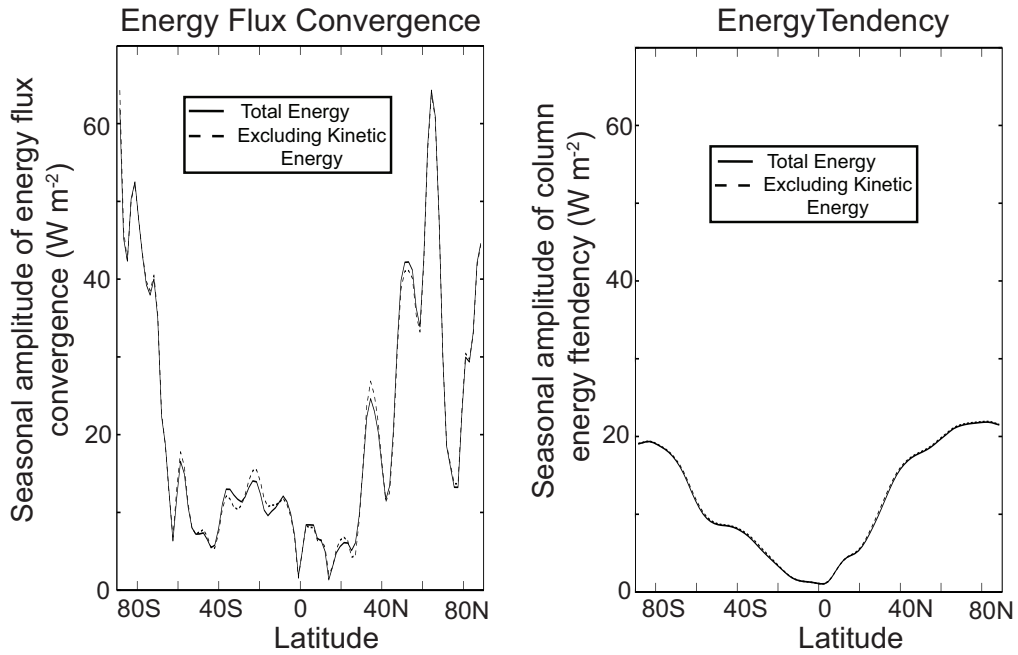


FIG. 12. Effect of neglecting the kinetic energy on the energy fluxes in equation A1. The seasonal amplitude (the amplitude of the annual Fourier harmonic in phase with the insolation) of the horizontal energy flux convergence (left panel) and the energy tendency are shown. Solid lines represent calculations that include kinetic energy; dashed lines do not include kinetic energy. All calculations were done from Fasullo and Trenberth's (2008b) study.