

1 **Seasonal asymmetries in the lag between insolation and surface temperature**

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ABSTRACT

15 We analyze the temporal structure of the climatological seasonal cycle in
16 surface air temperature across the globe. We find that, over large regions of
17 the Earth, the seasonal cycle of surface temperature departs from an annual
18 harmonic: the duration of fall and spring differ by as much as two months. We
19 characterize this asymmetry by the metric ASYM, defined as the phase lag of
20 the seasonal maximum temperature relative to the summer solstice minus the
21 phase lag of the seasonal minimum temperature relative to winter solstice. We
22 present a global analysis of ASYM from weather station data and atmospheric
23 reanalysis and find that ASYM is well represented in the reanalysis. ASYM
24 generally features positive values over land and negative values over the ocean
25 indicating spring has a longer duration over the land domain whereas fall has
26 a longer duration over the ocean. However, ASYM also features more pos-
27 itive values over North America compared to Europe and negative values in
28 the polar regions over ice sheets and sea ice. Understanding the root cause of
29 the climatological ASYM will potentially further our understanding of con-
30 trols on the seasonal cycle of temperature and its future/past changes. We ex-
31 plore several candidate mechanisms to explain the spatial structure of ASYM
32 including: (i) modification of the seasonal cycle of surface solar radiation by
33 the seasonal evolution of cloud thickness; (ii) differences in the seasonal cycle
34 of the atmospheric boundary layer depth over ocean and over land; and (iii)
35 temperature advection by the seasonally evolving atmospheric circulation.

36 **1. Introduction**

37 The climatological seasonal cycle of surface (2m) daily average air temperature (hereafter T_{2m})
38 in the extratropics is forced by the seasonal cycle of insolation and moderated by atmospheric ad-
39 vection and radiative damping (North and Coakley 1978; Donohoe and Battisti 2013). Throughout
40 the extratropics, seasonal variations in insolation are nearly sinusoidal (Berger 1978). Hence, the
41 seasonal cycle of T_{2m} is well represented by the amplitude and phase of the annual harmonic
42 (Wallace and Osborn 2002; Thomson 1995; McKinnon et al. 2013). The spatial structure of the
43 amplitude and phase of the annual harmonic in T_{2m} provide insights into the underlying climate
44 processes including:

- 45 • the large amplitude of T_{2m} in the high latitudes driven by the large magnitude of seasonal
46 variations in insolation (Von Hann and Ward 1903; Peixoto and Oort 1992);
- 47 • the small amplitude and large phase lag in T_{2m} over the oceans due to the high heat capacity
48 of the ocean mixed layer and the efficient thermal coupling between T_{2m} and sea surface
49 temperature (SST; Prescott and Collins 1948);
- 50 • the large amplitude and small phase lag of T_{2m} over continental regions far downwind of the
51 ocean due to the small heat capacity of the land surface and the atmospheric boundary layer
52 relative to that of the ocean mixed layer (Brooks 1917).

53 In addition, continental regions that are downwind of the ocean represent a mixture of the phase
54 and amplitude of T_{2m} between the above mentioned marine and continental cases (Stine et al.
55 2009).

56 Additional insights into the processes driving the seasonal cycle have been gleaned from the
57 spatial patterns of the change in the phase and amplitude of the annual harmonic of T_{2m} and
58 atmospheric temperature due to global warming that include:

- 59 • reduced amplitude and delayed phase of T_{2m} in regions where sea ice melt exposes the atmo-
60 sphere to the thermal reservoir of the ocean (Dwyer et al. 2012);
- 61 • enhanced amplitude and reduced phase lag of mid-tropospheric temperature (Donohoe et al.
62 2013) due to enhanced shortwave absorption in the atmospheric column associated with at-
63 mospheric moistening (Donohoe and Battisti 2013);
- 64 • phase changes in T_{2m} over land are moderated by the strength of mixing between marine and
65 continental air associated with annular mode changes (Stine and Huybers 2012).

66 Finally, the spatial pattern of observed changes in the annular harmonic of T_{2m} provide a de-
67 tectable fingerprint of anthropogenic forcing (Santer et al. 2018).

68 Are there additional insights to be gleaned from the climatological seasonal cycle in T_{2m} that
69 extend beyond the amplitude and phase of the annual harmonic? For example: is the cooling of
70 the surface air temperature in the fall more rapid than the warming in the spring? Is the answer the
71 same over the ocean and land domains or between the different continents? Annual harmonic anal-
72 ysis is poorly suited to answer these questions because the cold and warm seasons are constructed
73 to be symmetric in amplitude, duration and timing.

74 For illustrative purposes, consider the seasonal cycle of T_{2m} at Seattle-Tacoma international
75 airport (SEATAC; Fig. 1) that inspired this work – due in part to the authors’ familiarity with this
76 region. The minimum temperature of the smoothed seasonal cycle (see methods in Section 2B)
77 occurs on January 4th, 14 days after the winter solstice (the timing of the T_{2m} extrema and solstices
78 are shown by the vertical black and red lines, respectively) whereas the maximum temperature
79 occurs on August 4th, 44 days after the summer solstice. This (30 day) seasonal asymmetry of
80 the temperature lag (hereafter ASYM) relative to insolation suggests that the seasonal warming
81 into the peak of summer takes almost 2 months longer than the seasonal cooling into the peak of

82 winter. This point is visualized by comparing the observed seasonal cycle of T_{2m} to an annual
83 harmonic fit to the data (c.f. the thick black line to the thick dashed blue line in Fig. 1). The rate
84 of cooling (warming) during the autumn (spring) is far greater (less) than that expected from the
85 annual harmonic. The annual harmonic analysis of T_{2m} cannot capture these substantial seasonal
86 timing asymmetries and instead produces a phase of temperature that splits the differences between
87 the phase of the maximum and minimum.

88 Departures of T_{2m} from the annual harmonic that are characterized by ASYM have received little
89 attention in the literature but have the potential to provide new insights into the physics of climate
90 and its future changes. van den Broeke (1998) note that T_{2m} over Antarctica has a significant
91 semi-annual component with a slow cooling of temperatures during the winter and fast warming
92 during the spring. This seasonality is understood to result from the seasonal “shape” of insolation
93 which follows the top half of a sine curve in the summer but flat-lines to zero during winter and
94 thus projects onto the higher order harmonics via Gibbs overshoot. May et al. (1992) produced a
95 global analysis of T_{2m} at 122 stations including the (separate) timing of the seasonal maximum and
96 minimum temperature during the smoothed seasonal cycle (see their Figs. 89 and 91). They noted
97 that T_{2m} departs from the annual harmonic in monsoonal regions where the onset of the monsoons
98 initiates cooling prior to the summer solstice (Hurley and Boos 2015) but commented little on the
99 large scale patterns of ASYM. More recently, SheridanDodds et al. (2016) analyzed the timing of
100 minimum and maximum temperatures at weather stations in the United states as well as trends in
101 the seasonal timing over the duration of the observational record.

102 The purpose of this manuscript is threefold: (i) to present a comprehensive global analysis of the
103 climatological ASYM from historic station observations; (ii) establish that the ASYM in observa-
104 tions is captured in the ERA-Interim reanalysis product and use the latter product to document the
105 global three dimensional structure of ASYM (including the seasonal cycle of temperature in the

106 troposphere above the boundary layer); and (iii) explore possible physical mechanisms that govern
107 the global structure of ASYM.

108 **2. Data and methods**

109 *a. Data sets*

110 **Station Data:** Historical station data of near surface (2 meter) air temperature from the Global
111 Historical Climatology Network (GHCN Peterson and Vose 1997) is the primary source of data
112 used in our analysis. This data set has daily maximum (T_{MAX}) and minimum temperature (T_{MIN})
113 for nearly 100,000 stations with some stations reporting nearly continuous data for 175 years and
114 others containing less than 1 year of data. We thin the number of stations used in our analysis
115 using the following criteria: (i) for stations within 0.1 degrees (latitude and longitude) of each
116 other, we remove all data except the longest continuous record – this eliminates locations that have
117 registered secondary or tertiary data; (ii) stations with less than 20 years of data are eliminated; (iii)
118 stations that have less than 5 years of data on any calendar day are eliminated; and (iv) stations
119 equatorward of 25° latitude for which the seasonal cycle is small in magnitude and more semi-
120 annual in nature are eliminated. These criteria reduce the number of stations to 3,096 stations
121 dispersed globally across land areas. We analyze the seasonal cycle of T_{2m} which is defined as
122 the average of T_{MAX} and T_{MIN} . While this method of averaging is biased relative to true time-
123 averages, it is recommended for the sake of consistency when comparing historical datasets such
124 as station observations that often only provide daily extrema (World Meteorological Organization
125 2011). All results presented here are nearly unchanged if we analyze the seasonal timing of T_{MAX}
126 or T_{MIN} in isolation.

127 ASYM is also calculated from in situ T_{2m} (and sea surface temperature – SST) data at marine
128 stations. Marine observations come from Ocean Weather ships that were deployed in the World
129 War II era by the UK (Downes 1977) and United States of America (US Weather Bureau 1977)
130 at fixed locations primarily in the North Atlantic but secondarily in the North Pacific. T_{2m} and
131 SST records of sufficient length to define a smoothed climatological seasonal cycle were retrieved
132 from Stations Alpha, Bravo, Charlie, Delta, Echo, India, Juliet, Kilo and, Mike in the Atlantic, and
133 Stations Extra, Hotel, Nan, Papa (Belka et al. 2014) and Victor in the Pacific.

134 Additionally, in situ data from Washington state is used to analyze the seasonal timing of tem-
135 perature and surface solar radiation. ASYM is calculated from near surface air temperature data
136 from Mount Rainier (D’Amico 2019), Washington, USA (Paradise Ranger Station, altitude 1,646
137 meters and Camp Muir Climbing camp, altitude 3,105 meters). Finally, we also examine the sea-
138 sonal cycle of total, direct and diffuse solar radiation measured on the roof of the Department of
139 Atmospheric Sciences at the University of Washington in Seattle, approximately 15 miles from
140 SEATAC airport (there is no solar data at SEATAC); these data have been measured at 1 minute
141 intervals since 2000.

142 **Gridded Temperature Data:** While our primary focus here is on ASYM derived from histor-
143 ical observations, we demonstrate that the regional structure of ASYM seen in the station data is
144 well represented in the reanalysis data which allows for a spatially continuous global analysis of
145 ASYM. We consider two sets of daily reanalysis temperature data: the ERA Interim (Dee et al.
146 2011) 2 meter air temperature at a resolution of 1.5° latitude (or roughly 150 km), and the NCEP
147 reanalysis 2 meter air temperature at a resolution of 2° latitude (or roughly 200 km). Results using
148 the higher resolution ERA data are shown in the manuscript figures; results using NCEP are shown
149 in the appendix (Fig. A2). The ERA reanalysis product we use spans the time period 1979-2015

150 and the NCEP data spans 1970-2012. The primary conclusions are insensitive to the choice of
151 reanalysis product.

152 In some instances, geographic features that are smaller than the ERA-Interim gridbox can intro-
153 duce differences between the seasonal cycle of T_{2m} (and potentially ASYM) measured at stations
154 and its representation at the nearest ERA-Interim gridbox. For example, shown in the lower panel
155 of Fig. 1 is the T_{2m} from the Seattle station data (SEATAC) and the temperature from the ERA
156 reanalysis gridbox that includes Seattle. The temporal “shape” of the seasonal evolution of T_{2m}
157 and the timing of the maximum and minimum are nearly identical in the ERA and station data (c.f.
158 the vertical blue and black lines that are nearly on top of each other). However, in absolute terms,
159 the T_{2m} at the nearest neighbor ERA gridbox is approximately 2.5°F colder than that measured
160 at SEATAC throughout the year. This absolute mismatch should be expected from the approxi-
161 mately 150 km horizontal resolution of the ERA data because the gridbox that includes Seattle
162 also extends across the Cascade mountains into Central Washington. Thus, the ERA gridbox that
163 includes Seattle also includes the town of Ellensburg, Washington located east of the Cascade
164 Mountains. The February average minimum temperature is 19°F (-7°C) in Ellensburg and 35°F
165 (2°C) in Seattle. Thus, we should not expect absolute agreement between the reanalysis and sta-
166 tion data – especially in the presence of sharp sub-grid scale topographic features. However, we
167 show below that, in general, the spatial structure of ASYM primarily varies at the continental scale
168 and, thus, the reanalysis data is well suited for assessing its spatial variability. Given the strong
169 correspondence between the large scale patterns of ASYM in the surface station temperature data
170 and surface reanalyses products, we can use the reanalysis data to probe the vertical structure of
171 ASYM.

172 **Solar Radiation data:** The seasonal phasing of solar radiation at the top of atmosphere and
173 at the surface is analyzed using satellite derived global gridded solar radiation data . We use data

174 from the Clouds and the Earth’s Radiant energy system experiment (CERES) SYN version 4 (Kato
175 and Coauthors 2018; Loeb and Coauthors 2018). The CERES data is derived from radiometer
176 measurements on the Aqua and Terra satellites that are assimilated using a radiative transfer model,
177 resulting in a self consistent surface radiative flux measurements. This data set has daily temporal
178 resolution over the time period 2000-2018 and provides global gridded coverage at 1° spatial
179 resolution.

180 **Sea Surface Temperature Data:** While our focus is on ASYM calculated from near surface air
181 temperature (T_{2m}), we supplement our analysis by calculating ASYM from daily SST data. SST
182 data is from NOAA High-resolution Blended Analysis Version 2 on a 0.25° global grid over the
183 time period 1981-2016 (Reynolds et al. 2007).

184 **Energy fluxes in the atmospheric and oceanic column:** The seasonal input of energy into the
185 atmospheric and oceanic columns is calculated using the methodology described in Donohoe and
186 Battisti (2013). These calculations use radiative fluxes at the top of atmosphere and surface from
187 the Clouds and Earth’s Radiant Energy System experiment 4.0 (CERES Loeb and Coauthors 2018;
188 Kato and Coauthors 2018), atmospheric energy flux convergence and column energy tendencies
189 calculated from ERA reanalysis (Dee et al. 2011). The turbulent energy exchange between the
190 surface and the atmosphere as a residual of the atmospheric energy budget.

191 *b. Methodology for defining the seasonal asymmetry in the lag between T_{2m} and insolation*

192 The methodology for calculating the seasonal asymmetry of the temperature time lag (ASYM)
193 from both station based T_{2m} data and reanalysis data is described here. Smoothed climatological
194 seasonal cycles of T_{2m} are created by first plotting all years of daily data on a domain of days
195 past January first of that year (black dots in Fig. 1) padded periodically on both sides. A spline
196 with 6 evenly spaced knots within the year is fit to the data (thick black line). Then the *timing*

197 *of the temperature maximum* (τ_{MAX}) is defined as the number of days past the summer solstice
198 on which the smoothed T_{2m} reaches its maximum; similarly, τ_{MIN} is defined as the number of
199 days past the winter solstice on which the smoothed T_{2m} reaches its minimum. Hence, positive
200 values of τ_{MAX} and τ_{MIN} correspond to the T_{2m} extremes that lag the solstices, and negative values
201 correspond to T_{2m} extremes that precede the solstices. Finally, ASYM is defined as τ_{MAX} minus
202 τ_{MIN} . Positive values of ASYM correspond to a maximum T_{2m} that lags summer solstice more
203 than minimum T_{2m} lags winter solstice. Hence, a positive ASYM indicates a prolonged spring and
204 hurried autumn (i.e. in the same sense of the ASYM seen at SEATAC in Fig. 1); negative ASYM
205 indicates a hurried spring and prolonged autumn. ASYM is calculated using the spline fit to the
206 data because, even with 70 years of data (e.g. at the SEATAC station shown in the top of Fig. 1),
207 the climatological average for a given calendar day (thin black line) is fairly jumpy from day to
208 day due to random sampling of synoptic variability and this would cause the timing of the extrema
209 to have some random component about the longer time scale seasonal evolution of temperature.
210 We chose a spline with 6 knots because this choice removes the high-frequency structure of the
211 synoptic variability while retaining the temporal structure of the seasonal evolution. Alternative
212 choices for the number of knots between 3 to 12 knots do not significantly change the results
213 presented here. *At some locations, the smoothed seasonal cycle of T_{2m} has more than one local*
214 *temporal maxima and minima. This occurs at 8 of the 3,096 (0.3%) stations and <2% of the area*
215 *of extratropics in the ERA data – denoted by gray crosses over the stations and gray stippling over*
216 *the ERA data in Fig. 2. In these cases, τ_{MAX} and τ_{MIN} are defined by the timing of the absolute*
217 *extrema in the smoothed T_{2m} . In these regions, τ_{MAX} and τ_{MIN} are potentially sensitive to small*
218 *differences (spatially or over different time periods used to define the climatology) in the temporal*
219 *structure of T_{2m} that change which local extrema is the absolute extrema used to define τ_{MAX} and*
220 *τ_{MIN} .*

221 The seasonal asymmetry in downward surface solar radiation is calculated using the same
222 methodology used to calculate the seasonal asymmetry of the temperature lag.

223 3. Global structure of ASYM

224 In this section, we present global maps of ASYM calculated from the comprehensive set of
225 surface station data and from ERA reanalysis. We also present the vertical structure of ASYM
226 derived from reanalysis and highlight some curious features.

227 *a. Correspondence between τ_{MAX} , τ_{MIN} and, ASYM calculated from station and reanalysis data*

228 Maps of τ_{MAX} , τ_{MIN} and ASYM calculated using the station data and calculated using the ERA
229 reanalysis data are shown in Fig. 2. Dots represent the station values and the underlying colors
230 represent the ERA values. For each of the three quantities, values calculated from the station data
231 are in good agreement with those calculated using the ERA reanalysis data (particularly over land
232 where it is difficult to discern the two). We quantify the correspondence between the station based
233 metrics and those calculated in the ERA reanalysis using the mean absolute difference between the
234 station calculation and that in the nearest-neighbor ERA gridpoint. The mean absolute difference
235 is 3.4 days in τ_{MAX} , 3.4 days in τ_{MIN} and 3.9 days in ASYM. The spatial correlation between
236 station based and ERA based metrics is 0.88 for τ_{MAX} , 0.85 for τ_{MIN} and, 0.89 for ASYM. A
237 scatter plot of ASYM calculated at each station versus that calculated from the nearest neighbor
238 ERA reanalysis gridpoint (Fig. A1) shows strong correlation with a slope near 1, and a small
239 ERA offset in ASYM (ASYM in the station data is 0.2 days greater than that in ERA on average).
240 Similar results (with slightly lower correlations and mean absolute differences) are found when the
241 ERA reanalysis is replaced with T_{2m} from the NCEP reanalysis product (Appendix Fig. A2). There
242 are large differences between ASYM calculated from the 14 ocean weather ship stations (OWS)

243 data and that found in the co-located ERA data: ERA based ASYM values are predominantly
244 negative over the ocean – with the exception of the western side of the ocean basins – whereas the
245 OWS data features values of ASYM near zero or positive, especially in the North Atlantic. We
246 return to this discrepancy in [section 4B](#).

247 Given the good correspondence between ASYM identified in the surface reanalysis data and that
248 from the T_{2m} station observations at the regional scale, we will extend the analysis of ASYM into
249 the troposphere using the reanalysis data to help illuminate the relevant mechanisms that affect
250 ASYM. We note that, with the exception of several high altitude stations, we have very few direct
251 observational constraints on ASYM calculated above the surface and must rely on the reanalysis
252 products for this calculation.

253 *b. Surface timing of seasonal cycle*

254 In general, τ_{MAX} and τ_{MIN} both show short lags over continental regions and long lags over
255 marine regions (Fig. 2) – differences that are expected due to the large thermal inertia of the ocean
256 versus the small thermal inertia of land and ice (Stine et al. 2009; Hasselmann 1976).

257 Westerly winds advect ocean-influenced air onto the western side of the continents, resulting
258 in larger values of τ_{MAX} and τ_{MIN} on the western side of the continents than on the eastern side.
259 Characteristic values of τ_{MAX} and τ_{MIN} are around 50 days in the western continents and around
260 20 days in the interior and along the eastern edge of the continents (Fig. 2). Westerly winds advect
261 continental-influenced air over the western ocean basins, resulting in τ_{MAX} and τ_{MIN} values that
262 are around 40 days, shorter than those over the central and eastern ocean basin (≈ 80 days; Fig.
263 2).

264 In addition to the large-scale similarities in the spatial maps τ_{MAX} and τ_{MIN} associated with the
265 contrasting heat capacity of land and ocean, there are substantial regional scale differences in τ_{MIN}

266 and τ_{MAX} . Figure 2 features the expected gradient from large lags on the western side of continents
267 to smaller lags in the continental interior and east side of the continents that is especially evident
268 in τ_{MAX} over North America and Eurasia and in τ_{MIN} over northern Europe. In contrast over North
269 America τ_{MIN} *increases* gradually eastward with values around 10 days along the west coast of the
270 US and values of around 30 days on the east coast and even larger values around the Great Lakes
271 (Fig. 2b). Hence, the west-east gradient of τ_{MIN} is opposite of that seen in τ_{MAX} and opposite of
272 what is expected based on the advection of maritime air into the continents by the climatological
273 westerly winds. A similar feature is seen over the Southwest Europe with small values of τ_{MIN} (\approx
274 10 days) just downwind of the Atlantic and τ_{MIN} increasing (to ≈ 40 days) in eastern Europe.

275 There are pronounced differences between τ_{MAX} and τ_{MIN} in the high latitudes of both hemi-
276 spheres. τ_{MAX} has small values over ice covered regions (≈ 10 -20 days), especially over the
277 Greenland and Antarctic ice sheets while τ_{MIN} has large values (≈ 60 days). The small values of
278 τ_{MAX} are likely due to the poor conduction of heat through ice which effectively isolates the atmo-
279 sphere from the underlying surface and, additionally, to the fact that temperatures in the marginal
280 ice zone are pegged near the freezing point in the summer. The larger values of τ_{MIN} result from
281 gradual cooling during the polar night until the sun rises again in the spring. Over northern India,
282 τ_{MAX} has values of less than -20 days indicating that the maximum in T_{2m} leads the summer sol-
283 stice by more than 20 days whereas τ_{MIN} in the same region is of order +20 days as one might
284 expect from a continental interior. We argue below that the ASYM in this region is a consequence
285 of the onset of the monsoon which results in thick cloud cover that halts the seasonal warming.

286 The seasonal asymmetries in the lag between T_{2m} and insolation are best highlighted by the
287 metric ASYM (Fig. 2c). In both the station data and reanalysis data, ASYM typically shows:

- 288 ● negative values over the ocean domain, indicating the winter minimum in T_{2m} is delayed more
289 than the summer maximum (exceptions include the northwestern Pacific, North Atlantic south
290 of Greenland, Gulf of Mexico, the west coast of Africa and the New Zealand sector which is
291 discussed further in the appendix);
- 292 ● positive values over land, indicating that summer maximum temperatures lag more than win-
293 ter minimum temperatures (with the exception of northern Europe which shows little differ-
294 ence between τ_{MAX} and τ_{MIN});
- 295 ● positive values of order 30 days over western North America that are in contrast to the near
296 zero values in the same latitudes (45°N to 55°N) over western Europe, even though both
297 regions are downwind of the ocean and are characterized by a maritime climate;
- 298 ● negative values in the Indian Monsoon region due to τ_{MAX} preceding the summer solstice;
299 and
- 300 ● negative values in the high latitudes of both hemispheres due to the long delay of winter
301 minimum temperatures relative to the delay of the summer maximum.

302 ASYM has a spatial pattern that is continental in scale and has magnitudes on the order of 30
303 days (Fig 2c), comparable in magnitude to the seasonal asymmetry seen in Seattle, and corre-
304 sponding to a two month difference between the length of autumn and spring. Largest positive
305 ASYM values are found primarily on the west coast of North America (particularly in California)
306 with ASYM values in San Diego, Long Beach, Santa Barbara, San Luis Obispo and San Francisco
307 all exceeding 40 days. The largest negative values of ASYM are found at stations in Antarctica
308 and India with values less than -50 days (offscale in Fig. 2) in five Antarctica stations including
309 the South Pole and four stations in India.

310 *c. Vertical structure of seasonal cycle timing*

311 Immediately downwind of the ocean, advection of air that has been influenced by interaction
312 with the ocean surface damps the seasonal amplitude of T_{2m} and increases the phase lag. This
313 advection occurs primarily in the mid and upper troposphere where zonal winds are strongest and,
314 hence, the amplitude (see Fig. 8 of Donohoe and Battisti 2013) and phase of atmospheric temper-
315 ature above the boundary layer are more zonally homogenous (see Fig. 3 of Stine and Huybers
316 2012) than at the surface. In order for zonal advection to influence T_{2m} overland, the upper level
317 seasonal temperature tendencies must be communicated downward via turbulence and radiative
318 heating¹. A central result from the previous section – positive ASYM over the western conti-
319 nents – requires that near surface zonal advection and/or the subsequent vertical communication
320 from the mid-troposphere to the surface does not operate symmetrically for seasonal cooling and
321 warming; we examine the vertical structure of ASYM to understand why.

322 The land-ocean contrast in τ_{MAX} that is prevalent at the surface is nearly zonally and vertically
323 invariant above the boundary layer with near uniform values around 45 days except for lower
324 values over the Arctic and Siberia (Fig. 3 left column and Fig. 4 top). The sharp gradient in τ_{MIN}
325 seen at the surface near the west coast of the United States is present throughout the troposphere.
326 Interestingly, the dipole between large τ_{MIN} to the west and small τ_{MIN} to the east over the western
327 United States at 400 hPa is shifted eastward relative to the surface dipole – so much so that along
328 130°W, τ_{MIN} aloft is delayed from that at the surface by almost 50 days. We speculate that the
329 upper level gradient in τ_{MIN} spanning the eastern Pacific Ocean and western US is caused by the
330 stationary high pressure ridge that is set up by the jet interacting with the Rocky Mountains during
331 December, January and, February (Held 2001; Eliassen and Palm 1961). Southerly advection of
332 warm air west of the ridge delays the seasonal minimum temperature until the ridge breaks down

¹This argument neglects the role of barotropic mountain waves

333 in February; similarly, northerly advection of cold air east of the ridge accelerates the cooling
334 leading to a phase advance of τ_{MIN} (i.e. $\tau_{MIN} < 0$).

335 There is also a north-south dipole in upper level (400 hPa) τ_{MIN} in both the North Pacific and
336 North Atlantic storm track regions with $\tau_{MIN} > 60$ days on the equatorward flank and τ_{MIN} of
337 approximately 30 days on the poleward flank (Fig. 3). We speculate this timing results from the
338 equatorward shift of the jet, storm track and baroclinity during the winter months (which is farthest
339 equatorward in February (Barnes and Simpson 2017): enhanced cooling propagates equatorward
340 with the equatorward migration of the storm track, thus the minimum temperature is delayed near
341 the equatorward limit of the storm track. The upper level τ_{MIN} in the Southern Hemisphere storm
342 track does not display the same equatorward-poleward dipole, perhaps because the seasonal move-
343 ment of the storm track is damped and more semi-annual in nature in the Southern Hemisphere
344 due to the strengthening and separation of the subtropical and sub-polar jets during the winter
345 (Trenberth 1991). Because the upper level map of τ_{MIN} has much more structure than that of
346 τ_{MAX} , upper level ASYM primarily reflects the structure in τ_{MIN} .

347 Fig. 4 shows longitudinal cross sections of the τ_{MAX} , τ_{MIN} and, ASYM, averaged across the lat-
348 itude band from 47°N to 57°N which was chosen to highlight the contrast in structure between the
349 west coast of North America and Europe (Fig. 4). As previously noted, above 700 hPa the zonal
350 profile of τ_{MAX} is much more homogenous than at the surface. Across Eurasia, τ_{MAX} is nearly
351 vertically invariant from the surface to the tropopause and has a value of approximately 30 days –
352 indicative of continental air. This result suggests that convection from the solar heated land mass
353 influences the upper atmosphere during the summer. Except for over the eastern Pacific Ocean,
354 τ_{MIN} is more zonally homogenous aloft than at the surface, with values of approximately 40 days.
355 The wintertime stationary wave generated by the Rocky Mountains is responsible for large (≈ 70
356 day) τ_{MIN} that extends vertically throughout the troposphere in the eastern Pacific. Interestingly,

357 this homogenized sector of high τ_{MAX} slants eastward with altitude and, thus, explains the reversal
358 of ASYM with height seen over the Pacific Northwest US.

359 Over the oceans, the vertical structure of τ_{MAX} in Fig. 4a shows a strong vertical gradient at
360 about 900 hPa with less delayed and vertically homogenous τ_{MAX} above 900 hPa. In contrast,
361 the vertical structure of τ_{MAX} over land is vertically homogenous from the surface all the way
362 through the troposphere suggesting some vertical communication between the surface and upper
363 atmosphere during the summer. The vertical structure of τ_{MIN} shows a strong vertical gradient
364 between 900 hPa and 700 hPa everywhere except over the eastern Pacific – suggesting that the
365 surface is isolated from the upper atmosphere over most regions during the winter.

366 Surface station data (circles circumscribed in white in Fig. 4) and marine observations (circles
367 circumscribed in gray) in the latitude band between 47°N to 57°N confirm the strong zonal gra-
368 dients seen in τ_{MIN} and ASYM in the reanalysis data including the following features: (i) the
369 strong gradient in τ_{MIN} seen along the West coast of North America; (ii) the increasing lag of τ_{MIN}
370 moving from the west coast of the US toward the interior and east coast; (iii) the large, continuous
371 gradient in τ_{MIN} between the ocean domain and land domain over northern Europe. Also shown in
372 Fig. 4 are τ_{MAX} , τ_{MIN} and, ASYM from the high altitude stations on Mount Rainier in Washington
373 state, USA (46.88°N, 121.7° W – plotted at the pressure level of the mean surface pressure at these
374 stations). Data from these stations show delays in τ_{MIN} that are greater than that at the surface and
375 closer to the values seen in the reanalysis atmospheric temperature at the same pressure. Specif-
376 ically, the T_{2m} at Camp Muir (680 hPa) has a prolonged cold season temperature with a delayed
377 τ_{MIN} (≈ 45 days) that is halfway between that at the same pressure in the reanalysis (≈ 60 days)
378 and that at the surface (≈ 20 days—Appendix Fig. A3).

379 **4. Mechanisms of asymmetry in the seasonal cycle of temperature**

380 We now explore several candidate mechanisms to explain the large scale patterns of climatolog-
381 ical ASYM and analyze the support for and shortcomings of each mechanism.

382 *a. Impact of clouds on the seasonal cycle of surface solar radiation*

383 The seasonal cycle of surface temperature is primarily driven by the amount of downwelling
384 solar radiation reaching the surface (DSR). DSR is a function of insolation at the top of atmosphere
385 – dictated by Earth-Sun geometry – and atmospheric shortwave transmissivity which is governed
386 by the concentration of atmospheric constituents that absorb (i.e. water vapor and ozone) and
387 reflect (i.e. clouds) solar radiation. Clouds play the dominant role in determining the spatial and
388 seasonal variability of atmospheric shortwave transmissivity (Taylor et al. 2007; Donohoe and
389 Battisti 2011). Here, we explore whether seasonal variations in cloud cover can explain the spatial
390 pattern of ASYM by way of modifying the seasonal timing of maximum and minimum DSR
391 relative to the insolation.

392 We consider the seasonal cycle of DSR observed in Seattle (Fig. 5a) as an illustrative example
393 and note at the outset that while the seasonal shape of T_{2m} in Seattle does follow (with uniform
394 lag) the seasonal evolution of DSR, this result is limited to the region and does not apply more
395 generally across the globe. Using the same periodic spline fit with 6 knots that was previously
396 applied to temperature, the date of maximum DSR is July 6th, 14 days after the summer solstice,
397 and the minimum DSR occurs on December 11th, 10 days before the winter solstice. We define
398 the timing of the DSR maximum and minimum relative to the solstices as RAD_{MAX} and RAD_{MIN}
399 respectively. The offset of the DSR extrema from the solstices results from the seasonal cycle of
400 atmospheric transmissivity that most likely is due to seasonal cycle in clouds (e.g. cloud fraction
401 and optical depth); this interpretation is supported by analysis of the direct versus diffuse DSR

402 which shows direct DSR is minimized on the 1st of December and is maximized on July 25th
403 (not shown) and is also consistent with the seasonality of precipitation in Seattle, which peaks in
404 November and is a minimum in late July.

405 Unlike the relationship between TOA insolation and T_{2m} at SEATAC, Fig. 5b shows that T_{2m}
406 lags DSR by a consistent phase throughout the seasonal cycle; τ_{MIN} lags RAD_{MIN} by 24 days
407 ($14 - (-10) = 24$ days) and τ_{MAX} lags RAD_{MAX} by 30 days ($44 - 14 = 30$ days). Thus, the seasonal
408 phasing of Seattle temperature is consistent with what is expected in a continental climate due to
409 local (1D) processes forced by DSR with seasonally invariant heat capacity and local feedbacks.
410 Under these assumptions, the temperature response lags the DSR by a constant phase across the
411 seasonal cycle – with phase determined by the ratio of the heat capacity to the damping (Crowley
412 and North 1988). Indeed, the smoothed seasonal cycle of SEATAC temperature (green line in Fig.
413 5b) appears to lag the DSR (black line) by a fixed phase throughout the seasonal cycle. This is
414 quantified in the inset of Fig. 5b by comparing DSR (ordinate) to the temperature (abscissa) at
415 SEATAC with the latter lagged by a (seasonally invariant) time lag of 25 days; the two time series
416 are strongly correlated at $R^2 = 0.99$ indicated by a straight line (black dots in the inset of Fig. 5b).
417 In contrast the same plot of temperature versus insolation – with the optimal time lag recalculated
418 (32 days) – produces an hourglass structure (red dots) that is indicative of a time lag between
419 insolation and temperature that varies with season.

420 Seasonal variations in atmospheric shortwave transmissivity play a first order role in the season-
421 ality of DSR in Seattle; transmissivity varies by a factor of 2 from 0.3 to 0.6 between summer and
422 winter, respectively, which is smaller than – but still significant when compared to – the seasonal
423 cycle of daily mean insolation, which varies by more than a factor of 4 (between 105 and 480 W
424 m^{-2}).

425 These results motivate the following questions: How much does the seasonality of cloud cover
426 alter the seasonal timing of the DSR that drives surface temperature changes across the globe?
427 Can cloud induced shifts in the DSR extrema relative to the solstices explain the large scale pat-
428 terns of ASYM in the T_{2m} shown in Fig. 2? To answer these questions, we quantify the seasonal
429 asymmetry of DSR with the metric RAD_{ASYM} – defined to be the time of the DSR maximum
430 relative to summer solstice (RAD_{MAX}) minus the time of the DSR minimum relative to winter sol-
431 stice (RAD_{MIN}). RAD_{ASYM} can be positive or negative with positive (negative) values indicating a
432 summer maximum that is more (less) delayed than the winter minimum. For example, RAD_{ASYM}
433 has a value of $(14 - (-10)) + 24$ days for Seattle.

434 In situ DSR measurements of sufficient length to establish a global climatology of RAD_{ASYM}
435 do not exist. Thus, we calculate RAD_{ASYM} from the satellite derived gridded climatological DSR
436 record from CERES satellite data. The CERES data from the nearest gridpoint to Seattle (blue dots
437 and lines in Fig. 5a) are in excellent agreement with the local radiometer data in absolute mag-
438 nitude, internal variability and, most importantly, the phasing of the seasonal cycle in DSR, with
439 RAD_{MIN} and RAD_{MAX} differing by less than 3 days between the two data sets. In the appendix
440 we show that RAD_{ASYM} calculated from CERES is in close agreement with that calculated using
441 in situ DSR observations at other stations in the National Solar Radiation Database of the United
442 States of America (see Fig. A5 Sengupta et al. 2018), albeit with substantial disagreement on the
443 amplitude of the seasonal cycle in DSR. Specifically, station data confirm the positive RAD_{ASYM}
444 values seen in the northern latitudes of the United States extending from the Pacific coast to the
445 mid-west and the negative RAD_{ASYM} values seen in southwest region (due to the influence of the
446 North America monsoon), the southeast (i.e. Florida) and in Alaska.

447 The global pattern of RAD_{MAX} (Fig. 6a) shows patterns that are regional in extent and have
448 values with magnitudes of approximately 30 days. RAD_{MAX} is positive over most of the United

449 States with largest values on the west coast. RAD_{MAX} is positive in the Mediterranean region
450 and near zero in Northern Europe. Negative values of RAD_{MAX} are seen in the monsoon regions
451 of India, South-East Asia, Africa and South-West North America where cloud cover increases
452 well before the summer solstice and is sustained thereafter, resulting in maximum DSR prior
453 to monsoon onset. Over the oceans, there is a pronounced North-South dipole of RAD_{MAX} in
454 the North Pacific, North Atlantic, South Pacific and South Atlantic with positive values on the
455 equatorward flank of the storm track and negative values on the poleward flank. The sharp north-
456 south dipole in RAD_{MAX} is a genuine feature of the data that we illustrate in the inset of Figure
457 6a by plotting the time series of DSR at two points on different sides of the dipole in the North
458 Pacific (indicated by the red and blue circles on the map). To the south of the nodal line, RAD_{MAX}
459 occurs 30 days after the summer solstice (red time series) whereas to the north RAD_{MAX} occurs
460 30 days before the summer solstice (blue time series). This dipole reflects a poleward shift of the
461 storm track from spring to summer, with reduced cloud cover prior to the shift (i.e. in the spring)
462 on the poleward side and after the shift (i.e. in the late summer) on the equatorward flank.

463 The negative values of RAD_{MAX} in the marginal ice zones, especially in the Southern Ocean,
464 reflect the retreat of sea ice prior to the summer solstice by the following mechanism: the high
465 surface albedo of the ice in the spring reflects solar radiation incident on the surface and a portion
466 of this upwelling solar radiation from the surface is reflected back to surface by clouds (Taylor
467 et al. 2007; Donohoe and Battisti 2011). The presence of ice in the spring enhances both upwelling
468 solar radiation at the surface and DSR with a net decrease in solar radiation absorbed at the surface.
469 Thus, the negative values of RAD_{MAX} in the marginal ice zones do not accelerate the timing of
470 τ_{MAX} since the phase advanced DSR is not indicative of the timing of *net* solar input to the surface.

471 The global pattern of the timing of the DSR minimum relative to the winter solstice (RAD_{MIN}
472 – Fig. 6b) has much smaller amplitudes than that of RAD_{MAX} . We find no obvious explanation

473 for this result. Because RAD_{MIN} is everywhere smaller in magnitude than RAD_{MAX} , RAD_{ASYM}
474 strongly resembles the map of RAD_{MAX} (c.f. Fig. 6c and a,b).

475 We now evaluate whether the seasonal timing of DSR and temperature are correlated at the
476 global scale by comparing the spatial structure of RAD_{ASYM} (colors in Fig. 6c) and ASYM (con-
477 tours with the same colorbar). On the west coast of North America, the region of highest positive
478 ASYM values and RAD_{ASYM} are co-located over the Pacific Northwest and along the Pacific coast
479 down to the Mexican border. In these regions, T_{2m} lags DSR by ≈ 15 days throughout the sea-
480 sonal cycle – consistent with a local continental response to solar forcing. Similarly, over western
481 Europe ASYM and RAD_{ASYM} are both generally positive over the Mediterranean region and are
482 near zero to the north. This finding suggests that the seasonal cycle of cloudiness may explain the
483 contrasting ASYM between the west coast of North America and Europe by way of modifying the
484 phasing of the seasonal cycle of surface heating relative to the phasing of insolation.

485 On the global scale, however, ASYM and RAD_{ASYM} are not significantly spatially correlated.
486 This is especially true over the oceans where the strong north-south dipole in RAD_{ASYM} seemingly
487 has no impact on ASYM which is predominantly negative throughout the oceans except for posi-
488 tive values in the northwest Pacific. The spatial correlation between ASYM and RAD_{ASYM} is also
489 not significant over the global land masses considered collectively due to the lack of correlation
490 over the eastern half of Eurasia. These results suggest that while RAD_{ASYM} may offer an explana-
491 tion for the contrasting ASYM in the otherwise similar maritime climates of western Europe and
492 western North America, RAD_{ASYM} is *not* the primary cause of ASYM at the global scale. Thus,
493 we look for alternative mechanistic controls of ASYM.

494 *b. Land-ocean contrast in ASYM*

495 1) THE SEASONAL EVOLUTION OF T_{2m} OVER OCEAN AND LAND

496 The prevailing (first-order) large scale pattern of ASYM consists of negative values over the
497 ocean domain and positive values over the land domain, indicative of winter minimum tempera-
498 tures that are more delayed than summer maximum temperatures over the ocean and vice versa
499 over land. We visualize the contrasting seasonal shape of T_{2m} by co-plotting temperature data
500 from a station representative of a extratropical continental climate (Tynda in the Siberia region
501 of Russia, 55°N, 125°E, 1600 feet elevation, green line, Fig. 7a) with that of a marine climate
502 (Ocean Weather Station Papa in the North Eastern Pacific, 50°N, 145°W, purple line). For point of
503 reference, ASYM for the continental record is +12 days whereas that of the marine record is -20
504 days which are typical continental and marine ASYM values across the globe (see Fig. 2c). The
505 continental station is colder on average than the marine station (mean of -4.7°C versus 8.3°C), has
506 a much larger amplitude seasonal cycle (annual harmonic amplitude of 24.2°C versus 3.8°C) and
507 has a smaller time lag between T_{2m} and insolation (annual harmonic phase lag of 35 days versus
508 87). To easily visualize differences in the shape of the seasonal cycle in these two locations, we
509 rescale T_{2m} so that curves have the same mean vertical position and distance between temperature
510 extremes, and shift the continental record to right (more delay) by 52 days (equal to the phase
511 difference of the respective annual harmonics) so that seasonal maxima are nearly aligned (dashed
512 red line). Comparing the dashed green and solid purple lines in Fig. 7a, we see (a) T_{2m} in the
513 continental record lingers in the vicinity of the summer maximum whereas the T_{2m} in the marine
514 record is sharply peaked in summer with a delayed and more rapid temperature increase in the
515 spring, and (b) the winter minimum in the continental record is much more sharply peaked than
516 that in the marine record, which cools slowly during the late autumn and winter compared to the

517 land station. The substantially different temporal evolutions seen at Tynda and Ocean Station Papa
518 are typical of those in continental and marine environments, respectively.

519 2) SEASONAL EVOLUTION OF ATMOSPHERIC AND OCEANIC BOUNDARY LAYER DEPTH

520 We argue below that the ASYM contrast between land and ocean domains is due to the different
521 seasonality in the depth of the atmosphere and ocean that is greatly influenced by the surface
522 forcing – hereafter referred to as the boundary layer depth. The seasonality of boundary layer depth
523 differs between the atmosphere and ocean due to the contrasting influence surface solar heating has
524 on the stability of the two fluids: increased surface solar heating in summer (compared to winter)
525 stabilizes the ocean boundary layer causing it to thin (shoal), whereas increased surface solar
526 heating destabilizes the atmosphere over land and thus greatly deepens the atmospheric boundary
527 layer (by convection) in the summer compared to winter. The seasonality of ocean mixed layer
528 depth is amplified by the seasonal cycle in surface wind stress with reduced surface wind speed in
529 the summer resulting in further shoaling of the oceanic boundary layer (Montegut et al. 2004). The
530 lag between surface temperature and solar heating increases with increased boundary layer depth
531 because: (i) a deeper layer has a greater heat capacity (thermal inertia) which is more important
532 in the ocean and (ii) a deeper layer is coupled to advective processes away from the surface which
533 is more important in the atmosphere. Together, this causes: an earlier τ_{MAX} and delayed τ_{MIN} in
534 SST (Fig. 8b,d) resulting in a positive ASYM in SST throughout the extratropics (Fig. 8f); and
535 a delayed τ_{MAX} and an earlier τ_{MIN} in surface air temperature over land (Fig. 2a,b) and hence
536 positive ASYM (Fig. 2c).

537 The negative ASYM in SST is due to slower cooling of the SST in the winter and early spring
538 (blue time series in Fig. 9) in contrast to a rapid warming of SST in the fall. This temporal evolu-
539 tion of SST is ubiquitous in the mid-latitude oceans, is well represented in the gridded reanalysis

540 (cyan time series in Fig. 9) and it differs substantially from an annual sinusoid. We illustrate
 541 the impact of seasonal variations in mixed layer depth on the seasonal evolution of SST using a
 542 simplified thermodynamic budget of the ocean mixed layer (Hasselmann 1976):

$$\frac{dH(t)SST'}{dt} = S'(t) - \lambda SST'. \quad (1)$$

543 In Eq. 1, SST is the ocean mixed layer temperature, $H(t)$ is the seasonally dependent heat capac-
 544 ity of the oceanic mixed layer, $S(t)$ is the solar radiation entering the ocean, λ is the sensitivity
 545 of turbulent fluxes (latent plus sensible) to surface temperature changes and primes ($'$) indicate
 546 anomalies from energetically balanced annual mean quantities. Note that we have ignored the
 547 role of the (seasonal anomalies in) oceanic heat flux convergence and the impact of atmospheric
 548 temperature changes on the downward surface heat fluxes. A sinusoidal solar forcing results in a
 549 sinusoidal SST response for a given temporally invariant value of H with amplitude (denoted by
 550 vertical brackets, $\|$):

$$\|SST\| = \|S\| / \sqrt{(H\sigma)^2 + \lambda^2}, \quad (2)$$

551 where $\frac{2\pi}{\sigma} = 365d$. The phase lag (ϕ) relative to the solar forcing increases with increasing H :

$$\phi = \text{atan} \left(\frac{\sigma H}{\lambda} \right). \quad (3)$$

552 However, H is much deeper in the winter than in the summer. Hence, one would expect the
 553 decrease in SST in the winter to be less than the increase in SST in the summer and the lag be-
 554 tween the minimum insolation and minimum SST (τ_{MIN}) would be greater than the lag between
 555 the maximum insolation and maximum SST (τ_{MIN}). This characteristic slowly evolving (i.e. de-
 556 layed) minimum SST is seen at all OWS stations (Fig. 9 – blue lines). Numerical solutions with

557 seasonally varying H demonstrate that this expected seasonal dependence of phasing (eq. 3) and
558 amplitude (eq. 2) of SST roughly holds for reasonable choices of temporally variant H and λ and,
559 furthermore, that the slowly evolving SST minimum seen in Fig. 9 is replicated in these numerical
560 simulations of Eq. 1 (not shown).

561 We now turn to the seasonal cycle of the surface air temperature (T_{2m}) overlying the ocean.
562 In the interior of the ocean basins, the seasonal evolution of T_{2m} mirrors that of SST with τ_{MAX}
563 nearly in-phase in the two fluids and τ_{MIN} within 2 weeks of one another (c.f. the red and blue
564 time series at OWS stations Echo and Papa in Fig. 9a,d). An in phase relationship between SST
565 and T_{2m} is expected for an isolated (no advection, far downwind of land) atmospheric column
566 coupled to an ocean. In such a system, the atmosphere is primarily heated seasonally by the
567 absorption of solar radiation in the atmospheric column and damped by turbulent exchange with
568 the ocean – with seasonal energy storage in the atmospheric column being non-trivial but smaller
569 in magnitude (Fig. 7 of Donohoe and Battisti 2013). In such a system, the time scale of adjustment
570 between the atmosphere and the ocean is given by the atmospheric heat capacity ($\frac{C_p P_s}{g}$) divided
571 by the sensitivity of turbulent fluxes to the contrast between SST and T_{2m} ($\lambda_{SST} \approx 30 \text{ W m}^{-2}$
572 K^{-1}) which is approximately 4 days. Thus the atmosphere and ocean adjust in concert to seasonal
573 insolation changes and are driven from the top down.

574 We now turn to the impact of seasonal variations in the atmospheric boundary layer depth on the
575 seasonal evolution of T_{2m} over land masses (the green line in Fig. 7a). During the summer, solar
576 heating of the surface destabilizes the surface atmosphere resulting in convection and an efficient
577 connection between the surface and mid-troposphere, where zonal advection of marine-influenced
578 air ameliorates the amplitude and alters the phase of T_{2m} (Stine and Huybers 2012). In contrast,
579 during the winter, the reduced surface solar heating results in net longwave cooling of the surface
580 that stabilizes the atmospheric boundary layer and effectively isolates the air in the boundary

581 layer from the influence of the air aloft. The result is a rapid cooling of the surface, nearly in
582 phase with winter solstice due to the low heat capacity of the surface and lower atmosphere.
583 The thermodynamic de-coupling between the atmospheric boundary and mid-troposphere over
584 the continents during the winter can be seen in the cross sections in Fig. 4: τ_{MIN} in the surface
585 boundary layer is distinct from that in the free troposphere. In contrast, τ_{MAX} is nearly vertically
586 homogenous throughout the atmosphere over the continents suggesting strong coupling between
587 the boundary layer and mid troposphere during the summer.

588 3) SEASONAL CYCLE OF T_{2m} AND AND SST IN THE WESTERN OCEAN BASINS

589 In contrast to the in-phase relationship between T_{2m} and SST seen in the interior of the ocean
590 basin, immediately downwind of the continents, the winter minimum in T_{2m} is almost 4K colder
591 than and precedes the minimum in SST by over a month (e.g. at Station Bravo and India –Fig.
592 9c,f). We demonstrate here that the seasonal evolution of both SST and T_{2m} in this region are pro-
593 foundly impacted by atmospheric advection off the continent and, thus, depart from the expected
594 behavior of an coupled atmospheric-oceanic column responding to seasonal changes in insolation
595 that was developed in the previous subsection.

596 Eq. 1 assumes that the seasonal cycle of SST is driven by solar radiation at the surface, damped
597 by turbulent energy fluxes (via λ_{SST}) and ignores changes in the turbulent and downwelling long-
598 wave radiation that are *driven*, in part, by changes in the near surface atmospheric temperature.
599 This assumption fails on the western side of the North Atlantic and North Pacific ocean basins
600 where the seasonal input of energy into the ocean surface – defined as the amplitude of the sea-
601 sonal anomaly that is in phase with the insolation (see Donohoe and Battisti 2013) – is primarily
602 associated with the *atmosphere fluxing energy into (out of) the ocean* column during the warm
603 (cold) season (c.f Fig. 10a and b). This strong seasonal heating and cooling of the ocean im-

604 mediate downwind of the continents results from atmospheric advection off the continent that
 605 amplifies the seasonal variations in ocean energy content: (i) cold air advection during winter pulls
 606 energy out of the ocean (via turbulent energy fluxes) further cooling the SST and (ii) warm air ad-
 607 vection during summer pushes (anomalous) energy into the ocean, further warming the SST. In
 608 contrast, on the eastern side of the North Atlantic and Pacific and throughout the Southern Ocean,
 609 the seasonal input of energy into the ocean column is dominated by solar heating of the surface.
 610 The contrasting source of seasonal oceanic heating can be seen from a plot of the ratio of the sea-
 611 sonal contribution of atmosphere-ocean exchange to that of solar heating (Fig. 10c) which is >1
 612 on the western half of the basins and <1 on the eastern half. This ratio is an excellent predictor of
 613 the seasonal amplitude of SST (shown in contours in Fig. 10c) which suggests that the amplitude
 614 of seasonal variations in SST is set by seasonal variations in atmospheric advection and is largely
 615 independent of seasonal variations in oceanic circulation. Because the seasonal atmospheric ad-
 616 vection off the continent is nearly in phase with insolation, the phasing of the seasonal cycle of
 617 SST is well captured by Eq. 3 with the departures from a sinusoid (and the differing values of
 618 τ_{MAX} and τ_{MIN}) due to the seasonal evolution of H .

619 The dominant atmospheric energy balance on seasonal timescales over the western ocean basins
 620 is between atmospheric advection and surface heat fluxes (c.f. Fig. 10d and e). In physical terms,
 621 during the winter, westerly winds bring cold air from the continent that strongly cools (magnitude
 622 $> 150 \text{ W m}^{-2}$ – Fig. 10e) the atmosphere over the western ocean basins. T_{2m} is 4K colder than
 623 the SST with the contrast between SST and T_{2m} peaking in February (Fig. 9c,f). The cold air
 624 pulls energy out of the warmer ocean via upward turbulent energy fluxes ($\approx 150 \text{ W m}^{-2}$,
 625 not shown) – consistent with estimates of λ_{SST} . For regions in which the seasonal heating by
 626 atmospheric advection exceeds that of heating by atmospheric solar absorption (denoted by ratios
 627 > 1 in Fig. 10f), atmospheric heat flux divergence is balanced by upward turbulent energy fluxes.

628 Because the seasonal evolution of SST is more gradual (smaller in magnitude) than that of T_{2m} ,
629 the turbulent exchange is proportional to and in-phase with T_{2m} . Thus, the minimum in T_{2m} occurs
630 during the maximum in cold air advection (atmospheric heat flux divergence) which occurs in
631 January and February when the lateral gradient of T_{2m} between the ocean and land and the zonal
632 wind strength are greatest. In contrast, the SST will continue to cool until the radiative heating
633 exceeds the atmospheric energy flux divergence, which occurs approximately a month later.

634 From a Lagrangian perspective, the seasonal disequilibrium between T_{2m} and SST results from
635 air being advected to a region of substantially different SST on the timescale of atmospheric ad-
636 justment to the SST via the turbulent energy fluxes (4 days) due to the strong lateral gradient of
637 SST over the western ocean basin. The lateral SST gradient in the winter is a consequence of the
638 zonal flux of cold air off the continent (Fig. 10c); thus, the τ_{MIN} contrast between T_{2m} and SST
639 results from atmospheric advection.

640 In the eastern ocean basins there is substantial disagreement between τ_{MIN} calculated from the
641 OWS T_{2m} and that calculated from the ERA T_{2m} at the nearest neighbor (c.f. the dots and colors
642 in Fig. 9E). This discrepancy is associated with the seasonal minimum in T_{2m} in the reanalysis
643 being more in phase with the SST minimum as compared to the OWS T_{2m} , which is coldest a full
644 two months before the SST minimum at stations such as OWS India (Fig. 9F) and is unrelated
645 to biases in the SST assimilated into the ERA reanalysis. One possible cause is that winter-time
646 cold air advection off the continents in the ERA model does not extend far enough east across the
647 ocean basin.

648 *c. Seasonal timing in the high latitudes*

649 ASYM is negative in the high latitudes of both hemispheres (Fig. 2c) with values of <-30 days
650 over the Arctic sea ice and the Greenland and Antarctic ice sheets. This seasonal asymmetry of

651 surface temperature results from the relatively short lag between the summer solstice and time of
652 maximum temperature ($\tau_{MAX} = 20-30$ days) typical of a continental climate with small surface
653 heat capacity, and long τ_{MIN} values (> 60 days). These regions are effectively isolated from
654 marine influence at all times of year. The relatively long delay in seasonal minimum temperature
655 results from the temporal evolution of insolation which deviates greatly from the annual harmonic
656 above the Arctic and Antarctic circle, where insolation flat lines at zero during the polar night (red
657 line in Fig. 7b). The definition of the winter solstice is fairly meaningless in these regions since
658 the minimum insolation is realized over the course of several months. We use the temperature
659 record at Greenland Summit Station (Shuman et al. 2001, 72.5°N and 38.5°W) over the time
660 period 2008-2018 to demonstrate the characteristic shape of the seasonal cycle of T_{2m} over the
661 high latitude ice sheets (black line in Fig. 7b). After a well defined peak in temperature in mid-
662 July, temperatures decrease steadily until the beginning of the polar night in November. During the
663 polar night, temperature continues to decrease slowly until just after the Sun rises again in March.
664 τ_{MIN} occurs just after the end of the polar night – timing that is analogous to the timing of the
665 minimum temperature in the diurnal cycle, which generally occurs just after sunrise because the
666 surface continues to cool until the absorbed solar radiation exceeds the emitted longwave radiation
667 (Holmes et al. 2013).

668 *d. Impact of monsoons on the seasonal cycle of temperature*

669 The onset of the Indian monsoon (around the beginning of June) heralds increased evaporative
670 cooling of the northern Indian Ocean and an increase in cloud cover and precipitation over the
671 continent, thereby reducing DSR and increasing the soil moisture available for evaporation and
672 cooling the surface (Gadgil 2003). The Indian monsoon region, along with a small section of the
673 North American monsoon, are the only regions on Earth where τ_{MAX} occurs before the summer

674 solstice. The timing of the temperature maximum relative to the onset of monsoonal precipitation
675 is analyzed from the station data in New Delhi, India which is representative of the climate in the
676 region (Fig. 7d). Temperature rapidly increases from the seasonal minimum in December until
677 the 1st of June, when the precipitation begins to ramp up (blue line). Thereafter, the temperatures
678 gradually decline.

679 The beginning of this cooling period corresponds to the onset of the Monsoon in the region
680 (India Meteorological Department 2019; Webster et al. 1998) which, in the climatological average,
681 occurs in the middle of May in Southwest India and approximately a month later in Northern India
682 (see Fig. 3 of Krishnamurthy and Shukla 2000). The onset of the monsoon is coincident with an
683 increase in atmospheric water vapor (see Fig. 3 in Singh et al. 2004), a migration of the clouds
684 over the region (Sikka and Gadgil 1980) causing a reduction in downwelling solar radiation at the
685 surface (Yasunari 1979) of order 60 W m^{-2} (see Fig. A6), and an increase in soil moisture (Nayak
686 et al. 2018). The reduction in surface solar radiation and enhanced soil moisture act in concert to
687 cool the surface via reduced energy input to the surface and an increase in surface energy loss via
688 an increased upward latent heat flux. This cooling tendency prior to the summer solstice results in
689 negative values of τ_{MAX} .

690 In contrast to the negative τ_{MAX} values in the Indian monsoon region, τ_{MAX} in the East Asian
691 Monsoon region are of order +30 days. The East Asian Monsoon (north of 25N) is characterized
692 by the Meiyu Front and by persistent rain and clouds within and to the south of the Front; these
693 rains last from early Spring to early Summer. In mid-July, when insolation is nearly at peak, the
694 Meiyu Front breaks up and clouds and rain become more episodic; as a result, the temperature rises
695 and reaches a maximum 30 days after solstices, in late July. We note that we do not consider the
696 seasonal timing of temperature in the Southeast Asian monsoon in this study because this region

697 is primarily equatorward of 25° and, thus, has a significant semi-annual component of insolation,
698 T_{2m} and precipitation (Misra and DiNapoli 2014) that makes the definition of τ_{MAX} problematic.

699 **5. Summary and conclusions**

700 The phase and amplitude of the annual harmonic in surface temperature has been used exten-
701 sively in the existing literature to probe the underlying climate physics responsible for seasonality
702 and its changes under global warming (Santer et al. 2018; Laepple and Lohmann 2009; Dwyer
703 et al. 2012; Stine et al. 2009; Stine and Huybers 2012; Donohoe and Battisti 2013). We have
704 demonstrated that the departures of T_{2m} from the annual harmonic are substantial, vary greatly at
705 the regional scale across the globe and are well encapsulated by the metric ASYM – the seasonal
706 asymmetry in the lag between the solstices and the temperature extrema. Table 1 provides a sum-
707 mary ASYM values identified in different regions, the proposed relevant physical mechanism and
708 the section of the manuscript where we discuss the mechanism.

709 To emphasize how different the temporal evolution of the seasonal cycle can be between different
710 regions of the Earth, we draw the reader’s attention to the panels of Fig. 7 which shows disparate
711 but emblematic examples of contrasting seasonal cycles (scaled and shifted for ease of compar-
712 ison). Fig. 7a shows a typical maritime seasonal cycle of T_{2m} (station PAPA, with a prolonged
713 seasonal minimum – purple line) and a continental seasonal cycle of T_{2m} (Tynda, Siberia, with
714 a prolonged seasonal maximum – green line). Fig. 7c shows two maritime climates on the west
715 coast of the continents – one with a long duration warming period in North America (San Fran-
716 cisco, CA, USA – green line) and one with a short duration warming period in Europe (Bergen,
717 Norway – purple line). The contrasting shape of the seasonal cycle in T_{2m} between San Francisco
718 and Bergen highlights that continentality is not the only control of the seasonal phasing of T_{2m} .
719 The phasing of T_{2m} in Europe a North America differ due to the contrasting seasonality of cloud

720 cover – specifically the persistent summer stratus decks on the west coast of the United States that
721 have no counterpart in western Europe. Additionally, the seasonal phasing of T_{2m} is influenced by
722 the contrasting seasonal migration of the Atlantic and Pacific storm tracks: the Pacific storm track
723 is centered on Alaska while Bergen is stormy throughout the year.

724 We note that the substantial differences evident in the time series highlighted in Fig. 7a and
725 7c correspond to modest values of ASYM (-20 days for the maritime record, +12 days for the
726 continental record, +44 days for the North American record and -1 day for the European record).
727 These examples of contrasting seasonal shapes are not extreme cases; they are representative of
728 differences in the shape of the seasonal cycle encapsulated by the amplitude of large scale patterns
729 of ASYM seen across the globe. Some of these points have been made previously at the local
730 scale (van den Broeke 1998) and noted briefly in station data on a global scale (May et al. 1992)
731 but the present work provides a comprehensive global analysis of ASYM based in both reanal-
732 ysis and station data and extending throughout the troposphere. Much physical insight has been
733 gleaned from analysis of the phase and amplitude of the annual harmonic in T_{2m} that has provided
734 insights that are both fundamental and applicable to the long-term climate processes (i.e. climate
735 feedbacks) relevant to global warming. We hope that the analysis of ASYM presented here will
736 inspire future work probing the underlying physics that will lead to new insights relevant for large
737 scale processes at climate timescales.

738 We have identified in sections 4 and 5 several candidate mechanisms that explain the spatial
739 pattern of ASYM seen in various locations including: the atmospheric versus oceanic contrast in
740 seasonal boundary layer depth in response to solar heating, the phase differences in surface solar
741 radiation and top of atmosphere insolation due to clouds, and the impact of seasonal evolving
742 atmospheric circulations such as the monsoon and stationary waves. This list is by no means

743 exhaustive and we find no single mechanism that adequately explains the broad scale features of
744 ASYM.

745 We emphasize that much of our analysis is exploratory. In future work we hope to evaluate if
746 ASYM is realistically simulated in comprehensive models and in a hierarchy of lower complexity
747 models (i.e. uncoupled, slab ocean, simplified radiative code, single column models) to isolate
748 the fundamental processes responsible for ASYM (and its spatial variability). An observational
749 pathway forward would be to analyze whether the phasing of the diurnal cycle varies seasonally in
750 the same sense of ASYM. This result would suggest that single column energetics associated with
751 the atmospheric depth over which surface heating is communicated is an essential component in
752 determining ASYM (and its spatial variability).

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754 and Dennis D’Amico for assistance with Camp Muir temperature data. AD’s work was partially
755 funded by the NSF Antarctic Program Grant Number PLR 1643436.

756 APPENDIX

757 Five supplemental figures to support the findings of the main text are shown in this appendix.
758 Fig. A1 show a global comparison of ASYM calculated from station data (abscissa) versus that
759 calculated from the nearest-neighbor ERA reanalysis. Fig. A2 shows the same analysis of the
760 seasonal timing of T_{2m} presented in Fig. 2 using the NCEP reanalysis 2 meter air temperature
761 instead of the ERA-interim data. Fig. A3 compares the seasonal cycle of T_{2m} at various altitudes
762 in the Seattle area using high elevation weather stations on Mount Rainier to probe the strong
763 vertical gradients in τ_{MIN} seen in Figs. 3 and 4.

764 In Fig. A4 we present an explanation for the curious sharp west-east gradient in τ_{MIN} over the
765 ocean south of Australia in both reanalysis products (Fig. 2 and A2); τ_{MIN} values to the west

766 of Australia are typically > 70 days while those to the east are ≈ 20 days. This is an artifact,
767 however. Throughout the region there is a 100 day period where temperature is very close to the
768 seasonal minimum with a slight warming ($\approx 0.1\text{K}$) in the middle of this period (see Fig. A4)
769 resulting in two local temporal minima of T_{2m} at locations throughout the region. East (west) of
770 Australia, the absolute minimum temperature occurs early (late) in this 100 day interval. (The
771 minor local maxima in the middle of this 100 day interval is likely due to seasonal variations in
772 upwelling). We view this spatial discontinuity of τ_{MIN} as a shortcoming of the methodology used
773 here to characterize the seasonal cycle of T_{2m} which was intended to represent a seasonal cycle
774 with only one maximum and minimum.

775 Figure A5 compares the phasing of DSR calculated at stations in the United States to that cal-
776 culated at the nearest-neighbor gridpoint from the CERES data. Station data confirm the positive
777 RAD_{ASYM} values seen in the northern latitudes of the United States extending from the Pacific
778 coast to the mid-west and the negative RAD_{ASYM} values seen in southwest region (influence of
779 the North America monsoon), the southeast (i.e. Florida) and in Alaska.

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894 mary mechanisms responsible for τ_{max} and τ_{min} , and the section of the paper
895 that discusses the emblematic region. 42

TABLE 1: Summary of emblematic regions, their typical values of τ_{max} and τ_{min} , the primary mechanisms responsible for τ_{max} and τ_{min} , and the section of the paper that discusses the emblematic region.

T_{2m}			
Domain	Delay	Mechanisms	Section
Mid-continent	$\tau_{max} \sim 30 d$ $\tau_{min} \sim 20 d$	Low heat capacity of land; seasonal cycle in atmospheric boundary layer depth	4b1, 4b2
Midlatitude ocean far from land	$\tau_{max} \sim 60 d$ $\tau_{min} \sim 80 d$	High heat capacity; seasonal cycle in ocean boundary layer depth	4b1,4b2
Midlatitude ocean immediately downwind of land	$\tau_{max} \sim 50 d$ $\tau_{min} \sim 40 d$	T_{2m} and SST out of phase due to atmospheric advection	4b3
Polar regions; sea ice	$\tau_{max} \sim 30 d$ $\tau_{min} \sim 60 d$	6 months of darkness, low heat capacity in summer	4c
India	$\tau_{max} \sim -20 d$ $\tau_{min} \sim 20 d$	Reduced mid and late summer DSR associated with monsoon onset	4d
Seattle	$\tau_{max} = 44 d$ $\tau_{min} = 14 d$	Seasonal cycle of DSR due to seasonality of clouds	4a

SST			
Domain	Delay	Mechanisms	Section
Midlatitude ocean (far from land)	$\tau_{max} \sim 60 d$ $\tau_{min} \sim 90 d$	Large heat capacity; seasonal cycle in boundary layer thickness	4b1, 4b2
Midlatitude ocean immediately downwind of land	$\tau_{max} \sim 60 d$ $\tau_{min} \sim 70 d$	Atmospheric advection; high heat capacity; seasonal cycle in boundary layer thickness	4b3

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900 thick black line shows the spline fit to the data. The vertical black lines show the date of
901 the spline fit maximum (τ_{MAX}) and minimum (τ_{MIN}) and the vertical red lines show the date
902 of summer and winter solstices. The dashed blue line shows the annual harmonic fit. Ticks
903 on the x-axis are on the start of each calendar month. (Bottom Panel) As in the top panel
904 but superimposed with T_{2m} taken from with the nearest grid point to SEATAC in the ERA-
905 Interim reanalysis product. The light blue shaded area represents the one standard deviation
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1002 5,400 feet) and (blue) Camp Muir (elevation 10,188 feet) with that near sea level at (black)
1003 SEATAC. The gray dashed line is the seasonal cycle of SST at ocean station PAPA shifted
1004 and rescaled for visual comparison. The vertical lines indicate the timing of τ_{MAX} and τ_{MIN}
1005 in each time series given by the same color. 58

1006 **Fig. A4.** Time series of ERA surface air temperature at gridpoints separated by 3° longitude spanning
1007 the discontinuity in ASYM south of Australia seen in Fig. 2c at 55°S . The blue time series
1008 is from 55°S , 145°E with ASYM = -44 days and the red time series is from 55°S , 148°E
1009 with ASYM = +27 days. The dashed vertical lines show the timing of τ_{MIN} and τ_{MAX}
1010 for the time series of the same color. 59

1011 **Fig. A5.** Comparison of RAD_{MAX} , RAD_{MIN} and RAD_{ASYM} calculated in the gridded CERES data to
1012 that calculated using DSR measurements from the US (Sengupta et al. 2018). The dots show
1013 the observed daily mean DSR and the solid lines show the smoothed climatological seasonal
1014 cycle. The direct radiation is shown in red, the diffuse in green and the total observed DSR in
1015 black. The dashed black line shows the smoothed seasonal cycle of total DSR from CERES
1016 data at the nearest neighbor gridbox. Vertical lines show the time of RAD_{MAX} and RAD_{MIN}
1017 identified in the smoothed timeseries of the same linetype. The purple vertical lines show
1018 the winter and summer solstices. The inset map shows the RAD_{ASYM} values at each of the
1019 stations (colored dots circumscribed by gray) overlaid on (colors) RAD_{ASYM} calculated
1020 from the CERES DSR data. 60

1021 **Fig. A6.** **Change in surface solar radiation associated with Monsoon onset. Map of difference in tem-**
1022 **porally smoothed (spline with 10 knots) downwelling solar radiation at the surface between**
1023 **after summer solstice (July 10) and 20 days prior to summer solstice (June 1st) from CERES**
1024 **data.** 61

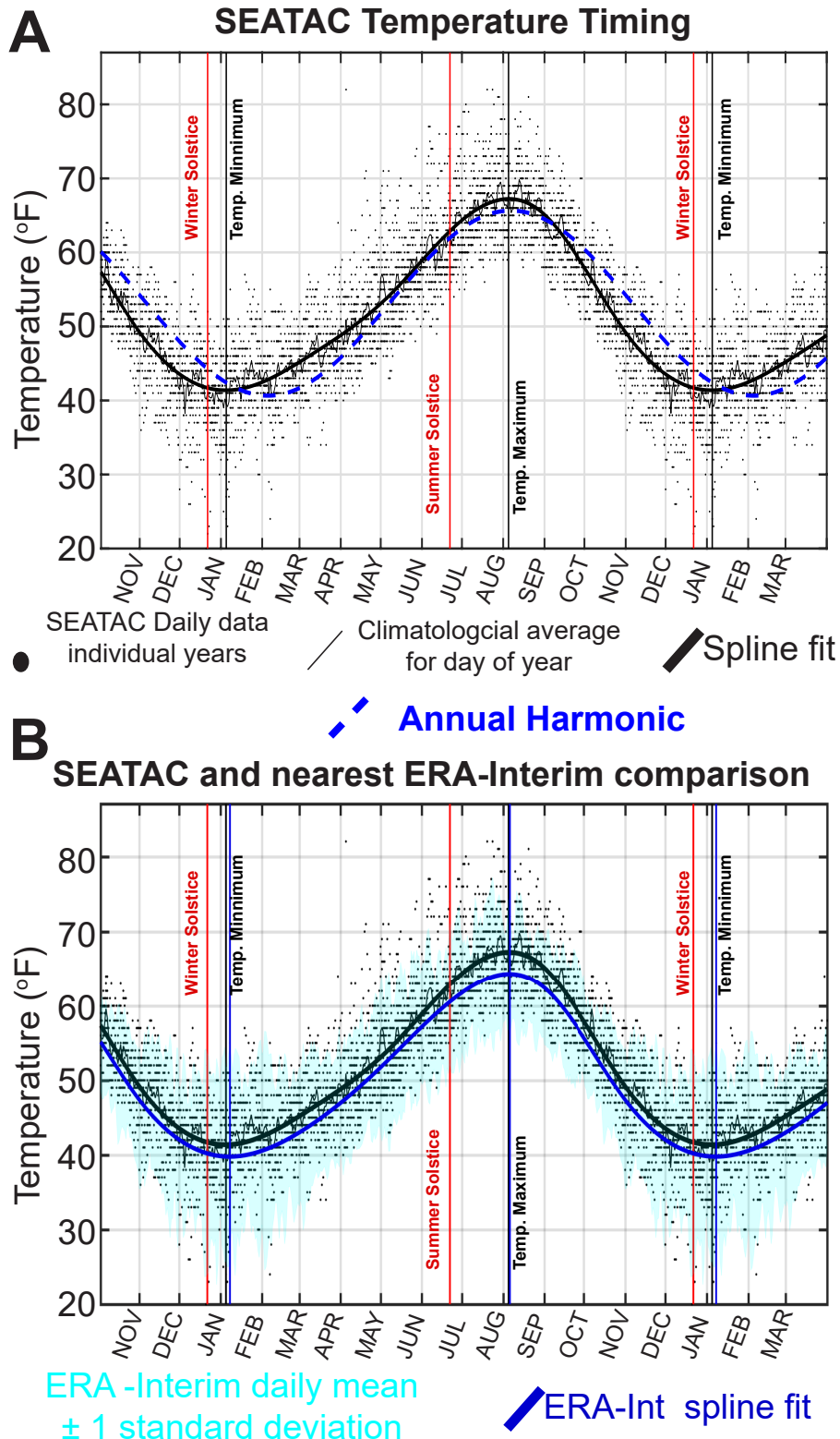
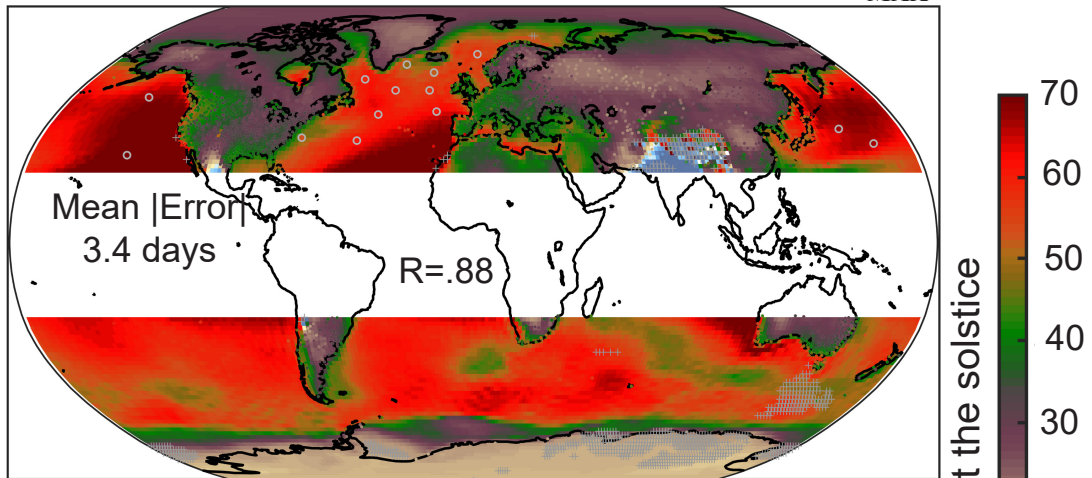
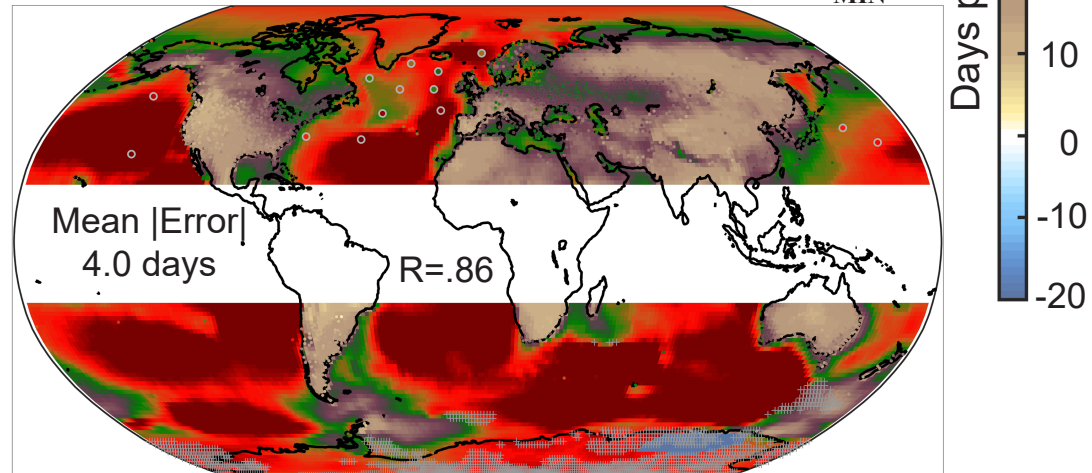


FIG. 1: (Top Panel) Seasonal cycle of daily average surface (2m air) temperature (T_{2m}) at SEATAC airport on a partially repeated periodic domain. The dots show daily data for individual years. The black thin lines show the climatological average for each calendar day and the thick black line shows the spline fit to the data. The vertical black lines show the date of the spline fit maximum (τ_{MAX}) and minimum (τ_{MIN}) and the vertical red lines show the date of summer and winter solstices. The dashed blue line shows the annual harmonic fit. Ticks on the x-axis are on the start of each calendar month. (Bottom Panel) As in the top panel but superimposed with T_{2m} taken from with the nearest grid point to SEATAC in the ERA-Interim reanalysis product. The light blue shaded area represents the one standard deviation of the ERA data. The solid blue line is the spline fit to the ERA data and vertical blue lines show the date of the τ_{MAX} and τ_{MIN} . The monthly ticks along the abscissa indicate the first day of the month here and elsewhere in this paper.

A Time lag of maximum temperature-- τ_{MAX}



B Time lag of minimum temperature-- τ_{MIN}



C Asymmetry of Seasonal Time lag (A-B)

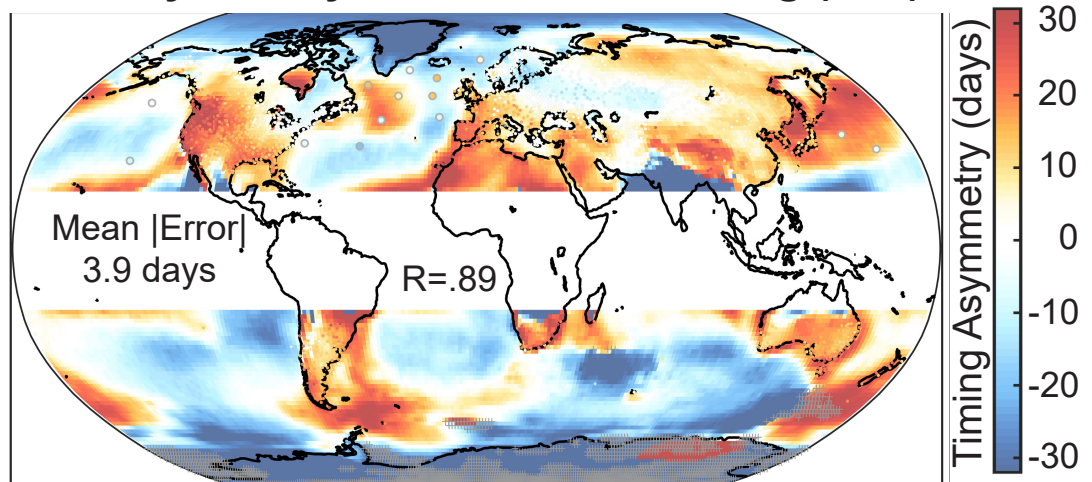


FIG. 2: (a) Time lag of maximum temperature relative to the summer solstice (τ_{MAX}). Colors show results from the gridded ERA T_{2m} data and the individual dots show the location of the station data color coded by the time lag given by the common colorbar in the upper right. The gray stippling indicates ERA gridpoints with more than one temporal minima/maxima in the smoothed T_{2m} timeseries and the gray crosses through a station indicate the same for station data. (b) As in (a), but for the time lag of the minimum temperature relative to the winter solstice (τ_{MIN}). (c) The asymmetry (ASYM) in the timing of the seasons defined as the top panel minus the middle panel. This panel uses the colorbar on the lower right. The spatial correlation (R) between values at all stations and that at the nearest gridpoint is shown in the middle of each plot along with the mean absolute difference.

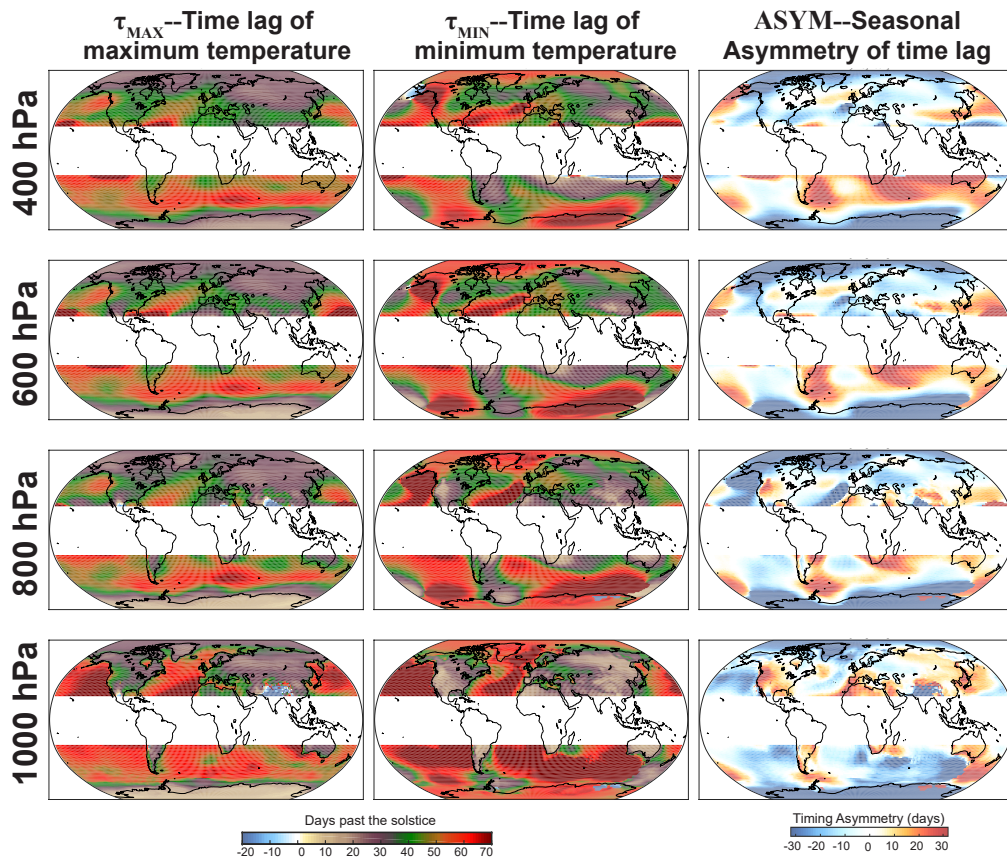


FIG. 3: Vertical structure of the seasonal timing of atmospheric temperature at various pressure levels in the ERA reanalysis. Time lag of the seasonal maximum temperature (τ_{MAX} , left column), minimum temperature (τ_{MIN} , left column) and, timing asymmetry (ASYM, right column) defined as the left column minus the right column. Each row is a different pressure level with the surface (1000 hPa) at the bottom to 800, 600 and, 400 hPa (top row).

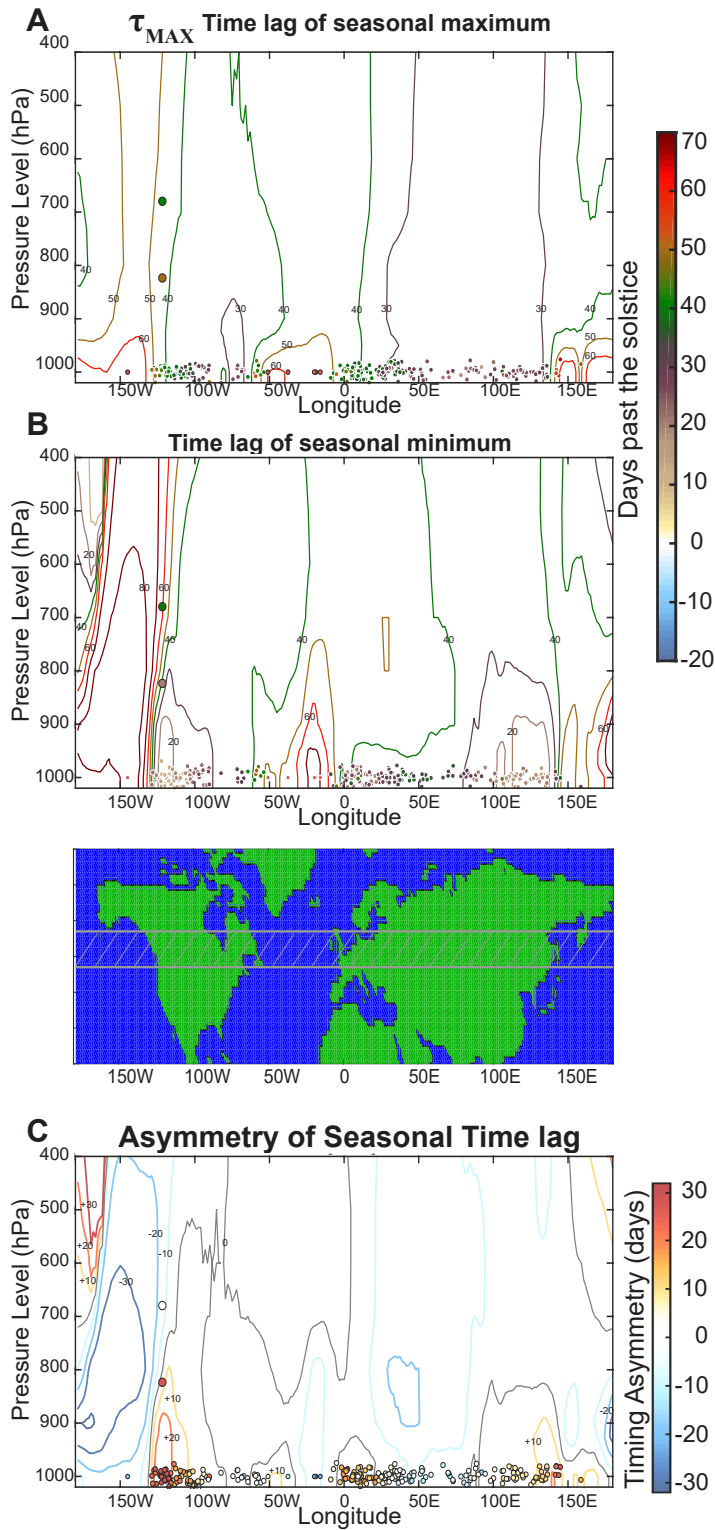


FIG. 4: Longitude/altitude cross sections of seasonal timing averaged between 47°N and 57°N . (a) The time lag of the seasonal maximum relative to the summer solstice; (b) the time lag of the seasonal minimum relative to the winter solstice; and (c) the seasonal asymmetry in the timing defined as the difference between (a) and (b). The contours show the results from the ERA reanalysis with contour interval 10 days with color values shown in the color bar (zero contour in gray). The color filled circles show results from surface stations and Ocean Weather Ships (jittered about 1000 hPa for visual purposes) and the dots at higher elevations show the high altitude stations on Mt. Rainier, WA, USA.

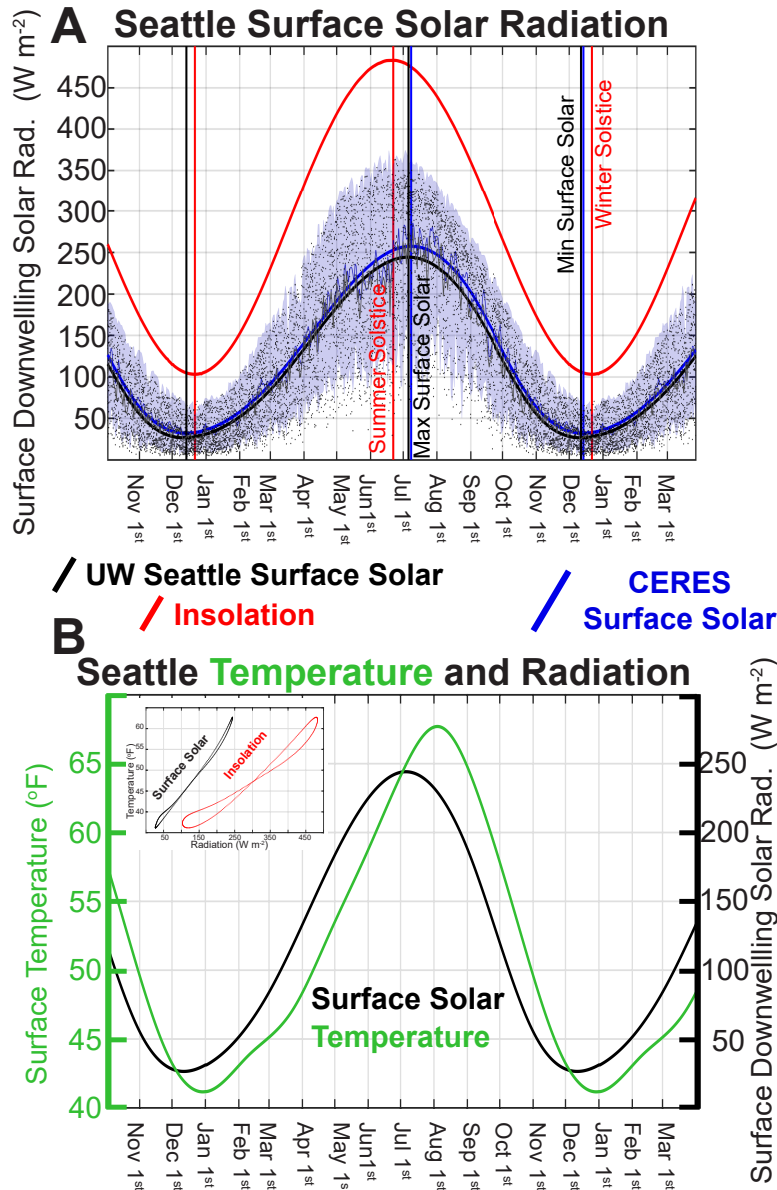
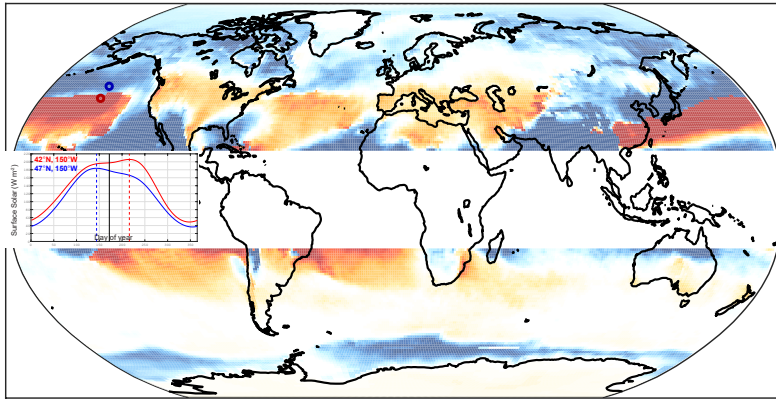
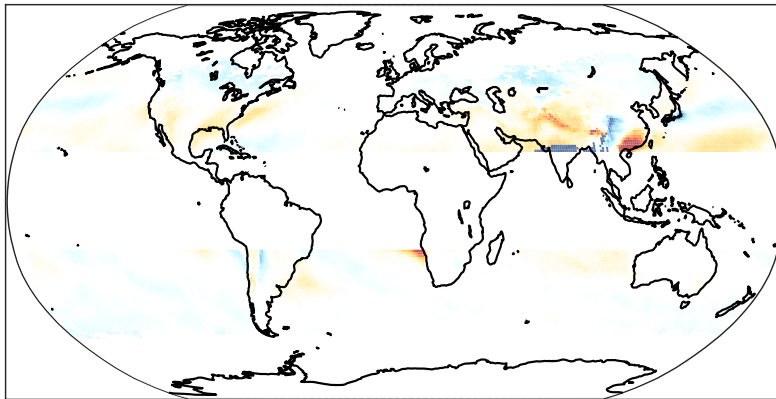


FIG. 5: (a) Seasonal time series of the downwelling surface solar radiation (black) measured at the University of Washington (UW ATG), Seattle and at (blue) the nearest gridpoint of the CERES data. Black dots show the individual daily mean values at UW ATG, the thin black line is the climatological mean for the calendar day and the thick black line is the spline fit. The shaded blue area is bounded by the maximum and minimum CERES values for the calendar day over the record (2000-2018), the thin blue line is the climatological mean for the calendar day and the thick blue line is the spline fit. The red line shows the (top of atmosphere downward) insolation from CERES. Vertical black and blue lines show the timing of the (spline fit) maximum and minimum surface solar radiation in the CERES and UW ATG data respectively and the vertical red lines show the summer and winter solstice. (b). Seasonal time series of the spline fit (black) UW ATG surface solar radiation and (green) T_{2m} at SEATAC. The inset shows the relationship between T_{2m} at SEATAC and (black) surface solar radiation at an optimal lead of 25 days and (red) top of atmosphere insolation at an optimal lead of 32 days.

A Timing of maximum surface solar



B Timing of minimum surface solar



C Seasonal Asymmetry of Surface Solar

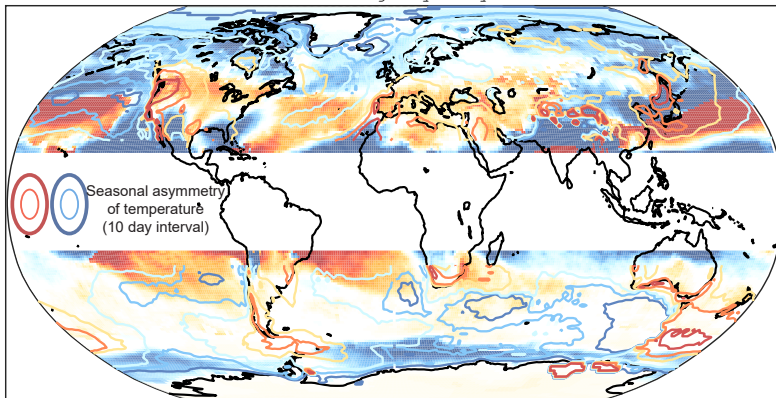


FIG. 6: (a) Timing of maximum in surface solar radiation relative to the summer solstice from the CERES dataset. (b) Timing of seasonal minimum of surface solar radiation relative to the winter solstice. (c) Seasonal asymmetry of surface solar radiation defined as the difference between the top and middle panel. The contours show the seasonal asymmetry of surface temperature (ASYM) from the ERA surface data (repeated from Figure 3) with contour interval of 10 days using the same colorbar as the surface solar timing (zero contour is omitted). In panel (a), the inset shows the time series of downwelling surface solar radiation at two gridpoints straddling the strong north-south gradient in the North Pacific (marked by circles on the map) with timing of the maximum shown by the vertical lines.

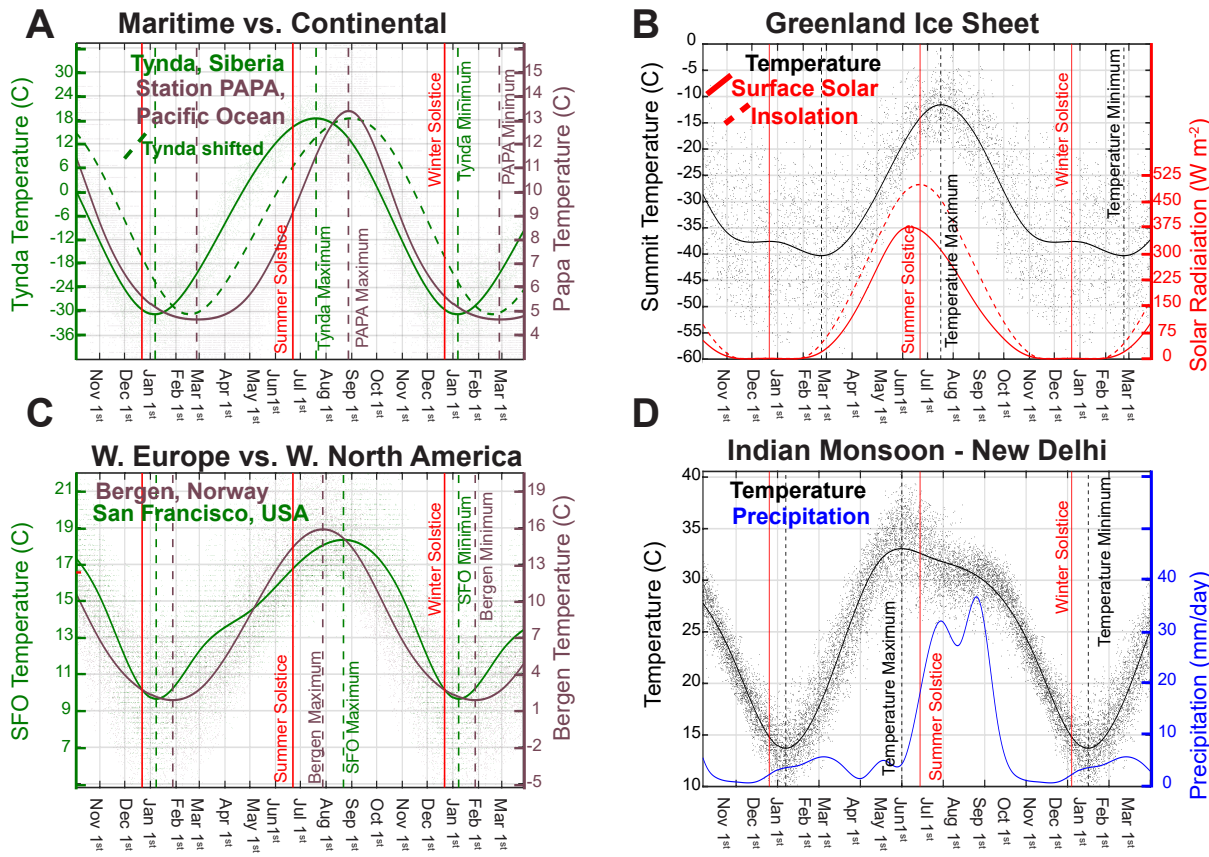


FIG. 7: (a) Comparison of the seasonal cycle of surface temperature at a typical continental location (Tynda, Siberia – green) and a maritime location (Weather Station Papa – purple). The maritime record has been rescaled by a factor of 6 (c.f. the left, green and right, purple y-axes) so that the seasonal cycles are visible on the same axis. Dots show the daily data and the lines show the spline fit to the data; vertical lines show the timing of the extrema. The dashed green line shows the continental record lagged by 42 days (such that the maximum are in phase) to highlight the different shapes of the seasonal cycle over land and ocean. (b) Seasonal times series of temperature (black) and solar radiation at the surface (dashed red) and insolation at TOA (solid red) at Greenland Summit Station. (c) As in panel B except a comparison on maritime climates in western North America (San Francisco airport, USA, green) and western Europe (Bergen, Norway, purple). (d) Seasonal time series of temperature (black) and precipitation (blue) at New Delhi, India.

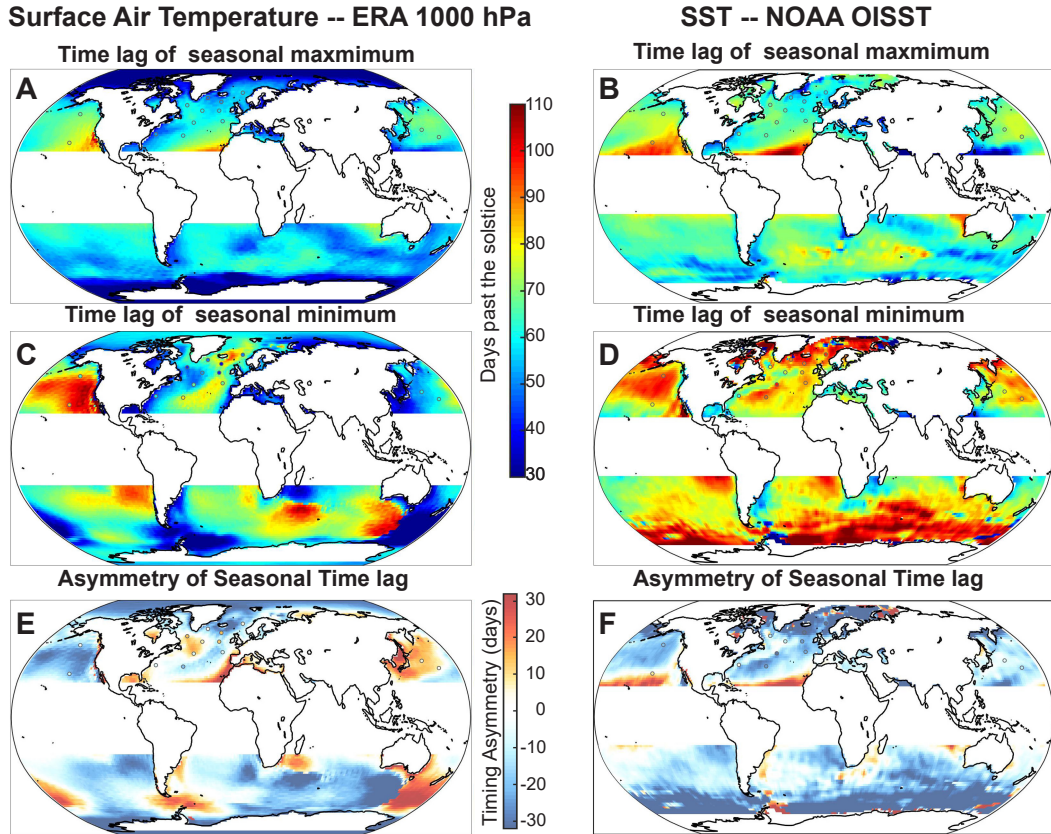


FIG. 8: Comparison of seasonal timing of T_{2m} and SST over the oceans. The left panels show τ_{MAX} , τ_{MIN} and ASYM defined from surface air temperature (from ERA 1000 hPa) and the right panels show the same quantities calculated from SST data (NOAA OISST). The dots show the ocean weather ship locations and are color coded by the timing of the station measured T_{2m} in the left panel and SST in the right panel.

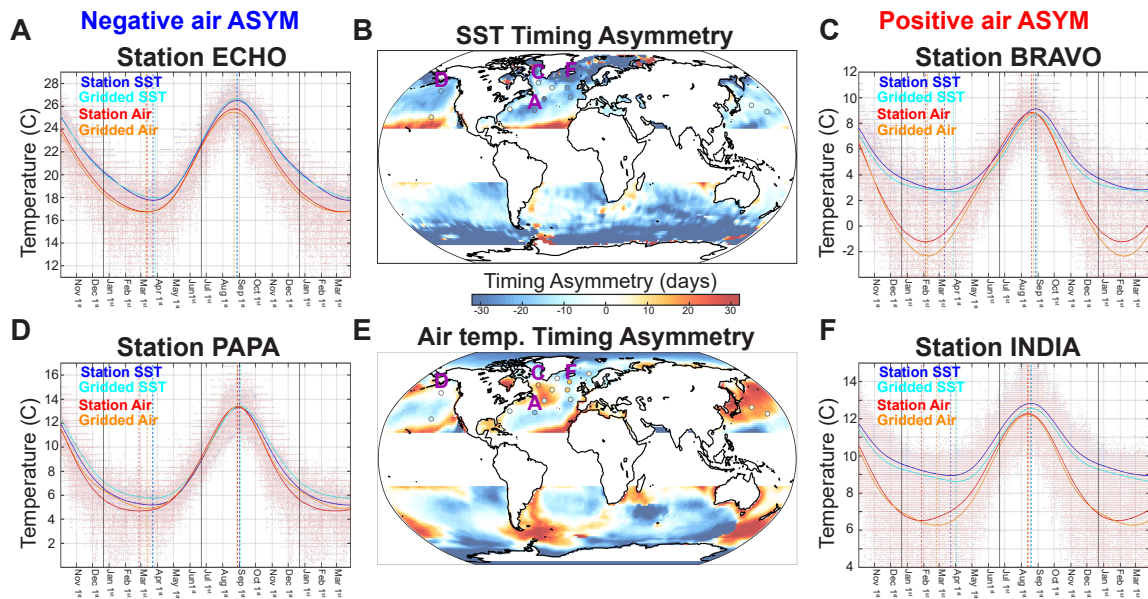


FIG. 9: (a,c,d,f) Time series of smoothed seasonal cycles of (blue) SST and (red) surface air temperature (T_{2m}) at selected OWS locations. The cyan line shows the NOAA OISST data and the nearest neighbor gridpoint. The orange line shows the ERA reanalysis T_{2m} data at the nearest neighbor gridpoint. The stations are organized by those with negative (positive) ASYM in T_{2m} on the left (right) and those where the timing is well (poorly) replicated in the gridded data on the top (bottom). The maps of ASYM defined from the SST data (NOAA OISST) and the ERA surface air temperature data respectively (previously shown in Fig. 8) are shown in panels b and e with the highlighted stations labeled by the letter of the panel they appear in (just above the station location) to orient the reader.

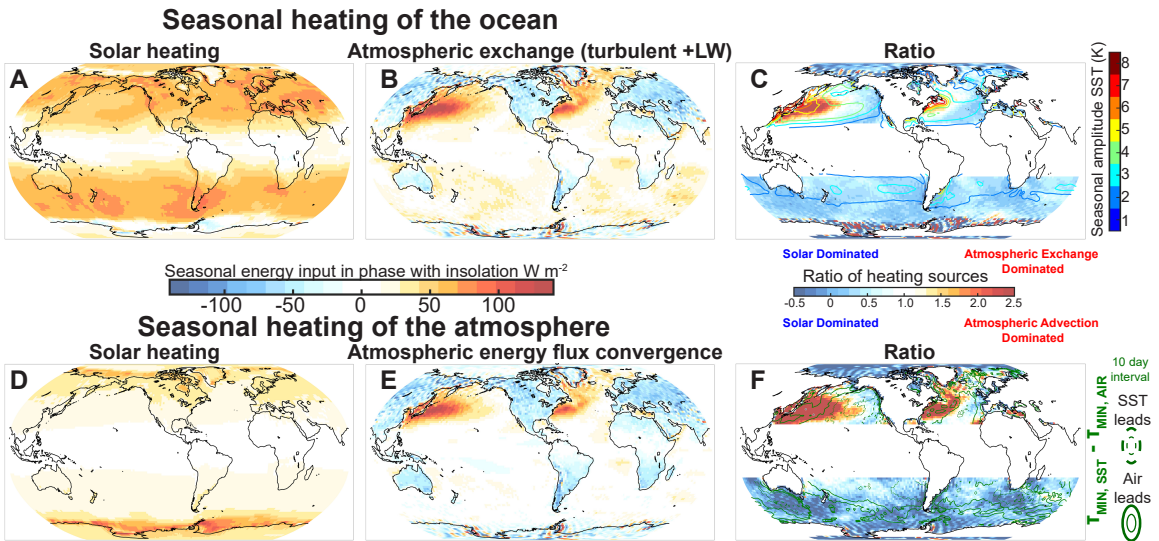


FIG. 10: The source of seasonal heating of the (top) ocean and (bottom) atmosphere defined as the seasonal amplitude of energy input phase with the insolation. The oceanic energy input is broken down into (a) surface solar radiation and (b) energy exchange with the atmosphere. Panel (c) shows the ratio (atmospheric exchange divided by solar heating); the seasonal amplitude of SST is overlaid in contours. The atmospheric energy input is broken down into (d) solar radiation absorbed in the atmospheric column and (e) atmospheric energy flux convergence (advection). Panel (f) shows the ratio (advection to solar); the phase advance of the atmospheric minimum temperature relative to the SST minimum is shown in green contours.

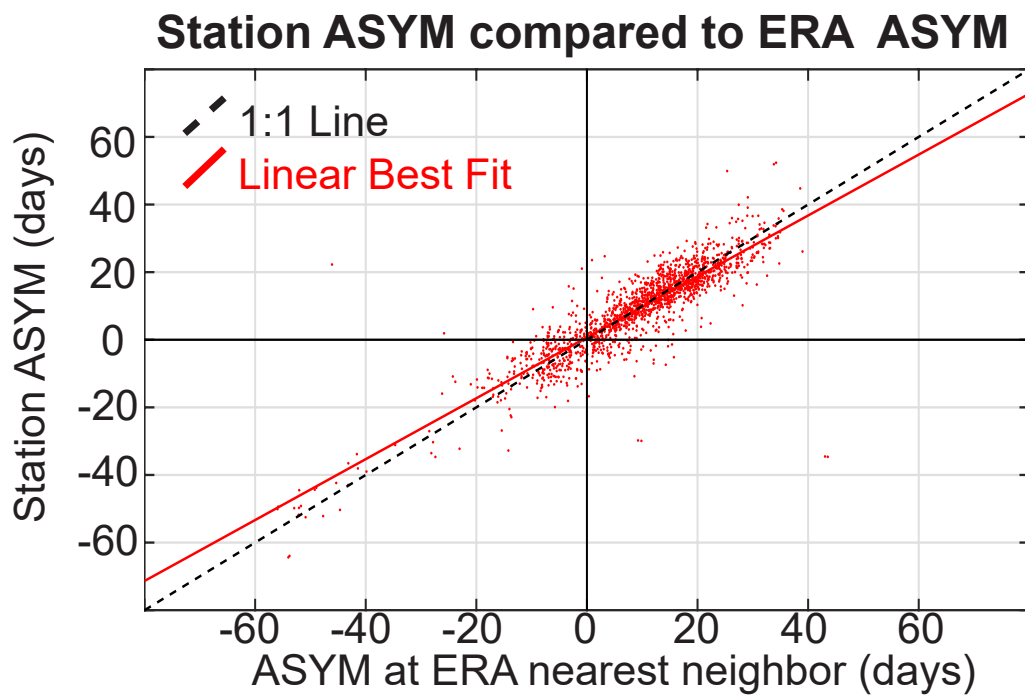
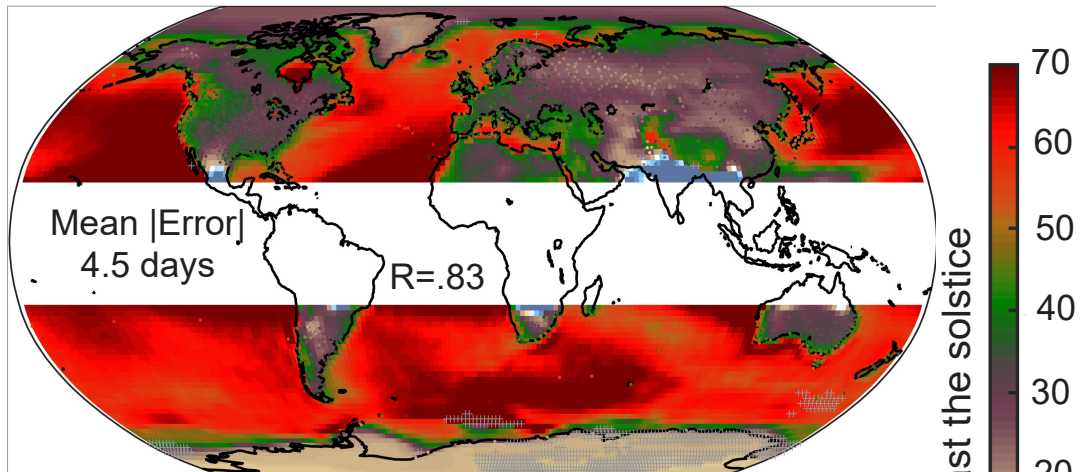
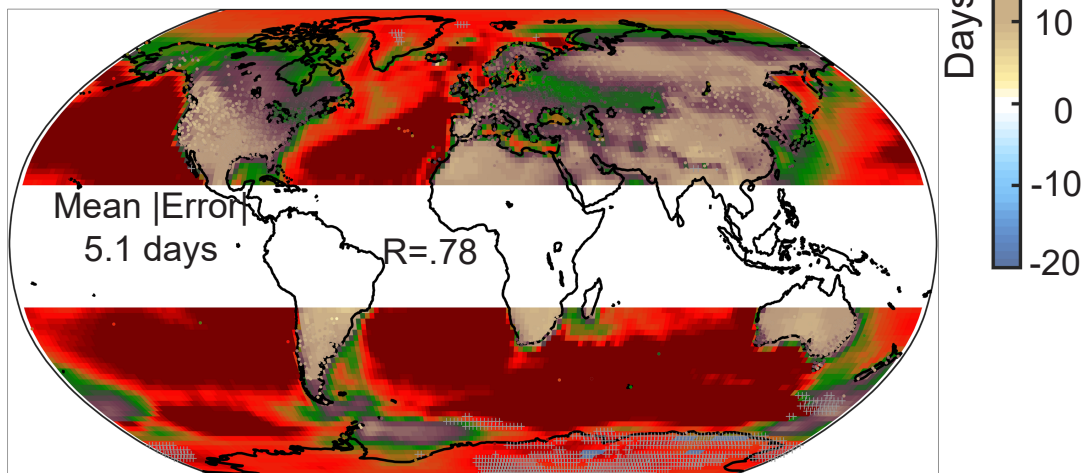


Fig. A1: Scatter plot of (ordinate) ASYM calculated from the station data versus (abscissa) the ASYM calculated from the ERA T_{2m} at the nearest grid point. The dashed black line is the 1:1 line and the red line is the linear best fit.

A Time lag of seasonal maximum



B Time lag of seasonal minimum



C Asymmetry of Seasonal Time lag (A-B)

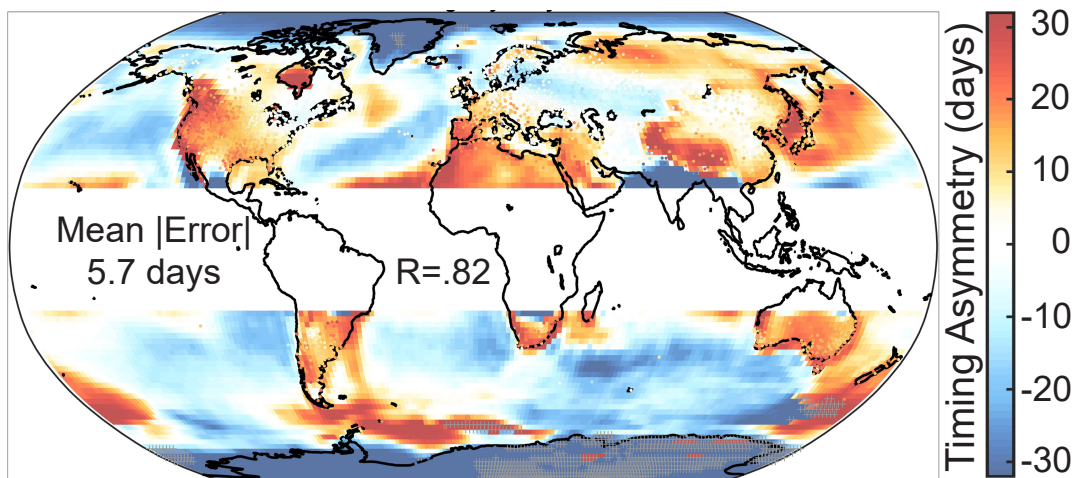


Fig. A2: As in Fig. 2 except for the NCEP reanalysis. (a) Time lag of seasonal maximum temperature relative to the summer solstice. Colors show results from the gridded 2 meter air temperature from the NCEP reanalysis and the individual dots show the location of the station data color coded by the time lag given by the common colorbar to the right. (b) As in (A) except for the time lag of the minimum temperature relative to the winter solstice. (c) The asymmetry in the timing of the seasons (ASYM), defined as (a) minus (b). This panel uses the colorbar on the lower right. The gray stippling indicates regions with more than one temporal maxima/minima in the smoothed seasonal cycle of T_{2m} .

Seattle and Mt. Rainier Seasonal Cycle Temperature

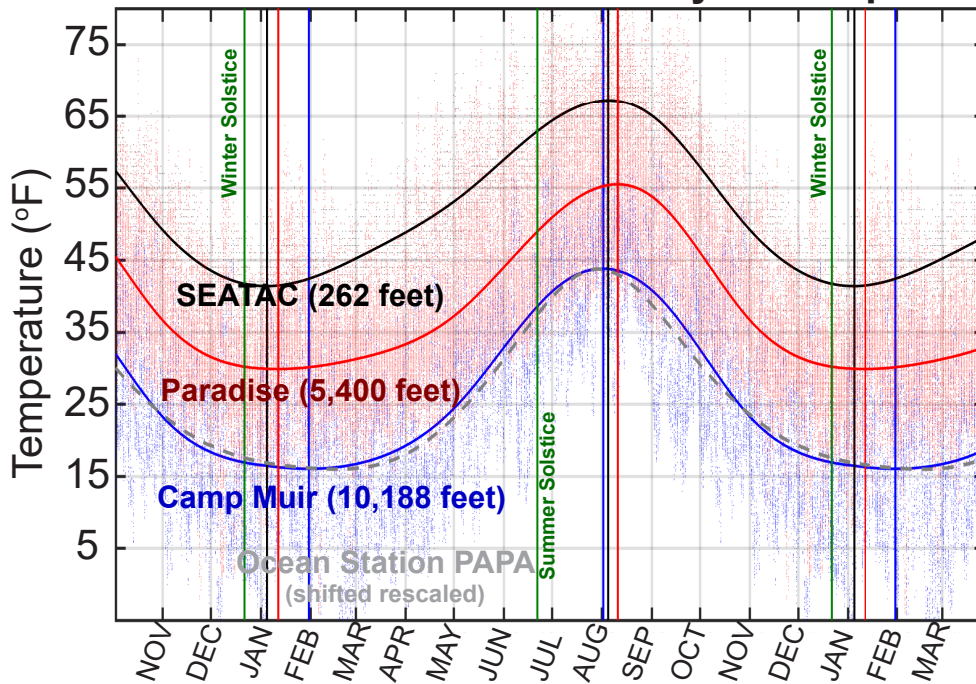


Fig. A3: Comparison of seasonal cycle of temperature on Mount Rainier at (red) Paradise (elevation 5,400 feet) and (blue) Camp Muir (elevation 10,188 feet) with that near sea level at (black) SEATAC. The gray dashed line is the seasonal cycle of SST at ocean station PAPA shifted and rescaled for visual comparison. The vertical lines indicate the timing of τ_{MAX} and τ_{MIN} in each time series given by the same color.

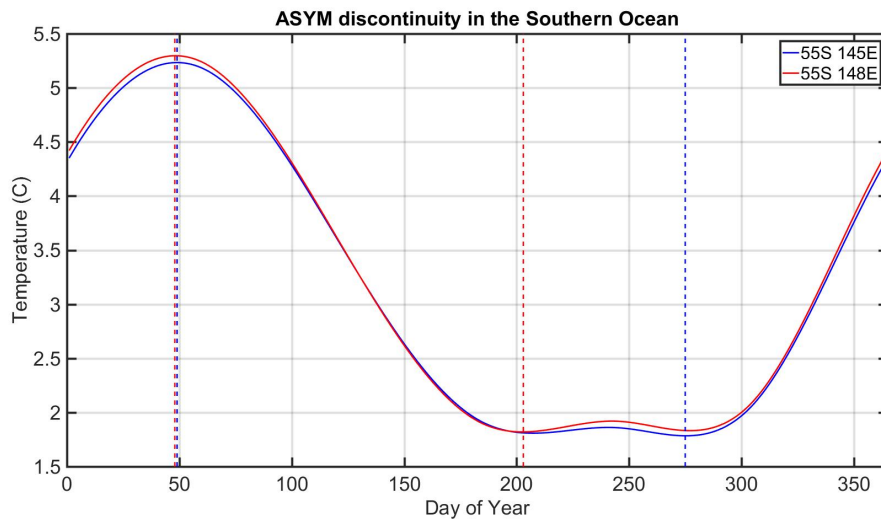


Fig. A4: Time series of ERA surface air temperature at gridpoints separated by 3° longitude spanning the discontinuity in ASYM south of Australia seen in Fig. 2c at 55°S . The blue time series is from 55°S , 145°E with $\text{ASYM} = -44$ days and the red time series is from 55°S , 148°E with $\text{ASYM} = +27$ days. The dashed vertical lines show the timing of τ_{MIN} and τ_{MAX} for the time series of the same color.

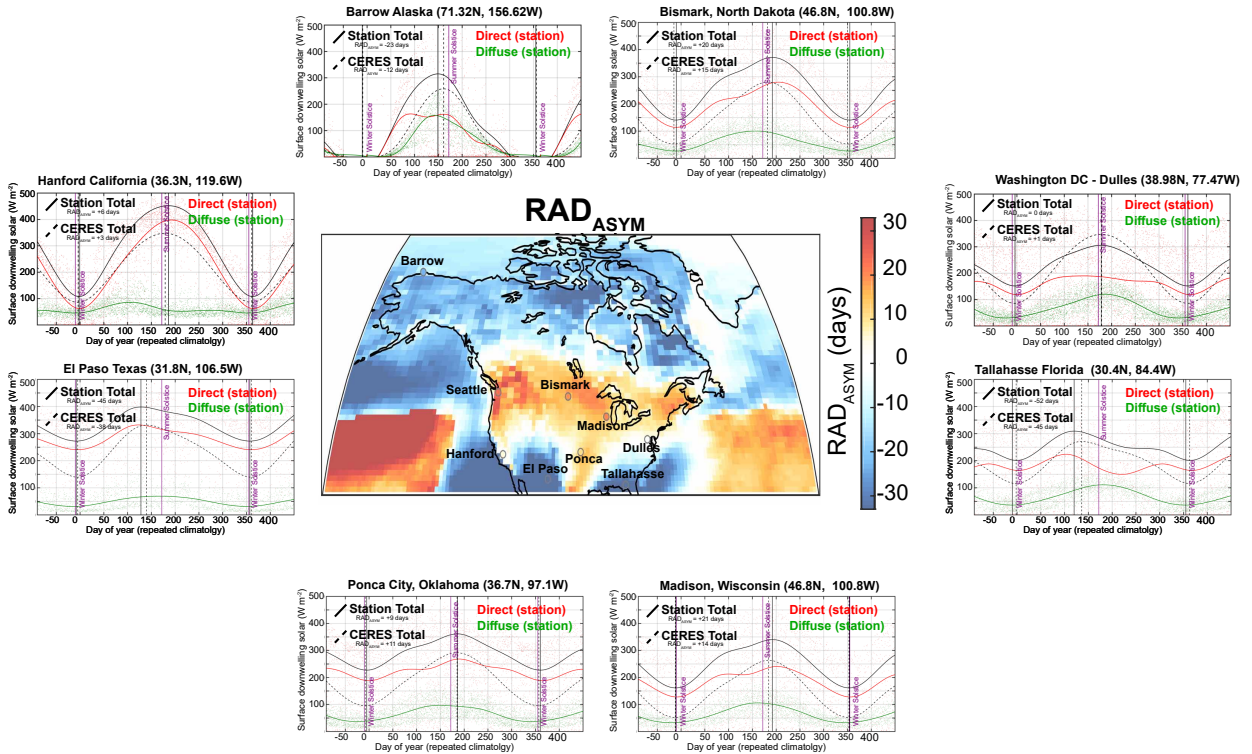
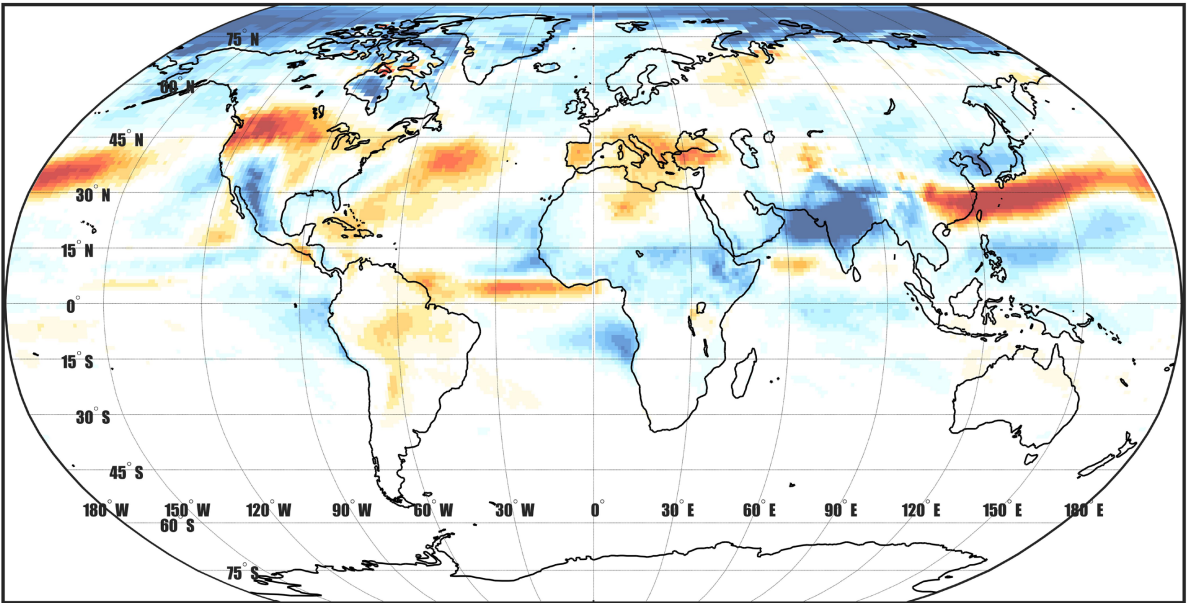


Fig. A5: Comparison of RAD_{MAX} , RAD_{MIN} and RAD_{ASYM} calculated in the gridded CERES data to that calculated using DSR measurements from the US (Sengupta et al. 2018). The dots show the observed daily mean DSR and the solid lines show the smoothed climatological seasonal cycle. The direct radiation is shown in red, the diffuse in green and the total observed DSR in black. The dashed black line shows the smoothed seasonal cycle of total DSR from CERES data at the nearest neighbor gridbox. Vertical lines show the time of RAD_{MAX} and RAD_{MIN} identified in the smoothed timeseries of the same linetype. The purple vertical lines show the winter and summer solstices. The inset map shows the RAD_{ASYM} values at each of the stations (colored dots circumscribed by gray) overlaid on (colors) RAD_{ASYM} calculated from the CERES DSR data.



July minus June surface
downwelling solar radiation (W m^{-2})

-60 -40 -20 0 20 40 60

Fig. A6: Change in surface solar radiation associated with Monsoon onset. Map of difference in temporally smoothed (spline with 10 knots) downwelling solar radiation at the surface between after summer solstice (July 10) and 20 days prior to summer solstice (June 1st) from CERES data.