The Simulated Impact of the Snow Albedo Feedback on the Large-Scale Mountain–Plain Circulation East of the Colorado Rocky Mountains

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ABSTRACT

The Front Range mountain–plain circulation (FRMC) is a large-scale diurnally driven wind system that occurs east of the Colorado Rocky Mountains in the United States and affects the weather both in the Rocky Mountains and Great Plains. As the climate warms, the snow albedo feedback will amplify the warming response in the Rocky Mountains during the spring, increasing the thermal contrast that drives the FRMC. In this study, the authors perform a 7-yr pseudo–global warming (PGW) regional climate change experiment along with an idealized PGW “fixed albedo” experiment to test the sensitivity of the FRMC to the snow albedo feedback (SAF). The authors find a mean increase in the springtime FRMC strength in the PGW experiment that is primarily driven by the snow albedo feedback. Furthermore, interannual variability of changes in FRMC strength is strongly influenced by interannual variability in the SAF. An additional case study experiment configured with a much higher resolution is performed to examine the finescale details of how the SAF and the FRMC interact. This experiment includes a passive tracer to investigate subsequent impacts on pollution transport. The case study reveals that loss of snow cover causes an increase in the strength of the FRMC. Advection by the strengthened FRMC increases the concentration of tracers emitted over the Great Plains in the boundary layer over the Front Range mountains.

1. Introduction

Mountain–plain circulations are broad terrain-driven diurnal wind systems that form in response to surface heating and cooling of large mountain slopes relative to the free troposphere over adjacent lowland plains. While the dynamics that drive these circulations are analogous to those that drive smaller-scale valley-and-slope flows, these circulations are generally considered distinct from these flows as they are much larger, spanning several hundreds of kilometers, and their impacts extend well beyond the immediate mountain region (e.g., Weissmann et al. 2005; Zardi and Whiteman 2013). Globally, these circulations play important roles in transporting moisture and pollutants into mountain regions (e.g., Steyn et al. 2013; Sullivan et al. 2016) and in initiating convection and modulating propagating mesoscale convective systems (e.g., Tucker and Crook 1999; Carbone and Tuttle 2008). Previous studies indicate that these circulations are strongly sensitive to characteristics of the land surface that affect diurnal heating and stability of the lower atmosphere (e.g., Ookouchi et al. 1984; Segal et al. 1991; Bossert and Cotton 1994b; Wolyn and Mckee 1994; de Wekker et al. 1998; Chase et al. 1999; Letcher and Minder 2016).

The Front Range mountains of Colorado mark the easternmost mountain barrier separating the Rocky Mountains to the west from the gradually sloping plains to the east (Fig. 1). During the warm season, under weak synoptic conditions (e.g., light winds aloft and clear skies), a large mountain–plain circulation develops during the day in response to the thermal forcing. This circulation, referred to as the Front Range mountain–plain circulation (FRMC), is typically observed by surface stations east of the Rocky Mountains as a twice-daily wind reversal that extends over 300 km east of the mountains (Toth and Johnson 1985). Numerical simulations of the FRMC performed by Bossert and Cotton (1994a,b) showed that the upslope...
easterly winds span the depth of the boundary layer and the return westerly flow aloft extends to 4.5–5 km ASL. These simulations reproduced the observed wind patterns near the surface; however, there were no observations available at the time to evaluate the vertical structure of the simulated FRMC. However, recent lidar observations of the FRMC presented in Sullivan et al. (2016) show a vertical structure generally consistent with the numerical simulations from Bossert and Cotton (1994a,b).

This circulation episodically advects pollution sourced from the urban and agricultural corridors east of Colorado’s Front Range into the mountains. Of particular interest is the pollution originating from agricultural activity (most prominently, cattle feedlots), which has been found to contribute significantly to the production of volatile organic compound gases and aerosols, especially nitrogen species resulting from fertilization (e.g., Clements et al. 2016). This feedlot pollution is often advected well into the Front Range mountains, where it is known to affect water quality and soil chemistry, including algal growth in streams and rivers (Wolfe et al. 2001; Piña 2013).

Motivated by these results, we investigate how the FRMC is affected by snow loss and the snow albedo feedback (SAF) as the global climate warms and how these changes in the FRMC can impact pollution transport from the Great Plains into the mountains. The SAF is a positive climate feedback mechanism whereby a warming climate results in reduced snow cover and, subsequently, a reduced surface albedo, thereby increasing the absorbed solar radiation and, as a result, warming at the surface. It is expected that as the climate warms, the SAF will amplify warming over mountain regions with transient snow cover (e.g., Giorgi et al. 1997; Salathé et al. 2008; Gao et al. 2011; Kotlarski et al. 2012; Letcher and Minder 2015). In the Rocky Mountains, large snowpacks accumulate during the winter and often persist into early summer. During the spring ablation season [March–June (MAMJ)], the SAF is particularly strong because of the elevated sensitivity of snow cover to warming and high solar insolation (e.g., Letcher and Minder 2015). While numerous studies have focused on the impacts of climate change on snow in this region (e.g., Barnett et al. 2008; Pierce et al. 2008; Rauscher et al. 2008; Viviroli et al. 2011; Rasmussen et al. 2014), there are currently no studies that investigate how SAF-enhanced warming over the Rocky Mountains impacts atmospheric conditions over adjacent regions (e.g., the Great Plains).

It is expected that under future warming, the SAF will preferentially increase warming over the elevated terrain of the Rockies where snow cover is reduced (e.g., Fyfe and Flato 1999; Letcher and Minder 2016), thereby increasing the thermal contrast that drives the FRMC. We hypothesize that FRMC will become stronger during the middle to late spring in response to this increased thermal contrast. This hypothesis is motivated by the Bjerknes circulation theorem, which, neglecting planetary rotation and viscous forces, relates changes in a circulation $C$ to baroclinicity along a closed circuit within a fluid:

$$\frac{DC}{Dt} = -RT \ln p,$$  \hspace{1cm} (1)

where $T$ and $p$ are temperature and pressure and $C$ is defined as the tangential wind velocity $V$ integrated around a closed circuit:

$$C = \oint V \, dl.$$  \hspace{1cm} (2)

A simplified two-dimensional version of this framework is often applied to describe the generation of a
sea-breeze circulation over a vertically oriented rectangular-shaped circuit with isobaric upper and lower branches (e.g., Holton and Hakim 2012), yielding

$$\frac{D\langle v \rangle}{Dt} = \frac{R \ln(p_0/p_1)}{2(h + L)} (\overline{T}_1 - \overline{T}_2),$$

(3)

where $\langle v \rangle$ is the mean speed of the circulation, $h$ and $L$ are the vertical and horizontal dimensions of the circuit, $p_0$ and $p_1$ correspond to the pressure of the lower and upper legs of the circuit, respectively, and $\overline{T}_1$ and $\overline{T}_2$ are the vertically averaged temperatures on the two vertical branches of the circuit. Using this simplified framework, a rough estimate of the sensitivity of a mountain–plain circulation to changes in the thermal contrast between the mountains and the plains ($T_1 - T_2$) can be obtained. For example, assuming that a mountain–plain circulation follows a box circuit 200 km in length and 3 km deep and that the lower and upper legs of this circuit follow isobaric paths at 800 and 600 hPa, the mean velocity of this circulation is increased by approximately 2 m s$^{-1}$ after 3 h for a 1-K increase in the thermal contrast across the circuit. This simple calculation suggests that the mountain–plain circulation is highly sensitive to small changes in the thermal contrast. In reality, the wind acceleration around the circuit will be counteracted by frictional forces (Holton and Hakim 2012); thus, this framework likely predicts too strong a response. In addition to friction, this simple framework leaves out processes that may also influence the FRMC (e.g., Coriolis force; cross-isobar flow on horizontal branches). Numerical models can account for all of the processes neglected in the simple framework and can therefore be used to study how a mountain–plain circulation will respond to changes in the driving thermal contrast.

Ookouchi et al. (1984) tested the sensitivity of mesoscale mountain-breeze circulations to variations in soil moisture using simple 2D models over highly idealized terrain. They showed that spatial variations in soil moisture along a mountain slope had a substantial influence on the strength of the mountain-breeze circulation by modulating the driving thermal contrast. Building upon this, Bossert and Cotton (1994b) ran similar idealized simulations over realistic 3D terrain and showed that the mountain–plain wind systems on both the eastern Front Range and western slope of the Colorado Rockies were strongly sensitive to surface temperature anomalies caused by changes in soil moisture concentrations. Chase et al. (1999) was the first study to link possible changes in the FRMC to regional climate changes. In their study, they argued that regional cooling trends over the western Great Plains linked to increased irrigation and agricultural activity decreased the strength of the FRMC by reducing the surface sensible heating on the Great Plains. More recently, Letcher and Minder (2016) used multiyear high-resolution regional climate modeling experiments to investigate the relationship between the SAF and local diurnal wind systems in the Rocky Mountains. They showed that enhanced warming caused by the SAF increased the thermal contrast between the elevated mountains and surrounding valleys and plains, thereby increasing the strength of the local mountain-breeze circulations. Here, we expand on the analysis of Letcher and Minder (2016) to explore the sensitivity of the larger-scale FRMC and pollution transport to changes in Rocky Mountain snow cover using more-controlled climate experiments.

In this paper, we use medium-resolution multiyear regional climate simulations to assess the role of the SAF and snow cover in modulating the FRMC. Furthermore, we use a case study experiment to investigate the detailed structure of FRMC changes associated with the SAF and how these changes potentially affect pollution transport from the Great Plains to the Rocky Mountains. For simplicity, we focus primarily on the dry circulation response to the SAF, although there are likely impacts of FRMC changes on moist convection as well.

In section 2, we present the data and methods. In section 3, we show the results from the multiyear regional climate experiment. In section 4, we show results from the high-resolution case studies. Concluding remarks are presented in section 5.

2. Data and methods

a. Multiyear regional climate simulations

We perform two sets of regional climate model (RCM) experiments using version 3.7 of the Weather Research and Forecasting (WRF) Model (ARW core; Skamarock et al. 2008). One experiment mimics the RCM experiments performed over the headwaters domain by Rasmussen et al. (2014). In our experiment, WRF simulations are performed over the domain marked by the terrain shading in Fig. 1a. WRF is initialized and forced on the lateral boundaries using the North American Regional Reanalysis (NARR) dataset (Mesinger et al. 2006). The model is interactively coupled to the land surface using version 3.6 of the standard Noah land surface model (LSM). Additional model details are presented in Table 1. We chose to use 12-km horizontal grid spacing instead of a finer resolution to minimize computational expense. While this resolution is not convection permitting, it is sufficient
to resolve the FRMC, as well as the bulk characteristics of the SAF over the Rocky Mountains (Letcher and Minder 2015). Some potential sources of model uncertainty include the LSM, as well as the surface and boundary layer parameterizations. We chose to use the Noah LSM in lieu of an LSM with a more sophisticated snow model to simplify the sensitivity experiments employed in this study. Several upgrades were made to the Noah, version 3.6, snow model to address several of its well-documented deficiencies (Livneh et al. 2010; Barlage et al. 2010). Furthermore, a brief sensitivity experiment comparing the simulated diurnal wind response of mountain-breeze circulations to the SAF between Noah and the more sophisticated Noah LSM with multiparameterization options (Noah-MP) showed remarkably similar results, suggesting that the use of Noah is appropriate for this study (Letcher 2017). The extent to which the simulated FRMC is sensitive to surface and boundary layer parameterizations was not investigated as part of this study and may represent a source of uncertainty in our results.

To examine the sensitivity of the FRMC to snow-cover changes associated with future anthropogenic climate change, we make use of a pseudo–global warming (PGW) experiment (e.g., Schär et al. 1996; Rasmussen et al. 2011). In this experiment, a mean climate perturbation derived from an ensemble of global climate model simulations is applied to the lateral boundary conditions of the regional model. This downscaling technique ensures a high signal-to-noise ratio of the climate response relative to natural internal variability as it largely preserves the synoptic variability from the control simulation. We use the same PGW perturbation applied to the set of RCM simulations described in Liu et al. (2016): the RCP8.5 CMIP5 ensemble-mean difference between a future state, the 30-yr mean over 2070–2100, and a present state, the 30-yr mean over 1975–2005. The climate perturbations applied are spatially variable, monthly mean changes of temperature, relative humidity, horizontal winds, and geopotential height. These perturbations were interpolated in time to match the reanalysis forcing and added to the initial and lateral boundary conditions. Additionally, a radiative forcing is added that is representative of a CO2 concentration of 800 ppm, the global-mean prescribed CO2 concentration at 2085 from the RCP8.5 emissions scenario.

To isolate the role of the SAF in modifying the FRMC, we perform a third simulation in which the PGW simulation described above is repeated with one important difference: the surface albedo is held fixed to that of the control simulation. We refer to this as the “fixed albedo” (FA) experiment. In this simulation, snow cover is allowed to evolve freely in response to the applied climate perturbation; however, the seasonal and interannual variations of surface albedo from the control simulation are preserved. This configuration is similar to the configuration described by Hall (2004). By comparing the FA and control experiment, the climate change response without an SAF is obtained. Then by comparing the PGW and FA experiments, all direct impacts of the climate forcing and feedbacks unrelated to surface albedo are removed, including non-albedo-related snow feedbacks, thereby isolating the effects of the SAF.

We limit our analysis to the months of April and May to focus on the time when a strong SAF and a pronounced FRMC are likely to overlap, following the methodology of Letcher and Minder (2016). Furthermore, we disregard the first year of model output to allow for LSM spinup, resulting in a 7-yr spring climatology.

b. High-resolution case study simulations

To further investigate the impacts of the SAF on the strength of the FRMC and the associated impacts on pollution transport, a single-day high-resolution case study experiment was performed. This model simulation was performed from 1200 UTC 19 May to 0600 UTC 21 May 2005. The first 18 h of the model simulation were disregarded to allow for model spinup, leaving the 24-h period beginning at 0600 UTC 20 May

<table>
<thead>
<tr>
<th>Model setup</th>
<th>RCM</th>
<th>Case study</th>
</tr>
</thead>
<tbody>
<tr>
<td>Time</td>
<td>0000 UTC 1 Oct 2000–0000 UTC 1 Oct 2008</td>
<td>0600 UTC 20 May–0600 UTC 21 May 2005</td>
</tr>
<tr>
<td>Δx</td>
<td>12 km</td>
<td>12 km (d01 only)</td>
</tr>
<tr>
<td>Microphysics</td>
<td>Thompson (Thompson et al. 2004)</td>
<td>Thompson</td>
</tr>
<tr>
<td>Lateral boundaries</td>
<td>NARR (Mesinger et al. 2006)</td>
<td>NARR</td>
</tr>
<tr>
<td>Radiation</td>
<td>CAM3 (Collins et al. 2004)</td>
<td>RRTM for GCMs (RRTMG; Iacono et al. 2008)</td>
</tr>
<tr>
<td>Boundary layer</td>
<td>Yonsei University (YSU; Hong et al. 2006)</td>
<td>YSU</td>
</tr>
<tr>
<td>Cumulus</td>
<td>Betts–Miller–Janjic (BMJ; Betts 1986; Betts and Miller 1986)</td>
<td>Kain–Fritsch (KF; d01 only; Kain and Fritsch 1993)</td>
</tr>
<tr>
<td>LSM</td>
<td>Noah (Chen and Dudhia 2001; Ek et al. 2003)</td>
<td>Noah-MP (Niu et al. 2011; Yang et al. 2011)</td>
</tr>
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</table>
for analysis. This time period is characterized by having both weak synoptic conditions and anomalously high snow cover over the Rockies; thus, we expect a robust FRMC and a potentially strong response to reductions in snow cover.

WRF was configured using a two-way triple-nested telescoping grid configuration (Fig. 2a). The parent domain was run with 12-km horizontal grid spacing and forced using data from the NARR. Additional details about the parent domain are presented in Table 1. The two inner domains were convection permitting, with horizontal grid spacings of 4 and 1.3 km. To obtain a high-resolution realistic representation of snow cover without requiring extensive spinup time, snow variables are initialized in the inner two domains using output from the 4-km WRF continental United States (CONUS) simulation described in Liu et al. (2016). The CONUS simulation is a multiyear convection-permitting regional climate simulation coupled to the Noah-MP (Niu et al. 2011; Yang et al. 2011). For consistency with the snow initialization, our case study simulation was coupled to the Noah-MP rather than the standard Noah LSM as in our regional climate experiments. To assess the impact of snow loss on the FRMC, this simulation was repeated but with the snow initialized using output from the reduced-snow-cover CONUS PGW simulation. This simulation is hereafter referred to as the PGWSNO simulation.

Simulated pixel-average snow-cover fraction \( f_{sn} \) is evaluated against the Moderate Resolution Imaging Spectroradiometer (MODIS) satellite-derived snow-cover area and grain size (MODSCAG; Painter et al. 2009) pixel-average snow-cover product (Figs. 3a,b). In general, \( f_{sn} \) from the control simulation matches \( f_{sn} \) from MODSCAG. This is especially true along the Front Range mountains. Because of the generally good agreement, we feel these simulations are appropriate for examining the impact of snow loss. The PGWSNO snow cover (Fig. 3c) is substantially reduced over the region, restricted to the highest elevations, and almost entirely eliminated on the Front Range, resulting in large changes in the surface energy budget over the high terrain that may influence the FRMC.

To explore how changes in the FRMC may impact regional pollution transport, we add a passive tracer source to the case study simulations. Here, a massless passive tracer representing agricultural-based pollution is emitted at a constant rate in the innermost domain into the first model level from the region shown by the rectangular area on the Great Plains in Fig. 2b. This tracer is transported by both advection and sub-grid mixing and has no sinks (e.g., wet scavenging). This rectangle roughly outlines a region with a high concentration of farms, including numerous large cattle feedlots. Tracer emission begins after model initialization, 3 h prior to analysis time (0300 UTC May 20) to avoid changes in tracer transport during the spinup period.

c. Quantifying the relationship between the FRMC and the thermal contrast

The circulation of the FRMC is quantified by integrating the wind velocity around the closed, meridionally averaged \( x-z \) circuit marked by the blue volume in Fig. 1b. To avoid interpolation, the circuit path follows the native WRF terrain-following coordinate. The circulation is normalized by the length of the circuit providing a mean wind speed for the FRMC \( V_{FRMC} \).

The thermal contrast between the Rocky Mountains and the atmosphere over the Great Plains \( \theta_{can} \) is used as an index for the thermal forcing that drives the FRMC:
where the angle brackets indicate the spatial mean, the subscripts MTN and PLN correspond, respectively, to the regions marked by boxes 1 and 2 indicated in Fig. 1a, \( \theta_{\text{MTN}} \) is the surface potential temperature, and \( \theta_{\text{PLN}} \) is the mean potential temperature below \( \approx 2700 \) m ASL, which is the approximate mean terrain elevation of box 1.

To assess the impacts of the SAF on the strength of the FRMC, changes in FRMC strength \( \Delta V_{\text{FRMC}} \) are compared to changes in the thermal contrast \( \Delta \theta_{\text{con}} \). Both \( V_{\text{FRMC}} \) and \( \theta_{\text{con}} \) are 6-h averages centered on 1500 LST to focus this analysis on the daytime response when the FRMC is at peak strength. To more directly attribute these changes to the SAF, changes in fractional snow cover \( f_{\text{sn}} \) averaged over region 1 are also compared to \( \Delta \theta_{\text{con}} \).

3. Regional climate simulation results

a. Observational comparison

A brief evaluation of the model-simulated diurnal cycle against a collection of surface stations is presented in Fig. 4. This analysis compares the 7-yr April–May mean diurnal cycle of hourly 2-m temperature and near-surface wind data from the control simulation to regional surface observations from the remote automated weather stations (RAWS) and Automated Surface Observing System (ASOS) networks. There is a clear diurnal cycle in temperature with a mean amplitude of approximately 15 K. At nearly all stations, the winds follow a daily transition from downslope flow overnight to upslope flow during the day. For example, the winds at Grand Junction (KGJT) on the western slope transition from easterly overnight to westerly during the day, and the winds at Pueblo (KPUB), east of the Front Range, transition from northwesterly at night to southeasterly during the day.

In general, the model is able to represent the diurnal cycle of temperature and wind at all sites; however, the maximum temperatures are slightly underestimated at the higher-elevation sites [e.g., Meeker (KEEO) and Cortez (KCEZ)]. This underestimate is likely associated with deficiencies in the Noah LSM’s treatment of snow (e.g., Chen et al. 2014; Letcher and Minder 2016). Nevertheless, the model is generally able to capture the diurnal cycle of near-surface wind, suggesting that the well-documented deficiencies in the Noah LSM do not have a large negative impact on the dynamic processes at the spatial and temporal scales most relevant to this study. However, the downslope-to-upslope transition in the winds is clearer and better represented over the western sites than over the Great Plains.

b. Structure of FRMC changes

Figure 5 summarizes mean changes variables related to the SAF and FRMC. Here, April–May mean PGW – control (ctrl), FA – ctrl, and PGW – FA changes in albedo \( \Delta \alpha \), temperature \( \Delta T_s \), and 10-m winds \( \Delta V_{10} \) averaged between 1500 and 2100 LST are compared.

The similar warming patterns and the spatial coupling between \( \Delta T_s \) and \( \Delta \alpha \) found in the PGW – ctrl and
PGW – FA differences suggest that almost all of the spatial variability in warming is associated with the SAF (Figs. 5b,h). This is supported by the absence of spatial structure in warming when the SAF is suppressed in the FA – ctrl differences (Fig. 5e).

In response to the warming, there is a clear increase in easterly winds over the Great Plains in the PGW experiment (Fig. 5c). This increase attains a strength of nearly 0.5 m s\(^{-1}\) near the foot of the Front Range and is caused by the SAF, as inferred by the similar patterns of \(\Delta V_{10}\) seen in the PGW – FA differences (Fig. 5i). In fact, the wind changes in the absence of the SAF are nearly opposite in direction to the changes in the PGW – ctrl difference (Fig. 5f), suggesting that the thermal forcing caused by the SAF is strong enough to overcome large-scale geopotential height changes in the PGW boundary forcing that would otherwise lead to enhanced southwesterly flow over the region. On the western slope of the Front Range mountains, there is a similar, stronger upslope wind response leading to an increase in horizontal convergence near the summit of the Front Range mountains.

To illustrate the diurnal cycle of this response, Hovmöller diagrams showing the diurnal evolution of mean zonal wind \(u\) are presented in Fig. 6. In the control simulation, upslope flow first develops over a narrow region along the eastern slopes of the Front Range at local noon (Fig. 6a). This upslope flow extends east over the Great Plains as the afternoon progresses, maximizing around 1900 LST before weakening and becoming downslope overnight. The PGW – ctrl difference shows that upslope flow is enhanced first along the eastern slopes of the Front Range starting around 1000 LST (Fig. 6b). The upslope flow anomalies strengthen and expand east as the SAF-enhanced warming nears its daytime maximum shortly after peak insolation. By late afternoon, the anomalous upslope flow covers most of the western Great Plains. The upslope flow anomalies persist well after the sun goes down before weakening overnight as the SAF-enhanced warming decreases. These changes can be largely attributed to the SAF, as indicated by the similar patterns of \(\Delta u\) in the PGW – FA difference (Fig. 6d) as well as
the weaker and less organized FA – ctrl difference (Fig. 6c).

To gain insight into the three-dimensional structure of the FRMC and changes that occur in response to the SAF, vertical cross sections of 6-h-average $T$, $u$, and $w$ centered at 1500 LST, meridionally averaged over the gray dashed volume in Fig. 1b, are analyzed (Fig. 7). In the control simulation, the isotherms are sloped downward from west
to east crossing the isobars, indicating a thermal contrast between the mountains and the plains favorable for the FRMC (Fig. 7a). The daytime diurnal zonal wind anomalies over the Great Plains are easterly within the boundary layer and westerly aloft, indicative of the FRMC (Fig. 7b). Along the eastern slope and near the peak elevations of the Front Range mountains, there is a narrow and deep strip of strong vertical motion marking the western boundary of the FRMC (Fig. 7c). The easterly anomalies over the Great Plains extend to higher altitudes than might be expected from thermal forcing alone. We speculate that this is due to daytime-enhanced boundary layer drag counteracting the synoptic flow above the boundary layer.

The role the SAF plays in modulating the thermal contrast between the mountains and the plains is apparent by comparing $\Delta T$ for the PGW – ctrl, FA – ctrl, and PGW – FA differences (Figs. 7d,g,j). The SAF-enhanced warming over the mountains is within the range of 1–2 K and is mostly confined to the lowest 1 km AGL.

In comparing the PGW – ctrl change in $u$ (Fig. 7e), there is a large area of anomalous low-level easterlies below mountaintop height, with anomalously westerly flow aloft over the Great Plains, consistent with an anomalously strong FRMC. The anomalous easterly winds attain a magnitude of nearly 1 m s$^{-1}$ and extend more than 200 km east of the Front Range. A similar, but opposite, pattern in $\Delta u$ present on the western slope of the Rocky Mountains suggests that the western slope mountain–plain circulation is affected in a similar way. The anomalously strong lower branches of the eastern and western mountain–plain circulations converge near the highest terrain, where there is an increase in mean vertical motion of about 2–3 cm s$^{-1}$ (Fig. 7f). In comparing the PGW – ctrl changes in $u$ and $w$ to the FA – ctrl (Figs. 7h,i)
and the PGW – FA (Figs. 7k,l) changes, it is clear that the changes in the zonal and vertical winds below 4 km ASL seen in the PGW – ctrl difference are caused almost entirely by the SAF, since they are absent in the FA – ctrl differences (Figs. 7h,i). These results suggest that the regionally confined SAF has much broader impacts that extend far beyond the region of enhanced warming.

To quantitatively investigate the bulk relationship between SAF-enhanced warming and FRMC changes, the afternoon mean tangential wind velocity integrated around the circuit shown in Fig. 7 is calculated and compared to changes in the afternoon thermal contrast between the mountains and plains $D_{u_{con}}$ and snow loss in the mountains $D_{f_{sn}}$. For this analysis, the data are grouped into individual monthly mean values for April and May each year, yielding a sample size of 14 for statistical analysis. These results are presented in Fig. 8 and Table 2.

In the control simulation, there is a strong and statistically significant correlation ($p < 0.05$) between the thermal contrast and mountain snow-cover anomalies, indicating that during the spring, seasonal, and interannual variability in the thermal contrast is strongly connected to the amount of snow cover in the Rockies (Table 2). However, the relationship between variability in the thermal contrast and variability in the FRMC is not significant. While this suggests that variability in FRMC strength is unrelated to variability in the thermal contrast between the mountains and the plains, we do not believe this to be the case. Instead, we speculate that the lack of a significant relationship is because the metric we use to quantify the FRMC is not capturing interannual variability in the thermally driven FRMC; rather, it is more likely capturing interannual variability in the synoptic westerly flow near the top of the closed circuit. This is because, even under weak synoptic forcing, the wind velocity near the top of the circuit is the same order of magnitude as the thermally driven easterly flow near the surface, and therefore, variability in synoptic forcing can dominate variability in the FRMC metric when it is applied to the control simulation. This metric performs better when comparing the PGW and ctrl simulations as synoptic variability is held constant.

In both the PGW – ctrl and PGW – FA experiments, interannual variability in $D_{f_{sn}}$ is strongly and significantly correlated with interannual variability $D_{con}$ (Table 2; Figs. 8a,c). Furthermore, there is a significant relationship between $D_{con}$ and $\Delta V_{FRMC}$ (Table 2; Figs. 8b,d). This supports the hypothesis that seasonal and interannual variability of changes in the strength of the FRMC are largely controlled by variability in the

![Fig. 7. Vertical cross sections of 7-yr April and May mean (a),(d),(g),(j) $T$, (b),(e),(h),(k) $u$, and (c),(f),(i),(l) $w$. Cross sections are meridionally averaged over the gray volume shown in Fig. 1. Isobars are contoured in (a). For the ctrl simulation, $u$ and $w$ are anomalies relative to their diurnal averages. All values are 6-h means centered over 1500 LST, representing the daytime branch of the FRMC. The black circuit indicates the bounding circuit used for calculating $V_{FRMC}$ presented in Fig. 8. Black terrain outline indicates ctrl $f_{sn} > 0.2$ and green terrain outline indicates PGW $f_{sn} > 0.2$.](image-url)
thermal contrast caused by variability in SAF-enhanced warming. The correlation between $\Delta \theta_{\text{con}}$ and $\Delta V_{\text{FRMC}}$ is positive and of similar magnitude in both the PGW$^2$ ctrl and PGW$^2$ FA comparisons. This positive correlation is consistent with the hypothesis that an increase in the thermal contrast increases the strength of the FRMC. Since these relationships are not present in the FA$^2$ ctrl comparison, both changes in $\theta_{\text{con}}$ and $V_{\text{FRMC}}$ can be specifically attributed to the SAF (Table 2).

Overall, the analysis presented in this section provides evidence that the SAF substantially increases the strength of FRMC in a warmed climate and that the SAF influences changes in atmospheric dynamics far removed from the mountains where it originates.

FIG. 8. Scatterplots comparing changes in fractional snow cover $f_{\text{sn}}$, FRMC strength $V_{\text{FRMC}}$, and the thermal contrast between the mountains and the plains $\Delta \theta_{\text{con}}$ averaged between 1500 and 2100 LST. (a),(b) PGW$^2$ ctrl and (c),(d) PGW$^2$ FA. (a),(c) $\Delta \theta_{\text{con}}$ vs $\Delta f_{\text{sn}}$; (b),(d) $\Delta V_{\text{FRMC}}$ vs $\Delta \theta_{\text{con}}$. Each point is a single monthly mean value for a single year. The circles and stars represent April and May, respectively. The gray dashed lines show linear regression; related statistics are presented in Table 2.

### Table 2. Comparison of FRMC strength and $\Delta V_{\text{FRMC}}$ to the thermal contrast $\Delta \theta_{\text{con}}$ and fractional snow cover $f_{\text{sn}}$. Subscript 1 indicates linear regression between $V_{\text{FRMC}}$ and $\theta_{\text{con}}$. Subscript 2 indicates linear regression between $\theta_{\text{con}}$ and $f_{\text{sn}}$. Boldface text indicates statistical significance above the 95% level using a two-tailed $t$ test.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Mean $\theta_{\text{con}}$ (K)</th>
<th>Mean $f_{\text{sn}}$ (–)</th>
<th>Mean $V_{\text{FRMC}}$ (m s$^{-1}$)</th>
<th>Slope$^1$ (m s$^{-1}$ K$^{-1}$)</th>
<th>Slope$^2$ (K %$^{-1}$)</th>
<th>$R_1^2$</th>
<th>$R_2^2$</th>
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<tr>
<td>ctrl</td>
<td>1.61</td>
<td>0.35</td>
<td>9.99</td>
<td>0.18</td>
<td>$-0.10$</td>
<td>0.01</td>
<td>0.80</td>
</tr>
<tr>
<td>PGW$^2$ ctrl</td>
<td>1.13</td>
<td>$-0.21$</td>
<td>1.59</td>
<td>$0.59$</td>
<td>0.05</td>
<td>0.57</td>
<td>0.83</td>
</tr>
<tr>
<td>FA$^2$ ctrl</td>
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<td>$-0.16$</td>
<td>0.78</td>
<td>0.26</td>
<td>$-0.02$</td>
<td>0.12</td>
<td>0.12</td>
</tr>
<tr>
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<td>$-0.05$</td>
<td>0.81</td>
<td>$0.42$</td>
<td>0.17</td>
<td>0.80</td>
<td>0.92</td>
</tr>
</tbody>
</table>
complement the results from the 12-km RCM simulations described in this section, we use results from a much higher-resolution experiment to examine the detailed circulation response to the SAF and the potential impacts of these changes on pollution transport.

4. Case study results

Here, we examine results for the high-resolution case study simulation described in section 2b. A brief overview of the synoptic evolution of the event is presented in Fig. 9. This event is characterized as having a large upper-level ridge with widespread anticyclonic flow that persists throughout the event and serves to minimize cloud cover and maximize diurnal forcing. In the lower levels, a trough develops over the elevated terrain of the Rockies and strengthens throughout the day. At the surface, upslope flow initially develops over the Great Plains but then transitions to more southerly flow during the late afternoon and early evening as the west-to-east pressure gradient increases in response to a synoptic high that develops and strengthens over Kansas. The synoptic evolution of this case study is broadly characteristic of fair-weather days favorable to the development of the FRMC and generally similar to that of other FRMC case studies (e.g., Chase et al. 1999; Bossert and Cotton 1994b).

The diurnal evolution of the modeled and observed near-surface winds for this event is shown in Fig. 10. In this figure, the modeled 10-m winds are compared to all available hourly near-surface wind observations from the MesoWest dataset (Horel et al. 2002), regardless of station network. In the model, there is a transition from weak downslope flow in the predawn hours (0600 LST) to weak, though geographically broad, upslope flow by midday (1000–1400 LST). During the late afternoon (1600 LST), the winds east of the Front Range become largely southerly, with upslope confined to the eastern slope of the Front Range mountains. This shift likely reflects the mixing of synoptic flow to the surface in combination with a tightening low-level pressure gradient between the thermal trough over the Rockies and the ridge over the Great Plains. This diurnal evolution is broadly seen in the observations, especially near the Front Range mountains. However, there are many places where the model occasionally contradicts the observed winds. Notably, there is a lack of daytime upslope flow observed at the stations in northeast Colorado.

Figure 11 illustrates how the tracer concentration $T_t$ evolves in response to the development of the FRMC throughout the day. During the early morning (0700 LST),
the tracer is largely confined to the lowest 0.5 km of the atmosphere well east of the Front Range and east of the source region because of persistent katabatic flow overnight. As the boundary layer deepens throughout the day and the FRMC matures (1000–1300 LST), the tracer disperses vertically and expands west up the slope of the Front Range. In the late afternoon (1600 LST), as the southerly component of the wind becomes dominant and synoptic winds mix down to the surface, the FRMC weakens and the tracer plume is advected north and east away from the mountains.

To investigate how circulation changes caused by snow loss influence tracer transport, Fig. 11 was recreated for PGWSNO – ctrl anomalies (Fig. 12). Over the morning hours (0700–1300 LST), a general increase in up slope flow develops from the initially noisy field of $\Delta V_{10m}$ east of the Front Range mountains. During the afternoon (1300–1600 LST), the increased up slope flow persists and extends well away from the mountains. This increase in up slope flow is better visualized in the vertical cross sections, which show a confined region of increased easterly winds near the summits of the Front Range, which expands eastward as the day progresses and the warming intensifies. At 1300 and 1600 LST, the pattern of $\Delta u$ is broadly similar to the pattern seen in Figs. 5–7, with increased easterly flow within the boundary layer east of the Front Range and increased westerly flow to the west.

In response to the changes in the wind patterns, a clear dipole pattern of anomalous tracer concentration emerges, indicating a westward shift in the main tracer plume. This dipole is generally uniform in the vertical and spans the depth of the boundary layer. Interestingly, while there is a substantial increase in tracer concentration in the mountains, the westward edge of the tracer plume [as defined by $\log_{10}(T_r) = -3.5$; indicated by the thick lines in Fig. 12] in the PGWSNO simulation is only shifted approximately 5.5 km west from its maximum extent in the control run despite persistent increases in easterly up slope winds.
One reason for the relatively minor change in westward plume extent is that, in the westward plume, extent is limited by the terrain rather than the strength of the FRMC. Since the FRMC is already transporting the tracer plume to the crest of the Front Range Mountains, an increase in the strength of the FRMC does not substantially change its geographic extent. To help illustrate this, Hovmöller diagrams comparing tracer concentrations from the control simulation and changes in low-level zonal wind are presented in Fig. 13. The westward expansion of the tracer plume from its source region during the midday and afternoon hours as the FRMC develops is clearly identified in Fig. 13a. The western edge of the control tracer plume roughly marks the western edge of easterly flow associated with the FRMC as inferred from the wind vectors in Fig. 13a. The western boundary of the tracer plume in the PGWSNO simulation is shifted slightly farther west from the control for most of the day, consistent with a stronger FRMC. However, once the western boundary of the plume in the PGWSNO simulation reaches the crest of the Front Range mountains, it remains generally constant in time until it begins to shift back east, indicative of the constraints posed by the mountain crest and opposition from the mountain–plain circulation on the western side of the Front Range.

There is considerable similarity between the diurnal cycle of $\Delta u$ from the case study (Fig. 13b) and the diurnal cycle of mean $\Delta u$ in the PGW – FA experiment (Fig. 6c), especially in terms of the eastward progression of the easterly wind anomalies throughout the afternoon. The good agreement between the case study experiment and the RCM experiments suggests that the $\Delta u = 12$-km simulations are representing the changes in the FRMC properly.

Throughout this case, changes in tracer concentration associated with snow loss are largely shaped by the changes in the horizontal winds that advect tracer into and out of the mountain region. Other mechanisms that control the tracer concentration within the mountain region are changes in the boundary layer depth and ventilation of tracer through the boundary layer top into the free troposphere.

To quantify the relative importance of these mechanisms, we define a control volume over the Front Range mountains as the volume horizontally bounded by the geographic area within the dashed red box in Fig. 11 and west of the solid blue line. The model diagnostic boundary layer top is used as the top of the control volume and is allowed to vary in time as the boundary layer grows and retracts throughout the day. We
compute the tracer concentration within the control volume as it changes in time over the mountains and compare this with the integrated horizontal and vertical tracer fluxes directed into the control volume following

\[
\frac{dT_r}{dt} = \left( \int_{y_s}^{y_e} \int_{z_{dlc}}^{z_{top}} v T_r \, dx \, dz \right)_{y=y_s} - \left( \int_{y_s}^{y_e} \int_{z_{dlc}}^{z_{top}} v T_r \, dx \, dz \right)_{y=y_n} + \left( \int_{y_s}^{y_e} \int_{z_{dlc}}^{z_{top}} u T_r \, dy \, dz \right)_{x=x_w} - \left( \int_{y_s}^{y_e} \int_{z_{dlc}}^{z_{top}} u T_r \, dy \, dz \right)_{x=x_e} - \left( \int_{x_s}^{x_n} \int_{y_s}^{y_e} w T_r \, dx \, dy \right)_{z=z_{top}} + \left( \int_{x_s}^{x_n} \int_{y_s}^{y_e} w T_r \, dx \, dy \right)_{z=z_{sfc}} \right) V_{ctrl}^{-1}, \tag{5}
\]

where the subscripts \(x\) and \(y\) denote the zonal and meridional axes; the subscripts \(e\), \(w\), \(n\), and \(s\) denote the eastern, western, northern, and southern limits of the control volume; \(V_{ctrl}\) is the physical volume of the control volume; and \(z_{sfc}\) and \(z_{top}\) are the physical surface and the top of the boundary layer. In the following analysis, Eq. (5) is integrated in time. The tracer flux on the western boundary is negligible and not presented in the analysis. Additionally, we combine the fluxes on the northern and southern boundaries into a single term for simplicity. This framework allows for a comparison of horizontal and vertical tracer advection along with the effects of dilution within the boundary layer and how these interact with changes in the FRMC to determine differences in tracer concentrations in the Front Range mountains.

Time series comparing the evolution of tracer concentration contoured as a function of height, as well as the mean boundary layer tracer concentration, between ctrl and PGWSNO simulations are shown in (Fig. 14). In the mountain region, the tracer plume is largely confined to the boundary layer in both simulations, as evidenced
by the shading in Figs. 14a and 14b, indicating a limited role of ventilation. However, there is a vertical expansion of tracer in the PGWSNO simulation, illustrating the role of dilution. Despite dilution effects, there is up to a doubling of the mean tracer concentration within the boundary layer in the PGWSNO simulation compared to that of the control. Furthermore, the tracer remains in the control volume approximately 2 h longer in the PGWSNO simulation than in the control.

A comparison of the fluxes in the control simulation shows that tracer concentration within the mountains grows throughout the morning as the FRMC advects tracer through the eastern boundary of the control volume. The tracer concentration peaks at 1400 LST and then decreases throughout the afternoon as southerly and westerly flow advect tracer out of the control volume. Compared to the horizontal fluxes, the vertical tracer flux is small, indicating that tracer ventilation through the boundary layer plays a minor role.

The PGWSNO – ctrl change (Fig. 15b) shows a doubling of the westward tracer flux into the domain and a comparable increase in the tracer concentration. Consequently, there is a larger meridional flux of tracer out of the domain. The change in the vertical tracer flux is almost negligible despite substantial increases in tracer concentration, suggesting a diminished role of ventilation in the PGWSNO simulation.

The comparatively minor change in the vertical tracer flux is associated with changes in boundary layer development. While there is a clear increase in the vertical extent of the tracer plume in the PGWSNO simulation (Fig. 14b), the increase in boundary layer height limits increases in the vertical tracer flux through the boundary layer top. We quantify this reduction by taking the “ventilation ratio” (i.e., the ratio between the accumulated vertical and westward tracer fluxes). The ventilation ratio (computed as an average between 1300 and 2100 LST) was reduced by approximately 20% from 0.09 in the control simulation to 0.07 in the PGWSNO simulation. To quantify the effects of dilution, we take the ratio between the mean change in the control volume due to the increased boundary layer depth and the

Fig. 13. Hovmöller diagrams showing the diurnal evolution of (a) log10($T_{max}$) from the ctrl simulation and (b) $\Delta u$. The vectors in (a) show the 10-m horizontal wind vectors. The thick solid black lines in (b) show $\Delta T_r$. The thick dashed lines indicate the −3.5 contour of log10($T_{max}$) for the ctrl (black) and PGWSNO (red). The black lines on the right side of the panels show the mean downwelling TOA solar radiation. (c) A cross section of terrain, with a color-shaded line showing the mean fractional snow loss. The vertical red and blue lines correspond to the maximum western extent of the tracer plume [as defined by the log10($T_r$) = −3.5 contour].
control volume from the control simulation. This ratio is 0.07, indicating that the effects of dilution reduce the tracer concentration within the boundary layer by an average of 7%. Taken together, the combined effects of reduced ventilation and increased dilution act to reduce the tracer concentration a mere 5% compared to a >100% increase in tracer concentration caused by the increased westward tracer flux, indicating these effects are secondary.

Overall, these results support the conclusions of section 3 by showing similar dynamic changes in response to a snow-cover decrease. It should be noted that these results are specific to this 1-day case study and likely vary from case to case according to factors such as the background synoptic conditions and snow-cover extent. An additional caveat to these results is that there is no consideration regarding wet or dry deposition of the passive tracer. Finally, this analysis was performed for a specific source region of pollution, and these results would likely be very different when considering other distributions of sources. However, this analysis provides some preliminary insight into how dynamic changes east of the Front Range caused by the SAF may interact with regional pollution sources, suggesting that an increase in FRMC strength will likely increase the transport of pollution from the western Great Plains into the Front Range during the spring.

5. Conclusions

The RCM simulations presented here indicate that locally enhanced warming associated with the loss of snow cover over the Rocky Mountains causes a stronger diurnally driven mountain–plain circulation east of the Rocky Mountains by increasing the thermal contrast between the mountain atmosphere and the free troposphere over the Great Plains. The simulated FRMC circulation strength is increased by up to 2 m s$^{-1}$ east of the Front Range mountains. These changes are directly attributed to the SAF through use of the fixed-albedo experiment. Furthermore, these impacts extend 200 km east of the mountains, suggesting that the SAF over the mountains has wide-ranging impacts on regional circulations. Accordingly, accurately representing terrain features and snow processes over mountainous terrain in regional models may also affect their performance in simulating the climate response far from the the mountains. This has particular implications for the configuration of regional climate modeling studies. Additionally, GCM simulations that cannot properly resolve the terrain and its influence on the regional SAF will not be able to capture this feedback fully.

In a high-resolution, convection-permitting case study experiment, reducing mountain snow cover caused the FRMC to strengthen in a similar way as in the RCM experiments, supporting the conclusions drawn from the RCM experiment. In exploring the changes in transport of tracers emitted from the plains east of the Rockies, it was shown that the increase in FRMC strength advected more tracer into the mountains, increasing mean boundary layer tracer concentration by up to 100%, suggesting that SAF-forced increases in FRMC strength may bring more agricultural pollution into the mountains. Furthermore, it was shown that changes in boundary layer ventilation and dilution were secondary to that of increased westward tracer advection via the FRMC, suggesting that the additional pollution brought into the mountains largely stays concentrated within the boundary layer, where it is more likely to affect the regional ecology.
We chose to focus on the dry dynamic response of the FRMC to the SAF; however, it is plausible that the SAF-induced changes in the FRMC could have a substantial impact on the evolution of diurnal moist convection on the Front Range and perhaps even on the evolution of more organized convection downstream over the Great Plains. For instance, a stronger FRMC increases the convergence near the summit of the Front Range, which could lead to an increase in convective initiation there. Additional analysis could help clarify the role that the SAF and its effects on thermally driven circulations have in modulating precipitation changes in the Rocky Mountains as well as other mountain regions around the world.

As climate modeling expands to include an increased focus on high-resolution regional climate simulations, studies investigating how regional climate feedbacks such as the SAF modulate the local and regional atmospheric dynamics will be useful in helping to determine the mesoscale impacts of climate change. Studies such as this help illuminate some of the potential anthropogenic impacts associated with these mesoscale dynamic responses to climate changes, especially in ecologically and water-sensitive mountain regions.

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