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Key Points:

- Clouds can have a warming or cooling effect at the surface in the Upper Colorado River Basin (UCRB)
- The sign (warming or cooling) is highly sensitive to the surface albedo
- The albedo dependence implies that reductions in snowpack duration/extent will lead to a stronger and more persistent cooling effect

Supporting Information:

Supporting Information may be found in the online version of this article.

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Seasonality and Albedo Dependence of Cloud Radiative Forcing in the Upper Colorado River Basin

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Abstract Mountains create and enhance their own clouds, which both scatter and absorb shortwave radiation from the sun and absorb and re-emit land surface and atmospheric longwave radiation. However, the impacts of clouds on the surface radiation balance in high elevation snowy mountain terrain are poorly explored. In this study, we use data collected by the SAIL field campaign and partner organizations in the upper elevations (2,880 m.a.s.l.) of the Upper Colorado River Basin (UCRB) over a 21-month period from September 2021 to June 2023 to estimate Cloud Radiative Forcing (CRF) in the shortwave, longwave, and the net effect. Longwave warming effects dominate during the winter when snow albedos are high (0.8–0.9) and the background atmospheric precipitable water vapor is low (<0.5 cm), yielding a maximum monthly average net CRF of +34.7 W·m^{−2}, meaning that clouds increase the net radiation relative to clear skies during this time period. The sign of net CRF switches in the warm season as snow recedes, sun-angles increase, and the North American monsoon arrives, yielding a minimum monthly average net CRF of −47.6 W·m^{−2} with hourly minima of −600 W·m^{−2}. The sign of net CRF is typically positive, even at solar noon, when the surface is snow covered, except for a brief period over melting, low-albedo snow (0.5–0.6) impacted by dust impurities. Sensitivity tests elucidate the role of the surface albedo on the net CRF. The results suggest that net CRF will increase in magnitude and lead to a more persistent cooling effect on the surface net radiation budget as the snow cover declines.

Plain Language Summary The impact of clouds on the radiative energy at the Earth's surface in Upper Colorado River Basin is important, but their impact varies across seasons: clouds warm the surface in the winter and cool the surface in the summer. We show that the presence or absence of snow has a major impact on when and how clouds warm or cool the surface. Since less snow is predicted in the future, clouds will cool the surface more than they do now.

1. Introduction

Clouds both heat the surface by absorbing and emitting longwave radiation and cool the surface by attenuating or reflecting shortwave radiation from the sun. Globally, the cooling effect dominates, and clouds exert a net cooling effect on the Earth's climate of approximately 20 W·m^{−2} (Dessler, 2010). However, the radiative effects of clouds remain among the most significant sources of climate change uncertainty (Klein et al., 2018; Lutsko et al., 2021; Stephens, 2005). At the regional scale, the instantaneous radiative effects of clouds and their sign (warming, i.e., increasing net radiation at the surface; or cooling, i.e., decreasing the net radiation at the surface) depend on a number of factors, namely the large scale atmospheric structure, the micro and macrophysical properties of the cloud itself (Arking, 1991; Chen et al., 2000; Hartmann, 1993; Ramanathan et al., 1989; Shupe & Intrieri, 2004), and as the reflectance of the land surface (Ambach, 1974).

Mountains are unique environments for examining cloud effects and the terrestrial surface energy balance (SEB). For one, mountains can create or enhance clouds through a variety of dynamical mechanisms (Houze, 2012; Kirshbaum et al., 2018). The atmosphere above high mountains is typically dry given the thermodynamic constraints on atmospheric water vapor (Smith, 2019), and mountain surfaces are often covered by reflective snow and ice (Bales et al., 2006). Clouds only precipitate for a short portion of their life cycle (if at all). Meanwhile, their impacts can be felt at the surface at all times they are present through modulation of net radiation flux (R_{NET}) terms of the SEB. The SEB of a snow-covered surface is governed by the balance between R_{NET} and the sum of turbulent heat exchanges (sensible and latent), the ground heat flux, the energy lost to snowmelt phase transition, and the internal energy change within the snowpack (Armstrong & Brun, 2008).

Many studies have quantified the radiative effects of clouds at the Earth surface in the Arctic (Cox et al., 2016; Dong et al., 2010; Intrieri, 2002; Miller et al., 2015; Ryan et al., 2022; Sedlar et al., 2011; Shupe & Intrieri, 2004) and other regions (Berg et al., 2011; Dong et al., 2006; McFarlane et al., 2013), but few have focused on mountain regions (Conway et al., 2015; Marty et al., 2002; Nyeki et al., 2019).

The lack of investigations into clouds and R_{NET} is perhaps nowhere more glaring than the mountainous Upper Colorado River Basin (UCRB). The UCRB is undergoing a multidecadal “millennium drought” (Schmidt et al., 2023) that is likely driven by a number of nonexclusive factors (Bryant et al., 2013; Deems et al., 2013; Hogan & Lundquist, 2024; Milly & Dunne, 2020; Skiles et al., 2015; Udall & Overpeck, 2017; Xiao et al., 2018). The prospects of earlier slower snowmelt (Musselman et al., 2017) and a future with no or minimal snow accumulation (Siirila-Woodburn et al., 2021) highlight the hydroclimatic changes that are underway and are expected to accelerate. Like other high elevation regions, the UCRB is likely warming at rates exceeding that of global land areas (Byrne et al., 2024; Pepin, 2005; Rangwala & Miller, 2012). The causes of elevation-dependent warming are still not well understood (Mountain Research Initiative EDW Working Group et al., 2015; Naud et al., 2013; Pepin et al., 2022; Rangwala & Miller, 2012). Whether or not clouds are acting to oppose or enhance warming in mountain regions is an open question (Byrne et al., 2024; Chimborazo et al., 2022). The limited scrutiny of cloud effects may be due, in part, to the challenges that satellite remote sensing faces for distinguishing clouds versus snow (Stillinger et al., 2019) and observing multi-layer clouds (Mace et al., 2009; Naud et al., 2013; Wang et al., 2000), in addition to the more general challenge of maintaining high quality surface radiometric observations in mountains (Foken, 2008; Lehner & Rotach, 2018).

To meet these challenges, this study uses surface radiation and cloud cover data collected by the Surface Atmosphere Integrated Field Lab (SAIL) campaign (Feldman et al., 2023) to quantify Cloud Radiative Effect (CRE) and Cloud Radiative Forcing (CRF) in the East River Watershed (ERW), located near Crested Butte, Colorado, in the headwaters of the UCRB.

The goals of this study are to investigate the following high-level questions: (a) What is the magnitude, sign, and seasonality of CRF during the SAIL period? (b) How does CRF change from cold, snow-dominated season to the warm, snow-free season in conjunction with the spring ablation of snow cover? In order to fully contextualize the results from this inquiry, we also examine synoptic controls on cloud fraction using ERA5 reanalysis fields (Hersbach et al., 2020), and examine the SAIL results alongside independent, coarser spatial resolution estimates of CRE and CRF from the CERES SYN1deg Ed.4a product (Rutan et al., 2015; Wielicki et al., 1996). The observational data reported here will serve as benchmarks for atmospheric model competence in this region, in addition to providing a basic understanding of the radiative effects of clouds in the UCRB.

2. Methods

2.1. Study Site and SAIL Campaign

The ERW is a 300 km², high elevation (2,440–4,350 m.a.s.l.) subalpine watershed in the UCRB. The ERW flows into the larger Gunnison River Watershed (Figures 1a and 1b), which is the second most significant contributor to Colorado River streamflow (Xiao et al., 2018). The ERW is characterized by steep topographic relief and was heavily glaciated during the Pleistocene epoch until approximately 13 ka B.P. (Quirk et al., 2024). The valley floor is covered with a mixture of conifers, aspen, shrubs and grasses (Figure 1c), and Gothic Mountain casts shadows in the early morning and evening hours (Figure 1d).

The hydrometeorology of the ERW has been intensively studied by the SAIL (Feldman et al., 2023), NOAA SPLASH (de Boer et al., 2023), SOS (Lundquist et al., 2024), and Watershed Function SFA (Hubbard et al., 2018, 2020) field campaigns in addition to long-term ecological and climate monitoring activities (Prather et al., 2023). Recent observations of near surface air temperatures show a range between −35° during cold season nights and warm season maxima of 25–30°C (de Boer et al., 2023). In the upper elevations, over 800 mm of precipitation accumulates annually (Rudisill et al., 2023), and much of it as a seasonal snowpack beginning in October that melts in the valley floor by June. Broadly speaking, at least three primary storm regimes impact the ERW: winter baroclinic frontal systems associated with Pacific extratropical cyclones; remnant atmospheric river moisture (Kirk & Schmidlin, 2018; McGinnis, 2000; Xiao & Lettenmaier, 2021); and warm season convective events sometimes associated with moisture export from the North American monsoon (Adams & Comrie, 1997).

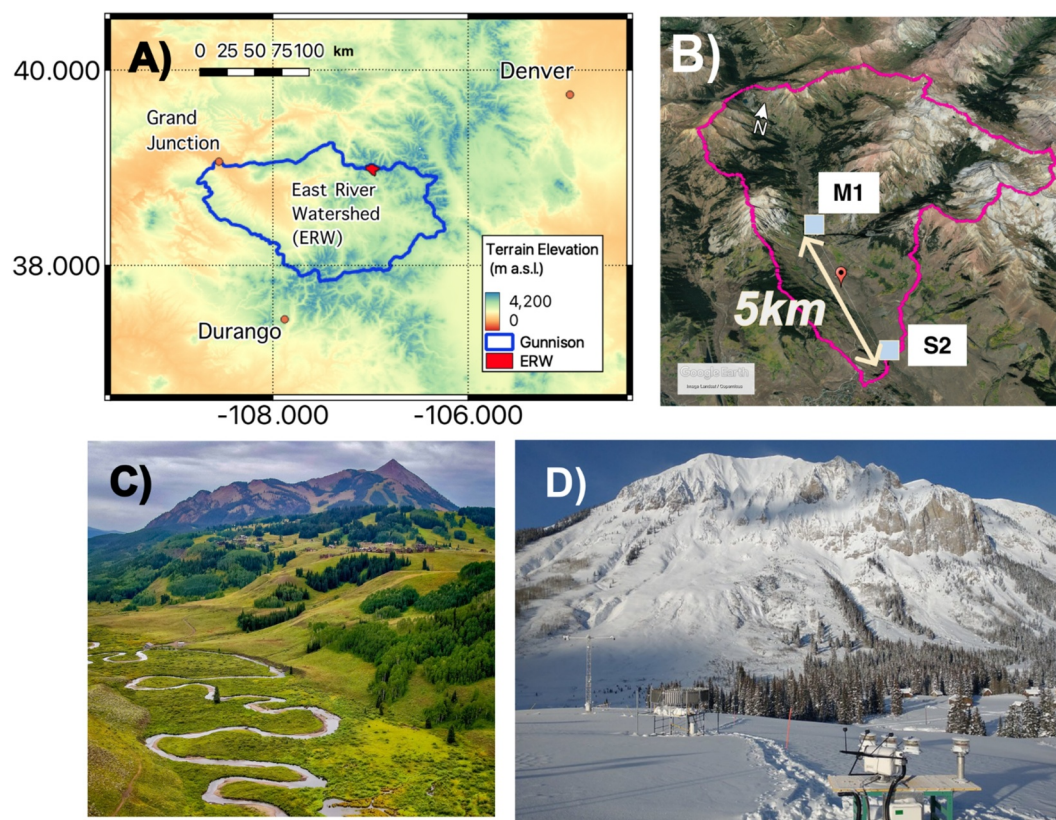


Figure 1. An overview of the East River Watershed (ERW) study region: (a) The topography of the ERW, surrounding mountains, and the greater Gunnison river watershed. (b) A zoom-in view of the ERW, with the primary SAIL site (M1) and supplemental site (S2) highlighted. (c) A summer view of the ERW valley, approximately looking south from the S2 site. (d) Mid-winter view of the M1 site looking approximately west toward Gothic Mountain. The radiometer system used in this study is visible in the bottom right. The image in panel (c) is available courtesy of Ken Williams, Lawrence Berkeley National Laboratory, and the image in panel (d) is available courtesy of the U.S. Department of Energy Atmospheric Radiation Measurement user facility.

SAIL was a deployment of the U.S. Department of Energy's Second Atmospheric Radiation Measurement (ARM) Mobile Facility (AMF-2) (Mather & Voyles, 2013) from September 2021 to June 2023 and included measurements of clouds, radiation, and surface properties using both passive and active ground based remote sensing. The ARM program uses well developed frameworks to create best estimate products from multi-sensor raw data streams (Mather & Voyles, 2013; Shupe et al., 2016; Stokes & Schwartz, 1994; Xie et al., 2010), termed value-added product (VAP). Section 2.3 describe the datastreams used in this study. We use data collected primarily from two different observing sites in the ERW across a span of 5 horizontal km (Figure 1b). The SAIL M1 site (2,880 m.a.s.l) was the primary observing site and the location of twice daily radio soundings (Figure 1b) in addition to radiation and meteorological observations. The S2 site south of the ERW hosts (among other instruments) a continuously monitoring webcam facing north toward the M1 site that provides a qualitative view of cloud cover.

2.2. Quantifying Cloud Radiative Forcing at the Surface

The effect of clouds on the surface radiation budget can be quantified using the CRF concept (Ramanathan et al., 1989). An estimate of equivalent clear-sky radiation, obtained by an empirical estimate or radiative transfer model, is subtracted from the observed radiation field, which may or may not contain clouds.

The transmittance by shortwave radiation through a cloud layer is a function of many factors including the cloud microphysical characteristics, the solar zenith angle, and the albedo of the surface. The attenuation of shortwave radiation by the clear-sky is primarily a function of water vapor, oxygen, ozone, aerosols, and Rayleigh scattering. Longwave effects are primarily a function of cloud temperature and the microphysical properties of that cloud,

particularly the presence of liquid water (Shupe & Intrieri, 2004; Stephens, 1978). Since a cloud layer is a similar temperature to the background atmosphere at the same level, the height of the cloud layer inherently influences longwave radiative effects.

We adopt the general approach of Miller et al. (2015), Cox et al. (2015), and Ramanathan et al. (1989) to quantify CRF, which is given by:

$$CRF_F = F_{all-sky} - F_{clear} \quad (1)$$

$$F^x = f^x \downarrow - f^x \uparrow, \quad (2)$$

where F is the net flux at the surface for either broadband shortwave (SW) or broadband longwave (LW), $f^x \downarrow$ and $f^x \uparrow$ are the upwelling and downwelling hemispherically integrated components of their respective fluxes. The subscript refers to “all-sky” conditions, which may or may not have a cloud present (e.g., SAIL observations of radiation), and “clear” refers to the radiation estimate from an atmosphere with no clouds present. The net CRF is, therefore, the sum of the longwave and shortwave components, or $CRF_{NET} = CRF_{SW} + CRF_{LW}$. The terms can also be rearranged to show $CRF_{NET} = R_{NET}^{all-sky} - R_{NET}^{clear}$. Thus, a positive CRF_{NET} indicates that clouds increase the surface net radiation compared to clear skies, and, and a negative CRF_{NET} indicates less energy available at the surface under cloudy conditions.

Some simplifications of CRF_{SW} are possible. The shortwave albedo of the land surface is defined as the ratio of the upwelling and downwelling shortwave radiation terms. Albedo (α) is not an invariant property of the surface, but can depend on cloudiness, particularly over snowpacks (Male & Granger, 1981), but this work treats α as invariant regardless of cloud cover (see in Supporting Information S1 for additional discussion). If we assume that the albedo does not change between clear and cloudy scenes, it follows that

$$CRF_{SW} = (1 - \alpha)[SW_{all-sky} \downarrow - SW_{clear} \downarrow]. \quad (3)$$

The land surface both reflects and emits radiation in the longwave spectrum, complicating the treatment of the upwelling terms of CRF_{LW} . We explicitly account for the difference in longwave emission from the surface between clear and cloudy scenes following Dong et al. (2010) and Cox et al. (2016). We use an XGboost machine learning regression model (Chen & Guestrin, 2016) to estimate clear-sky upwelling longwave based on near surface air temperature, humidity, surface albedo, and the solar zenith angle. Additional details are described in the Supporting Information S1. Ultimately, the choice to explicitly account for changing surface emission has a small impact on the overall magnitude and sign of CRF.

In addition to CRF, we also compute the cloud radiative effect (CRE),

$$CRE_f = f_{all-sky} \downarrow - f_{clear} \downarrow \quad (4)$$

which is simply the difference between the all-sky and clear-sky downwelling radiation fluxes.

2.3. Observations

2.3.1. Surface Radiation

Radiation fluxes are observed using the ARM SkyRad instrument collection (Andreas et al., 2018) which consist of Eppley Laboratory, Inc. Pyranometers with a spectral bandpass between 295 and 3,000 nm. The pyrgeometers have a spectral bandpass between 3.5 and 50 μ m. The raw data from the sensors are quality-controlled by the “QCRAD” VAP procedure (Long & Shi, 2006). QCRAD applies a correction for shortwave infrared loss and flags suspect data based on estimated limits for the given climatology of the region. QCRAD uses the sum of the diffuse and direct components to represent $SW \downarrow$ whenever possible (Long & Shi, 2006).

Technicians physically checked the radiometers for each day of the SAIL campaign and recorded maintenance notes. Cameras recorded continuous imagery and were used to check for snow accumulation on the radiometer domes. Manual checks of the imagery showed that snow accumulation of any amount in the vicinity of the radiometer domes occurred for fewer than 4% of the observation days, so the total number of affected data hours is

much less than that. Riming can also be a significant challenge for taking accurate radiation measurements in cold-regions (Cox et al., 2021; Miller et al., 2015). Although there are likely times when riming did occur, we do not believe that rime was a significant problem at the SAIL site based on the technician notes and our own review of the data. Less than 1% of the 1-min QCRAD downwelling short and longwave data failed the quality control checks during the study period. We ultimately examine data at the hourly average timescale. Given that the observed radiation data coverage is quite high, we simply omit unobserved minutes from averaging calculations. Ad-hoc tests showed this decision had little impact on the overall interpretation of the results.

2.3.2. Clouds and Atmospheric Structure

Cloud occurrence and cloud properties are measured in several ways. To examine cloud base height (CBH) and cloud fraction (C_{frac}), we use the Active Remote Sensing of Cloud Locations (ARSCL) VAP. The ARSCL algorithm (Clothiaux et al., 2000) merges data from a ceilometer, Ka-Band radar, micropulse lidar, microwave radiometer, and surface instruments to retrieve cloud information. The ARSCL CBH is determined using a combination of ceilometer and micropulse lidar and should be unaffected by attenuation from rain or drizzle, whereas cloud top (and therefore cloud thickness) requires identification of the cloud top radar signature, which is less-certain during rainy conditions or periods with multi-layer clouds. The ARSCL product produces CBH estimates every 4 s. A separate value is recorded to indicate if no cloud base is detected. We define C_{frac} temporally, as the number of cloud detections with a cloud base less than 6 km above the valley floor, divided by the number of total measurements per hour. The 6 km threshold was chosen based on subsequent analyses, since we found little impact on the surface radiation when clouds are above this height. Data resampling is performed using the Pandas python library.

In addition, we examine data from the hemispheric Total Sky Imager (TSI) (Morris, 2005), which uses visible photography to estimate hemispheric cloud fraction during the daytime and acts as a secondary check on C_{frac} computed from ARSCL data.

2.4. Estimating Clear-Sky Radiation

The Rapid Radiative Transfer Model for GCMs (RRTMG) longwave (RRTMG_{LW}) and shortwave (RRTMG_{SW}) radiative transfer models, version 4 (Iacono et al., 2008) are used to estimate clear-sky downward radiation fluxes. RRTMG_{SW} uses a correlated-k approach with 14 spectral bands. The RRTMG_{LW} model has a spectral range of 3.08–100.0 μm divided into 16 spectral bands. The models are configured by inputting hourly vertical profiles of interpolated radiosoundings of atmospheric temperature, barometric pressure, and water vapor at 100 evenly spaced vertical levels up to 10 hPa. The total water vapor path of each hourly sounding is corrected by the precipitable water vapor (PWV) retrieved from the sky brightness temperatures at 23.8 and 31.4 GHz frequencies measured by a microwave radiometer (Morris, 2019). These data are processed as part of the ARM “interponde” VAP (Jensen et al., 2021). Quantities at the lowest model level are taken from automated weather stations at a two-meter observation height. Ozone profiles come from data collected by the NOAA ozone monitoring laboratory, located approximately 200 km east of the ERW (Sterling et al., 2018). RRTMG_{LW} is run with hourly inputs and RRTMG_{SW} is run with a 10-min frequency. The results are averaged to the hourly level to better account for nonlinearity introduced by the changing solar zenith angle.

Since the observations are located in complex terrain (Figure 1), some additional factors must be taken into account. The M1 site was located on a flat slope, but is surrounded by steep slopes in all directions. We ignore all timesteps where the sun's direct beam is shadowed by surrounding terrain for the computation of CRF_{SW} and CRE_{SW}. Instances when the radiometers are shadowed is computed using a high resolution digital elevation model and solar position calculations (see in Supporting Information S1). Ultimately, the morning and evening periods where the radiometers are shadowed account for less than 9% of the total daytime hours examined, and these hours correspond with less than 5% of the total TOA shortwave radiation flux averaged over the SAIL period. Other complex terrain related processes (Feldman et al., 2022; Lee et al., 2019; Manners et al., 2012; Matzinger et al., 2003) that affect SW $\downarrow$ and LW $\downarrow$ are not directly accounted for, including multiple scattering shortwave radiation from nearby snow covered sloping terrain, as workable corrections are not well established.

Despite these potentially complicating factors from complex terrain, the RRTMG calculations validated well against observations of radiation during clear skies, excluding periods when the radiometer is shadowed by terrain (Figure 2). We test both models against observations during clear-sky conditions ($C_{\text{frac}} < 1\%$), and find that

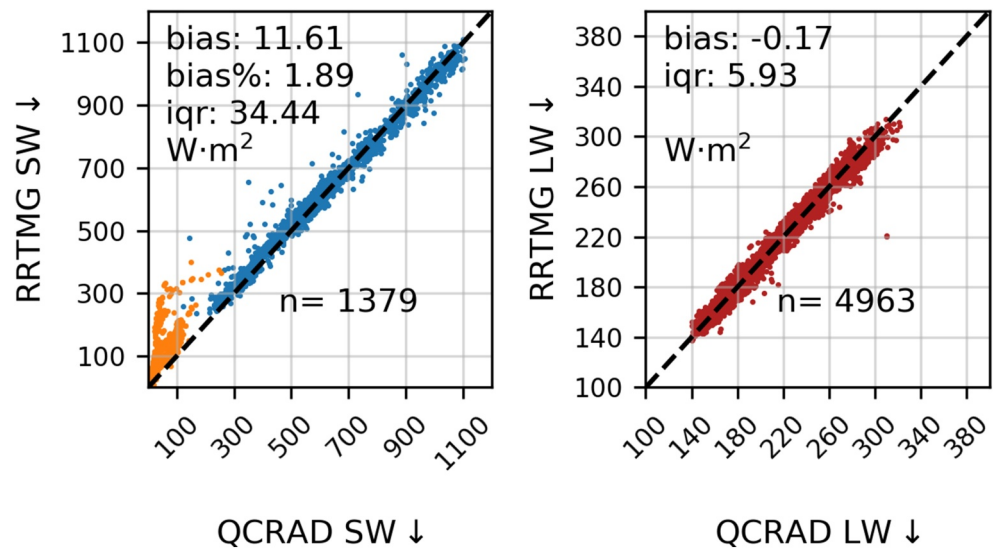


Figure 2. Clear-sky Rapid Radiative Transfer Model for GCMs (RRTMG) radiation transfer calculations of broadband hemispheric irradiance for SW↓ and LW↓ compared against observed QCRAD data at the SAIL site during clear-sky conditions. The bias and interquartile range of the error distribution are also shown. Timesteps where the radiometers are shadowed by surrounding terrain are shown in orange and are not considered in the bias or interquartile range calculation.

downwelling shortwave has a mean bias of $11.6 \text{ W}\cdot\text{m}^{-2}$ with a bias interquartile range of $34.4 \text{ W}\cdot\text{m}^{-2}$. This bias represents 2% of the clear-sky, nonshadowed, SW↓ averaged across the entire study period. Downwelling longwave has a mean bias of $-0.2 \text{ W}\cdot\text{m}^{-2}$ with an interquartile range of $5.9 \text{ W}\cdot\text{m}^{-2}$. The low bias of the longwave model is perhaps not surprising, given that longwave radiation is primarily a function of the temperature at the lowest atmospheric levels which are input into the model.

The clear-sky model errors are comparable to similar studies (Aebi et al., 2017), and appropriate given the target observation accuracies of $\pm 2\text{--}5.0 \text{ W}\cdot\text{m}^{-2}$ for downwelling shortwave and $\pm 9.0 \text{ W}\cdot\text{m}^{-2}$ for downwelling longwave (Augustine et al., 2000; McArthur, 2005). Moreover, comparisons against an independent ARM radiometer system (Cook, 2024) located nearby shows that the modeled shortwave estimate is in fact slightly closer to the QCRAD observation than the other system. For these reasons, we conclude that the clear-sky models are sufficient for the purposes of estimating CRE or CRF at this particular location, but keeping in mind that the CRF estimate is perhaps off by as much as 2% (relative to the average SW↓) due to the bias of the shortwave model.

2.5. CERES SYN1deg Ed.4a Data

We also examine data from the NASA Clouds and the Earth's Radiant Energy System (CERES) Synoptic (SYN) 1-degree (1deg) data product (Rutan et al., 2015; Wielicki et al., 1996), hereafter referred to as CERES SYN1deg Ed.4a. This hourly, $1 \times 1^\circ$ product merges geostationary satellite observations, TOA broadband radiance measurements from the polar orbiting Terra and Aqua satellites, and several other observations of the atmospheric and surface state, combined with radiative transfer modeling to produce estimates of shortwave and longwave fluxes for both all-sky and clear-sky conditions at the surface. Data from 2000 to 2023 are used to compute independent estimates of CRE, CRF, and C_{frac} for the grid cell nearest the ERW, as well as the entirety of North America land areas between 25 and 60°N . Since the data resolution is coarse, comparisons against CERES SYN1deg Ed.4a are not intended as a validation of either model, but rather serve as an independent data point for contextualizing the SAIL results beyond the SAIL time period and location.

3. Results

3.1. Seasonal and Diurnal Characteristics of Clouds in the ERW

C_{frac} evolves seasonally during the SAIL period and responds to both large scale synoptic forcing and local thermodynamic forcing during the warm season (Figure 3). The highest C_{frac} occurred during the cold season (ONDJFM) of 2023. This year also experienced greater snow accumulation compared to the 2022 (not shown).

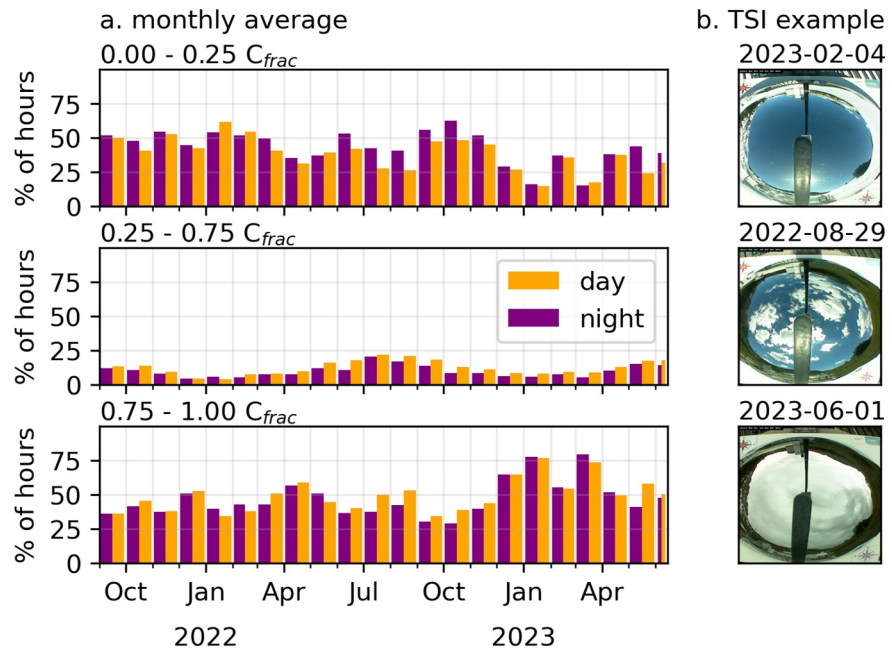


Figure 3. (a) The monthly percentage of hourly scenes with 0–0.25, 0.25–0.75, and 0.75–1.0 C_{frac} during the SAIL period. (b) Example daytime scenes for each category from the Total Sky Imager are depicted on the right hand side of each row.

The distribution of C_{frac} is highly bimodal (close to 0.0 or 1.0), though intermediate C_{frac} values become increasingly common during the warm season. These intermediate C_{frac} are characteristic of shallow cumuli that develop from surface based convection either locally or nonlocally (Figure 3, middle panel), and are also visible in the TSI imagery.

By grouping C_{frac} by hour-of-day and splitting the data into the cold and warm seasons, we can observe additional details about the diel cycle of clouds in the ERW (Figure 4). For each hour-of-day, the mean C_{frac} is estimated using one hundred independent bootstrap samples with replacement. We find a more distinct diel cycle of C_{frac} during the warm season, where peak C_{frac} occurs approximately 2 hours after noon. During the cold season, C_{frac} has a weaker diel cycle and is generally between around 0.5–0.6, with a modest increase in likelihood in the afternoon. The lowest likelihood of detecting a cloud is at 7:00 a.m. local time during the warm season (0.35–0.5).

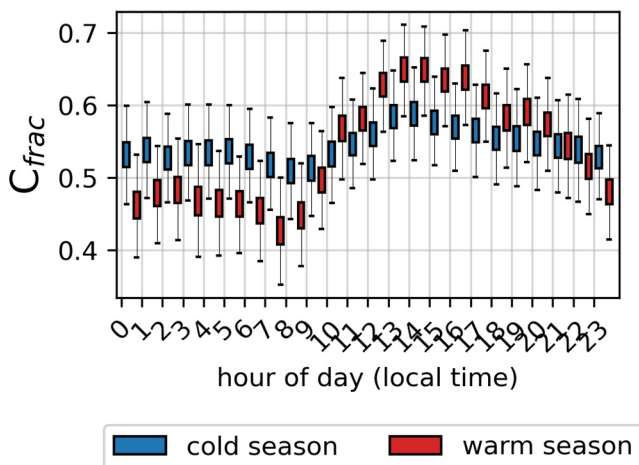


Figure 4. Warm (AMJJAS) and cold season (ONDJFM) C_{frac} as a function of the time of day (local time). Boxplots show estimates of the mean C_{frac} generated by one hundred bootstrap samples with replacement.

Examining the synoptic regimes that can influence C_{frac} , shows, perhaps unsurprisingly, that clear skies (Figures 5a–5d) correspond with high pressure synoptic-scale anomalies, and cloudy periods are found during low pressure anomalies (Figures 5b and 5c). Composites are computed from 6-hourly ERA5 geopotential height fields (Hersbach et al., 2020) using daily average heights provided by Lee et al. (2023) to compute the anomalies. The direction of 500 mb flow over the ERW is primarily from the southwest during cloudy periods and the west or northwest during clear periods; however, the cloudy and cloud-free wind direction regimes show better separation during the cool versus warm season (Figure 5a–5d). The cloudy, cold season pattern most closely matches the Pacific ridge regime described in Lee et al. (2023) (their Figure 10) and the clear-sky pattern is most similar to the Pacific trough regime. The cloudy composite regime during the warm season is indicative of monsoon circulation, with flow from the southwest and ridging in the south Pacific. The warm season clear-sky composite is characterized by westerly flow and an Alaskan trough.

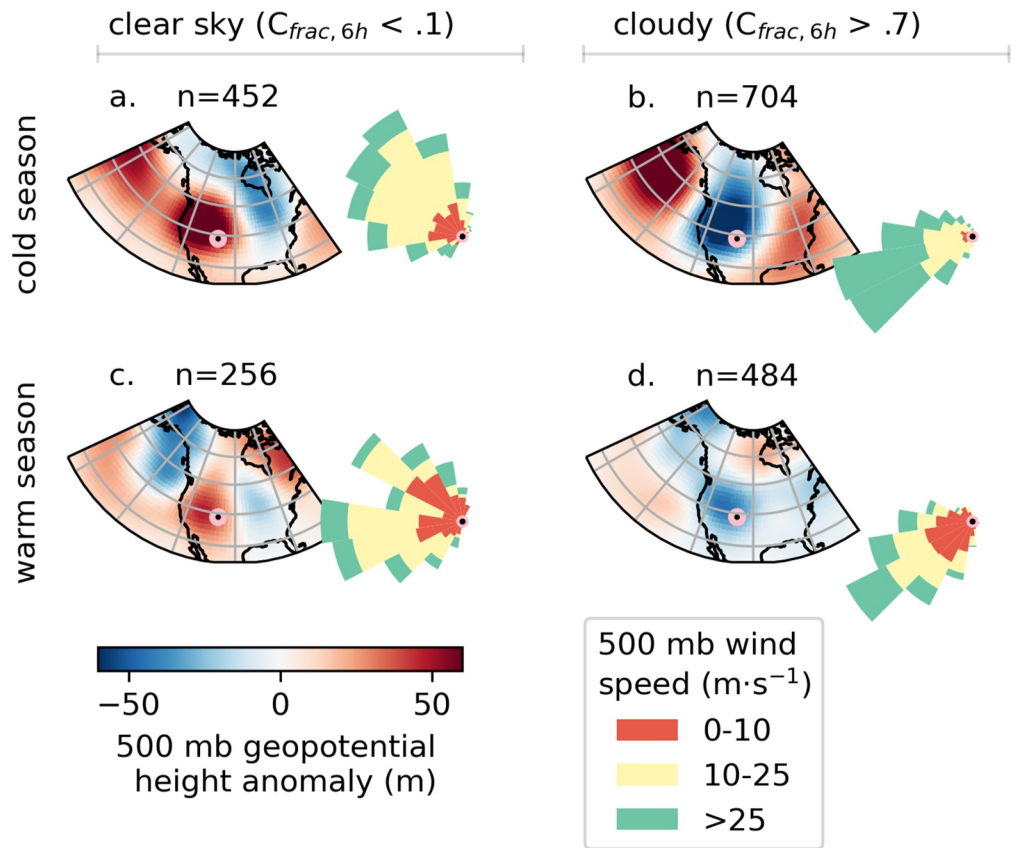


Figure 5. Synoptic controls on clouds in the East River Watershed (ERW)—the ERA5 500 mb geopotential height anomaly, and 500 mb wind rose plots (right of each respective anomaly map plot) for the grid box containing the ERW (pink circle). Anomalies and wind roses are computed for clear ((a) and (c); defined as 6-hourly observed cloud fractions less than 0.1 C_{frac}) and cloudy ((b) and (d); greater than 0.7 C_{frac}) conditions for the cold (ONDJFM; top row) and warm (AMJJAS; bottom row) seasons. The number (n) of 6-hourly observations retained for each plot are shown.

3.2. Differences Between All-Sky and Clear-Sky Radiation

We can now examine CRE during the SAIL period (Section 2.2). An example of the CRE calculation is depicted in Figure 6 with concurrent images from the webcam (panels a–d), located at S2 (Figure 1), looking north toward the main SAIL site (pink open circle). Clouds are not detected until 9:00 a.m. local time, and the skies are completely cloudy after 2:00 p.m. Correspondingly, CRE_{LW} reaches $+60 \text{ W} \cdot \text{m}^{-2}$, whereas CRE_{SW} reaches a minimum of less than $-300 \text{ W} \cdot \text{m}^{-2}$.

Extending this analysis to the entire SAIL period (Figures 7a–7d) shows that clouds overwhelmingly have a negative CRE_{SW} (Figure 7b) and a positive CRE_{LW} (Figure 7c). The hourly CRE_{SW} is closely tied to incident TOA radiation (Figure 7a), with the most negative periods occurring near the summer solstice ($-1,000 \text{ W} \cdot \text{m}^{-2}$) when the sun is at its highest elevation and most intense. Comparing the TOA SW radiation (gray curve) against the $RRTMG_{SW}$ modeled clear-sky radiation at the surface shows that the clear-sky radiation decreases with respect to TOA radiation during the warm season. This is due largely to the seasonality of aerosols and water vapor, which both increase during the warm season (Figure S1 in Supporting Information S1). There are also notable periods during the summer where clouds cause a positive CRE_{SW} , and in some cases even exceed the expected TOA radiation. This phenomenon has been demonstrated elsewhere and often occurs from forward scattering by clouds when the sun's direct beam is unblocked (Berg et al., 2011; Cordero et al., 2023; Damiani et al., 2018; Gueymard, 2017; Pfister et al., 2003).

Compared to the hourly averages, the monthly average values of CRE_{SW} and CRE_{LW} reflect both the individual radiative effects and the frequency of cloud occurrence. The monthly averaged CRE_{SW} reaches a minimum approaching $-100 \text{ W} \cdot \text{m}^{-2}$ during the warm season, and CRE_{LW} reaches a maximum of almost $50 \text{ W} \cdot \text{m}^{-2}$ during

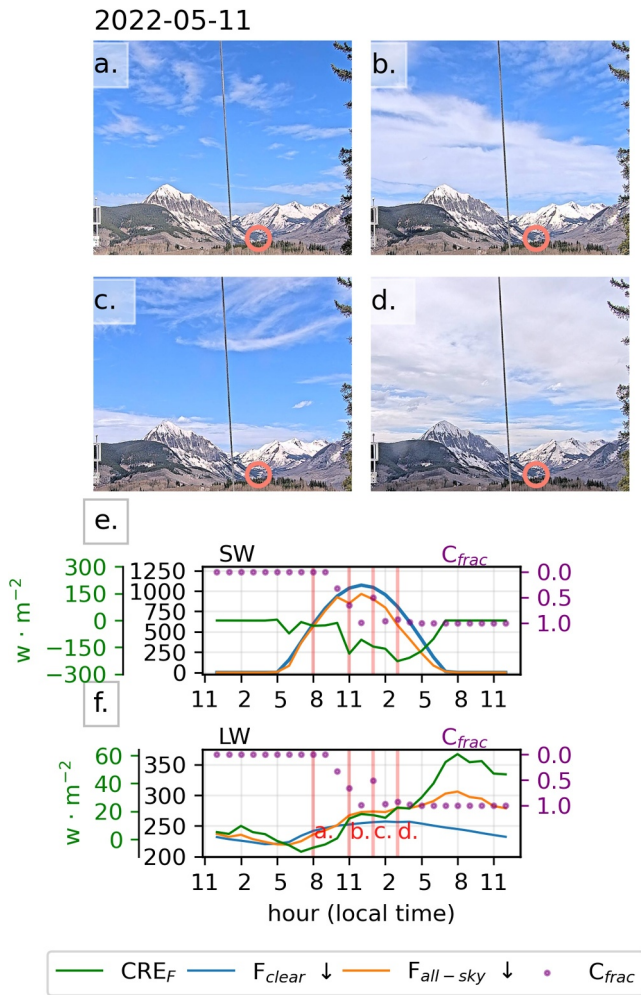


Figure 6. An example of shortwave and longwave CRE from 11 May 2022. Web camera images looking north toward the SAIL site (pink circle). (a–d) Qualitative views of the evolution of clouds (e) CRE_{SW} , (f) CRE_{LW} .

the cold season (Figure 7d). The monthly averaged CRE_{LW} is substantially higher in the cloudier winter of 2023 compared to 2022 due to the more frequent occurrence of storms in 2023 (Figure 3).

3.2.1. Diminishing Longwave Effect During the Warm Season

The maximum hourly CRE_{LW} tends to decrease during the warm season (Figure 7). To assess possible causes, we isolate periods where clouds are present using a threshold of 0.9 hourly C_{frac} (43% of all hourly scenes), and examine CBH, PWV, and CRE_{LW} during these periods (Figure 8).

PWV increases markedly in the warm season, from a mean value of 0.40 cm during the cold season to a mean of 0.95 cm during the warm season (Figure 8a) with a monthly maximum of over 1.25 cm in July (Figure S1 in Supporting Information S1). Cloud bases likewise increase in height during the warm season to a modal height of 2,500 m above the valley floor (Figure 8b). These factors create a distribution of CRE_{LW} with two peaks—one characteristic of high clouds and a weak longwave effect and a second peak composed of low clouds (Figures 8c and 8d). The highest values of CRE_{LW} are found during the cold season, and are characteristic of a low PWV background state with low cloud bases (Figure 8d).

There are likely two effects that contribute to the shift in CRE_{LW} to lower values during the summer. First, it should be recognized that, when looking upwards from the surface, a cloud layer will appear very similar to a high emissivity greybody emitter at the same temperature of the atmosphere at that level. The same is not true for the clear-sky atmosphere. The spectroscopic absorption of water vapor modulates CRE_{LW} through rotational band absorption in the far-infrared at wavelengths longer than 15 μm , continuum absorption in the mid-infrared window from 8 to 12 μm , and mid-infrared vibration-rotational band absorption at around 6 μm (Harries et al., 2008). Because the absorption coefficients are larger in the rotational band than the mid-infrared window, when PWV > 1 cm the rotational band is saturated by atmospheric gases alone and no cloud forcing is possible at these frequencies, whereas the mid-infrared remains semi-transparent and additional forcing by clouds is observable. In the winter, however, the atmosphere is much drier and the far-infrared is semi-transparent (Feldman et al., 2014) so CRE_{LW} occurs as a result of the absorption of clouds over a much broader range of

longwave emission, including emission close to the peak of Planck function emission of the surface and atmosphere. This phenomenon is complicated by the fact that the integral of the Planck function is larger in summer, since the atmosphere is warmer, which permits the mid-infrared cloud forcing to be potentially larger in summer than it is in winter. Generally, correlated increases in temperature and water vapor will reduce CRE_{LW} for PWV > 1 cm while compensating effects between mid- and far-infrared stabilize relatively large CRE_{LW} for PWV < 1 cm (Cox et al., 2015).

The height of the cloud base also matters: it increases substantially from the cold season, with a modal peak of 1,000 m.a.g.l., to the warm season, with a modal peak of 2,500 m.a.g.l. (Figure 8b). Since the radiating temperature diminishes as clouds increase in height, a higher cloud will be much colder than the near surface where most of the clear-sky longwave flux originates (Ohmura, 2001), and CRE_{LW} may be lessened relative to a lower cloud base with a warmer radiative temperature. As a point of reference, a similar relationship between water vapor and CRE_{LW} is also described in Aebe et al. (2017) at a site in the Swiss Alps, and the distribution of CRE_{LW} has a very similar modal value.

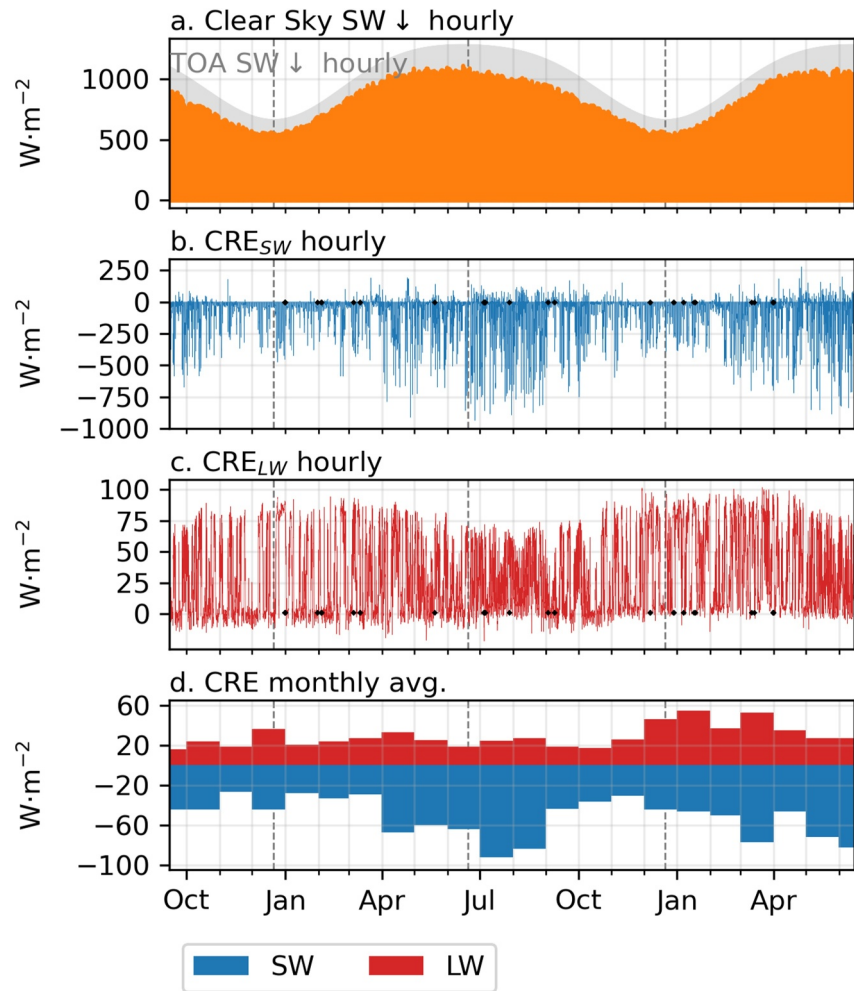


Figure 7. (a) Hourly TOA incident shortwave radiation and downwelling clear-sky shortwave radiation at the surface, (b) hourly average CRE_{SW} , (c) hourly average CRE_{LW} , and (d) monthly average CRE_{LW} and CRE_{SW} observed at the SAIL site. Black dots indicate missing hourly values. Dashed lines indicated the winter and summer solstices, respectively.

3.3. Cloud Radiative Forcing

The previous discussion of CRE only considers the differences in the downwelling terms of atmospheric radiation (Equation 4) between all-sky and clear-sky conditions. The following section examines CRF_{NET} , which also takes into account upwelling terms of R_{NET} (Figures 9a–9c).

The pattern of CRF_{NET} is similar to the pattern of CRE_{SW} , and closely follows the solar ephemeris, but the seasonality of the surface albedo, driven by the presence or absence of snow, also plays an important role (Figure 9b). At an hourly timescale, clouds have the most significant effect during the warm season when the albedo is low (0.2) and the sun is at high elevation angles near the summer solstice, yielding a minimum CRF of $-600 \text{ W}\cdot\text{m}^{-2}$. The hourly maximum CRF_{NET} is smaller in magnitude seldomly exceeds $100 \text{ W}\cdot\text{m}^{-2}$ (Figure 9b).

Separating the monthly averaged CRF_{NET} into daytime and nighttime helps distinguish the role of changing day-lengths and solar insolation on CRF_{NET} . We find clouds always have a net warming effect at night, when shortwave effects are by definition zero. However, during the cold season and over snow, CRF_{NET} remains positive even during the daytime hours (Figure 9c). Moreover, not only is the average CRF_{NET} positive when averaged over all daytime hours, but CRF_{NET} remains positive even at solar noon for DJFM and DJFMA (for 2022 and 2023, respectively) when the incident downwelling shortwave radiation is the most intense.

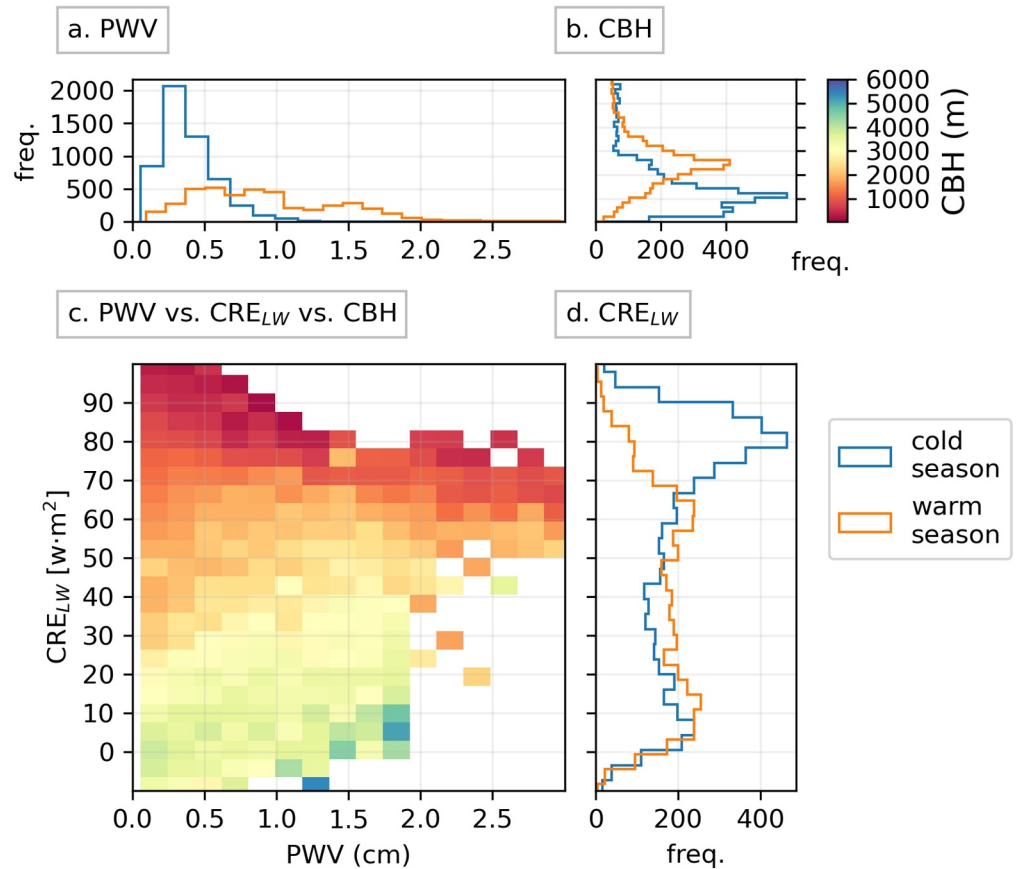


Figure 8. The relationship between precipitable water vapor (PWV), cloud base height (CBH), and CRE_{LW} for time periods where C_{frac} is greater than 0.9 during the SAIL campaign. (a) Histogram of PWV, (b) histogram of CBH, (c) gridded heatmap of average CBH as a function of PWV and CRE_{LW} , and (d) histogram of CRE_{LW} . The y-axis of panel (b) serves as the color ramp for panel (c).

Ultimately clouds warm the surface on average for 5 months of the first year of SAIL and 6 months of the second year. The JFM monthly average ranges between 14.3 and 34.7 $W \cdot m^{-2}$, and the JAS monthly average ranges between -18.5 and -47.6 $W \cdot m^{-2}$ (Table 1). Averaging across the entirety of 2022 (a complete 12 months of data, as opposed to data from 2023) yields a CRF_{NET} of -7.9 $W \cdot m^{-2}$. The sign switch from positive to negative occurred in May of both years and switched from negative to positive in December and November for 2021 and 2022, respectively.

3.3.1. The Influence of Albedo on CRF_{NET}

The transition from cooling to warming coincides with the loss of snow and albedo decline. Revisiting Equation 3 helps explain the CRF_{NET} dependence on albedo, and specifically the $1-\alpha$ coefficient, also called absorption or co-albedo (Bair et al., 2019). Although a cloud layer may scatter/absorb a significant amount of shortwave radiation away from the surface, the net effect on the surface is diminished by bright snow, which only absorbs 10%–20% of the incident radiation regardless. A small change in albedo causes a large increase in absorption. For instance, a 12% reduction in albedo from 0.8 to 0.7 (characteristic of snow aging or the introduction of impurities) yields a 50% increase in the absorbance, and therefore magnitude of CRF_{SW} based on Equation 3. The relationships between albedo and CRF are explored further in Section 3.5 and 3.4.

3.3.2. Clear-Sky Upwelling Longwave Radiation

In addition to the downwelling terms, CRF_{LW} is modulated by the land or snow surface longwave emission, which also responds to the presence or absence of clouds. The estimates of clear-sky upwelling longwave (Section 2.2) show, as expected, that the clear-sky upwelling emission is generally higher during the daytime (i.e., the ground

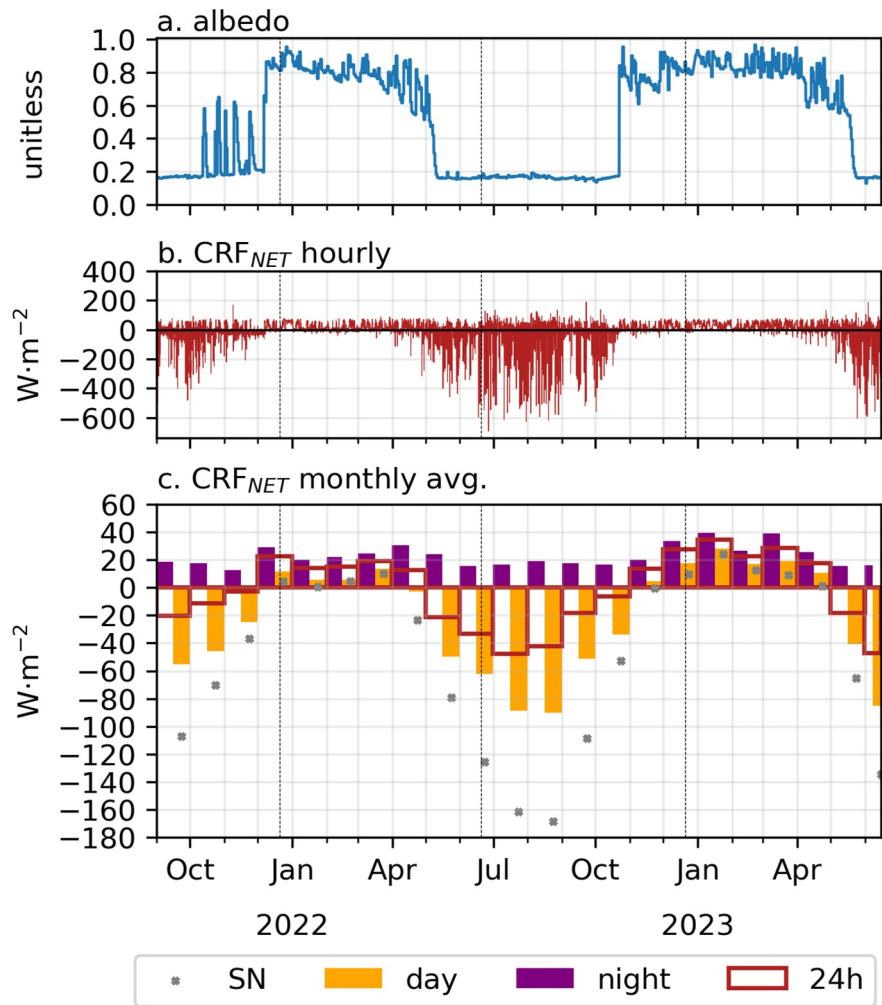


Figure 9. (a) Land surface albedo, (b) hourly average CRF_{NET} , and (c) monthly average CRF_{NET} for the hour nearest solar noon (SN), day, night, and full 24-hr average time periods. Daytime is defined as all times when the solar altitude angle is greater than zero (middle). The daytime and nighttime bars are unweighted by the length of the day, but the total bars are. Black dots indicate missing hourly values. The dashed vertical lines indicate the winter and summer solstices, respectively.

surface has a higher radiative temperature), but lower at night, when compared to cloudy conditions (Figure S4 in Supporting Information S1). Ultimately, the model predicts that the hourly maximum difference between all-sky and clear-sky scenes increases markedly during the warm season compared to the cold season (Figure S6 in Supporting Information S1), but the difference is almost zero at a monthly average. Moreover, the model predicts substantially less clear-sky upwelling longwave during cold season of 2023 (i.e., equivalent clear-sky scenes have colder radiative temperatures compared to all-sky), on the order of $10 \text{ W} \cdot \text{m}^{-2}$ per month, which acts to oppose the positive downwelling CRE_{LW} . Averaged across the entire period, the clear-sky upwelling longwave is $4 \text{ W} \cdot \text{m}^{-2}$ less than the all-sky upwelling longwave, so the total effect the upwelling longwave term is small compared to other terms of CRF_{NET} .

3.4. Case Study: Dust-on-Snow in the Spring of 2023

It is well known that dust deposition on snow surfaces is a major factor that depresses the albedo of snowpacks in the western US and in the UCRB in particular (Painter et al., 2007, 2010; Skiles et al., 2015). Dust was observed at the SAIL site during the spring of 2023, and the observed spectral albedo of the snow surface confirms that dust deposition played an important role on

Table 1
Summary Statistics of Monthly Average CRF_{NET} Aggregated by Season

	Season	Min	Max	Mean
Cold Season	OND	−11.4	27.6	7.2
	JFM	14.3	34.7	22.3
Warm Season	AMJ	−47.4	17.6	−15.0
	JAS	−47.6	−18.5	−32.2

Note. The units are $\text{W} \cdot \text{m}^{-2}$. The same data are presented in Figure 9c.

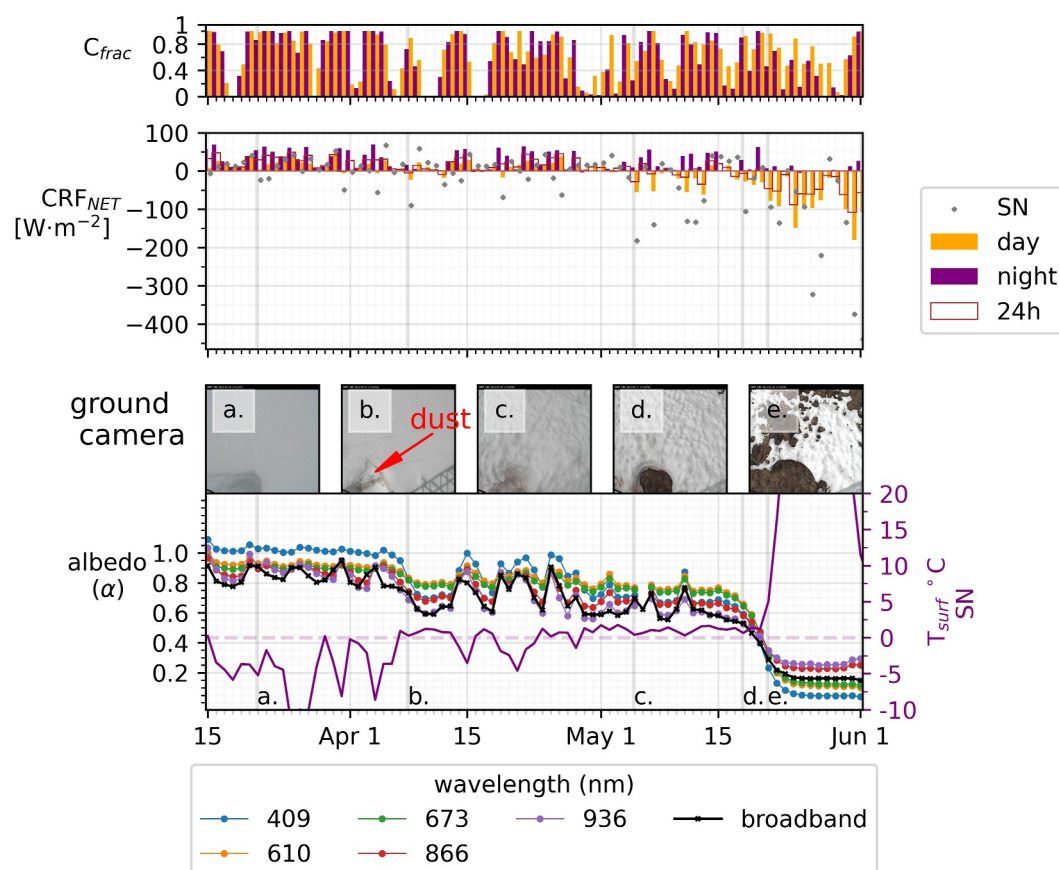


Figure 10. The impact of dust-on-snow on albedo and CRF_{NET} . From top to bottom: daytime C_{frac} , nighttime C_{frac} , CRF_{NET} for solar noon (SN), day (yellow) and night (purple), camera images of the ground surface near the radiometer at times (a–e), and the spectral albedo of the surface in five discrete wavelengths and broadband shortwave (black line). The surface temperature (T_{surf}) at SN is also shown (purple line).

both depressing the surface albedo and changing the sign of CRF_{NET} (Figure 10). The spectral albedo of the snow surface is computed at 409, 610, 673, 866, and 936 nm wavelengths using observations from the SAIL multi-filter rotating shadowband radiometer (Hodges & Michalsky, 2016). The spectral albedo from each band is computed using the ratio of upwelling relative to downwelling terms near solar noon. On 7 April, the reflectance in the 409 wavelength declines relative to the other bands. Because dust preferentially absorbs visible light in the shorter blue and violet wavelengths relative to red and near-infrared, this spectral albedo change signifies the presence of dust on the snow surface (Painter et al., 2007), which is also visible from the webcam images.

At the same time, the afternoon snow surface temperature measured by a co-located infrared thermometer (Morris, 2018) reached 0°C indicating melt, and the broadband albedo dropped from approximately 0.8 to 0.6. The daily average CRF_{NET} switched sign from the previous day (positive) to negative, though the magnitude of CRF_{NET} was small. CRF_{NET} at solar noon, on the other hand, reached a value just above $-100 \text{ W}\cdot\text{m}^{-2}$, the most negative value of the year to-date. A clear-sky period followed, and then subsequent snowstorms on 15 April, 19–20 April, and 25 April refreshed the snow albedo. Dust on the snow surface reemerged on the 28th, after which the snowpack began consistently melting midday. Clouds had a significant daytime cooling effect on the melting snowpack during the first and second weeks of May. Finally, CRF_{NET} immediately decreased as snow melts out exposing the underlying soil and grass, concurrent with a dramatic rise in land surface temperature at midday.

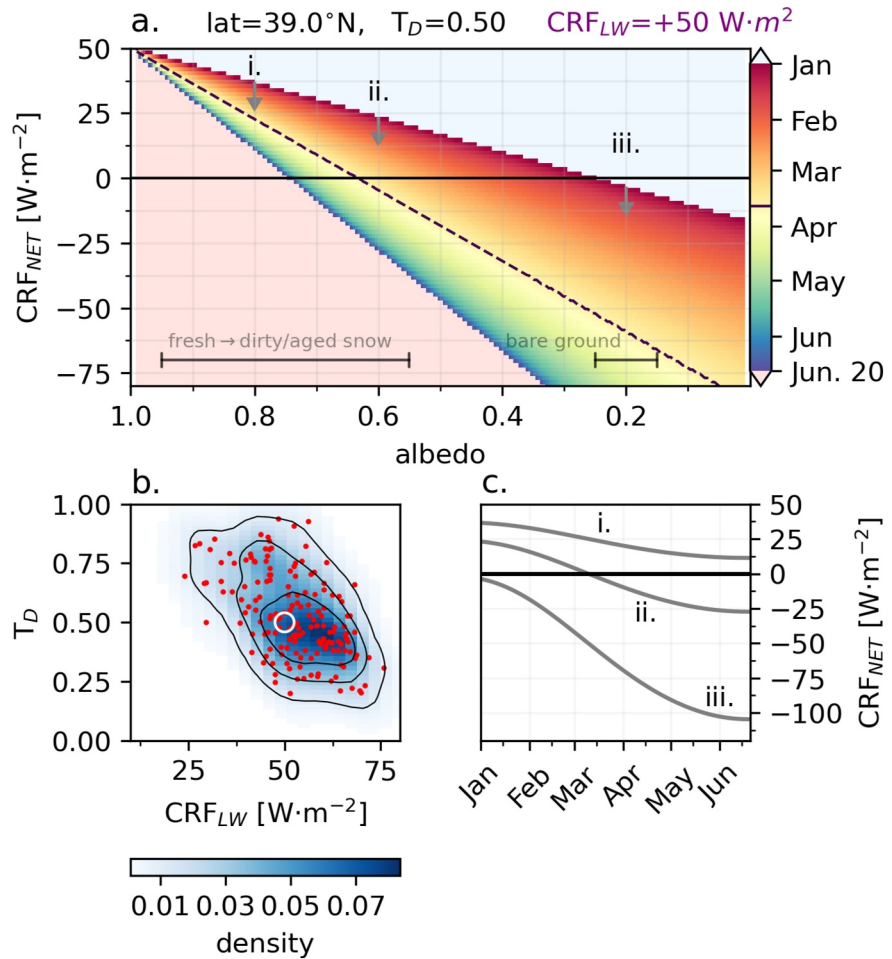


Figure 11. (a) An idealized model of daily average CRF_{NET} as a function of day of year between the winter solstice and summer solstice and albedo. The shortwave transmissivity (T_D) is 0.5, the latitude is $38.9^\circ N$, and CRF_{LW} is set to $+50 W \cdot m^{-2}$. The dashed diagonal line indicates the equinox date. (b) The relationship between T_D and CRF_{LW} based on SAIL observations for days where C_{frac} is >0.9 . Red dots are individual days, whereas color shading represents data density from a fitted gaussian kernel density estimate. The value of T_D and CRF_{LW} are highlighted with the open circle. (c) Examples CRF_{NET} for three different albedo values (0.8, 0.6, and 0.2).

3.5. Idealized Model of the Springtime CRF Sign Transition

Using the observation presented previously, we can construct an idealized model of the sign-transition of daily CRF_{NET} as a function of day of year and surface albedo and focus on the spring transition period. The idealized model follows Shupe and Intrieri (2004) and uses the relationship:

$$CRF_{SW}(k) = (1 - \alpha) S_k T_0 \overline{\cos[\theta]}_k (T_D - 1), \quad (5)$$

where α is the land surface albedo, T_0 is the bulk clear-sky transmissivity (the fraction of incident clear-sky radiation at the surface relative to the TOA incident radiation), S_k is the solar constant ($1,361 W \cdot m^{-2}$) adjusted by the time-varying Earth-Sun distance, T_D is bulk cloud layer transmissivity (the fraction of all-sky radiation relative to the clear-sky radiation at the surface) and the k subscript indicates a daily index. The average cosine of the solar zenith angle term ($\overline{\cos[\theta]}_k$) is purely a function of latitude and day of year. For illustrative purposes, we choose a fixed T_D of 0.5 and CRF_{LW} of $+50 W \cdot m^{-2}$, which are reasonable values based on the observed SAIL data (Figure 11b) for periods where the daily C_{frac} is greater than 0.9 (representing 25.4% of all observation days). T_0 is set at a constant 0.8 (acknowledging that this does change throughout the year; Figure

S1 in Supporting Information S1). We then vary the surface albedo from 0.0 to 1.0, and examine the resulting CRF_{NET} as a function of day of year Figure 11a.

The effect of a higher/lower CRF_{LW} parameter can be represented visually by translating the cone of possible CRF_{NET} values up/or down along the y-axis Figure 11a. The maximum shortwave effect is bounded by the summer solstice line (the daily average cosine of the solar zenith angle is maximized), and is likewise bounded by the minimum at the winter solstice. The effect of increasing the value T_D will widen the cone of possible CRF_{NET} values, and a given date-line will cross the zero line at a higher α value.

As an example, consider an albedo of 0.8 that is constant in time. This value will yield a CRF_{SW} never lower than $-50 \text{ W}\cdot\text{m}^{-2}$, even when the sun is at maximum strength, so a net cooling effect never occurs given a $+50 \text{ W}\cdot\text{m}^{-2}$ CRF_{LW} (line i. In Figure 11c). Conversely, a surface albedo of 0.6 yields a transition from warming to cooling that occurs in late mid March (line ii. In Figure 11c). From these relationships, we can also see that for the typical range or snow albedos (0.55–0.90), only a relatively weak CRF_{LW} is required to result in a positive CRF_{SW} regardless of the time of year. Again, an albedo of 0.8 would only require a CRF_{LW} of $25 \text{ W}\cdot\text{m}^{-2}$ (which is on the low end of the observations) to balance the shortwave effect on the equinox. And finally, even the weak early winter sun (Figure 11c) would yield a net cooling effect when the albedo is characteristic of bare ground (line iii. In Figure 11c).

Finally, and perhaps most importantly, we can see that an instantaneous reduction in albedo can flip not only the sign but also increase the magnitude of the radiative forcing of clouds in the spring. Observations show that the daily averaged CRF_{LW} is capped at $75 \text{ W}\cdot\text{m}^{-2}$, and rarely exceeds $100 \text{ W}\cdot\text{m}^{-2}$ even at the hourly timescale (Figure 11b; Figure S6 in Supporting Information S1) However, all else being equal, the counteracting CRF_{SW} forcing can be much larger in absolute value for low-albedo conditions (Figure 9). For example, with the parameters defined here, bare ground conditions on the equinox would yield a CRF_{NET} of less than $-50 \text{ W}\cdot\text{m}^{-2}$, but fresh snow covered conditions (albedo of 0.8) yield a CRF_{NET} of $+25 \text{ W}\cdot\text{m}^{-2}$. This represents a 100% increase in the absolute value of CRF_{NET} due to a change in surface albedo alone. This is a conservative value, since snow persists well past the equinox in this location (Figure 9a). At the same time, an instantaneous snow loss in the early winter can have the opposite effect, and decrease the magnitude of CRF_{NET} .

3.6. Climatic and Regional CRF_{NET} From CERES SYN1deg Ed.4a

To better contextualize the observations from the SAIL period into a climatic and regional context, we can view the SAIL based estimates of CRF alongside independent data from CERES SYN1deg Ed.4a (Figures 12a and 12b). We caution that this comparison is not intended as a validation of either CRF estimate, but rather an exercise to yield synergistic insights from both products. For comparison purposes, a locally estimated scatterplot smoother (LOESS; Cleveland (1979)) is applied to both SAIL and CERES data using the python “statsmodels” package (<https://github.com/statsmodels/statsmodels>; last accessed 4 Dec. 2024) with the “frac” parameter set to 0.035. The 1×1 degree grid cell nearest the SAIL site is selected for comparison.

The CERES climatological average CRF_{NET} and the smoothed SAIL period data are positive on the order of $20 \text{ W}\cdot\text{m}^{-2}$ during January and -40 to $-50 \text{ W}\cdot\text{m}^{-2}$ in May to August (Figure 12c). The sign transition occurs approximately 1 month earlier than the SAIL estimate. An examination of the CERES surface albedo (not shown) indicates much lower values than those observed by SAIL. This finding is consistent with larger grid sizes that include lower elevation, snow-free conditions and a mix of land and vegetation cover types.

The monsoon in July and August is also visible in the climatological average of CRF_{NET} and C_{frac} from the CERES data, and also the SAIL data to a limited extent. CRF_{NET} cooling declines along with the reductions in pre-monsoon C_{frac} in June, but increases again in July and August. The minima of the smoothed CRF_{NET} observed by SAIL occurs (Figures 9c and 12) very close to the climatological peak of the monsoon as revealed from the CERES record.

Finally, the CERES data show that, for its latitude, the high elevation and high albedo UCRB is somewhat unique in North America (Figure 12a). Clouds have a stronger warming effect at higher latitudes, which is likely due to both decreasing solar insolation with latitude and increasing frequency of high albedo snow surfaces. Areas south of 35°N seldom experience a net warming effect from clouds. For an equivalent latitude to the ERW, the fraction of the year where clouds have a net warming effect declines along with topographic elevation, even though C_{frac} stays the same or increases in both directions.

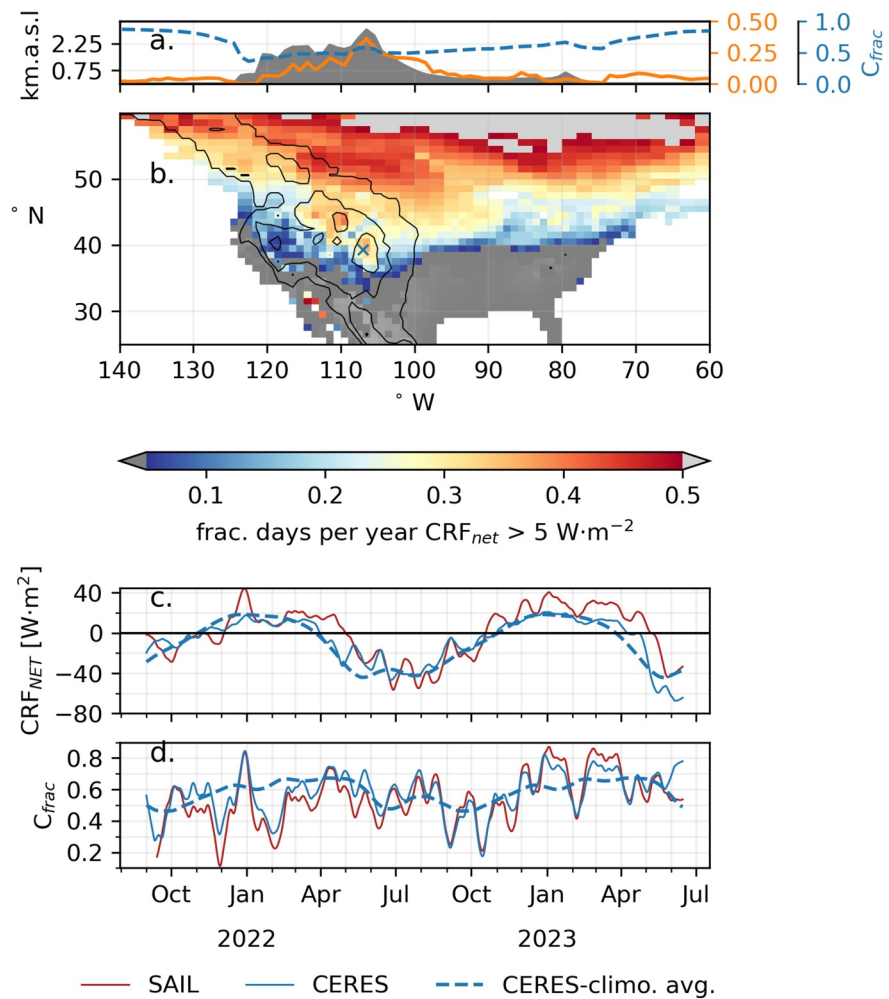


Figure 12. (a) The 39° N cross section of topographic elevation (km a.s.l.; gray shading), C_{frac} (blue line), and the fraction of the year where the daily CRF_{NET} is greater than $5 \text{ W}\cdot\text{m}^{-2}$ (orange line) for the CERES data record (2000–2023). (b) The same statistic as the latter in panel (a), but as a map-view of North America land area between 25 and 60°N. The 1, 2, and 2.5 km (a)s.l. elevation contours are depicted with black lines. The location of the East River Watershed is shown with the gray "x" marker. (c) SAIL (red) and CERES (blue) CRF_{NET} for the SAIL period smoothed by a LOESS filter. The dashed line indicates the climatological daily average CRF_{NET} from CERES. (d) The same as (c), but for C_{frac} .

4. Discussion

In this study, we quantified CRE and CRF for a single, intensively monitored, high elevation (2,880 m.a.s.l.) site in the headwaters of the UCRB (Figure 1). Although this site was relatively flat, nearby sloping mountain surfaces may have a much different surface radiation budget, given the slope-dependency of incident solar radiation and other longwave related terrain effects (Feldman et al., 2022; Lee et al., 2019; Manners et al., 2012). The radiometers were shadowed by terrain during the early morning and late evening (Figure S2 in Supporting Information S1). Although these times were omitted from the calculation of CRF/CRE, it is likely that clouds have a positive effect when the radiometer is shadowed, since clouds will increase the diffuse fraction of radiation relative to clear skies. We elected not to apply a correction for the radiometer sky view factor (Manners et al., 2012) to the clear-sky $LW\downarrow$ or diffuse component of $SW\downarrow$ radiation, but doing so may improve the biases in the clear sky model (Figure 2). More work is required to account for the full 3-D nature of radiation, and interactions with cloud cover including multiple scattering between clouds and the terrain surface, on the surface radiative balance in snow covered complex terrain.

The similarity of the seasonal sign of CRF_{NET} from the coarse $1 \times 1^\circ$ CERES SYN1deg Ed.4a across the UCRB suggests that SAIL results are not just specific to peculiarities of this particular location in the ERW, but that other

high elevation regions in the UCRB undergo a similar transition of clouds warming to cooling the surface. Furthermore, these results lend confidence to the latter's use applications in data sparse mountain regions (Hinkelman et al., 2015; Lapo et al., 2017).

4.1. Comparisons With Previous Studies

Relatively few studies have examined cloud radiative effects in mid latitude mountain environments, but those that have show results complementary to what is presented here. Aebe et al. (2017) similarly examined relationships between CRE_{LW} , CBH, and water vapor, but relied in part on radiative transfer calculations due to limitations of co-located observational data. They report CRE_{LW} values similar to what is presented here, and likewise found an inverse relationship between CRE_{LW} , PWV, and CBH. Marty et al. (2002) examined cloud radiative effects across a gradient of terrain elevations in the Swiss Alps, and computed seasonal CRF estimates using averages of observations during clear-sky and all-sky periods, as opposed to the radiative transfer modeling approach adopted in this study. They determined a net CRF of roughly $40 \text{ W} \cdot \text{m}^{-2}$ during the winter, $-50 \text{ W} \cdot \text{m}^{-2}$ during the summer, and between 15 and $20 \text{ W} \cdot \text{m}^{-2}$ as an annual mean for their site closest in altitude to SAIL. Although the seasonal sign is consistent, a combination of methodological and climatic differences likely explains the lower magnitudes estimated at SAIL.

The relationship between CRF_{NET} and the snow energy balance in the UCRB warrants additional investigation. It is well recognized that clouds enhance melt on the Greenland ice sheet (Ryan et al., 2022; Van Tricht et al., 2016) as well as some alpine glaciers (Conway et al., 2015), but the same is likely not true here. We show that, over melting snowpacks with an albedo of 0.5 – 0.6 , CRF_{NET} is commonly negative, meaning less R_{NET} is available for meltwater production. But clouds do increase the R_{NET} over high albedo snowpacks (0.8 – 0.9), even at solar noon during mid-winter (Figure 12c). More work, including modeling and information about the internal snowpack thermodynamic state, is required to fully interrogate the relationship between CRF and snowmelt processes. Although not explicitly studying CRF, Sedlar et al. (2024) quantified the importance of clouds on other meteorological states and terms of the SEB, as well as feedbacks between the lower atmosphere and the diurnal pattern of C_{frac} .

Moreover, the importance of dust on reducing snow albedo and accelerating ablation has been well articulated in the UCRB (Deems et al., 2013; Painter et al., 2010; Skiles et al., 2015). We also observe that dust reduces the snow albedo and likely contributes to the transition from CRF_{NET} warming to cooling in the spring of 2023 (Figure 10). Although the interaction between surface albedo and CRF_{NET} is not a new result (e.g., Ambach, 1974), we believe this is the first time the connection between the sign of CRF_{NET} and dust-on-snow processes have been articulated in the literature in the UCRB. Snow aging and grain-growth processes also contribute to declining snow albedo and have a characteristic spectral signature (Skiles et al., 2018), but we have not isolated these processes in the current study. However, snow aging processes are likely evident in the lower albedo values of the 2022 cold season compared to 2023 (Figure 12a). Further analyzing interactions between snow properties (grain size, impurities) the spectral character of snow albedo, and CRF, are left for future work.

4.2. CRF and Declining Snow Cover

This work has focused on the present-day characteristics of CRF in the UCRB, leading to the natural question of how/if CRF is related to ongoing drought in the UCRB (Section 1) and how CRF is projected to change in the future as the climate warms. This study has not attempted to answer these questions, though some insights can be gained. It is well established that snow cover is declining across the globe (Notarnicola, 2020; Siirila-Woodburn et al., 2021), and that snowmelt is projected to initiate earlier in the season and occur more slowly (Musselman et al., 2017). Our simple model (Section 3.5) suggests that in such a case, CRF_{NET} transitions from warming to cooling earlier in the year as snow cover diminishes, and also that the magnitude of CRF_{NET} would increase. The latter suggests that clouds may have an increasingly important role on the surface radiation budget in snow covered mountain areas, with potential implications for other terms of the SEB, such as energy available for evapotranspiration. However, this simple model does not take into account potential feedbacks between the surface and C_{frac} (e.g., Sedlar et al., 2024; Wallace & Minder, 2021), or other factors such as changing cloud microphysical characteristics and atmospheric environment. Consequently, the relationship between the timing of snow ablation and albedo, the diurnal character of C_{frac} (Figure 4), and large scale dynamic forcing (Figure 5)

warrants additional scrutiny at longer spatiotemporal horizons, particularly in the spring, when recent work has demonstrated robust trends of declining storm systems in the UCRB (Hogan & Lundquist, 2024).

5. Conclusions

We have shown, through the combined analysis of field observations and radiative transfer calculations, that the transition from clouds warming the surface (positive CRF, or increasing R_{NET} relative to clear skies) to clouds cooling the surface (negative CRF, or decreasing R_{NET} relative to clear skies) during the SAIL campaign (2021–2023) depends on the albedo of the land, with clouds generally warming surfaces covered by fresh snow and generally cooling surfaces covered by impure snow or bare ground. The relationship is true for both daily averages of CRF_{NET} as well as at solar noon when the sun is at maximum strength. At monthly scales, CRF_{NET} averages $22.3 \text{ W}\cdot\text{m}^{-2}$ during JFM when the background PWV is low (less than 0.5 cm) snow albedos are high (0.8–0.9), and nighttime clouds are common (Table 1). For a short window during snowmelt, when dust impurities contribute to broadband snow albedo reductions to ~ 0.6 , clouds have a net cooling effect, indicating that clouds inhibit melt for only a short period in the spring. Clouds yield an average net cooling effect of $32.2 \text{ W}\cdot\text{m}^{-2}$ during JAS, when shortwave insolation is high, the ground is snow-free (albedos of 0.2), higher cloud bases and a moister atmosphere reduce the longwave CRE (Section 3.2.1), and the North American monsoon is active.

Given these factors, the strong dependence on the surface albedo suggests a stronger and more persistent springtime cooling effect of clouds as snow cover diminishes across the UCRB (Section 3.5). More work is needed to better quantify the interactions of snow thermodynamics, snowmelt, snow aging, and dust-on-snow processes, and the sign and magnitude of CRF in the UCRB and mid latitude snow climates in general.

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Data Availability Statement

All of the SAIL data sets used in this study are publicly available from the ARM data discover page (<https://adc.arm.gov/>; date last accessed January 2025), including the QCRAD (Zhang, 2021), ARSCL (Johnson et al., 2021), intersonde (Jensen et al., 2021), TSI (Flynn & Morris, 2021), ground facing infrared thermometer (Shi et al., 2021), multi-filter rotating shadowband radiometer (Hodges et al., 2021), and surface meteorological data (Kyröuac et al., 2021). We used the RRTMG-LW v4.85 and RRTMG-SW v4.0. We use the “climlab” python model wrapper to run the radiative transfer models (Rose et al., 2022). The CERES data were downloaded from <https://ceres-tool.larc.nasa.gov/ord-tool/jsp/SYN1degEd41Selection.jsp> (date last accessed July 2024). The codes used to produce the figures and analysis in this study are available on GitHub (Rudisill, 2025). ERA5 geopotential height and wind data are available from the Copernicus Climate Data Store (<https://cds.climate.copernicus.eu/>; date last accessed August 2024), and the anomalies were prepared using data from Lee et al. (2023).

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