

Understanding the Extratropical Liquid Water Path Feedback in Mixed-Phase Clouds with an Idealized Global Climate Model

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Abstract

A negative shortwave cloud feedback associated with higher extratropical liquid water content in mixed-phase clouds is a common feature of global warming simulations, and multiple mechanisms have been hypothesized. A set of process-level experiments performed with an idealized global climate model show that the common picture of the liquid water path (LWP) feedback in mixed-phase clouds being controlled by the amount of ice susceptible to phase change is not robust. Dynamic condensate processes—rather than static phase partitioning—directly change with warming, with varied impacts on liquid and ice amounts. Here, three principal mechanisms are responsible for the LWP response, namely higher adiabatic cloud water content, weaker liquid-to-ice conversion through the Bergeron-Findeisen process, and faster melting of ice and snow to rain. Only melting is accompanied by a substantial loss of ice, while the adiabatic cloud water content increase gives rise to a net increase in ice water path (IWP) such that total cloud water also increases without an accompanying decrease in precipitation efficiency. Perturbed parameter experiments with a wide range of climatological LWP and IWP demonstrate a strong dependence of the LWP feedback on the climatological LWP and independence from the climatological IWP and supercooled liquid fraction. This idealized setup allows for a clean isolation of mechanisms and paints a more nuanced picture of the extratropical mixed-phase cloud water feedback than simple phase change.

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24 **1. Introduction**

25 With atmospheric warming from greenhouse gases, cloud properties would vary in manifold
26 ways, resulting in further changes in radiative fluxes and climate. Despite the recent advances
27 in mechanistic understanding, the so-called cloud feedback is widely considered to be the largest
28 contributor to the uncertainties in climate sensitivity and model projection of future warming
29 (Sherwood et al. 2020). Ceppi et al. (2017) identifies three robust components of cloud feedback
30 in comprehensive global climate models (GCMs): a positive longwave feedback from rising free
31 tropospheric clouds, a positive shortwave (SW) feedback from decreasing subtropical low cloud
32 fraction, and a negative SW feedback from increasing extratropical cloud optical depth.

33 Uncertainty associated with cloud feedback is dominated by the SW components (Soden and
34 Vecchi 2011; Vial et al. 2013). Among these, this study focuses on the component that affects
35 radiation through altering cloud optical depth or brightness (as opposed to cloud fraction). This
36 cloud optical depth feedback is robustly negative in the Coupled Model Intercomparison Project
37 Phase 5 (CMIP5) GCMs (Zelinka et al. 2016), though it may be artificially tuned to a small range
38 (McCoy et al. 2016), and mechanistic uncertainty still abounds (Gettelman and Sherwood 2016;
39 Ceppi et al. 2017; Korolev et al. 2017). Observations have shown that in pure liquid and mixed-
40 phase (liquid and ice co-existing) clouds, cloud optical depth is primarily controlled by liquid
41 water path (LWP), which is the vertically integrated cloud liquid (Stephens 1978). Ice affects cloud
42 optical depth to a lesser extent owing to larger sizes of ice particles and ice water path (IWP) being
43 generally smaller than LWP (Pruppacher and Klett 2010; McCoy et al. 2014; Cesana and Storelmo
44 2017). GCMs predict a robust extratropical LWP increase in response to global warming, which
45 is thought to be the main driver of the negative SW cloud feedback (e.g. Ceppi et al. 2016).

46 Recent modeling studies have highlighted the need to improve GCM representation of the
47 extratropical cloud feedback. Zelinka et al. (2020) showed that the increased climate sensitivity
48 in CMIP6 models relative to CMIP5 is largely due to changes in this feedback. The multi-model
49 ensemble mean changes from negative in CMIP5 to slightly positive in CMIP6 presumably due to
50 model physics differences. Therefore, it is critical to delineate the underlying mechanisms of the
51 extratropical cloud feedback and its various components.

52 Multiple pathways have been proposed to explain the extratropical increase (Ceppi et al. 2017) in
53 liquid cloud condensate. The first is an increase in the adiabatic cloud water content. With warming,
54 the amount of water condensed in saturated updrafts increases (Tselioudis et al. 1992; Gordon and
55 Klein 2014); the fractional change is greater at colder temperatures (Betts and Harshvardhan 1987;
56 Somerville and Remer 1984). The second mechanism involves phase change in mixed-phase clouds
57 (e.g., Mitchell et al. 1989; Senior and Mitchell 1993; McCoy et al. 2015; Storelvmo et al. 2015; Tan
58 et al. 2018), which occurs only at temperatures below freezing. As isotherms shift upward with
59 warming, the liquid-to-ice ratio at a given mixed-phase cloud location is likely to increase (Tan et al.
60 2016), thereby increasing cloud optical depth. An implication of this phase change mechanism is
61 that since liquid precipitates less efficiently than ice, total cloud water content may increase (Klein
62 et al. 2009; McCoy et al. 2015; Ceppi et al. 2016; McCoy et al. 2018). This work will address
63 both mechanisms and their impacts on LWP and IWP. A third potential mechanism frequently
64 mentioned in the literature is poleward jet shifts. As this effect is highly model dependent and
65 unlikely to be dominant (Kay et al. 2014; Ceppi and Hartmann 2015; Wall and Hartmann 2015;
66 Ceppi et al. 2016), it is not explored here.

67 The relative importance of the proposed mechanisms is still unclear. LWP itself is robustly linked
68 to temperature in both models (Ceppi et al. 2016) and observations (Terai et al. 2019), hinting at the
69 potential for emergent constraints on the negative SW cloud feedback. McCoy et al. (2016) noted

70 that among CMIP5 GCMs, T5050, the diagnosed temperature at which liquid and ice exists in
71 equal amounts globally, is strongly anti-correlated with LWP, but positively correlated with cloud
72 fraction despite the lack of a physical explanation. At the same time, the range of T5050 (as well as
73 a similarly-defined 90% glaciated temperature) estimated from space-borne observations is much
74 lower than that diagnosed from CMIP5 models, suggesting that the models tend to freeze liquid
75 at temperatures that are too high (Cesana et al. 2015; McCoy et al. 2016). Multiple GCM studies
76 (McCoy et al. 2014; Tan et al. 2016; Frey and Kay 2018) have shown that increasing the ratio of
77 supercooled liquid to total water (the so-called supercooled liquid fraction or SLF) in mixed-phase
78 clouds decreases the SW negative feedback, and thus increases climate sensitivity. These results
79 have been attributed to models with higher T5050 having more susceptible ice (McCoy et al. 2018),
80 which is hypothesized to control the feedback strength (as in Tan et al. 2018). Improvements in
81 understanding the governing mechanisms are especially important as some modeling studies with
82 observationally-based constraints have suggested that the negative SW cloud optical depth feedback
83 is too strong or even of the wrong sign in GCMs, implying that the actual climate sensitivity may
84 be underestimated (e.g. Tan et al. 2016; Terai et al. 2016).

85 This work utilizes an idealized model to probe the physical mechanisms underlying the extratrop-
86 ical cloud water feedback. Idealized models complement comprehensive GCMs (Held 2005, 2014)
87 since their workings are relatively easy to understand (Pierrehumbert et al. 2007). This is particu-
88 larly true as previous studies of mixed-phase clouds are hindered by the non-linear complexity of
89 cloud microphysics and the potential for unrealistic interactions between different parameterized
90 processes (Ceppi et al. 2017). We seek to test the plausibility of the leading hypotheses in the
91 mixed-phase cloud feedback literature including the simple conceptual picture of liquid increasing
92 at the expense of ice with warming, which has fueled the notion of the extratropical LWP feedback
93 being controlled by the amount of susceptible ice. As mentioned above, more ice in the control cli-

94 mate is thought to cause a greater increase in liquid with warming. The main supporting evidence
95 is the positive correlation between the LWP feedback and climatological SLF or T5050 (McCoy
96 et al. 2018; Tan et al. 2018). With a set of targeted, process-level experiments, we seek to explore
97 the complexity of the mixed-phase cloud feedback. We also use a perturbed parameter ensemble of
98 experiments with varied cloud physics settings to investigate the feasibility of predicting the LWP
99 feedback from the control climate.

100 This paper is arranged as follows. Section 2 outlines the methodology. Section 3 presents the
101 results from process-level and perturbed parameter experiments. Section 4 compares with previous
102 studies with the goal of examining the plausibility of the phase change mechanism and other related
103 arguments. Section 5 concludes as to rethinking the physical picture of the extratropical mixed-
104 phase cloud feedback and suggests a path for future research.

105 **2. Methodology**

106 The idealized GCM used here combines a simple dry GCM with passive water and clouds as
107 described in detail in Ming and Held (2018). The core is Held-Suarez dry dynamics (Held and
108 Suarez 1994) at a T42 horizontal resolution (about 2.8° spacing) with 20 equally spaced vertical
109 layers. Passive water vapor and cloud tracers (specific humidity, cloud liquid mixing ratio, cloud
110 ice mixing ratio, and cloud fraction) are included, but are not allowed to feedback on the dynamics
111 (i.e., no latent heating or cloud radiative effects). The cloud tracers evolve following a prognostic
112 large-scale cloud scheme with bulk single-moment microphysics. The sub-grid-scale total-water-
113 based relative humidity (RH) is assumed to follow a beta distribution, which is a function of the
114 grid-mean RH. The beta distribution is designed such that a grid box with a mean total-water-based
115 RH value above a certain threshold value (RH_c , 83.3% at the default half-width of 0.2) would
116 have sub-grid-scale RH over 100%, thus producing clouds. The role of surface evaporation to

117 create the water vapor tracer is mimicked by nudging air parcels below 850 hPa toward saturation
118 as in Galewsky et al. (2005). As clouds are completely decoupled from dynamics, this model
119 a unique tool for isolating individual mechanisms in a clean fashion without circular feedbacks.
120 With no convective parameterization, the application of the cloud scheme is limited to stratiform
121 clouds (and not any mixed-phase clouds formed in shallow convection). Yet, as noted in Ming
122 and Held (2018), while idealized, this model provides strong representation of cloud distribution
123 in the extratropical free troposphere. The control simulation (Ctrl) is the model's default climate.
124 For Ctrl and all perturbation experiments, the atmospheric state (e.g., temperature and winds) is
125 identical at every time step. All model simulations include a 300-day spin-up, and the next 1000
126 days are averaged for analysis.

127 The bulk microphysics scheme has separate but interconnected treatments of liquid and ice based
128 on Rotstajn (1997) and Rotstajn et al. (2000). The same scheme is also used in the GFDL AM2.1
129 model, one of the two models compared in Ceppi et al. (2016). As shown in Fig. 1, water vapor forms
130 cloud liquid and ice through condensation and deposition, respectively. The initial partitioning of
131 cloud liquid and ice is based entirely on temperature. All condensate at temperatures greater than
132 -40°C is formed as liquid based on the consideration that ice nuclei are generally limited in the
133 atmosphere (Rotstajn et al. 2000). Supercooled liquid (existing between 0° and -40°C) can then be
134 converted to ice principally through the Bergeron-Findeisen (BF) process (and without an explicit
135 treatment of heterogeneous ice nucleation). In the control climate, the primary sink of water vapor
136 (98.8% globally) is conversion to cloud liquid. Microphysical sources of water vapor come from
137 cloud liquid (evaporation), cloud ice (ice sublimation), rain (rain evaporation), and snow (snow
138 sublimation). Together, rain evaporation and snow sublimation, the most significant microphysical
139 sources, comprise 22.3% of all water vapor sources. Surface evaporation (a non-microphysical
140 source) constitutes the main supplier of water vapor (76.4%).

141 Cloud liquid forms rain through autoconversion and accretion. To facilitate conversion of cloud
142 liquid to ice through the BF process, a minimum amount of ice crystal mass (10^{-12} kg) on which
143 deposition can occur is assumed to be always present. (Note that the BF process is not formulated
144 to be explicitly linked to aerosols.) Cloud liquid is also converted to cloud ice through riming
145 (accretion of cloud liquid by ice) and homogeneous freezing (colder than -40°C). Overall, 68.2% of
146 cloud liquid sinks are to rain and 30.9% to cloud ice. Cloud ice is lost almost completely (98.3%)
147 to snow through ice settling. In the microphysics scheme, cloud ice and snow are treated effectively
148 as one species, experiencing the same fall rate, and are only distinguished by their location in or
149 outside of a cloud. Ice and snow can melt into rain: if this takes place in a cloud, it is considered
150 melting of ice; if it takes places outside of a cloud, it is considered melting of snow. Cloud ice is
151 also lost to water vapor through sublimation.

152 The *process-level experiments* involve increasing the temperature field fed to certain parts of
153 the microphysics scheme or the formulation of surface evaporation by 2 K (summarized in Table
154 1). These isolated warming experiments are designed after Ceppi et al. (2016). Here, in the
155 microphysics scheme (same as that used in the AM2.1 aquaplanet in Ceppi et al. (2016)), there
156 are at least four explicitly temperature-dependent processes: partitioning of newly formed cloud
157 condensate, the BF process, homogeneous freezing, and melting of ice and snow. When water
158 vapor experiences condensation/sublimation at the beginning of the microphysics scheme, it is
159 initially partitioned into cloud liquid and ice based solely on temperature. Only liquid is created at
160 temperatures warmer than -40°C , and only ice otherwise. Supercooled liquid can be converted to
161 ice through the BF process, homogeneous freezing, and riming. For the BF process, temperature
162 affects whether or not the process occurs (below 0°C) as well as the rate of cloud liquid being
163 converted to cloud ice, which is greater at lower temperatures (see Eqn. A8). These two effects are
164 tested in combination (BF2K, subjecting the *BF* process to a 2-K warming). (By contrast, riming

165 is not directly dependent on temperature; see Eqn. A10.) Homogeneous freezing of cloud liquid to
166 ice occurs only when the temperature is less than -40°C and converts all cloud liquid to ice. Ice and
167 snow melt into rain only when the temperature is higher than 0°C , with the melting being limited
168 to the amount that would restore the grid-box temperature to 0°C . Melting of ice and snow are
169 tested in combination (ME2K, subjecting *melting* to a 2-K warming). All of these microphysical
170 processes—initial partitioning, the BF process, homogeneous freezing, and melting—are also
171 perturbed in tandem in MI2K (2-K warming of *microphysics*).

172 A significant influence of temperature in the cloud scheme is in the calculation of the saturation
173 specific humidity (q_s) and related variables (the T derivative of q_s , the psychrometric constant,
174 and the sum of the vapor diffusion and thermal conductivity factors) that are used in many parts
175 of the scheme. Since surface evaporation is also formulated in parallel based on q_s , q_s for
176 microphysics and surface evaporation are perturbed simultaneously in Qse2K (2-K warming of q_s
177 for the stratiform cloud scheme and *evaporation*). This experiment enables us to study the effect
178 of the adiabatic cloud water content increase. Finally, to cover all the aforementioned effects of
179 temperature as well as any other effects (such as the influence of temperature on air density), a 2-K
180 temperature increase is fed to the cloud scheme and surface evaporation to create the Tse2K (full
181 warming) experiment.

182 To develop a predictive theory of the extratropical mixed-phase cloud feedback that is applicable
183 to a wide range of control states, a set of *perturbed parameter experiments* (also summarized in
184 Table 1) are created by systematically modifying three key parameters of the cloud scheme. The
185 first two have been suggested as significant for the mixed-phase cloud feedback: the strength of
186 the BF process may be too efficient (Tan et al. 2016) and RH_c too high (McCoy et al. 2016). To
187 vary the strength of the BF process, the formula for the conversion rate is altered arbitrarily by
188 multiplying with a constant (0.25, 0.5, 2 or 4). The corresponding experiments are labeled as

189 quarBF, halvBF, doubBF and quadBF. Note that these adjustments do not result in actual changes
 190 in the BF rate as large as those imposed. The effective RH_c (83.3% in Ctrl) is varied from 76.7% to
 191 90.0% at increments of $\sim 3.3\%$ (rh767, rh800, rh867, and rh900) by altering the half-width of the
 192 sub-grid-scale RH beta distribution. Finally, a third parameter is chosen to cleanly affect the mean-
 193 state amount of cloud ice: the fall speed of cloud ice (relative to the large-scale vertical motion) is
 194 perturbed by multiplying with a constant (0.5, 0.75, 1.25 or 1.5). The corresponding experiments
 195 are v050, v075, v125 and v150. For each of these states, a Tse2K simulation (increasing the
 196 temperature field fed to the cloud scheme and surface evaporation by 2 K) is created, and the
 197 response (for example, rh767_Tse2K minus rh767) analyzed.

198 The key to understanding the steady-state mixing ratios of cloud liquid and ice (q_l and q_i ,
 199 respectively) and their responses to the warming is how they are related to the time tendencies of
 200 the aforementioned microphysical processes. To illustrate the point, let us write the time derivative
 201 of a variable q (q_l or q_i) as:

$$\frac{dq}{dt} = s - aq^b, \quad (1)$$

202 where s is the source term, and the sink term is parameterized as a power-law function of q with a
 203 and b as constants. It follows that the fractional change of q can be related to the fractional change
 204 of s by:

$$\frac{\delta q}{q} = \frac{1}{b} \frac{\delta s}{s}. \quad (2)$$

205 The formulation and behavior of the autoconversion parameterization (Eqn. A1) are discussed
 206 in Golaz et al. (2011) (see their Equations 12-14). Although the rate is nominally proportional
 207 to $q_l^{7/3}$, it is effectively controlled by a numerical limiter (Eqn. A3), which tends to set q_l at a
 208 critical value (q_{crit}) determined by a tunable threshold droplet radius (r_{thresh}) and droplet number
 209 concentrations (N). Since neither r_{thresh} nor N changes in this study, q_l should be close to q_{crit}

210 when autoconversion is the dominant process. By contrast, accretion is proportional to q_l and the
211 flux of rain (Eqn. A4). The BF rate (Eqn. A8) is effectively independent of q_l , but conditionally
212 proportional to $q_i^{1/3}$. Riming (Eqn. A10) is proportional to q_l and the flux of settling ice, which is
213 related to the fall speed and q_i . Similarly, ice settling (Eqn. A6) at a specific level is determined
214 by the fall speed and vertical gradient of q_i ($\partial q_i / \partial p$, where p denotes pressure). If q_i is altered by
215 the same ratio throughout the column, an assumption that holds approximately for the simulations
216 examined here, the fractional change in the ice settling rate would be the same as that in q_i . The
217 microphysical tendency equations are listed in the Appendix for reference. Condensation and
218 deposition, the main sources of cloud liquid and ice, are not directly related to q_l or q_i .

219 The analysis focuses on two variables: LWP and IWP, which are, respectively, vertically inte-
220 grated cloud liquid and cloud ice in units of g m^{-2} . Absolute and fractional changes in LWP and
221 IWP are normalized by warming and thus given in units of $\text{g m}^{-2} \text{K}^{-1}$ and $\%$ K^{-1} , respectively.
222 Due to the highly simplified nature of the boundary layer in this model (i.e., surface evaporation
223 saturating the air below 850 hPa), for the purposes of this analysis the vertical integral has a lower
224 boundary of 850 hPa such that LWP and IWP only represent the cloud condensate above 850 hPa.
225 Similarly, specific humidity and cloud condensate tendency terms, when column-integrated, only
226 represent values above 850 hPa. 30° to 60° and 60° to 90° are considered the mid-latitudes and
227 high-latitudes, respectively, and together they are considered the extratropics. Data is averaged
228 between the two hemispheres because of the hemispheric symmetry of the simulated climate. The
229 supercooled liquid fraction (SLF) is calculated as the ratio of cloud liquid to total cloud water
230 (liquid and ice). The daily SLF is binned as a function of temperature at an interval of 0.1 K
231 for each grid box in the extratropical region above 850 hPa with the temperature at which SLF is
232 closest to 50% considered to be T5050 (liquid and ice partitioned equally).

233 3. Results

234 a. Process-level Experiments

235 Fig. 2 shows the zonal-mean LWP and IWP (averaged between hemispheres) in the control
236 case (Ctrl), yielding a picture of the model's default climate [see Ming and Held (2018) for other
237 related variables including RH and CF]. Here, LWP dominates IWP equatorward of the storm
238 tracks (at around 45°); note that this LWP/IWP ratio is not directly comparable with full GCMs
239 as here the boundary layer is excluded in the calculation of LWP and IWP. In the total warming
240 experiment (Tse2K), the general features, including the location of the storm tracks, remain the
241 same. Both LWP and IWP are higher at all latitudes in the warmer climate. The increase in LWP
242 is more pronounced than that in IWP in the mid-latitudes, while they are more comparable in the
243 high-latitudes.

244 Table 2 and Fig. 3 break down the LWP and IWP feedbacks seen in Tse2K. The increase in
245 LWP (Fig. 3a) in the extratropics is dominated by the microphysical component (MI2K) with a
246 much smaller (slightly less than 20%) contribution from the increased q_s (Qse2K). MI2K and
247 Qse2K combine nearly linearly to produce the full Tse2K increase in LWP suggesting that Tse2K
248 does not add any significant temperature-affected processes beyond those perturbed in MI2K and
249 Qse2K. The LWP feedback from the adiabatic water content increase is stronger in the high-
250 latitudes ($5.2\% \text{ K}^{-1}$) than in the mid-latitudes ($1.6\% \text{ K}^{-1}$), as one would expect from the nonlinear
251 temperature-dependence of the Clausius-Clapeyron relation.

252 Within the combined microphysical component, the BF process (BF2K) is responsible for most
253 of the LWP increase, with a smaller contribution from melting (ME2K) present only in the mid-
254 latitudes (Fig. 3b), and homogeneous freezing and initial phase partitioning producing negligible
255 results (presumably because of the small amount of cloud condensate present near -40°C). The

256 BF effect is realized through the temperature-dependence of the conversion rate, as opposed to the
257 temperature threshold at which the BF process takes control. LWP increases as the BF process
258 slows down, converting less liquid to ice. The melting of snow to rain dominates the melting of ice
259 to rain in terms of their effects in enhancing LWP. As discussed later, this can be conceptualized as
260 a consequence of weaker riming since there is less snow (falling ice) to collect cloud liquid. Thus,
261 we conclude that the increase in LWP with warming results primarily from a significant weakening
262 of the BF process.

263 The IWP feedback is more nuanced. As shown in Fig. 3c, Qse2K and MI2K produce opposite
264 effects: IWP increases at all latitudes in the former, while it decreases in the mid-latitudes with no
265 significant change in the high-latitudes in the latter. In Qse2K, the normalized fractional increase
266 in the high-latitude IWP ($7.9\% \text{ K}^{-1}$) is greater than the mid-latitude counterpart ($6.7\% \text{ K}^{-1}$),
267 consistent with the adiabatic water content increasing with temperature at a faster rate at colder
268 temperatures. The net result in Tse2K, to which Qse2K and MI2K add effectively linearly, is an
269 increase in IWP, principally poleward of 45° . The relative importance of the BF process versus
270 melting is reverse to the LWP feedback. The microphysical effect is dominated by ME2K (Fig. 3d);
271 the enhanced melting of snow contributes to the lowering of IWP more than that of cloud ice. By
272 contrast, BF2K gives rise to very little change in IWP. The fact that a weakening of the BF process
273 causes a large increase in LWP, but no concurrent decrease in IWP is somewhat counter-intuitive,
274 a point to which we will return later in this section when discussing the BF2K results in detail. (As
275 with LWP, perturbing homogeneous freezing or initial phase partitioning produces no significant
276 change in IWP.)

277 Fig. 4 shows the vertical structures of the changes in the mixing ratios of cloud liquid and ice.
278 To better understand the underlying physical mechanisms, the main tendency terms driving the
279 steady-state cloud liquid and ice are plotted in Figs. 5 and 6, respectively. No appreciable change

280 in q_l is present below the freezing line in any experiment (Fig. 4) even when there are large local
281 changes in cloud liquid tendencies, as is the case for condensation in Qse2K (Fig. 5a). It is also
282 clear from Fig. 5 that autoconversion and accretion are the principal sinks of q_l above 0°C in
283 Ctrl, with autoconversion slightly stronger. As explained in Section 2, q_{crit} exerts a strong control
284 over q_l when autoconversion dominates. By contrast, the BF process and riming take over in the
285 mixed-phase cloud temperature range (between 0° and -40°C). Both the BF process and riming
286 increase with the enhanced condensation in Qse2K (Fig. 5m and q). While the BF process is
287 independent of q_l , since riming is proportional to q_l steady-state q_l increases (Fig. 4a). On the ice
288 side, faster riming acts to increase q_i (Fig. 6e). Moreover, the increased condensation leads directly
289 to higher q_i through the BF process (Fig. 6a), which is conditionally proportional to $q_i^{1/3}$. The
290 resulting higher flux of settling ice, which is formulated to be approximately proportional to q_i ,
291 tends to further accelerate riming, but lower q_l . This cancels out much of the increase in q_l caused
292 by the increased condensation (Fig. 4a). The end result is that the normalized fractional increase
293 in the extratropical IWP ($6.8\% \text{ K}^{-1}$) is much greater than the LWP counterpart ($1.7\% \text{ K}^{-1}$).

294 The imposed warming to the BF process (BF2K) slows down the BF conversion from liquid
295 to ice (Fig. 5n). Since autoconversion and accretion play limited roles in the mixed-phase cloud
296 regime, an acceleration of riming (Fig. 5r) is the only way to re-establish the q_l tendency balance,
297 causing a significant increase in q_l (Fig. 4b). This re-balancing can be conceptualized as a weaker
298 BF process producing more cloud liquid to be scavenged by falling ice through riming. Since
299 the q_l and q_i tendencies (and their changes) are of the same magnitude but opposite signs for the
300 BF process and riming, the effect of the two processes on q_i is dictated by the balance of their
301 q_l counterparts (Fig. 6b and f). Because the effects of q_i are of opposing sign, there is near-zero
302 net change in cloud ice (Fig. 4f). This somewhat counterintuitive result emphasizes the need to
303 evaluate changes in q_l and q_i based on process changes and a dynamic re-balancing of sources

304 and sinks. For example, when weakened BF process (as through warming) experiments were run
305 with the riming process entirely removed from the microphysics scheme, instead of BF process q_i
306 tendency change being balanced principally by enhanced riming with little change in ice settling
307 (as shown in Fig. 6f and j), without riming, the tendency change was principally balanced by
308 significantly weakened ice settling.

309 The melting perturbation (ME2K) is unique in the sense that the resulting changes in cloud liquid
310 and ice are of mirror image in terms of spatial structure (Fig. 4c and g). The main reason is that the
311 melting perturbation effects are relatively confined to a narrow domain of a few degrees above the
312 time-averaged freezing line. The warming-induced additional melting acts to increase the flux of
313 rain and decrease the flux of settling ice simultaneously. Both factors have implications for q_l . The
314 former tends to accelerate accretion with an effect of decreasing the q_l tendency, while the latter
315 acts to slow down riming which increases the q_l tendency. The simulation shows a net increase of
316 q_l , suggesting that the latter factor prevails over the former. The signs of the simulated rate changes
317 are consistent with the expectations, and they largely balance out each other (Fig. 5k and s), with
318 a weaker contribution from autoconversion (Fig. 5g). On the ice side, the reduced supply of ice
319 from riming is balanced entirely by lowering q_i and thus settling (Fig. 6g and k). The role of the
320 BF process here is negligible as it is relatively ineffective at temperatures within a few degrees of
321 0 °C.

322 This process-level analysis illustrates why the principal components of the full warming (Tse2K)
323 simulation, namely Qse2K, BF2K, and ME2K, increase q_l and hence LWP, as summarized schemat-
324 ically in Fig. 7. Although they all point in the same direction, the microphysical warming com-
325 ponents (BF2K and ME2K) are a stronger contribution to the LWP feedback than the macro-
326 physical/thermodynamic component (Qse2K). The extratropical IWP feedback stems from a broad
327 increase in q_i from Qse2K being offset partially by a decrease near the freezing line from ME2K.

328 The results underscore that multiple processes with distinct characteristics are influential in shaping
329 the LWP and IWP responses, and contradict the common picture suggested in mixed-phase cloud
330 feedback literature of an effective trade-off between ice and liquid. Here, the dominant processes
331 which increase LWP with warming in mixed-phase clouds are not doing so at the expense of ice, so
332 the actual picture is more complicated than a (direct or indirect) conversion from ice to liquid with
333 warming. Liquid and ice in mixed-phase clouds are not in a static equilibrium; rather, they exist in
334 a dynamic balance of sources and sinks. These source and sink processes are directly changed by
335 warming as opposed to a simple temperature-dependent phase partitioning.

336 *b. Perturbed Parameter Experiments*

337 To further explore the sensitivity of the LWP and IWP feedbacks, a set of alternative control
338 states was created by altering three key aspects of the cloud scheme, namely the value of RH_c ,
339 the strength of the BF process and the fall speed of ice (v_{fall} , Eqn. A7), summarized in Table
340 1. As shown in Fig. 8, the first two changes produce a wide range of the climatological LWP
341 (approximately a factor of 2), but little variation in IWP. Lower RH_c or weaker BF process leads
342 to higher LWP. While these experiments are not designed to fully explain the insensitivity of IWP
343 to RH_c or the BF process in more detail than the previous section, the broad principle is that
344 steady-state values are determined by a dynamic balance of continuing phase conversion, not a
345 static equilibrium. And, ice changes are harder to manufacture using local processes (like the BF
346 process) when ice is so strongly controlled by gravitational settling. In the v_{fall} perturbations,
347 IWP varies widely (a factor of more than 3) with higher fall speed giving rise to lower IWP but
348 with little spread in the climatological LWP.

349 All of these perturbed parameter experiments are subjected to 2-K warming in a way analogous
350 to Tse2K. The resulting normalized LWP and IWP changes (δLWP and δIWP , respectively) are

351 plotted against their climatological counterparts in Fig. 9. Ranging from 2.6 to 3.4 g m⁻² K⁻¹,
352 relative to 3.0 g m⁻² K⁻¹ in Tse2K (Table 2), the LWP feedback is positively correlated with the
353 climatological LWP (Fig. 9a). The best linear fit yields that $\delta\text{LWP} = 0.045 \text{ LWP} + 1.60$, with an R^2
354 of 0.98. Thus, the fractional change can be written as $\delta\text{LWP}/\text{LWP} = 0.045 + 1.60/\text{LWP}$, suggesting
355 that the marginal gain decreases with increasing LWP. Since the four experiments targeting the BF
356 process, namely {quar, halv, doub, quad}BF, effectively demonstrate the basic behavior of the LWP
357 feedback, we start by focusing on them in the effort to explain the latter. As shown above, the main
358 sink terms for cloud liquid in the mixed-phase regime are the BF process and riming. As the BF
359 process becomes stronger from quarBF to quadBF, riming has to weaken if the total sink is constant,
360 giving rise to lower climatological LWP, in line with the model simulations. Recall that the riming
361 rate is proportional to cloud liquid. The process-level experiments suggest that the warming effect
362 is realized mostly through the BF process. In these experiments, the warming-induced perturbation
363 to the BF process is roughly proportional to its baseline rate (not shown). Therefore, the lower
364 the climatological LWP is, the stronger the baseline BF rate and associated perturbation are. The
365 combination translates into higher fractional change in LWP with lower climatological LWP (from
366 a stronger BF process).

367 Lowering RH_c tends to increase LWP by enhancing condensation in a way similar to Qse2K.
368 They differ in that the former causes a large increase in autoconversion, but without any substantial
369 change in accretion or riming, while all three processes increase in the latter. As explained
370 before, autoconversion can adjust to forced changes such as those resulting from warming without
371 perturbing cloud liquid. As a result, a control state with enhanced autoconversion should be less
372 sensitive to warming (in this limited context; other feedbacks in complex models such as that
373 noted in Mülmenstädt et al. (2021) may complicate this picture). This explains why lowering RH_c
374 gives rise to larger LWP, but smaller fractional increases in response to warming. Of interest is

375 the minimal effect on the extratropical climatological LWP and δ LWP from drastically changing
376 the climatological IWP (or susceptible ice) in the ice fall speed experiments. Clearly, the LWP
377 feedback is correlated with the climatological LWP, but not the climatological IWP. The preceding
378 analysis also holds when the LWP feedback is further divided into the mid- and high-latitude
379 components (not shown).

380 The IWP feedback is correlated strongly with the climatological IWP (Fig. 9b). Note that the
381 variation in the IWP feedback is almost exclusively from the ice fall speed experiments (ranging
382 from 0.57 to 1.70 g m⁻² K⁻¹). An inspection of the best linear fit result (δ IWP = 0.023·IWP
383 + 0.031, with an R^2 of 1.00) indicates that the intercept is so small that the warming-induced
384 change in IWP is effectively proportional to the climatological IWP. In other words, the normalized
385 fractional change is constant at 2.3% K⁻¹. This relatively simple relation reflects the fact that
386 gravitational settling is the main process through which cloud ice can be adjusted to re-establish
387 the mass balance. As seen both from the process-level experiments and the BF-series parameter
388 perturbation experiments, the amount of cloud ice is not sensitive to the BF process. In the
389 meantime, riming is under the strong control of the cloud liquid balance. This leaves gravitational
390 settling as the only way to alter cloud ice without affecting other processes substantially. Note
391 that similar linear relationships hold if the climatological LWP and IWP are computed only for the
392 mixed-phase temperature range (between 0 and -40°C), confirming the independence of the LWP
393 feedback from the climatological IWP (or susceptible ice).

394 **4. Discussion**

395 As noted in the introduction, much of the existing literature on the extratropical mixed-phase cloud
396 feedback centers on the correlation between the climatological SLF/T5050 and LWP feedback.
397 Specifically, the lower SLF is or the higher T5050 is, the stronger the LWP feedback is (Tan et al.

398 2016; Frey and Kay 2018; McCoy et al. 2018). The presumption is that the phase change mechanism
399 plays a crucial role, meaning that ice would be statistically replaced by liquid as isotherms shift
400 with warming. Thus, the climatological susceptible ice or IWP is thought to be predictive of the
401 feedback strength, forming the basis of potential emergent constraints (Tan et al. 2016). A related
402 argument is that the phase change would give rise to a decrease in precipitation efficiency and a
403 net increase in total water path (TWP, the sum of LWP and IWP) as liquid is less efficient than
404 ice in forming precipitation (McCoy et al. 2018). While it is clear from the previous section that
405 the mixed-phase cloud feedback is much more complicated than simple phase change, we further
406 test the validity of both claims—SLF/T5050 as a predictor and decreased precipitation efficiency
407 increasing TWP—against our results.

408 The climatological T5050 in the perturbed parameter experiments spans a wide range (~ 15 K)
409 (Fig. 10). Stronger BF process and higher RH_c favor lower LWP (or SLF) and higher T5050. The
410 normalized δLWP , however, is strongly anti-correlated with T5050 ($R^2 = 0.92$, Fig. 10) as it is
411 positively correlated with the climatological LWP (Fig. 9a). The T5050/ δLWP anti-correlation is
412 opposite to that expected if susceptible ice drove the LWP feedback and is contrary to the findings
413 of Tan et al. (2016) and Frey and Kay (2018) based on the CAM5 model and of McCoy et al.
414 (2018) based on CMIP5 models. Furthermore, as shown in Fig. 8, the climatological IWP is
415 effectively constant for these experiments. This calls into question the hypothesis that susceptible
416 ice controls the strength of the LWP feedback. As another evidence against the hypothesis, if the
417 v_{fall} perturbations are included, the predictive power of T5050 is significantly diminished ($R^2 =$
418 0.76 , Fig. 10). The large variations in the climatological IWP, which drive the spread in T5050
419 in the v_{fall} perturbations, do not affect δLWP significantly. Thus, any connection here between
420 T5050 and the LWP feedback is not derived from the climatological ice but rather the climatological
421 liquid. This finding suggests that it is important to, when showing correlation between changes in

422 T5050 (or SLF) and LWP feedback or climate sensitivity, also consider the independent roles of
423 changes in climatological liquid or ice as potentially meaningful in addition to their ratio.

424 To understand why a T5050/LWP feedback connection might be present in some models but
425 not others, we consider the dissection of mechanisms for LWP increase in aquaplanet versions
426 of CAM5 and AM2.1 in Ceppi et al. (2016). AM2.1 uses virtually the same large-scale cloud
427 parameterizations as our idealized model, and the AM2.1 results documented in Ceppi et al.
428 (2016) are in excellent agreement with ours despite numerous differences in model setup and
429 experimental design, a testament to the central role of cloud parameterizations in determining the
430 feedback. Whereas both CAM5 and AM2.1 yield higher LWP in response to warming, their IWP
431 changes differ in sign (see their Figure 2). IWP decreases in CAM5, but increases in AM2.1.
432 Moreover, microphysical processes, especially the BF process, are responsible for the majority
433 of the LWP increases, but cannot even account for the signs of the combined extratropical IWP
434 changes (their Figure 7): the microphysically-induced IWP change is an increase in CAM5 and a
435 decrease in AM2.1. Note that CAM5 implements the Morrison-Gottelman microphysics scheme
436 (Morrison and Gottelman 2008), which differs significantly from the Rotstayn-Klein microphysics
437 scheme (Rotstayn 1997) used in AM2.1 and our model, particularly in the treatment of ice and
438 snow. As noted previously, the Rotstayn-Klein scheme treats cloud ice and snow indistinguishably
439 and therefore lacks direct representation of cloud ice autoconversion and accretion by snow (though
440 tuning of the ice fall speed can indirectly account for these sinks of cloud ice). Additionally, the
441 Morrison-Gottelman scheme includes a representation of ice nucleation, which was found in Tan
442 and Storelvmo (2016) to have an even stronger effect than ice fall speed on LWP and IWP. In
443 this sense, it is not inconceivable to see microphysically-induced IWP changes being qualitatively
444 different between the two models. Clearly, the large discrepancy in IWP response to warming merits

445 further analysis and evaluation of both microphysics schemes, especially given the important role
446 of ice cloud microphysics for Arctic cloud feedback (Tan and Storelvmo 2019).

447 Beyond the microphysical feedback, in the Ceppi et al. (2016) study, if one assumes linear
448 additivity (which appears to hold) the non-microphysical component of the IWP change would be
449 a net loss in CAM5 and a net gain in AM2.1. Our results demonstrate that the non-microphysical
450 enhancement of IWP in AM2.1 is attributable to the adiabatic cloud water content increase, a
451 possibility noted in Ceppi et al. (2016). Thus, attempting to reconcile this work with others raises
452 the intriguing question of what factors could outweigh the adiabatic cloud water content effect
453 (however strong it is) and cause the net loss seen in CAM5. These factors (perhaps related to
454 convective ice) should be further explored in complex GCMs and the adiabatic ice effect evaluated
455 for robustness. From the process dissection in Ceppi et al. (2016), it appears that the considerable
456 loss of cloud ice in the warming experiments conducted with CAM5 in Tan et al. (2016) and Frey
457 and Kay (2018) is not microphysical (stratiform) in origin, and thus should not be interpreted
458 as being related to the concurrent increase of cloud liquid, which roots in microphysics. This
459 mechanistic understanding casts further doubt on the susceptible ice hypothesis and other related
460 arguments. From a broader perspective, Ceppi et al. (2016) also noted a robust extratropical
461 LWP increase with warming in the CMIP5 model ensemble mean, without a compensating large
462 decrease in IWP. This is consistent with other studies showing diverse extratropical LWP and IWP
463 feedbacks in models beyond the two highlighted by Ceppi et al. (2016). For example, Lohmann
464 and Neubauer (2018), using ECHAM6-HAM2 with microphysics after Lohmann and Roeckner
465 (1996), found no increase in ECS with increased SLF (unlike the relation found in Tan et al. 2016).
466 McCoy et al. (2021) showed that among CMIP5 and CMIP6 GCMs, most show an increase in
467 liquid along with a slight reduction in ice.

468 Having seen no evidence of the utility of SLF/T5050 as a predictor here for LWP feedback, we
469 now consider whether decreased precipitation efficiency contributes here to the increase in TWP.
470 We calculate the large-scale precipitation efficiency as defined in Zhao (2014), which is the ratio
471 of the total cloud condensation rate (the sum of condensation and deposition fluxes) to surface
472 precipitation and represents the fraction of the condensate that subsequently rains out. There is a
473 slight increase in precipitation efficiency with warming (80.5% in Ctrl versus 81.1% in Tse2K).
474 This results from microphysical increases (80.7% in BF2K and 80.8% in ME2K) being offset by a
475 macrophysical decrease (80.0% in Qse2K). All changes are on the order of 1% or less. Critically,
476 no evidence of an increase in cloud lifetime is present, with precipitation efficiency increasing
477 rather than decreasing. Another measure of a precipitation efficiency effect is surface precipitation
478 normalized by TWP (P/TWP) as in McCoy et al. (2015), which can be thought of as the inverse
479 of the cloud water residence time. Following the Clausius-Clapeyron relation, the extratropical
480 surface precipitation increases by 6.9% K^{-1} in Tse2K and Qse2K, but remains essentially constant
481 in the microphysical experiments. P/TWP increases by 1.9% from 1.03 hr^{-1} in Ctrl to 1.05 hr^{-1}
482 in Tse2K. Again, the net result is a slight decrease in the cloud water residence time or a slight
483 increase in precipitation efficiency. These results do not support a precipitation efficiency effect
484 with warming here as widely claimed (e.g., at the heart of the argument of Bjordal et al. 2020).

485 This finding does not mean a precipitation efficiency feedback is not present in reality, but it
486 may not be present in models as assumed. Mülmenstädt et al. (2021) showed that when warm rain
487 parameterizations are adjusted to better simulate reality in a complex GCM (ECHAM-HAMMOZ),
488 a large negative cloud lifetime effect becomes present. Here we show that other mechanisms can
489 explain a significant increase in LWP and TWP, emphasizing the need to carefully diagnose
490 mechanisms to explain model results which may not contain a significant precipitation efficiency
491 feedback without a warm rain efficiency adjustment. In our model, the weakening of the BF process

492 (BF2K) increases TWP while keeping precipitation nearly constant, suggesting that the BF process
493 alone could affect precipitation efficiency, and thus should be the focus of research to improve
494 its representation in models in addition to the need for improvement in warm rain efficiency as
495 highlighted by Mülmenstädt et al. (2021)

496 Here, in the absence of a precipitation efficiency-mediated strong phase change effect, the adia-
497 batic cloud water content effect is shown to be responsible for increasing TWP by enhancing both
498 liquid and ice. McCoy et al. (2015) observed that increasing TWP was a significant contribution to
499 increased extratropical LWP in CMIP5 models, with only 20–80% of the LWP increase being due
500 to phase re-partitioning. Using observations and modeling, McCoy et al. (2019) highlighted the
501 primacy of the adiabatic cloud water content effect in explaining the increase in LWP with warm-
502 ing in extratropical cyclones. It was found that more than 80% of the enhanced Southern Ocean
503 extratropical cyclone LWP in GCMs from warming can be predicted based on the relationship
504 between the climatological warm conveyor belt moisture flux and cyclone LWP and the change in
505 moisture flux with warming (see also McCoy et al. 2020). While phase change may play a role in
506 the remaining unexplained LWP increases, especially in the poleward half of cyclones, it is clearly
507 a secondary mechanism. A ground-based observational study (Terai et al. 2019) found that both the
508 moist adiabatic scaling and phase partitioning mechanisms are equally important for explaining the
509 increase in LWP with warming at cold temperatures. A complementary space-based observational
510 study (Tan et al. 2019), however, suggests phase change is more important than the adiabatic cloud
511 water content increase in explaining the increase in cloud optical depth with cloud top temperature.
512 Between these observational studies, the GCM studies referenced in this Discussion section, and
513 the idealized modeling results presented herein, it is clear that more research is clearly needed
514 for elucidating the relative importance of the two mechanisms. These mechanisms, as well as a

515 potential precipitation efficiency mediated effect, should be carefully diagnosed in future GCM
516 research as an important step in constraining the mixed-phase cloud feedback.

517 **5. Conclusions**

518 This study used an idealized GCM to perform a set of process-level experiments which delin-
519 eated three key mechanisms of the extratropical LWP feedback involving mixed-phase clouds:
520 higher adiabatic cloud water content, weaker liquid-to-ice conversion through the BF process, and
521 strengthened melting of ice and snow to rain with associated impacts on riming. Over half of
522 the extratropical LWP increase can be attributed to the weakening of the BF process, without a
523 corresponding decrease in IWP. The extratropical IWP in fact increases with warming due to the
524 adiabatic cloud water effect, with a small offset caused by stronger melting. Warming experiments
525 in a perturbed parameter ensemble demonstrate a strong dependence of the LWP feedback on the
526 climatological LWP and independence from the climatological IWP. T5050 is anti-correlated with
527 δ LWP and is therefore only useful as a predictor insofar as it represents the climatological LWP as
528 opposed to the climatological IWP. No associated decrease in precipitation efficiency is found in
529 this modelling setup.

530 The overarching goal of this study is to improve mechanistic understanding of the extratropical
531 mixed-phase cloud feedback. Our results help refine the current physical conceptualization of the
532 LWP feedback as more nuanced than simple phase change, involving impacts of higher adiabatic
533 cloud water content, weaker cloud liquid sinks such as the BF process, and indirect phase changes
534 moderated by precipitation processes (especially riming). Liquid and ice in mixed-phase clouds
535 are in a dynamic equilibrium with microphysical process efficiencies defining time-averaged phase
536 partitioning and its change with warming. These results are helpful for guiding efforts to constrain
537 mixed-phase parameterizations in GCMs through process-oriented diagnostics. In particular, the

538 effect of warming on the BF process, which is at the heart of mixed-phase cloud microphysics,
539 should be better understood and represented in GCMs (see Tan and Storelvmo 2016). In addition
540 to the BF process, the climatological LWP needs to be better constrained. Not only is it shown here
541 to be predictive of the LWP feedback, but also the radiative impact of increases in LWP is highly
542 dependent on the control state (Bodas-Salcedo et al. 2016, 2019). Finally, similar process-based
543 studies, especially among varying microphysics schemes, are vital, as cloud water source and sink
544 efficiencies define the mixed-phase cloud phase partitioning (Ceppi et al. 2016). Mixed-phase cloud
545 studies should show results at the process level to better conclude as to the driving mechanisms and
546 implications for climate sensitivity. Because of complex interactions in full GCMs when mixed-
547 phase physics are perturbed (as in Tan et al. 2016; Frey and Kay 2018), idealized setups such as
548 that utilized here present a clean, complementary approach for elucidating causal relationships.

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553 *Data availability statement.* The output from the simulations described in this manuscript is
554 archived at the Geophysical Fluid Dynamics Laboratory and is available upon request.

555 APPENDIX

556 **Microphysical Transformation Equations**

557 The following equations are those parameterized in the microphysical scheme used herein (after
558 Rotstayn 1997; Rotstayn et al. 2000).

559 *a. Precipitation Formation Processes*

560 *Autoconversion*: the time rate change of grid mean liquid from autoconversion is parameterized

561 as:

$$\frac{\partial q_l}{\partial t} \Big|_{au} = -q_a \times \left(\frac{0.104 g \rho^{4/3} E_{c,au}}{\mu (N \rho_l)^{1/3}} \right) \times (q_l/q_a)^{7/3} \times H(r_d - r_d^{au}) \quad (A1)$$

562 where μ is the dynamic viscosity of air, $E_{c,au}$ is the mean collection efficiency of the autoconversion

563 process, ρ_l is the density of pure liquid, and N is the number of cloud droplets per unit volume. In

564 the Heaviside function, H , r_d^{au} is a critical drop radius that the mean volume radius of cloud drops,

565 r_d , must exceed for autoconversion to occur, where:

$$\rho q_l / q_a = 4\pi N \rho_l r_d^3 / 3 \quad (A2)$$

566 Autoconversion is limited to that which would decrease q_l to the threshold:

$$MAX \left(- \frac{\partial q_l}{\partial t} \Big|_{au} \right) = \ln \left(\frac{\rho q_l / q_a}{4\pi N \rho_l (r_d^{au})^3 / 3} \right) \times \frac{q_l}{\Delta t_{cld}} \quad (A3)$$

567 *Accretion*: the time rate change of grid mean liquid from accretion is parameterized as:

$$\frac{\partial q_l}{\partial t} \Big|_{acc} = -a_{rain}^{cld} \times 65.8 E_{c,acc} (R_{rain}^{cld} / \rho_l a_{rain}^{cld})^{7/9} \times (q_l / q_a) \quad (A4)$$

568 where R_{rain}^{cld} is the grid mean flux of rain entering the rid box from above that enters saturated air,

569 a_{rain}^{cld} is the portion of the grid box that this occurs in, and $E_{c,acc}$ is the collection efficiency between

570 rain drops and cloud droplets which is parameterized as:

$$E_{c,acc} = r_d^2 / (r_d^2 + 20.5 \mu^2) \quad (A5)$$

571 *Gravitational Settling*: the sink of cloud ice due to gravitation settling is:

$$\frac{\partial q_i}{\partial t} \Big|_{gr} = - \frac{\partial}{\partial p} \{ q_a \times \rho g V_f \times (q_i / q_a) \} \quad (A6)$$

572 where V_f is the fall speed the cloud ice fall as relative to the large-scale vertical motion and is
 573 parameterized as:

$$V_f = 3.29(\rho q_i/q_a)^{0.16} \quad (\text{A7})$$

574 *b. Conversions between Liquid and Ice*

575 *BF Process*: the time rate change of the Bergeron-Findeisen process (growth of an ice crystal
 576 from preferential condensation) is parameterized as:

$$\left. \frac{\partial q_l}{\partial t} \right|_{\text{berg}} = - \frac{q_a \times (N_i/\rho)^{2/3} \times 7.8 \times (\text{MAX}(q_i/q_a, M_{i0}N_i/\rho))^{1/3}}{(\rho_i)^{2/3} \times (A + B)} \quad (\text{A8})$$

577 where N_i is the number of ice nuclei per unit volume, M_{i0} is the mass (10^{-12}) of an initial
 578 crystal assumed to always be present, ρ_i is the mass density of pristine ice crystals. Additionally,
 579 $A = (L_v/K_a T) \cdot ((L_v/R_v T) - 1)$ and $B = R_v T / \chi e_s$, where K_a is the thermal conductivity of air, χ
 580 is the diffusivity of water vapor in air, and R_v is the gas constant for water vapor. The ice nuclei
 581 density, N_i , is parameterized assuming the air is a liquid water saturation:

$$N_i = 1000 \exp \left[12.96 \frac{(e_{sl} - e_{si})}{e_{si}} - 0.639 \right] \quad (\text{A9})$$

582 where e_{sl} and e_{si} are the saturation vapor pressures over liquid and ice, respectively.

583 *Riming*: the time rate change of riming (falling ice colliding and coalescing with cloud droplets)
 584 is parameterized as:

$$\left. \frac{\partial q_l}{\partial t} \right|_{\text{rim}} = -a_{\text{snow}}^{\text{cld}} \times \lambda_f E_{c,\text{rim}} (R_{\text{snow}}^{\text{cld}} / 2\rho_i a_{\text{snow}}^{\text{cld}}) \times (q_l/q_a) \quad (\text{A10})$$

585 where ρ_i is the assumed density of falling ice crystals, $R_{\text{snow}}^{\text{cld}}$ is the grid mean flux of settling ice
 586 entering the rid box from above that enters saturated air, $a_{\text{snow}}^{\text{cld}}$ is the portion of the grid box that this
 587 occurs in, $E_{c,\text{rim}}$ is the collection efficiency for the riming process (fixed), and λ_f is parameterized
 588 as a function of temperature:

$$\lambda_f = 1.6 \times 10^3 \cdot 10^{0.023(276.16K - T)} \quad (\text{A11})$$

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TABLE 1. Description of the experiments.

Name(s)	Perturbation(s)
Ctrl	the control with $RH_c = 83.3\%$
Tse2K	2-K warming applied to the temperature seen by the (stratiform) cloud scheme and surface evaporation
Process-level Experiments (Section 3a)	
Qse2K	2-K warming applied to calculation of q_s for the cloud scheme and surface evaporation
MI2K	2-K warming applied to <i>microphysical</i> processes: BF process, melting, homogeneous freezing, and initial phase partitioning
BF2K	2-K warming applied to the <i>BF</i> process
ME2K	2-K warming applied to <i>melting</i>
Perturbed Parameter Experiments (Section 3b)	
{quar, halv, doub, quad}BF	the BF conversion rate multiplied by {0.25, 0.5, 2, 4}
rh{767, 800, 867, 900}	$RH_c = \{76.7\%, 80\%, 86.7\%, 90\%\}$
v{050, 075, 125, 150}	the ice fall speed multiplied by {0.5, 0.75, 1.25, 1.5}
{name}_Tse2K	the corresponding Tse2K experiment for {name} (e.g., quarBF_Tse2K)

740 TABLE 2. Normalized changes in LWP and IWP ($\text{g m}^{-2} \text{K}^{-1}$) in the process-level experiments. The normalized
 741 fractional changes ($\% \text{K}^{-1}$) are in parentheses. The climatological values (g m^{-2}) in Ctrl are also given.

	Extratropics		Mid-Latitudes		High-Latitudes	
	LWP	IWP	LWP	IWP	LWP	IWP
Ctrl	29.9	35.6	38.3	42.7	4.6	14.1
Tse2K	3.0 (9.9)	0.9 (2.4)	3.6 (9.3)	0.8 (1.9)	1.1 (24.2)	1.0 (7.2)
Qse2K	0.5 (1.7)	2.4 (6.8)	0.6 (1.6)	2.8 (6.7)	0.2 (5.2)	1.1 (7.9)
MI2K	2.2 (7.4)	-1.4 (-4.0)	2.6 (6.9)	-1.9 (-4.4)	0.9 (19.4)	0.0 (-0.3)
BF2K	1.7 (5.5)	-0.1 (-0.2)	1.9 (5.0)	-0.1 (-0.2)	0.9 (18.7)	0.0 (0.2)
ME2K	0.6 (2.1)	-1.4 (-3.9)	0.8 (2.1)	-1.8 (-4.2)	0.0 (0.8)	-0.1 (-0.5)

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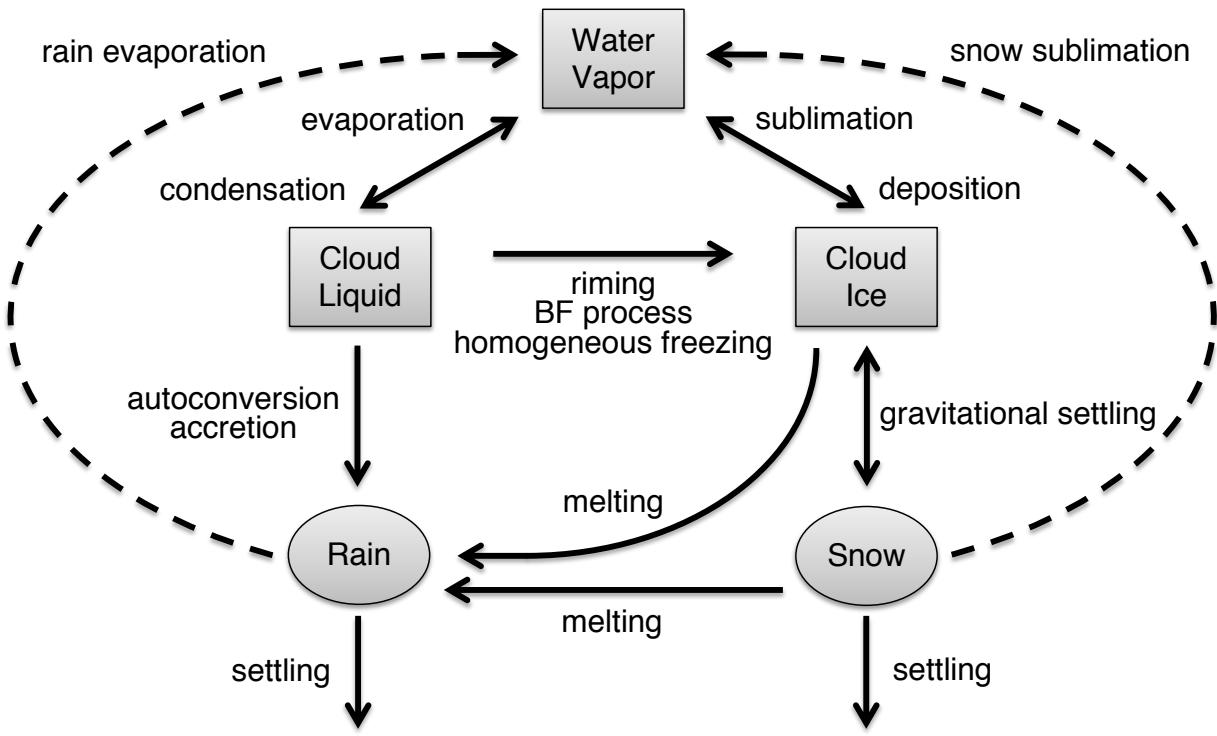
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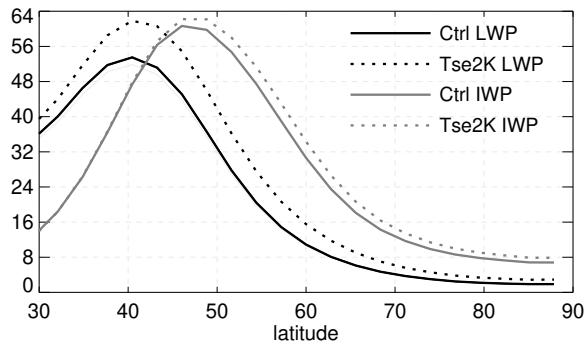
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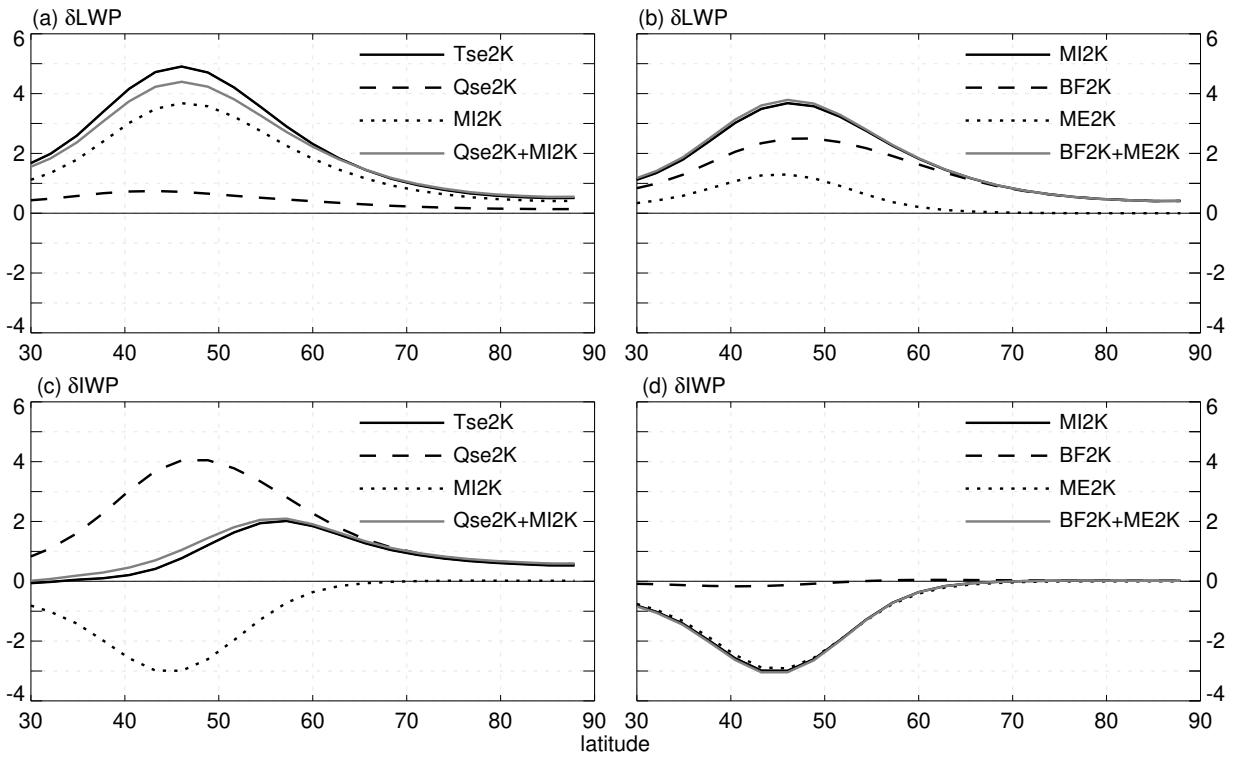
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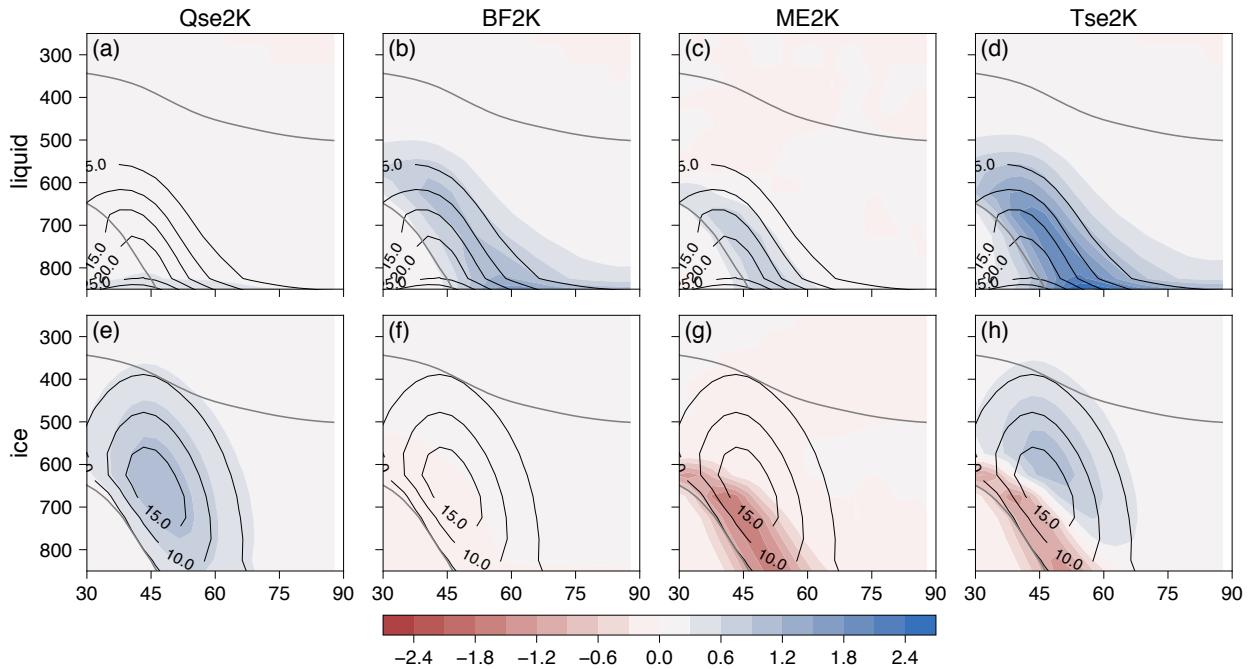
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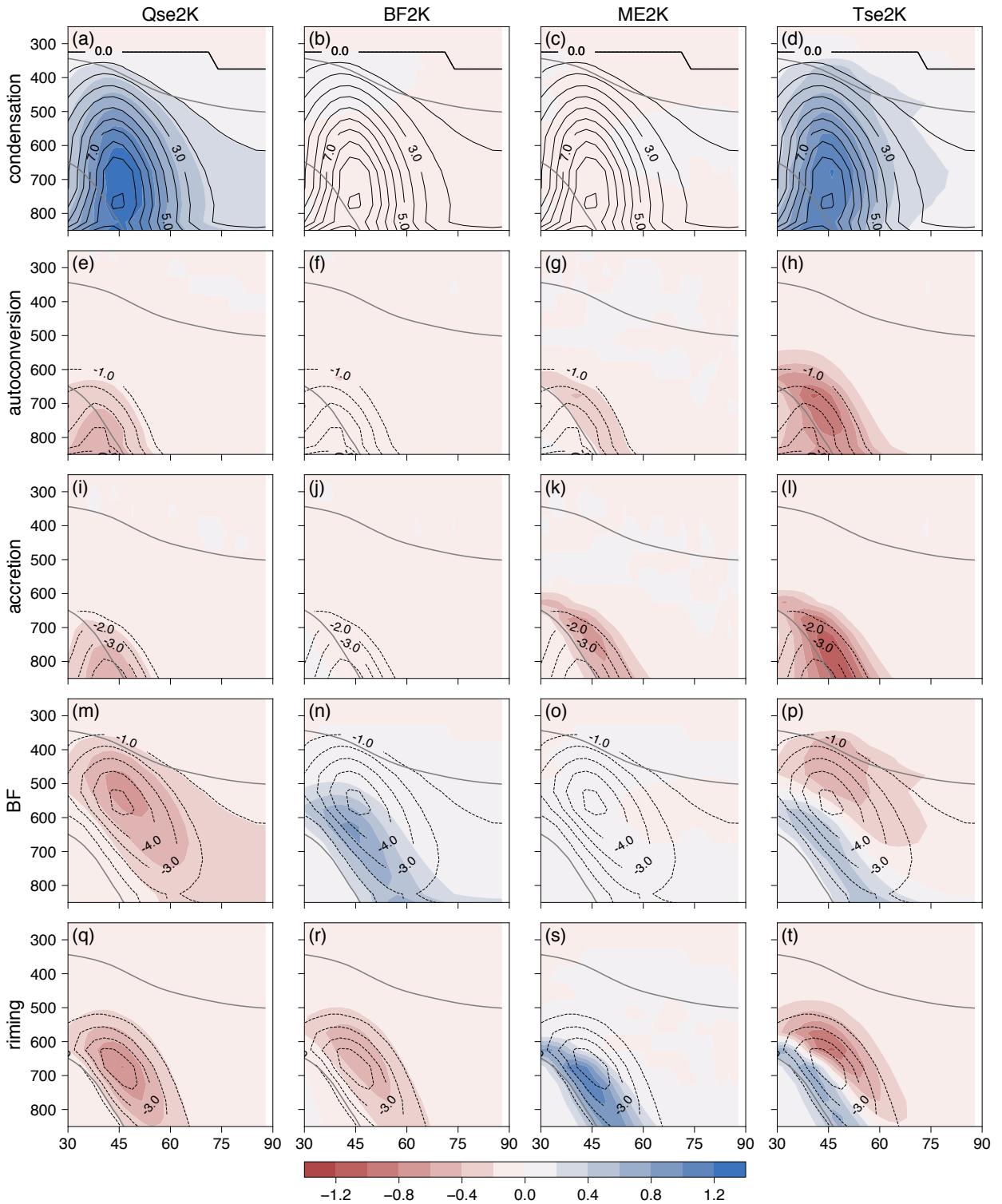
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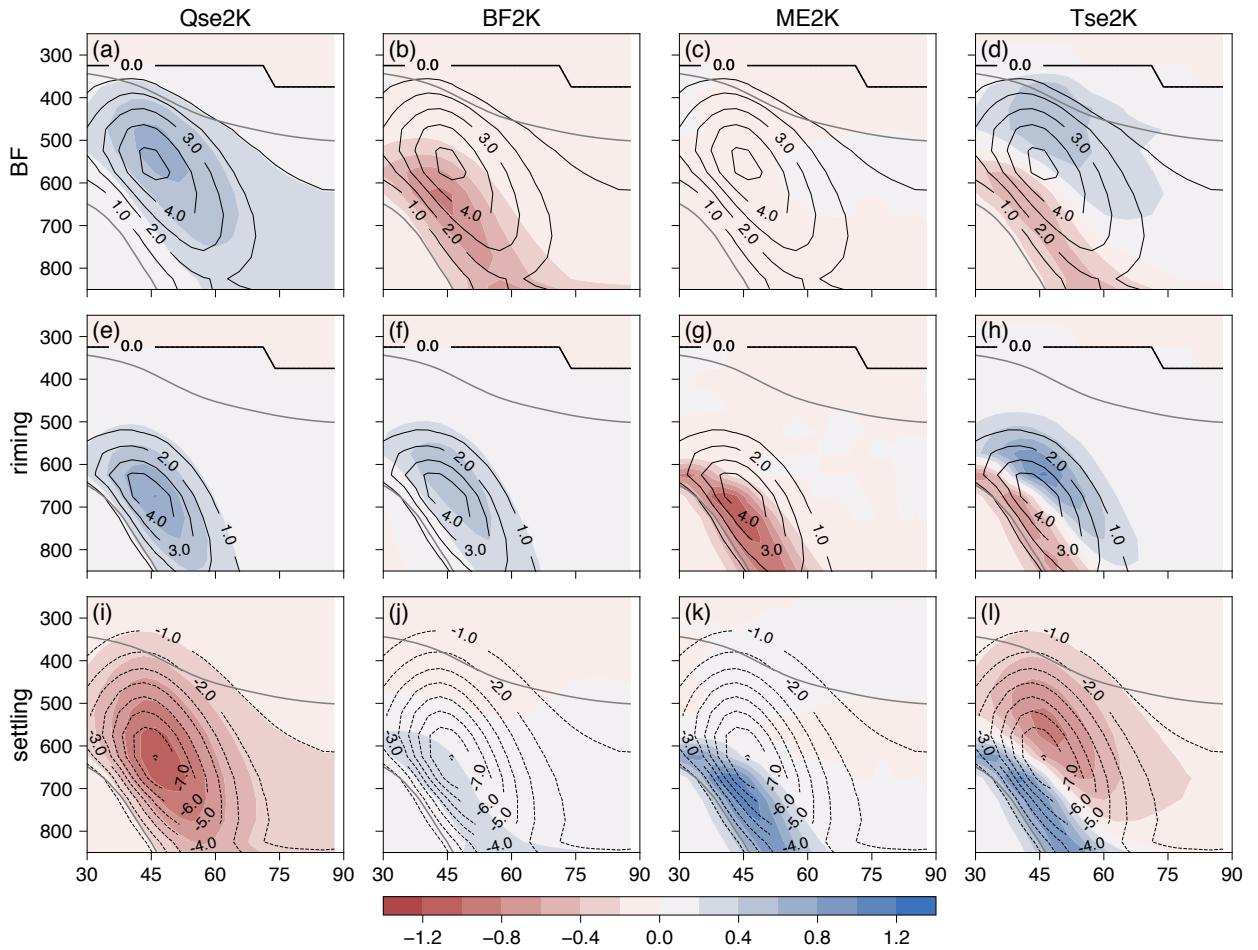
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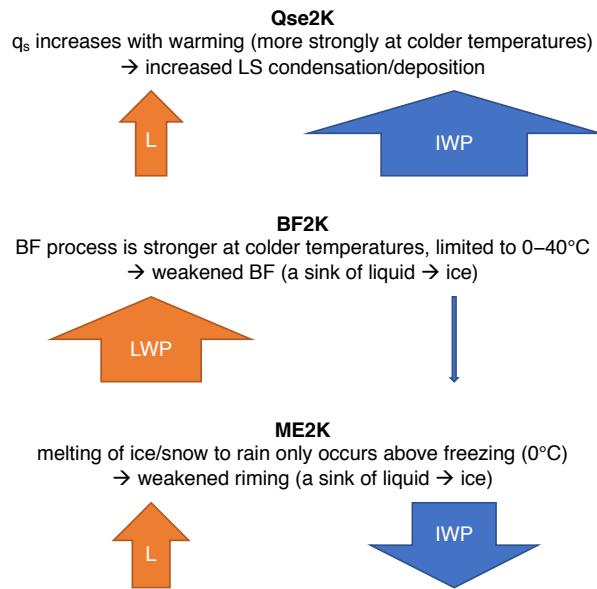
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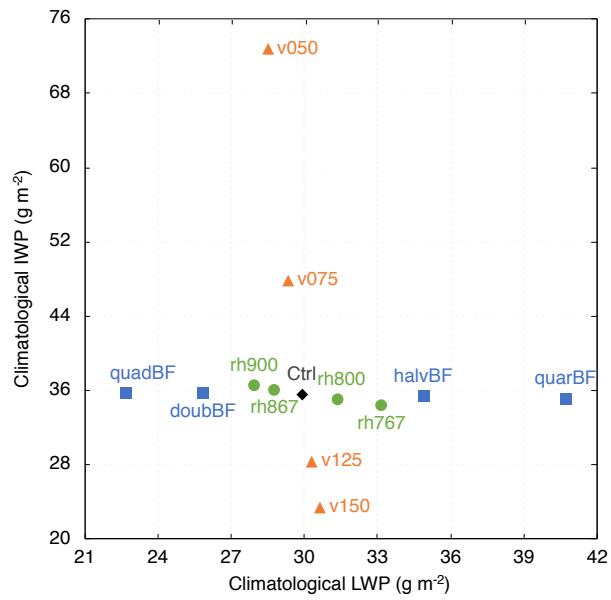
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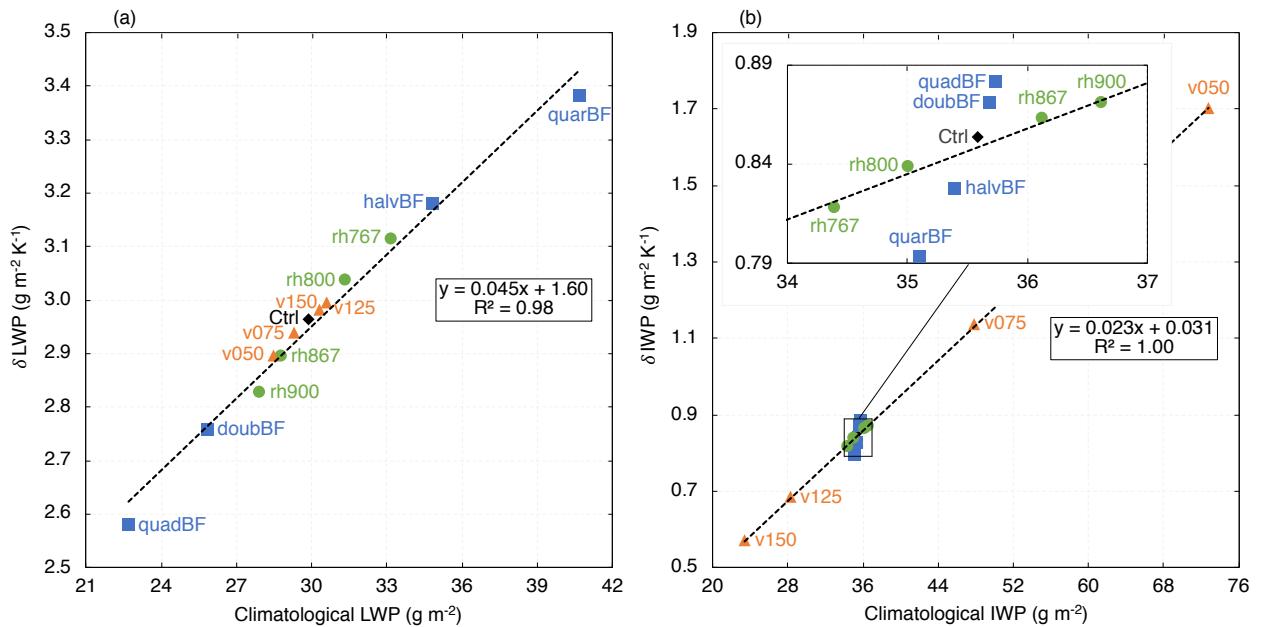
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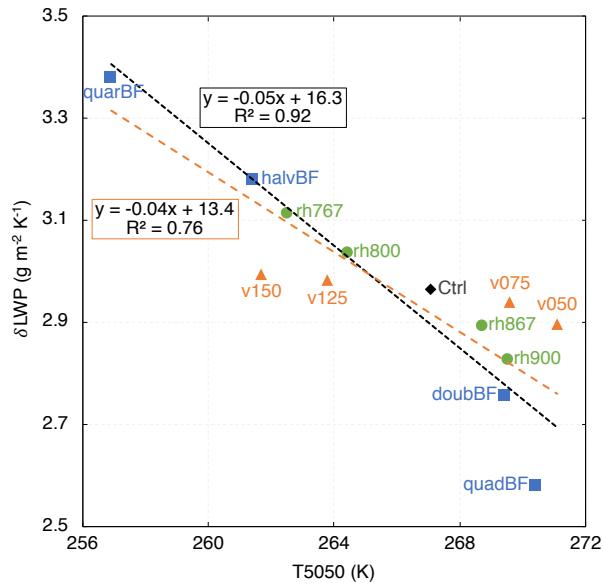
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