

1 **The effect of atmospheric transmissivity on model and observational**
2 **estimates of the sea ice albedo feedback?**

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ABSTRACT

15 The sea-ice-albedo feedback (SIAF) is the product of the ice sensitivity
16 (IS) – how much the surface albedo in sea-ice regions changes as the planet
17 warms– and the radiative sensitivity (RS) – how much the top of atmosphere
18 radiation changes as the surface albedo changes. We demonstrate that the RS
19 calculated from radiative kernels in climate models is reproduced from calcu-
20 lations using the “approximate partial radiative perturbation” method that uses
21 the climatological radiative fluxes at the top of atmosphere and the assumption
22 that the atmosphere is isotropic to shortwave radiation. This method facilitates
23 the comparison of RS from satellite-based estimates of climatological radia-
24 tive fluxes with RS estimates across a full suite of coupled climate models
25 and, thus, allows model evaluation of a quantity important in characterizing
26 the climate impact of sea ice concentration changes. The satellite based RS
27 is within the model range of RS that differs by a factor of two across climate
28 models in both the Arctic and Southern Ocean. Observed trends in Arctic
29 sea ice are used to estimate IS which, in conjunction with the satellite-based
30 RS yields an SIAF of $0.16 \pm 0.04 \text{ W m}^{-2} \text{ K}^{-1}$. This Arctic SIAF estimate
31 suggests a modest amplification of future global surface temperature change
32 by approximately 14% relative to a climate system with no SIAF. We calcu-
33 late the global albedo feedback (GAF), including changes in snow cover over
34 land, in climate models using model specific RS and IS and find a model mean
35 feedback parameter of $0.37 \text{ W m}^{-2} \text{ K}^{-1}$ which is 40% larger than the IPCC
36 AR5 estimate based on using RS calculated from radiative kernel calculations
37 in a single climate model.

38 1. Introduction

39 Sea ice area is expected to decrease as the climate system warms, and this in turn will lead to
40 a darker surface, and increase in solar radiation absorbed by the climate system. This additional
41 radiative input reinforces the initial warming providing a positive climate feedback often termed
42 the sea-ice-albedo feedback (SIAF). Early literature on climate stability in simplified models sug-
43 gested that SIAF could cause abrupt and dramatic climate state transitions under smoothly varying
44 external forcing (North 1984; Budyko 1969) or produce multiple equilibria in more comprehensive
45 coupled climate models (Ferreira et al. 2011). More modest estimates of the global albedo feed-
46 back (including changes associated in surface albedo over land) were found in coupled climate
47 models (Stocker et al. 2013; Bony et al. 2006; Soden and Held 2006a), producing an IPCC AR5
48 ensemble mean global SIAF of $0.26 \text{ W m}^{-2}\text{K}^{-1}$ (Flato et al. 2013) leading to a 22% increase
49 in the global climate response to external forcing (Roe 2009) relative to system with no surface
50 albedo feedback. Pistone et al. (2014, 2019) used the co-variance of year-to-year sea ice anoma-
51 lies and satellite radiation to produce an observationally based estimate of SIAF with a similar
52 magnitude for the Arctic sea ice ($0.31 \text{ W m}^{-2}\text{K}^{-1}$) and pointed out this additional radiative input
53 to the climate system due to Arctic ice melt to date 25% the anthropogenic forcing. There is still a
54 substantial ($\pm 0.1 \text{ W m}^{-2}\text{K}^{-1}$) inter-model spread in strength of the SIAF (Yu et al. 2006; Hall and
55 Qu 2006) that is understood to be the leading cause of inter-model differences in the high latitude
56 climate response (polar amplification) (Hall 2004; Kay et al. 2012) (Holland and Bitz 2003).

57 SIAF measures how much additional radiative energy the Earth system gains due to sea ice
58 loss as the planet warms, which amplifies the warming relative to a system with no SIAF. SIAF
59 is quantified as the global (area weighted) average of $\text{RI}_{TOA,\alpha}$, the Radiative Impact of sea ice

60 change (the local TOA radiative flux change due to surface albedo changes (α) from sea ice loss
 61 per degree of global averaged surface temperature change):

$$SIAF = [RI_{TOA,\alpha}(x,y)] \quad (1)$$

62 where $[\]$ brackets indicate a global average. Following Yu et al. (2006, Eq. 1), the spatial map of
 63 $RI_{TOA,\alpha}(x,y)$ is the product of two quantities (Soden and Held 2006b; Shell et al. 2008): 1.) the
 64 surface albedo change due to sea ice loss per unit of global mean surface temperature change, $[dT_S]$,
 65 ($\frac{d\alpha_{SI}}{[dT_S]}$), and 2.) the sensitivity of top of atmosphere (TOA) radiation to surface albedo ($\frac{\partial RAD_{TOA}}{\partial \alpha}$)
 66 that we hereafter refer to as radiative sensitivity (RS) :

$$RI_{TOA,\alpha}(x,y) = \underbrace{\frac{d\alpha_{SI}}{[dT_S]}}_{IS(x,y)} \underbrace{\frac{\partial RAD_{TOA}(x,y)}{\partial \alpha(x,y)}}_{RS(x,y)}. \quad (2)$$

67 The normalization of $RI_{TOA,\alpha}(x,y)$ by global mean temperature (T_S) change is integrated into the
 68 IS term and RS is defined as the local radiative change at the TOA per unit of surface albedo
 69 change. This study considers only the radiative impact of α changes in high latitudes (poleward of
 70 60°N and 55°S , in the northern (NH) and Southern (SH) hemispheres respectively) over oceans,
 71 and calculations of SIAF exclude the impact of changes in terrestrial snow cover. **RS and α**
 72 **changes are calculated for each month and then their product is time averaged.** Changes in α_{SI}
 73 are calculated over the ocean and capture both the impact of sea ice loss and changes in surface
 74 albedo over sea ice (i.e. snow and melt ponds). Hall and Qu (2006) **claim** that RS varies very
 75 little between climate models. As a result, much of the literature on SIAF uncertainty has focused
 76 on processes controlling sea ice albedo changes (Horvat et al. 2019) and the sensitivity of **sea ice**
 77 **concentration (SIC)** to warming (Yu et al. 2006; Qu and Hall 2005; Curry et al. 1994) which both
 78 vary substantially between models. The IPCC estimate of SIAF (Flato et al. 2013; Soden and Held

79 2006b) used a RS calculated from a single model, neglecting inter-model differences and biases
80 (relative to observations) and assuming RS does not contribute to SIAF uncertainty. We assess the
81 validity of this assumption below.

82 RS depends primarily on cloud reflectivity; clouds impede the amount of downwelling solar
83 radiation reaching the surface and also reduce the amount of solar radiation reflected by the surface
84 from reaching the TOA (Taylor et al. 2007; Donohoe and Battisti 2011) leading to a quadratic
85 dependence of RS on cloud reflectivity. High latitude cloud properties vary substantially between
86 models and exhibit many biases relative to observations (Gorodetskaya et al. 2006; Vavrus et al.
87 2009; Trenberth and Fasullo 2010). Cloud differences can contribute to model differences in RS
88 that in turn influence: (i) the sensitivity of sea ice loss to future warming (Hwang et al. 2011) via
89 local positive radiative feedbacks and (ii) the impact of sea ice loss on the global energy budget
90 and, thus, the global climate sensitivity to external forcing.

91 This study assesses inter-model differences in RS and consistency compared to estimates from
92 satellite observations. We also identify relative contributions of IS and RS to model spread and
93 biases (relative to observations) in the amplification of global warming by SIAF, and evaluate the
94 impact of using RS from a single climate model to calculate SIAF across models as was done in
95 Soden and Held (2006b) and the IPCC AR5 estimate of surface albedo feedback.

96 The manuscript is organized as follows: section 2, outlines how a simplified isotropic model
97 often discussed in textbooks on radiative transfer, and further developed by Taylor et al. (2007)
98 can be used to calculate RS from standard climate model output and demonstrates that the method
99 reproduces results from more computationally demanding radiative kernel techniques. This fa-
100 cilitates further evaluation of inter-model spread in RS in the coupled models participating in the
101 Coupled Model Inter-comparison Project (CMIP3 and CMIP5 Meehl et al. 2007; Taylor et al.
102 2012). Most importantly, this method also provides an observational estimate of RS from satellite

103 data (Section 3). These estimates of RS along with the sea ice response over the historical period
104 are used to calculate an observational SIAF (Section 4). The observational SIAF is compared to
105 that in model simulations under historical forcing and 4XCO₂ and the model spread and biases are
106 decomposed into contributions from RS and IS (Section 5). A summary and discussion follows.

107 **2. The impact of surface albedo changes on TOA radiation in radiative kernels and a simpli-** 108 **fied model**

109 *a. Radiative kernels*

110 The impact of surface albedo changes on TOA radiation (RS) has been rigorously calculated
111 using radiative kernel techniques in a small number of climate models (Smith et al. 2018; Pen-
112 dergrass and Vitt 2018; Shell et al. 2008; Block and Mauritsen 2013; Soden and Held 2006b;
113 Previdi 2010). RS can be calculated directly from offline radiative model calculations by pre-
114 scribing changes to the surface albedo (α) at each grid point and then running the radiative code
115 with all other fields unchanged – a technique referred to as a radiative kernel calculation (Soden
116 and Held 2006b; Shell et al. 2008). Radiative kernels are generally calculated **at each gridpoint**
117 **over a global domain** by perturbing the surface albedo at each grid point by a specified amount
118 (independent of whether that surface albedo change is feasible) using atmospheric models with
119 prescribed historical climatological (seasonally varying) sea surface temperatures. We use kernel
120 calculations (for specific models) provided by: 1.) Karen Shell, NCAR CAM3 (Shell et al. 2008),
121 2.) Karoline Block, MPI ECHAM6 (Block and Mauritsen 2013), 3.) Angie Pendergrass, NCAR
122 CAM5 (Pendergrass and Vitt 2018), 4.) Chris Smith, UKMO HadGEM2 (Smith et al. 2018), 5.)
123 Brian Soden, GFDL AM2p12b (Soden and Held 2006b).

124 RS is reported in $\text{W m}^{-2} \%^{-1}$ where the $\%$ refers to a 0.01 unit change in surface albedo
125 (independent of the climatological surface albedo). Summertime (MJJA) daily-averaged TOA
126 insolation in the Arctic (defined as the region poleward of 60°N) is of order 420 W m^{-2} , and a
127 $4.2 \text{ W m}^{-2} \%^{-1}$ RS would be expected in a completely transparent atmosphere. Radiative kernel
128 calculations produce an RS Arctic average of $1.63 \text{ W m}^{-2} \%^{-1}$ across the 4 different models
129 (numbers in the upper right of each panel in Fig. 1) indicating that the atmosphere attenuates
130 the surface contribution to reflected radiation at the TOA by a factor of ~ 2.6 ($4.2/1.63$). Kernel
131 estimates of RS in Arctic summer (May-June-July-August) are largest over Greenland ($2\text{-}3.5 \text{ W}$
132 $\text{m}^{-2} \%^{-1}$), smallest in the Greenland Iceland Norwegian (GIN) Seas ($0.5\text{-}1 \text{ W m}^{-2} \%^{-1}$) with
133 intermediate values in the central Arctic ($1\text{-}2.5 \text{ W m}^{-2} \%^{-1}$ – Upper panels of Fig. 1). This
134 spatial structure primarily reflects the climatological pattern of solar radiation reaching the surface
135 in the Arctic (Lindsay et al. 2014). Highest RS values are found where cloud cover and water
136 vapor are low over the high topography of Greenland. Moderate RS values are seen in the central
137 Arctic due to the thin but persistent cloud cover over the perennial sea ice. RS is smallest in the
138 GIN seas due to abundant thick clouds.

139 There is remarkable inter-model spread in Arctic RS across the different radiative kernel calcu-
140 lations, especially over the central Arctic where the models differ by a factor of **two**. As shown
141 below, the diversity of RS across the different kernel calculations is a consequence of inter-model
142 differences in the mean state cloudiness and *not* due to differences in radiative transfer code or the
143 methodology used to calculate the kernels between the different groups.

144 In the SO, RS during the Austral summer (NDJF) calculated from radiative kernels shows a
145 zonally annular structure in all models with smaller values over the cloudy storm track region
146 equatorward of the ice edge and larger values over the sea ice (upper panels of Fig. 2). However,
147 the models differ to first order on the magnitude of RS over the open ocean and on the location

148 and aerial extent of the region of larger RS adjacent to the Antarctic continent. In HADGEM2,
149 the value of RS over the open ocean is $2 \text{ W m}^{-2} \%^{-1}$ whereas in NCAR CAM3 RS is 1 W m^{-2}
150 $\%^{-1}$ over the same region. In NCAR CAM5, the region of high RS adjacent to the Antarctic coast
151 extends substantially into the SO whereas in NCAR CAM3 and ECHAM6 the high RS region
152 is confined to the coast itself with the exception of the Weddel and Ross Seas. The inter-model
153 differences in the aerial extent of the high RS region roughly correspond to inter-model biases in
154 summertime ice extent; the gradient in atmospheric transmissivity is linked to the sea ice edge via
155 cloud coverage and atmospheric water content although in some models the gradient in cloudiness
156 is significantly poleward of the ice edge (i.e. NCAR CAM3) while in other models the cloud
157 gradient is co-located with the ice-edge (i.e. NCAR CAM5). Overall, the SO domain average RS
158 (excluding the Antarctic continent – to focus on the sea ice) ranges from 1.29 to $1.75 \text{ W m}^{-2} \%^{-1}$
159 (as shown by the values in the upper right corner of Fig. 2).

160 *b. Isotropic single layer model*

161 Taylor et al. (2007) – hereafter T07– developed a model (hereafter the isotropic model) that can
162 be used for approximating RS from the climatological radiative fluxes at the TOA and surface
163 and some basic assumptions about shortwave radiative transfer in the atmosphere. Part of the T07
164 derivation is repeated here for clarity with a few modifications to variable names. Of the incident
165 shortwave radiation at the TOA (S), assume a fraction (A) is absorbed in the atmosphere above
166 cloud top and a fraction R of the radiation incident on cloud top is reflected back to space (Fig.
167 3). This resultant downwelling radiation at the surface is $S(1-A)(1-R)$. A fraction (α , equal to
168 the surface albedo) of this downwelling radiation is reflected upwards. Of this surface upwelling
169 radiation, R is reflected back (downward) to surface with the remainder ($S[1-A][1-R]^2$) transmit-
170 ted to space. Reflections and transmissions are continued indefinitely subject to the three primary

171 assumptions: (i) cloud optical properties can be represented by a single layer ,(ii) cloud reflection
 172 is isotropic – the same fraction (R) of broadband shortwave radiation incident on the cloud layer
 173 is reflected independent of the direction (upwelling/downwelling) and how many previous inter-
 174 actions with the surface and cloud occur and (ii) all of the atmospheric absorption occurs above
 175 cloud top on the first downward pass which is apt for describing SW absorption by ozone in the
 176 stratosphere (Chou and Lee 1996). We further analyze the limitations of these assumptions at the
 177 end of this subsection.

178 In the isotropic model, loss of shortwave radiative energy from the climate system due to surface
 179 albedo is a three step process: (i) insolation must be transmitted to the surface then (ii) reflected
 180 by the surface and finally (iii) transmitted from the surface to the TOA. Mathematically, upwelling
 181 SW radiation at the TOA that results from reflection off the surface is equal to the insolation
 182 (S) times the downwelling transmissivity ($[1 - A][1 - R]$) times the upwelling transmissivity (1-
 183 R). The isotropic model also includes higher order reflections where the SW radiation reflected
 184 at the surface is reflected back to the surface off clouds and thereafter will contribute additional
 185 upwelling SW fluxes at the TOA with each subsequent reflection equal to the value of the previous
 186 order contribution times αR . These terms form an infinite geometric series that converges to the
 187 expression:

$$SW \uparrow_{TOA} = \underbrace{SR(1-A)}_{SW \uparrow_{TOA,atmos}} + S\alpha \underbrace{\frac{(1-A)(1-R)^2}{1-\alpha R}}_{SW \uparrow_{TOA,surf}}, \quad (3)$$

188 where $SW \uparrow_{TOA,atmos}$ and $SW \uparrow_{TOA,surf}$ indicates the upwelling radiation at the TOA that was de-
 189 rived from atmospheric and surface reflection respectively. Thus, if the values of R and A along
 190 with α and S are known, the contribution of the surface to the SW flux at the TOA can be cal-
 191 culated. In our case, the isotropic model provides equations relating 3 satellite derived quantities

192 ($SW \uparrow_{TOA}$, $SW \uparrow_{SURF}$ and, $SW \downarrow_{SURF}$) in terms of 3 unknown variables (A, R, α) and the satellite
 193 measured S . The result is a determined set of 3 equations in terms of 3 variables. Thus, the clima-
 194 tological radiative fluxes allow the calculation of the single pass A and R for each climate model.
 195 We can then calculate the expected change of $SW \uparrow_{TOA}$ as α changes with all else being equal by
 196 taking the partial derivative of Eq. 3 with respect to α .

$$RS = \frac{\partial SW \uparrow_{TOA}}{\partial \alpha} = S \frac{(1-A)(1-R)^2}{1-\alpha R} \left(1 + \frac{R\alpha}{1-R\alpha} \right). \quad (4)$$

197 This provides an alternate method for calculating RS that relies only on readily available model
 198 output at monthly resolution that can also be compared with the RS calculated from radiative
 199 kernel techniques.

200 The lower panels of Fig. 1 show the RS in the Arctic summer calculated from Eq. 4 applied to
 201 the monthly climatological output from the same control simulations that were used to calculate
 202 the radiative kernels. The RS calculated from the isotropic model is in good agreement with that
 203 calculated from radiative kernels in terms of the spatial pattern of RS and inter-model differences.
 204 Spatial correlation between RS in the isotropic model and radiative kernel calculation for each
 205 model is high with an R^2 that exceeds 95% in all but NCAR CAM3. The inter-model differences
 206 in domain average of RS is within 10% in the absolute sense and captures the rank of RS in models
 207 (c.f. the adjacent upper and lower panels of Fig. 1 with R^2 listed in the middle). The isotropic
 208 model explains 94% of the variance in MJJA RS calculated from radiative kernels considered
 209 across models and over all Arctic grid points collectively with a root mean squared (RMS) error
 210 of $0.15 \text{ W m}^{-2} \%^{-1}$ (top panel of supplemental Fig. A2). As a basis for comparison, if one used
 211 the spatial pattern of MJJA RS calculated using radiative kernels from one model to predict the
 212 kernel based RS in a different model – as was done in the IPCC estimate of SIAF– the RS variance
 213 explained is 21% with a RMS error of $0.67 \text{ W m}^{-2} \%^{-1}$ (bottom panel of Fig. A2). Thus, the

214 isotropic model offers a factor of 4 improvement on the practice of applying RS calculations from
215 a single climate model.

216 The isotropic model also captures the spatial pattern and inter-model spread of the kernel cal-
217 culated RS in the SO (Fig. 2) although the absolute values of RS differs by as much as 20% (in
218 the HADGEM2 model). The isotropic model explains 96% of the variance in NDJF kernel RS
219 across models over the SO (top panel of supplemental Fig. A3) with a root mean squared (RMS)
220 error of $0.23 \text{ W m}^{-2} \%^{-1}$. When radiative kernels from one model are used to predict the kernel
221 based NDJF RS in a different model the variance explained is 71% with a RMS error of 0.47 W
222 $\text{m}^{-2} \%^{-1}$ (bottom panel of Fig. A3). Thus, the isotropic model offers a factor of 2 improvement
223 on the practice of applying RS calculations from a single climate model in the SO. These results
224 indicate that the isotropic model captures the essential SW radiative processes that determine the
225 RS of surface albedo changes, and that the inter-model spread in RS is determined by the climato-
226 logical cloud reflectivity which is adequately calculated from the modeled TOA and surface fluxes
227 according to Eq. 3.

228 The isotropic model tends to bias the RS high relative to the radiative kernel (c.f. the domain av-
229 erage values listed in the upper right of the map in the upper and lower panels of Figs. 1 and 2) and
230 we speculate this results from the simplifying assumption that atmospheric absorption only occurs
231 during the first pass as this allows more of the radiation reflected off the surface to be transmitted to
232 space than would occur if the atmosphere absorbed upwelling solar radiation. Alternative formula-
233 tions of similar isotropic models (Donohoe and Battisti 2011) assume the atmospheric absorption
234 occurs in the same layer as the cloud reflection and occurs on all passes through the atmosphere to
235 account for shortwave absorption by water vapor that occurs throughout the troposphere (Donohoe
236 and Battisti 2013). This model better matches the RS calculated by radiative kernels in the tropics
237 and mid-latitudes but substantially underestimates RS relative to the radiative kernel derived value

238 at high latitudes (Appendix Fig. A1). We speculate that in the dry Arctic, the atmospheric absorp-
239 tion is primarily by stratospheric ozone whereas in the lower latitudes water vapor also contributes.
240 For this reason, we choose to assume the absorption occurs only on the downward pass and return
241 to possible impacts and improvements of this method in the discussion section.

242 *c. Causes of inter-model spread in RS*

243 What processes are responsible for the factor of 2 spread in modeled RS in Figs. 1 and 2? The
244 ability of the isotropic model to reproduce the kernel based RS calculated for each model demon-
245 strates that the mean state atmospheric opacity is the primary determinant. Generally speaking,
246 RS is determined by how much insolation is transmitted to the surface and thus how much impact
247 surface albedo changes have on reflected solar radiation. More specifically, RS is proportional
248 to the atmospheric transmissivity squared with higher order modifications due to the impact of
249 multiple reflections (Eq. 4). What then causes the inter-model spread in atmospheric opacity?

250 Clear sky surface albedo kernels (Fig. 4) have much larger magnitudes than their all-sky coun-
251 terparts. The very similar spatial structures and absolute values in the four models with available
252 kernel calculations have domain averages that differ by 2% from the multi-model mean, indicat-
253 ing that 1) clear-sky processes are not responsible for the inter-model spread in all-sky RS and
254 2) the different radiative transfer codes used in the climate models find a similar RS for a similar
255 (clear-sky) mean-state.

256 The atmospheric opacity parameters – reflectivity and absorptivity – **calculated by the** isotropic
257 model applied to the mean states of the different climate models are shown in Fig. 5. The all-sky
258 reflectivity is subdivided into a clear-sky and cloud component by applying the isotropic model
259 to the clear-sky mean state radiative fields (as in T07) to define a clear-sky reflectivity, and the
260 cloud reflectivity is then defined as the all-sky minus clear-sky reflectivity. All climate models

261 have very similar and nearly spatially uniform clear-sky reflectivity and all-sky absorptivity with
262 Arctic domain average absolute differences from the model mean of order 0.02 fractional units.
263 The slight spatial structure in clear-sky reflectivity and absorptivity is consistent between climate
264 models. Clear-sky reflectivity is larger near the North Pole consistent with enhanced Rayleigh
265 scattering due to the shallower angle of incidence with latitude. Absorptivity is smaller over the
266 thinner atmosphere above topography and drier continents consistent with reduced absorption by
267 water vapor. In contrast to the consistency of absorption and clear-sky reflection between models,
268 the cloud reflectivity differs substantially between models in both spatial structure and domain
269 average values (which differ between models by over 0.20 fractional units). In general, regions of
270 stronger cloud reflectivity have smaller RS values consistent with less downwelling solar radiation
271 at the surface in cloudy regions. However, the anti-correlation between the spatial variability in RS
272 and cloud reflectivity is significant but far from perfect ($R \approx -.60$) within a given climate model
273 due to the (comparable in magnitude) impact of the spatial structure of mean state albedo (Eq. 4)
274 on the multiple reflection contribution to RS. On a broader scale, the Arctic domain average cloud
275 reflectivity is very strongly anti-correlated ($R=-0.99$) with the domain average RS indicating that
276 Arctic averaged RS is primarily determined by the mean state cloud reflectivity.

277 **3. Observational estimate of radiative sensitivity to surface albedo changes and comparison** 278 **to coupled models**

279 Given the strong correspondence between RS calculated from radiative kernels and the isotropic
280 model (Figs. 1 and 2), we can use the isotropic model to calculate RS from observational estimates
281 of radiative fluxes at the TOA and surface and use these same fields (routinely available from model
282 simulations) to assess model biases in RS and diagnose their role in the SIAF.

283 Observational estimates of climatological radiative fluxes are taken from the CERES EBAF
284 surface product version 4.0 (Loeb and Coauthors 2018; Kato and Coauthors 2018) between 2000
285 and 2018. Climate model RS is estimated using the isotropic model for the last decade (1995-
286 2005) of historical CMIP5 (Taylor et al. 2012) climate simulations forced. ¹

287 Maps of summer (MJJA) RS estimated from satellite products and models are shown in Fig. 6.
288 Three spatial averages of RS are also provided: (i) the whole domain poleward of 60°N (upper
289 left corner in black) – observational value of $1.79 \text{ W m}^{-2} \%^{-1}$; (ii) the Arctic ocean excluding
290 land masses (lower right in blue) – observational value of $1.68 \text{ W m}^{-2} \%^{-1}$ and; (iii) the spatial
291 average over the sea ice (the spatial footprint and region varies between models, lower left in
292 purple) – observational value of $1.92 \text{ W m}^{-2} \%^{-1}$. The observational RS is very similar to multi-
293 model mean values ($1.72, 1.65$ and, $1.79 \text{ W m}^{-2} \%^{-1}$ over the entire Arctic domain, Arctic ocean
294 and sea ice regions respectively). The models and observations generally agree on the spatial
295 pattern of RS over the Arctic with high values over the Greenland ice sheet where the reduced
296 mass of the atmosphere above the high topography is associated with enhanced atmospheric SW
297 transmissivity, lower RS values over the GIN Sea and more spatially uniform RS values over the
298 Central Arctic. The magnitude of RS differs substantially across models with domain average RS
299 varying by almost a factor of two between the models, consistent with results from the radiative
300 kernel based RS calculation (Fig. 1). The inter-model (2σ) spread in Arctic average RS is 0.57,
301 0.53 and, $0.64 \text{ W m}^{-2} \%^{-1}$ over the full Arctic domain, Arctic ocean and climatological sea ice
302 respectively.

303 The SO observational estimate of summertime (NDJF) RS is similar but slightly lower (domain
304 average excluding the Antarctic continent of $1.56 \text{ W m}^{-2} \%^{-1}$) than the multi-model mean (1.71
305 $\text{W m}^{-2} \%^{-1}$). All models and observations show an annular structure in RS with smaller values

¹Most of the radiative kernel calculations discussed in Section 2 used “modern”, slightly differing time periods.

306 in the storm track region and larger values adjacent to the Antarctic continent over the sea ice
307 (Fig. 7). RS differs substantially between models (order factor 2) in the storm track region and
308 on the location and lateral extent of the high RS region adjacent to the continent. Some models
309 (i.e. CSIRO MK5) also have zonal asymmetries in RS that are best characterized as a zonal
310 wavenumber 1 pattern. The domain average RS values differ by less than the factor of 2 differences
311 seen in the Arctic, but the local RS difference between models –especially in the storm track region
312 – are of order a factor of 2. The inter-model (2σ) spread in SO average RS is $0.54 \text{ W m}^{-2} \%^{-1}$,
313 comparable in magnitude to that over the Arctic domain and Arctic ocean.

314 These results collectively suggest that while CMIP5 ensemble average RS of high latitude ice
315 loss is quite similar to that implied from observational constraints, models diverge substantially
316 on the radiative impact of ice loss because of differences in atmospheric optical properties (i.e.
317 clouds).

318 **4. Observational estimate of ice albedo feedback**

319 The Arctic sea-ice-albedo feedback (SIAF) is (the spatial average of) the product of the RS –
320 the TOA radiative impact of surface albedo changes – and the ice sensitivity (IS) – the surface
321 albedo change due to Arctic sea ice loss per unit of global warming (Eqs. 1,2). Thus, the RS
322 calculated from the climatological radiative fluxes and the isotropic model in the previous sections
323 along with estimates of IS from the observational record provide an observational estimate of the
324 SIAF that can be compared to the SIAF calculated using the same methodology applied to CMIP5
325 simulations with historical and long term forcing. Furthermore, we can explicitly ask if the model
326 spread (and potential bias relative to observations) in SIAF is explained by RS or IS spread.

327 The observational estimate of IS is calculated from the changes in decadal surface albedo of
328 the Arctic ocean from 1982 to 2016 (2007-2016 average minus 1982-1991 average – Fig. 8)

329 during each summer month divided by the global mean surface temperature (T_S) change over the
 330 same time period. We use two different observational based data sets to calculate the change in
 331 surface albedo over this time period : (i) sea ice concentration calculated by the National Snow
 332 and Ice Data Center (Cavalieri et al. 1996) from passive microwave brightness measured by the
 333 Nimbus 7 satellite available from 1979-2016 and (ii) broadband (all-sky) surface albedo measured
 334 by the Advanced Very High Resolution Radiometer (AVHRR) Polar Pathfinder (APP-x) extended
 335 data set (Wang and Key 2005) that covers the 1982-2017 time period. The central estimate of
 336 our observational based IS is the average of calculations from these two data sets (elaborated on
 337 below) and our uncertainty estimates account for differences across the two data-sets.

338 The NSIDC sea-ice concentration changes are converted to a surface albedo change record by
 339 multiplying the SIC changes by the albedo contrast between sea ice and open ocean ($\Delta\alpha$), which
 340 is assumed to be spatially and temporally invariant:

$$IS = \frac{dSIC}{[dT_S]} \Delta\alpha. \quad (5)$$

341 Eq. 5 assumes that changes in α_{SI} are isolated to regions of sea-ice melt. NSIDC monthly maps of
 342 the decadal average change in sea ice concentration are multiplied by an assumed surface albedo
 343 contrast between the open ocean and sea ice ($\Delta\alpha$) of 0.54 -assuming a typical ice α of 0.6 (Hummel
 344 and Reck 1979) and an ocean albedo of 0.06 (Hansen et al. 1983). This choice of typical ice
 345 albedo is an average of snow covered sea ice found during the late spring and sea ice with melt
 346 ponds in the late summer (see Fig. 9 of Perovich et al. 2002). This map of monthly NSIDC
 347 ice-concentration derived surface albedo change and those derived from the APP-x (also monthly)
 348 data are averaged to produce the observational best estimate of change in surface albedo (Fig. 8C)
 349 – hereafter referred to as the Observational Best Estimate (OBE). Both products produce similar
 350 estimates of surface albedo changes (Appendix Fig. A4). We use differences between the two

351 surface albedo data sets as well as the intra-decadal variability within each data set to calculate the
352 uncertainty in observational IS (Fig. 8D) as outlined in the Appendix.

353 Observational IS is calculated by normalizing OBE surface albedo changes by a global surface
354 temperature change of $0.7 \pm 0.1\text{K}$ over the 1982-2016 time period. The central estimate and
355 uncertainty in global mean surface temperature change come from the average and standard devi-
356 ation of the mean across three different global surface temperature data sets: (i) National Centers
357 for Environment Prediction (NCEP) reanalysis surface air temperature (Kalnay et al. 1996), (ii)
358 the Goddard Institute for Space Studies Surface Temperature Analysis (GISTEMP) (Hansen et al.
359 1999), and (iii) the modification by Cowtan and Way (Cowtan and Way 2014) of the Met Office
360 Hadley Centre surface temperature dataset (Morice et al. 2012) version 4 (HadCRUT4).

361 The monthly IS is then multiplied by the monthly RS derived from CERES data, and then time
362 averaged (over the summer months) to produce a map of radiative impact of sea ice changes (Fig.
363 8E). While the previous figures showed MJJA average in the NH and NDJF in the SH, figure 8E ex-
364 tends the summertime season to include the six months centered on the summer solstice (AMJJAS
365 in NH and ONDJFM in the SH) since previous work (Flanner et al. 2011) found an appreciable
366 contribution to the SIAF during the shoulder seasons especially in April. The uncertainty in the
367 $RI_{TOA,\alpha}$ is (Fig. 8F) is assessed from a Monte Carlo simulation that takes into account 3 different
368 uncertainties in the input data sets propagated (in quadrature) onto the calculation of $RI_{TOA,\alpha}$ (see
369 Appendix for details): (i) the uncertainty in RS – due to uncertainty in the climatological radia-
370 tive fluxes, (ii) uncertainty in the surface albedo change – due to both intra-decadal variability
371 and differences between the APP-x and NSIDC ice concentrations data sets and (iii) uncertainty
372 in the global mean temperature change that goes into the calculation of IS. We note that $RI_{TOA,\alpha}$
373 in Fig. 8E is, by definition, the radiative impact of sea ice changes normalized by global mean
374 surface temperature change ($=0.7\text{K}$) and has a summertime (AMJJAS) Arctic domain average of

375 $4.9 \pm 1.4 \text{ W m}^{-2} \text{ K}^{-1}$ which translates to an absolute change in summertime radiation of $3.4 \pm$
376 1.0 W m^{-2} over the Arctic. To convert this number to a global and annual mean radiative impact,
377 one must weight this number by the ratio of summer months to the year ($\frac{6}{12}$) and the spatial area
378 of the Arctic (poleward of 60°N) divided by that of the globe (.065) resulting in a global TOA
379 radiative change of 0.11 W m^{-2} over the 1982-2016 period. This translates to a global radiative
380 feedback (divide by $0.7 \text{ K global } T_S \text{ change}$) of $0.16 \pm 0.04 \text{ W m}^{-2} \text{ K}^{-1}$ given the observed global
381 surface temperature change over the same period. The uncertainties cited above reflect 2 standard
382 deviations.

383 We do not estimate the observational based surface albedo feedback in the SO because the
384 change in SO sea ice concentration over the observational period is not statistically significant
385 above the year-to-year variability (Jones and Coauthors 2009). We also note that this estimate is
386 isolated to the Arctic ocean (we have masked the APP-x albedo changes over land) and, thus, does
387 not include the impact of changes in snow cover over land.

388 **5. Comparison of observational and model SIAF and decomposition of inter-model spread** 389 **of SIAF into RS and IS**

390 We now compare the observational Arctic SIAF derived above with that derived by the same
391 methodology in historical CMIP5 simulations. The RS for each climate model that was calculated
392 using the isotropic model in the previous section (from the climatology at end of the historical
393 simulation – 1995 to 2005) is multiplied by the decadal average surface albedo change –calculated
394 as the ratio of upwelling to downwelling broadband shortwave radiation at the surface – over
395 the historical simulation (1995 to 2005 minus 1975 to 1985). We note that this time period was
396 chosen to correspond to the end of the historical simulations and differs from the 1982 to 2016
397 period used for the observational calculations. The RS and surface albedo changes are calculated

398 for each month and the product is spatially averaged over the Arctic ocean to calculate the SIAF;
399 we exclude the impact of changes in snow cover over land from our calculations. For simplicity,
400 we will only discuss the annual and global mean of the calculations normalized by the global
401 mean surface temperature change over the same time period, as we did for the observations. The
402 CMIP5 ensemble mean Arctic SIAF in the historical simulations is $0.12 \text{ W m}^{-2} \text{ K}^{-1}$ with a spread
403 (2 standard deviations, σ) of $0.13 \text{ W m}^{-2} \text{ K}^{-1}$ (gray histogram in Fig. 9A with wide bars). The
404 ensemble mean is slightly smaller than the observational estimate (c.f the solid and dashed vertical
405 black lines in Fig. 9) but the large inter-model spread indicates that the models differ in either RS
406 and/or IS. We now ask how much RS and IS contribute to the inter-model differences in Arctic
407 SIAF.

408 To estimate the IS contribution to the SIAF spread, the calculation of SIAF is repeated but the
409 model specific RS is replaced with the observational based RS value. The resulting distribution of
410 SIAF (blue histogram in Fig. 9A) shows the spread produced by biases and inter-model differences
411 in IS. The mean value of SIAF in the fixed RS distribution ($0.12 \text{ W m}^{-2} \text{ K}^{-1}$ – Table 1) is nearly
412 equal to that of the full SIAF calculation (c.f. the blue and black vertical lines). The CMIP5
413 ensemble average SIAF is quite insensitive to RS model biases, and it is lower than the observed
414 estimate because the modeled IS is smaller than the observational estimate. Furthermore, the
415 spread in the fixed RS distribution is only slightly smaller than that of the full SIAF calculation
416 ($2\sigma = 0.12 \text{ W m}^{-2} \text{ K}^{-1}$) indicating that the majority of the inter-model spread in SIAF calculated
417 from the historical simulation is a result of the IS differences between models.

418 A similar analysis can be made to estimate the impact of biases (relative to observations) and
419 inter-model RS differences on the calculated SIAF by replacing the model specific IS with that
420 derived from observations (red histogram in Fig. 9A). The CMIP5 ensemble average SIAF of the
421 fixed IS distribution ($0.16 \text{ W m}^{-2} \text{ K}^{-1}$ – Table 1) is larger than that of the full SIAF calculation

422 (c.f. the red and black vertical lines in Fig. 9) indicating that the CMIP5 ensemble average IS
423 is smaller than that observed (the OBE value) a result also found by Rosenblum and Eisenman
424 (2016). The inter-model spread in SIAF in the fixed IS experiment ($2\sigma = 0.04 \text{ W m}^{-2} \text{ K}^{-1}$)
425 is smaller than that of the full calculation and fixed RS experiment indicating that inter-model
426 differences in RS play a smaller but not insignificant role in the SIAF spread calculated over the
427 historical simulations. A summary of the role of biases and inter-model differences in RS and IS
428 in determining the model distribution of SIAF is provided in Table 1.

429 This partitioning of SIAF differences in contributions from RS and IS takes spatial and temporal
430 co-variances of ice loss and RS into account by weighting the ice loss to the RS at that location
431 and time. Similar results for the impact of IS and RS on the total spread in SIAF are obtained by
432 simply noting the fractional spread (relative to the ensemble mean) of summertime Arctic domain
433 average RS and IS between models. The ratio of domain and summertime average inter-model
434 spread (2σ) to the ensemble mean domain and summertime average of RS is 40% whereas that
435 of IS is 107% roughly scaling with the fractional contribution to SIAF spread calculated above.
436 This result suggests that inter-model differences in IS and RS are fairly spatially and temporally
437 homogenous and the resultant inter-model spread in SIAF is independent of the spatiotemporal
438 co-variability of RS and IS. Previous work has found similar large magnitude inter-model spread
439 in IS in CMIP3 (Mahlstein and Knutti 2012) and CMIP5 (Stroeve and Notz 2015) linked to the
440 spread in the magnitude of Arctic amplification.

441 The sea ice retreat over the historical record represents the superposition of the response to
442 climate forcing and natural variability and, thus, the inter-model spread in IS calculated over the
443 30 years of historical simulations is expected to exceed that in response to long-term sustained
444 forcing. [Schneider et al. \(2018\)](#) found that decadal trends in sea-ice during periods when global
445 mean temperatures increased by more than 0.5K provided good estimates of the long-term SIAF in

446 an ensemble of climate models. Other studies suggest that as much as 50% of the observed Arctic
447 sea loss since 1979 could be a result of the natural variability of atmospheric circulation (Ding
448 et al. 2017, 2019; Kay et al. 2011). To reduce the amount of internal variability relative to the
449 forced component, we also look at the contribution of RS and IS to the inter-model spread in SIAF
450 in response to an abrupt and sustained quadrupling of atmospheric CO₂ where the forced climate
451 change signal is expected to be larger than the natural variability. The IS in the CO₂ quadrupling
452 simulations is calculated from the change in surface albedo and global mean surface temperature
453 between the PI and the average over years 50-100 since CO₂ quadrupling. The RS used to calculate
454 the SIAF is calculated from the PI climatological fields in the same model. The ensemble average
455 Arctic SIAF calculated from the 4XCO₂ simulations is $0.13 \pm 0.09 \text{ W m}^{-2} \text{ K}^{-1}$ (uncertainty is
456 2σ) and is in close agreement with the ensemble average of the historical simulation (0.12 ± 0.13
457 $\text{W m}^{-2} \text{ K}^{-1}$) with reduced inter-model spread. The central estimate and range of SIAF from all
458 model simulations – calculated from 2σ of the mean – is $0.13 \pm 0.02 \text{ W m}^{-2} \text{ K}^{-1}$ and is slightly
459 smaller than but not statistically different from the observational estimate ($0.16 \pm 0.04 \text{ W m}^{-2}$
460 K^{-1}). Because the 4XCO₂ ice response primarily reflects the forced response, the similarity of
461 the ensemble average SIAF diagnosed from historical and 4XCO₂ simulations suggests that the
462 same physics responsible for the long-term SIAF are evident in historical simulations despite the
463 additional statistical noise from internal variability.

464 When the model specific RS is replaced by the observational estimate of RS the resultant Arctic
465 SIAF for the CO₂ quadrupling simulations is $0.13 \pm 0.08 \text{ W m}^{-2} \text{ K}^{-1}$ and when the model specific
466 IS is replaced by the observational estimate of IS the resultant SIAF is $0.16 \pm 0.04 \text{ W m}^{-2} \text{ K}^{-1}$
467 (lower left panel of Fig. 9 and Table 1). These results suggest that in the long-term response to
468 sustained anthropogenic forcing: 1.) the CMIP5 ensemble average RS (spatially and temporally
469 weighted by the relevant regions of ice loss) is very near the observational estimate, 2.) the CMIP5

470 ensemble average IS (spatially and temporally weighted by structure of RS) is slightly smaller
471 than the observational estimate and is responsible for the model SIAF being smaller than the
472 observational estimate and 3.) inter-model differences in IS contribute twice as much to the inter-
473 model spread in SIAF (63% of the ensemble average value) than do inter-model differences in
474 RS (30% of the ensemble average value). We note that, the inter-model spread in IS and RS are
475 significantly ($R=0.54$) correlated (at 95% confidence interval) and return to the implication of this
476 result in the discussion section.

477 A similar analysis can be performed for the 4XCO₂ simulations in the SO (poleward of 55°S) to
478 indicate an ensemble average SIAF of $0.08 \pm 0.13 \text{ W m}^{-2} \text{ K}^{-1}$ (Fig. 9C, Table 1). The SO SIAF
479 is negative in a single model (GFDL ESM2G) that simulates sea ice growth in the Weddell Sea
480 under 4XCO₂. When the model specific RS is replaced by the observational estimate of RS, the
481 calculated SO SIAF is $0.07 \pm 0.11 \text{ W m}^{-2} \text{ K}^{-1}$ suggesting the the ensemble average RS is slightly
482 larger than that estimated from the observations, consistent with Fig. 7. Because no observational
483 estimate of SO IS is available, we probe the sensitivity of SO SIAF to RS by replacing the model
484 specific IS with the ensemble average IS resulting in a calculated SIAF of $0.08 \pm 0.04 \text{ W m}^{-2} \text{ K}^{-1}$;
485 inter-model differences in RS result in inter-model differences in SO SIAF of magnitude 50% the
486 ensemble mean estimate. However, the contribution of inter-model spread in RS to SIAF spread is
487 dwarfed by the impact of inter-model differences in IS which produces inter-model differences in
488 SO SIAF exceeding the central estimate by almost a factor of 1.5 (160%). This result is consistent
489 with the large inter-model differences in SO ice response to global warming reported by Shu et al.
490 (2015); Polvani and Smith (2013).

491 *a. Global surface albedo feedback – comparison to IPCC AR5 value*

492 The IPCC AR5 estimated a global surface albedo feedback of $0.26 \text{ W m}^{-2} \text{ K}^{-1}$ based on the
493 calculations of Soden and Held (2006b) which use a single RS – derived from kernel calculations
494 in the GFDL model (Fig. 1) – applied the surface albedo change in each CMIP3 model. These
495 calculations are global and include the impact of α changes over land (due to changes in snow
496 cover) in addition to the sea-ice related changes considered up to this point and we term this com-
497 bined contribution of land and sea ice changes the global albedo feedback (GAF). More recently,
498 Schneider et al. (2018) presented a CMIP5 ensemble mean GAF $0.40 \text{ W m}^{-2} \text{ K}^{-1}$ using NCAR
499 CAM5 based kernels. It is unclear if this discrepancy results from the different RS used in these
500 studies or the IS in different GCM ensembles. Here, we compare the GAF produced using the
501 (kernel based) RS from a single model to that calculated using a model specific RS derived from
502 the isotropic model.

503 Our GAF calculations are based upon surface albedo change calculated from the 4XCO_2 sim-
504 ulations minus that in the PI simulation normalized by the global mean surface temperature (TS)
505 change in that model – a quantity akin to IS in Eq. 2 but including the albedo changes over land.
506 This albedo change is multiplied by RS estimated two ways: (i) using the method introduced in
507 this study, where RS is calculated from the isotropic model (Eq. 4) using radiative fluxes from
508 appropriate model specific PI simulation and (ii) using the method introduced by Soden and Held
509 (2006b) where the GFDL surface albedo kernel (Fig. 1) is used to estimate RS for all models. We
510 separate the GAF calculation into hemispheres. In the NH, the GAF calculated in this study is
511 larger than that calculated using the GFDL kernel in all models (all the red dots fall below the 1:1
512 line in the upper left panel of Fig. 10) as would be expected from the GFDL RS being at the very
513 low end of the model range especially over the ocean domain. In the CMIP5 ensemble average,

514 the NH **GAF** is $0.20 \text{ W m}^{-2} \text{ K}^{-1}$ using the GFDL kernel as compared to $0.27 \text{ W m}^{-2} \text{ K}^{-1}$ using
515 the isotropic model methodology (35% greater – Table 2). The NH **GAF** has 64% more spread
516 using the model specific RS because: (i) the ensemble mean RS is larger than the GFDL kernel RS
517 and (ii) the inter-model spread in RS contributes to the **GAF** spread as discussed in the previous
518 subsection. If we restrict the calculation to the Arctic ocean poleward of 60°N (as was done in
519 Sections 4 and 5) we find a CMIP5 ensemble average SIAF of $0.09 \text{ W m}^{-2} \text{ K}^{-1}$ using the GFDL
520 kernel compared to the $0.13 \text{ W m}^{-2} \text{ K}^{-1}$ (Table 1) using the isotropic model methodology (45%
521 greater). This result suggests that approximately half of the **GAF** is due to α changes over land as
522 found by Flanner et al. (2011).

523 In the Southern Hemisphere, the **GAF** estimates from the two methods are in closer agreement;
524 the dots cluster along near the 1:1 line in the upper right panel of Fig. 10 with the exception of
525 the models producing the highest **GAF**. This result is expected since the GFDL RS is near the en-
526 semble mean over the SO (Fig. 2 and Fig. 7). The ensemble average **GAF** in the SH is, therefore,
527 very similar when using the methodology in this study ($0.09 \text{ W m}^{-2} \text{ K}^{-1}$) as compared to that
528 calculated using GFDL RS only ($0.08 \text{ W m}^{-2} \text{ K}^{-1}$) – Table 2. Globally, we calculate an **GAF**
529 of $0.37 \text{ W m}^{-2} \text{ K}^{-1}$ which is 30% greater than the same result found applying the GFDL RS to
530 CMIP5 4XCO₂ simulations of $0.29 \text{ W m}^{-2} \text{ K}^{-1}$. We note the the IPCC AR5 cites a global **GAF** of
531 $0.26 \text{ W m}^{-2} \text{ K}^{-1}$ derived from the GFDL kernel and CMIP4 simulations and, thus, our estimate
532 is 40% larger than the AR5 value. We attribute 30% of this increase to improved methodology
533 of using model specific RS and 10% to the difference between CMIP4 and CMIP5 model char-
534 acteristics. **Importantly, the IPCC diagnosis of the overall climate sensitivity of climate models is**
535 **unaffected by our revised more positive GAF. Rather, our results suggest that the shortwave cloud**
536 **feedback should be revised downward by the same amount because cloud feedbacks are diagnosed**

537 from all-sky minus clear-sky TOA radiation adjusted by all-sky minus clear-sky radiative kernel
538 calculations.

539 **6. Summary and discussion**

540 We have shown that the radiative impact of surface albedo changes (RS) calculated using of-
541 fline radiative transfer models (radiative kernels) can be closely replicated using a single layer
542 isotropic SW radiation model applied to the climatological radiative fluxes at the TOA and sur-
543 face. This procedure allows estimates of SIAF to be conveniently calculated from observational
544 data sets and standard model output without use of a kernel calculation, facilitating a comparison
545 of observational and model estimates of SIAF. It also allows the differences between models and
546 observations based calculations to be decomposed into contributions from RS and IS. The multi-
547 model mean of RS is close to the observational estimate in the Arctic and only slightly larger than
548 the observational estimate in the SO. However, the inter-model spread in RS (Figs. 6 and 7) is
549 substantial, producing inter-model differences in SIAF estimates that are 30% and 50% the mag-
550 nitude of the ensemble mean SIAF in the Arctic and SO respectively. In agreement with Sledd
551 and L'Ecuyer (2019), high latitude clouds tend to mask the impact of surface albedo variations on
552 the TOA albedo by a factor of 2-3 in observational estimates. Differences in climate model clouds
553 influence the degree of cloud masking.

554 Our results indicate that inter-model differences in IS are more important than RS in explaining
555 the inter-model spread in SIAF. However, IS is not statistically independent of RS ($R = 0.54$). It
556 is possible that inter-model differences in RS contribute to inter-model difference in IS because
557 models that have a larger radiative response to sea ice loss will tend to have greater sea ice loss
558 due to a stronger positive feedback between initial ice loss and radiative heating. In this sense, the
559 contribution of RS to inter-model differences in SIAF of $0.04 \text{ W m}^{-2} \text{ K}^{-1}$ both in the Arctic and

560 SO can be thought of as a lower bound on the contribution of mean state radiative biases to the
561 SIAF. We hope to explore the impact of mean state radiative biases (RS) on IS and the persistence
562 of sea ice loss events in future work.

563 We estimate an observationally based global, and annually averaged increase in TOA radiation
564 of 0.11 W m^{-2} from Arctic sea ice changes over the 1982-2016 time period using observationally
565 based estimates of sea ice changes and the CERES derived radiative sensitivity (RS) implying a
566 SIAF of $0.16 \pm 0.04 \text{ W m}^{-2} \text{ K}^{-1}$. Flanner et al. (2011) found a Northern Hemisphere average
567 "cryospheric radiative forcing" of 0.45 W m^{-2} over the 1979-2008 time period about half of
568 which (0.22 W m^{-2}) was attributed to sea ice changes – the other half was attributed to snow
569 changes over land. Thus, the Flanner et al. (2011) result converted to a global average ($0.22/2$
570 $=0.11 \text{ W m}^{-2}$) agrees very well with our findings. Similarly, (Cao et al. 2015) found a Northern
571 Hemisphere SIAF of $0.25 \text{ W m}^{-2} \text{ K}^{-1}$ using observed surface albedo change and RS estimated
572 using model based kernels derived from GFDL (Soden and Held 2006b) and CAM3 (Shell et al.
573 2008). This result translates to a global feedback of Arctic changes of $0.12 \text{ W m}^{-2} \text{ K}^{-1}$ which is
574 smaller than our central estimate and we speculate this result follows from the lower than observed
575 RS in the CAM3 kernel (Fig. 1).

576 Pistone et al. (2014, 2019) calculated a substantially larger SIAF ($0.31 \pm 0.04 \text{ W m}^{-2} \text{ K}^{-1}$) from
577 the inter-annual covariance of sea ice concentration and TOA radiation measured by CERES. We
578 speculate that some of the TOA radiative variability that coincides with ice loss events in Pistone
579 et al. (2014) is not directly a consequence of (i.e. geographically co-located with and/or a radiative
580 consequence) surface albedo changes but, rather, is a consequence of atmospheric optical proper-
581 ties (i.e. clouds, water vapor, etc) that co-vary with Arctic sea ice concentration. A central question
582 moving forward is whether the atmospheric changes (and the associated radiative anomalies) ac-
583 companying Arctic sea ice loss over the limited historical period result from natural variability of

584 atmospheric circulation initiated by tropical and mid-latitude processes or are a direct result of sea
585 ice loss and, thus, should be expected to also apply to future climatological changes. Additionally,
586 how accurately does the observational IS calculated over the historic record represent the expected
587 relationship between future changes in Arctic ice concentration and global mean temperature?

588 Pistone et al. (2014) suggest that the SIAF alone results in a 25% enhancement of global warm-
589 ing via radiative feedbacks, a value they derive from the ratio of their calculated radiative impact of
590 historic ice loss divided by the anthropogenic climate forcing to date. We offer two modifications
591 as updates to their calculation: (i) a significantly lower estimate of the radiative impact of Arctic
592 sea ice loss outlined above and (ii) consideration of how the implied feedback relates to equilib-
593 rium climate sensitivity, noting that the climate system is not currently in equilibrium with the
594 anthropogenic forcing to date. For the latter reason, the feedback gain of the Arctic SIAF should
595 be calculated by comparing the SIAF to the equilibrium radiative feedback of all other radiative
596 processes as opposed to the ratio of the transient radiative impact of ice loss to date to the applied
597 forcing. Given observational central estimates of the total equilibrium feedback parameter of -1.19
598 $\text{W m}^{-2} \text{K}^{-1}$ (Armour 2017) and our observational estimate of the Arctic SIAF ($\lambda_{SIAF} = +0.16 \pm$
599 $0.04 \text{ W m}^{-2} \text{K}^{-1}$) the implied feedback parameter of all processes excluding the SIAF (λ_0) satis-
600 fies the equation $-1.19 \text{ W m}^{-2} \text{K}^{-1} = \lambda_0 + 0.16 \text{ W m}^{-2} \text{K}^{-1}$. This implies that λ_0 (the reference
601 climate feedback parameter of a system with no SIAF) is $-1.35 \text{ W m}^{-2} \text{K}^{-1}$. We note that the
602 reference climate feedback parameter is more negative than that of a system with a SIAF implying
603 a smaller climate sensitivity of the reference system relative to the full system with a SIAF as is
604 expected for the positive SIAF. The fractional amplification of global mean temperature changes
605 – the feedback gain, G_{SIAF} – due to the SIAF is then (Roe 2009):

$$G_{SIAF} = \frac{1}{1 + \frac{\lambda_{SIAF}}{\lambda_0}} = 1.14 \pm .04. \quad (6)$$

606 Thus, our analysis suggests that the Arctic SIAF amplifies global warming by 14% (2σ range be-
 607 tween 10 and 19%) at the equilibrium timescale and is a more modest amplifier of global warming
 608 than the 25% suggested by Pistone et al. (2014).

609 The IPCC AR5 report (Flato et al. 2013) points out a discrepancy between the observational
 610 based SIAF of Flanner et al. (2011) and the model based estimate of Soden and Held (2006b)
 611 and speculates that models are biased toward low IS, but the role of inter-model spread and biases
 612 in RS were neglected. While we find no ensemble mean model bias in Arctic RS (Fig. 6), the
 613 model estimate of RS used in Soden and Held (2006b) is taken from radiative kernel calculations
 614 in a single (GFDL) model and then applied to the IS across models. The RS from that model
 615 (Fig. 1) is biased low relative to both the observational based RS (by 46% of the kernel RS in
 616 the Arctic average) and the CMIP5 ensemble mean. As a result, the AR5 estimate of the global
 617 surface albedo feedback of $0.26 \text{ W m}^{-2} \text{ K}^{-1}$ based on the calculations of Soden and Held (2006b)
 618 is substantially lower than our calculated value of $0.37 \text{ W m}^{-2} \text{ K}^{-1}$ which uses model specific RS
 619 estimates. This result suggests that at least some part of the low model bias identified in the IPCC
 620 AR5 is a consequence of using a RS that is inconsistent with some climate models. We recommend
 621 using model specific RS derived from the isotropic model as a better practice to applying radiative
 622 kernels across models. Additionally, our results identified no discernible model bias in the SIAF
 623 at least when considering like quantities over the Arctic ocean domain.

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632 APPENDIX

633 We describe the methodology used to calculate the uncertainty in our observational estimates
634 of the RS, IS and $RI_{TOA,\alpha}$ the spatial average of which gives the resultant SIAF (Eq. 1). We
635 do so by first bootstrapping (random re-sampling with replacement) the original observational
636 data into subsets half the temporal length of the original data to produce an ensemble of records.
637 For example, in the CERES data used to calculate the RS, we produce an ensemble of radiative
638 climatologies derived from random selections of 9 years of the 18 years of data. This procedure
639 queries how sensitive the radiative climatologies are to the limited length of the CERES record.
640 Similarly, the surface albedo changes are calculated from the difference of random selections of
641 5 year averages within the period 1982-1991 and 2007-2016. We then use the re-sampled data to
642 calculate the RS – using the isotropic model– and IS in a Monte-Carlo simulation. We calculate
643 100 different estimates of RS and 100 different estimates of IS with 50 derived from re-sampled
644 NSIDC ice concentration data and 50 derived from re-sampled APP-x data. Thus, our estimates
645 of IS (Fig. 8D) account for two sources of uncertainty: (i) the impact of intra-decadal variability
646 on calculating longer term changes in surface albedo and (ii) instrumental uncertainty.

647 The within data set intra-decadal variability of surface albedo contributes more to the IS uncer-
648 tainty than the differences between APP-x and NSIDC sea ice concentration data sets; the standard
649 deviation in IS calculated from ensembles of just the 50 NSIDC or 50 APP-x data is similar to that

650 derived from the 100 member ensemble considered collectively. Given that the NSIDC estimate
651 of surface albedo change is derived from sea ice concentration changes only and does not account
652 for changes in the albedo over ice, the similarity of the NSIDC and APP-x derived IS suggest that
653 albedo changes are primarily associated with changes in ice area, in opposition to the findings of
654 (Horvat et al. 2019). The uncertainty in RS (taken as 2 standard deviation across the re-sampled
655 ensemble) is approximately 10% of the mean RS with larger values in the vicinity of sea ice edge
656 (Fig. 8B) suggesting that the cloud properties that determine the RS are fairly constant from year-
657 to-year. In contrast, the uncertainty in the IS (Fig. 8D) is approximately 60% of the mean value
658 with particularly large uncertainties in the Beaufort Sea suggesting that the intra-decadal variabil-
659 ity and measurement uncertainty of sea ice changes substantially hinders the calculation of long
660 term IS over the relatively short observational record.

661 We now describe how we use the uncertainty in IS and RS to calculate the uncertainty in $RI_{TOA,\alpha}$,
662 the spatial average of which gives the SIAF uncertainty. We diagnose uncertainty $RI_{TOA,\alpha}$ by
663 convoluting the 100 estimates of RS and the 100 estimates in IS to produce 10,000 estimates of
664 $RI_{TOA,\alpha}$. This procedure accounts for the spatial co-variance of IS and RS uncertainty and central
665 estimates. For example, the uncertainty in IS will have a larger impact in the regions and seasons
666 where RS is largest. The uncertainty in the $RI_{TOA,\alpha}$ looks like and is comparable in fractional
667 magnitude to that in surface albedo change with a slight modification by the spatial pattern of
668 the mean RS. The spread in the spatial average of these 10,000 $RI_{TOA,\alpha}$ is combined with the
669 uncertainty in global mean temperature changes – propagated in quadrature since both quantities
670 are scalars– to produce a probability distribution function of SIAF (dark black distribution in left
671 panels of Fig. 9). These calculation give an Arctic SIAF of $0.14 \pm 0.4 \text{ W m}^{-2} \text{ K}^{-1}$ where the
672 uncertainty is taken as 2σ .

673 The uncertainty in the observational global SIAF can be decomposed into contributions from the
674 RS and IS uncertainty as follows: (i) the contribution of RS is calculated as 2σ of the distribution
675 derived from the 100 estimates of RS and multiplied by the OBE IS and (ii) the contribution of IS
676 is calculated as 2σ of the distribution derived from the 100 estimates of IS and multiplied by the
677 mean RS. The uncertainty in the observational SIAF is almost entirely ($\pm 0.04 \text{ W m}^{-2} \text{ K}^{-1}$) due
678 to uncertainty in the IS (dark blue narrow distribution in Fig. 9) whereas the uncertainty in the RS
679 contributes very little to the global uncertainty in the SIAF ($\pm 0.003 \text{ W m}^{-2} \text{ K}^{-1}$ – the very narrow
680 dark red distribution in the left Fig. 9A).

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 824 shows the full calculation using the model specific RS and IS. The second row
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TABLE 1: SIAF values (in $\text{W m}^{-2} \text{K}^{-1}$) for the (top) Arctic and (bottom) Southern Ocean derived from (left) Observations and model simulations of (middle) 4XCO₂ and (right) historical simulations. Each value shows the central estimate and 2σ range across the bootstrapping Monte-Carlo simulations for the observations and inter-model spread for the models. The top row in each hemisphere shows the full calculation using the model specific RS and IS. The second row shows the impact of inter-model differences in IS as calculated using the model specific IS and the observed RS. The third row shows the impact of inter-model differences in RS as calculated using the model specific RS and the observed IS.

Arctic			
	Observations	4XCO ₂	Historical
Full Calculation	0.16 ± 0.04	0.13 ± 0.09	0.12 ± 0.13
IS contribution $RS_{OBS} \times IS$		0.13 ± 0.08	0.12 ± 0.12
RS contribution $RS \times IS_{OBS}$		0.16 ± 0.04	0.16 ± 0.04
Southern Ocean			
	Observations	4XCO ₂	Historical
Full Calculation		0.08 ± 0.13	
IS contribution $RS_{OBS} \times IS$		0.07 ± 0.11	
RS contribution $RS \times IS_{OBS}$		0.08 ± 0.04	

TABLE 2: **Global albedo feedback (GAF)** in CMIP5 climate models calculated using the methodology of this study – with a model specific RS from the isotropic model– compared to that calculated using RS from the GFDL surface albedo kernel for all models. The CMIP5 ensemble mean and 2σ are shown for each hemisphere and divided into ocean and full domains.

Northern Hemisphere		
	Ocean Domain	Total
This Study	$0.15 \pm .10$	0.27 ± 0.18
GFDL RS kernel	0.11 ± 0.06	0.20 ± 0.11
Southern Hemisphere		
	Ocean Domain	Total
This Study	0.09 ± 0.16	0.10 ± 0.17
GFDL RS kernel	0.08 ± 0.14	0.09 ± 0.15
Global		
	Ocean Domain	Total
This Study	0.24 ± 0.15	0.37 ± 0.19
GFDL RS kernel	0.19 ± 0.11	0.29 ± 0.13

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858 been repeated from Fig. 1. The dark purple line shows the sea ice edge designated by the
859 MJJA 50 % sea ice concentration contour. The full domain spatial average is shown in the
860 upper left corner of each panel in black, the Arctic ocean average is shown in the lower right
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868 note the models for which radiative kernel calculations are available that have been repeated
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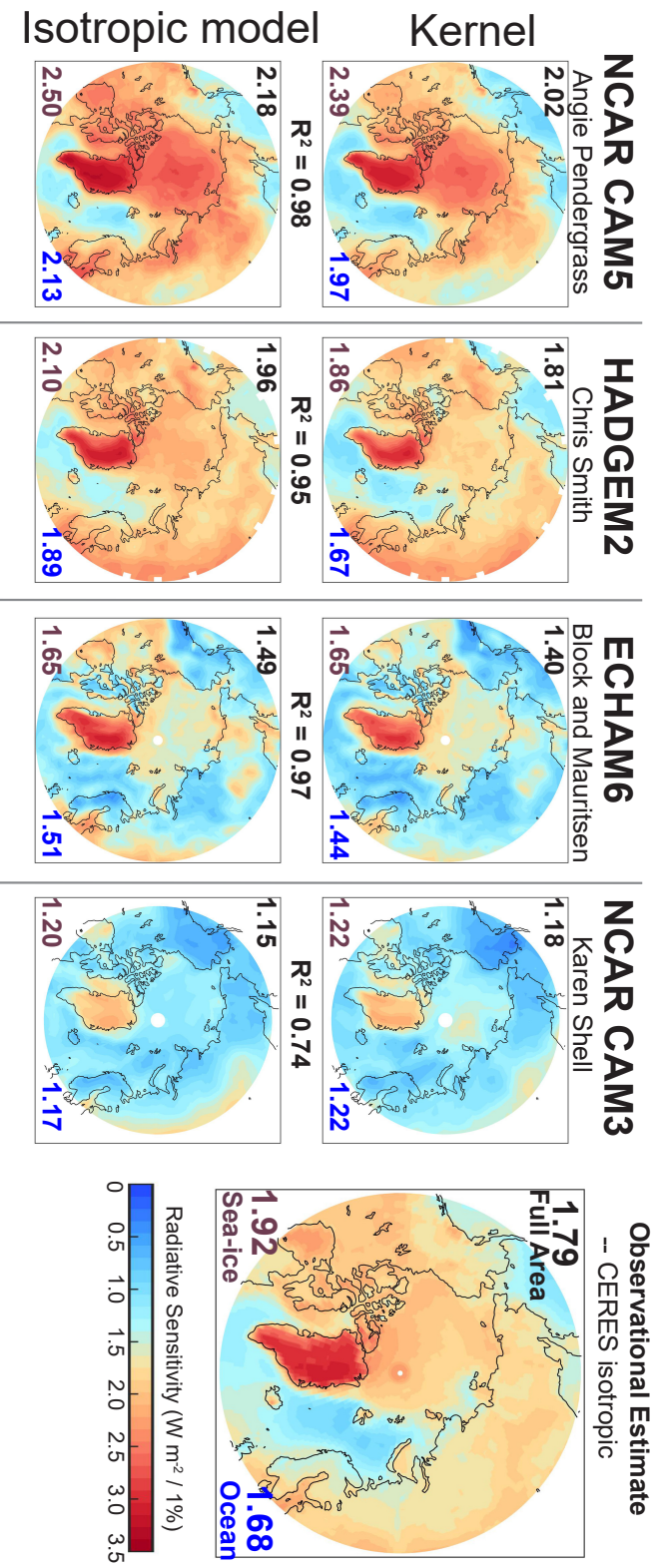


FIG. 1: Arctic summertime (MJA) surface albedo radiative sensitivity (RS) calculated from radiative kernels (upper panels) and estimated from the climatological radiative fields using the idealized isotropic radiation model (lower panels) in the same models. The squared spatial correlation coefficient between the kernel isotropic methods in the same model are provided in the middle and the Arctic domain averaged values are shown in the upper right of each panel. Note that there is no RS calculation from the GFDL model because mean state fields from this simulation were not saved.

Observational estimates from CERES EBAF satellite data and the isotropic model are shown to the right.

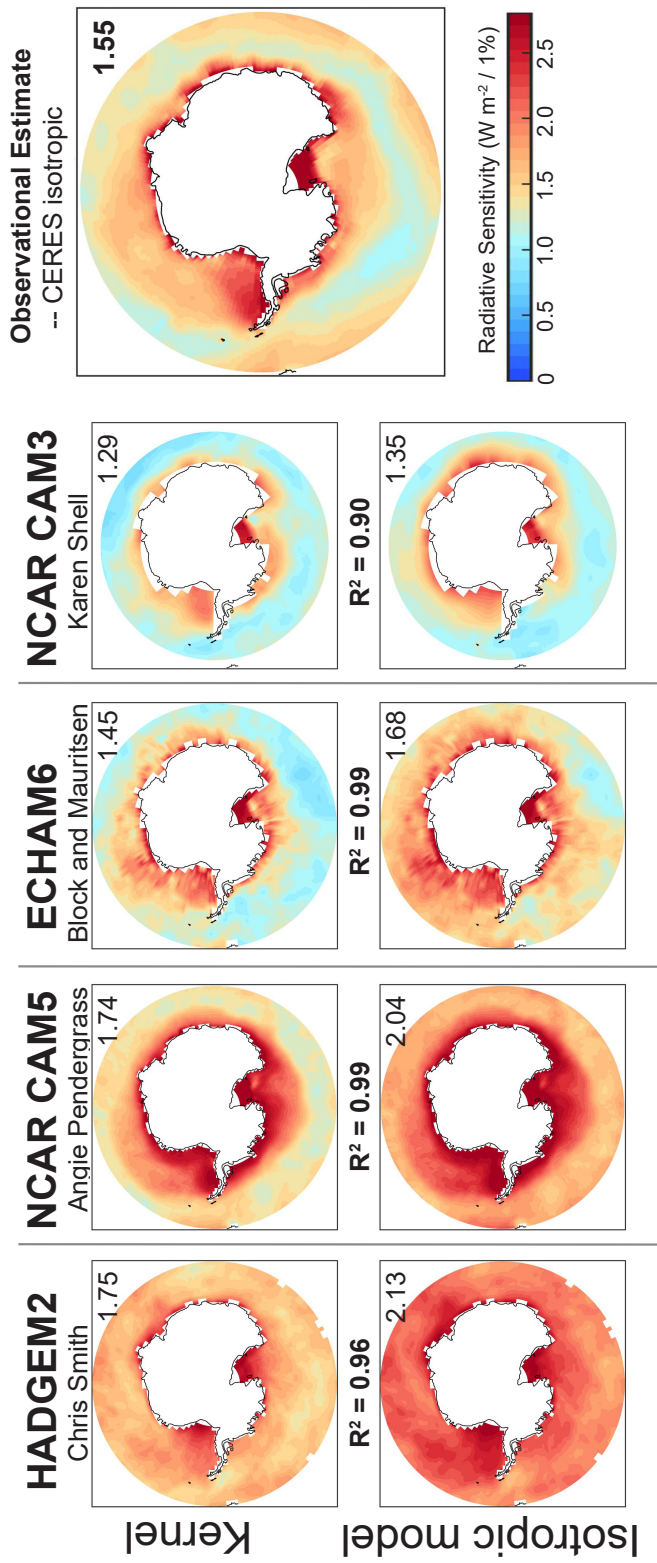


FIG. 2: As in Fig. 1 but for the NDJF RS in the Southern Ocean. Domain averaged surface albedo feedbacks exclude the Antarctic continent. Note that the figures are ordered by domain average RS over the Southern Ocean and this order differs from Fig. 1.

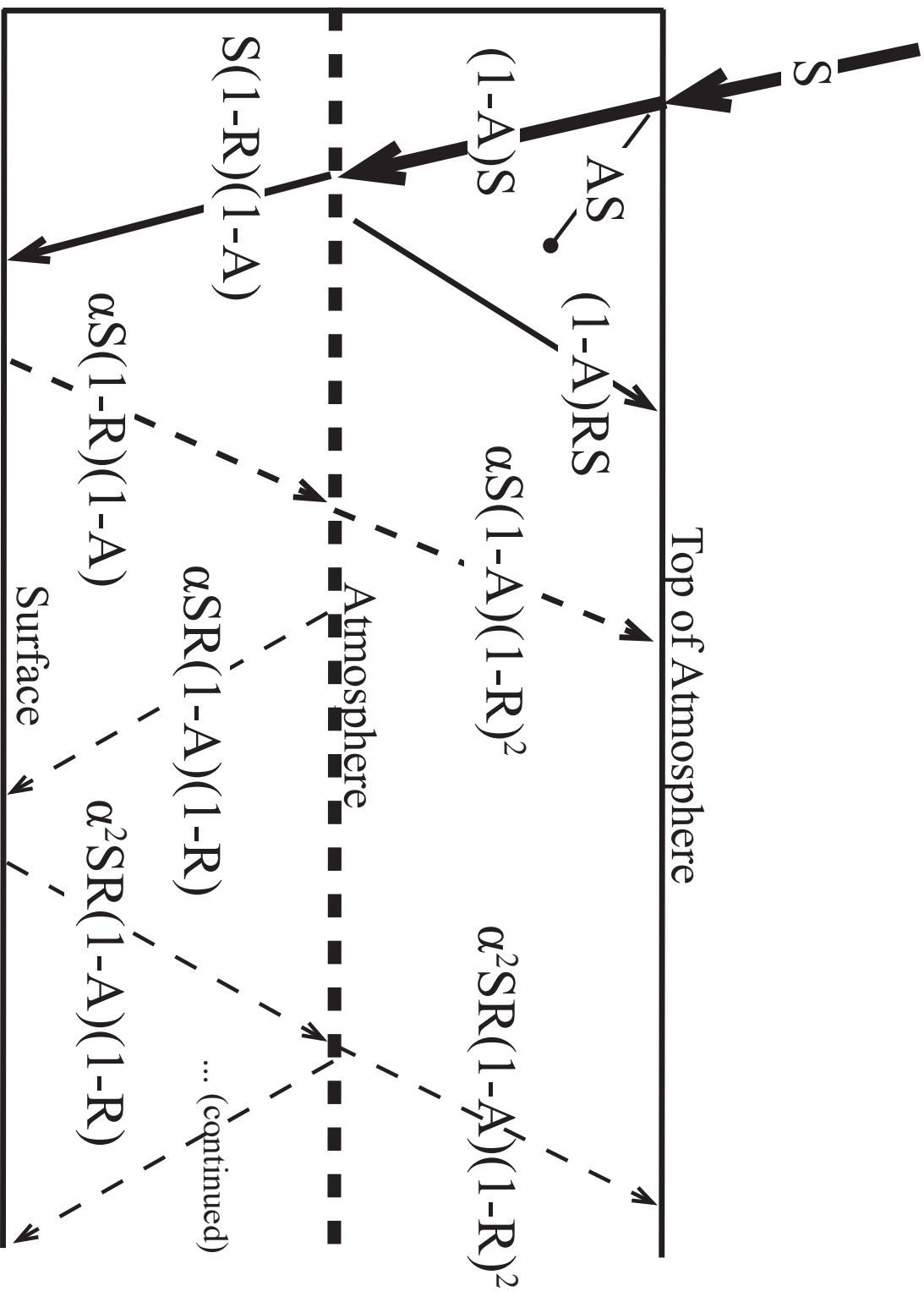


FIG. 3: Schematic of the single layer isotropic model modified from T07.

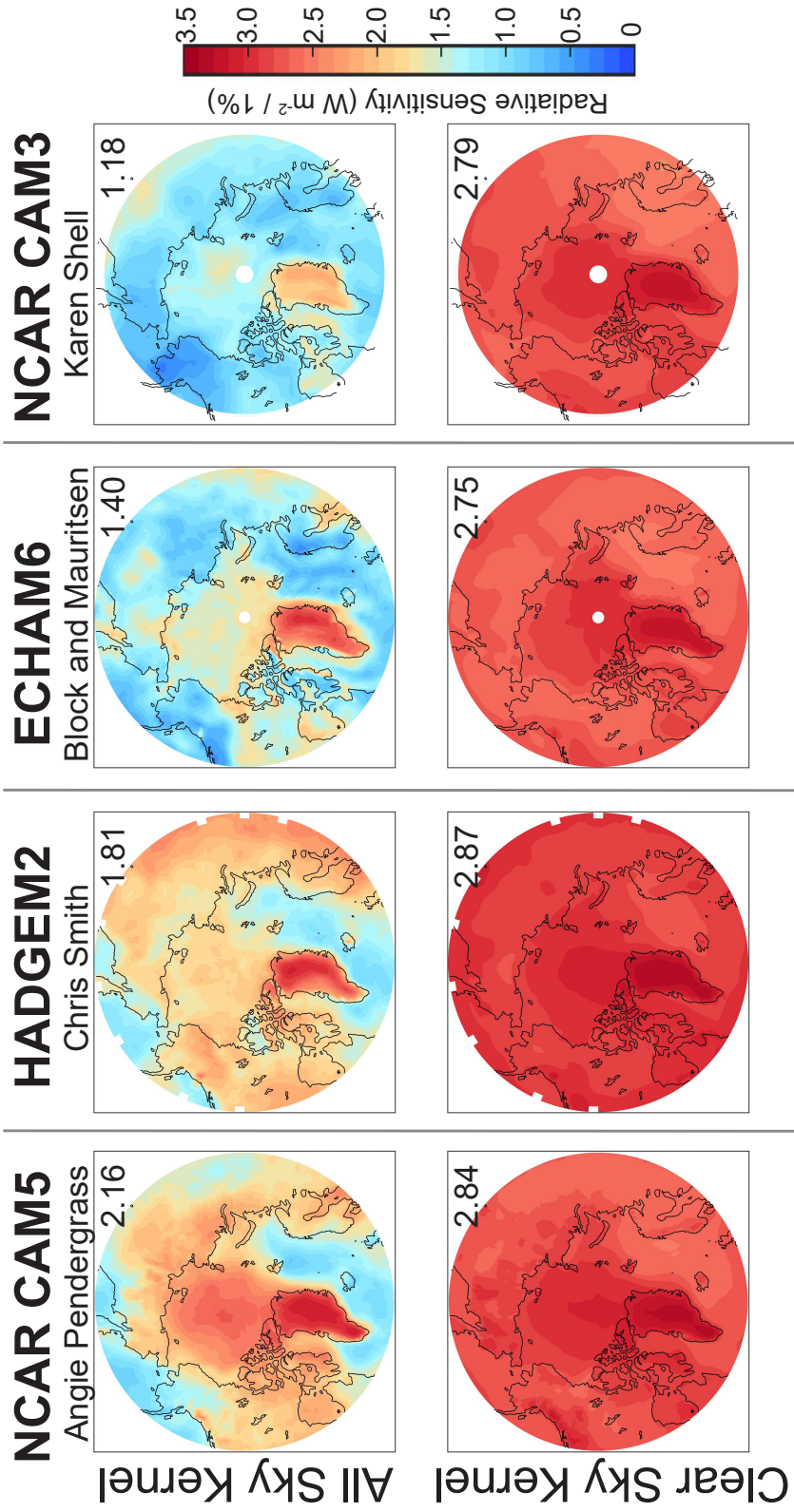


FIG. 4: Comparison of Arctic Summertime (MJJJA) full sky and clear sky surface albedo kernels.

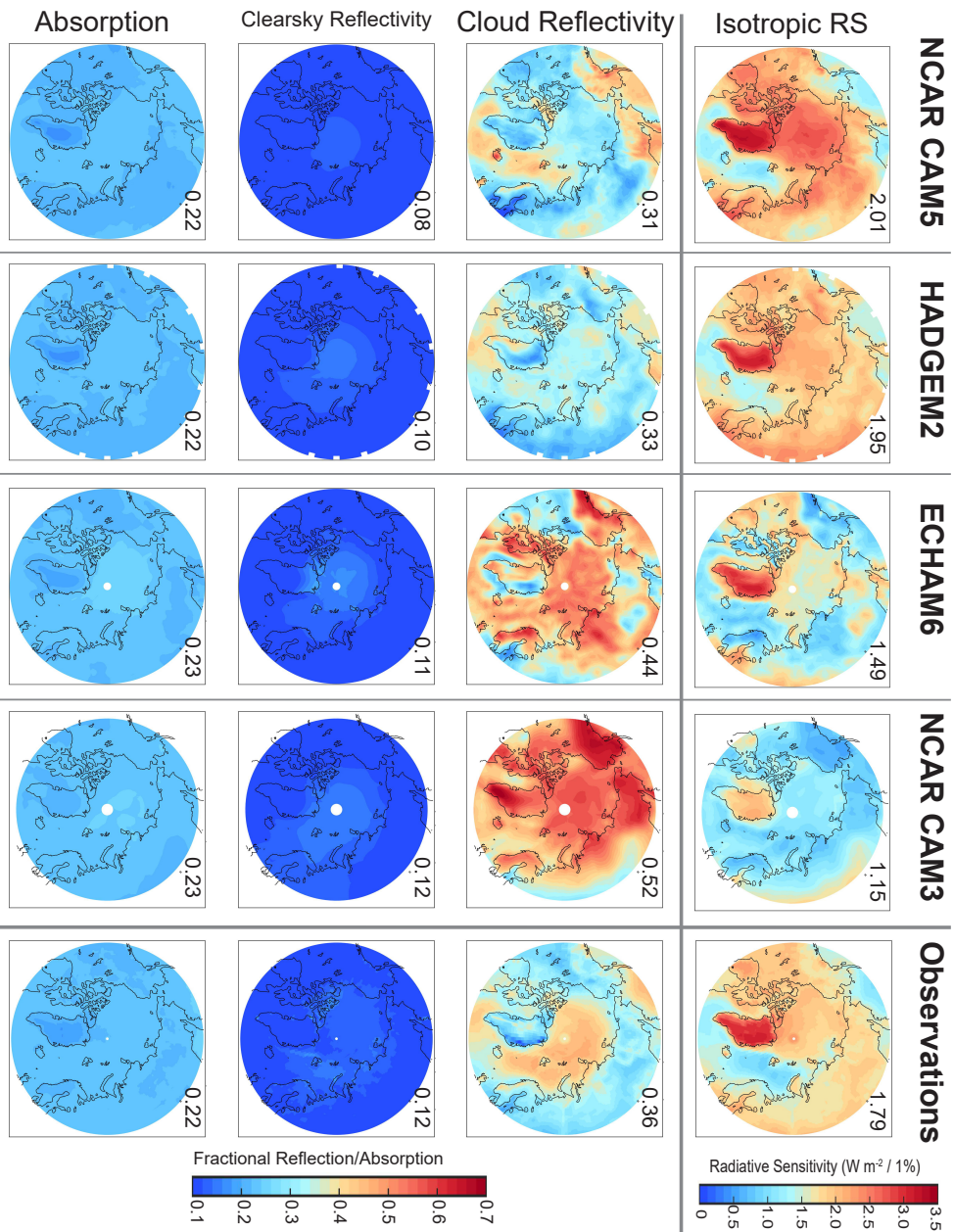


FIG. 5: Comparison of the atmospheric opacity parameters that result from the application of the isotropic model to the Arctic summertime (MJJJA) mean state radiative fields in the different climate models and observations. (Top Row) All-sky RS repeated from Fig. 1. (Second Row) Cloud reflectivity defined as the isotropic reflectivity applied to the all-sky radiative fields minus that defined from the clear-sky fields with the latter shown in the third row. (Bottom row) All-sky absorptivity. The (full Arctic) domain average is shown in the upper right of each panel. The four models for which kernels are available are shown to the left columns and the observational calculation from CERES data is shown to the right.

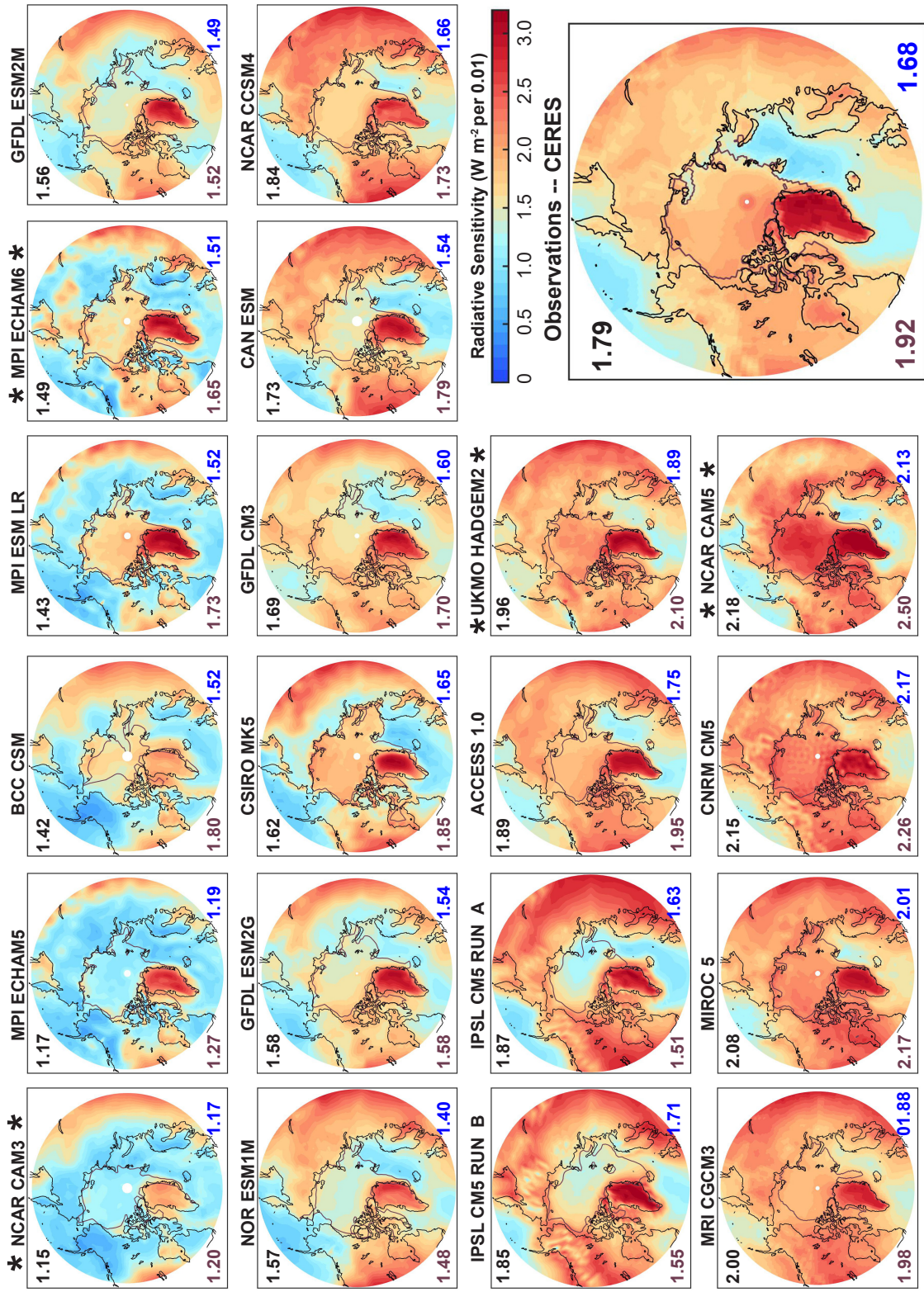


FIG. 6: Arctic summertime (MJJA) radiative sensitivity estimated using the isotropic model and the climatological radiation fields for CMIP5 historical simulations. Models are ordered as in reading a book (left to right then down) according to the domain average albedo feedback. Asterisks denote the models for which radiative kernel calculations are available that have been repeated from Fig. 1. The dark purple line shows the sea ice edge designated by the MJJA 50 % sea ice concentration contour. The full domain spatial average is shown in the upper left corner of each panel in black, the Arctic ocean average is shown in the lower right corner in blue and, the spatial average over the sea ice is shown in the lower left corner in purple. Observational estimates from CERES satellite data are shown in the bottom right panel.

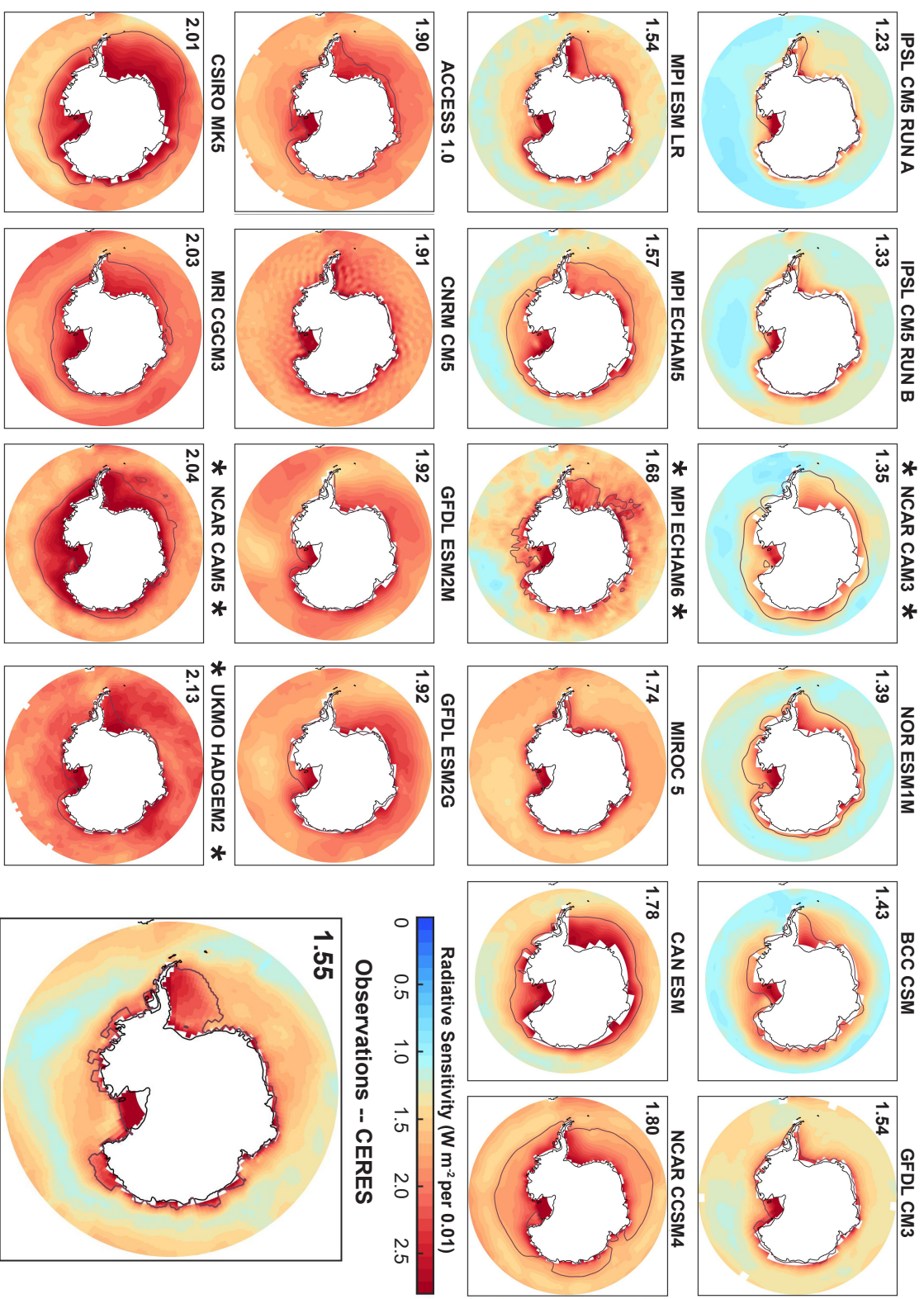


FIG. 7: As in Fig. 6 but for the NDJF Southern Ocean RS. Domain averaged surface albedo feedbacks exclude the Antarctic continent. The dark purple line shows the sea ice edge designated by the NDJF 50 % sea ice concentration contour. Note that models are ordered by Southern Ocean domain averaged RS and this order differs from that in Fig. 6. Asterisks denote the models for which radiative kernel calculations are available that have been repeated from Fig. 2.

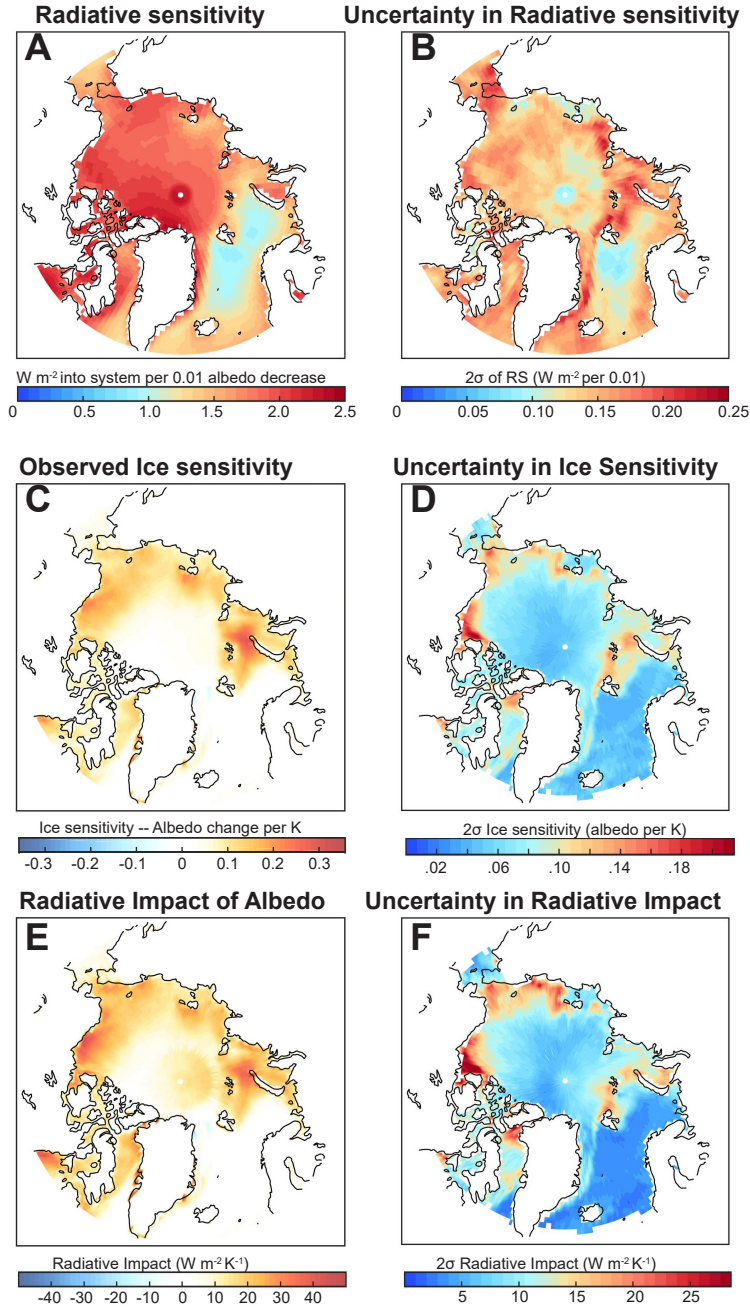
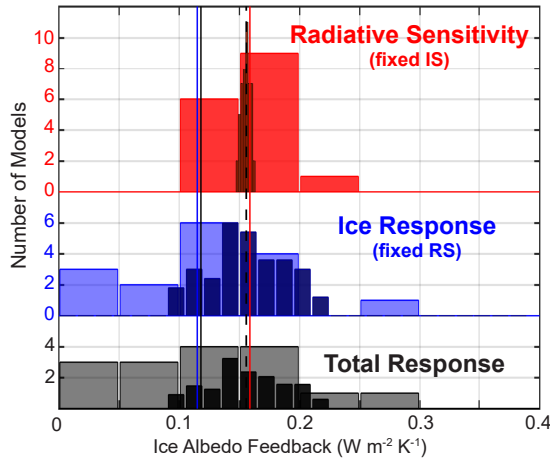
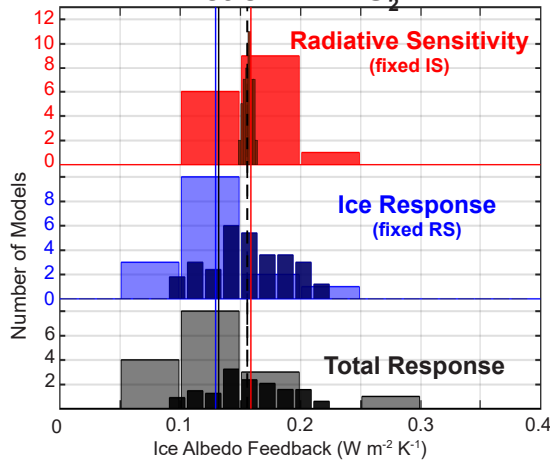


FIG. 8: Spatial maps of observational estimates of summertime (MJJA) radiative sensitivity (RS, top), ice sensitivity (IS, middle) and the radiative impact of surface albedo change ($RI_{T_{OA},\alpha}$, bottom). The RS is calculated from the isotropic shortwave model applied to the CERES data. The IS is calculated from Observational Best Estimate (OBE) surface albedo change between 1982 and 2016 divided by the global mean surface temperature change. The left panels show the central estimates of each quantity and the right panels show the uncertainty (2 standard deviations, σ) calculated from a Monte Carlo bootstrapping re-sampling with replacement as described in the Appendix.

A Arctic -- Historical Simulation



B Arctic -- 4XCO₂



C Southern Ocean -- 4XCO₂

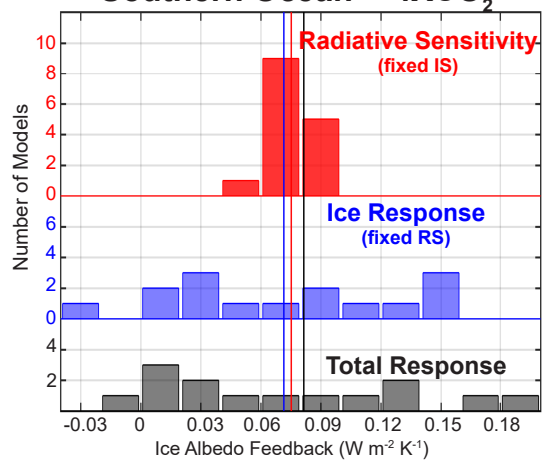


FIG. 9: Estimates of global (and annual) SIAF from climate models and observations using the radiative sensitivity (RS) from the isotropic model applied to the climatology and the change in surface albedo under external forcing normalized by the global mean temperature change. (Upper left) Arctic sea ice changes over the historical (2007 to 2016 minus 1982 to 1991 averages). The black bars show the CMIP5 model distribution using the climate model specific radiative sensitivity and ice changes, the blue bars show the distribution using the model specific sea ice changes and observational RS and the red bars show the distribution using the observational sea ice change and model specific radiative sensitivity. Solid vertical lines show the model mean of each distribution. The dashed vertical line shows the observational estimate. The overlaid dark and thinner distribution shows the histogram of observational estimates of ice albedo feedback calculated from a Monte Carlo re-sampling of subsets of the ice albedo data and radiative data; the black distribution shows the impact of uncertainties in the observational RS and IS combined, the blue distribution shows the impact of the IS uncertainty only and the red shows the impact of the RS uncertainty only. (Lower Left) As in the above panel except using the modeled changes in the 4XCO₂ simulations. (Lower Right) Distribution of surface albedo feedback in the Southern Ocean diagnosed from 4XCO₂ normalized sea ice changes. Because the observational estimate of sea ice changes over the historical simulation is not statistically significant, the red distribution is calculated from the model specific radiative sensitivity and the model mean normalized sea ice change.

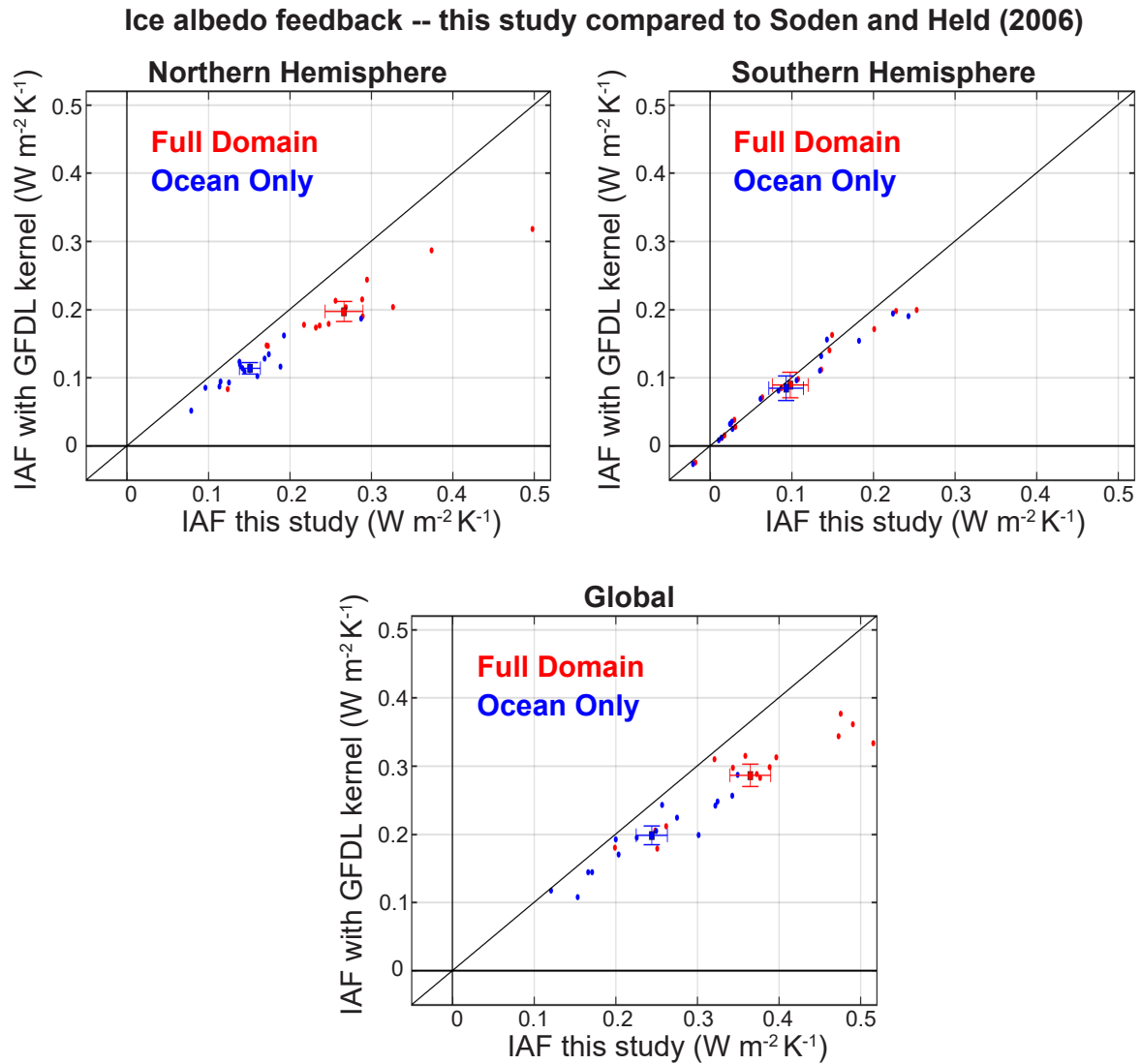


FIG. 10: Comparison of ice albedo feedback calculated from CMIP5 4XCO₂ using (ordinate) the method of Soden and Held (2006b) with RS in all models set to the GFDL surface albedo kernel versus (abscissa) the method introduced here with RS calculated from the model specific climatological radiative fluxes via the isotropic model. The blue markers show the contribution of the ocean domain only and the red markers show the full domain. All values shown are the contribution to the global mean. Dots show individual models and filled squares show the ensemble average with bars showing one standard deviation of the mean. The upper left shows the NH, the upper right the SH and the bottom the global mean. The black line is the 1:1 line.

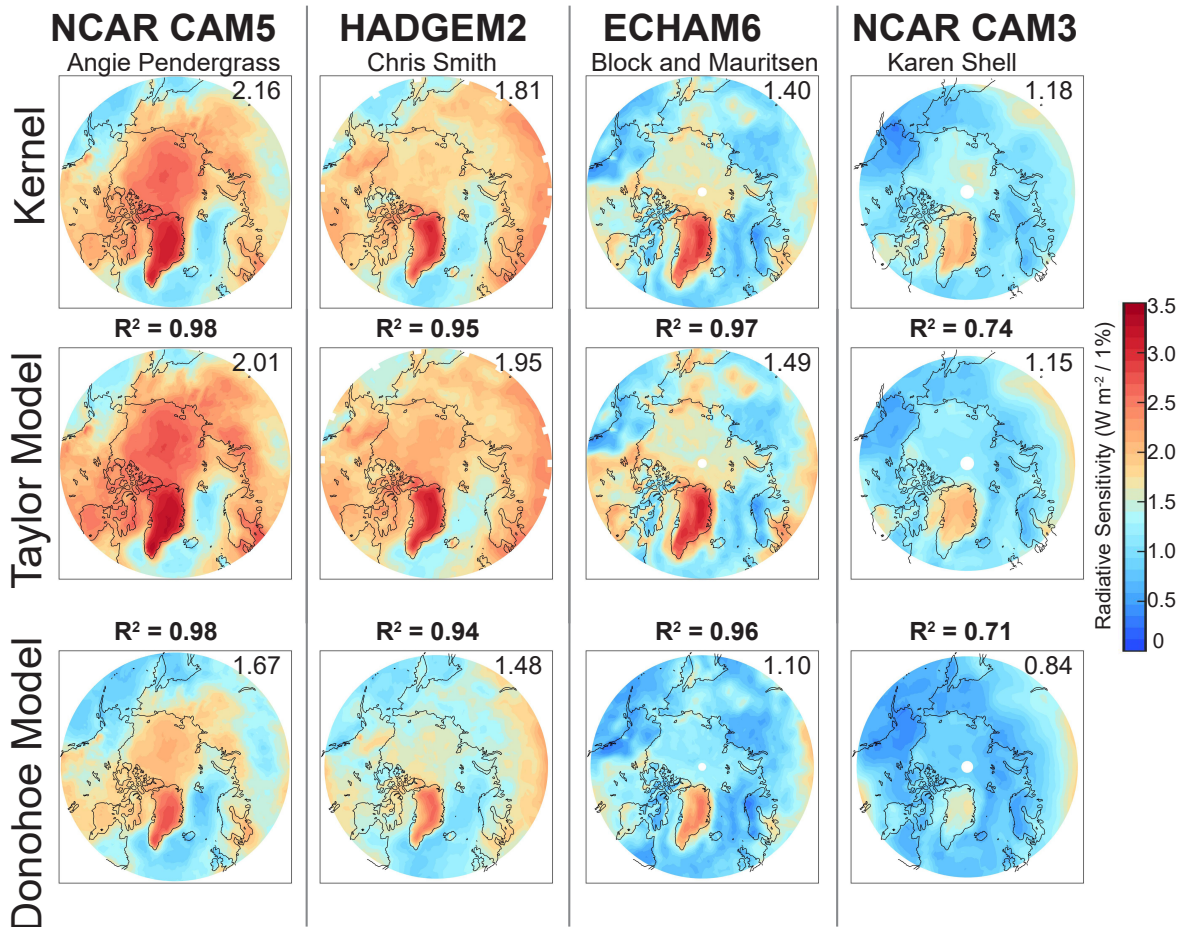


Fig. A1: Arctic summertime (MJJA) surface albedo radiative sensitivity (RS) calculated from radiative kernels (upper panels) and estimated from the climatological radiative fields using the idealized isotropic radiation model of T07 (middle panels) and (Donohoe and Battisti 2011) in the same models (bottom panels). The squared spatial correlation coefficient between the kernel isotropic methods in the same model are provided in the middle and the Arctic domain averaged values are shown in the upper right of each panel.

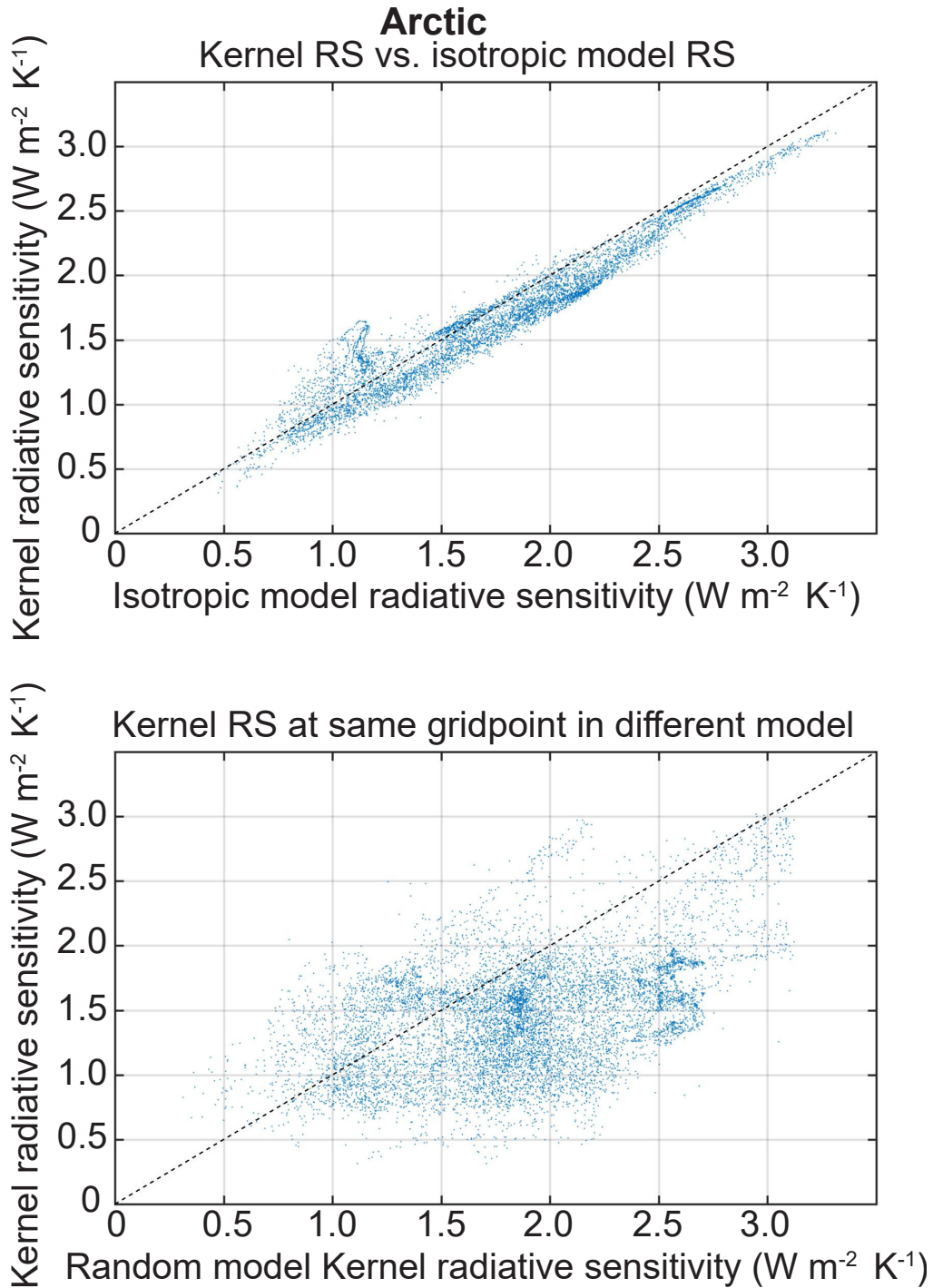


Fig. A2: (Top panel) Scatter plot of MJJA radiative sensitivity calculated by (ordinate) radiative kernels and (abscissa) the isotropic model from the mean state in the same climate model. All four climate models and Arctic gridpoints considered collectively. (Bottom panel) Scatter plot of MJJA radiative sensitivity calculated from radiative kernels in one model versus the radiative sensitivity calculated from radiative kernels in a different model (selected at random). The dashed black line shows the 1:1 line.

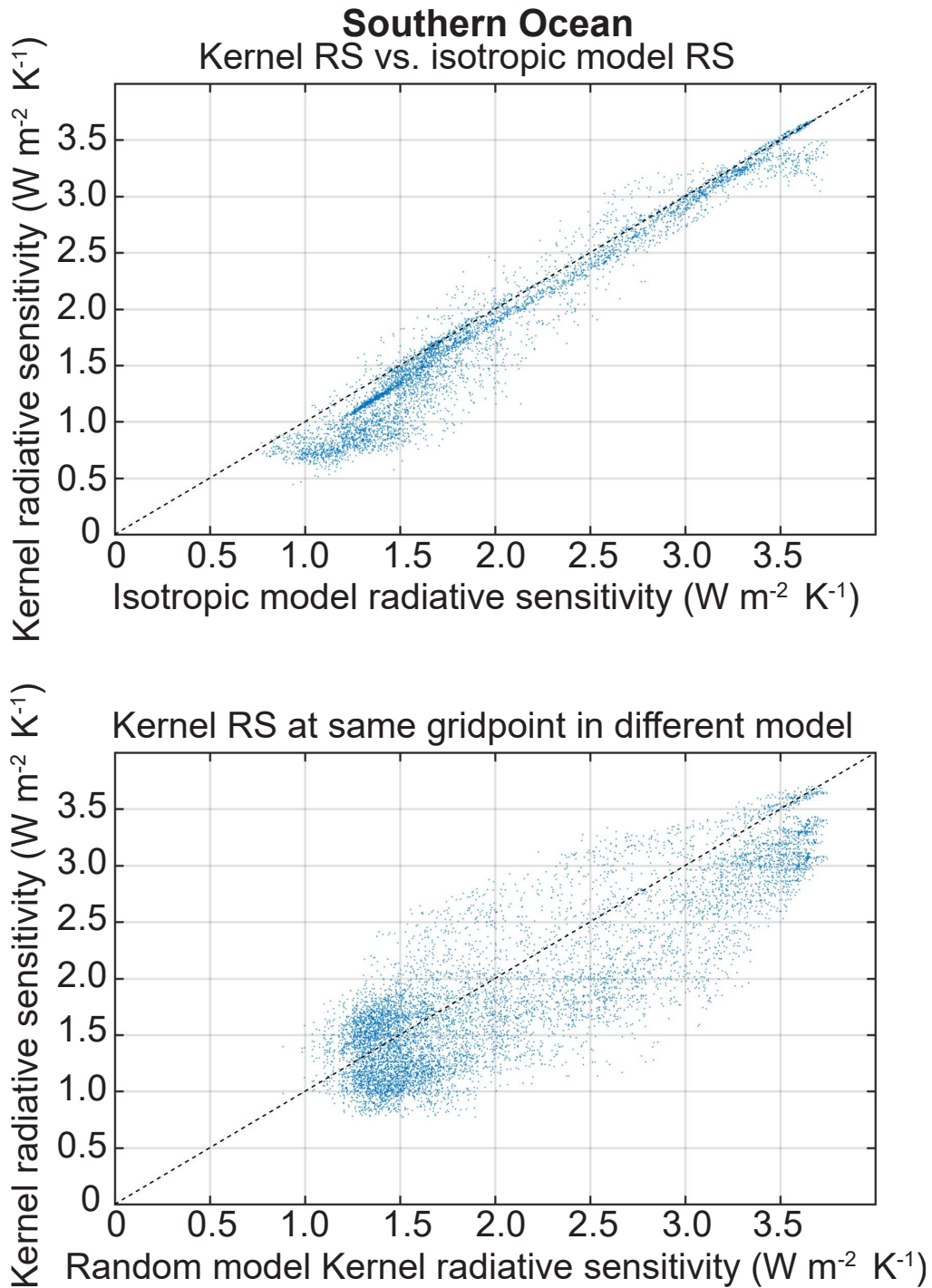


Fig. A3: (Top panel) Scatter plot of NDJFM radiative sensitivity calculated by (ordinate) radiative kernels and (abscissa) the isotropic model from the mean state in the same climate model. All four climate models and Southern Ocean gridpoints considered collectively. (Bottom panel) Scatter plot of NDJF radiative sensitivity calculated from radiative kernels in one model versus the radiative sensitivity calculated from radiative kernels in a different model (selected at random). The dashed black line shows the 1:1 line.

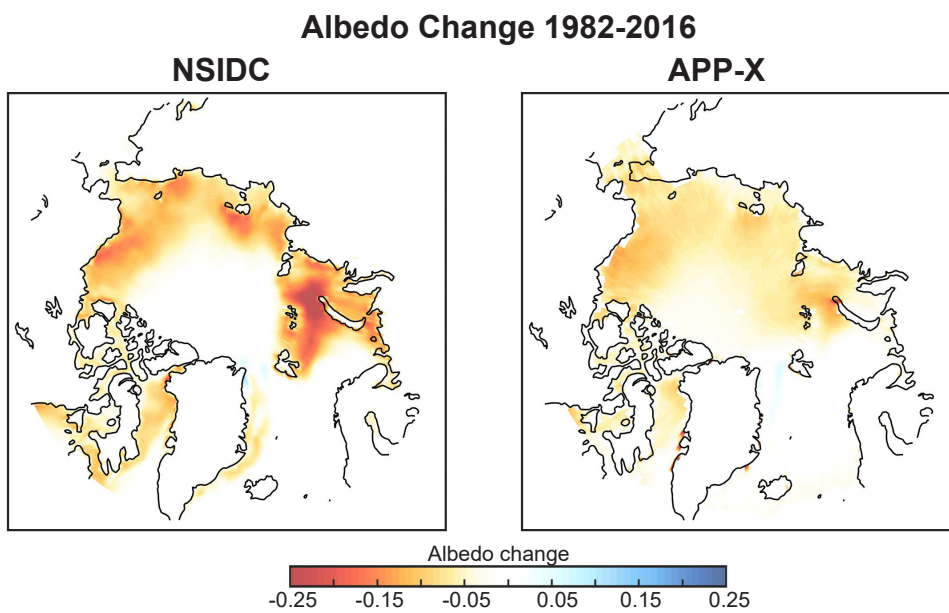


Fig. A4: Comparison of the (MJJA) surface albedo changes (1982-2016) calculated from the NSIDC sea ice concentration data (left) and the APP-x surface albedo data (right).