ABSTRACT: We explore the possible role of plant–atmosphere feedbacks in accelerating forest expansion using a simple example of forest establishment. We use an unconventional experimental design to simulate an initial forest establishment and the subsequent response of climate and nearby vegetation. We find that the forest’s existence produces favorable nearby growing-season conditions that would promote forest expansion. Specifically, we consider a hypothetical region of forest expansion in modern Alaska. We find that the forest acts as a source of heat and moisture for plants to the west, leading them to experience earlier springtime temperatures, snowmelt, and growth. Summertime cooling and cloud formation over the forest also drive a circulation change that reduces summertime cloud cover south of the forest, increasing solar radiation reaching plants there and driving warming. By isolating these vegetation–atmosphere interactions as the mechanisms of increased growth, we demonstrate the potential for forest expansion to be accelerated in a way that has not been highlighted before. These simulations illuminate two separate mechanisms that lead to increased plant growth nearby: 1) springtime heat advection and 2) summertime cloud feedbacks and circulation changes; both have implications for our understanding of past changes in forest cover and the predictability of biophysical impacts from afforestation projects and climate change–driven forest-cover changes. By examining these feedbacks, we seek to gain a more comprehensive understanding of past and potential future land–atmosphere interactions.

SIGNIFICANCE STATEMENT: This study investigates whether the emergence of a high-latitude forest could influence the way water and energy are exchanged between the land and atmosphere in a way that impacts nearby growing conditions and subsequent forest expansion. We use a computer model to simulate a climate with and without forest establishment in the high latitudes and test the response of plants surrounding the forest to the two different climates. We find that a forest is indeed able to spur neighboring plant growth by modifying regional climate and producing more favorable growing conditions for surrounding vegetation. Specifically, forest establishment can bring better growing conditions to plants adjacent to it by warming the air and altering nearby circulation and cloud cover.

KEYWORDS: Atmosphere-land interaction; Biosphere-atmosphere interaction; Cloud cover; Ecosystem effects; Paleoclimate

1. Introduction

Changes in high-latitude vegetation are known to have strong local and remote effects in the atmosphere (Bonan et al. 1992; Swann et al. 2010; Kim et al. 2020). Specifically, a change in vegetation can modify surface temperatures through both surface albedo (Bonan 2008) and atmospheric water vapor changes (Swann et al. 2010). Modeled high-latitude systems have been shown to be particularly sensitive to these two feedback mechanisms (Bonan et al. 1992). Paleoclimate simulations have repeatedly found that high-latitude vegetation responses to external forcing have the potential to be amplifying agents of a change in climate. Examples of vegetation feedbacks appear in the recent past resulting from anthropogenic forcing (Browkin et al. 1999; Snyder et al. 2004; Levis et al. 2000), during the mid-Holocene (Foley et al. 1994; Harrison et al. 1998; Texier et al. 1997), during the last glacial maximum (Kubatzki and Claussen 1998; Levis et al. 1999), and during interglacial time periods (Crucifix and Loutre 2002; Meisner et al. 2003; Kubatzki et al. 2000; de Noblet et al. 1996).

Vegetation influences its surrounding climate by modulating many physical properties of the land surface. Changes in a given land surface property stemming from a change in vegetation shift the balance of energy fluxes between the atmosphere and the land surface (Bonan 2008; Manabe 1969). Surface albedo modulates how much solar radiation is absorbed by the surface. Plant types differ (among themselves and from other land-cover types) in how they partition latent and sensible heat fluxes. Changing land surface roughness
alters the amount energy exchanged between the land and the atmosphere via turbulence, and plants determine evaporative resistance of the surface, which affects the release of energy as latent heat from the surface. Sensible and latent heat fluxes are important determinants of leaf and surface temperature and as a result alter net outgoing longwave radiation emitted from the surface (Lagué et al. 2019). Additionally, changes in surface vegetation are inherently coupled to the global climate system through the carbon cycle. In effect, changing surface vegetation induces changes in overall climate, and in particular, biome shifts alter the way energy and water are exchanged between the land surface and the atmosphere and are especially powerful in affecting local climate.

Lagué et al. (2019) quantify how atmospheric feedbacks amplify the temperature change that occurs due to a change in various surface properties, including albedo and evaporative resistance. They show that, while the atmosphere generally exhibits feedbacks everywhere, the complete climate impact of a change in any surface property will depend on the context of its regional climate—in particular, water availability. For example, in a water-limited region, a decrease in albedo might cause a warming effect, while in wet regions, it may lead to more energy released through evaporative cooling. In the high latitudes, they find that the decrease in albedo associated with switching from a grassland (albedo of 0.2 on average) to a needleleaf evergreen forest (0.1 on average) is able to produce a strong warming feedback that is primarily due to the regional importance of surface albedo and atmospheric water vapor feedbacks but is also due to nonlocal feedbacks.

Boreal vegetative changes that emerge from orbital forcing have been shown to compound local warming and by consequence induce regional climate changes would influence nearby growing conditions. The extent to which nearby plant species (and surface properties) could respond and accelerate forest expansion has not been quantified. In this paper, we investigate the potential for vegetation–atmosphere feedbacks to accelerate biome shifts by changing nearby growing conditions. In particular, we investigate how orbitally driven boreal forest establishment in the high latitudes drives accelerated forest expansion solely through surface–atmosphere interactions. By dissecting the land–atmosphere interactions at play locally that could lead to forest expansion, we hope to build a more comprehensive understanding of the land-cover change dynamics and ultimately set a clearer stage for these global and long-range dynamics.

Previous work has shown that 1) shifts in climate are able to shift biome distributions and that 2) past vegetation shifts may have been able to influence local climate. However, what is not known is whether the shift in vegetation could be self-perpetuating because of its influence on the atmosphere. Vegetation models that dynamically predict vegetation cover or communities are challenging to constrain with observations and the way vegetation cover and community composition are represented may not reflect actual mechanistic drivers (Fisher et al. 2015). Therefore, we devised an innovative experimental design to test whether the establishment of forest, motivated by an orbitally driven shift in climate, could by its own influence on the local atmosphere accelerate the expansion of forest nearby. We employ an experimental design that isolates the atmospheric influences on nearby growth that occur as a result of forest establishment. Quantifying this potential for a surface vegetation change to elicit downstream vegetation changes enriches our understanding of past climate and vegetation changes, broadens our knowledge of the Earth system as a whole, and could inform land-use management strategies.

The main goal of this paper is to isolate and examine the functionality of forest-driven acceleration of forest expansion. This allows us to conceive of and quantify the effect in isolation. Note that in this study we do not represent any specific forest establishment that occurred in the past but place our experimental forest in the high latitudes to place it in a context where surface changes have the potential to effect strong climate responses (Swann et al. 2010).

2. Methods

a. CESM

We perform a set of coupled experiments and subsequently a set of land-only experiments, all in the Community Earth System Model (CESM2.1.0) using the Community Atmosphere Model, version 6 (CAM6.0; Danabasoglu et al. 2020); the Community Land Model, version 5 (CLM5.0; Lawrence et al. 2019), a slab ocean model (Neale et al. 2010) with prescribed heat fluxes taken from a CESM2.1 preindustrial control (PI-Control) run, and an interactive sea ice model, the Los Alamos Sea Ice Model, version 5 (CICE5.0; Hunke et al. 2017). We run our simulations globally at a resolution of 1.9° latitude by 2.5° longitude.

1) LAND MODEL DETAILS

We use CLM5.0 to simulate both the forest’s influence on the climate (phase 1) and plant responses to the altered climate (phase 2). CLM5.0 represents major plant processes including “big leaf” photosynthesis, respiration, plant phenology, and nutrient cycling, all of which influence surface water and energy fluxes (Lawrence et al. 2019). Using this model is preferable to using a dynamic global vegetation model (DGVM), which would predict vegetation patterns and associated biophysical processes in response to climate but lack representation of the actual mechanisms controlling changes in plant cover and are challenging to constrain with observations (Fisher et al. 2015). Additionally, while DGVMs do explicitly represent more terrestrial plant processes, such as plant competition for sunlight and plant recruitment and survival based on statistical climate envelopes of modern biomes, our experimental design does not require dynamic vegetation to assess if forest expansion could have been accelerated. However, our design does include certain ecological assumptions that are explored in the discussion and conclusions (section 4).

Land model grid cells are composed of subgrid land units, which include vegetation, urban, glacier, lake, and cropland units. Within a vegetation land unit, the model simulates
vegetation properties according to the specified classification of plant type, or plant functional type (PFT). For both the control and experiment simulations, vegetation type is prescribed to be constant in time; while individual PFTs are free to respond to the climate forcing, for example, by growing more or fewer leaves, the percent of each grid cell covered with each PFT is fixed in time. The same atmospheric forcing is used to force all subgrid units within a land grid cell, though plant activity within each may increase or decrease. In a coupled simulation, surface variables and fluxes are passed to the atmosphere model as an average of the subgrid quantities weighted by their fractional areas within a land model grid cell, but in a land-only simulation, no information is passed from the land to the atmosphere.

In both the coupled and land-only runs, CLM5.0 was run with active biogeochemistry (BGC), which allows the model to simulate biogeochemical cycling (including carbon and nitrogen cycling) for each PFT. For our purposes, the primary importance of this setting is to allow leaf area to respond dynamically to changes in climate. Irrigation of crops was turned off.

Including bare ground, there are 16 possible PFTs in the model. Each uses one of three possible phenology algorithms: evergreen, for which a fraction of annual leaf growth persists for longer than one year; seasonal-deciduous, which only grows during part of the year, determined by growing degree-days (GDD); and stress-deciduous, which grows according to temperature and soil moisture conditions and may start and stop growing multiple times throughout the year (Lawrence et al. 2019).

Surface albedo for vegetation units in CLM is determined by the optical properties of the PFTs present. These properties dictate the flux of solar radiation (per watt received at the surface) reflected, transmitted, and absorbed by the components of the vegetated land surface [leaves, stems, or ground; see Sellers et al. (1986) and Lawrence et al. (2019)]. (Specific values for leaf-level albedo \( \alpha \) and total scattering \( \omega \) for visible wavelengths are given in Table A1 in the appendix.) In addition to their optical properties, each PFT in CLM has an associated canopy height that dictates interception, resultant snow cover and surface albedo, in addition to shading of trees through the canopy (Lawrence et al. 2019). There is no explicit modeling of lateral redistribution of snow in CLM5.0, but there is representation of snow layers as well as various distinct compaction processes (Lenaerts and van den Broeke 2012; van Kampenhout et al. 2017).

On average in the high latitudes, vegetated surfaces absorb more direct and indirect incident light than nonvegetated surfaces since vegetation is darker than snow. Snow albedo varies by snow type and contamination but is always higher than vegetation. Soil albedos increase with dryness and vary by soil type (which determines color in CLM). Depending on saturation and color, dry soil albedos can range from 0.08 to 0.36, and saturated soil albedos can range from 0.04 to 0.25 (Lawrence et al. 2019).

2) ATMOSPHERE AND OCEAN MODEL DETAILS

To simulate the atmospheric response to changes in vegetation, we use CAM6.0 [described in Bogenschutz et al. (2018) and Gettelman et al. (2019)]. In this study, we are concerned with understanding water and energy budget changes that take place because of atmospheric responses to a vegetation change. To that end, carbon dioxide (CO\(_2\)) and aerosol concentrations are not interactive in the coupled simulations and do not respond to land changes. CO\(_2\) is held constant at preindustrial levels (284.7 ppm), and aerosols are prescribed. Since we hold CO\(_2\) constant in both the control and experiment simulations, the specific value of CO\(_2\) concentration does not affect our results. As a result, changes in atmospheric reflectivity are solely due to changes in cloud formation and changes in water vapor. Cloud changes are due to dynamical changes induced by the vegetation changes we impose on the ground, including changes in vertical motions, temperature, and moisture.

In all simulations, we use a slab ocean model (SOM), which assumes ocean circulation does not evolve; fixed monthly heat fluxes are prescribed for each ocean grid cell, representing horizontal and vertical energy transport within the ocean and a specified mixed layer of the ocean. This allows sea surface temperatures (SSTs) to change and energy to be exchanged with the atmosphere. The role of oceans in propagating land surface change impacts on global climate has been previously demonstrated (e.g., Bonan et al. 1992; Davin and de Noblet 2010; Swann et al. 2012). By allowing ocean temperatures to respond to changes in energy fluxes from the atmosphere, SOMs allow atmospheric signals to propagate farther than fixed SST models but are much less computationally expensive than fully dynamic ocean models and prevent the propagation of climate signals driven by variability in ocean circulation. As such, the SOM provides a good choice for studying the impacts of changes in the land surface on atmospheric circulation.

Our experiments focus on the impact of a vegetation change in modern Alaska. CESM ranks among the top-performing models in representing present-day temperature and precipitation in this region (Walsh et al. 2018; Bhatt et al. 2021). While CAM6.0 generally performs better than CAM5.0, removing a winter cold bias, it maintains a cold bias in summer and a warm bias in winter in the Arctic stemming from yearlong overestimation of clouds (Baek et al. 2022). This bias would exist in both our control and experiment atmospheric runs, but it would not functionally affect the change between runs. Our use of low horizontal resolution does, however, introduce uncertainty in circulation response. See further discussion in section 4.

b. Experimental design

We allow the atmosphere, land, and sea ice models to run interactively at a global scale to simulate the atmospheric response to forest establishment in Beringia, which we refer to as the “coupled” runs [described in detail in section 2b(1) below]. We then use the modified atmospheric states produced in the coupled runs to drive a set of global land-only simulations to assess the productivity changes in neighboring vegetation to determine whether the forest interactions lead to more favorable growing conditions on the border of the forest [described in detail in section 2b(2) below]. We perform all experiments in CESM2.1 at a global scale to ascertain the dominant
large-scale vegetation–atmosphere interactions. A schematic of the two-phase experimental setup is shown in Fig. 1, and more details about each simulation are given in Table 1.

1) COUPLED LAND–ATMOSPHERE EXPERIMENTS: PHASE 1

To test the response of the local and regional climate to an idealized boreal forest establishment, we simulate the emergence of a region of 100% needleleaf evergreen boreal trees in present-day northwest Canada and eastern Alaska. Directly to the east of this region, we impose an ice sheet on the land surface (Fig. 1).

The region of forest establishment extends across all land units from 61.5° to 73°N and from 135° to 157°W. The ice sheet is roughly the same size as the imposed forest and extends from 61.5° to 73°N and from 115° to 135°W. Both regions (not including ocean grid cells) contain about 850,000 km² of land area. In the experiment simulation, 100% needleleaf evergreen boreal trees are imposed in the modified vegetation region, while the control experiment is initialized with bare ground in the same location; no other plant type is allowed to grow in this region in either the experiment or control simulation. We use bare ground as the control land surface so as to simply compare between nonvegetated and vegetated land, and, while the transition does not represent a specific past vegetation pattern, it allows us to compare atmospheric responses with and without plant feedbacks.

The ice sheet region is prescribed to have 100% glacier land-type coverage; the ice sheet appears in both experiment and control simulations, and is flattened in both in order to remove any climate responses to the orography of a melting ice sheet. Since all other land grid cells are held constant, the atmosphere responds to only the forest establishment land-cover change and no other associated ecosystem dynamics. This idealized experimental setup is intended to distinguish the atmospheric response to forest establishment from the atmospheric response without forest.

Additionally, to reduce limitation of plant growth by solar radiation, we set orbital forcing in both simulations to be consistent with 6000 years ago (simulated summer insolation hitting the prescribed vegetation area averages 443 W m⁻², which is about 6.7% above modern). Outside our prescribed vegetation in this region, 1850 default vegetation maps are used for both runs.

We run each coupled simulation for a total of 80 years, discarding the first 40 years as spinup, and analyze the last 40 years. In the 40-yr time period, short time scale plant carbon pools, including leaf area, come to equilibrium, though slower soil carbon pools, which are not included in our analysis, are not equilibrated. The land model does not consider time-evolving vegetation structure (i.e., plant heights and diameters are fixed in time) but does include dynamically time-varying leaf area. Therefore, the time scales of changes in ecosystem structure do not impact our simulations.

2) LAND-ONLY EXPERIMENTS: PHASE 2

In phase 2, we perform a set of experiments intended to assess each plant response separately to assess how high-latitude plant systems might respond to nearby forest establishment.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Atmosphere</th>
<th>Land</th>
</tr>
</thead>
<tbody>
<tr>
<td>Phase 1: Coupled simulations to drive modified atmosphere</td>
<td>CAM6</td>
<td>CLM5—bare ground patch + ice sheet patch; 1850 vegetation surrounding</td>
</tr>
<tr>
<td>Control</td>
<td>CAM6</td>
<td>CLM5—needleleaf evergreen boreal forest patch + ice sheet patch; 1850 vegetation surrounding</td>
</tr>
<tr>
<td>Experiment</td>
<td>Fixed atmosphere from coupled control simulation</td>
<td>CLM5—needleleaf deciduous boreal trees</td>
</tr>
<tr>
<td>Control</td>
<td>Fixed atmosphere from coupled experiment simulation</td>
<td>CLM5—broadleaf deciduous boreal trees</td>
</tr>
<tr>
<td>Experiment</td>
<td>Fixed atmosphere from coupled experiment simulation</td>
<td>CLM5—broadleaf deciduous boreal shrubs</td>
</tr>
<tr>
<td>—</td>
<td>—</td>
<td>CLM5—Arctic C3 grass</td>
</tr>
</tbody>
</table>

TABLE 1. Experimental setup for coupled (phase 1) and subsequent land-only (phase 2) experiments.
We expose each possible plant type to atmospheric conditions generated from phase 1, diagnostically assessing the potential for nearby plant growth. Because we do not know the exact type of plant that would grow there, as a way of assessing change in vegetation potential, we perform a set of land-only simulations in CLM5.0 where we prescribe all land (global) to be uniformly occupied by a single PFT and repeat the process for every possible plant type in the region (five different PFTs) (see Table 1). To find out how initial forest establishment impacts potential plant growth nearby, the land-only simulations are forced with the atmospheric state (last 40 years) from the coupled simulations, repeated twice in a row (for a total of 80 years of land-only simulations for each PFT). We run five land-only “control” simulations where plants see the climate state of bare ground nearby, in which atmospheric forcing data comes from the coupled (CLM-CAM) simulation with bare-ground land cover next to the ice sheet in the experimental region; and we run five land-only “experiment” simulations where plants see the climate state of a forest in which atmospheric forcing data from the coupled simulation with forest cover next to the ice sheet in the experimental region.

In each of the land-only simulations, one PFT is uniformly imposed over all land, replacing all land cover from the coupled experiments; this PFT uniformly replaces the forest/bare ground, the ice sheet, and all land units across all land areas globally. This experimental setup allows us to diagnose only the plant response and therefore isolate and attribute the climate impact of forest establishment on the growth of plants nearby. We focus our analysis on growth changes adjacent to the imposed forest as a way of understanding how a forest might accelerate expansion.

c. Analytical methods

1) ASSESSING EXPANSION POTENTIAL

In the land-only simulations, we quantify the plant responses to the two different climate states by examining differences in surface energy fluxes, surface temperature, and gross primary productivity (GPP). Plant responses in these simulations do not feed back on the atmosphere. Since the differences in atmospheric state are driven only by the establishment of the forest, changes to plant growth in the land-only experiments can be interpreted as being driven by the influence of forest establishment on the growth of plants nearby. We focus our analysis on growth changes adjacent to the imposed forest as a way of understanding how a forest might accelerate expansion.

The result is that, by comparing responses across plant types and identifying the atmospheric changes (driven by forest establishment) that led to those increases, we can reach a conclusion about how atmospheric changes affect the potential for growth of multiple plant types present in high-latitude ecosystems.

The means are calculated over 40-yr periods with an assumed 20 degrees of freedom (i.e., allowing for an autocorrelation of 2 years), but our results are relatively insensitive to small changes in length of time series (i.e., 30- and 50-yr trends yield similar results). Averages of land variables are weighted by gridcell land fraction and latitude (i.e., gridcell area).

2) ATTRIBUTING CLOUD CHANGES

We follow the example of Kim et al. (2020) and use relative humidity as a proxy for cloudiness to investigate differences in cloud formation between the two experiments, especially to the south of the established forest. Changes in moisture or temperature can independently lead to cloud responses, and changes due to each can be examined separately. We partition the difference between simulations into two terms: one from differences in temperature \( T \) and another from differences in specific humidity \( q \):

\[
\Delta RH_T = \frac{q_{ctl}}{qsat_{ctl}} - \frac{q_{exp}}{qsat_{ctl}} \quad \text{and} \quad (1)
\]

\[
\Delta RH_q = \frac{q_{exp}}{qsat_{ctl}} - \frac{q_{ctl}}{qsat_{ctl}}, \quad (2)
\]

where \( q_{ctl} \) and \( q_{exp} \) are the specific humidity and saturated specific humidity of the bare ground-run atmosphere (atm_ctl), and the \( q_{ctl} \) and \( q_{exp} \) are the specific humidity and saturated specific humidity of the forest-run atmosphere (atm_exp). Equation (1) estimates the expected magnitude and sign of the change in the relative humidity profile between the simulations given the change in atmospheric temperature alone, and Eq. (2) estimates the impact of the difference in specific humidity alone.
3) ATTRIBUTING HEATING

Plant growth, as measured by GPP, is strongly affected by temperature. We examine the source of changes in temperature using the thermodynamic energy equation:

\[ \frac{\partial T}{\partial t} = -u \frac{\partial T}{\partial x} - v \frac{\partial T}{\partial y} + \delta S_p + Q. \]  (3)

where \( Q \) is the diabatic heating rate and \( S_p \) is the static stability for an isobaric system:

\[ S_p = \frac{RT}{cp} - \frac{\partial T}{\partial p}. \]  (4)

As a simplified way to determine the source of the change in temperature between a control and experiment simulation, we decompose the change in temperature into its component parts: zonal temperature advection, meridional temperature advection, vertical thermal advection and associated adiabatic expansion and compression, and diabatic heating [terms 1–4, respectively, on the right side of Eq. (3)]. By comparing the change in each of these terms on the right-hand side of Eq. (3) between the two simulations, we can estimate the attributable source of an air parcel’s change in temperature.

3. Results

a. Overall, forest establishment drives nearby warming and productivity increases

GPP increases throughout the year when plants replace bare ground in the forest establishment region (Fig. 2). Across the box where forest is imposed, spatial heterogeneity in leaf area index (LAI) occurs due to the region’s terrain. At the peak of the growing season, the forest reaches a maximum monthly mean GPP of about 8 gC m\(^{-2}\) per day (Fig. 2). This biological activity, the incumbent surface property changes, and their combined effects are the drivers of a modified local atmosphere state. They determine the balance of surface–atmosphere water and energy fluxes. In the subsequent land-only experiments, across all five PFTs, climate modified by the forest’s influence drives increased GPP relative to the control inside the region where a forest was imposed in the coupled experiments—a location referred to as the Forest Box (green outline in Figs. 4 and 5, which are described in more detail below).

In addition to increased growth within the Forest Box, neighboring regions to the west and south of the forest saw higher GPP across all PFTs (Figs. 3 and 4). These two adjacent locations that experience increased growth will be referred to as the West and South Boxes (outlined in yellow and red, respectively, in Figs. 3 and 4).

In general at high latitudes, productivity is primarily limited by temperature and sunlight; GPP therefore increases with higher growing season temperatures or longer growing seasons (i.e., earlier springtime warming) (Nemani et al. 2003; Chapin et al. 2000). In both the West and the South Boxes, the largest increases in GPP magnitude take place during the springtime, with overall productivity peaking in June and the largest variability in mean growth occurring at the beginning and end of the growing season.

In the Forest Box, throughout the growing season, imposing a forest lowers surface albedo and increases solar radiation absorbed by the land surface. At the same time, increased cloud formation works to reduce incident solar radiation; that is, planting a forest decreases evaporative resistance and allows more energy to be released in the form of latent heat from the surface to the atmosphere (provided there is enough soil moisture), and this increase in latent heat flux causes low clouds to form and block incoming solar radiation from reaching the surface in the Forest Box. While both of these effects take place during both spring and summer, in the spring months [March–May (MAM)], the effect of lowering the land surface albedo is dominant (Fig. 3). Slightly less solar radiation reaches the land surface, but more of it is absorbed, and in net, more solar radiation is absorbed by the forest than the bare ground during the spring months (Fig. 5). By contrast, in the summer months [June–August (JJA)] the formation of shortwave-blocking clouds is dominant (Fig. 4); less solar radiation
FIG. 3. Changes in atmospheric conditions in the West Box: (a) Change in springtime (MAM) near-surface (reference height) air temperature. (b) Change in MAM near-surface (below 820 hPa) water vapor concentration. Arrows in (a) and (b) show near-surface wind (below 820 hPa). (c) Change in MAM fraction of ground covered by snow in land-only experiment run with broadleaf deciduous boreal shrubs (PFT 11). (d) Change in MAM land surface albedo. Nonsignificant changes have been masked out in (a)–(d). (e) Change in monthly mean GPP for all simulated PFTs in the West Box. Stars indicate significance using Student’s t test. (f) Change in May GPP in the West Box vs shortwave absorbed by vegetation for representative PFTs.
FIG. 4. Changes in atmospheric conditions in the South Box: (a) Change in summertime (JJA) near-surface (reference height) air temperature. (b) Change in JJA near-surface (below 820 hPa) water vapor concentration. Arrows in both (a) and (b) show near-surface wind (below 820 hPa). (c) Change in JJA low cloud cover (clouds occurring below 750 hPa). (d) Change in JJA incident shortwave radiation. Nonsignificant changes have been masked out in (a)–(d). (e) Change in seasonal cycle of monthly mean GPP for all simulated PFTs in the South Box. Stars indicate significance using Student’s t test. (f) Change in August GPP in the South Box vs shortwave absorbed by vegetation for needleleaf evergreen boreal trees (dark blue), needleleaf deciduous boreal trees (teal), and broadleaf deciduous boreal trees (green).
radiation reaches the surface, and in net, less solar radiation is absorbed by the forest than bare ground (Fig. 6). As a result, the Forest Box absorbs more energy during the spring and subsequently experiences a seasonal average surface warming of over 4 K (Fig. 3), which can be seen in the increase in longwave radiation leaving the surface (Fig. 7). The mirrored decrease in outgoing longwave during the summer (Fig. 7) is a result of a 3-K decrease in surface temperature due to cloud formation (Fig. 4).

Soil moisture changes were small in both the West and South Boxes relative to the surface albedo and low cloud changes. We do not see any changes in high-latitude transpiration stress due to changes in soil moisture. While the coupled experiments were run with a flat ice sheet imposed to the east of the forest, the ice sheet was removed for the land-only runs, and plants were prescribed in its place. However, no PFT consistently displayed an increase in GPP where the ice sheet had been in the coupled simulations; the atmosphere maintained temperatures that were too cold.

Although the increases in productivity in the South and West Boxes can both be attributed to higher temperatures, we propose that each of the locations has a unique warming mechanism, and that the impact of warming on productivity is predominantly occurring in the spring in the West Box, and in the summer in the

Fig. 5. Changes in growing conditions in the West Box during the growing season (March–August) for broadleaf deciduous boreal shrubs, the dominant plant type in this region: (a) Spring and (b) summer changes in solar radiation at the surface, with negative values indicating into the ground. Yellow bars show the negative change in downwelling solar radiation at the surface. If negative, more solar radiation reaches the surface in the experiment simulation run than in the control. Light-blue bars show the change in solar radiation reflected by snow. Green bars show the change in solar radiation reflected by snow-free ground. Orange bars show the total change in solar radiation absorbed by the surface (sum of the preceding bars); if negative, more net solar radiation is absorbed by the surface. (c) Spring and (d) summer changes in surface snow fraction (purple bars), percent changes in total latent heat flux released from the surface (blue bars), and changes in vegetation temperature (red bars). (e) Spring and (f) summer percent changes in exposed LAI (dark-green bars) and GPP (light-green bars). One asterisk indicates significance at the 90% confidence level, and two asterisks indicate significance at the 95% confidence level.
South Box. The atmospheric changes that drive GPP changes in the South Box are statistically significant at the 90% confidence level, while in the West Box they are significant at the 95% confidence level. We discuss the mechanisms in more detail below.

b. West Box: Forest drives plant growth through spring warming driven by heat and moisture advection and snowmelt

In the land-only experiments, plants in the West Box experience the strongest increase in growth during the spring. To explain this increase, we first discuss atmospheric changes driven by the forest in the coupled experiments by considering the changes in the Forest Box that drive regional change.

1) IMPOSED FOREST WARMS ATMOSPHERE OVER THE FOREST BOX IN COUPLED SIMULATIONS IN THE SPRING

In the spring, the dominant effect of replacing bare ground (where snow is exposed) with trees (where trees mask snow) is a lowering of land albedo; surface albedo in the Forest Box drops by a seasonal average of 0.4 (from 0.53 to 0.13), while top of atmosphere albedo increases by only 0.07 (from 0.43 to 0.50), leading to more overall shortwave absorbed by the surface when a forest is in place (Fig. 7). The land surface within the Forest Box absorbs an average of 42.5 W m\(^{-2}\) more total energy. This excess energy warms the land surface and near-surface atmosphere and is balanced by an increase in outgoing longwave radiation from the surface as well as increases in latent and sensible heat fluxes (see Fig. 7).

In addition to decreasing albedo, as discussed above, imposing a forest within the Forest Box decreases surface evaporative resistance as a result of increased leaf area and root water extraction (or plant transpiration), allowing more energy to be released in the form of latent heat as plants transpire (as evident in both spring and summer latent heat flux changes in Fig. 7). This increase in transpiration leads to higher water vapor in the atmosphere regionally (Fig. 3), consistent with prior work by Swann et al. (2010) and Kim et al. (2020). The resultant change (between bare ground and...
A forest) in total latent heat flux reflects the pattern of the growing season, peaking in June when plant transpiration hits a maximum with a 33.4 W m⁻² increase in latent heat flux from the surface to the atmosphere. Throughout the spring, the land covered in forest releases an average of 9 W m⁻² more latent heat than its bare ground counterpart. There is also a 10.4 W m⁻² increase in outgoing longwave radiation (red dashed), changes in latent heat flux (blue dotted), and changes in sensible heat flux (green dot-dashed); changes in net solar flux are repeated (gold solid).

Adding a forest changes the amount of incident solar radiation at the land surface in the coupled experiments. Subsequently, each plant type in the land-only experiments is exposed to this difference in shortwave reaching the surface and determines the net solar radiation absorbed by the land according to their unique properties (Table A1 in the appendix), but the response of the plants in the Forest Box in the land-only experiments does not then alter the atmosphere and thus does not feed back onto downward shortwave fluxes beyond what is driven in the coupled simulations. The forest-driven atmospheric changes remain the principal and independent driver of changes in nearby plants’ GPP.

In the land-only control simulations, in the West Box, the spatial patterns of snow depth and snow cover have north-south gradients, with higher amounts of snow at higher latitudes as well as increased snow amount and a longer snowy season over regions of elevated topography such as the Brooks Range. When a forest is established, the latitude where 50% of any grid cell is covered in snow shifts northward, causing surface albedo to drop for much of the West Box. The decrease in snow cover and snow depth is especially apparent for broadleaf deciduous shrubs. In the West Box, the neighboring forest establishment caused an additional 4.8 W m⁻² of springtime solar radiation to be absorbed by the land surface primarily by reducing surface albedo (Fig. 5). Atmospheric changes were responsible for blocking less than half as much solar radiation (MAM average of 1.08 W m⁻²).

In the West Box, we decompose the change in air temperature into its component parts (see section 2) and find that across the atmospheric profile the largest differences occur below 800 hPa (near the surface), with the largest (positive) changes between the two experiments coming from the zonal heat transport and to a lesser extent the diabatic heating, illustrating the role of heat and moisture flux from the forest westward over the West Box. The integrated change in zonal heat transport $\langle -u(\partial T/\partial x) \rangle$, between 1000 and 200 hPa (where the temperature increases) shows the largest positive change between the two experiments (Table 2). Here, angle brackets denote the vertical integral across the column such that

$$\langle A \rangle = \frac{-1}{g} \int_{p_{\text{surf}}}^{p_{\text{TOA}}} (A) dp.$$  

The snowmelt in the West Box is a result of near-surface warming, which comes from a number of sources. First, the land surface in the West Box is warmer due to the neighboring forest in the coupled experiments, causing snowmelt in the subsequent land-only experiments. A maximum heating rate of 0.25 K kg m⁻² s⁻¹ can be attributed to this process. Second, wind at low levels advects warm air from the Forest Box to the West Box (Fig. 3). Mean winds in the spring at the surface travel east to west over the Forest Box bringing warm, wet air to the West Box. Heating through this process reaches a maximum of 1 K kg m⁻² s⁻¹ near the surface (Fig. A2 in the appendix). Third, reduced cloud fraction in the springtime leads to an overall increase in net shortwave radiation absorbed by the surface in the West Box, contributing to overall warmer springtime temperatures (Fig. 3).
TABLE 2. The thermodynamic energy equation [Eq. (3)] can be rearranged to isolate temperature tendency \( \hat{\theta} T / \hat{t} \), which in equilibrium should approach zero. The terms on the right side of the equation balance differently in each simulation. Below are the column-integrated mean values for the various terms in the control simulation and their change when the forest is imposed (\( \Delta \)) for each region of interest in its respective season of interest. The terms that exhibit the largest positive change between simulations are in boldface type.

<table>
<thead>
<tr>
<th>Region</th>
<th>Integrated components of temperature tendency (K s(^{-1}) (1000–200 hPa)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(-u(\partial T / \partial x))</td>
</tr>
<tr>
<td>Spring, West control</td>
<td>0.65</td>
</tr>
<tr>
<td>(\Delta) spring, West</td>
<td>0.59</td>
</tr>
<tr>
<td>Summer, South control</td>
<td>0.94</td>
</tr>
<tr>
<td>(\Delta) summer, South</td>
<td>-0.11</td>
</tr>
</tbody>
</table>

Overall, increased temperatures are additionally seen during the summer over the South Box. Since mean state snow cover during the summer is negligible in this region, there is no snow-driven change in albedo. While changes in surface optical properties during summer are small (Fig. 6), atmospheric optical properties show more changes and make a comparatively larger difference for plants in the South Box. Namely, in contrast to the cloud increases seen over the Forest Box, forest establishment reduces cloud cover over the South Box (Fig. 4). This decrease occurs despite increases in latent heat flux that occur in the South Box as a result of higher GPP during the summer months. This reduction in clouds allows more

FIG. 8. Decomposition of relative humidity change in the South Box during the summer months (JJA). The yellow dashed line shows the change in relative humidity across the atmospheric column that would take place solely due to changes in temperature across the South Box. The blue dotted line shows the comparable change in relative humidity due to moisture changes. The solid green line shows the net change across the column and is representative of the change seen in the model output. Across most of the atmosphere, the effect of temperature dominates, showing the dominant effect of temperature in decreasing cloud cover over the South Box during the summer.

c. South Box: Forest drives plant growth through summer warming and cloud reduction driven by circulation changes

Warmer springs also drive increases in GPP (albeit smaller increases relative to the West Box) across all PFTs in the South Box. However, the South Box experiences additional climatic changes—in particular, summertime cloud loss—which extends growth increases longer throughout the growing season. The increase in downwelling solar radiation allows PFTs to experience increased growth during the summer, a mechanism we will focus on in this section. The land grid cells contained in the South Box are outlined in red in Figs. 3 and 4.

1) IMPOSED FOREST COOLS ATMOSPHERE OVER THE FOREST BOX IN COUPLED SIMULATIONS IN THE SUMMER

During summer inside the Forest Box, increased atmospheric reflectivity because of low-cloud formation outweighs the darkening effect of adding trees, leading to a decrease in net shortwave radiation absorbed at the surface. Surface optical changes lead to an average summertime drop in surface albedo of 0.03, while cloud formation leads to a 0.02 increase in atmospheric albedo (Fig. 7). We define atmospheric albedo as the fraction of solar radiation that does not reach the surface at all (blocked by the atmosphere) such that the sum of atmospheric albedo and surface albedo is the planetary (TOA) albedo. As a result, in the Forest Box, shortwave radiation absorbed by the surface decreases by 7.4 W m\(^{-2}\); outgoing longwave drops by 14.8 W m\(^{-2}\); sensible heat flux drops by 16.9 W m\(^{-2}\), while latent heat flux increases by 27.1 W m\(^{-2}\).

Thus, in the summer, the formation of clouds leads to reduced incoming shortwave energy at the surface combined with increased outgoing latent heat flux that leads to a near-surface cooling of 1.5 K over the Forest Box.

As in the West Box, the South Box experiences large-scale warming effects from the neighboring established forest acting as a springtime energy source for the region as a whole (Fig. 4). This springtime warming helps to establish an earlier start to the growing season, just as in the West Box. This early growing season warming helps establish higher leaf area (measured LAI) for each PFT, and by itself initiates increased GPP for all PFTs throughout the summer, regardless of meteorological changes (Fig. 6).
shortwave to be absorbed by the surface and warmer temperatures during the summer (Fig. 4).

To diagnose if the cloud increase in the South Box is due to atmospheric changes in temperature or moisture, we use relative humidity as a proxy for cloudiness in the model and compare the contributions from temperature changes and moisture changes with relative humidity changes (see section 2). During the summer months over the South Box, changes in specific humidity and temperature counteract one another; however, the effect of the temperature change wins out (Fig. 8).

2) WARMING DUE TO INCREASE IN ZONAL HEAT AND MOISTURE TRANSPORT FROM FOREST BOX TO SOUTH BOX

As was done to diagnose the cause of springtime heating in the West Box, we can examine the changes to the various terms in Eq. (3) to isolate the heating mechanism in the South Box. The largest change over the South Box in the summer is in the meridional temperature advection term, \(-\partial_t \bar{T} \bar{y}\), which is the only term that contributes positively to heating across the entire column. Below 200 hPa, \(-\partial_t \bar{T} \bar{y}\) uniquely contributes consistently positively to the change in balance of temperature, and this is also the section of the atmospheric profile where temperature increases consistently (Table 2; Fig. A3 in the appendix). Other contributions below 200 hPa, including zonal advection of heat, are reduced by smaller amounts as can be seen in Table 2. We can therefore attribute the temperature increase in the atmosphere to enhanced northward temperature advection.

Enhanced northward heat transport leads to reduced cloud cover, increased insolation, and consequent forest expansion in the South Box. A further examination of changes in meridional heat transport shows that changes in \(v\), rather than changes in \(\partial_t \bar{T} \bar{y}\), lead to a circulation change resembling a stationary wave that drives stronger northward winds over the South Box, bringing warm winds from the Gulf of Alaska. (See the appendix for further discussion of this mechanism.)

d. Mechanisms for expansion associated with coastal boundary

Both heat and moisture advection play important roles in making adjacent land areas more suitable for forest expansion, and more work is needed to determine how important the coastal location of our established forest is. Specifically, with respect to productivity increases to the south, it is natural to question whether the establishment of a forest elsewhere in the region would have produced the same cloud losses. While the existing meridional temperature and moisture gradients are relatively strong in this location when compared with similar locations around the same latitude band, poleward heat transport from the equator would be likely to increase no matter where any given energy source (i.e., forest emergence) were placed in a similar experiment (with the exception of the poles). Furthermore, it was the dominance of temperature over moisture that in the end determined that a forest might reduce clouds southward in the high latitudes; reducing the availability of moisture by placing the forest farther inland could be imagined to amplify this effect.

Having the Bering Sea to the west and the Gulf of Alaska to the south may have contributed to the existing wind patterns that enabled heat and moisture to be easily transported to our regions of interest; however, in a paleoclimate context, these wind patterns would have been considerably different with a tall ice sheet to the east, and more work is needed to investigate the role of coastal configuration and orography in climate-enabled forest self-expansion.

In conclusion, in this study, we have shown that surface–atmosphere interactions are strong enough to cause changes in plant productivity locally and in neighboring regions. Forest growth and expansion could be perpetuated by feedbacks like the ones we have identified. The mechanisms we have simulated are inspired by past instances of high-latitude forest growth and can be used to conceptualize the processes at play in this region. The increases in productivity that take place in regions neighboring initial forest establishment are driven by a balance of a variety of mechanisms, implying that the balance of forces that determine whether a forest will be catalyzed or unaffected would be sensitive to the initial forest’s location (proximity to the coast), as well as other factors we have intentionally excluded, such as orographic changes and changes in CO2.

e. Application to real-world forest expansion

This study is an exploration of the climate impacts of high-latitude forest expansion. It is not representative of a specific instance of forest expansion in Earth’s past but is inspired by the high-latitude changes in response to orbital forcing as seen in paleorecords (Bigelow et al. 2003; Williams et al. 2004; Bartlein et al. 2011) and by previous climate simulations of vegetation–climate interactions that highlight the power of high-latitude responses (Kutzbach et al. 1998; Foley et al. 1994; Swann et al. 2010). With our example high-latitude forest, we are able to learn that through albedo and evapotranspiration changes as well as consequent temperature and moisture advection, boreal forests are able to help themselves expand, suggesting that these mechanisms are relevant in the context of rapid climate changes and deglaciation. We have taken advantage of the flexibility of a climate model to decouple establishment and expansion, first by quantifying the local and large scale climate response to a newly established forest, and second by quantifying the impact of that climate response on neighboring plants to indicate the influence establishment could have on expansion. Thus, we are able to isolate the role of plant feedbacks in accelerating their own expansion.

This study isolates the ability for forest establishment to influence nearby plants only through atmospheric influences. Our results show that land–atmosphere effects of forest establishment can lead to significant benefits to plants and potential forest expansion adjacent to an area of forest emergence.

4. Discussion and conclusions

In this analysis we use an unconventional experimental design to quantitatively attribute how vegetation–atmosphere interactions drive forest expansion. While simplified, the demonstration of this effect is novel and moves beyond either quantifying forest influences on climate or reconstructing past forest expansion.
We use this simplified example inspired by paleoclimate to build our understanding of how vegetation–atmosphere interactions play out in the climate system. We can use it to interpret biophysical impacts from climate projections and afforestation projects in addition to past forest establishment.

The use of an Earth system model to comprehensively simulate vegetation climate interactions advances our understanding of ecosystem–climate interactions of a biome already known to have an outsized influence on global climate (Bonan et al. 1992; Swann et al. 2010), in a region already experiencing climate change (Pearson et al. 2013; Chapin et al. 2000). While highly idealized, our setup allows us to isolate and examine the influence of boreal forest establishment on forest expansion by assessing the response of nearby plants to the individual climate impacts caused by the forest and allows us to identify a set of dominant dynamics at play, building understanding of the Earth system, and setting the stage for complementary assessments using high-resolution regional models that would yield information about how specific instances of forest establishment could shift biome boundaries and influence expansion. Our approach illuminates vegetation–climate dynamics and potential for improved growing conditions in a way that is representative of the region but is not able to map edge effects in this way.

In this study we find that high-latitude forest establishment has the capacity to facilitate forest expansion through multiple pathways: It can improve growth in nearby plants by driving warmer springtime temperatures especially through zonal temperature advection, which may magnify earlier snowmelt and reduced springtime albedo. Additionally, forest establishment may warm nearby temperatures throughout spring and summer, allowing neighboring plants to attain higher leaf area earlier (Figs. 3, 5 and 6). Increased summertime cloud formation over an established forest has the potential to drive reductions in summertime cloudiness to the south by strengthening meridional temperature advection. Over the land to the south, stronger northward winds can draw warmth (and to a lesser extent, moisture) from the Gulf of Alaska and Bering Sea, which reduces overall relative humidity and leads to increased summertime solar radiation incident at the surface, leading to warmer temperatures for plants (Figs. 4 and 8; appendix Fig. A3).

It is possible that a different global land-cover distribution could elicit different responses from local perturbations in plants. However, we expect this to be a secondary effect to the perturbations themselves. We can offer some interpretation from Lagué et al. (2019) regarding the collateral atmospheric differences between what we simulated and what one may see with a different vegetation distribution: considering a mid-Holocene example, orbitally forced vegetation simulations from Kutzbach et al. (1998) show that the distinguishing features of vegetation from 6000 years ago (as compared with modern vegetation) include more extensive taiga and cold deciduous forest in the high latitudes generally in addition to more moisture-demanding vegetation in the southeastern region of the modern Sahara desert as well as in central Australia. Lagué et al. (2019) show that increases in vegetation in these places, and concomitant changes in surface characteristics (lowering albedo, increasing evaporative resistance, and increasing vegetation height) would likely incur warming feedbacks.

Our findings have several other important caveats. Seed dispersal is not represented in CLM5.0; in these simulations, we assume no competition between plants and no limit to seed dispersal. We additionally do not account for soil carbon stocks (which would not come to equilibrium in these simulations and do not account for any response to changes in atmospheric carbon dioxide, which is held constant).

We perform our experiments at low spatial resolution and in doing so restrict our results to large-scale effects that can be reasonably generalized. The use of a low-resolution global climate model does introduce considerable uncertainty. For this reason, we focus our attention on the role of large-scale land-cover changes, examining broadscale impacts on local climate and do not explicitly account for edge effects. This region’s climate in reality depends on the fine-scale topography of its coastline and terrain. Regional circulation patterns, such as atmospheric rivers, that respond to larger climate modes are not well represented by CESM at low resolution (Liu et al. 2022). Modern Alaskan circulation and precipitation patterns are highly dependent on the fine-scale orography of Alaska, which is poorly resolved in coarse-scale climate models. In particular, in drawing conclusions about plant responses in the West Box, real-world plant growth is affected by other important climatological constraints, including waterlogging, the spatial pattern of permafrost, fire disturbance, microclimates associated with subgrid elevation, and other aspects of the coastal landscape that are included but may be poorly represented. Additionally, our results that show plant growth improvements in the South Box due to increased heat and moisture advection from the Gulf of Alaska are subject to uncertainty due to the coarse representation of Alaskan orography. Despite these limitations, the novel experimental framework presented in this study allows us to draw broad conclusions about the ability of high-latitude forest expansion to enhance growing conditions for vegetation in the surrounding region.

Acknowledgments. Authors Shum and Swann acknowledge support from NSF Grant 1553715 to the University of Washington. This research was done while author Lagué received postdoctoral funding support from the James S. McDonnell Foundation Postdoctoral Fellowship in Dynamic and Multiscale Systems and NSERC Grant PGSD3-487470-2016 and while author Rushley received postdoctoral funding support from an NRC Research Associateship award at the Naval Research Laboratory. We acknowledge high-performance computing support from Cheyenne (https://doi.org/10.5065/D6RX9X9H) provided by NCAR’s Computational and Information Systems Laboratory, sponsored by the National Science Foundation.

Data availability statement. Model results are available through the University of Washington Libraries ResearchWorks digital repository (http://hdl.handle.net/1773/49485). Source code for CESM, the model used in this study, is archived and publicly accessible online (https://doi.org/10.5281/zenodo.3895306), with development code publicly available on github (https://escomp.github.io/CESM/release-cesm2/downloading_cesm.html) for CESM.
APPENDIX

Surface Representation in CLM, Heat Transport Analysis, and Additional PFT Responses

a. Optical properties of vegetation and soil

Surface albedo for vegetation land units in CLM is determined by the optical properties of the PFTs on the ground. These properties are used to calculate the flux of solar radiation (per watt received at the surface) reflected, transmitted, and absorbed by the components of the vegetated land surface (leaves, stems, or ground) following Sellers et al. (1986). The specific optical properties for leaf-level albedo $a$ and total scattering $v$ for visible wavelengths are given in Table A1 according to Lawrence et al. (2019).

b. Diagnosis of changes in meridional heat transport to South Box

In the South and West Boxes, we diagnose the source of warming using the thermodynamic energy equation [Eq. (3)], separating zonal heat advection, meridional heat advection, vertical motions, and diabatic heating (Figs. A1; A2). Calculations of $\partial T / \partial y$ are performed across the grid cells’ southern boundaries, and calculations of $\partial T / \partial x$ are done across the grid cells’ eastern boundaries.

We can distinguish the influence of changes in wind from changes in the meridional temperature gradient by expanding $\Delta[-\nu(\partial T / \partial y)]$ in the following way:

$$
\Delta[-\nu(\partial T / \partial y)] = - \left( \nu_{\exp \beta_{y_{\text{cell}}}} - \nu_{\text{cell}} \beta_{y_{\text{cell}}} \right) - \left( \nu_{\text{cell}} \beta_{y_{\text{exp}}} - \nu_{\text{cell}} \beta_{y_{\text{cell}}} \right).$

(A1)

In the equation above, changes in meridional wind are captured in the first term A and changes in the temperature gradient are captured in the second term B. The contribution of each of these terms can be seen in Fig. A3. The computation of $\partial T / \partial y$ is computed as the change in temperature across the southern boundary of each grid cell in the South Box (including the border between ocean grid cells directly to the south of the box).

We find that across the South Box, changes in $\nu(\partial T / \partial y)$ are driven more by changes in $\nu$. The meridional temperature gradient works in the opposing direction, as more warming is seen across the atmospheric column in the northern part of the South Box (Fig. A3).

Summertime cooling and subsidence over the Forest Box contribute to a near-surface anticyclonic circulation shift around the Forest Box, driving convergence near the surface and divergence aloft, at least partially leading to more warm air being advected northward over the South Box. The South Box is also the region with the largest increases in meridional heat and moisture transport. We conclude that the establishment of forest led to increases in meridional wind in the South Box during the summer, driven by a stationary-wave-like circulation change. This increase in $\nu$ leads to increased advection of heat and moisture from the Bering Sea and Gulf of Alaska,

Table A1. Optical properties/characteristics of land surfaces used in CLM, where $a$ is PFT leaf reflectance and $\omega$ is the scattering coefficient and in plants is equal to $a + \tau$, where $a$ is the leaf-element reflectance and $\tau$ is the leaf-element transmittance. PFT reflectances have a narrow range from 0.07 (needleleaf), to 0.10 (broadleaf) to 0.11 (grass). Values are taken from the CLM5.0 technical documentation (Lawrence et al. 2019).

<table>
<thead>
<tr>
<th>PFT</th>
<th>Reflectance $a$</th>
<th>Scattering coef $\omega$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Needleleaf evergreen boreal tree leaf</td>
<td>0.07</td>
<td>0.12</td>
</tr>
<tr>
<td>Needleleaf deciduous boreal tree leaf</td>
<td>0.07</td>
<td>0.12</td>
</tr>
<tr>
<td>Broadleaf deciduous boreal tree leaf</td>
<td>0.10</td>
<td>0.15</td>
</tr>
<tr>
<td>Broadleaf deciduous boreal shrub leaf</td>
<td>0.10</td>
<td>0.15</td>
</tr>
<tr>
<td>C3 Arctic grass leaf</td>
<td>0.11</td>
<td>0.16</td>
</tr>
<tr>
<td>Intercepted snow</td>
<td>—</td>
<td>0.8</td>
</tr>
<tr>
<td>Soil</td>
<td>—</td>
<td>0.04–0.36</td>
</tr>
</tbody>
</table>
ultimately leading to cloud losses and more incoming solar radiation at the land surface (and warming).

Overall, these circulation-driven changes to cloud cover and temperature can explain statistically significant GPP increases with 90% confidence, meaning that the explanation for GPP increases during the summer cannot be restricted to an analysis of summertime climate and that increases in LAI during the spring allowed a marginally improved climate to yield higher rates of productivity later in the season (Fig. 6).

c. Growth responses by PFT

In our analysis, we selected PFTs with the strongest presence in each region to illustrate a representative growth response for the West and South Boxes. We include in Figs. A4–A11 all responses for each PFT in each region.

![Fig. A1](image1.png)

**Fig. A1.** (a) Mean change in atmospheric temperature over the South Box. (b) Decomposition of the thermodynamic energy equation [Eq. (3)] in the South Box during experiment and control simulations. (c) Change in components of the thermodynamic energy equation, showing that the change in \(-v(\partial T/\partial y)\) contributes dominantly to the change in atmospheric temperature.

![Fig. A2](image2.png)

**Fig. A2.** (a) Mean change in atmospheric temperature over the West Box. (b) Decomposition of the thermodynamic energy equation [Eq. (3)] in the West Box during experiment and control simulations. (c) Change in components of the thermodynamic energy equation, showing that the changes in \(-u(\partial T/\partial x)\) and \(Q\) contribute dominantly to the change in atmospheric temperature.
FIG. A3. (a) Change in zonal mean atmospheric temperature over the South Box. The pattern of warming shows more dramatic temperature change in northern grid cells, explaining the negative contribution to $\Delta [-\psi (\partial T / \partial y)]$. (b) Change in zonal mean meridional wind over the South Box. [Nonsignificant changes are masked out in (a) and (b).] (c) Change in components of meridional temperature advection in Eq. (A1) (the dotted purple line shows the wind-driven component, term A; the dashed red line shows the temperature-gradient-driven component, term B. Changes in wind contribute dominantly to upper-atmosphere $\psi (\partial T / \partial y)$ changes.
FIG. A4. Changes in growing conditions in the West Box during the growing season (March–August) for needleleaf evergreen boreal trees; (a) spring and (b) summer changes in solar radiation at the surface are shown, with negative values indicating into the ground.
FIG. A5. As in Fig. A4, but for needleleaf deciduous boreal trees.
EARTH INTERACTIONS VOLUME 27

Changes in Energy and Growing Conditions in the South Box
pft03 (needleleaf deciduous boreal tree)

FIG. A6. As in Fig. A4, but in the South Box for needleleaf deciduous boreal trees.
FIG. A7. As in Fig. A4, but for broadleaf deciduous boreal trees.
Changes in Energy and Growing Conditions in the South Box

**a** Spring Solar Changes
- Solar blocked by atmosphere: -0.02, 1.32, 6.9
- Solar reflected by snow: 10.2, 11.4, 6.9

**b** Summer Solar Changes
- Solar absorbed by land surface (neg into ground): 7.3, 6.4, 5.0

**c** Spring Surface Changes
- Delta Snow Fraction: -0.02, -0.02, 0.00
- Delta Latent Heat Flux: 5.3, 0.65, 0.00
- Delta Vegetation Temperature: 6.2, 0.23, 0.00

**d** Summer Surface Changes
- Delta Snow Fraction: 5.9, 6.9
- Delta Latent Heat Flux: 5.8, 3.5
- Delta Vegetation Temperature: 5.8

**e** Spring Growth Changes
- March: 21.6
- April: 17.4
- May: 12.0

**f** Summer Growth Changes
- June: 6.2
- July: 3.3
- August: 5.8

---

**FIG. A8.** As in Fig. A4, but in the South Box for broadleaf deciduous boreal trees.
FIG. A9. As in Fig. A4, but in the South Box for broadleaf deciduous boreal shrubs.
Changes in Energy and Growing Conditions in the West Box
pH12 (arctic C3 grass)

Fig. A10. As in Fig. A4, but for Arctic C3 grasses.
REFERENCES


