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1 **Sea Ice Loss, Water Vapor Increases, and Their Interactions with**
2 **Atmospheric Energy Transport in Driving Seasonal Polar Amplification**

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7 ABSTRACT: The ice-albedo feedback associated with sea ice loss contributes to polar amplifi-
8 cation, while the water vapor feedback contributes to tropical amplification of surface warming.
9 However, these feedbacks are not independent of atmospheric energy transport, raising the pos-
10 sibility of complex interactions that may obscure the drivers of polar amplification, in particular
11 its manifestation across the seasonal cycle. Here, we apply a radiative transfer hierarchy to the
12 idealized Isca climate model coupled to a thermodynamic sea ice model. The climate responses
13 and radiative feedbacks are decomposed into the contributions from sea ice loss, including both
14 retreat and thinning, and the radiative effect of water vapor changes. We find that summer sea ice
15 retreat causes winter polar amplification through ocean heat uptake and release, and the resulting
16 decrease in dry energy transport weakens the magnitude of warming. Moreover, sea ice thinning
17 is found to suppress summer warming and enhance winter warming, additionally contributing to
18 winter amplification. The water vapor radiative effect produces seasonally symmetric polar warm-
19 ing via offsetting effects: enhanced moisture in the summer hemisphere induces the summer water
20 vapor feedback and simultaneously strengthens the winter latent energy transport in the winter
21 hemisphere by increasing the meridional moisture gradient. These results reveal the importance
22 of changes in atmospheric energy transport induced by sea ice retreat and increased water vapor to
23 seasonal polar amplification, elucidating the interactions among these physical processes.

24 **1. Introduction**

25 Both observations (Serreze et al. 2009; Screen and Simmonds 2010a) and simulations forced by
26 increased CO₂ concentrations (Manabe and Wetherald 1975; Holland and Bitz 2003; Hahn et al.
27 2021) exhibit a strong surface warming in the polar regions compared to the global average. This
28 phenomenon known as polar amplification is seasonally asymmetric, reaching its maximum during
29 winter and minimum during summer in observations (Serreze et al. 2009; Screen and Simmonds
30 2010b) and general circulation models (GCMs; Deser et al. 2010; Hahn et al. 2021). Since
31 the projected polar changes are large and are posited to have consequences for global climate,
32 understanding their causes is a central goal of climate science. However, the mechanisms that
33 promote the dramatic seasonality of polar amplification, and the interactions among processes
34 across the seasonal cycle, are still under debate.

35 Sea ice processes have long been suggested to be the dominant driver of polar amplification via
36 the ice-albedo feedback (Budyko 1969; Manabe and Wetherald 1975; Taylor et al. 2013). Under
37 global warming, sea ice retreat reduces the surface albedo, leading to a greater surface shortwave
38 absorption that warms the polar surface. Additionally, the stable lower troposphere in the polar
39 region inhibits vertical mixing. The resulting surface-amplified warming produces less outgoing
40 longwave radiation than vertically uniform warming, which requires a strong surface warming in
41 order to balance the given forcing at the top of atmosphere, a positive lapse rate feedback. At a
42 global scale, the combination of a positive lapse rate feedback at high latitudes and a negative lapse
43 rate feedback at low latitudes indicates a contribution to polar amplification (Pithan and Mauritsen
44 2014; Goosse et al. 2018; Stuecker et al. 2018; Boeke et al. 2021). Furthermore, since the polar
45 lapse rate feedback is highly correlated to surface warming, some studies emphasize the combined
46 mechanism of the positive lapse rate feedback and ice-albedo feedback (Graversen et al. 2014;
47 Feldl et al. 2017, 2020). Notably, these feedbacks operate in different seasons and are linked by
48 seasonal ocean heat storage and release (Boeke and Taylor 2018; Dai et al. 2019; Feldl et al. 2020;
49 Shaw and Smith 2022) and changes in effective surface heat capacity as melting ice transitions to
50 open ocean (Manabe and Stouffer 1980; Dwyer et al. 2012; Hahn et al. 2022).

51 Preferential increases in tropical humidity amplify warming in the tropics relative to the high
52 latitudes, thus acting against polar amplification. Nevertheless, polar increases in water vapor
53 promote polar warming, and that warming may in turn activate other feedbacks. Results from a

54 single column model suggest that the water vapor feedback induced by increased specific humidity
55 from remote sources produces considerable near-surface warming in high latitudes (Henry et al.
56 2021), which would manifest as a positive lapse rate feedback. Similarly, in a moist energy
57 balance model, the water vapor feedback amplifies other positive feedbacks in the polar region,
58 thus rising to predominance in driving polar amplification (Beer and Eisenman 2022). Through
59 idealized modeling, these studies suggest the role of water vapor in polar amplification may be
60 underappreciated: interactions among the water vapor feedback and other processes promote polar
61 warming regardless of the meridional structure of the feedback considered in isolation.

62 Although diagnostics applied to coupled model experiments indicate that the total atmospheric
63 heat transport makes small contributions to polar amplification (Pithan and Mauritsen 2014; Goosse
64 et al. 2018), the dry and moist components of energy transport changes are large and compensating
65 (Stuecker et al. 2018; Graversen and Langen 2019; Feldl et al. 2020; Hahn et al. 2021; Henry
66 et al. 2021; Taylor et al. 2022). Idealized model experiments have shown that a stronger surface
67 polar heat source, such as associated with a stronger surface albedo feedback, produces a stronger
68 decrease in poleward dry energy transport, offsetting polar warming (Feldl et al. 2017; Henry et al.
69 2021). This is consistent with the well-known anti-correlation between polar amplification and
70 changes in atmospheric energy transport (Hwang et al. 2011). On the other hand, an increase in
71 latent energy transport may have an outsized warming impact on the polar regions via the “water
72 vapor triple effect”, which includes the greenhouse effect of increased moisture and cloudiness, as
73 well as the latent heat release of water vapor condensation (Graversen and Burtu 2016; Baggett
74 and Lee 2017; Yoshimori et al. 2017; Graversen and Langen 2019; Taylor et al. 2022, their figure
75 12). Slightly counteracting this warming effect, a strong increase in the moist component of energy
76 transport, and a weak decrease in the dry component, reduces the polar lapse rate feedback (Feldl
77 et al. 2020).

78 While these prior works suggest an underappreciated polar warming role for water vapor and the
79 moist component of atmospheric energy transport, it remains unclear how they manifest across the
80 seasonal cycle and interact with the ice-albedo and lapse rate feedback. Here we use an idealized
81 GCM coupled to a thermodynamic sea ice model as a minimal model for probing the mechanism
82 of seasonal polar amplification, including the crucial effects of radiative feedbacks, sea ice retreat
83 and thinning, and atmospheric energy transport. Aquaplanet simulations with thermodynamic sea

84 ice, though relatively unexplored, have yielded important insights about the seasonality of polar
 85 amplification (Feldl and Merlis 2021) and the midlatitude storm track response to sea ice loss
 86 (Shaw and Smith 2022). We further advance the model configuration by pairing it with a radiative
 87 transfer hierarchy to isolate the roles of water vapor increases and sea ice loss, as well as their
 88 interactions with atmospheric energy transport across the seasonal cycle in driving polar amplified
 89 warming.

90 **2. Models and Method**

91 *2.1 Models and experimental design*

92 We use the Isca modeling framework (Vallis et al. 2018), based on the Geophysical Fluid
 93 Dynamics Laboratory (GFDL) dynamical core. The resolution of the atmospheric component is
 94 T42 ($\sim 2.8^\circ \times 2.8^\circ$) with 30 evenly spaced vertical levels, and it is coupled to a thermodynamic
 95 sea ice model following Zhang et al. (2022) and a 60-m mixed layer slab ocean. The convection
 96 scheme is a simplified Betts-Miller scheme described in Frierson et al. (2007). The insolation
 97 includes a seasonal cycle of a 360-day year with the equinoxes at days 90 and 270 but no diurnal
 98 cycle. The solar constant is 1360 W m^{-2} , with a circular Earth’s orbit (eccentricity=0) and an
 99 Earth-like obliquity of 23.439° . Our model configuration is an idealized aquaplanet with neither
 100 clouds, continents, nor ocean heat transport (i.e., Q-flux is zero). All experiments in this study are
 101 run for 70 years with 30 years of spin-up.

102 In order to investigate the role of the water vapor feedback, we conduct experiments with
 103 a hierarchy of radiative transfer schemes: gray radiation and Rapid Radiative Transfer Model
 104 (RRTM). The gray radiation scheme follows Frierson et al. (2006), in which longwave optical
 105 depth depends on latitude (ϕ) and pressure (p):

$$\tau_0 = \tau_e + (\tau_p - \tau_e) \sin^2 \phi ; \tau = \tau_0 \left[f_1 \frac{p}{p_s} + (1 - f_1) \left(\frac{p}{p_s} \right)^4 \right] \quad (1)$$

106 where τ_e and τ_p are surface values of longwave optical depth at the equator and pole, respectively,
 107 f_1 is the linear absorption factor, and p_s is the surface pressure. We tune the gray radiation
 108 parameters to achieve the same climatology as the RRTM simulations. Specifically, optical depth
 109 parameters are set to $(\tau_p, \tau_e) = (2.5, 5.8)$, and $f_1 = 0.2$. A top-of-atmosphere (TOA) coalbedo

110 profile, $0.7535 - 0.0345(\frac{3\sin^2\phi-1}{2})$, is applied to the insolation to scale the TOA net shortwave
 111 flux to match that in RRTM. The global warming scenario is simulated by multiplying τ_0 by a
 112 tuning value of 1.155 to reproduce the radiative forcing in the $4 \times \text{CO}_2$ scenario in RRTM. For
 113 the RRTM radiative transfer scheme (Mlawer et al. 1997), we follow the setup of the Model of an
 114 Idealized Moist Atmosphere (MiMA; Jucker and Gerber 2017). Importantly, since the calculations
 115 of radiative heating are based on the atmospheric temperature and humidity structure, this model
 116 includes a water vapor feedback. The CO_2 concentration is 300 ppm in the control run. We
 117 perform the global warming scenario by quadrupling the CO_2 concentration. The magnitude of the
 118 radiative forcings in the gray radiation and RRTM schemes are 9.29 and 9.28 W m^{-2} , respectively,
 119 calculated as the TOA radiative flux change in the perturbation simulations relative to simulations
 120 with the surface boundary fixed (i.e., the fixed-SST forcing; Hansen et al. 2005).

121 We implement sea ice thermodynamics from Zhang et al. (2022), which is based on Semtner
 122 (1976), in our slab ocean boundary rather than the Isca default prescribed sea ice representation,
 123 in which ice distribution does not depend on atmospheric or oceanic temperature. The sea ice
 124 thickness (h_i) is governed by the following equation:

$$L_i \frac{dh_i}{dt} = F_{\text{atm}} - F_{\text{base}} \quad (2)$$

125 where $L_i = 3.0 \times 10^8 \text{ J m}^{-3}$ is latent heat of fusion of ice, F_{atm} is the net energy flux exchange
 126 between surface and atmosphere, including radiative, sensible, and latent heat fluxes (F_{rad} , F_{SH} ,
 127 and F_{LH} , respectively), and F_{base} is the basal heat flux from the ocean mixed layer into the ice.
 128 F_{base} depends linearly on the temperature gradient between the mixed layer (T_{ml}) and the ice
 129 base (freezing point, $T_{\text{base}} = 273.15 \text{ K}$): $F_{\text{base}} = F_0(T_{ml} - T_{\text{base}})$, where $F_0 = 120 \text{ W m}^{-2} \text{ K}^{-1}$ is the
 130 ocean-ice heat exchange coefficient. The ocean mixed-layer temperature (T_{ml}) is determined by

$$\rho_w c_w h_{ml} \frac{dT_{ml}}{dt} = \begin{cases} -F_{\text{base}} & \text{where ice is present} \\ -F_{\text{atm}} & \text{under ice-free condition} \end{cases} \quad (3)$$

131 where $\rho_w = 1035 \text{ kg m}^{-3}$ is the density of sea water, $c_w = 3989.24 \text{ J kg}^{-1} \text{ K}^{-1}$ is the specific heat
 132 of water, and $h_{ml} = 60 \text{ m}$ is the ocean mixed layer depth. When ice is present, conductive heat flux
 133 through ice F_i is given by

$$F_i = k_i \frac{T_{\text{base}} - T_s}{h_i}, \quad (4)$$

134 and the surface temperature T_s is determined by a balance between the surface flux F_{atm} , which is
 135 a function of T_s , and the conductive heat flux through the ice F_i :

$$F_{\text{rad}} + F_{\text{SH}} + F_{\text{LH}} = k_i \frac{T_{\text{base}} - T_s}{h_i} \quad (5)$$

136 where $k_i = 2 \text{ W m}^{-1} \text{ K}^{-1}$ is the thermal conductivity of sea ice. Lastly, ice fraction (f_{ice}) is set to
 137 1 where the ice thickness (h_i) is greater than 0 and is set to 0 elsewhere. The surface albedo is a
 138 linear function of ice fraction: $\alpha = (1 - f_{\text{ice}})\alpha_{\text{ocn}} + f_{\text{ice}}\alpha_{\text{ice}}$; we use values of ocean and ice albedo
 139 $(\alpha_{\text{ocn}}, \alpha_{\text{ice}}) = (0.22, 0.45)$.

140 In addition to the fully interactive sea ice thermodynamics implementation described above, we
 141 apply the direct ice-nudging method to lock the ice thickness (and fraction) to deactivate sea ice
 142 loss, and lock only the ice fraction to disable the sea ice-albedo feedback. Ice fraction-locked
 143 simulations are run for RRTM control and forced CO_2 . Ice thickness-locked simulations are run
 144 for both radiation models (RRTM and gray) and for both control (CTL) and perturbation (PTB;
 145 forced CO_2 or optical depth) configurations. All locking experiments are locked at every model
 146 time step to the ice thickness and fraction from the RRTM control simulation with interactive sea
 147 ice. These eight simulations, illustrated schematically in Fig. 1, enable us to isolate different
 148 physical components of the forced climate response. The total response is comprised of the
 149 response to CO_2 , the water vapor radiative effect (WV), and interactive sea ice loss (box a;
 150 $RRTM_{\text{ice}} \text{PTB} - RRTM_{\text{ice}} \text{CTL}$). The response to CO_2 and WV in the absence of ice loss
 151 (box b) is $RRTM_{\text{ice-lock}} \text{PTB} - RRTM_{\text{ice-lock}} \text{CTL}$. The response to CO_2 , WV, and ice thickness
 152 changes (box c) is $RRTM_{\text{fraction-lock}} \text{PTB} - RRTM_{\text{fraction-lock}} \text{CTL}$. The response to CO_2 (box d) is
 153 $Gray_{\text{ice-lock}} \text{PTB} - Gray_{\text{ice-lock}} \text{CTL}$. All of these responses occur in the presence of temperature
 154 feedbacks (Planck and lapse rate).

155 The response to WV and sea ice loss in isolation from radiative forcing can then be determined
 156 by (b)-(d) and (a)-(b), respectively. Strictly, the response to the water vapor radiative effect will also
 157 include any differences in temperature feedbacks between RRTM and the gray radiation schemes.
 158 The temperature feedbacks in $RRTM_{\text{ice-lock}}$ are close to that in $Gray_{\text{ice-lock}}$ in the polar regions

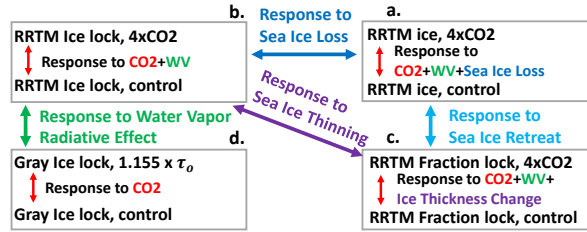


FIG. 1. The schematic of simulations.

159 but about 61% weaker (less negative) in the tropics (Fig. S1), comparable in magnitude to the
 160 water vapor feedback itself in the tropics. However, as detailed below in Section 2.2, the warming
 161 contributions associated with each feedback are calculated as the radiative response normalized
 162 by the global-mean Planck feedback and, in Sections 3.2 and 3.3, we will demonstrate that the
 163 contributions of the temperature feedbacks are within 0.5 K in both radiative schemes. The response
 164 to sea ice loss can further be decomposed into a response to sea ice retreat and the response to sea
 165 ice thinning, determined by (a)-(c) and (c)-(b), respectively. The response to sea ice thinning arises
 166 due to the difference between two experiments with locked ice fraction, but in one, ice thickness is
 167 free to change. The response to sea ice retreat captures the effect of albedo change in the absence
 168 of ice thickness change.

169 We note that sea ice interventions typically neglect to conserve energy or freshwater and have been
 170 shown to impose artificial heating in an idealized model (England et al. 2022). Given the present
 171 study’s specific objective of fixing ice fraction, and ice fraction and thickness, the ice-nudging
 172 method is the most straightforward approach. Nudging the ice to remain at its climatological state
 173 disables a latent heating mechanism, and we will show that this leads to a larger surface warming in
 174 summer than the simulations with interactive ice. We posit that this surplus warming is physically
 175 meaningful; in addition to the radiative feedbacks associated with sea ice retreat, the absorption of
 176 latent heat by melting ice influences the surface temperature response, and we quantify this effect
 177 via mechanism denial. Moreover, since the component responses sum to the total response with
 178 interactive sea ice—by construction with no residual—this effect is part of the total response.

179 *2.2 Warming contribution method*

180 The local atmospheric energy balance is used to quantify the contribution of physical mechanisms
 181 to spatial patterns of warming (Feldl and Roe 2013; Pithan and Mauritsen 2014; Goosse et al. 2018;
 182 Stuecker et al. 2018; Hahn et al. 2021):

$$F + (\lambda_p + \sum_i \lambda_i) \delta T_s + \delta \text{AHT}_d + \delta \text{AHT}_q + \delta \text{SEB} + \delta R_{\text{res}} = 0. \quad (6)$$

183 It includes radiative forcing (F), the radiative response associated with the Planck feedback ($\lambda_p \delta T_s$)
 184 and with other climate feedbacks ($\lambda_i \delta T_s$), anomalous dry and moist components of atmospheric
 185 heat transport ($\delta \text{AHT}_d, \delta \text{AHT}_q$), anomalous surface energy budget (δSEB), and a residual term
 186 (δR_{res}), all in units of W m^{-2} . As a reminder, there is no ocean heat transport in the aquaplanet and
 187 hence no associated warming contribution. The warming contributions are obtained by dividing
 188 each term in Eq.(6) by the global- and annual-mean Planck feedback ($\overline{\lambda_p}$, in $\text{W m}^{-2} \text{K}^{-1}$).

$$\delta T_s = -\frac{F}{\lambda_p} - \frac{\lambda'_p \delta T_s}{\lambda_p} - \frac{\sum_i \lambda_i \delta T_s}{\lambda_p} - \frac{\delta \text{AHT}_d}{\lambda_p} - \frac{\delta \text{AHT}_q}{\lambda_p} - \frac{\delta \text{SEB}}{\lambda_p} - \frac{\delta R_{\text{res}}}{\lambda_p} \quad (7)$$

189 where $\lambda'_p = \lambda_p - \overline{\lambda_p}$ is the departure of the Planck feedback from its global, annual mean. The
 190 surface energy budget term is given by the change in surface radiative and turbulent energy
 191 fluxes and represents, in the aquaplanet, changes in ocean heat uptake. We calculate the monthly
 192 contributions for each term. The winter is defined as December to February (DJF) for the northern
 193 polar regions (60° to 90°), while the summer is defined as June to August (JJA). The range of
 194 tropics is within 30° .

195 Climate feedbacks are calculated using the radiative kernel method (Shell et al. 2008; Soden
 196 et al. 2008), which decomposes the changes in TOA radiation into the contribution of individual
 197 climate variables. We calculate a temperature kernel for the gray radiation scheme and, separately,
 198 temperature and water vapor kernels for RRTM. Each kernel calculation is based on the respective
 199 control climate with interactive sea-ice simulated at 8x daily resolution. Temperature is perturbed
 200 by 1 K in each atmospheric layer and at the surface, and specific humidity by the logarithm of
 201 the anomaly corresponding to a 1-K warming, assuming constant relative humidity. The resulting

202 monthly kernels are weighted by the pressure thickness of each layer relative to 100 hPa, such that
 203 the units are $\text{W m}^2 \text{K}^{-1} 100 \text{ hPa}^{-1}$. The ice-albedo feedback is calculated using the approximate
 204 partial radiative perturbation (APRP) method (Taylor et al. 2007) instead of the kernel method.

205 Following Donohoe et al. (2020), the seasonal total atmospheric heat transport (AHT) is calcu-
 206 lated by the difference between monthly TOA and surface energy fluxes

$$\text{AHT}(\phi) = 2\pi a^2 \int_{-\frac{\pi}{2}}^{\phi} (F_{\text{TOA}} - F_{\text{sfc}} - \text{Storage}_{\text{atm}}) \cos \phi' d\phi' \quad (8)$$

207 where F_{TOA} and F_{sfc} are downward TOA radiative flux and the surface energy flux, respectively,
 208 and a is the radius of Earth. The atmospheric energy storage term ($\text{Storage}_{\text{atm}}$) is governed by the
 209 vertical integral of sensible and latent heat:

$$\text{Storage}_{\text{atm}} = \frac{1}{g} \int_0^{P_s} \frac{d}{dt} (c_p T + Lq) dp \quad (9)$$

210 where c_p is the specific heat of air at constant pressure and L is the latent heat of vaporization.
 211 The seasonal moist atmospheric heat transport (AHT_q) is calculated by monthly evaporation (E)
 212 minus precipitation (P):

$$\text{AHT}_q(\phi) = 2\pi a^2 \int_{-\frac{\pi}{2}}^{\phi} L(E - P - \text{Storage}_{\text{atm},q}) \cos \phi' d\phi' \quad (10)$$

$$\text{Storage}_{\text{atm},q} = \frac{1}{g} \int_0^{P_s} \frac{dq}{dt} dp. \quad (11)$$

213 The dry component is determined by the difference between total and moist AHT:

$$\text{AHT}_d(\phi) = \text{AHT}(\phi) - \text{AHT}_q(\phi). \quad (12)$$

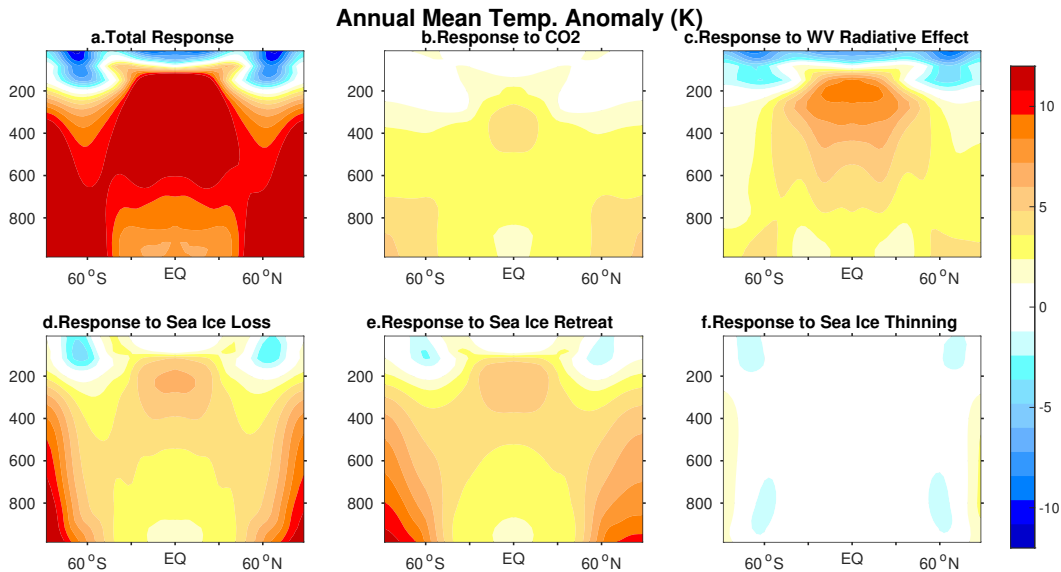
214 The global mean value of each term in Eq.(8) and Eq.(10) are removed to ensure zero transport at
 215 poles.

216 **3. Results**

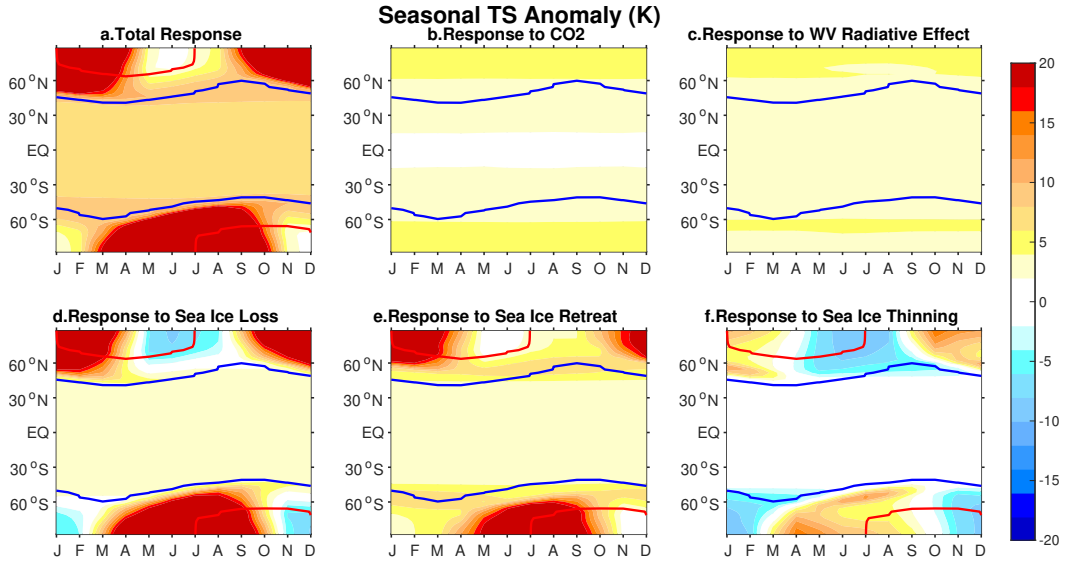
217 *3.1 Warming pattern and its seasonality*

218 The anomalous air temperature is decomposed into the warming associated with CO₂, with
 219 the water vapor radiative effect, and with sea ice loss (Fig. 2). In the annual mean, the total
 220 response (Fig. 2a) presents a classic global warming pattern, including tropical warming in the
 221 upper troposphere and polar amplification. The former is mainly contributed by the response to
 222 the water vapor radiative effect (Fig. 2c), and the latter is mainly contributed by the response to
 223 sea ice loss (Fig. 2d). The response to sea ice loss (Fig. 2d) is dominated by the response to sea
 224 ice retreat (Fig. 2e), which presents a strong surface-amplified and polar-amplified warming. By
 225 construction, the individual responses add to the total response with no residual warming. Fig. 3
 226 shows the seasonal anomalous surface temperature. As expected, the surface manifestation of the
 227 polar-amplified atmospheric warming occurs during winter (Fig. 3a). Since the responses to CO₂
 228 and water vapor radiative effect (Fig. 3b,c) have no apparent change throughout the seasonal cycle,
 229 the seasonality of polar amplification is dominated by the response to sea ice loss (Fig. 3d).

230 The asymmetric surface temperature pattern includes summertime cooling and wintertime warm-
 231 ing, which is comprised of the response to sea ice retreat and response to sea ice thinning. The



230 FIG. 2. Zonal and annual mean atmospheric temperature response (K) to (a) all drivers, (b) CO₂, (c) water
 231 vapor radiative effect, (d) sea ice loss, (e) sea ice retreat, and (f) sea ice thinning.



232 FIG. 3. Zonal mean seasonal surface temperature response (K) to (a) all drivers, (b) CO₂, (c) water vapor
 233 radiative effect, (d) sea ice loss, (e) sea ice retreat, and (f) sea ice thinning. Solid contours show the ice edge,
 234 defined where ice fraction is marginally non-zero (i.e., 1 %), in the $RRTM_{ice}$ control (blue) and $4 \times CO_2$ (red)
 235 simulations.

238 strong wintertime polar amplification presented in response to sea ice retreat (Fig. 3e) is initiated
 239 by ice-albedo feedback, which will be discussed in section 3.3. The response to sea ice thinning
 240 (Fig. 3f) presents summer cooling and winter warming. In summer, latent heating associated with
 241 sea ice melt is expected to inhibit the surface warming. The ice-locking simulation excludes this
 242 physical process, allowing a greater temperature increase with CO₂ forcing. Thus, the difference
 243 between the simulation in which only ice fraction is locked and the simulation in which ice thickness
 244 and fraction are locked presents a summer cooling response. In winter, for the surface temperature
 245 below the freezing point, increased conductive heat flux caused by thinning ice (not shown) results
 246 in surface warming, consistent with Hahn et al. (2022). The latent heat absorption by melting ice
 247 and increased conduction through thinner ice together explain the asymmetric seasonality of polar
 248 surface temperature when the climate warms.

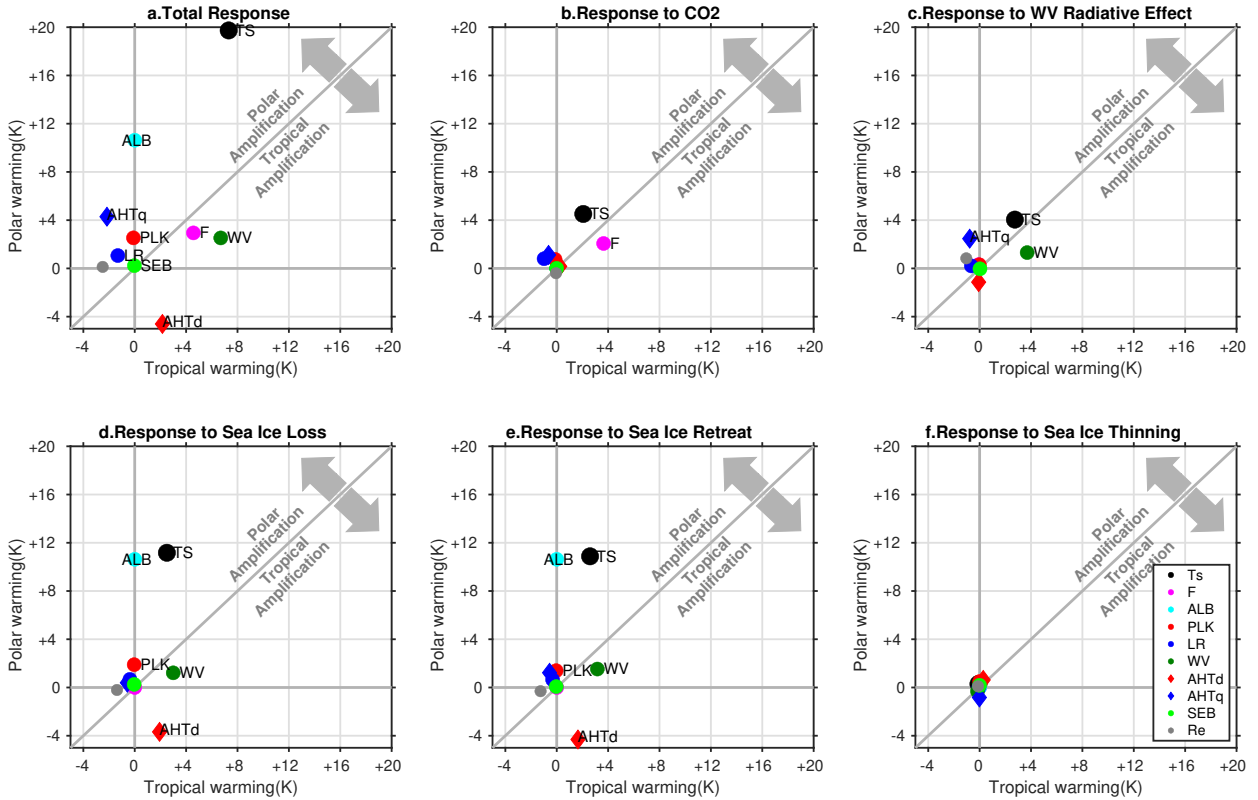
249 3.2 Annual mean warming contribution

250 To thoroughly investigate the detailed mechanisms of polar amplification characterized in the
 251 previous section, we use the warming contribution method (see section 2.2). Fig. 4 shows the

252 annual mean warming contributions associated with individual feedbacks and atmospheric energy
253 transports, with the total warming indicated by the black marker. The dominant contributor to
254 the total response is the ice-albedo feedback (Fig. 4a); in addition, the latent energy transport
255 contributes strongly to polar warming. The water vapor feedback and dry energy transport con-
256 tribute to tropical warming. Unlike previous studies that consider lapse rate feedback a primary
257 role in polar amplification (Pithan and Mauritsen 2014; Goosse et al. 2018; Stuecker et al. 2018;
258 Boeke et al. 2021), it contributes little in our simulations. Consistent with this modest contribution,
259 the vertical warming structure (Fig. 2a) in our model is relatively uniform throughout the polar
260 troposphere. Though the warming profile is bottom-heavy in DJF (Fig. S2), which would be
261 expected to promote a positive lapse rate feedback, the RRTM radiative kernel in that season is
262 weak (not shown) and hence lower tropospheric warming has little impact of the TOA radiative
263 flux. The weak kernel is likely related to the idealizations of our model, in particular the lack of
264 clouds. The warming profile is top-heavy in JJA (Fig. S2), producing a seasonally negative lapse
265 rate feedback.

273 The tropical amplification due to the water vapor primarily comes from the response to water
274 vapor radiative effect, as does the increase in poleward latent energy transport (Fig. 4c). Physically,
275 increased water vapor in the tropics supports both a tropically amplified water vapor feedback and
276 an increase in poleward latent energy transport via an enhanced meridional humidity gradient.
277 Although the water vapor feedback contributes to tropical warming, the increasing poleward latent
278 energy transport contributes to polar warming concomitantly. Hence, the combination of the two
279 causes a nearly uniform surface warming that slightly supports polar amplification. Moreover, the
280 Planck and lapse rate feedbacks contribute negligibly here, illustrating that the main difference
281 between RRTM and gray radiative schemes is dominated by the water vapor radiative effect rather
282 than temperature feedbacks, as mentioned in section 2.1.

283 Sea ice loss (Fig. 4d) is the primary physical mechanism leading to polar amplification. The
284 ice-albedo feedback plays the predominant role in polar amplification while the decreased poleward
285 dry transport, consistent with the reduced meridional temperature gradient, contributes to tropical
286 amplification. In addition, the enhanced water vapor feedback slightly supports tropical warming
287 because of the strong moistening for a given warming in higher initial temperature regions from
288 the Clausius-Clapeyron relation. Comparing Fig. 4e and Fig. 4f, we confirm that, in the annual

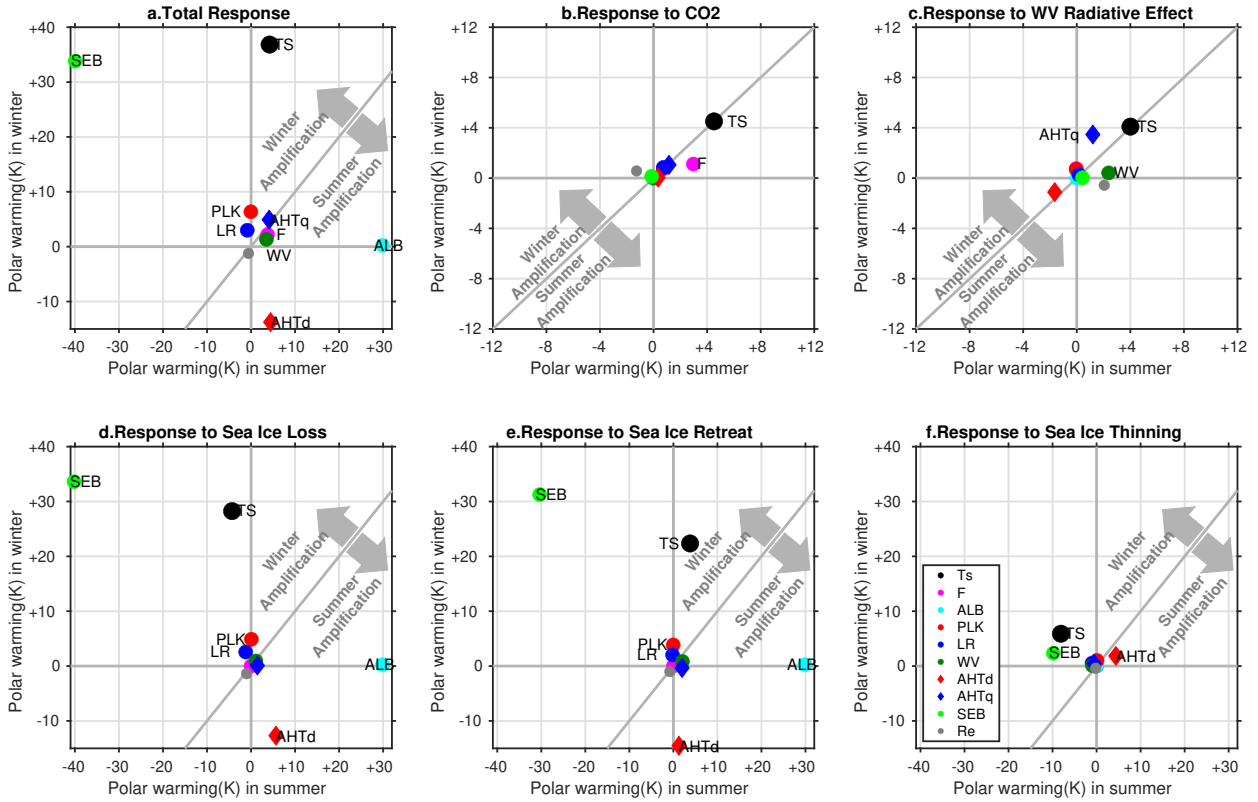


266 FIG. 4. Annual mean warming contributions (K) to (a) the total surface temperature change and the surface
 267 temperature change due to (b) CO₂, (c) water vapor radiative effect, (d) sea ice loss, (e) sea ice retreat, and (f)
 268 sea ice thinning. Warming contributions are shown for forcing (F), albedo feedback (ALB), Planck feedback
 269 (PLK), lapse rate feedback (LR), water vapor feedback (WV), change in dry and moist atmospheric heat transport
 270 (AHT_d and AHT_q), surface energy budget (SEB), and residual term (Re). The contributors above (below) the
 271 one-to-one line contribute to polar (tropical) amplification. The polar region is defined as 60°N to 90°N and the
 272 tropical region is defined within 30°.

289 mean, the mechanisms that produce surface warming in response to sea ice loss arise entirely from
 290 sea ice retreat. Sea ice thinning activates no annual-mean feedbacks.

291 3.3 Seasonal warming contribution

299 The seasonal polar warming contributions are calculated to identify the processes that promote
 300 polar amplification as a wintertime phenomenon (Fig. 5). In the total response (Fig. 5a), an
 301 increase in upward surface energy fluxes leads to winter warming. In contrast, the ice-albedo
 302 feedback and dry energy transport are the main contributors to summer amplification, and the rest



292 FIG. 5. Seasonal warming contribution (K) to (a) the total polar surface temperature change and the polar
 293 surface temperature change due to (b) CO₂, (c) water vapor radiative effect, (d) sea ice loss, (e) sea ice retreat, and
 294 (f) sea ice thinning. Warming contributions are shown for forcing (F), albedo feedback (ALB), Planck feedback
 295 (PLK), lapse rate feedback (LR), water vapor feedback (WV), change in dry and moist atmospheric heat transport
 296 (AHT_d and AHT_q), surface energy budget (SEB), and residual term (Re). The contributors above (below) the
 297 one-to-one line contribute to winter (summer) amplification. The polar region is defined as 60°N to 90°N. Note
 298 that the axes in b and c are different from the others.

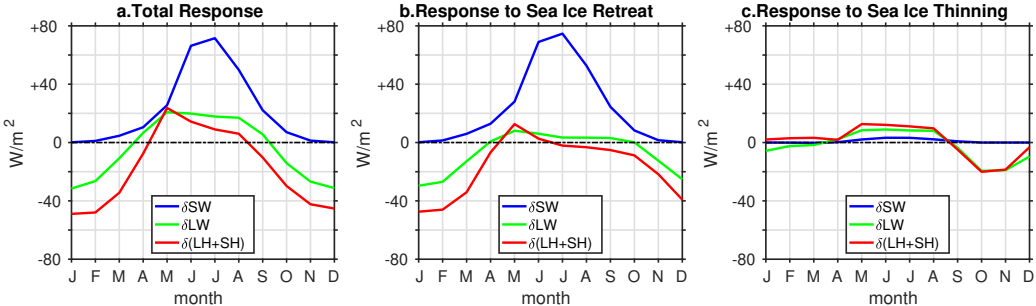
303 of the contributors play a minor role. This pattern of the surface energy budget overcompensating
 304 the ice-albedo feedback and dry energy transport arises from the response to sea ice loss (Fig.
 305 5d), and specifically the response to sea ice retreat (Fig. 5e). Physically, the ice-albedo feedback
 306 contributes to summer warming; however, it is balanced by ocean heat uptake, and the release of
 307 energy back to the atmosphere results in polar amplification during winter. The mechanism of
 308 seasonal ocean storage and release triggered by summertime ice-albedo feedback has also been
 309 found in fully coupled models (Feldl et al. 2020; Hahn et al. 2021; Jenkins and Dai 2021). The

310 decrease in dry energy transport stems from the reduced meridional temperature gradient associated
311 with wintertime polar warming.

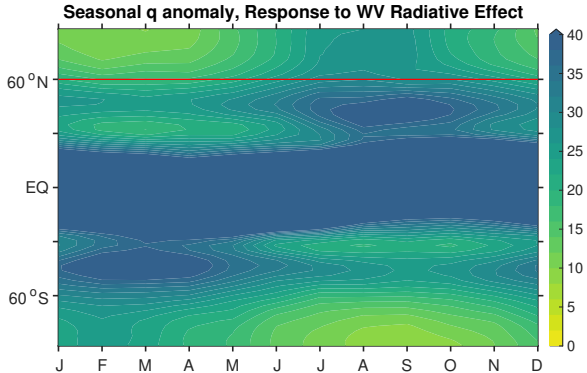
312 The surface energy budget in the polar region (poleward of 60°N) is further decomposed to
313 provide insights into the mechanisms of seasonal ocean-atmosphere exchange associated with sea
314 ice loss (Fig. 6). When sea ice retreats, the ocean absorbs a large amount of surface shortwave
315 flux during summer (blue line in Fig. 6b) and releases heat during winter. The enhanced winter
316 warming heats the lower atmosphere through both turbulent heat flux (latent and sensible heat flux)
317 and longwave radiation. The seasonal surface energy budget response to sea ice loss is consistent
318 with the ECHAM6 slab ocean model (Shaw and Smith 2022) and CESM1 coupled model (Jenkins
319 and Dai 2021). In contrast, sea ice thinning has no equivalent shortwave absorption, and the
320 small summer downward and winter upward anomalous longwave and turbulent heat flux (Fig. 6c)
321 are induced by the summer cooling and winter warming pattern in response to sea ice thinning.
322 Specifically, the response to sea ice thinning produces moderate winter amplification stemming
323 from the surface energy balance term (Fig. 5f), which is consistent with an increased conductive
324 heat flux through ice in winter and suppressed warming in summer. As discussed in Section 3.1,
325 the suppression of summer warming manifests here as a surface cooling because of the strong
326 warming in the ice-locking simulations. Summer cooling also slightly increases the poleward dry
327 energy transport.

328 The seasonal response to water vapor radiative effect (Fig. 5c) is much smaller than that to
329 sea ice loss (note the different axes in Fig. 5b and 5c). Intriguingly, the cancellation between
330 the water vapor feedback and latent energy transport previously identified in the annual mean
331 analysis is also evident in the seasonal polar warming. We interpret this in terms of seasonal
332 vertically integrated atmospheric humidity changes in response to the water vapor radiative effect.
333 As shown in Fig. 7, the water vapor changes feature enhanced moisture in summer and an increased
334 meridional humidity gradient in winter. The change in meridional humidity gradient at 60°N is
335 $0.0041 \text{ g/m}^2/\text{km}$ in DJF, compared to $0.0023 \text{ g/m}^2/\text{km}$ in JJA. Though the water vapor feedback
336 and increased latent energy transport warm the polar regions in all seasons, the former is stronger
337 in summer than winter, while the latter is stronger in winter than summer, leading to no seasonal
338 asymmetry in polar warming in response to water vapor radiative effect (black circle in Fig. 5c
339 falls on 1:1 line). From a global perspective, this response is tied to the seasonal progression of

340 humidity changes, as the moistening tropics shift poleward toward the summer hemisphere and
 341 away from the winter hemisphere. Lastly, the seasonal contributions of Planck and lapse rate
 342 feedbacks are extremely small, as mentioned in section 2.1 and consistent with our interpretation
 343 of this response as due to the water vapor radiative effect.



344 FIG. 6. The decomposition of seasonal surface energy budget ($W m^{-2}$) response to (a) all drivers, (b) sea
 345 ice retreat, and (c) sea ice thinning in the polar region. The surface energy budget is comprised of shortwave
 346 (SW, blue), longwave (LW, green), and turbulent heat flux (latent and sensible heat, LH+SH, red). Downward is
 347 defined as positive.



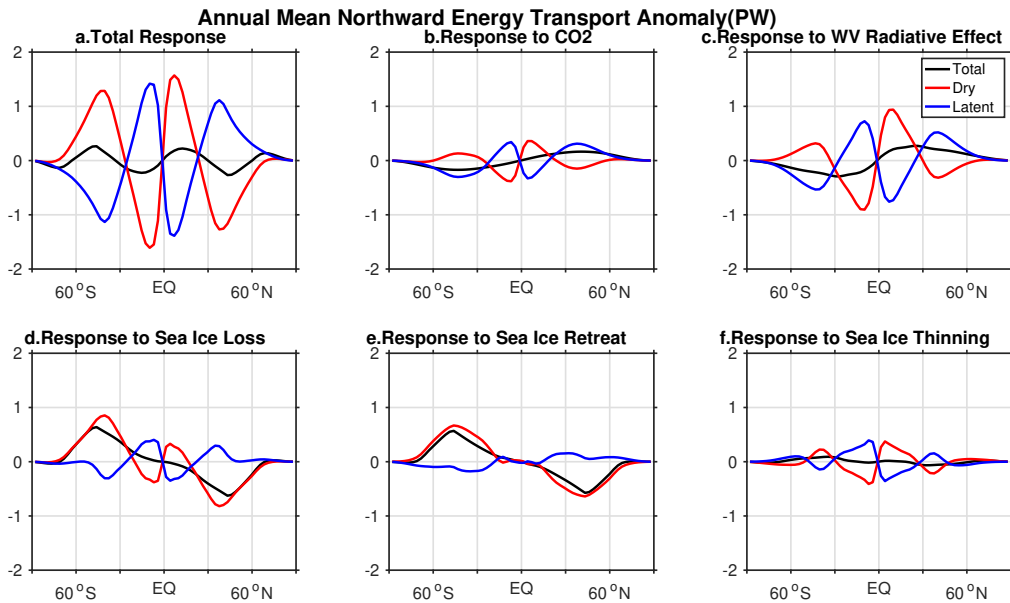
348 FIG. 7. Zonal mean seasonal vertically integrated specific humidity (g/m^2) response to water vapor radiative
 349 effect. The red line marks $60^\circ N$.

350 *3.4 Energy transport*

354 To further understand the role of remote interactions with polar amplification, the annual mean
 355 and seasonal atmospheric energy transport are diagnosed. Fig. 8 shows the annual mean northward
 356 atmospheric energy transport. In the total response (Fig. 8a), the total poleward atmospheric energy

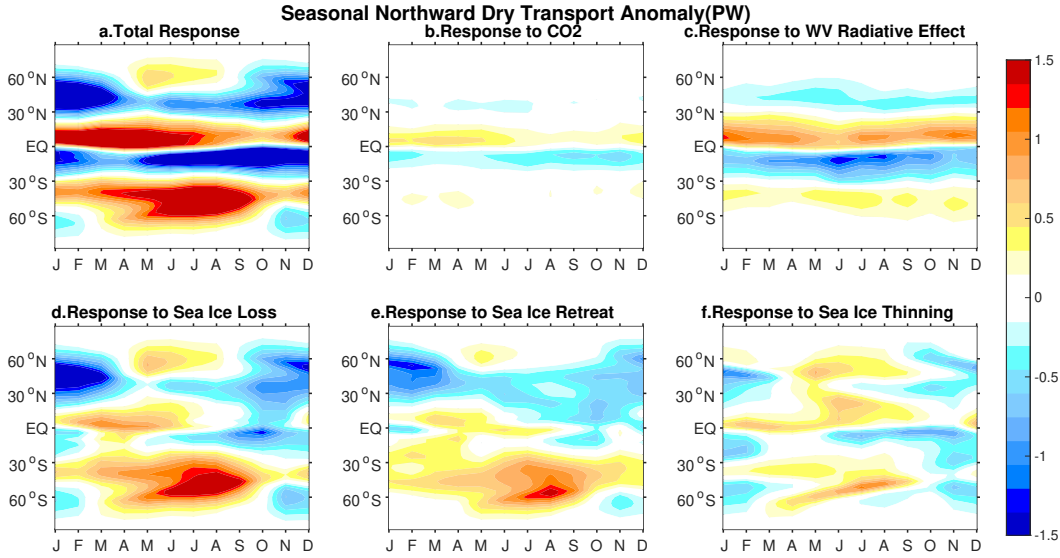
357 transport into the polar regions (at 60°) is close to zero, with dry and moist components tending to
 358 compensate one another, consistent with Hwang et al. (2011) and Graversen and Langen (2019),
 359 even without oceanic energy transport in our models. By separating the atmospheric heat transport
 360 into component responses, we find that the CO_2 and water vapor radiative effect (Fig. 8b and 8c)
 361 explain the increase in total heat transport, and the sea ice retreat (Fig. 8e) explains the decrease.
 362 This behavior is due to an overcompensation by the latent energy transport in response to CO_2
 363 and water vapor radiative effect and an overcompensation by the dry energy transport in response
 364 to sea ice retreat. Audette et al. (2021) further explain the decrease in poleward heat transport in
 365 response to sea ice loss as a warming of the returning moist isentropic circulation at high latitudes,
 366 while sea surface warming instead strengthens the moist isentropic circulation and hence poleward
 367 heat transport. Sea ice thinning does not strongly impact annual-mean poleward energy transport
 368 (Fig. 8f).

369 Fig. 9 and Fig. 10 show the seasonal poleward dry and moist components of atmospheric heat
 370 transport. The seasonality of dry energy transport in the total response (Fig. 9a) in the mid to
 371 high latitudes arises from the response to sea ice loss (Fig. 9d), with a substantial winter decrease



351 FIG. 8. Annual mean northward atmospheric energy transport (PW) response to (a) all drivers, (b) CO_2 , (c)
 352 water vapor radiative effect, (d) sea ice loss, (e) sea ice retreat, and (f) sea ice thinning. Total, dry, and moist
 353 component are shown in black, red, and blue, respectively.

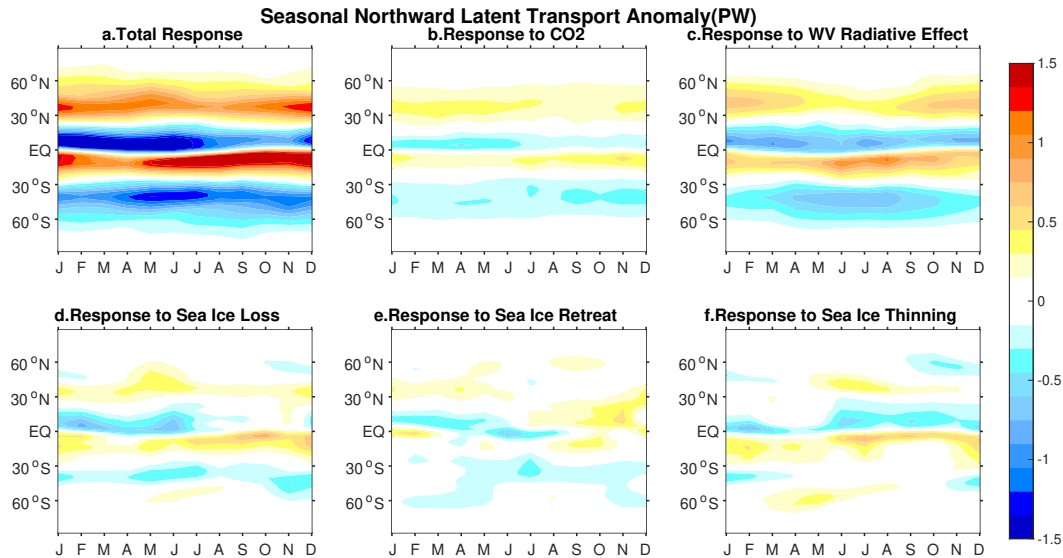
372 associated with sea ice retreat (Fig. 9e) and a summer increase and winter decrease associated with
 373 sea ice thinning (Fig. 9f). The seasonal pattern of dry energy transport is highly connected to the
 374 seasonal warming pattern. The strong winter weakening in response to sea ice retreat results from
 375 the strong winter amplification of polar warming (Fig. 3e). The weaker summer strengthening
 376 and winter weakening in response to sea ice thinning trace back to the weaker summer cooling and
 377 winter warming (Fig. 3f). In sum, the dry energy transport changes can be interpreted in terms
 378 of the seasonal polar warming in a straightforward manner. For total changes in latent energy
 379 transport (Fig. 10a), the seasonality in the mid to high latitudes is dominated by the water vapor
 380 radiative effect (Fig. 10c). In particular, the slight winter maxima that emerges in the absence of
 381 sea ice loss is consistent with the seasonally enhanced meridional moisture gradient (Fig. 7), and
 382 is also apparent in the change in northward energy transport at 60°N (Fig. S3).



383 FIG. 9. Seasonal northward dry component of atmospheric energy transport (PW) response to (a) all drivers,
 384 (b) CO₂, (c) water vapor radiative effect, (d) sea ice loss, (e) sea ice retreat, and (f) sea ice thinning.

387 4. Summary and Discussion

388 We use the Isca modeling framework (Vallis et al. 2018) coupled to a thermodynamic sea ice
 389 model (Semtner 1976; Zhang et al. 2022) to investigate the cause of polar amplification forced
 390 by increased greenhouse gas concentrations. By comparing the model with different radiative
 391 schemes and sea ice locked to its seasonally varying climatological control state, we separate the



385 FIG. 10. Seasonal northward moist component of atmospheric energy transport (PW) response to (a) all
 386 drivers, (b) CO₂, (c) water vapor radiative effect, (d) sea ice loss, (e) sea ice retreat, and (f) sea ice thinning.

392 total climate response into the response to CO₂, water vapor radiative effect, sea ice retreat, and
 393 sea ice thinning. Furthermore, for each response, a feedback analysis is performed to quantify
 394 the warming contributions associated with particular physical processes, such as the ice-albedo
 395 feedback, the water vapor feedback, ocean heat uptake, and dry and latent atmospheric energy
 396 transports.

397 The ice-albedo feedback plays the dominant role in polar amplified warming. The summertime
 398 ice-albedo feedback contributes to wintertime polar amplification through ocean heat uptake and
 399 release. Accompanied by the exposed ocean, the enhanced upward longwave and turbulent heat
 400 flux during winter heats the lower troposphere and induces a strong decrease in poleward dry
 401 energy transport and a weakly positive polar lapse rate feedback. The minor contribution of the
 402 lapse rate feedback to polar warming, in contrast to previous studies (Pithan and Mauritsen 2014;
 403 Goosse et al. 2018; Stuecker et al. 2018; Boeke et al. 2021), likely results from more uniform
 404 atmospheric warming in our idealized model and a weak TOA radiative response to that warming
 405 (i.e., a weak temperature kernel). A more realistic lapse rate feedback may support a stronger polar
 406 amplification via the response to sea ice retreat, especially in winter, offset to some extent by a
 407 correspondingly larger decrease in dry energy transport. However, we do not expect it to become
 408 a strong contributor to the response to water vapor radiative effect, which exhibits only modest

409 surface-amplified warming in high latitudes (Fig. 2c). Overall, these ocean-atmosphere coupled
410 mechanisms are consistent with previous studies: The seasonal polar amplification is supported by
411 a more positive winter lapse rate feedback (Boeke et al. 2021; Feldl et al. 2020) and reduced by a
412 decreasing poleward dry energy transport (Jenkins and Dai 2021) via increased sensible and latent
413 heat fluxes into the lower atmosphere in winter (Boeke and Taylor 2018; Dai et al. 2019; Feldl et al.
414 2020; Shaw and Smith 2022).

415 Though sea ice thinning does not contribute to annual-mean polar amplification or substantial
416 poleward energy transport changes, it does shape the seasonal climate changes. We interpret the
417 suppressed summer warming that occurs in response to sea ice melt as a consequence of latent heat
418 absorption, while winter sea ice thinning causes warming through increasing conductive heat flux.
419 Sea ice thinning thus drives a stronger seasonality of polar warming, consistent with more idealized
420 studies using an energy balance model and gray radiation GCM in Feldl and Merlis (2021) and a
421 single-column sea ice model in Hahn et al. (2022). Furthermore, the seasonal temperature pattern
422 induced by sea ice thinning, which can only be simulated in a sea ice model that includes ice
423 thermodynamics, contributes to seasonality in the dry energy transport changes. A unique aspect
424 of our study is that, by separately locking ice thickness and fraction and locking ice fraction alone,
425 we are able to cleanly isolate and quantify the effect of ice albedo changes and ice thinning on the
426 surface temperature and atmospheric energy transport response across the seasonal cycle.

427 The water vapor feedback is widely considered a tropical amplification contributor (Pithan and
428 Mauritsen 2014; Goosse et al. 2018; Hahn et al. 2021). However, we find that the poleward
429 latent energy transport induced by enhanced water vapor at lower latitudes contributes to polar
430 amplification. The magnitude of the contribution of increased latent energy transport is comparable
431 to (even slightly larger than) that of the water vapor feedback in our annual mean analysis, leading
432 to a net polar amplification. These results are supported by Beer and Eisenman (2022), who
433 apply a feedback locking method in a moist energy balance model and find that the water vapor
434 feedback becomes the primary factor of polar amplification, due to its interaction with other positive
435 feedbacks. Russotto and Biasutti (2020) also conclude that latent energy transport, interacting with
436 water vapor feedback, plays an essential role in polar amplification. For our seasonal polar warming
437 diagnosis, summer amplification by the water vapor feedback is compensated by its induced increase
438 in poleward latent energy transport, which is larger in winter. Since this mechanism is illuminated

439 by ice-locking simulations, the water vapor feedback is precluded from causing sea ice loss in
440 our diagnostic framework, though, because it warms the polar regions in summer, it may help to
441 initiate the ice-albedo feedback. Notably, sea ice retreat provokes only small changes in latent
442 energy transport in our annual mean and seasonal analysis, with the largest transport increase into
443 the Arctic in late summer (Fig. S3e).

444 The increase in latent energy transport into polar regions in our simulations is largely seasonally
445 invariant. The total response of Arctic transport is comprised of a modest winter peak in response
446 to the water vapor radiative effect and a modest late summer peak in response to sea ice retreat
447 (Fig. S3). This suggests that the summer maximum in latent energy transport increases identified
448 in previous studies (McCrystall et al. 2021; Kaufman and Feldl 2022) is associated at least in
449 part with sea ice loss. However, we also note that our idealized aquaplanet configuration includes
450 a 60-m deep slab ocean and hence has a higher heat capacity than a model with continents. A
451 higher heat capacity leads to a delayed phase shift of the tropical temperature maximum (Dwyer
452 et al. 2012), the tropical moisture maximum (via the Clausius-Clapeyron relation), and all else
453 equal, the midlatitude moisture gradient. In response to the water vapor radiative effect, for
454 instance, a shallower mixed layer would likely produce a somewhat earlier peak in the latent energy
455 transport increase. While this may lead to stronger, summer-dominated seasonality of latent energy
456 transport increase in the total response, we do not expect it to alter our main result that the water
457 vapor radiative effect produces seasonally symmetric polar warming, because the peak water vapor
458 feedback would also be shifted to earlier in the season.

459 The idealized GCM used in this study does not include a representation of clouds and hence
460 our analysis omits the cloud feedback. Only about half of the literature supports that cloud
461 feedback contributes to polar amplification, while the other half supports that it contributes to
462 tropical amplification or is unsure (Previdi et al. 2021). This uncertainty comes from the complex
463 interaction between clouds and other processes. Low cloud formation by increasing turbulent heat
464 fluxes over the newly exposed ocean enhances the downward longwave flux (Kay and Gettelman
465 2009), which may lead to stronger polar warming in fall and winter. The condensational heating of
466 clouds by increasing turbulent heat fluxes during winter also mediates the impact of sea ice loss on
467 the vertical structure of Arctic warming (Kaufman and Feldl 2022). Clouds may also enhance the
468 impact of latent energy transport increases through their longwave effect, contributing to winter

469 polar warming (Taylor et al. 2022; Dimitrelos et al. 2023). Besides the local impacts, the negative
470 shortwave cloud feedback in the polar region may strengthen the poleward energy transport by
471 increasing the meridional temperature gradient.

472 In conclusion, we separate the effect of CO₂, water vapor, sea ice retreat, and sea ice thinning
473 to polar amplification in a hierarchy of idealized models. We confirm that the ice-albedo feedback
474 is the primary factor in annual-mean polar warming, ocean heat uptake accounts for the seasonal
475 delay, and the dry atmospheric heat transport is a passive response to surface warming patterns in all
476 cases. Sea ice loss is the essential physical process in shaping the seasonality of polar warming, with
477 about 57% of winter amplification contributed by sea ice retreat and 43% of winter amplification
478 contributed by sea ice thinning (calculated as the distance from the 1:1 line in Fig. 5e,f). Both
479 the water vapor feedback and latent energy transport changes are manifestations of a preferential
480 tropical and summer-hemisphere increase in humidity, with the winter hemisphere experiencing a
481 corresponding increase in the meridional moisture gradient. Thus the secondary contribution of
482 latent energy transport to annual-mean polar amplification is intrinsic to the tropical amplification
483 effect of the water vapor feedback. Our results highlight the importance of the interaction between
484 feedbacks and atmospheric energy transports on the seasonality polar amplification, and thus
485 improve understanding of its mechanisms.

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492 *Data availability statement.* Data and code supporting this study will be made available at
493 <https://github.com/pochunchung/Isca> upon publication.

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