

1           **Seasonal cycle of idealized polar clouds: large eddy**  
2                           **simulations driven by a GCM**

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10           **Key Points:**

- 11           • LES driven by time-varying large-scale forcing from an idealized GCM is used to  
12           simulate the seasonal cycle of Arctic clouds
- 13           • Simulated low-level cloud liquid is maximal in late summer to early autumn, and  
14           minimal in winter, consistent with observations
- 15           • Large-scale advection provides the main moisture source for cloud liquid and shapes  
16           its seasonal cycle

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**Abstract**

The uncertainty in polar cloud feedbacks calls for process understanding of the cloud response to climate warming. As an initial step, we investigate the seasonal cycle of polar clouds in the current climate by adopting a novel modeling framework using large eddy simulations (LES), which explicitly resolve cloud dynamics. Resolved horizontal and vertical advection of heat and moisture from an idealized GCM are prescribed as forcing in the LES. The LES are also forced with prescribed sea ice thickness, but surface temperature, atmospheric temperature, and moisture evolve freely without nudging. A semigray radiative transfer scheme, without water vapor or cloud feedbacks, allows the GCM and LES to achieve closed energy budgets more easily than would be possible with more complex schemes; this allows the mean states in the two models to be consistently compared, without the added complications from interaction with more comprehensive radiation. We show that the LES closely follow the GCM seasonal cycle, and the seasonal cycle of low clouds in the LES resembles observations: maximum cloud liquid occurs in late summer and early autumn, and winter clouds are dominated by ice in the upper troposphere. Large-scale advection of moisture provides the main source of water vapor for the liquid clouds in summer, while a temperature advection peak in winter makes the atmosphere relatively dry and reduces cloud condensate. The framework we develop and employ can be used broadly for studying cloud processes and the response of polar clouds to climate warming.

**Plain Language Summary**

The polar regions are changing rapidly. Clouds and their feedbacks remain uncertain due to small-scale unresolved processes in climate models, which contributes to uncertainties in polar climate projection. In order to understand the mechanisms that control polar clouds, we focus on their seasonal cycle in the current climate. We adopt an idealized framework for driving high-resolution simulations by a global climate model. With minimal components represented, we find similar features between the simulated and observed polar clouds. In particular, liquid clouds reach maximum in summer, which coincides with the summer peak in moisture advection from lower latitudes. Therefore, projection of polar clouds will depend on future changes in heat and moisture advection. This framework will allow us to study the response of polar clouds to climate warming.

## 48 1 Introduction

49 As the Arctic warms and sea ice cover declines, it is pressing to reduce the uncer-  
50 tainties associated with polar climate change. One of the processes that contributes to  
51 Arctic climate change is the cloud radiative feedback (Holland & Bitz, 2003; Vavrus, 2004;  
52 Graversen & Wang, 2009). Clouds, depending on their amount, phase composition (liq-  
53 uid and/or ice), and altitude have different radiative effects. Cloud feedbacks in polar  
54 regions differ from their frequently studied low-latitude counterparts because polar re-  
55 gions have little to no incoming shortwave radiation in winter, they generally have a high  
56 surface albedo from ice cover, and even low clouds in polar regions are often mixed-phase  
57 clouds. As a result, the net cloud radiative effect at the surface is positive. i.e., clouds  
58 warm the surface because their longwave radiative effect dominates, unlike in low lat-  
59 itudes, where their predominant effect is a cooling of the surface (Shupe & Intrieri, 2004).  
60 How this cloud radiative effect changes with climate, and thus feeds back onto climate  
61 change, importantly influences the trajectory of Arctic climate change, including the Arc-  
62 tic amplification of climate change (Pithan & Mauritsen, 2014; Kay et al., 2016).

63 The polar regions are characterized by large insolation variations and hence dis-  
64 play a robust seasonal cycle. During the polar night, convergence of advective heat fluxes  
65 and surface turbulent heat fluxes become the dominant energy sources for the polar at-  
66 mosphere. By contrast, insolation is a dominant factor during the polar day. The mag-  
67 nitude of Arctic amplification in response to increased greenhouse gas concentrations also  
68 displays marked seasonality. Reanalysis and climate models show the largest surface warm-  
69 ing in winter (Serreze et al., 2009; Screen et al., 2012), when shortwave feedbacks, for  
70 example, from ice or clouds are weak or absent. Models suggest that a positive longwave  
71 feedback from clouds contributes to the maximum winter warming (Bintanja & van der  
72 Linden, 2013; La n e et al., 2016; Yoshimori et al., 2014).

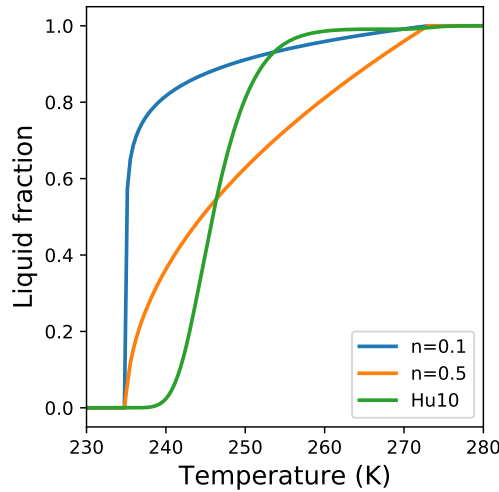
73 Early studies of Arctic clouds were often limited by the scarcity of observations.  
74 However, they have laid the groundwork for characterizing Arctic clouds and their sea-  
75 sonal cycle. For instance, Beesley and Moritz (1999) attempted to explain the seasonal  
76 variability of Arctic low clouds using a single-column model. In the model, large-scale  
77 forcing based on reanalysis for summer and winter produced a cloudy summer and a clear  
78 winter boundary layer (BL), which is consistent with the observed seasonal cycle of Arc-  
79 tic clouds. They also found that artificially shutting off surface evaporation in summer

80 does not eliminate low clouds. This suggests an important role for large-scale forcing in  
81 providing moisture and shaping the seasonal cycle of Arctic clouds. It is also essential  
82 to have the correct temperature dependency of cloud liquid and ice partitioning, as cloud  
83 ice crystals have a shorter residence time than liquid droplets.

84 Advances in satellite observations over the past decade have provided unprecedented  
85 3D coverage of clouds in polar regions. It is now known that liquid clouds persist through-  
86 out the year over the Arctic Ocean, and the low-level liquid-containing cloud fraction is  
87 highest in summer and autumn. Ice-dominated clouds, on the other hand, show max-  
88 imum cloud fraction in the winter upper troposphere (Cesana et al., 2012). Consistently,  
89 liquid water path reaches its maximum in August–September and minimum in winter  
90 (Lenaerts et al., 2017). However, it remains challenging for GCMs to correctly simulate  
91 the present-day seasonal cycle of clouds in the Arctic (Karlsson & Svensson, 2013; Tay-  
92 lor et al., 2019; Kretzschmar et al., 2019; Lenaerts et al., 2017). Recently, Baek et al. (2019)  
93 have shown that improved representation of atmospheric heat transport in a GCM al-  
94 leviate Arctic cloud biases in simulations.

95 Most studies on polar cloud feedbacks have used GCMs. However, because GCMs  
96 rely on cloud and turbulence parameterizations that are often tuned to observations in  
97 low latitudes (Brient et al., 2016), the reliability of inferences about polar cloud feed-  
98 backs from GCMs is questionable. Here we adopt a complementary approach that uses  
99 high-resolution large eddy simulations (LES) to explicitly resolve clouds and turbulence  
100 in the polar troposphere. Although LES have been frequently used to study the Arctic  
101 boundary layer (Klein et al., 2009; Morrison et al., 2011; Ovchinnikov et al., 2014; Savre  
102 et al., 2015), they have been rarely used to simulate the entire Arctic troposphere. The  
103 challenge is that LES alone cannot support large-scale circulations because of their lim-  
104 ited domain size. We use output from a GCM to provide the large-scale forcing neces-  
105 sary to drive LES. The idea is similar to using GCM output or reanalysis to drive a single-  
106 column model (e.g., Dal Gesso & Neggers, 2018), but without relying on cloud param-  
107 eterizations.

108 As a first step, we choose an idealized approach that only captures essential pro-  
109 cesses, including large-scale circulations, a closed surface energy budget, sea ice, and mixed-  
110 phase microphysics. Following Shen et al. (2020), we use a GCM with simple radiation  
111 and convection schemes but without clouds, to provide horizontal and vertical advection



**Figure 1.** Liquid fraction  $\lambda(T)$  as a function of temperature  $T$  used in one-moment bulk microphysics scheme.

112 of heat and moisture resolved by the GCM as forcing terms in the LES. Therefore, we  
 113 can treat each LES as an idealized single GCM column, with turbulent fluxes resolved  
 114 rather than being parameterized as in the GCM. The simplification in radiation allows  
 115 the two models to achieve closed energy budgets easily so that they have energetically  
 116 consistent, though not necessarily realistic, mean state climates. The LES can also pro-  
 117 vide training data for developing and refining GCM parameterizations (Schneider et al.,  
 118 2017; Shen et al., 2020).

119 We address the following questions: Can we reproduce the observed seasonal cycle  
 120 of Arctic clouds with our approach? How is the seasonal cycle influenced by large-scale  
 121 forcing and surface fluxes? In what follows we describe the modeling setup (section 2),  
 122 followed by results (section 3), discussion (section 4), and conclusions (section 5).

## 123 2 Model Setup

### 124 2.1 GCM

125 We use an idealized moist GCM to simulate large-scale dynamics of an Earth-like  
 126 atmosphere (Frierson et al., 2006, 2007; O’Gorman & Schneider, 2008). The GCM solves  
 127 the hydrostatic primitive equation with T42 spectral resolution in the horizontal and 32  
 128 unevenly spaced vertical sigma levels. The lower boundary of the GCM is a 5-m thick

129 mixed-layer ocean, and the surface energy budget is closed so that evaporation changes  
130 are constrained energetically by changes in other surface energy fluxes. Clouds are not  
131 represented in the GCM. Any grid-scale supersaturation is removed immediately to pre-  
132 cipitation, and there is no reevaporation of condensate. The GCM uses a gray radiation  
133 scheme with prescribed longwave optical thickness. The longwave optical thickness does  
134 not vary with water vapor content of the atmosphere, likewise for the shortwave radi-  
135 ation. Therefore, the GCM does not capture water vapor nor cloud feedbacks. The de-  
136 fault surface albedo in the aquaplanet configuration is 0.38, but in our case, it also de-  
137 pends on the presence of sea ice. We set the surface albedo to 0.3 for open water, and  
138 to 0.5 for sea ice. The surface roughness length is set to  $5 \times 10^{-3}$  m for momentum, and  
139 to  $1 \times 10^{-3}$  m for scalars.

140 One modification of the GCM specific to the current study is the saturation va-  
141 por pressure calculation. In order to obtain consistent thermodynamics, especially at low  
142 temperatures, we implemented a look-up table in the GCM to get saturation vapor pres-  
143 sure and its temperature derivatives, instead of using the default formulation in O’Gorman  
144 and Schneider (2008). The look-up table is obtained by integrating the Clausius-Clapeyron  
145 equation with specific latent heats that depend on temperature (see Equation (1) below).  
146 At GCM runtime, the values are determined by linearly interpolating the closest look-  
147 up table values. This treatment of saturation vapor pressure is consistent with the LES  
148 used in this study (Pressel et al., 2015).

149 We run the GCM with an obliquity of  $23.5^\circ$ , zero orbital eccentricity, and a sea-  
150 sonal cycle that has a period of 200 days per year. The seasonal cycle is shortened in or-  
151 der to reduce the computational cost of the LES simulations. We refer to the four sea-  
152 sons as the corresponding 50-day averages (e.g., spring is the first 50 days, summer is  
153 day 51–100, etc.). We set the longwave optical thicknesses at the equator to  $\tau_e = 7.2$   
154 and at the pole to  $\tau_p = 1.8$ . We run the GCM for 11 years into an approximate sta-  
155 tistical equilibrium and use the last year to provide forcing for the LES.

## 156 2.2 LES

157 We work with the Python Cloud Large Eddy Simulation code (PyCLES) (Pressel  
158 et al., 2015). The model uses an anelastic framework, and it ensures closed total water  
159 specific humidity  $q_t$  and specific entropy  $s$  budgets. PyCLES has been used successfully

160 to simulate subtropical marine BL clouds (Tan et al., 2016, 2017; Pressel et al., 2017;  
 161 Schneider et al., 2019) and deep convective clouds (Shen et al., 2020).

162 We use a one-moment mixed-phase microphysics scheme that follows Kaul et al.  
 163 (2015) and solves prognostic equations for snow and rain water specific humidity sep-  
 164 arately. Cloud condensates are diagnosed through a saturation adjustment procedure  
 165 from  $q_t$ . To partition the total condensate (saturation excess) between liquid and ice,  
 166 we use a phase partition function that depends on temperature  $T$  alone

$$167 \quad \lambda(T) = \begin{cases} 0 & \text{for } T < T_{\text{cold}}, \\ \left( \frac{T - T_{\text{cold}}}{T_{\text{warm}} - T_{\text{cold}}} \right)^n & \text{for } T_{\text{cold}} \leq T \leq T_{\text{warm}}, \\ 1 & \text{for } T_{\text{warm}} < T, \end{cases} \quad (1)$$

168 where  $T_{\text{warm}} = 273$  K and  $T_{\text{cold}} = 235$  K are the threshold temperatures for homoge-  
 169 neous melting and freezing (Kaul et al., 2015). The exponent  $n$  in the liquid fraction  $\lambda$   
 170 is taken to be 0.5 (instead of 0.1, a typically used value for Arctic stratocumulus, see Kaul  
 171 et al. (2015)). The corresponding liquid fraction is shown in Figure 1. Also plotted for  
 172 comparison is the observationally-derived curve from Hu et al. (2010). Using the latter  
 173 does not change the simulated seasonal cycle of clouds qualitatively, as will be discussed  
 174 in section 4.3.

175 Because the simulations are not limited to Arctic boundary layer clouds, we mod-  
 176 ified several processes in the microphysics scheme to be applicable to tropospheric clouds.  
 177 The slope parameter of the particle slope distribution function (PSDF) for snow uses the  
 178 default formulation in Grabowski (1998) instead of the empirical expression in Morrison  
 179 et al. (2011) (see also Appendix A in Kaul et al. (2015)). The intercept parameter of the  
 180 snow PSDF follows the expression in Sekhon and Srivastava (1970).

181 The LES uses the same gray radiation scheme as the GCM. Because the LES ref-  
 182 erence pressure can differ substantially from the GCM pressure at the same altitude, we  
 183 use the GCM pressure and air density to calculate the radiative tendency in the LES.  
 184 All LES simulations were conducted with a horizontal resolution of 400 m and a verti-  
 185 cal resolution that varies from 74 m near the surface to 420 m at the domain top. The  
 186 three-dimensional LES domain is 25.6 km wide and 18 km high. A sponge layer of 6 km  
 187 at the top of the domain is implemented to damp the velocity and scalar fluctuations  
 188 toward the domain-mean values. Simulated clouds below 10 km are insensitive to the  
 189 sponge layer depth. Therefore, we focus on the representation of the bottom 10 km of

190 the model domain. Like the idealized GCM, the lower boundary of LES is a 5-m thick  
 191 mixed-layer ocean with closed surface energy budget.

### 192 **2.3 Sea Ice Model**

193 We implemented a thermodynamic sea ice model similar to the Semtner (1976) “zero  
 194 layer” model. This model was initially developed for a GCM, but we now have imple-  
 195 mented in the LES too; however, we prescribe ice thickness in the LES using the GCM  
 196 output (see Section 2.5). This treatment approximates the specific heat of the ice to be  
 197 negligible, which implies that the temperature profile within the sea ice remains linear.  
 198 The present model differs from Semtner (1976) in that for simplicity the freshwater value  
 199 for the freezing point,  $T_m = 273.16$  K, is used at the surface and base of the ice, and  
 200 a constant latent heat of fusion of ice of  $L_i = 3.0 \times 10^8$  J m<sup>-3</sup> is adopted. Sea ice grows  
 201 at the base in winter, and ablation occurs at both the surface and the base in summer.  
 202 There is no surface snow layer and no horizontal sea ice motion.

203 Where the surface is ice covered ( $h_i > 0$ ), the sea ice thickness evolves according  
 204 to

$$205 \quad L_i \frac{dh_i}{dt} = F_{\text{atm}} - F_{\text{base}}. \quad (2)$$

206 Here the flux exchange between surface and atmosphere  $F_{\text{atm}}$  includes radiation and tur-  
 207 bulent sensible and latent heat fluxes ( $F_{\text{rad}}$ ,  $F_{\text{SH}}$ , and  $F_{\text{LH}}$ , respectively), defined to be  
 208 positive upward,

$$209 \quad F_{\text{atm}} = F_{\text{rad}} + F_{\text{SH}} + F_{\text{LH}}. \quad (3)$$

210 The basal heat flux  $F_{\text{base}}$  from the ocean mixed layer into the ice is taken to depend lin-  
 211 early on the temperature gradient between the mixed layer (at  $T_{\text{ml}}$ ) and the ice base (at  
 212 the melting temperature  $T_m$ ),

$$213 \quad F_{\text{base}} = F_0(T_{\text{ml}} - T_m),$$

214 using the coefficient  $F_0 = 120$  W m<sup>-2</sup> K<sup>-1</sup> as in Eisenman (2007). The surface tem-  
 215 perature of the ice  $T_s$  is determined implicitly by a balance between the surface flux  $F_{\text{atm}}$   
 216 (which is a function of  $T_s$ ) and the conductive heat flux through ice,

$$217 \quad F_{\text{atm}} = k_i \frac{T_m - T_s}{h_i},$$

218 except where this gives  $T_s > T_m$ , in which case instead we set

$$219 \quad T_s = T_m,$$



220 representing surface melt (Eisenman & Wettlaufer, 2009).

221 The ocean mixed-layer temperature  $T_{\text{ml}}$  is determined by

$$222 \quad \rho_w c_w h_{\text{ml}} \frac{dT_{\text{ml}}}{dt} = -F_{\text{atm}} \quad (4)$$

223 under ice-free conditions and

$$224 \quad \rho_w c_w h_{\text{ml}} \frac{dT_{\text{ml}}}{dt} = -F_{\text{base}} \quad (5)$$

225 where ice is present. Here  $\rho_w$  is the density of water,  $c_w$  is the specific heat of water, and  
 226  $h_{\text{ml}}$  is the constant ocean mixed-layer thickness. The representations of the surface fluxes  
 227 ( $F_{\text{rad}}$ ,  $F_{\text{SH}}$ , and  $F_{\text{LH}}$ ) do not explicitly depend on whether the surface is ice-covered or  
 228 ice-free, although they do depend on the surface temperature.

229 The transition from ice-free to ice-covered conditions happens when  $T_{\text{ml}}$  cools be-  
 230 low  $T_{\text{m}}$  during a GCM time step, in which case frazil ice growth is represented by set-  
 231 ting  $T_{\text{ml}} = T_{\text{m}}$  and assigning a positive value to  $h_i$  equal to this change in  $T_{\text{ml}}$  scaled  
 232 by  $L_i$ . Similarly, a transition from ice-covered to ice-free conditions occurs when  $h_i$  reaches  
 233 zero, at which point any additional net energy flux warms  $T_{\text{ml}}$ .

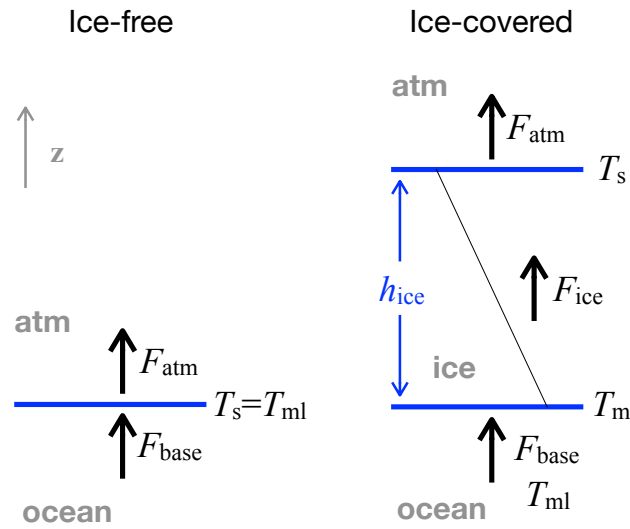
234 Note that because there is no lateral ocean energy flux (“ $Q$  flux”) in the present  
 235 setup,  $T_{\text{ml}}$  remains at  $T_{\text{m}}$  where ice is present, causing  $F_{\text{base}} = 0$ .

## 236 2.4 Large-Scale Forcing

237 In order to include large-scale dynamics in the limited-domain of LES, we use time-  
 238 varying large-scale fluxes simulated by the GCM. The details of the forcing framework  
 239 are described in Shen et al. (2020). In summary, we use LES to simulate a single grid  
 240 column of a GCM, but with processes that are parameterized in the GCM (e.g., convec-  
 241 tion, condensation, and boundary layer turbulence) resolved in the LES. The forcing terms  
 242 include horizontal and vertical advection of temperature and specific humidity, as well  
 243 as temperature tendencies due to numerical damping and spectral filtering in the GCM.

244 A major distinction between our forcing framework and that of Shen et al. (2020)  
 245 is the time-varying forcing. Instead of using the long-time mean tendencies, we use the  
 246 instantaneous tendencies from the GCM, updated every 6 hours. Therefore, the hori-  
 247 zontal advective  $q_t$  source term  $S_{\text{hadv}}$  becomes

$$248 \quad S_{\text{hadv}} = -\tilde{u}\partial_x\tilde{q}_t - \tilde{v}\partial_y\tilde{q}_t, \quad (6)$$



**Figure 2.** Schematics of the surface boundary conditions. In the GCM, the sea ice specific heat is taken to be zero, so that the temperature profile within the ice is linear.

249 and the vertical advective  $q_t$  source term  $S_{\text{vadv}}$  becomes

$$250 \quad S_{\text{vadv}} = -\tilde{w}\partial_z q_t. \quad (7)$$

251 Tildes ( $\tilde{\cdot}$ ) denote variables resolved on the GCM grid.

252 Like for the specific humidity, the horizontal advective temperature tendency  $J_{\text{hadv}}$   
253 is taken directly from the GCM,

$$254 \quad J_{\text{hadv}} = -\tilde{u}\partial_x \tilde{T} - \tilde{v}\partial_y \tilde{T}, \quad (8)$$

255 and the vertical advective temperature tendency  $J_{\text{vadv}}$  becomes

$$256 \quad J_{\text{vadv}} = -\tilde{w}\partial_z T - \tilde{w}\frac{g}{c_p}, \quad (9)$$

257 where  $g$  is the gravitational acceleration, and  $c_p$  is the specific heat of dry air. The source  
258 terms (6) and (7) are included in the prognostic equation for  $q_t$ , and the source terms  
259 (6)–(9) are included in the prognostic equation for  $s$  (Shen et al., 2020).

260 For horizontal momentum forcing ( $u$  and  $v$ ), we impose the GCM-resolved hori-  
261 zontal momentum tendencies on the LES momentum equations. This also differs from  
262 Shen et al. (2020), where the GCM large-scale pressure gradient is imposed.

263 The forcing fields are taken from GCM grid boxes closest to 70°N. This has more  
 264 relevance for the Arctic Ocean, given the aquaplanet nature of the idealized GCM. To  
 265 include synoptic-scale variability, we choose four grid points (0°, 90°, 180°, and 270° lon-  
 266 gitude) instead of using zonal-mean fields from the GCM. The results we present are av-  
 267 erages of the 4 simulated locations, which are statistically identical. We call this aver-  
 268 age the ensemble mean.

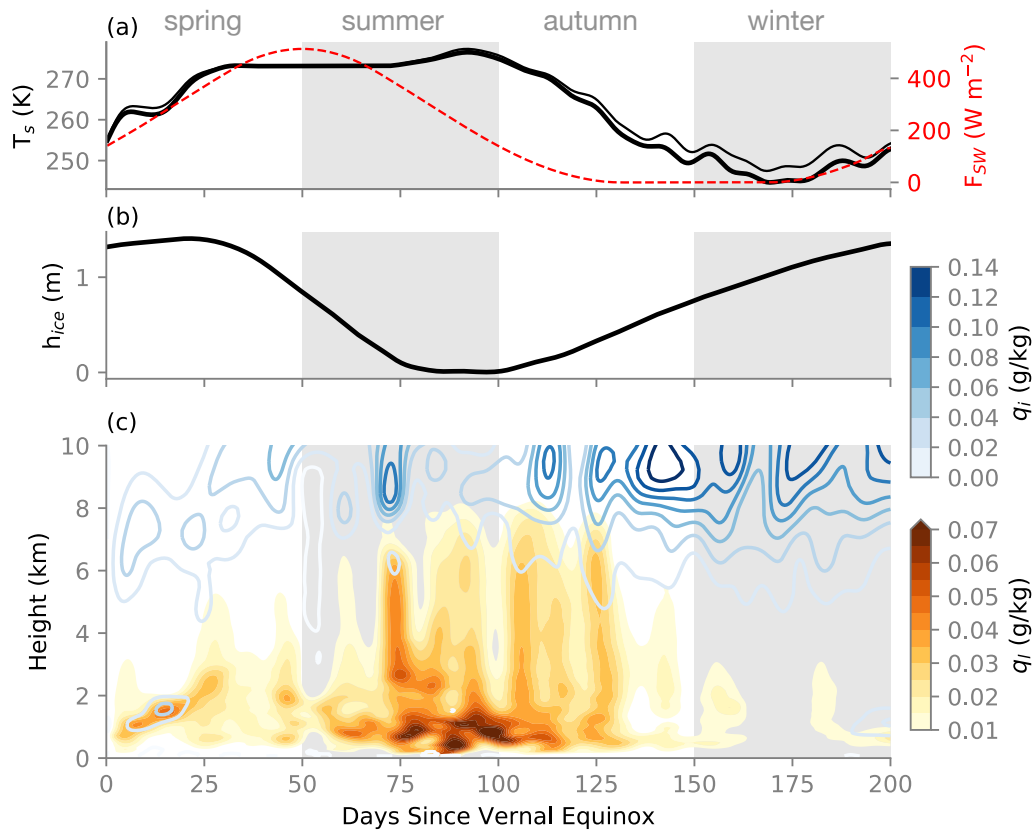
## 269 **2.5 Surface Forcing**

270 To have consistent surface states, we prescribe sea ice thickness in PyCLES from  
 271 GCM output, updated every 6 hours. This ensures consistent bottom boundary condi-  
 272 tions in the GCM and LES, and it indirectly constrains the turbulent heat fluxes. Sur-  
 273 face heat fluxes and temperatures are calculated interactively in the LES, thus slight dif-  
 274 ferences are present between the LES and GCM. We have tested directly prescribing sur-  
 275 face turbulent heat fluxes instead of sea ice thickness, which lead to unreasonable air tem-  
 276 peratures in the LES near the surface. We find that prescribing sea ice thickness is a good  
 277 compromise to obtain comparable surface conditions in the GCM and LES.

## 278 **3 Results**

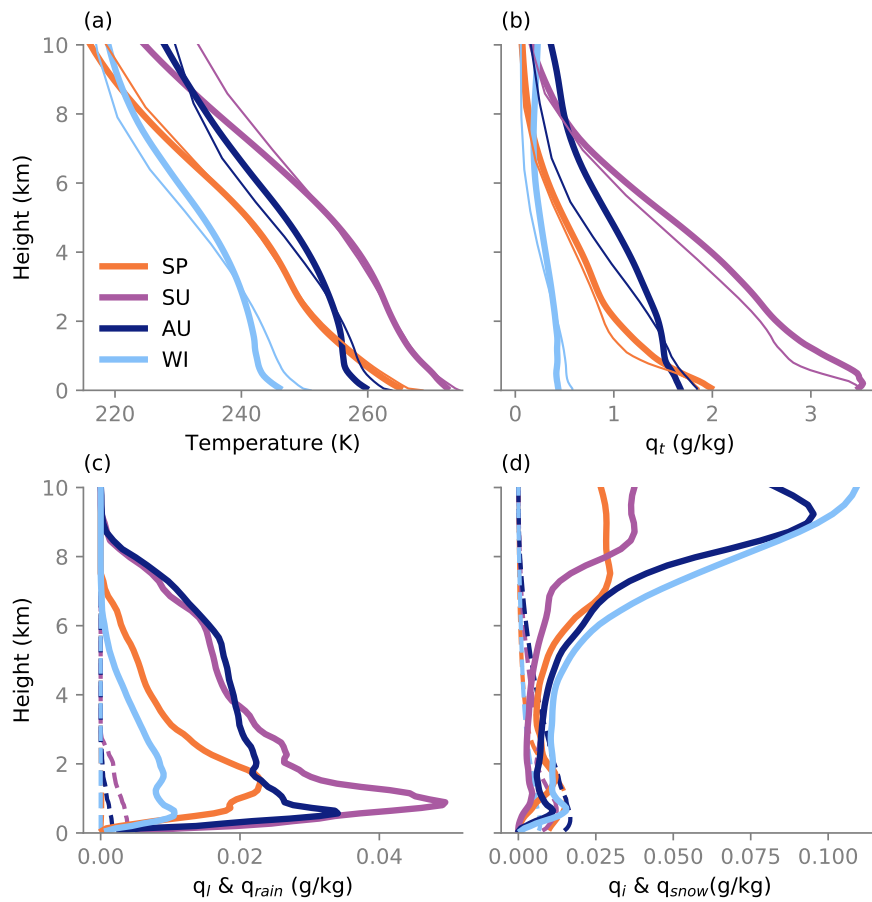
### 279 **3.1 Seasonal Cycle**

280 The high-frequency forcing introduces a large amount of variability in the LES sim-  
 281 ulations. For better visualization, we apply a 10-day lowpass 5th order Butterworth fil-  
 282 ter to smooth the 6-hourly LES output. Figure 3 shows the seasonal cycle of the sur-  
 283 face state and cloud condensates from the GCM-forced LES. Also shown is the insola-  
 284 tion forcing at TOA, which corresponds well with the increase of surface temperature  
 285  $T_s$  from mid winter to mid spring when ice thickness reaches its maximum of 1.4 m (Fig-  
 286 ure 3a and 3b). As  $T_s$  reaches the melting temperature, all shortwave forcing is used to  
 287 melt the sea ice, and the ice thickness declines into summer. Then  $T_s$  increases again above  
 288 the melting temperature, but quickly decreases as insolation declines and sea ice thick-  
 289 ens into winter. Overall, there is a good agreement between LES and GCM  $T_s$ , with the  
 290 largest difference of 5 K in winter. The variation of surface temperature is about 30 K,  
 291 which is within the observed range (26–36 K) of the annual variation of monthly-mean  
 292 near-surface temperatures in the Arctic (Persson, 2002).



**Figure 3.** LES ensemble-mean seasonal cycle of domain-mean (a) surface temperature and TOA shortwave radiative flux, (b) sea ice thickness, and (c) cloud condensate profiles (filled colors for liquid water, contours for ice). GCM surface temperature is shown as the thin black line. Data are smoothed by a 10-day lowpass filter.

293 The maximum cloud liquid is found within the boundary layer during summer and  
 294 and autumn, when the surface temperature  $T_s$  is high and ice thickness  $h_i$  is low (Figure 3c).  
 295 This is also when cloud liquid reaches the highest vertical extent at about 8 km. Cloud  
 296 liquid is present throughout spring, but with lower vertical extent, and it becomes in-  
 297 termittent during winter. Cloud ice, on the other hand, has its maximum in winter in  
 298 the upper troposphere, and it is present throughout the year. The general pattern of the  
 299 seasonal cycle of clouds resembles that of the observed Arctic Ocean cloud fraction (Cesana  
 300 et al., 2012): the maximum liquid cloud fraction is found in summer and early autumn,  
 301 though lower liquid cloud amount persists in winter in the lower troposphere; the high-  
 302 est liquid cloud tops are also found during summer and early autumn, reaching 8 km.



**Figure 4.** LES seasonal (50-day average) domain-mean profiles of (a) temperature, (b) total water specific humidity, (c) liquid water (solid) and rain (dashed) specific humidity, and (d) ice water (solid) and snow (dashed) specific humidity. Thin lines in (a) and (b) show the GCM values for comparison.

303 The polar region experiences large seasonal variations in its thermodynamic pro-  
 304 files, which is simulated by both the idealized GCM and the LES (Figure 4). In addi-  
 305 tion to the large differences in the temperature magnitudes across the seasons, the static  
 306 stability also differs substantially (Figure 4a). Although there is no temperature inver-  
 307 sion in the boundary layer, the lower troposphere is more stable in autumn and winter  
 308 when insolation is weaker, and is more convective in spring and summer when insola-  
 309 tion is stronger. The boundary layer is also moister in summer and spring, although in  
 310 autumn the free troposphere is moister than in spring (Figure 4b).

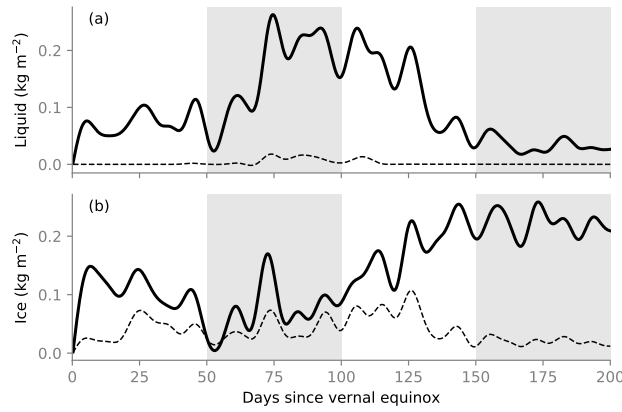
311 Cloud water profiles also display large seasonal variability. Liquid water specific  
 312 humidity  $q_l$  peaks in the lower troposphere below 2 km throughout the year. The  $q_l$  peak  
 313 in summer is five times the  $q_l$  peak in winter (Figure 4c). The  $q_l$  peak below 1 km in sum-  
 314 mer and autumn further indicates the presence of the stratiform layers (also apparent  
 315 in Figure 3c). In contrast, ice water specific humidity  $q_i$  peaks in the upper troposphere,  
 316 and maximizes in winter (Figure 4d). Rain is negligible, but there is a significant amount  
 317 of snow in the lower troposphere, with a magnitude that is comparable to  $q_l$ .

318 Most of the clouds contain ice at higher altitudes, as seen in Figure 3c. Low clouds,  
 319 on the other hand, are dominated by liquid except in winter. Ice clouds are mainly found  
 320 in the upper troposphere above the liquid-containing clouds. The  $q_i$  maximum is in the  
 321 upper troposphere throughout the year, from 7 km in spring to 10 km in winter. Although  
 322 the  $q_i$  maximum is about twice the  $q_l$  maximum, the ice water concentration ( $q_i \rho_{\text{air}}$ ) max-  
 323 imum is much lower than the liquid water concentration. The dominant precipitating  
 324 species in our simulations is snow. Most snow is found in spring and autumn, and the  
 325  $q_{\text{snow}}$  maximum is located at the base of the liquid stratiform layer, below 2 km. This  
 326 is consistent with simulations of Arctic stratocumulus. It suggests that  $q_{\text{snow}}$  forms mostly  
 327 from autoconversion of liquid water in the middle to lower troposphere, instead of from  
 328 ice water in the upper troposphere.

329 The seasonal cycle of condensed water paths integrated over the lower 10 km of the  
 330 LES domain is shown in Figure 5. Cloud liquid water path (LWP) exhibits a seasonal  
 331 cycle with a maximum of  $0.25 \text{ kg m}^{-2}$  in summer and a minimum of  $0.03 \text{ kg m}^{-2}$  in win-  
 332 ter (Figure 5a). Cloud ice water path (IWP) shows a shifted seasonal cycle that peaks  
 333 at  $0.25 \text{ kg m}^{-2}$  in winter (Figure 5b). Intuitively, LWP is the dominant cloud conden-  
 334 sate in summer, while IWP dominates in winter, due to the temperature dependency of  
 335 the liquid fraction shown by equation (1). The snow water path is nonzero throughout  
 336 the year and exceeds the rain water path.

### 337 **3.2 Estimating Cloud Radiative Effects**

338 Although the gray radiation scheme does not allow cloud-radiation interactions in  
 339 either the GCM or the LES, one can use an offline radiative transfer model to estimate  
 340 the radiative effects of the clouds in the LES. To do so, we use the Rapid Radiative Trans-  
 341 form Model for GCMs (RRTMG) (Iacono et al., 2008). Domain-mean profiles of 6-hourly



**Figure 5.** Seasonal cycle of ensemble-mean (a) liquid water path (solid) and rain water path (dashed), and (b) ice water path (solid) and snow water path (dashed). Data are smoothed by a 10-day lowpass filter.

342 temperature, specific humidity, pressure, density, and cloud condensates are used as in-  
 343 put fields for RRTMG. We define the longwave and shortwave cloud radiative effects (CREs)  
 344 as the difference between net all-sky fluxes and clear-sky fluxes, either at TOA or at the  
 345 surface:

346 
$$\text{LWCRE} = (\text{LW}_{\text{all-sky}}^{\downarrow} - \text{LW}_{\text{all-sky}}^{\uparrow}) - (\text{LW}_{\text{clear}}^{\downarrow} - \text{LW}_{\text{clear}}^{\uparrow}), \quad (10)$$

347 
$$\text{SWCRE} = (\text{SW}_{\text{all-sky}}^{\downarrow} - \text{SW}_{\text{all-sky}}^{\uparrow}) - (\text{SW}_{\text{clear}}^{\downarrow} - \text{SW}_{\text{clear}}^{\uparrow}), \quad (11)$$

348 
$$\text{CRE} = \text{LWCRE} + \text{SWCRE}. \quad (12)$$
  
 349

350 The annual-mean CRE at TOA and at the surface are summarized in Table 1, along with  
 351 the observed climatological values from CERES-EBAF averaged over 70–75°N. The ob-  
 352 served net effect of clouds at TOA is to cool the climate, dominated by SWCRE. For the  
 353 LES, when both cloud liquid and ice are included in the radiative transfer calculation,  
 354 the LWCRE term dominates because there is excessive cloud ice in the upper troposphere  
 355 in our simulations. If we only include cloud liquid water in the calculation, the annual-  
 356 mean values based on the LES are much closer to observations. Surface CRE is not as  
 357 sensitive to upper-tropospheric cloud ice, since cloud liquid in the lower troposphere is  
 358 already optically thick. The surface CRE based on our LES closely matches that observed.  
 359 Because of the closer match with observations, we focus on the liquid CRE in our anal-  
 360 ysis here, and defer the discussion on cloud ice bias to Section 4.

**Table 1.** Ensemble-mean annual-mean cloud radiative effect at TOA and surface. For comparison, we show the CERES-EBAF 4.0 climatology averaged from 07/2005 through 06/2015.

CRE ( $\text{W m}^{-2}$ )	TOA			SFC		
	LW	SW	Net	LW	SW	Net
CERES-EBAF	14	-25	-10	41	-27	15
Cloud liquid + ice	41	-33	7.9	45	-27	17
Cloud liquid only	14	-26	-12	39	-22	16

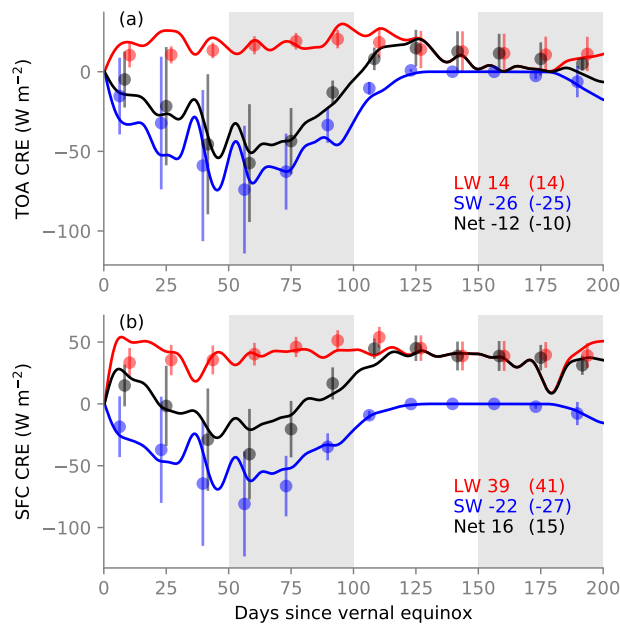
361 Figure 6 shows the seasonal cycle of CRE at TOA and at the surface using cloud  
 362 liquid only in the calculations. The ensemble mean CRE is the average of 4 offline ra-  
 363 diative transfer calculations from each LES simulation (as opposed to the offline calcu-  
 364 lation of the ensemble mean clouds). The seasonal cycle of TOA CRE is dominated by  
 365 the seasonality in SWCRE: Clouds have a strong cooling effect during the sun-lit part  
 366 of the year; during polar night, their longwave warming effect dominates, as expected  
 367 (Figure 6a). The seasonal cycle of LWCRE is much more muted than SWCRE, which  
 368 peaks in late summer at TOA. At the surface, the LWCRE seasonal cycle is damped com-  
 369 pared to TOA; SWCRE variability is weaker at the surface than at TOA, but still peaks  
 370 in late spring (Figure 6b). The net CRE at the surface is much higher than at TOA (16  
 371 versus  $-12 \text{ W m}^{-2}$ ), suggesting that polar clouds warm the surface both in observations  
 372 and in our LES.

## 373 4 Discussion

### 374 4.1 Comparison to Observations

375 An encouraging result of our experiment is the resemblance of the simulated liq-  
 376 uid clouds to observations. Although the model setup here is highly idealized, many pro-  
 377 cesses are absent, and detailed reproduction of the seasonal cycle is not a goal, the sim-  
 378 ulated seasonal cycle of clouds and CRE still resembles that observed. This suggests that  
 379 the minimal building blocks for the seasonal cycle are present in this idealized setup. For  
 380 example, Cesana et al. (2012) produced the seasonal cycle of cloud fraction averaged over  
 381 the Arctic Ocean ( $70\text{--}82^\circ\text{N}$ ) based on a space-borne lidar (CALIPSO-GOCCP). They  
 382 found the maximum frequency of occurrence of liquid clouds near the surface from May  
 383 to September, and the liquid cloud reaches its maximum vertical extent at 7.5 km alti-





**Figure 6.** Ensemble-mean seasonal cycle of CRE due to cloud liquid only at (a) top of atmosphere and (b) surface, estimated off-line with RRTMG and domain-mean profiles. Data are smoothed by a 10-day lowpass filter. Annual mean CRE values are shown in the lower right. Dots show the observed CERES-EBAF CRE monthly climatology averaged over 70–75°N, and error bars show the spatial standard deviation for each month. Annual mean CRE values are shown in the parentheses.

384 tude in July. During winter, the liquid cloud fraction is lower, but liquid clouds still per-  
 385 sist below 2 km. Ice cloud fraction is lower than liquid overall, and is zero below 4 km  
 386 during June to August. The ice cloud maximum occurs at 7 km in winter, while ice cloud  
 387 reaches as high as 11 km. These observations match well with the simulated seasonal cy-  
 388 cle of clouds in our LES (Figure 3c). However, it should be borne in mind that direct  
 389 comparisons between LES and observations are difficult because the spatial scales and  
 390 definitions of cloud fractions are different in LES and in satellite-derived observations  
 391 in Cesana et al. (2012). A more sophisticated comparison should involve satellite sim-  
 392 ulators that convert simulated thermodynamic fields to variables that are directly mea-  
 393 sured by satellites (Chepfer et al., 2008; Kay et al., 2016). Nonetheless, the similarity  
 394 of the LES to observations provides evidence for the physical relevance of our experiments.

395 We can also compare the integrated cloud condensates with satellite observations  
 396 over the Arctic Ocean north of 60°N (Figure 2 in Lenaerts et al. (2017)). The observed

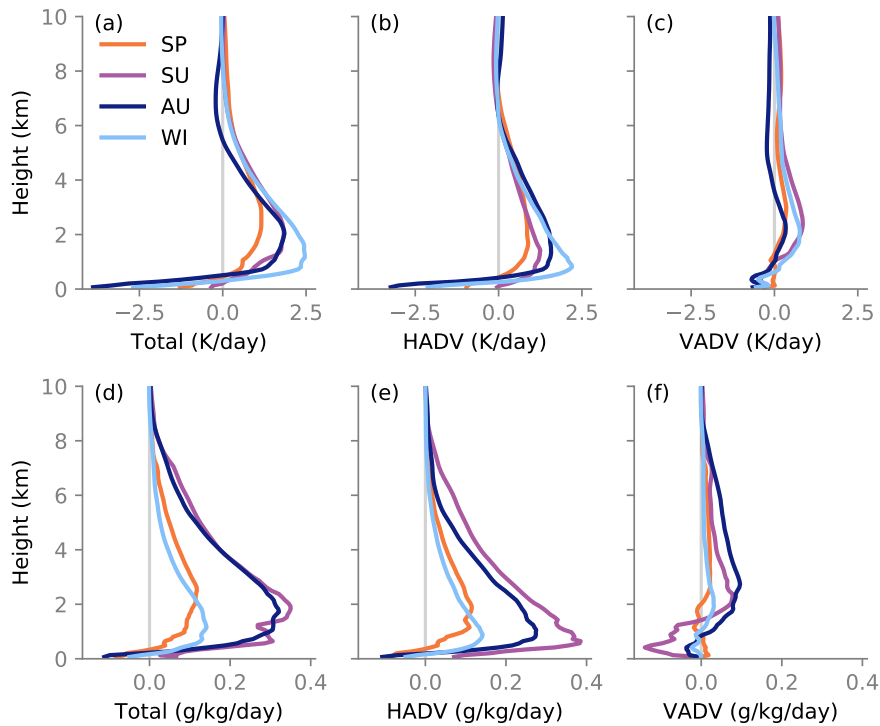
397 LWP ranges from 0.015 to 0.125 kg m<sup>-2</sup>, with the maximum occurring during late sum-  
398 mer and the minimum during winter. Although the maximum ensemble-mean LWP dur-  
399 ing summer in our LES is over-estimated (0.15 kg m<sup>-2</sup>), the timing of the maximum and  
400 minimum is consistent with the observed LWP in polar oceans (Figure 5a). Larger dis-  
401 crepancies are found in IWP. The observed IWP over the Arctic Ocean ranges from 0.01  
402 to 0.11 kg m<sup>-2</sup>. In the LES, the ensemble-mean IWP ranges from 0.07 to 0.4 kg m<sup>-2</sup>  
403 (Figure 5b), much higher than observed. The seasonal cycle of IWP is weak in obser-  
404 vations, and our results show a peak in IWP during winter. The cloud ice excess in the  
405 LES may be related to our simple treatment of ice microphysics and an inefficient re-  
406 moval of ice particles at high altitudes. Interestingly, comprehensive climate models of  
407 the CMIP5 generation tend to underestimate IWP (Lenaerts et al., 2017).

408 Being aware of the biases in our simulated cloud fields, we can compare our esti-  
409 mated liquid CRE to observations from CERES-EBAF (Loeb et al., 2017; Kato et al.,  
410 2018). We choose all longitudes in the latitude band 70–75°N to get average observed  
411 radiative fluxes. The selected domain covers the seasonal sea ice edge, providing the rel-  
412 evant comparison to our idealized experiment. The monthly data from CERES-EBAF  
413 are scaled in time to match the accelerated seasonal cycle of our LES (Figure 6). The  
414 observed SWCRE shows high standard deviations during sunlit months, but the observed  
415 LWCRE shows low standard deviations in warmer months. As a result, our simulated  
416 SWCRE is generally within the observed range during the highly variable spring and early  
417 summer months. Our simulated LWCRE is too strong in spring, and TOA SWCRE is  
418 stronger in late summer/early autumn compared to CERES-EBAF. Nonetheless, our sim-  
419 ulated annual-mean TOA LWCRE and SWCRE based on cloud liquid alone agree well  
420 with observations.

## 421 **4.2 Forcing and Clouds**

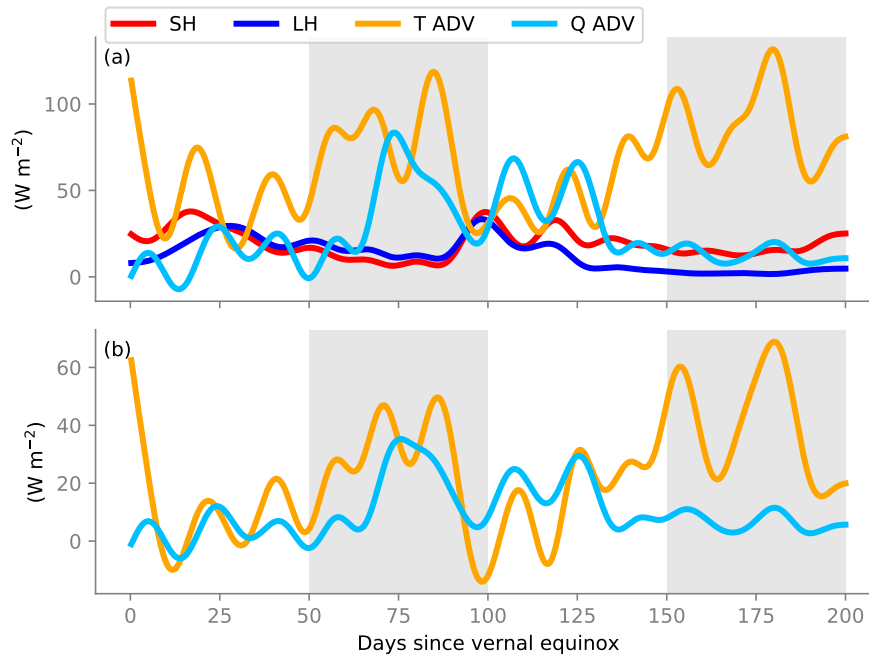
422 What determines the seasonal cycle of Arctic clouds? Radiation is the zeroth-order  
423 driver for any seasonal variability in temperatures in the Arctic. Moisture, on the other  
424 hand, comes from either large-scale advection or surface evaporation. Condensation de-  
425 pends on both temperature and moisture forcing.

426 The external non-radiative forcing for clouds in our LES includes two components:  
427 large-scale advection and surface fluxes. These two are not independent of one another



**Figure 7.** Seasonal average profiles of large-scale forcing of (a) total temperature advection, (b) horizontal temperature advection, (c) vertical temperature advection, (d) total specific humidity advection, (e) horizontal specific humidity advection, and (f) vertical specific humidity advection. Horizontal advection (HADV) is taken directly from the GCM, while vertical advection (VADV) is a hybrid of GCM and LES fields.

428 in the real climate system. Large-scale advection is more important at high latitudes than  
 429 at lower latitudes, because of the large atmospheric heat transport that balances the net  
 430 negative radiative forcing at TOA. Large-scale advection brings heat and moisture into  
 431 the high latitudes year-round (Figure 7a and 7d). For both temperature and specific hu-  
 432 midity advection, the horizontal advection terms dominate (Figure 7b and 7e). Temper-  
 433 ature advection is the strongest in winter, when the pole-to-equator temperature gra-  
 434 dient is the strongest. Summer temperature advection is weak, but it is associated with  
 435 the largest moisture advection. On the other hand, moisture advection is weak in win-  
 436 ter and spring, contributing to a polar atmosphere that is cold and dry. The moisture  
 437 advection seasonal cycle is consistent with the observed horizontal moisture advection  
 438 north of 70°N, but our simulations have peak values in summer that are twice the re-  
 439 analysis values (Serreze et al., 2007; Newman et al., 2012). At the surface, evaporation



**Figure 8.** (a) Seasonal cycle of vertically integrated total temperature and specific humidity advection (converted to dry and latent energy fluxes), as well as sensible and latent heat fluxes at the surface. (b) Seasonal cycle of total temperature and moisture advection integrated over the bottom 2 km. Data are smoothed by a 10-day lowpass filter.

440 is limited in winter but provides a significant source of lower-tropospheric water vapor  
 441 during summer and early autumn.

442 Figure 8a shows the seasonal cycle of the vertically integrated large-scale forcing  
 443 tendencies (with the moisture flux convergence expressed as a latent heat flux conver-  
 444 gence), along with turbulent fluxes at the surface. There is significant synoptic variabil-  
 445 ity in the large-scale advection terms from the GCM; here we focus on the overall bud-  
 446 get and have smoothed all fields with a 10-day lowpass filter.

447 For the entire LES domain and throughout the year, large-scale temperature ad-  
 448 vection is stronger than the surface sensible heat flux. However, if we focus on the low-  
 449 est 2 km (Figure 8b), the surface sensible heat flux is of comparable magnitude to the  
 450 large-scale temperature advection in the boundary layer. Moisture advection is strongest  
 451 in summer and autumn, both in the boundary layer and in the entire troposphere. In  
 452 spring when moisture advection reaches its minimum, surface latent heat flux becomes

453 the dominant moisture source. In winter, large-scale moisture advection contributes more  
454 to the moisture budget than the surface latent heat flux.

455 The concurrence between the moisture advection peak and cloud liquid maximum  
456 (Figure 8a and 5a) points to the dominant role that large-scale moisture advection plays  
457 in governing the seasonal cycle of cloud liquid in the polar region. In summer, air tem-  
458 peratures continue to rise and so does the saturation specific humidity. A moisture source  
459 is needed for condensation to occur during this period, and in our case the source comes  
460 from large-scale advection of water vapor. Air temperatures begin to decrease at the end  
461 of summer, which lowers the saturation specific humidity. Cloud condensates form in au-  
462 tumn due to both cooling and a continued supply of water vapor from large-scale ad-  
463 vection. In winter, the peak in large-scale temperature advection warms the troposphere,  
464 making it harder to form cloud condensates.

465 Beesley and Moritz (1999) tested the sensitivity to large-scale advection of mois-  
466 ture by swapping summer and winter moisture advection in a single-column model. They  
467 found little changes in the simulated cloud fraction. However, both liquid and ice wa-  
468 ter paths were doubled in winter when summer moisture advection is applied (roughly  
469 doubling the winter moisture advection). Their insensitivity of cloud fraction to mois-  
470 ture advection may be due to biases in the mean state, such as the lack of high-frequency  
471 variability in the forcing. In future work, we plan to analyze how large-scale advection  
472 from reanalysis and comprehensive GCMs affects LES cloud cover, to better assess the  
473 influence of forcing magnitude and frequency.

### 474 **4.3 Limitations**

475 Although the idealized GCM has been shown to capture many large-scale features  
476 of the atmospheric circulation, not all aspects are accurately simulated. Known biases  
477 such as jet stream biases and in the storm track response to warming exist (e.g., Tan et  
478 al., 2019). Furthermore, the GCM used in the study has a positive relative humidity  
479 bias in the polar regions. According to reanalysis, the climatological relative humidity  
480 in the free troposphere is between 65% and 70% at 70°N. In the idealized GCM, the rel-  
481 ative humidity is at least 10% higher. This leads to a moist bias in the LES, manifested  
482 in the excessive IWP (Figure 5b) and high ice water specific humidity in the upper tro-  
483 posphere (Figure 3c). The lack of continents may partly explain the over-estimated sum-

484 mer moisture advection into the polar region, as mentioned in section 4.2. We will ad-  
485 dress these issues in future revisions of the experimental design to improve our under-  
486 standing of polar cloud dynamics.

487 Our use of a one-moment bulk microphysics scheme can be limiting in reproduc-  
488 ing the observed cloud seasonal cycle, and especially the ice phase. IWP in our LES is  
489 about 4 times higher than what is seen in observations over the Arctic Ocean (Lenaerts  
490 et al., 2017). We tested the sensitivity of our results to the formulation of liquid frac-  
491 tion (Figure 1) by using the observationally derived formula in Hu et al. (2010), with higher  
492 liquid to ice ratio above 246 K, vice versa below 246 K, and the largest modification in  
493 liquid fractions at temperatures around 240 K (Figure S1). With this modification in  
494 the LES, we found the largest modification in  $q_l$  at temperatures above 240 K because  
495 of the exponential nature of the Clausius-Clapeyron relation. As a result, LWP is higher  
496 in summer to autumn and lower in winter in the simulation with Hu et al. (2010) liq-  
497 uid fraction (Figure S2). Its effect on liquid CRE is strongest in winter, because there  
498 is a cancellation in LW and SW during sunlit seasons. The lowered LWP in winter due  
499 to Hu et al. (2010) liquid fraction leads to a slight reduction of LWCRE, which domi-  
500 nates the net CRE change of  $-2.4 \text{ W m}^2$  in the annual mean (Figure S3).

501 The lack of water vapor and cloud feedbacks in our modeling framework becomes  
502 a major drawback when it comes to representing details of cloud structures and coupling  
503 between radiation and dynamics. For example, cloud-top radiative cooling imposes a dom-  
504 inant forcing to the dynamics of stratocumulus (Bretherton et al., 1999). Without it, the  
505 turbulence in the boundary layer is unlikely to be strong enough to produce a well-mixed  
506 layer and an inversion above the cloud tops. Lack of this radiation-dynamics coupling  
507 explains the structural differences between our simulated clouds and observed Arctic clouds.  
508 However, our GCM-forcing framework provides a clean setup to study the role large-scale  
509 advection plays in controlling the seasonal cycle of cloud liquid. In a follow-up paper,  
510 we will use the same framework to explore the response of polar clouds to climate warm-  
511 ing.

## 512 **5 Conclusions**

513 We adopted an idealized framework in which large eddy simulations are driven by  
514 large-scale forcing from a GCM in a high-latitude setting. Our approach encapsulates

515 components of first-order importance in the polar regions, such as large-scale advection  
 516 of heat and moisture, sea ice, and a simple representation of mixed-phase microphysics.  
 517 Water vapor and cloud feedbacks are not represented in the gray radiative transfer schemes  
 518 in both the GCM and the LES.

519 The seasonal cycle of simulated polar clouds resembles observations qualitatively.  
 520 In particular, maximum cloud liquid is found below 2 km in summer and autumn, and  
 521 it reaches minimum in winter. Cloud ice is found mostly in the upper troposphere. The  
 522 condensed water path is dominated by ice, which is overestimated compared to obser-  
 523 vations. LWP, on the other hand, agrees better with satellite-derived values over the Arc-  
 524 tic Ocean. Offline radiative transfer calculations of liquid cloud radiative effects also show  
 525 encouraging agreement with CERES-EBAF: the net liquid cloud radiative effect is to  
 526 cool the LES domain, but to warm the surface.

527 Analysis of the forcing budget points to the dominant role that large-scale advec-  
 528 tion of moisture plays in controlling the seasonal cycle of cloud liquid. In the boundary  
 529 layer, surface evaporation is of comparable magnitude to large-scale moisture advection.  
 530 The peak of large-scale temperature advection occurs in winter, when the pole-to-equator  
 531 temperature gradient is greatest. This warms the troposphere and reduces cloud con-  
 532 densates.

533 Our idealized framework provides an opportunity to study mechanisms of cloud-  
 534 climate feedbacks in the complicated polar climate system. In a follow-on paper, we will  
 535 look at the polar cloud response to climate warming caused by increased longwave op-  
 536 tical thickness of the atmosphere. We will also analyze how changes in large-scale ad-  
 537 vection with warming affect the simulated cloud amount, to pave the road for future stud-  
 538 ies with more realistic large-scale forcing.

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552 available online at <https://data.caltech.edu/records/1429>.

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