Diagnosing the Impacts of Northern Hemisphere Surface Albedo Biases on Simulated Climate[Ⓐ]

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ABSTRACT

Many Earth system models contain substantial biases in the magnitude and seasonal cycle of the albedo of snow-covered surfaces. Various structural and parametric deficiencies have been identified as potential causes of these albedo biases, related to vegetation distribution and abundance, snow albedo, and the representation of snow interception by forest canopies. There is, however, little understanding of how the albedo biases directly influence simulated climate because of difficulties in isolating them from other complex processes and feedbacks. In this study, we conduct a number of novel simulations using the National Center for Atmospheric Research Community Earth System Model (CESM), replacing the model's internal surface albedo calculation with values prescribed from observations or from other model simulations. Results show that while biases in surface albedo are largest in winter, those during spring have the greatest impact on surface climate because incoming solar radiation is much stronger. Correcting biases in the seasonal cycle of albedo in CESM reduces climatological temperature biases across the boreal region in spring and partially corrects Arctic sea level pressure biases, but due to compensating errors, overall climate biases are not always reduced. Additionally, we impose albedo patterns extracted from other climate models with large positive and negative albedo biases to illustrate the climate responses that can result. Prescribed surface albedo produces significant impacts on surface radiation, near-surface land temperatures, and, more rarely, atmospheric circulation. This is important because small changes to mean climate during spring can have major implications for the snow and surface radiation regimes.

1. Introduction

Surface albedo is an important characteristic of Earth's energy balance and therefore a key parameter in global climate models (GCMs). Much of the seasonal variability in Northern Hemisphere (NH) land surface albedo is due to changes in the presence of snow cover

(Roesch 2006). Its influence is exerted across much of the mid-high latitudes and in areas of high elevation, covering approximately 40% of NH land during late winter (Robinson and Frei 2000). The high albedo of snow acts to cool the climate by reducing incident radiation absorbed at the surface (Flanner et al. 2011), while the low thermal conductivity of snow cover insulates soil temperatures (Gong et al. 2004; Zhang 2005). Furthermore, Dutra et al. (2011) show that interannual variability in snow cover explains much of the variability in winter near-surface temperatures across snow-covered regions. Anomalous snow cover can also indirectly influence large-scale atmospheric circulation and NH winter climate (Saito and Cohen 2003; Gong et al. 2004; Fletcher

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et al. 2009; Allen and Zender 2011; Cohen et al. 2012). In extratropical land areas, surface albedo increases from its snow-free value (~ 0.08 – 0.15) in early fall to a maximum value in late winter (~ 0.3 – 0.8), followed by a rapid decline during spring melt (e.g., He et al. 2014). The peak snow-covered surface albedo over a region is strongly tied to its land cover, with lower values (more energy absorption) across dense forests and higher values (less energy absorption) in areas of low-lying or no vegetation (grasslands, croplands, tundra; Barlage et al. 2005).

Realistically simulating the seasonal evolution of snow-covered surface albedo has proven challenging for many climate models. The coarse nature of GCMs (~ 100 – 200 -km horizontal resolution) causes issues with the representation of topography, land cover, and snow distribution/properties, all of which are influential in determining surface albedo. Prior research has shown that several climate models have trouble with the magnitude and/or timing of seasonal changes in land surface albedo over the NH extratropics (Thackeray et al. 2014, 2015; Li et al. 2016; Wang et al. 2016). This albedo bias (relative to satellite-derived observations) is largely attributed to the oversimplified representation of key snow processes and vegetation, which drive errors in simulated snow-cover extent and snow-covered surface albedo.

One area with particularly large biases is the boreal evergreen forest, where nearly all models from phase 5 of the Coupled Model Intercomparison Project (CMIP5) overestimate peak winter albedo (Lorant et al. 2014; Thackeray et al. 2015). Simulating albedo can be difficult in this environment because of complex interactions between snow and forest cover, including the masking effect that the canopy has on underlying snow (Qu and Hall 2007; Essery 2013) and its role in intercepting snowfall (Thackeray et al. 2014; Bartlett and Verseghy 2015). Several models also exhibit substantial biases in their leaf area index (LAI), which is a key factor in determining snow-covered surface albedo (Lorant et al. 2014; Wang et al. 2016). Models with an LAI that is too low tend to have a snow-covered surface albedo that is biased high because of a larger canopy gap fraction (and vice versa). There is also a large spread in peak albedo over non-forested regions such as grasslands, shrublands, and tundra. In these landscapes, snow can completely cover most surface vegetation, resulting in very high peak albedo values (~ 0.6 – 0.8), which is better captured by CMIP5 models compared to the boreal forest.

Biases in peak albedo found in individual models create a knock-on effect whereby the simulation of albedo through the melt period is negatively impacted, with implications for snow albedo feedback (SAF). This is because SAF strength is largely controlled by the spring surface albedo contrast (snow-covered surface

albedo–snow-free albedo; Qu and Hall 2007; Fletcher et al. 2015). It is important to better understand model processes related to SAF (e.g., Thackeray et al. 2018) because it remains a significant source of uncertainty (40%–50% is attributable to SAF) in model projections of future spring warming over NH extratropical land (Qu and Hall 2014).

In this study, we directly perturb simulated surface albedo using climatological values derived from satellite observations and models to quantify the influence of albedo biases on climate. The land model [Community Land Model, version 4 (CLM4); Lawrence et al. 2011] has known issues related to the seasonality of albedo over the boreal forest (Thackeray et al. 2014). These issues are largely a result of how forest canopy hydrology is parameterized in the model. Notably, CLM4 removes all snow from the canopy when temperatures exceed the freezing point, which causes an early transition from a snow-covered to a snow-free environment. Model development to improve this issue in the latest version (CLM5) has resulted in an improved seasonal cycle of albedo.¹ Several other models also exhibit similar, although weaker, biases in the timing of albedo changes (Thackeray et al. 2015). Biases in albedo magnitude (rather than in seasonality) could be even more problematic for simulated climate. As a means of improving our understanding of these processes, we explore the importance of extreme CMIP5 albedo biases through a series of model experiments using the Community Earth System Model (CESM; of which CLM is the land component) including one that pushes CESM toward observed albedo.

Similar approaches to the one used here (section 2c) have been used to decouple the land surface from the atmosphere and investigate land–atmosphere interactions (Allen and Zender 2010, 2011). These studies predominantly prescribe soil moisture (e.g., Koster et al. 2000; Seneviratne et al. 2006) or snow-cover extent (Gong et al. 2002, 2004; Fletcher et al. 2007; Allen and Zender 2010; Sobolowski et al. 2010; Dutra et al. 2011). Many of the prescribed snow experiments attempt to identify linkages between fall snow cover and broad-scale atmospheric circulation patterns (such as the Arctic Oscillation) during winter (Gong et al. 2002; Fletcher et al. 2007, 2009; Orsolini and Kvamstø 2009; Allen and Zender 2011). In some cases, observed snow-cover extent was used to perturb the model (Orsolini and Kvamstø 2009; Douville 2010; Allen and Zender 2011), in which case one may expect results in an

¹ Available at http://www.cesm.ucar.edu/experiments/cesm2.0/land/diagnostics/clm_diag_PCKG.html.

improved albedo through a more realistic representation of the seasonal evolution of snow-cover extent. This does not, however, eliminate differences between simulated and observed snow-covered surface albedo. In particular, albedo biases driven by unrealistic vegetation masking, albedo parameters (e.g., fresh/old snow albedo), or prescribed land-cover classes are still present when simulated snow-cover extent is replaced with observations.

The primary goal of this work is to determine the influence of albedo biases on simulated Northern Hemisphere climate. This research addresses the question of whether correcting previously identified biases will move snow, temperature, and SLP simulations closer to, or further from, observations. We also establish the framework for performing similar simulations that will be run as a part of the ESM-SnowMIP project (Krinner et al. 2018), which seeks to better understand the variability in Earth system model simulations of snow. The data and methods are described in section 2. In section 3, we present the results from a series of uncoupled and coupled climate experiments. Last, section 4 highlights the key findings of this research and provides a discussion of how our findings relate to current literature.

2. Data and methods

a. Model description

The model used in this study is the CESM (Gent et al. 2011), version 1.0.4. The CESM configuration we use is composed of atmosphere [Community Atmosphere Model (CAM4); Neale et al. 2013] and land (CLM4; Lawrence et al. 2011) components, with specified sea surface temperature and sea ice. The horizontal resolution of the model is approximately 1° (0.9° latitude $\times$ 1.25° longitude), while the atmospheric vertical structure is made up of 26 levels. The land surface component is relatively sophisticated with detailed representations of key biogeophysical processes. Vegetation properties in the model are determined using the satellite phenology (SP) mode of CLM. The version of CESM used here includes CLM4, which simulates snow accumulation and melt processes along with compaction, aging, and the radiative impact of aerosols on snow. It has a multilevel snow scheme with up to five snowpack layers, depending on snow depth (Oleson et al. 2010). Because of the dominant role that snow cover plays in the seasonal cycle of NH surface albedo, it is important to have sophisticated snow processes. Prior research shows that issues with the representation of canopy snow in CLM (version 4.5 and earlier) cause surface albedo to decrease 2 months earlier than observed across forested regions (Thackeray et al. 2014).

b. Observational data

A blend of satellite-derived albedo products is used to perturb model simulations. The blended product (henceforth, OBSblend; fully described in section 2.2 of Thackeray et al. 2015) is an equally weighted mean of climatological monthly surface albedo from the Moderate Resolution Imaging Spectroradiometer (MODIS collection 5; Schaaf et al. 2002), the extended Advanced Very High Resolution Radiometer (AVHRR) Polar Pathfinder (APP-x) project (Wang and Key 2005), and the GlobAlbedo product (derived from MERIS, AATSR, and SPOT-VGT data; Muller 2013). The OBSblend product was previously limited by the spatial domain of the APP-x dataset, which extends between approximately 45° – 50° and 90° N (Thackeray et al. 2015), but here we also incorporate the albedo from MODIS outside of this region to obtain global coverage (including in the Southern Hemisphere). Since we are primarily interested in the Northern Hemisphere extratropics, OBSblend is a suitable choice. We find that daily albedo products have a high number of missing observations (primarily due to cloud cover obscuring the optical sensors), so monthly mean observational values are used to ensure complete coverage. The datasets are averaged over different temporal ranges (MODIS/GlobAlbedo: 2000–05; APP-x: 1982–2005), but their climatological values are not very sensitive to the time frame selected (Thackeray et al. 2015). Several other observation-based estimates are used to evaluate CESM's control climate. This includes near-surface air temperature and sea level pressure from the NCEP-II reanalysis (Kanamitsu et al. 2002) and snow water equivalent (SWE) from the Blended-5 dataset (described by Mudryk et al. 2015), which is a combination of five daily gridded datasets with consistent temporal and spatial coverage.

c. Experimental design

We conduct a novel set of model experiments to determine the impact of simulated albedo biases on climate. The experimental design overrides the land model's internal calculation of albedo and replaces it with albedo data prescribed from OBSblend, from another climate model, or from a different configuration of the same climate model (see below). By explicitly correcting the model biases using prescribed satellite data, we can directly relate the change in a model output variable (e.g., surface air temperature) relative to the uncorrected model to the albedo perturbation. Daily albedo perturbations are derived using linear interpolation between monthly values, thus leading to a smoothed seasonal cycle (e.g., gradual increase from October through November). The override procedure is

TABLE 1. Summary of CLM-CAM experiments. The MAM mean albedo over land areas poleward of 45°N is shown for each experiment. Note that the lowering of albedo over grasslands and tundra dominates the increased boreal albedo in OBSblend. The 2-m air temperature change (in relation to CTRL) for the same region is also shown for winter and spring. Temperature responses from CLM-OFF are much weaker: mean changes of -0.1 K for MAM and 0 K for DJF in OBSblend and -0.2 K for MAM and -0.1 K for DJF in HIGH.

Experiment	Albedo data	MAM albedo	MAM TAS change (K)	DJF TAS change (K)
CTRL	CLM-OFF CTRL	0.39	—	—
OBSblend	OBSblend	0.38	-0.4 ± 0.3	0.1 ± 0.4
HIGH	MIROC5	0.45	-1.8 ± 0.4	-0.8 ± 0.5
LOW	IPSL-CM5A-MR	0.33	-0.5 ± 0.3	0.0 ± 0.4

analogous to imposing prescribed sea surface temperatures (SSTs) to the atmospheric model, with the prescribed albedo value being passed to the land surface radiation code. Changes to albedo are implemented before other modules call this variable (e.g., for surface radiation calculations; see Text S1 in the online supplemental material for further details on the experimental setup). The impact of the surface albedo perturbation is largest over NH extratropical land (poleward of 30°N) because albedo biases in CESM outside of snow-covered regions are relatively small (not shown).

We perform sensitivity experiments using two configurations of CLM4: one in offline mode, driven by prescribed atmospheric forcing (temperature, precipitation, solar radiation, wind, pressure, and humidity) derived from reanalysis data (hereafter CLM-OFF), and one where CLM4 is coupled to the interactive atmospheric model CAM4 with prescribed sea ice conditions (thickness and area) and ocean surface temperatures (hereafter CLM-CAM).

Two control experiments (CTRL) of CLM-OFF that include the default internal albedo calculation from CLM4 are performed with different atmospheric forcing datasets. One experiment is forced by the Qian et al. (2006) dataset for the period of 1999–2004, while the other uses CRUNCEP (version 7) data from 1982 to 2012 (Viovy 2018). These two datasets differ in the long-term monthly time series of various climate records that are used in combination with short-term variations in NCEP reanalysis. This approach allows the impact of discrepancies in the forcing to be examined (see section 3a), an important step because the choice of forcing data is capable of dominating the response generated by land models (Ménard et al. 2015). Two other experiments (1999–2004) are then performed with CLM-OFF (driven by the Qian dataset; reasoning described in section 3a), each forced by a different perturbation to the surface albedo. The OBSblend experiment uses prescribed surface albedo derived from satellite observations to determine the effect of correcting biases in the default albedo calculation of CLM4 (recall from section 2a that CLM4 albedo is biased in the timing of the

seasonal cycle). The HIGH experiment seeks to determine the impact of large positive biases in the magnitude of snow-covered surface albedo by perturbing CLM4 with prescribed climatological monthly mean albedo (1980–2005) from the Model for Interdisciplinary Research on Climate, version 5 (MIROC5; Watanabe et al. 2010), which is one of the CMIP5 models with the highest peak annual mean surface albedo (Thackeray et al. 2015; Li et al. 2016). However, this is more complex than increasing albedo everywhere at all times, as MIROC5 simulates a slower-than-observed snow onset and in fact is one of the slowest in the CMIP5 ensemble [see Thackeray et al. (2015), their Fig. 2]. Therefore, this simulation explores the importance of winter/spring high bias in albedo, rather than during the early snow season. The albedo bias in MIROC5 is predominantly caused by structural errors in vegetation type and distribution (Lorant et al. 2014; Li et al. 2016; Wang et al. 2016), although some of it may arise from biases in simulated snow cover (not shown; snow cover is not perturbed in this experiment).

The list of experiments performed with CLM-CAM is shown in Table 1. A CTRL experiment of CLM-CAM uses prescribed climatological (1999–2004) albedo output from the Qian-forced CLM-OFF CTRL. Next, the OBSblend and HIGH experiments, described above, are repeated with CLM-CAM. In addition, the LOW experiment seeks to determine the impact of large negative biases in the magnitude of simulated snow-covered surface albedo by prescribing climatological (1980–2010) albedo output from the Institut Pierre Simon Laplace Climate Model version 5 with medium-resolution (IPSL-CM5A-MR) historical simulations (ensemble mean of three available members), which is one of the CMIP5 models with a very low annual mean peak surface albedo (Thackeray et al. 2015). In contrast, this model exhibits higher-than-observed snow-free surface albedo, which may act to balance the low bias where snow is present. The strength of each albedo perturbation is measured by the percent change in land albedo over areas poleward of 45°N (NH45) relative to the MAM mean in CTRL (average albedo shown in

Table 1). The spring albedo in OBSblend is 3% lower than CTRL, while much larger-amplitude perturbations are applied in HIGH (+15%) and LOW (−15%). All CLM-CAM simulations begin on 1 January and are integrated for 20 years and 1 month; the first month is discarded to account for spinup effects. The ocean and sea ice conditions are prescribed in a similar manner to the Atmosphere Model Intercomparison Project (AMIP; Gates et al. 1999). The ocean state uses climatological (1982–2001) annually repeating SSTs from the Hadley Centre Sea Ice and Sea Surface Temperature dataset (HadISST; Rayner et al. 2003). Because the atmosphere is free to interact with the land surface in CLM-CAM simulations, the surface response to albedo perturbations may be either amplified or damped by atmospheric feedbacks. However, it should be noted that using prescribed SSTs tends to damp the atmospheric circulation response (e.g., Liu and Wu 2004), and so our approach may still underestimate any potential circulation effects of changing surface albedo. Internal variability is even more damped in CLM-OFF than in CLM-CAM because the atmospheric, oceanic, and albedo conditions are all prescribed. Each of the 20 years in our CLM-CAM experiments represents an independent realization of an experiment, because the different years all have nearly identical boundary forcing (although soil temperatures do show minor drifts of $\pm 0.2^{\circ}\text{C decade}^{-1}$) and differ primarily in their atmospheric initial conditions. The use of seasonal means throughout the paper also increases the signal-to-noise ratio.

3. Quantifying the climate impact of surface albedo biases

a. Biases in CLM: Offline model simulations (CLM-OFF)

First, we briefly describe some aspects of the mean simulated CTRL climate in CLM-OFF that will be helpful for interpreting the results of later experiments. CLM-OFF slightly overestimates albedo at high latitudes in fall and winter (SON and DJF) and underestimates spring albedo across the boreal forest region, largely due to the timing of canopy snowmelt (Thackeray et al. 2015). Annual mean climatological near-surface air temperature from CLM-OFF with Qian forcing has a significant cold bias at high latitudes ($\sim 2^{\circ}\text{--}8^{\circ}\text{C}$ locally), with more pronounced biases in the winter and spring (Figs. 1a,c; note the differing scales). This cold bias is also a feature of the CRUNCEP-forced CLM-OFF simulation, which suggests that model processes may be responsible.

Next, we compare SWE from simulations with the two different forcing datasets (CRUNCEP and Qian). As shown in Fig. 2a, there are large regional biases

($>100\text{ mm}$) in climatological spring SWE relative to the Blended-5 product of Mudryk et al. (2015) when CRUNCEP forcing is used. CLM-OFF accumulates too much snow across midlatitude and northeastern North America and western Eurasia and features a major lack of snow across Alaska/Yukon and eastern Siberia. The low bias over Alaska/Yukon is due to regional underestimation of cold season precipitation in the CRUNCEP forcing dataset (Viovy 2018), and this has a significant impact on simulated albedo locally (not shown). Uncertainty in Arctic precipitation is a common issue due to sparse meteorological stations at high latitudes (Brown 2000). On the other hand, the Qian-forced (Qian et al. 2006) CTRL simulation shows a modest reduction in SWE biases across the northern extratropics, although the spatial pattern of biases is similar to those in the CTRL simulation using CRUNCEP forcing (cf. Figs. 2a,b). All subsequent analysis focuses on the Qian-driven simulations, which mitigates the impacts of SWE biases on albedo. Our aim is to determine whether correcting the albedo biases in CLM reduces the temperature and snow biases.

The upper panels of Fig. 3 show the magnitude of the perturbation to surface albedo in the OBSblend experiment: albedo is reduced by 0.05–0.25 in winter (DJF) over much of the extratropics and is increased by 0.05–0.15 in spring (MAM) across the boreal forest region. Although the forced changes to albedo are largest in DJF (Fig. 3b), they have little impact on simulated wintertime near-surface air temperature (Fig. 3e) because incoming solar radiation is very low. Springtime is expected to be the period when biases in surface albedo have the largest impact on climate because of the combination of widespread snow cover and higher insolation (Groisman et al. 1994; Qu and Hall 2014; Thackeray and Fletcher 2016). In OBSblend, the percent change in albedo averaged over NH45 during March–May (MAM) is slightly negative (−3%) because the lowering of albedo over grasslands and tundra outweighs the increased albedo in the boreal region due to vegetation masking the snow surface (Fig. 3c). Yet, correcting the albedo biases drives widespread springtime cooling that locally ranges from −0.1 to −0.5 K (Fig. 3f), but with little change in spring snow cover (not shown). This implies that the increased albedo over the boreal latitudes has more radiative impact than more poleward albedo decreases, where insolation is weaker.

The CLM-OFF results reveal relatively weak near-surface air temperature biases associated with the internal albedo biases in CESM. Similarly, the larger albedo perturbation from the CLM-OFF HIGH experiment shows more widespread, but small-magnitude, temperature responses (Fig. S1). However, the magnitude of near-surface

FIG. 1. (a),(b) Difference in climatological mean winter (DJF) and (c),(d) spring (MAM) 2-m air temperature (shading; K) between the CLM CTRL simulations and NCEP-II reanalysis over land. The CLM-OFF comparisons are for 1999–2004 (Qian forcing), while the CLM-CAM comparisons are for 1982–2001. Climatological SLP bias for CLM-CAM is shown by contours with an interval of 2 hPa, and negative contours are dashed. Note that the scale is doubled in (a).

air temperature changes is highly constrained in the offline configuration of CLM4 because although land surface temperature is prognostic, *atmospheric* temperatures are prescribed. The offline configuration also prevents any modulation of the response due to land–atmosphere interaction, and therefore, it is natural to ask whether this response will become weaker or stronger when coupling between the land and atmosphere is enabled.

b. Biases in CESM: The role of land–atmosphere interaction (CLM-CAM)

We first briefly examine the mean climate of the CTRL simulation with CLM-CAM (i.e., the configuration of CLM4 that is coupled to the atmospheric model

and forced with climatological albedo from CLM-OFF). This model exhibits biases in near-surface air temperature with DJF cold biases of between 2° and 8°C across northern Siberia, Arctic Canada/Alaska, and the central Arctic basin (Fig. 1b). Consistent with other studies using the same atmospheric model (NCAR-CAM4; Neale et al. 2013), CLM-CAM shows notable biases in its large-scale atmospheric circulation patterns over high latitudes in winter (Xie et al. 2012; De Boer et al. 2012; Grodsky et al. 2012). Figure 1b shows a pronounced negative bias in sea level pressure (SLP) over the polar cap and positive biases in the midlatitudes, indicative of a positive Arctic Oscillation (AO) pattern (Thompson and Wallace 2001). Weakened SLP over the polar cap is also a

FIG. 2. Bias in simulated climatological March SWE (units: mm). Two simulations of CLM-OFF with different forcing: (a) CRUNCEPv7 and (b) Qian. Bias is calculated with regards to the Blended-5 dataset (Mudryk et al. 2015).

feature of all six realizations of the 20-yr AMIP experiment performed using the CESM-CAM4 model,² indicating that this behavior is consistent in NCAR-CAM4. During spring, there are substantial warm biases across the mid–high latitudes along with weakened SLP over the pole (Fig. 1d). Biases of both signs in SWE over land remain large when compared to the Blended-5 dataset (up to 100 mm; not shown). Larger biases in snow are somewhat expected, because unlike in the CLM-OFF simulations, precipitation in CLM-CAM is prognostic based on physical parameterizations. A more detailed description of the global climate of CAM4 is given by Neale et al. (2013).

Figure 4 presents the near-surface air temperature and radiative and nonradiative energy flux responses for the OBSblend experiment with reference to CTRL. A dipole temperature response of up to $|2\text{ K}|$ is seen over North America and Eurasia in winter (NH45 land average is 0.1 K ; Table 1), with warming over central Asia and northwestern North America and cooling over western Asia and Scandinavia (Fig. 4a). The dominant response in spring is cooling of $1\text{--}2.5\text{ K}$ across the boreal region (Fig. 4b; NH45 land average is -0.4 K). This signals a 4–5 time amplification of the CLM-OFF response to the same perturbation (in reference to the zonal mean of TAS over $45^{\circ}\text{--}60^{\circ}\text{N}$ land). This response also seems to reduce the climatological warm bias in CLM-CAM across boreal latitudes (Fig. 1d; Fig. S3).

Diverse factors appear to be driving the near-surface air temperature responses in the two seasons. In winter, warming in central Eurasia is driven by increased solar

heating (shortwave tends to dominate the net radiative term), but the other features of the temperature response are not consistent with being radiatively driven (Fig. 4c). This is presumably a direct result of the lack of available insolation at high latitudes to translate the surface albedo perturbation to a response in the surface energy budget. We return later to the (nonradiative) causes of the cooling response over Scandinavia. In spring, the available insolation is much larger, and the temperature response (Fig. 4b) appears to be mostly radiatively driven. Cooling over the boreal forests of North America and Eurasia is associated with widespread decreases of $10\text{--}20\text{ W m}^{-2}$ in net radiation (Fig. 4d) and partially compensated by reductions in turbulent heat loss to the atmosphere (Fig. 4f).

The lack of a clear local radiative forcing in winter means that other processes are likely contributing to the temperature response across the continental high latitudes. There is some evidence to suggest that the temperature response in winter is generated through circulation changes that are driven by remote albedo-related forcing at lower latitudes. The largest radiative forcing in winter is situated over the Tibetan Plateau and adjacent areas (Fig. 4c), a region that is known to be a source of remote climate responses associated with interannual variability in snow cover (Liu et al. 2017). It should be noted that only a small portion of the overall atmospheric response in our experiments is statistically significant, but we explore the mechanisms here for potential impacts on surface temperature. The large-scale atmospheric circulation response in OBSblend during winter and spring is dominated by a weakening of the Azores high (significant at 95%) and Icelandic low (Figs. 5a,b) that projects onto the negative phase of the

² Available at <http://www.cesm.ucar.edu/experiments/cesm1.0/#amip>.

FIG. 3. Difference in (a)–(c) seasonal mean albedo and (d)–(f) near-surface air temperature between CLM-OFF OBSblend and CTRL experiments (using Qian forcing). Seasonal means for the fall (SON), winter (DJF), and spring (MAM) are shown.

North Atlantic Oscillation (NAO; Hurrell et al. 2003). This appears to reduce climatological SLP bias across the North Atlantic in CLM-CAM during winter (Fig. 1b; Fig. S3). To quantify this effect, we assess the difference in climatological mean DJF pressure between Lisbon and Reykjavik (23.4 hPa according to reanalysis) as a proxy for the NAO [similar to Visbeck et al. (2001); a stronger gradient occurs during positive NAO]. In OBSblend, the bias in this pressure gradient is reduced by 40% from CLM-CAM CTRL (CTRL bias: 7.8 hPa). We note that there is also a modest (>1.5 hPa) weakening of the Aleutian low, but this response, like much of the remaining circulation patterns, is not statistically significant.

In the context of these circulation changes, we can interpret certain aspects of the temperature responses in Figs. 4a and 4b as being dynamically driven. In both winter and spring, the region of cooling over

Scandinavia and northeastern Russia appears to be associated with increased advection of cold Arctic air by surface winds linked to the negative NAO pattern (Hurrell et al. 2003). Much of the warming signal in DJF across Alaska/western Canada is associated with a high-pressure ridge off the west coast of Canada that drives anomalous warming, predominantly during February (not shown). There is also a weak, but significant, poleward movement in the zonal mean zonal winds around 30° – 40° N over the North Atlantic sector (60° W– 40° E) during winter because of a reduced North Atlantic pressure gradient (Fig. 5c).

The mean cold season precipitation over northern extratropical land is largely unchanged on average despite these large-scale alterations to surface climate (not shown). However, increasing the albedo in spring over the boreal forest region (Fig. 3c) naturally reduces

FIG. 4. CLM-CAM response to OBSblend albedo forcing in seasonal mean (a),(b) 2-m air temperature (K), (c),(d) net radiation at the surface (positive downward; W m^{-2}), and (e), (f) sensible plus latent turbulent heat fluxes (positive upward; W m^{-2}). Seasonal means for the winter and spring are shown, while summer and fall are excluded because of less albedo perturbation. In (a),(b) stippling indicates regions of statistically significant temperature change ($p < 0.05$). Negative net radiation anomalies cool both the skin temperature (through shortwave and longwave) and near-surface air temperature (through longwave cooling), whereas a negative turbulent flux warms the skin temperature.

OBSblend - CTRL

FIG. 5. CLM-CAM climatological seasonal mean difference in (a), (b) SLP and (c), (d) zonal mean zonal wind across the North Atlantic (60°W–40°E) between the OBSblend and CTRL cases. Stippling indicates regions of statistical significance ($p < 0.05$). The climatological zonal mean wind strength from CTRL is shown by gray contours representing 5 m s^{-1} .

snowmelt by limiting shortwave radiation absorbed by the snowpack. This change, coupled with cooler near-surface air temperatures (Fig. 4b), leads to positive spring snow-cover fraction (SCF) anomalies (Fig. 6) that locally exceed +10%–20% for any given month (not shown). Prolonged snow-cover duration in turn influences the near-surface soil moisture during the following months (not shown), which can affect the surface energy budget and contribute to long-term atmospheric variability (Eltahir 1998; Liu 2003). Therefore, correcting surface albedo biases in CLM-CAM drives some large-scale atmospheric change and appears to reduce near-surface air temperature bias

across the boreal region in spring, while winter bias is amplified over Scandinavia and limited over northwestern North America.

c. Biases in other CMIP5 models

While the overall representation of surface albedo (magnitude and spatial pattern) in CLM-CAM is relatively good compared to other CMIP5 models, we use our prescribed albedo methodology with CLM-CAM to assess the possible influence of biases within the CMIP5 ensemble. CMIP5 models show substantial biases in snow-covered surface albedo, and in most models, this issue relates to the magnitude of albedo, rather than the

FIG. 6. Climatological monthly zonal mean spring snow-cover fraction in CLM-CAM OBSblend (blue lines), CTRL (red lines), and HIGH (gray lines). Snow cover is consistently greater in the OBSblend and HIGH cases than CTRL. Areas of permanent snow or ice cover are masked here (i.e., Greenland).

timing (as in CLM). This magnitude bias provides us with a further test of model sensitivity, and although we are unable to run these other models, some interesting results can be derived by forcing CLM-CAM with their albedo output (responses for CLM-OFF in this configuration are shown in Fig. S1, because they are weak due to the prescribed atmospheric forcing). First, we apply a strong perturbation (+15% NH45 MAM albedo relative to CTRL) in a high-magnitude experiment (HIGH) using the albedo pattern from MIROC5. The climatological mean albedo from MIROC5 is unrealistically high across most snow-covered latitudes in DJF and MAM (Figs. 7b,c), particularly in forested regions and over the Tibetan Plateau.

In the following, the HIGH experiment is compared to the OBSblend experiment, rather than CTRL (which is shown in Table 1), because we are interested in the impact of albedo biases relative to observations, not relative to the default (biased) internal albedo in CLM4. In the HIGH experiment—which represents a much stronger albedo perturbation than OBSblend

(Table 1)—significant anomalous near-surface cooling of 1–4 K is produced across the NH extratropics in winter (NH45 land average is -0.9 K) and spring (NH45 land average is -1.4 K; Figs. 8b,c). This equates to an average spring land cooling that is more than 4 times that in the OBSblend experiment (Table 1). The response in fall is relatively weak because the albedo perturbation in HIGH is primarily confined to the winter and spring (Fig. 7). Furthermore, the near-surface cooling is driven by wintertime albedo biases in the HIGH experiment that are large enough, and extend far enough south, that they induce a large reduction in the energy absorbed at the surface (more upward flux) that persists throughout winter and spring (not shown).

The enhanced surface cooling in HIGH is also associated with changes in the large-scale atmospheric circulation, as demonstrated by some significant responses in SLP over portions of the northern extratropics. During winter, the circulation response projects onto the positive phase of the NAO (Fig. 8e), with a poleward shift in the North Atlantic storm track ($\sim 60^\circ\text{N}$; Fig. 8h). Once again, winter changes are likely linked to remote albedo forcing, which is characterized by large decreases in absorbed shortwave at the surface during winter over the Tibetan Plateau (analogous to a high snow-cover year; Fig. S2). In spring, the response is dominated by a statistically significant deepening of the Aleutian low (Fig. 8f) and a significant weakening of the stratospheric polar vortex that extends down to the surface (Fig. 8i). These SLP changes are somewhat consistent with the findings of Liu et al. (2017), who show that anomalously high snow cover over the Tibetan Plateau can drive a deepening of the Aleutian low and SLP increases over the North Atlantic. As in OBSblend, the cold anomalies during spring in HIGH are also associated with extended snow-cover duration and large positive snow-cover anomalies (Fig. 6).

Last, we investigate the CLM-CAM response to prescribed low-albedo biases taken from the IPSL-CM5A-MR model (the LOW experiment; see Table 1), which exhibits a substantially darker surface than observations over most NH extratropical land when snow is present (Figs. 7e,f). The most notable exception is that the albedo across the Tibetan Plateau is *higher* in the LOW experiment than in OBSblend (Figs. 7d–f), likely as result of greater regional snow extent in the IPSL-CM5A-MR model (inferred from surface albedo; CMIP5 snow output is unavailable for the IPSL models). The albedo perturbation drives an increase in surface net shortwave radiation that is strongest during spring (6.2 W m^{-2}) and winter (2.2 W m^{-2}). By contrast, the response in fall is of opposite sign (a *decrease* of 3.5 W m^{-2}) because the snow-free land albedo in LOW is higher than observed

FIG. 7. Difference in seasonal mean albedo between the (a)–(c) HIGH and OBSblend experiments and (d)–(f) LOW and OBSblend experiments. Seasonal means for the fall (SON), winter (DJF), and spring (MAM) are shown.

(Fig. 7d). Consistent with the temporal variation in the albedo perturbation, LOW is cooler than OBSblend in fall, particularly over snow-free areas (Fig. 9a), while in winter there is strong warming over North America and cooling over Eurasia and a rather weak response everywhere in spring (Figs. 9b,c). The limited high-latitude warming in this experiment is somewhat surprising, given the darkened surface, but is likely a result of opposite-signed competing forces at lower latitudes (i.e., increased albedo over the Tibetan Plateau and snow-free areas). The strongest regional temperature response occurs over the northern Great Plains during winter, exceeding 3°C locally. This warming pattern is also accompanied by a significant deepening of the Aleutian low (Fig. 9e), which helps drive cool anomalies over Alaska and eastern Siberia. Other large-scale

differences include a strengthening Azores high and increased SLP over large portions of continental Eurasia (Fig. 9e). During other seasons, there is very little low-level atmospheric response of significance.

d. Isolating regional patterns

The experiments described so far change albedo over much of the NH extratropics, which makes it difficult to pinpoint which locations may be important in driving the generated responses. Here, we attempt to isolate the role that albedo biases across different regions may play in influencing our results. Figure 10 shows the monthly evolution of simulated surface albedo from each of the four CLM-CAM experiments over three distinct land-cover types (boreal forest, plains, and nonboreal tundra) and their temperature responses with respect to

FIG. 8. Response to high-magnitude albedo forcing in seasonal mean (a)–(c) 2-m air temperature (K), (d)–(f) SLP (hPa), and (g)–(i) NH zonal mean winds (m s^{-1}). Seasonal means for the fall, winter, and spring are shown, while summer is excluded because of less albedo perturbation. Stippling indicates regions of statistical significance ($p < 0.05$). The climatological zonal mean wind strength from OBSblend is shown by gray contours representing 5 m s^{-1} .

FIG. 9. As in Fig. 8, but for the LOW experiment relative to OBSblend.

OBSblend (black line). Peak albedo over the boreal forest in HIGH (MIROC5) is nearly double that of OBSblend and the other models. This coincides with much cooler near-surface air temperatures over the boreal in HIGH. On the other hand, early albedo

decreases in CLM-CAM (CTRL) result in significant spring warming across boreal regions compared to OBSblend. The mean regional temperature difference between these two extremes is 4 K. Across the plains, there is a similar temperature swing in spring (possibly

FIG. 10. Map of three distinct physiographic regions used here (boreal forest, plains, and nonboreal tundra) along with a regional breakdown of the monthly albedo and temperature responses. The temperature difference (bar plots) is with respect to the OBSblend experiment (black line). The nonboreal tundra and boreal forest regions are defined as by Thackeray et al. (2015), while the plains are composed of areas poleward of 40°N where grassland and cropland combine to exceed 75% coverage according to CESM land-cover data.

due to its close proximity to the boreal region), but much cooler conditions during the snow-free season in the LOW case. These cooler conditions are tied to a high bias in LOW's snow-free surface albedo. Finally, although there are very large albedo differences between experiments across the nonboreal tundra, the temperature response is relatively weak—likely a result of low insolation at these latitudes during most of the snow season.

For a more quantitative metric, we scatter the monthly albedo and temperature differences from Fig. 10 averaged across each of these regions and calculate the temperature change per unit of albedo forcing (Fig. S3). The differences are relative to OBSblend and include all months, meaning that we capture snow-covered and snow-free albedo biases. The average change across the three experiments is 0.11, 0.16, and 0.03 K per % over the boreal, plains, and tundra, respectively. This illustrates that albedo biases over the plains have the strongest influence on temperature,

while albedo biases over the tundra have the weakest link to temperature responses.

4. Discussion and conclusions

This study utilizes several novel climate model simulations, which prescribe surface albedo in an effort to better understand the potential impacts that Northern Hemisphere albedo biases have on simulated climate. We find that although extratropical winter albedo biases can be large, they have very little direct influence on climate because of low incoming radiation over most snow-covered areas. There is a suggestion that responses generated during winter are tied to remote influences from albedo forcing farther south (e.g., over the Tibetan Plateau). Remote atmospheric anomalies due to albedo forcing are consistent with prior work (e.g., Fletcher et al. 2009; Allen and Zender 2010; Liu et al. 2017). Alternatively, spring albedo biases are associated with significant local and remote NH climate responses in

near-surface air temperature, snow cover, and, less robustly, large-scale atmospheric circulation. Changes to temperature and snow are also greatly enhanced when land–atmosphere coupling is enabled (relative to CLM-OFF simulations).

Correcting the albedo in CESM toward satellite-derived values drives some significant large-scale changes in atmospheric circulation and appears to reduce near-surface air temperature bias across the boreal region in spring (Fig. S4). Winter temperature bias is slightly increased over Scandinavia and slightly reduced over northwestern North America (Fig. S4). Robust cooling is produced during spring across northern extratropical land (1–3 K; Fig. 4b), a feature that can be largely reproduced if we only apply albedo forcing across the boreal forest (Text S2). There is also a pattern reversal of the climatological biases in winter sea level pressure (Fig. 5a), reducing the model's tendency toward a positive NAO/AO (Fig. S4). We also find that correcting surface albedo has knock-on effects for land surface properties such as snow cover. Cold anomalies extend snow-cover duration and lead to greater snow-covered area during spring melt. This could partially explain why the representation of snow-cover extent is so variable during spring (Thackeray et al. 2016).

Other CMIP5 models feature more drastic albedo biases than CLM4 relative to the magnitude of observed snow-covered surface albedo. To demonstrate the potential importance of correctly representing albedo, we impose albedo patterns extracted from the CMIP5 models with the highest and lowest albedo biases. We find that these albedo perturbations can generate large climate impacts, raising the question of whether a significant portion of the climate difference between CMIP models may be due to their representation of land albedo. We find an approximately 4-K change in seasonal mean near-surface air temperature over the mid–high latitudes from the low forested albedo of CESM to the high albedo of MIROC5 (derived from the ensemble mean of five CMIP5 historical simulations). To put this into perspective, the CMIP5 ensemble has an inter-model spread in climatological (1980–2005) mean NH extratropical land spring temperature of nearly 5° (Fig. S5). Because spring temperatures are close to the freezing point, small changes in absolute temperature can have large impacts on the snow and surface radiation regimes. It is expected that correcting albedo biases will have a direct impact on seasonal temperature biases, but it may not be in the desired direction (toward observations). Therefore, models with substantial biases in snow-covered surface albedo are susceptible not only to modeling errors related to future climate tied to SAF strength, but also to mean seasonal climate.

In all experiments, much of the NH large-scale circulation response shows a lack of robustness due to large internal atmospheric variability, but we do find locally significant seasonal mean changes to SLP. In the HIGH experiment, there is significant strengthening of the Azores high during DJF and deepening of the Aleutian low during MAM (Figs. 8e,f). By contrast, the LOW experiment develops these same features during DJF (Fig. 9e). This result is slightly counterintuitive, because albedo perturbations of opposite sign are associated with similar SLP responses. We hypothesize that the winter SLP responses stem from the Tibetan Plateau and adjacent areas, where the albedo forcing is fairly consistent in the HIGH and LOW experiments during winter (Fig. S2). The proposed physical mechanism is that the albedo changes induce a near-surface temperature response that increases upward Rossby wave activity flux, which can alter stratospheric circulation, and these anomalies can then propagate down through the troposphere. Detailed investigation of these mechanisms in our experiments is beyond the scope of this study, but they do provide a possible explanation for some of the remote climate responses shown here. Alternatively, it is possible that these simulations are not long enough to fully separate forced and unforced climate variability when it comes to the SLP responses. Future studies should increase the length of similar experiments to further improve signal-to-noise ratios, although adding an additional 5 years to our experiments does not change the results. An additional limitation of this study is caused by the lack of visible and near-infrared streams of radiation from CMIP5 model output, which forces us to use the same albedo value for both components in our simulations (Text S1), when in reality they can be quite different over snow and vegetation.

Since we are interested in the sensitivity of the model to albedo biases, the brute force nature of our approach is effective. More elegant methodologies for incorporating observational constraints in models, for example, could involve data assimilation, but that was not necessary given the goals of this study. It should also be noted that the impact of albedo perturbations on climate is likely to vary in other CMIP5 models, but this is a useful first estimate from a widely used model.

The results shown here illustrate the surface climate impacts associated with simulated snow-covered surface albedo biases, which several recent studies have pointed out in the current generation of GCMs (Lorant et al. 2014; Thackeray et al. 2014, 2015; Li et al. 2016; Wang et al. 2016). We find that albedo biases across the NH extratropics from the CMIP5 ensemble can be influential for both local and remote climate features. This is consistent with prior studies linking anomalous snow

cover and snow albedo to contemporaneous and seasonally lagged climate features (e.g., Gong et al. 2004; Fletcher et al. 2007, 2009; Allen and Zender 2010; Liu et al. 2017). Moving forward, it is important for model development to reduce biases in snow-covered surface albedo for both its role in surface albedo feedback (Qu and Hall 2014; Fletcher et al. 2015) and the direct mean climate impacts discussed here. For some models, this means improving parameterizations of snow to reduce bias in snow cover (or canopy hydrology), while others may require reworking of how subgrid-scale lakes are treated (Verseghy et al. 2017) or changes to regional vegetation characteristics such as LAI or tree-cover fraction (Thackeray and Fletcher 2016). Interestingly, the results shown here suggest that correcting the albedo biases in CESM would partially fix spring temperature biases but may reinforce other existing temperature biases in winter. Therefore, making albedo more realistic in a model may not always have the desired effect on simulated temperature, but in this event, one should interpret this as an advancement for uncovering other model biases.

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