

# Response of marine post-frontal clouds to Gulf Stream variability

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23 **Key Points:**

- 24       • Stronger Gulf Stream SST gradients alter ocean-atmosphere interactions, producing  
25       deeper ice-phase clouds during cold air outbreaks.
- 26       • Stronger SST gradients drive more northwesterly cloud-top airmasses, while uniform  
27       SST increases vertical transport from the surface.
- 28       • Ocean-atmosphere coupling must be resolved to predict warming marine cloud  
29       feedback from distinct SST gradients and mean SST increases.

30

31 **Abstract**

32           Understanding how Gulf Stream variation influences cloud morphology is critical for  
33 evaluating cloud feedback in the western North Atlantic Ocean and beyond, where mesoscale  
34 air-sea interactions dominate. This study investigates the impact of altered mean sea surface  
35 temperature (SST) and SST gradients on post-frontal cloud characteristics during cold-air  
36 outbreaks, using the Weather Research and Forecasting (WRF) model. Three sensitivity  
37 experiments are conducted: a control simulation (default SST), Plus4 (uniform SST increase of  
38 4 K), and Gradplus (SST gradient enhanced by 25%, centered around mean SST). Results  
39 reveal distinctly different responses in boundary layer dynamics and cloud macro-physics. In  
40 Plus4, a warmer and moister boundary layer reduces total cloud cover but promotes larger  
41 cloud sizes and elongated cloud streets, with diminished liquid water and enhanced ice-phase  
42 hydrometeors. Conversely, Gradplus amplifies impacts in the upwind colder SST regions,  
43 yielding a drier, colder boundary layer, weaker energy transport, and higher liquid water path  
44 but reduced ice water content and cloud lines. Tracer analysis highlights that SST  
45 modifications alter air mass sources near cloud tops due to the entrainment of ambient air, with  
46 Plus4 amplifying boundary layer contributions to cloud-top regions. These findings underscore  
47 the spatially varying effects of SST gradients and mean SST on cloud organization and  
48 microphysics, emphasizing the need to resolve ocean-atmosphere coupling in global models  
49 to improve the prediction of marine cloud feedback under warming scenarios.

50

51 **Plain Language Summary**

52           This study uses weather model simulations to understand how warming and increasing  
53 temperature contrasts over the warm ocean current of the Gulf Stream affect clouds after cold  
54 air passes over the western North Atlantic. Warming the ocean surface by 4 degrees makes  
55 clouds larger but fewer, with a greater potential of increasing ice cloud content. Strengthening  
56 temperature contrasts near the Gulf Stream dries the air, weakening cloud-forming winds and  
57 slightly increasing the amount of liquid clouds while reducing ice clouds. Tracking airmass  
58 sources shows that warmer conditions draw more moisture from the ocean into clouds, while  
59 stronger temperature contrasts promote mixing in of drier air from above. These complex  
60 responses of clouds to the underlying surface conditions highlight why weather and climate  
61 models must better capture ocean-atmosphere interactions to improve predictions.

62

## 63 **1 Introduction**

64           The Gulf Stream (GS), a warm ocean current flowing northeastward off the North  
65 American coast in the Western North Atlantic Ocean (WNAO), plays a crucial role in modulating  
66 regional and global climate by transporting vast amounts of heat from the tropics to higher  
67 latitudes (Sweet et al., 1981; Kelly et al., 2010; Painemal et al., 2023). The strong sea surface  
68 temperature (SST) gradients associated with the GS drive mesoscale air-sea interactions that  
69 significantly influence atmospheric boundary layer dynamics and cloud formation (Minobe et al.,  
70 2008; Sorooshian et al., 2020). Recent studies have highlighted that variations in GS intensity  
71 and position, driven by anthropogenic warming and natural climate variability, can alter SST  
72 gradients and subsequently modify marine low-level cloud regimes (Joyce et al., 2009;  
73 Marshall et al., 2009; Smeed et al., 2018). These clouds, particularly post-frontal clouds (PFCs)  
74 formed during cold-air outbreaks (CAOs), are critical regulators of Earth's radiative balance, yet  
75 their response to changing GS characteristics remains inadequately understood (Tornow et al.,  
76 2021; Lamraoui et al., 2019; Li et al., 2022; Papritz et al., 2015; Naud et al., 2018).

77           Over the past decade, advances in high-resolution modeling and observations have  
78 improved our understanding of ocean-atmosphere coupling in the WNAO (Crosbie et al., 2024;  
79 Dadashazar et al., 2021; Dmitrovic et al., 2024; X.-Y. Li et al., 2022; Painemal et al., 2021;  
80 Sorooshian et al., 2020). Previous studies have demonstrated that warm fronts associated with  
81 the GS enhance turbulent heat fluxes, triggering secondary circulations that promote cloud  
82 organization (Lee et al., 2018; Liu et al., 2004; Liu et al., 2014; O'Neill et al., 2017; Sullivan et al.,  
83 2021). Minobe et al. (2008a) found a rain band over the GS in both satellite and general  
84 circulation model (GCM) outputs; however, this band would disappear when the input SST field  
85 is smoothed in the simulations, likely associated with a reduced SST gradient over the GS  
86 region in the model. Li et al. (2004) observed a cloud line over the GS region, which is formed  
87 by mesoscale solenoidal circulation induced by large surface thermal gradients. Also, Painemal  
88 et al. (2021) observed a convergence zone following the meandering path of GS and the area  
89 with the strongest SST gradients. However, most prior studies focused on the immediate cloud  
90 response to SST gradients, with limited attention to how long-term GS modifications—  
91 comprising both SST warming and gradient weakening—can affect cloud macro- and  
92 microphysical properties.

93           The challenges in exploring the impact of long-term GS variations on clouds are partly  
94 due to uncertainties arising from the coarse resolution of climate models and inadequate

95 satellite data resulting from shorter coverage period. Lee et al. (2018) examined several  
96 mechanisms through which SST bias in a spatially small region influences global model results.  
97 They found that, by imposing an SST bias between +6 K and +6.75 K from the atmosphere-  
98 ocean coupled model into the atmosphere-only model, the SST bias correction enhances the  
99 atmospheric updraft persistently even though the bias correction is only imposed in a small  
100 region. This suggests that exploring the impacts from decadal variations or climate changes of  
101 SST on tropospheric clouds in climate models is challenging, given the general bound of the  
102 SST variations is up to about 4 K, smaller than the coupled climate model bias (Pastor, 2022;  
103 Smeed et al., 2018). Chellappan et al. (2021) found that reanalysis data are representative for  
104 boundary layer height and surface fluxes but have inherited biases due to the too-wide GS in  
105 reanalysis data. Their findings also support that reanalysis data are good to initialize large-eddy  
106 simulations that can be used to reveal mechanisms linking the GS changes to PFC morphology,  
107 which are still unclear. High-resolution Weather Research and Forecasting (WRF) simulations,  
108 such as those conducted by Chen et al. (2022), are able to capture GS-induced SST anomalies  
109 that shape the morphology of cloud streets via BL instabilities, but the use of static SST  
110 conditions neglects realistic perturbations at decadal or shorter timescales of climate change.

111 Another often explored factor that affects clouds over the WNAO is the aerosol type,  
112 serving as cloud condensation nuclei, which are critical to cloud nucleation and properties  
113 (Petters & Kreidenweis, 2007). The WNAO region, located downwind of major coastal cities  
114 such as New York, has many different aerosol sources leading to distinct aerosol types  
115 dependent on season and atmospheric circulation patterns (Andrea F. Corral et al., 2021; H.  
116 Liu et al., 2025; Seckar-Martinez et al., 2025; Sorooshian et al., 2019). For instance, the  
117 wintertime has more offshore flow from the U.S. East Coast leading to urban emissions  
118 influence over the WNAO (Dadashazar et al., 2022), whereas spring and summer months are  
119 characterized by more influence from biomass burning emissions (Edwards et al., 2021; Mardi  
120 et al., 2021) and dust including from Africa (Ajayi et al., 2024; Aldhaif et al., 2020). H. Liu et al.,  
121 (2025) showed simulated aerosols over the WNAO and evaluated the simulation results against  
122 Aerosol Cloud meTeorology Interactions oVer the western ATlantic Experiment (ACTIVATE)  
123 field campaign measurements in 2020. They found that sea salt dominates BL aerosol mass  
124 concentration and optical depth, peaking in summer, followed by organics (for mass) and  
125 sulfate (for optical depth). Also, adjusting wildfire smoke injection heights to the mid-  
126 troposphere (as opposed to the BL top) improves model agreement with observations of

127 western U.S. wildfire smoke aerosols transported to the WNAO in summer. Given that  
128 predominant aerosol types are highly related to air mass origin, it is critical to characterize the  
129 nature of air masses over the WNAO for PFC studies. Previous studies mentioned that SST  
130 variations may alter boundary layer turbulence (e.g., Chen et al., 2022), but it remains unclear  
131 whether the representation of the SST field in those models leads to considerable changes of  
132 air mass sources impacting simulated PFCs.

133 In the above context, this study focuses on three scientific questions: (1) How does an  
134 increase in mean SST (e.g., +4 K, global warming) and an enhanced SST gradient (i.e., induced  
135 by enhanced GS strength) individually modify the PFC liquid water path (LWP), cloud fraction,  
136 and phase partitioning over the GS? (2) What are the dominant mechanisms—boundary layer  
137 turbulence, large-scale ascent, or microphysical feedbacks—mediating these PFC responses?  
138 and (3) Do air mass sources shift under the altered GS conditions?

139 To answer these questions, this study employs high-resolution WRF (v4.2) simulations  
140 with tailored SST perturbations (+4 K mean warming and 25% SST anomalies increase relative  
141 to the control experiment) to systematically evaluate PFC responses. Building on Chen et al.  
142 (2022), we introduce two key methodological innovations: (1) Isentropic analysis of SST impact  
143 zones to isolate regional energy transport, and (2) Lagrangian tracer analysis for quantifying  
144 air mass source changes near cloud-top heights. Our approach aims to provide mechanistic  
145 implications into how GS changes may alter PFCs under future warming scenarios.

## 146 **2 Methods**

### 147 **2.1 Case Overview**

148 This study focuses on a CAO event over WNAO observed on 1 March 2020 during the  
149 Aerosol Cloud Meteorology Interactions over the western Atlantic Experiment (ACTIVATE)  
150 field campaign (Sorooshian et al., 2019). The CAO event was preceded by two cold front  
151 passages on 27 February (primary) and 29 February (secondary), with northerly winds  
152 advecting dry, cold air over the warm GS water. This triggered intense surface heat fluxes  
153 ( $>300 \text{ W m}^{-2}$ ) over the GS and mesoscale solenoidal circulation. Post-frontal clouds developed  
154 after the frontal passage, emerged at 0800 Eastern Standard Time (EST) on 1 March 2020, and  
155 persisted for several hours. Chen et al. (2022) validated their WRF simulations against GOES-  
156 16 satellite retrievals and ACTIVATE dropsonde measurements (0945–1045 EST),  
157 demonstrating consistency in cloud morphology (e.g., cloud street alignment) and boundary

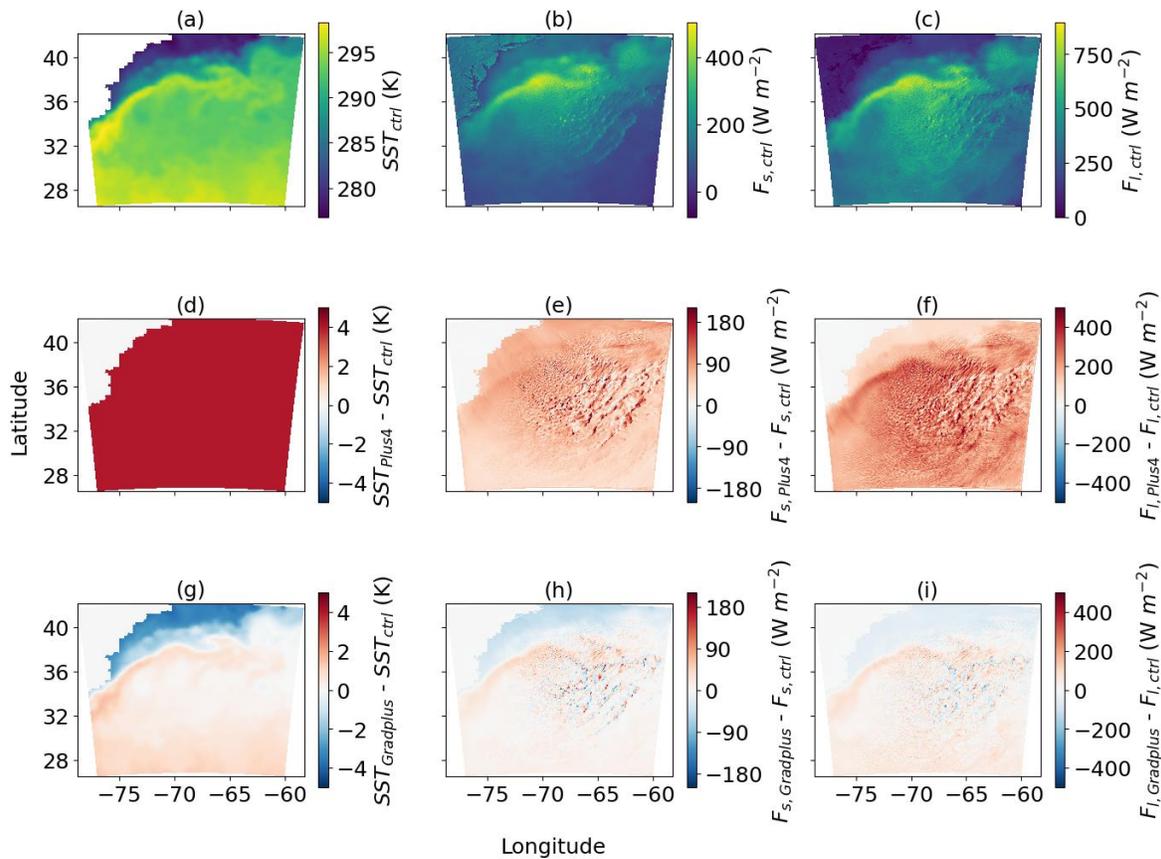
158 layer thermodynamics (e.g., inversion height and moisture profiles), despite a slight cold bias in  
159 the simulated near-surface temperatures that were attributed to the ERA5 reanalysis inputs.

160 This study builds on the same CAO case analyzed by Chen et al. (2022), leveraging its  
161 robust validation with both satellite and in situ observational datasets. The well-documented  
162 cloud evolution over the GS, characterized by cloud structures influenced by SST structure, for  
163 this case provides an ideal framework for investigating SST-driven impacts on PFC  
164 morphology.

165

## 166 **2.2 WRF Setups**

167 The Weather Research and Forecasting (WRF v4.2) model was configured with two  
168 nested domains. The outer domain covers a 1650 km × 1650 km region of the WNAO, with  
169 simulations conducted at 3 km horizontal resolution. The inner domain corresponds to a 450  
170 km × 450 km at 1 km horizontal resolution, and specifically intended for studying the cloud  
171 street dynamics. The simulations are conducted at 150 vertical levels (up to 100 hPa), with finer  
172 resolution in the lower troposphere (~46 m layer thickness up to 6 km) versus above. The  
173 model simulations were performed for the period of 0600 UTC March 1 – 0000 UTC March 2.  
174 Key physics schemes include the Morrison two-moment cloud microphysics scheme (Morrison  
175 et al., 2005), the Yonsei University PBL scheme (Hong & Lim, 2006), and the Rapid Radiative  
176 Transfer Model for GCMs (RRTMG) longwave and shortwave radiation schemes (Iacono et al.,  
177 2008). Subgrid convection parameterization is disabled in the simulations to better resolve  
178 mesoscale processes at gray-zone resolutions (Field et al., 2017). Initial and boundary  
179 conditions are derived from ERA5 reanalysis. This configuration aligns with the control  
180 simulation in Chen et al. (2022), which demonstrated strong consistency with satellite and  
181 dropsonde observations, thereby validating the model’s ability to capture PFC evolution. The  
182 other details of the simulation setups have been described in Chen et al. (2022).



183

184 **Figure 1.** (a) SST, (b) surface sensible heat flux, and (c) surface latent heat flux from the control  
 185 simulation of out domain at 0900 EST. Differences of (d) SST, (e) surface sensible heat flux, and  
 186 (f) surface latent heat flux between the Plus4 and Ctrl simulations. Differences of (g) SST, (h)  
 187 sensible heat flux, and (i) latent heat flux between the Gradplus and Ctrl simulations at 0900  
 188 EST on 1 March 2020.

189

190 Figures 1a-1c show the spatial distribution of SST and surface fluxes from the outer  
 191 domain. The SST enhancement associated with the GS leads to a large SST gradient at the  
 192 north edge of the GS. There is a strong surface flux band along the GS, and strong gradients  
 193 are present in both surface sensible heat flux ( $F_s$ ) and latent heat flux ( $F_l$ ). Hereafter, the  
 194 simulation with this SST setup is referred to as “Ctrl”.

195

## 196 2.3 Sensitivity Experiments

197 Two sensitivity experiments are conducted to isolate the individual aspects of SST changes  
198 and their impacts on PFCs:

- 199 1. Mean temperature experiment: SST is uniformly increased by 4 K across the domain,  
200 including the GS core region, to mimic extreme warming scenarios (the scenario with  
201 intensive fossil fuel burning and rapid economic growth), which is expected to enhance  
202 boundary layer moisture and reduce stability (hereafter referred to as “Plus4”).
- 203 2. Temperature gradient experiment: The magnitude of positive and negative SST  
204 anomalies, relative to the domain mean, is increased by 25% while the domain-mean  
205 SST is unchanged, intensifying thermal contrasts to evaluate the potential impact on  
206 circulation induced by SST gradients (hereafter referred to as “Gradplus” ).

207 Figures 1 (d-f) and (g-i) show the differences in SST and surface fluxes between the  
208 control simulation and the Plus4 and Gradplus simulations, respectively, at 0900 EST. The  
209 Plus4 simulation has a uniform SST increase. The values of  $F_s$  and  $F_l$  are highly enhanced by  
210 the SST increase of 4 K. The spatial variations in  $F_s$  arise from the cloud structure, which  
211 induces variations in near-surface horizontal winds, vertical velocity, and air temperature,  
212 thereby impacting the surface sensible heat flux. The spatial variations in  $F_l$  also show the  
213 structure influenced by clouds but the gradient at the north edge of the GS is larger than  $F_s$ ,  
214 indicating a larger effect on  $F_l$  from the warm water transported by the GS. The Gradplus  
215 simulation aims to increase the gradient of SSTs through their anomalies relative to the domain  
216 mean values. Thus, the warm region gets warmer and cold region gets colder, as reflected in  
217 Figure 1g. The spatial distributions of  $F_s$  and  $F_l$  driven by the changes in SST anomalies have  
218 some small-scale variations related to the turbulence above the ocean surface. In Section 3,  
219 we explore the responses of boundary layer and clouds to these two aspects of SST variations.

220

## 221 **2.4 FLEXPART Back Trajectory Tracking**

222 We also use the FLEXPART-WRF model (version 3.3.2) to track the backward trajectory  
223 of air mass reaching the model domain (Brioude et al., 2013). FLEXPART-WRF is a Lagrangian  
224 particle dispersion model coupled with WRF enabling high-resolution simulations of  
225 atmospheric transport and dispersion. By using WRF's output—including wind fields,

226 temperature, humidity, and turbulence parameters—FLEXPART-WRF tracks the trajectories of  
227 numerous air parcels (treated as particles) to model particle transport and to ultimately give  
228 insight into source-receptor relationships. For backward trajectory analysis, the model reverses  
229 the temporal integration of wind fields using the WRF output in the time frequency of 15 min,  
230 allowing particles to move backward in time from a receptor location (e.g., ACTIVATE aircraft  
231 sampling location) to identify potential emission sources. We use the turbulence option  
232 (TURB\_OPTION=1 and CBL\_SCHEME=1) that internally calculates boundary layer turbulent  
233 mixing using the Hanna turbulence scheme (Hanna, 1982) and allow the skewed turbulence  
234 (Brioude et al., 2013).

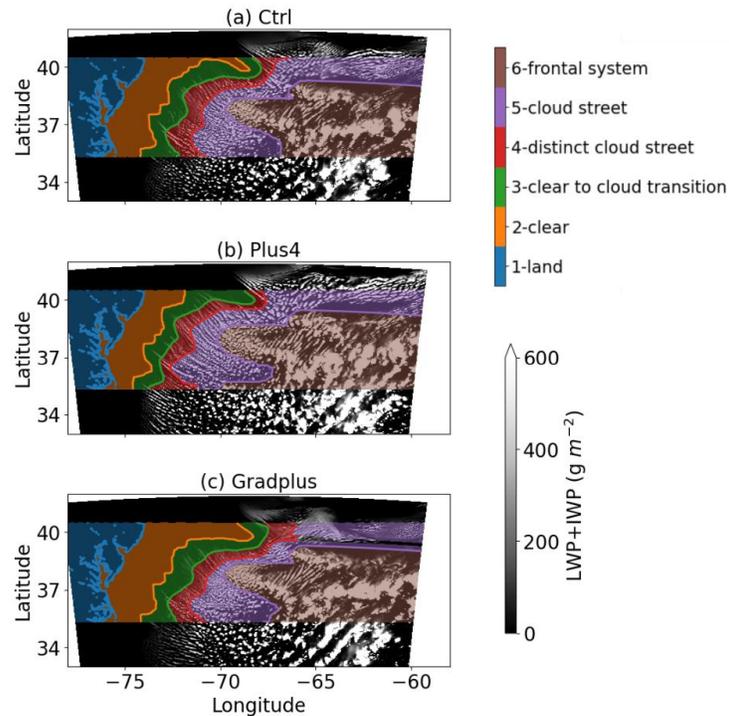
235

### 236 **3. Results**

#### 237 **3.1 Response of Cloud Morphology to SST Variations**

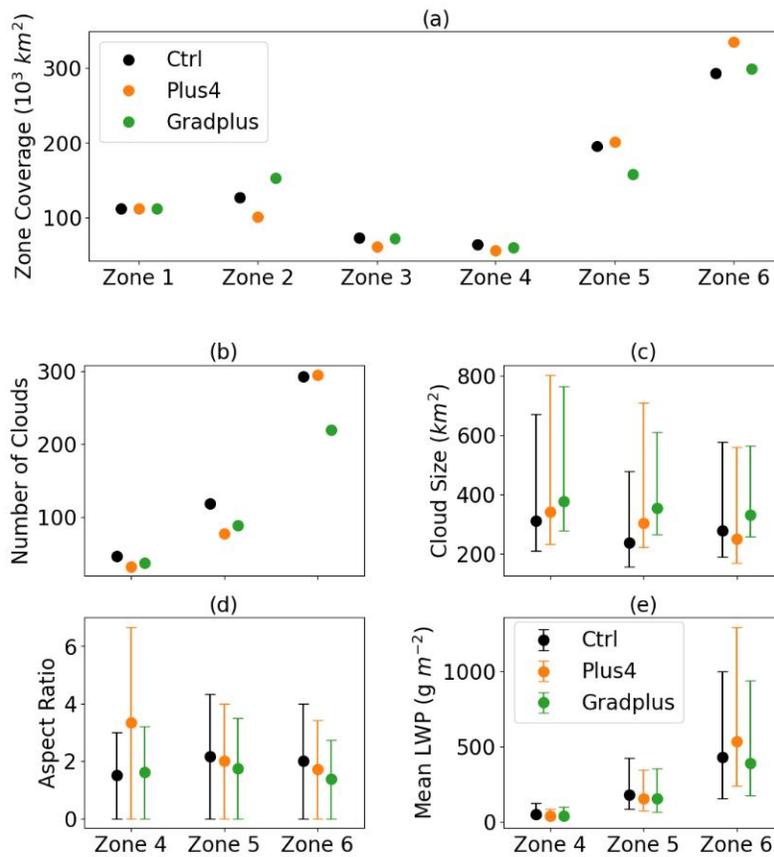
238 The cloud structure behind the frontal system exhibits diverse horizontal spatial  
239 patterns. Based on the classification criteria consistent with Chen et al. (2022), six zones are  
240 classified in each of the three experiments (Figure 2) according to thresholds of smoothed  
241 liquid water path. The cloud water path is smoothed by a uniform filter of 30 grids, which  
242 removes the high-frequency variations in the complex cloud structure. Then we separate out  
243 the clear coastal and zones 3 – 6 with the values of smoothed water path of <40, 40 – 200,  
244 200 – 500, and >500 g m<sup>-2</sup>, respectively. Zone 1 corresponds to the eastern coastal landmass  
245 of the United States. Zone 2 represents a cloud-free region over the ocean. Zones 3–5  
246 correspond to regions with PFCs, whereas zone 6 identifies the frontal system, which is  
247 dominated by extensive cloud cover.

248 PFCs west of the frontal zone (Zones 3–5) display distinct cloud morphologies.  
249 Between the clear-sky region (Zone 2) and the distinct cloud streets (Zone 4), Zone 3 serves as  
250 a transition zone where cloud streets appear from the upwind clear skies. Zone 4 features well-  
251 defined, continuous cloud street structures aligned with prevailing winds by a small angle  
252 (Chen et al., 2022). Near the frontal boundary (Zone 5), cloud streets transition into fragmented  
253 patches, manifested as open-cellular convection.



254  
 255 **Figure 2.** Spatial distribution of liquid water path (LWP) and ice water path (IWP) from (a) Ctrl,  
 256 (b) Plus4, and (c) Gradplus experiments (white shading) at 0900 EST on March 1, 2020.  
 257 Overlaid color shading indicates the zone classification also used by Chen et al. (2022).

258  
 259 We test the hypothesis that variations in SST patterns influence coverage of each zone  
 260 (Figures 2b, 2c) by comparing the different experiments. Relative to the Ctrl experiment, the  
 261 Plus4 simulation shows a 20.8% reduction in spatial area of Zone 2, a 16.3% reduction in Zone  
 262 3, a 12.7% reduction in Zone 4, a 2.6% expansion in Zone 5, and a 14.3% expansion in Zone 6  
 263 (Figure 3a). Also, the total area of PFCs (Zones 3–5) remains largely unchanged overall (only a  
 264 4.5% reduction), despite a spatial shift of clouds toward the coastline. These changes suggest  
 265 a significant enhancement in frontal cloud development (Zone 6) and a contraction of clear-sky  
 266 areas (Zone 2). This cloud redistribution is likely driven by warmer SSTs and elevated boundary  
 267 layer moisture, which promote convective cloud formation and frontal zone intensification.



268  
 269 **Figure 3.** (a) Areal coverage of PFC zones 1-6 from the control and two sensitivity simulations.  
 270 (b) number of cloud objects of PFC zones 4-6, (c) averaged cloud size of PFC zones 4-6, (d)  
 271 averaged aspect ratio of cloud horizontal mask of PFC zones 4-6, and (e) mean LWP of each  
 272 cloud object averaged in corresponding PFC zone at 0900 EST on 1 March 2020.

273

274 In the Gradplus experiment, the areal coverage of Zone 5 (cloud street zone) decreases  
 275 by 19.3%, while Zone 2 (clear zone) expands by 20.2%, reflecting a decline in organized PFC  
 276 structures and an increase in coastal clear-sky coverage. Zones 3, 4, and 6 show small  
 277 changes (1.5% reduction, 6.4% reduction, and 2.1% expansion, respectively), compared to  
 278 the Ctrl simulation. These results indicate that intensified SST gradients suppress cloud  
 279 organization within PFCs, favoring localized clear-sky expansion at the expense of coherent  
 280 cloud structures.

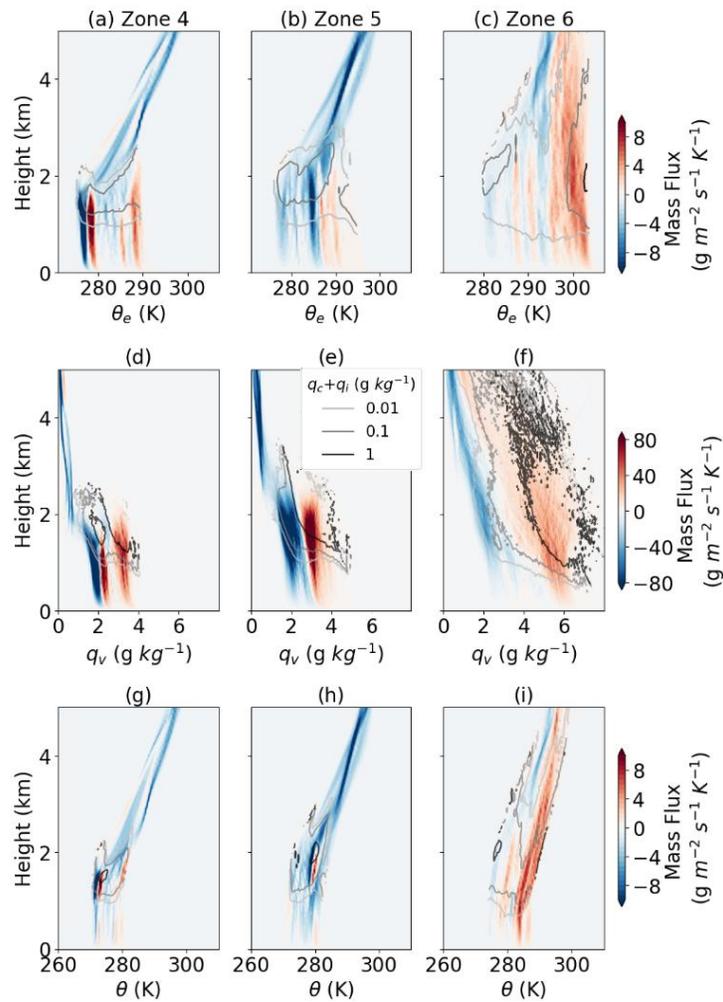
281 Figures 3b–3e illustrate changes in cloud object size under the altered SST patterns.  
 282 Cloud objects are defined here as regions where the ratio of LWP to smoothed LWP exceeds

283 0.8 (Chen et al., 2022). In the Plus4 experiment, PFCs exhibit a reduced number of cloud  
284 objects but a larger cloud size, while frontal clouds show minimal changes of cloud number  
285 despite a reduced cloud size. The aspect ratio in Figure 3d, which quantifies cloud street  
286 elongation, increases by 122.2% in Zone 4 (distinct cloud streets) indicating enhanced  
287 longitudinal organization of cloud streets in Zone 4. Mean LWP decreases by 26.0% in Zone 4  
288 and 15.5% in Zone 5 but rises sharply by 24.4% in Zone 6, reflecting more vigorous frontal  
289 clouds. These results suggest that warmer SSTs promote fewer but larger PFCs with elongated  
290 cloud streets in Zone 4, while frontal clouds become vertically deeper (evidenced by LWP  
291 increases) despite reduced horizontal extents.

292 In the Gradplus experiment, both PFCs and frontal clouds exhibit larger individual  
293 cloudy areas but a smaller number of clouds compared to the Ctrl simulation. The aspect ratio  
294 shows an overall reduction, except in Zone 4 where it demonstrates a slight increase. Mean  
295 LWP values remain consistently lower than those in the Ctrl simulation. The meteorological  
296 phenomena in Zones 5 and 6 may be attributed to their position within positive SST anomaly  
297 regions, where the primary SST gradients associated with the GS are displaced farther  
298 northwestward.

### 299 **3.2 Response of energy transport to SST variations**

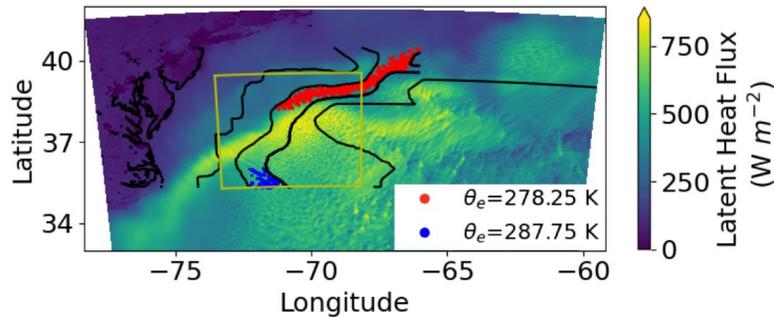
300 This section explores the mass and energy transport in frontal systems and PFCs.  
301 Previous studies have relied on isentropic analysis of moist convection to identify energy  
302 transport through diabatic processes, such as convection (Chen et al., 2023; Pauluis &  
303 Mrowiec, 2013). This method involves averaging the mass flux and other examined variables at  
304 the isopleths of equivalent potential temperature ( $\theta_e$ ) to isolate upward and downward motions  
305 that contribute to energy transport through their net effects. In other words, adiabatic vertical  
306 motions with oscillations (such as gravity waves) are filtered out by this averaging to derive  
307 irreversible energy/mass fluxes. Here, we apply the same method to Zones 4–6 in Figures 4a-  
308 4c to illustrate energy transport in these regions. We also show the mass flux at the isopleths  
309 of  $q_v$  (Figures 4d-4f) and  $\theta$  (Figures 4g-4i) to describe the properties of the air parcels.



310  
 311 **Figure 4.** Isentropic analysis of mass flux in (a) Zone 4, (b) Zone 5 and (c) Zone 6. (d-f) Mass  
 312 flux over isopleths of  $q_v$  in (d) Zone 4, (e) Zone 5, and (f) Zone 6. Mass flux over isopleths of  $\theta$   
 313 in (g) Zone 4, (h) Zone 5, and (i) Zone 6. All results are from the Ctrl simulation at 0900 EST on 1  
 314 March 2020.

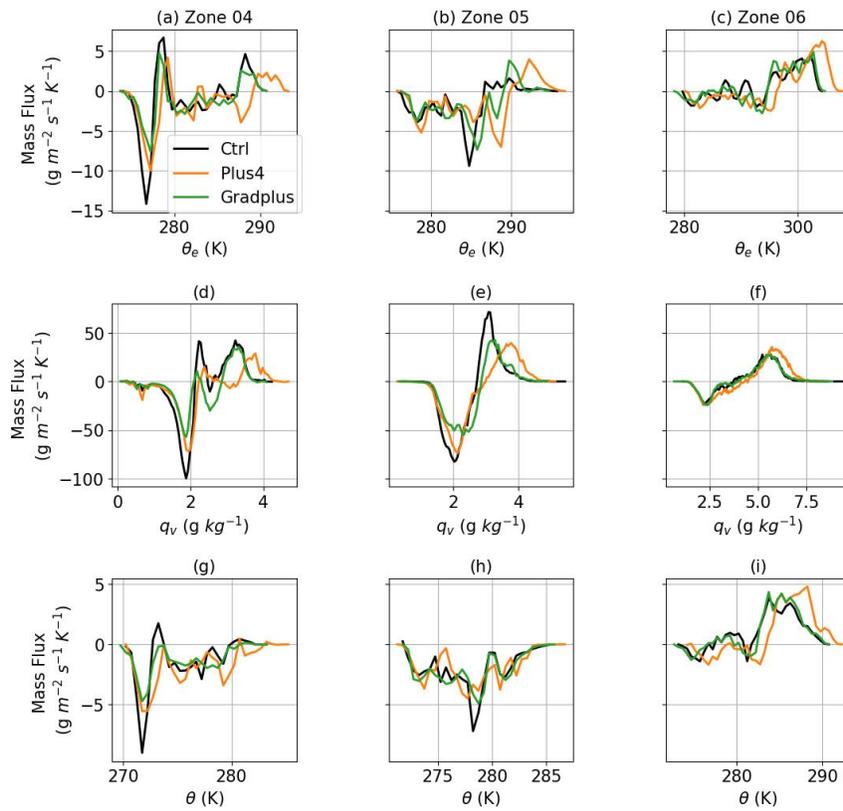
315  
 316 Zones 5–6 exhibit a primary upward band at high  $\theta_e$  and a downward band at low  $\theta_e$ ,  
 317 consistent with previous studies on moist convection (Chen et al., 2023; Pauluis & Mrowiec,  
 318 2013). As parcels with high  $\theta_e$  ascend and those with low  $\theta_e$  descend, the net energy transport  
 319 is upward. However, in Zone 4, two distinct upward bands are observed at different  $\theta_e$  values:  
 320 one at  $\sim 287.75$  K (similar to the upward band in Zone 5) and another at 278.25 K, which is  
 321 notably lower. We speculate these two bands relate to different processes, which will be

322 discussed below. Additionally, in Zones 4 and 5, a downward mass flux band is present above  
 323 the clouds, terminating at lower altitudes near the cloud top. This may be attributed to large-  
 324 scale downward motion within PFCs.



325  
 326 **Figure 5.** Spatial distribution of latent heat flux. Black contour denotes the zone classification  
 327 in Figure 2. Yellow box denotes the inner domain. Red and blue dots mark the location of the  
 328 two  $\theta_e$  peaks in Figure 4a at 0900 EST on 1 March 2020.

329  
 330 Since  $\theta_e$  comprises two components, potential temperature ( $\theta$ ) and water vapor mixing  
 331 ratio ( $q_v$ ), we also display the mass flux at isopleths of these variables in the second and third  
 332 rows of Figure 4, respectively. When using  $q_v$  isopleths, the mass flux magnitude is  
 333 significantly larger than with isentropic coordinates or  $\theta$  isopleths, suggesting a stronger  
 334 transport of  $q_v$  by mass flux than  $\theta$ , because upward and downward transports compensate  
 335 on isopleths, and the mass fluxes defined on isopleths represent the fluxes that effectively  
 336 transport the corresponding quantities. Cloud cores (black contours) appear directly above the  
 337 upward bands in the  $q_v$  analysis, implying that these clouds draw their moisture from the  
 338 boundary-layer upward motion. In Zone 4, two separate cloud cores align with the two upward  
 339 bands. When mass flux is averaged over  $\theta$  isopleths, it is large within and above the clouds,  
 340 indicating that thermal energy transport occurs primarily at the altitudes of clouds.



341  
 342 **Figure 6.** Averaged mass flux below 2 km at (a-c) isentropic coordinates, (d-f) isopleths of  $q_v$ ,  
 343 and (g-i) isopleths of  $\theta$  for (a, d, and g) Zone 4, (b, e, and h) Zone 5, and (c, f, i) Zone 6 at 0900  
 344 EST on 1 March, 2020.

345  
 346 The two separate upward mass flux bands in Figures 4a, 4d, and 4g suggest that Zone  
 347 4 involves distinct convection processes. We locate the two  $\theta_e$  values associated with large  
 348 mass fluxes on the spatial distribution of surface latent heat fluxes in Figure 5. The higher  $\theta_e$   
 349 band is located south of the GS region, at a distance from the GS itself, while the lower  $\theta_e$   
 350 band is positioned at the northern GS edge. Because the higher  $\theta_e$  value (287.75 K) in the  
 351 southern upward band is comparable to the  $\theta_e$  values in the upward bands of Zones 5 and 6,  
 352 we speculate that the southern upward energy transport band may represent an extension of  
 353 the frontal system, indicative of large-scale frontal influences. The low  $\theta_e$  upward mass flux  
 354 band at the northern GS edge is likely linked to local SST gradients, where mesoscale

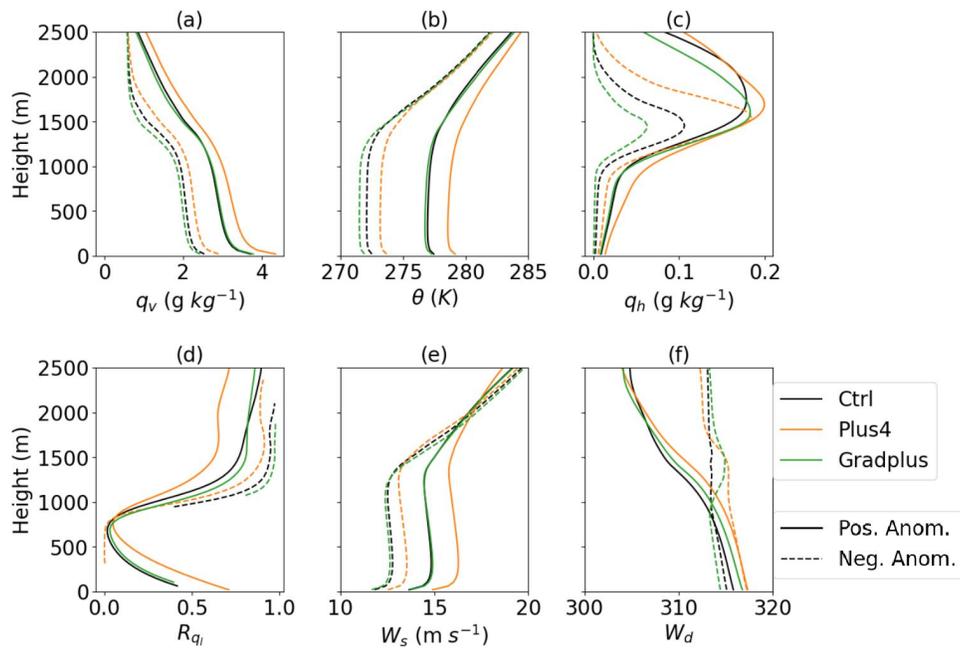
355 circulations may develop and influence the broader region, potentially triggering new  
356 convective upward energy transport. We further explore the secondary circulation associated  
357 with the SST gradients in Section 3.3.

358 In Figure 6, we explore the impacts of SST variations on energy transport in Zones 4–6,  
359 with mass flux averaged below 2 km. In the Plus4 experiments, energy transport bands shift to  
360 higher isopleth values of  $\theta_e$ ,  $q_v$  and  $\theta$  across all zones. This suggests that warmer SST  
361 scenarios drive convective energy transport to higher energy levels, accelerating the process.  
362 The energy transport magnitudes in the Gradplus experiments are similar to those in the Ctrl  
363 experiments, but mass flux peaks are reduced in the lower  $\theta_e$  band in Zone 4, indicating a  
364 weakened transport efficiency under the stronger SST gradients. The underlying mechanisms  
365 are discussed in Section 3.3. Zones 5 and 6 exhibit no significant differences between Ctrl and  
366 Gradplus experiments, likely because these zones lie outside the GS-influenced region.

367 In summary, the analysis of mass fluxes on isentropic coordinates indicates two distinct  
368 influences on the development of the cloud street zone: one influenced by the frontal system  
369 and the other by the SST gradient. The upward mass-flux band at the northern edge of the Gulf  
370 Stream illustrates the role of the SST gradient in affecting the post-frontal cloud system.

### 371 **3.3 Response of Boundary Layer Properties to SST Variations**

372 In this section, we focus on the inner domain where a contrast in SST anomalies is  
373 present. Figure 7 shows the vertical profiles of boundary layer properties in the negative-  
374 anomaly region (north of the GS) and the positive-anomaly region (south of the GS) at 0900  
375 EST. The Plus4 experiment results in higher boundary layer  $q_v$ ,  $\theta$ , hydrometeor mixing ratio,  
376 and wind speed in both regions (Figures 7a, 7b, 7c, and 7e). Additionally, the fraction of liquid-  
377 phase hydrometeors is reduced, indicating that clouds are elevated with more ice-phase  
378 hydrometeors in both regions (Figure 7d). Wind direction shifts clockwise with the increased  
379 SST (Figure 7f), likely due to a strengthened Coriolis effect by the enhanced wind speed.



380

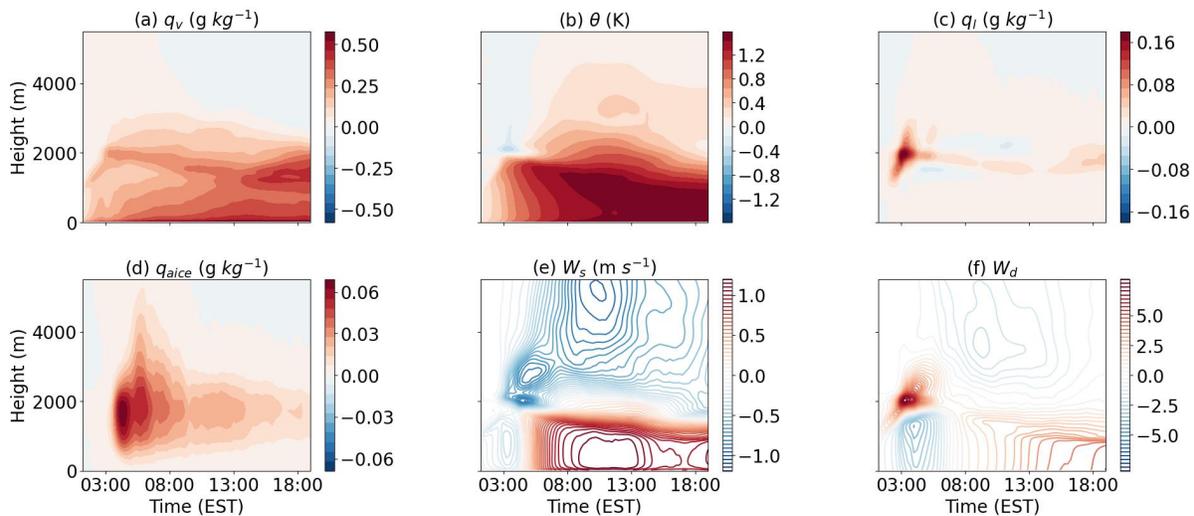
381 **Figure 7.** Vertical profile of (a) water vapor mixing ratio ( $q_v$ ), (b) potential temperature ( $\theta$ ), (c)  
 382 hydrometeor mixing ratio ( $q_h$ ), (d) the ratio of ice-phase hydrometeor to liquid-phase  
 383 hydrometeor ( $R_{qi}$ ), (e) wind speed ( $W_s$ ), and (f) wind direction ( $W_d$ ) at 0900 EST.

384

385 Figure 8 displays the time series of differences between the Plus4 experiment and the  
 386 Control simulation. The elevated boundary layer  $q_v$ ,  $\theta$ , cloud liquid and ice water mixing ratio,  
 387 and wind speed persist from sunrise to sunset. In the early morning before sunrise, liquid-  
 388 phase cloud content is slightly higher in the Plus4 experiment, but the differences diminish  
 389 after sunrise. The all-ice-phase hydrometeor content in Plus4 is significantly larger than in the  
 390 Control experiment, suggesting that the higher SST promotes deeper cloud development.

391

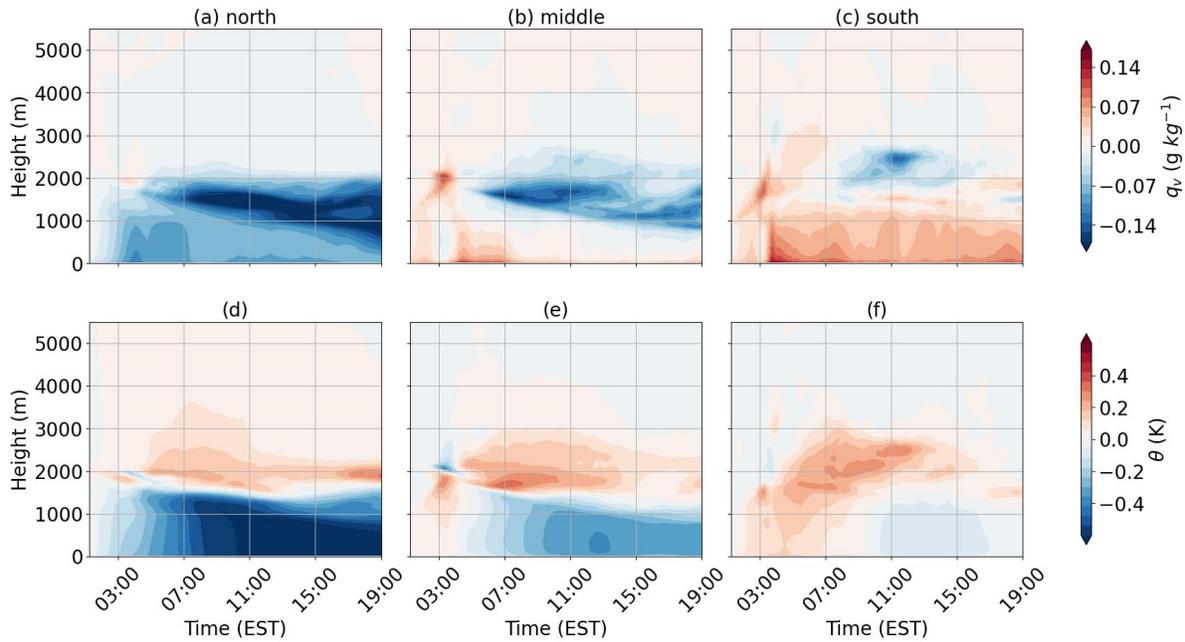
392 Figure 7 demonstrates that SST anomaly impacts in the Gradplus experiment are  
 393 notably weaker than in Plus4 at 0900 EST. The temporal variations in Gradplus are less  
 394 straightforward, as shown in Figures 9 – 11, with the magnitude of the differences being much  
 395 smaller than in the Plus4 experiment and the sign being less consistent. The GS region  
 396 ("middle" in Figures 9–11) corresponds to latitudes with maximum SST per longitude, with  
 areas north and south of it labeled accordingly.



397  
 398 **Figure 8.** Time series of the differences between the Plus4 and Ctrl experiments. Color shading  
 399 indicates (a)  $q_v$ , (b)  $\theta$ , (c) liquid water mixing ratio ( $q_l$ ), (d) mixing ratio of all ice ( $q_{ice}$ ), (e) wind  
 400 speed ( $W_s$ ), and (f) wind direction ( $W_d$ ).

401  
 402 In Figure 9, moving from north to south, the intensified SST gradient shifts their effects  
 403 from negative to positive in the values of  $q_v$  and  $\theta$  within the boundary layer, indicating  
 404 suppression of temperature and moisture in the north (due to a 25% SST reduction) and  
 405 enhancement in the south (due to an SST increase). At the top of the boundary layer, strong  
 406 negative effects on  $q_v$  are observed in the southern and middle regions, with altitude  
 407 decreasing over time, consistent with declining boundary layer heights (Chen et al., 2022).  
 408 Coupled with positive effects from Gradplus on wind speeds at these altitudes (Figures 10a–  
 409 10c), we speculate that drier air advection from the north causes the negative  $q_v$  anomaly at  
 410 higher altitudes. Additionally, strong divergence at the top of the boundary layer in Figures  
 411 10g–10i provides evidence of drier air transport by intensified northerly winds.

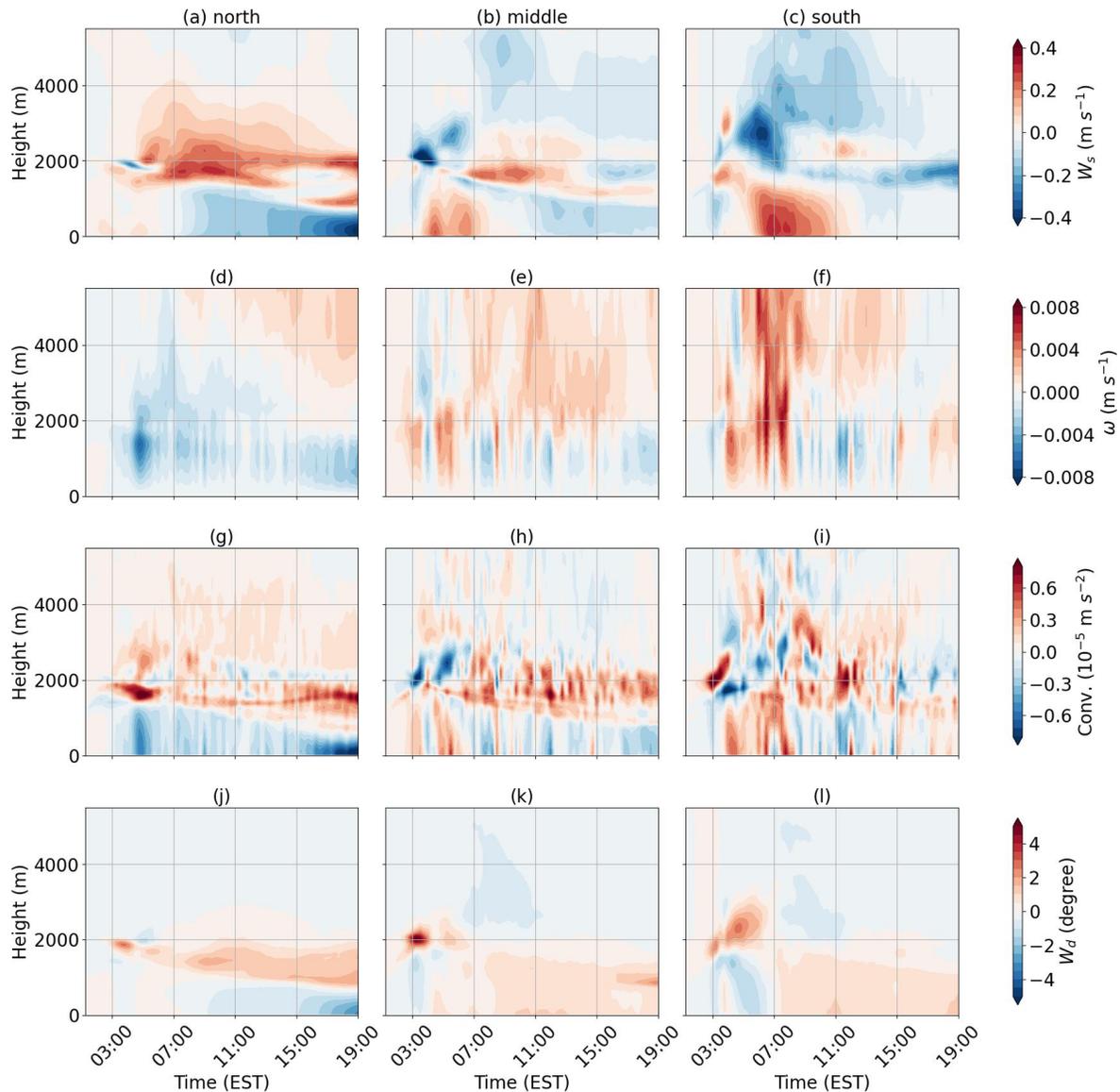
412



413  
 414 **Figure 9.** Time series of the differences of (a-c)  $q_v$  and (d-f)  $\theta$ , between Gradplus and Ctrl  
 415 experiments in the three regions: (a and d) north of the GS, (b and e) at the latitudes of GS, and  
 416 (c and f) south of the GS.

417  
 418 However, despite stronger advection observed above the boundary layer, vertical  
 419 transport of colder air is largely confined to the boundary layer in the northern region (Figure  
 420 9d), leaving positive anomalies at higher altitudes. Meanwhile, stronger convection lifts warm  
 421 air from the surface to above the boundary layer between 0400 and 0800 EST in the southern  
 422 region (Figures 10e–10f). Stronger convergence observed during 0400–0800 EST in the  
 423 southern region aligns with the enhanced vertical velocity (Figures 10g–10i). This suggests that  
 424 transport of thermal energy is from south to north, opposite to the direction of moisture  
 425 transport.

426

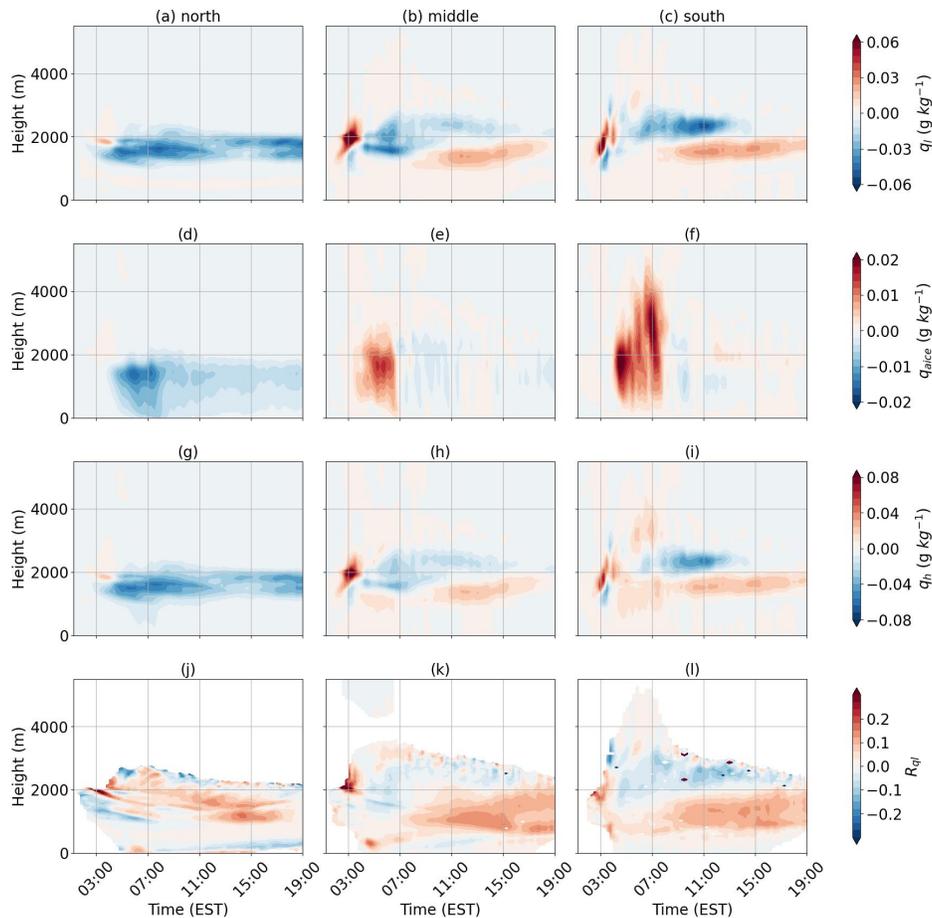


427  
 428 **Figure 10.** Time series of the differences of (a-c) wind speed, (d-f) vertical velocity, (g-i)  
 429 convergence, and (j-l) wind direction between Gradplus and Ctrl experiments in the three  
 430 regions: (a, d, g and j) north of the GS, (b, e, h and k) at the latitudes of GS, and (c, f, i and l)  
 431 south of the GS.

432

433 Two processes govern interactions between air above the negative SST anomaly region  
 434 in the north and the positive SST anomaly region in the south: 1) North-to-south advection of  
 435 smaller  $\theta$  and  $q_v$ , and 2) south-to-north diffusion of both  $\theta$  and  $q_v$  driven by their gradients  
 436 (i.e., larger values in the south). The net effects observed in Figure 9 result from these opposing

437 processes. In Section 3.2, we show that the one of the upward mass flux bands with cloud  
 438 formation is at the northern edge of GS, suggesting cloud processes interfere with the  
 439 interactions of  $\theta$  and  $q_v$  between regions north and south of the GS. We hypothesize that  
 440 higher  $q_v$  lifted from surface to above the boundary layer in the southern region by convection  
 441 is consumed via stronger phase changes with the intensified SST gradient, weakening the  
 442 south-to-north diffusion of  $q_v$ . Consequently, the north-to-south advection of drier air  
 443 dominates (Figures 9a–9c). We hypothesize that stronger phase changes associated with the  
 444 intensified SST gradient from water vapor to hydrometeors release more latent heat, warming  
 445 the air and intensifying south-to-north heat diffusion (Figures 9d–9f).



446

447 **Figure 11.** Same as Figure 9 but for (a-c)  $q_l$ , (d-f)  $q_{ice}$ , (g-i)  $q_h$ , and (j-l)  $R_{q_l}$ .

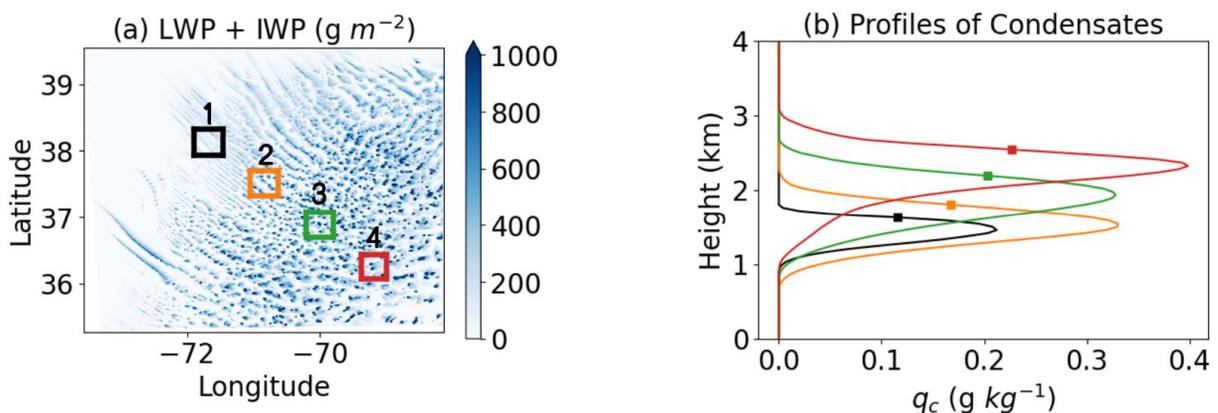
448

449 We examine the hypothesis of cloud interference described above in Figure 11, which  
 450 shows that a stronger low-level convergence (below  $\sim 2$  km) and larger vertical velocities before  
 451 08:00 EST in Gradplus (Figures 10f and 10i) closely associate with larger ice-phase  
 452 hydrometeor mixing ratios during the same period. An increase of liquid-phase hydrometeors  
 453 in Gradplus dominates within the boundary layer of the middle and southern regions (Figures  
 454 11k and 11l). These results suggest that, in the Gradplus experiment (vs. Ctrl), enhanced latent  
 455 heating from the increased ice-phase hydrometeors is key to modulating  $\theta$  and  $q_v$  interactions  
 456 across GS regions above the boundary layer. This implies that microphysical processes—  
 457 specifically, transitions from warm clouds to mixed-phase/ice-phase clouds—critically  
 458 influence  $\theta$  and  $q_v$  values above the boundary layer.

459

### 460 3.4 Response of the Sources of Airmass to SST Variations

461 Sections 3.1 – 3.3 demonstrate that the Plus4 and Gradplus experiments introduce  
 462 variations in local mesoscale circulation around the GS region. Beyond influencing thermal and  
 463 moisture fields, we hypothesize that SST variations also modify airmass sources, which  
 464 determine the type of aerosols (continental vs. marine) transported over the WNAO. To test this  
 465 hypothesis, we use Lagrangian tracers to track backward trajectories of airmasses ending up  
 466 within clouds. This approach aims to determine whether changes occur in the origin of  
 467 airmasses that ultimately affect clouds.



468

469 **Figure 12.** (a) Total (liquid and ice) water path at 0900 EST on March 1, 2020. Black, orange,  
 470 green, and red boxes are the four locations where tracers are released. (b) Averaged profiles of

471 cloud condensate mixing ratio ( $q_c$ ) within the four boxes in panel (a). Squares denote the  
472 heights where tracers are released in Figure 13.

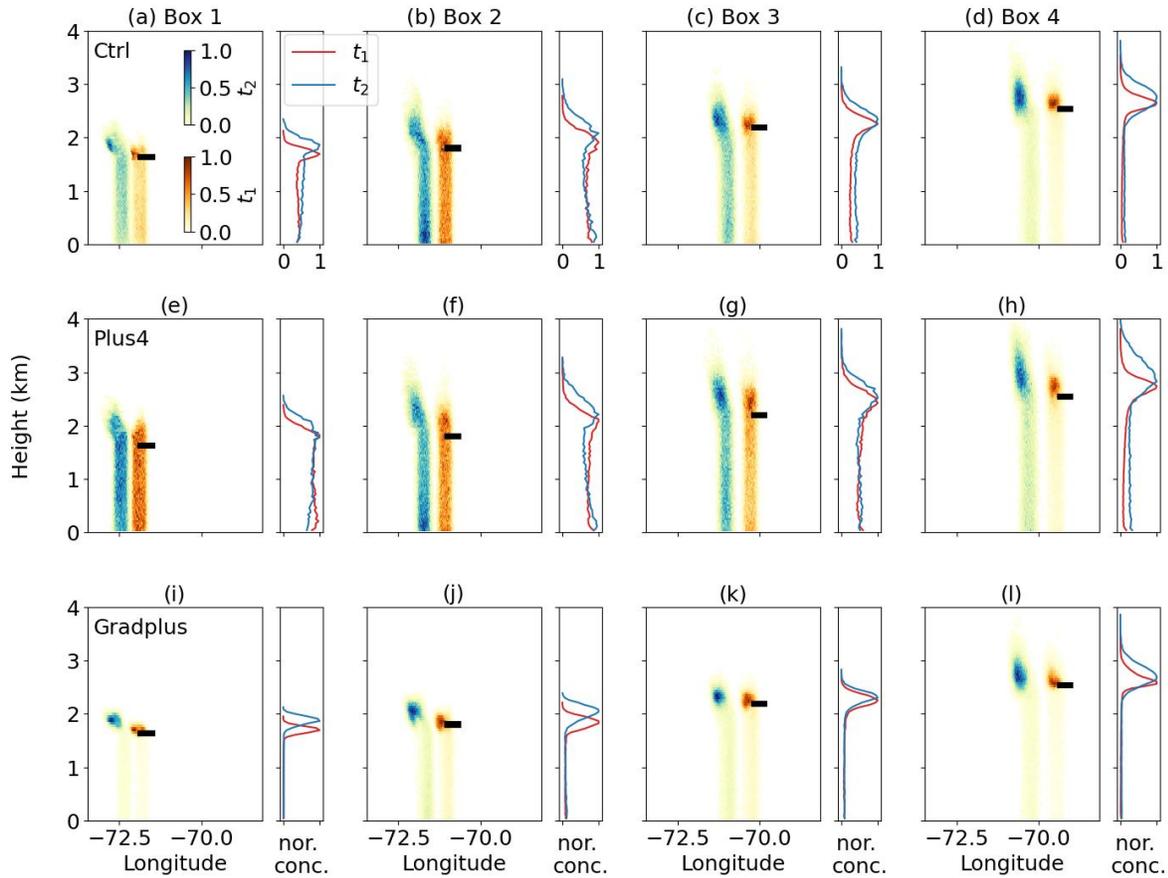
473 We release tracers at four locations (Figure 12a) and seven altitudes. The locations align  
474 with wind directions corresponding to cloud lines. The selected altitudes reflect their vertical  
475 position relative to cloud layers: (1) ocean surface, (2) 500 m above sea level and within the  
476 boundary layer, (3) cloud base, (4) midpoint between cloud base and the height of maximum  
477  $q_c$ , (5) height of maximum  $q_c$ , (6) midpoint between the height of maximum  $q_c$  and cloud top,  
478 and (7) cloud top. We calculate tracer concentrations for all 28 configurations of tracer release  
479 in the Ctrl, Plus4, and Gradplus experiments. Results show that, for six out of seven altitudes,  
480 tracer concentrations are similar at 30 minutes and 2 hours prior to release time between  
481 sensitivity and control simulations (not shown). The exception occurs for the altitude between  
482 the height of maximum  $q_c$  and cloud top (Squares in Figure 12b).

483 In the Ctrl experiment, the primary air source originates from west of the tracer release  
484 location, as tracer positions at 2 hours prior to release (cold color in Figure 13) are situated  
485 farther west compared to their positions at 30 minutes before release (warm color in Figure 13).  
486 When tracers are released within Box 2, two distinct air mass sources are observed: one from  
487 the cloud top and the other from the ocean surface. From Box 2 to Box 4, the cloud top  
488 air mass becomes increasingly dominant, suggesting that secondary circulation, associated  
489 with cloud streets that enhance tracer transport from the ocean surface, weakens progressively  
490 from Box 2 to Box 4.

491 In the Plus4 experiment, the air mass fraction originating from the ocean surface  
492 increases, likely due to the higher SSTs that drive a stronger turbulent transport within the  
493 boundary layer. In contrast, the contribution of air mass from the cloud top across all four  
494 boxes is enhanced in the Gradplus experiment, likely because stronger northwesterly winds  
495 above the boundary layer accelerate transport (Figures 10a-10c).

496 Aerosols linked to clouds in the examined regions are influenced by a mix of continental  
497 and marine sources. Our results regarding the air parcel origins implicitly suggest that SST  
498 variations associated with the GS trigger distinct shifts in aerosol composition. Specifically,  
499 marine aerosols influence Boxes 2-4 in Ctrl experiments and all four boxes in Plus4  
500 experiments via air parcels originated from sea surface, while continental aerosols influence the

501 other boxes through air parcels originated from continents. This subsequently alters aerosol-  
 502 cloud interactions over the WNAO region.



503  
 504 **Figure 13.** Profiles of normalized tracer concentration by the total number of each trajectory  
 505 run averaged at each longitude (left side of each panel a-l) and over the entire domain (right  
 506 side of each panel a-l) for tracers released midway between the maximum  $q_c$  and cloud top.  
 507 The horizontal black bar denotes the tracer release location. In each row, panels from left to  
 508 right correspond to release locations in Boxes 1–4 (see Figure 12a). Rows from top to bottom  
 509 represent results from the Ctrl, Plus4, and Gradplus experiments. The warm color (red)  
 510 represents 30 min prior to the release time ( $t_1$ ) and the cold color (blue) represents 2 hrs prior to  
 511 the release time ( $t_2$ ).

512

#### 513 **4 Conclusions**

514 This study investigates the impacts of altered Gulf Stream (GS) mean sea surface  
 515 temperature (SST) and gradients on post-frontal cloud (PFC) characteristics during cold-air

516 outbreaks (CAOs) over the western North Atlantic Ocean (WNAO). Using high-resolution WRF  
517 simulations of an ACTIVATE winter CAO event on 1 March 2020, combined with Lagrangian  
518 tracer analysis, we conduct two targeted sensitivity experiments: (1) a uniform +4K SST  
519 increase (Plus4; representing IPCC's A1F1 extreme warming scenario) to isolate mean warming  
520 effects, and (2) a 25% SST gradient enhancement (Gradplus; representing the strengthening of  
521 GS) with unchanged mean SST to examine gradient impacts. Three key findings emerge:

522 (1) Mean SST warming (+4 K) leads to a warmer, moister boundary layer, promoting  
523 larger cloud sizes. Enhanced ice-phase hydrometeors and elongated cloud streets are  
524 observed, with diminished liquid water path (LWP) in PFCs. Warmer SSTs amplify BL  
525 contributions of heat and moisture to cloud-top regions, altering air mass trajectories and  
526 favoring deeper, vertically extended clouds.

527 (2) Enhanced SST gradients (+25%) around the GS suppress cloud organization and  
528 increase clear-sky areal coverage. Stronger gradients intensify energy transport contrasts,  
529 leading to drier, colder BL conditions upwind of negative SST anomalies and enhanced liquid-  
530 phase hydrometeors in localized regions. Microphysical feedback, particularly ice-phase  
531 transitions, modulate thermal and moisture interactions across the GS, especially in the  
532 downwind positive SST anomaly region.

533 (3) Air mass source shifts under the SST perturbations are identified via Lagrangian  
534 tracers. Mean SST warming increases contributions of marine BL air to clouds, while enhanced  
535 GS gradients favor drier air entrainment from cloud-top regions. These shifts suggest potential  
536 alterations in aerosol-cloud interactions.

537 This study introduces two novel approaches: (1) isentropic analysis to isolate energy  
538 transport and (2) Lagrangian tracer tracking to quantify air mass sources. These methods reveal  
539 nonlinear PFC responses to SST variations. The findings highlight that the mean SST of the GS  
540 and SST gradients exert distinct controls on cloud macro- and microphysics, with implications  
541 for marine cloud feedback under global warming. Nevertheless, as this study examines a single  
542 case, extending these findings to long-term Gulf Stream variability requires further investigation.

543

544

545

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559

560 **Open Research**

561 ACTIVATE data are publicly available at  
562 <https://doi.org/10.5067/SUBORBITAL/ACTIVATE/DATA001>. The WRF community model was  
563 made available from the National Center for Atmospheric Research (NCAR) at  
564 <http://www2.mmm.ucar.edu/wrf/users/>. The post-processing code of FLEXPART-WRF was  
565 retrieved at <https://git.nilu.no/flexpart/flexpart-wrf/>. WRF model output (4 Tb) in this study is  
566 saved on PNNL's long-term storage system, Aurora ([rc-support@pnnl.gov](mailto:rc-support@pnnl.gov)). A subset of  
567 analyzed data and post-processing python code are also available from Zenodo (Chen and  
568 Wang, 2025).

569

570

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