Simple Land Interface Model (SLIM) Model Description

1.1 Introduction

The Simple Land Interface Model (SLIM) is a simple land model written to couple with a global Earth System Model (ESM). In particular, this model is currently written to couple to the Commu-
nity Earth System Model (CESM [Hurrell et al., 2013]) in place of the Community Land Model (CLM [Lawrence et al., 2018]).

This simple model bears strong resemblance to some of the early global land surface models, and draws heavily from the parameterizations set forth in models including the land surface model of Manabe [1969]; the Biosphere Atmosphere Transfer Scheme (BATS) [Dickinson et al., 1993]; the Land Surface Model version 1 (LSM 1.0) [Bonan, 1996]; and the Land Dynamics Model (LaD) [Milly and Shmakin, 2002a], which was used as the LM2 land surface model in the GFDL AM2LM2 model [Anderson et al., 2004].

1.2 Land Surface Model

The simple land model solves a linearized bulk surface energy budget coupled with soil temperatures and bucket hydrology. Various physical properties determine how energy is partitioned within the surface energy budget (see table 1). Hydrology is represented with a simple “bucket”, which has a prescribed capacity. Additionally, there is a simple snow model which allows for land-albedo feedbacks during winter months.

1.2.1 Land Surface Properties

The land model has several properties which are defined by the user for each land point. These variables are listed in table 1. The variables are provided to the model using a netcdf file provided by the user, where each value is specified for every terrestrial gridpoint.

The surface albedo determines how much incoming shortwave radiation is reflected from the land surface. The atmospheric model passes four different 'streams' of radiation to the land model: both a direct and diffuse value for visible light (0.2-0.7 µm) and near-infrared light (0.7 - 12.0 µm). Albedos for each of these radiative streams are prescribed both for bare ground and snow-covered ground. The emissivity $\varepsilon$ of the ground can be specified. Land surface emissivities range from 0.9 to 1.0 [Bonan, 2002]; if unspecified, and for the purposes of this study, it is assumed that $\varepsilon = 1$ over all land areas.

In order to calculate temperature profiles within the 10 soil layers, the soil thermal conductivity
κ and heat capacity $c_v$ must be specified. Over glaciated regions (specified using a user-defined glacier mask), the thermal resistance and heat capacity of ice rather than soil are used. The bucket model for hydrology requires a bucket capacity $W_{max}$ (the maximum amount of water each gridcell can hold) and a surface “lid” resistance to evaporation $r_s$. The aerodynamic roughness is calculated from the vegetation height $h_c$. A simple snow model is included in SLIM; as snow accumulates on the land surface, it begins to mask the albedo of the snow-free surface such that the surface albedo approaches that of snow.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Typical Value</th>
<th>Units</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\alpha_{gcd}$</td>
<td>0.2</td>
<td>[unitless]</td>
<td>Visible direct albedo for bare ground.</td>
</tr>
<tr>
<td>$\alpha_{svd}$</td>
<td>0.8</td>
<td>[unitless]</td>
<td>Visible direct albedo for deep snow.</td>
</tr>
<tr>
<td>$\alpha_{gnd}$</td>
<td>0.3</td>
<td>[unitless]</td>
<td>Near-infrared direct albedo for bare ground.</td>
</tr>
<tr>
<td>$\alpha_{snd}$</td>
<td>0.6</td>
<td>[unitless]</td>
<td>Near-infrared direct albedo for deep snow.</td>
</tr>
<tr>
<td>$\alpha_{gdf}$</td>
<td>0.2</td>
<td>[unitless]</td>
<td>Visible diffuse albedo for bare ground.</td>
</tr>
<tr>
<td>$\alpha_{svf}$</td>
<td>0.8</td>
<td>[unitless]</td>
<td>Visible diffuse albedo for deep snow.</td>
</tr>
<tr>
<td>$\alpha_{gnf}$</td>
<td>0.3</td>
<td>[unitless]</td>
<td>Near-infrared diffuse albedo for bare ground.</td>
</tr>
<tr>
<td>$\alpha_{snf}$</td>
<td>0.6</td>
<td>[unitless]</td>
<td>Near-infrared diffuse albedo for deep snow.</td>
</tr>
<tr>
<td>$M_s$</td>
<td>50</td>
<td>[kg/m²]</td>
<td>Snow-masking depth: mass of water required in snow bucket to fully mask the bare ground albedo.</td>
</tr>
<tr>
<td>$r_s$</td>
<td>100</td>
<td>[s/m]</td>
<td>“Lid” resistance to evaporation</td>
</tr>
<tr>
<td>$W_{max}$</td>
<td>200</td>
<td>[kg/m²] = [mm] Bucket capacity: maximum amount of water the soil can hold</td>
<td></td>
</tr>
<tr>
<td>$h_c$</td>
<td>0.1-20.0</td>
<td>[m]</td>
<td>Vegetation height; used to calculate roughness lengths for momentum and heat.</td>
</tr>
<tr>
<td>emissivity</td>
<td>0.9-1.0</td>
<td>[unitless]</td>
<td>Surface emissivity for longwave radiation</td>
</tr>
<tr>
<td>glc_mask</td>
<td>logical</td>
<td>[unitless]</td>
<td>Mask marking gridcells which should be treated as glaciers/ice sheets.</td>
</tr>
<tr>
<td>soil tk,1d</td>
<td>1.5</td>
<td>[W/m/K]</td>
<td>Thermal conductivity of soil (used for whole column).</td>
</tr>
<tr>
<td>soil cv,1d</td>
<td>2.0e6</td>
<td>[J/m3/K]</td>
<td>Heat capacity of soil (used for whole column).</td>
</tr>
<tr>
<td>glc tk,1d</td>
<td>2.4</td>
<td>[W/m/K]</td>
<td>Thermal conductivity of ice (used for whole column where glaciated).</td>
</tr>
<tr>
<td>glc cv,1d</td>
<td>1.9e6</td>
<td>[J/m3/K]</td>
<td>Heat capacity of ice (used for whole column where glaciated).</td>
</tr>
</tbody>
</table>

Table 1: Typical values for each of the model parameters in SLIM.
1.2.2 Atmospheric Fluxes

At each time step the land model is run, information is required about the state of the atmosphere. This information can come either from a data atmosphere model (e.g. reanalysis), or from a coupled atmospheric model such as the Community Atmosphere Model [Neale et al., 2012]. The variables required from the atmosphere by the land model are given in table 2.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Information required from atmosphere</th>
</tr>
</thead>
<tbody>
<tr>
<td>$SW_{nd}$</td>
<td>Direct, near-infrared incident solar radiation</td>
</tr>
<tr>
<td>$SW_{vd}$</td>
<td>Direct, visible incident solar radiation</td>
</tr>
<tr>
<td>$SW_{ni}$</td>
<td>Diffuse, near-infrared incident solar radiation</td>
</tr>
<tr>
<td>$SW_{vi}$</td>
<td>Diffuse, visible incident solar radiation</td>
</tr>
<tr>
<td>$LW$</td>
<td>Downwelling longwave radiation</td>
</tr>
<tr>
<td>$z_{ref}$</td>
<td>Height of reference level for atmospheric variables given at reference height</td>
</tr>
<tr>
<td>$T_{bot}$</td>
<td>Temperature at lowest level of atmosphere</td>
</tr>
<tr>
<td>$\theta_{ref}$</td>
<td>Potential temperature at reference height</td>
</tr>
<tr>
<td>$q_{bot}$</td>
<td>Specific humidity at lowest level of atmosphere</td>
</tr>
<tr>
<td>$u_{ref}$</td>
<td>Wind speed at reference height</td>
</tr>
<tr>
<td>$e_{ref}$</td>
<td>Vapor pressure at reference height</td>
</tr>
<tr>
<td>$p_{bot}$</td>
<td>Atmospheric pressure at lowest level of atmosphere</td>
</tr>
<tr>
<td>$p_{srf}$</td>
<td>Surface pressure</td>
</tr>
<tr>
<td>$\rho_{air}$</td>
<td>Density of air at reference height</td>
</tr>
<tr>
<td>$c_{p}$</td>
<td>Specific heat of air at constant pressure at reference height</td>
</tr>
<tr>
<td>rain</td>
<td>Liquid precipitation</td>
</tr>
<tr>
<td>snow</td>
<td>Frozen precipitation</td>
</tr>
</tbody>
</table>

Table 2: Table of values from the atmospheric model (or data atmosphere) required by the land model.

1.2.3 Surface Energy Budget

This model solves a linearized surface energy budget (eq 1) to calculate fluxes of energy and water to the atmosphere at each time step, and to calculate the surface temperature, temperature profile of the soil column, snow depth, and available water in each gridcell.
\[ SW^\downarrow + LW^\downarrow = SW^\uparrow + LW^\uparrow + LH + SH + G \]  

(1)

From the atmosphere, the land model receives the amount of downwards solar radiation \( SW^\downarrow \) for four radiation streams (visible direct, visible diffuse, near-infrared direct, and near-infrared diffuse), the amount of downwards longwave radiation \( LW^\downarrow \), and information about the temperature, humidity, and wind speed of the bottom of the atmosphere. The land model calculates the reflected shortwave radiation \( SW^\uparrow \), the upwards longwave radiation \( LW^\uparrow \), the sensible heat flux \( SH \), latent heat flux \( LH \), and ground heat uptake \( G \).

Equation 1 can be rewritten as

\[
(1 - \alpha)SW^\downarrow + \varepsilon LW^\downarrow - LW^\uparrow = LH + SH + G \\
R_{in} = LW^\uparrow + LH + SH + G \\
R_{net} = LH + SH + G
\]

(2)

where \( \alpha \) is the albedo of the surface. \( LW^\uparrow = \varepsilon \sigma T_s^4 \) is the longwave radiation emitted by the surface, which is a function of surface temperature \( T_s \) and surface emissivity \( \varepsilon \). \( R_{in} \) is the total absorbed radiative energy at the surface \( (SW_{abs} + LW_{abs}) \). \( R_{net} \) is the net radiative energy coming into the surface, which must be balanced by the turbulent energy fluxes (latent and sensible heat) and heat uptake by the land (soil or snow).

\( R_{in} \) can be directly calculated from the surface properties and atmospheric fluxes; \( LW^\uparrow = \varepsilon \sigma T_s^4 \), \( LH \), \( SH \), and \( G \) must each be found by evaluating the land model at each time step. In order to numerically calculate the balance of these fluxes at each time step, equation 2 is linearized around the change in surface temperature \( T_s \).

That is, we calculate a ‘first guess’ at a flux (using the surface temperature from the previous time step), as well as the derivative of that flux with respect to surface temperature. We proceed to calculate a new surface temperature for the current time step, then update the surface fluxes given the initial estimate and its derivative with respect to temperature. This is equivalent to taking a
first order Taylor expansion of each surface flux (equation 4), where some flux \( F \) at time \( i + 1 \) is approximated by its value at time \( i \) and its derivative with respect to surface temperature \( T_s \).

\[
F(T_s^{i+1}) = F(T_s^i) + \frac{\delta F(T_s^i)}{\delta T_s}(T_s^{i+1} - T_s^i) + \mathcal{O}(T^2)
\]  

(3)

\[
F(T_s^{i+1}) \approx F(T_s^i) + \frac{\delta F(T_s^i)}{\delta T_s}(T_s^{i+1} - T_s^i)
\]  

(4)

We solve the surface energy budget by linearizing each term with a first-order Taylor Expansion with derivatives w.r.t. surface temperature. i.e. for some surface flux \( S \), its value at the \( i + 1 \) time step is its value at the \( i \)th time step plus its derivative w.r.t. temperature times the change in surface temperature:

\[
S^{i+1} = S^i + \frac{\partial S}{\partial T_s}\Delta T_s.
\]  

(5)

For our longwave radiation, sensible heat flux, latent heat flux, and soil heat flux, this gives:

\[
LW_{\uparrow}^{i+1} = LW_{\uparrow}^{i} + \frac{\partial LW_{\uparrow}}{\partial T_s}\Delta T_s
\]  

(6)

\[
SH^{i+1} = SH^{i} + \frac{\partial SH}{\partial T_s}\Delta T_s
\]  

(7)

\[
LH^{i+1} = LH^{i} + \frac{\partial LH}{\partial T_s}\Delta T_s
\]  

(8)

\[
G^{i+1} = G^{i} + \frac{\partial G}{\partial T_s}\Delta T_s
\]  

(9)

\[
R_{in}^{i+1} = (1 - \alpha)SW_{\downarrow}^{i+1} - \varepsilon LW_{\downarrow}^{i+1}
\]  

(10)

Thus, in order to calculate the surface fluxes for the \( i + 1 \) time step, we must first calculate the change in surface temperature \( \Delta T_s \), and the derivative of each flux with respect to temperature.
1.3 Radiative Fluxes

The longwave radiation $LW^\uparrow$ [W/m²] emitted from the surface, and its temperature derivative, are given by equations 12-13 as a function of the surface temperature at the preceding timestep $T_s^i$, where $\varepsilon$ is the emissivity of the surface, and $\sigma = 5.670373 \times 10^{-8}$ W/m²/K⁴ is the Stefan-Boltzmann constant.

\[
LW^\uparrow = \varepsilon \sigma (T_s^i)^4 \tag{12}
\]
\[
\frac{\delta LW^\uparrow}{\delta T_s} = 4 \varepsilon \sigma (T_s^i)^3 \tag{13}
\]

The absorbed downwards longwave radiation is a direct function of the downwelling longwave radiation $LW^\downarrow$ and the surface emissivity $\varepsilon$. The absorbed shortwave radiation is a function of the downwelling shortwave radiation $SW^\downarrow$ in each of the four radiation streams, and the surface albedo for each corresponding radiative stream.

Bare-ground albedo is prescribed at each gridcell by the user. When a gridcell is free of snow, the bare-ground albedos are used. When there is snow on the ground ($S$, [kg/m²]), a blend of the bare-ground and snow-covered albedos are used, following equation 14 (this is the default implementation of snow-covered ground albedo in Anderson et al. [2004], Milly and Shmakin [2002b]). A snow-masking factor $M_s$ [kg] is used to define how steeply the bare-ground albedo should approach the snow-covered ground albedo (figure 1). A typical value of $M_s$ used in SLIM is 50 kg/m², which corresponds to roughly 25cm of snow (assuming a snow density of 200 kg/m³, typical of settled snow [Paterson, 1994]),

\[
\alpha_j = \begin{cases} 
\alpha_{soil,j} & S = 0 \\
(1 - \frac{S}{S+M_s})\alpha_{soil,j} + \frac{S}{S+M_s}\alpha_{snow,j} & 0 < S < M_s \\
\alpha_{snow,j} & S > M_s. \end{cases} \tag{14}
\]

So, the total amount of incoming radiative energy from the atmosphere at each time step can
be directly calculated as

$$R_{in}^{i+1} = \sum_j (1 - \alpha_j)SW_j^{i+1} + \varepsilon LW_j^{i+1}$$  \hspace{1cm} (15)

for the four shortwave radiative streams $j$.

1.4 Turbulent Heat Fluxes

The calculation of the turbulent heat fluxes (sensible and latent heat) relies on Monin-Obukhov theory [Monin and Obukhov, 1954]. Using the temperature, humidity, and wind speed of the bottom of the atmosphere, along with the temperature and humidity at the surface, the flux of heat and water can be calculated. These fluxes are influenced by the roughness of the land surface. The vegetation height $h_c$ [m] provided by the surface property netcdf file is used to calculate a displacement height $d$ (equation 16), a roughness length for momentum $z_{0m}$ (equation 17), and a
The above roughness lengths are used to calculate an Obukhov Length $L$, which in turn is used with the atmospheric temperature, humidity, and wind speed at the lowest atmospheric level to calculate an aerodynamic resistance for momentum $r_{am}$, heat $r_{ah}$, and moisture $r_{aw}$ (in [s/m]). The Obukhov Length is calculated iteratively, with an initial estimate $L_0$ used to calculate the next estimate $L_1$. In order to calculate the Obukhov Length, two intermediate functions $\psi_m$ (for momentum) and $\psi_h$ (for heat) are required (equations 20-21).

$$ y = (1 - 16x)^{0.25} $$

$$ \psi_m(x) = \begin{cases} 
2 \log\left(\frac{1+y}{2}\right) + \log\left(\frac{1+y^2}{2}\right) - 2 \arctan(y) + \frac{\pi}{2} & \text{if } x < 0 \\
-5x & \text{if } x \geq 0 
\end{cases} $$

$$ \psi_h(x) = \begin{cases} 
2 \log\left(\frac{1+y^2}{2}\right) & \text{if } x < 0 \\
-5x & \text{if } x \geq 0 
\end{cases} $$

We use the reference level (typically 10m) atmospheric winds $u_{ref}$, temperature $t_{ref}$, and water vapour $q_{ref}$, the surface temperature $t_s$ and water vapour $q_s$, as well as the dimensionless von Kármán constant $\kappa = 0.4$. Surface winds are assumed to be zero.

$$ u^* = \frac{u_{ref} \kappa}{\log\left(\frac{z_{ref} - d}{z_{om}}\right) - \psi_m\left(\frac{z_{ref} - d}{L_0}\right) + \psi_m\left(\frac{z_{om}}{L_0}\right)} $$

$$ t^* = \frac{(t_{ref} - t_s) \kappa}{\log\left(\frac{z_{ref} - d}{z_{oh}}\right) - \psi_h\left(\frac{z_{ref} - d}{L_0}\right) + \psi_h\left(\frac{z_{oh}}{L_0}\right)} $$
\begin{equation}
q^* = \frac{(q_{ref} - q_s)\kappa}{\log\left(\frac{z_{ref} - d}{z_0 h}\right) - \psi_h\left(\frac{z_{ref} - d}{L_0}\right) + \psi_h\left(\frac{z_{ref}}{L_0}\right)}
\end{equation}

\begin{equation}
t^*_v = t^* + 0.61 t_s q^*
\end{equation}

\begin{equation}
\theta_v = \theta_{ref}(1 + 0.61 q_{ref})
\end{equation}

Equations 22-26 are then used to make a new estimate of the Obukhov length,

\begin{equation}
L_1 = \frac{u^* \theta_v}{\kappa g t^*_v}.
\end{equation}

Additionally, we limit the Obukhov Length to keep it within a range that gives reasonable flux values, in the following manner; this capping is common in land models [Anderson et al., 2004, Lawrence et al., 2018].

\begin{equation}
\zeta_0 = \frac{z_{ref} - d}{L_1}
\end{equation}

\begin{equation}
\zeta = \begin{cases} 
\min(2, \zeta_0) & \text{if } \zeta_0 \geq 0 \\
\max(-2, \zeta_0) & \text{if } \zeta_0 < 0 
\end{cases}
\end{equation}

\begin{equation}
L_1 = \frac{z_{ref} - d}{\zeta}
\end{equation}

The above equations are iterated over until the difference between \(L_0\) and \(L_1\) is small, up to a maximum of 40 iterations per time step. If the above fails to converge in 40 iterations, the value of \(L_1\) with the smallest difference from its corresponding \(L_0\) is used.

Using the final value of \(L_1\), final values of \(u^*\) and \(t^*\) are obtained, which are used to calculate the aerodynamic resistance of momentum \(r_{am}\) and heat \(r_{ah}\) (in units of [s/m]). We also calculate the aerodynamic resistance for moisture \(r_{aw}\), by combining the evaporative resistance for heat with the prescribed evaporative resistance \(r_s\) that the user directly controls (comparable to a bulk stomatal resistance for a canopy - this is how the user directly controls how difficult it is to evaporate water out of the bucket). The aerodynamic resistances require the use of several variables
from the atmospheric reference height $z_{\text{ref}}$ (such as wind speed and air temperature):

$$r_{am} = \frac{u_{\text{ref}}}{u^*}$$  \hspace{1cm} (31)

$$r_{ah} = \frac{\theta_{\text{ref}} - T_s}{u^*T^*}$$  \hspace{1cm} (32)

$$r_{aw} = r_a + r_{ah}.$$  \hspace{1cm} (33)

The sensible heat flux $SH$ [W/m$^2$] is a function of the difference between the surface temperature $T_s^i$ and the potential air temperature at the reference height $\theta_{\text{ref}}^i$, as well as the roughness length for heat:

$$SH = c_{p,\text{air}}(T_s^i - \theta_{\text{ref}}^i)\frac{\rho_{\text{air}}}{r_{ah}}$$  \hspace{1cm} (34)

$$\frac{\delta SH}{\delta T_s^i} = c_{p,\text{air}}\frac{\rho_{\text{air}}}{r_{ah}}$$  \hspace{1cm} (35)

(where $c_{p,\text{air}}$ is the heat capacity of air and $\rho_{\text{air}}$ is the density of air).

The latent heat flux $LH$ [W/m$^2$] is a function of the difference between the surface humidity and the humidity in the atmosphere. It is further impacted by the evaporative resistance $r_s$, the aerodynamic resistance $r_{ah}$, and another term, $\beta$, which accounts for bucket fullness (equation 36).

$$\beta = \min\left(\frac{\text{water}}{0.75 \times W_{\text{max}}}, 1.0\right)$$  \hspace{1cm} (36)

$\beta$ is used to increase evaporative resistance under dry soil conditions. When the bucket is more than 75% full (i.e., the soil is moist), $\beta = 1$ (no additional resistance). When the bucket is less than 75% full, $\beta$ decreases linearly from 1 to 0; the smaller the $\beta$ term, the larger the resistance to evaporation.

The effective resistance of the land is a combination of the prescribed “lid” resistance $r_s$, the aerodynamic resistance due to the surface roughness $r_{ah}$ (see equation 33), and the $\beta$ value associ-
ated with how dry the soil is.

\[ LH = \rho_{air} \lambda (q_s^i - q_{ref}^i) \frac{\beta}{r_{aw}} \]  

(37)

\[ \frac{\delta LH}{\delta T_s} = \rho_{air} \frac{\delta q_s}{\delta T_s} \frac{\beta}{r_{aw}} \]  

(38)

In equation 38, \( \rho_{air} \) is the density of air, \( \lambda \) is the latent heat of vaporization (or sublimation, if temperatures are below freezing), \( q_s \) is the surface humidity, \( q_{ref} \) is the atmospheric humidity at some reference height, and \( T_s \) is the surface temperature. \( q_s, q_{ref}, \) and \( T_s \) are taken from the preceding time step \( i \). Note that if the latent heat flux term attempts to evaporate (or sublimate) more water than is available in the combined water and snow buckets, the latent heat flux term is set to the total water plus snow available, and \( \frac{\delta LH}{\delta T_s} = 0 \), and the excess energy that would have been used by LH if more water were available is instead partitioned to SH.

1.4.1 Ground Heat Flux

The change in heat uptake by the soil \( \frac{\delta G}{\delta T_s} \) requires solving the full temperature profile of the soil column. It is calculated using the energy imbalance of the other surface fluxes:

\[ \frac{dG}{dT_s} = \frac{\partial}{\partial T_s} (R_{in} - (LW^\uparrow + LH + SH)) \]

\[ = -\frac{\partial}{\partial T_s} LW^\uparrow - \frac{\partial}{\partial T_s} LH - \frac{\partial}{\partial T_s} SH, \]  

(39)

Heat transfer through the soil column is then calculated to get a new temperature for each soil layer, and a new surface temperature (section 1.4.3). Once the change in soil temperature at each soil layer (and specifically, the change in surface temperature \( T_s \)) is found, the total soil heat uptake \( G \) is given by

\[ G^{i+1} = G^i + \frac{dG}{dT_s} (T_s^{i+1} - T_s^i), \]  

(40)

where \( G^i \) is the energy flux into the soil at the previous time step \( i \), \( \frac{dG}{dT_s} \) is the derivative of the energy flux into the soil with respect to temperature. Here, \( G \) includes both the energy used to warm/cool the soil as well as any energy used to melt snow (\( G = G_{soil} + G_{snow} \)).
1.4.2 Hydrology

Water enters the land system by falling from the atmosphere as snow or rain. Water can fill up the bucket in each gridcell up to the bucket capacity $W_{max}$; if the amount of water in a gridcell exceeds $W_{max}$, the excess is moved into runoff. At present, the runoff is discarded; if this model were run coupled to a dynamic ocean model, runoff water should be routed through an appropriate river-runoff scheme and added to the ocean model. Water is removed from the bucket either through runoff or evaporation (latent heat flux). A baseline value of $W_{max} = 200\text{kg/m}^2$ is used, but can be modified spatially by the user. We use 200 kg/m$^2$ as it falls within the range of soil water capacities (assuming a 1m deep ‘bucket’) in the LM2 model, which range from 63 kg/m$^2$ for coarse soil to 445 kg/m$^2$ for peat.

Snow can also fall onto grid cells. There is no limit on the amount of snow which can be held on a gridcell (note - this means snow accumulates indefinitely over the ice caps - a glacier calving scheme would need to be implemented to counteract this effect if it was undesirable for some application). The heat flux into the land $G$ can be used to melt snow; melted snow flows into the water bucket.

The hydrology is a single layer ‘bucket’ with a prescribed capacity to hold water, and is not dependent on any specific soil properties. If the user wants this capacity to vary with geographically distinct soil types, they would need to feed the model a spatially varying map of maximum water content.

1.4.3 Soil Temperatures

In order to solve equation 2, we must find $\Delta T_s$. That is, we must calculate the new surface temperature. There are 10 soil layers in this model, with the midpoints of each soil layer given by equation 41:

$$z_i = -0.025 \times (\exp(0.5 \times (i - 0.5)) - 1.0) \quad i \in 1, 10. \quad (41)$$

The maximum soil depth is roughly 3.5m.

Soil temperatures are calculated using a simple heat diffusion equation through the soil layers,
with a zero-flux bottom boundary condition (no energy can go in or out of the soil column through the bottom) and an upper boundary condition given by the $G_{soil}$ term in the surface energy budget equation. Since the water in the bucket hydrology model is effectively isolated from the soil column, the amount of water in a given gridcell doesn’t influence the soil thermal properties. Thus, in addition to the prescribed heat capacity and thermal conductivity of the soil, there is a fixed density of freezable water in each soil layer, which is not coupled to the amount of water actually present and available for evaporation in that gridcell. The soil \textit{does} have a fixed density of freezable water in each layer (set by default to 300 kg/m$^3$). That is, the thin layers near the surface have a small amount of water in the soil layer which can be frozen/thawed using heat in that soil layer, while the deeper soil layers have a larger total volume of water available to freeze/thaw. This water is always present, and interacts with the soil \textit{only} in a thermal manner. The water in the soil layers does not interact with the hydrology portion of the model - that is, it is not moved up and down between soil layers, and cannot be evaporated. The primary reason to include this freezable water in each soil layer is to allow the model to have a more realistic timescale of soil temperature change during spring and fall at high latitudes, where it takes time for the ground to freeze and thaw. This is comparable to the representation of water and soil in the LM2 model [Anderson et al., 2004].

We use the surface energy fluxes to update the soil temperature at each layer $n = 1 : N$ in the soil column, using the equation for heat diffusion:

$$c_{v,n} \frac{\partial T}{\partial t} = - \frac{\partial F}{\partial z},$$

(42)

which can be re-arranged as

$$\partial T = - \frac{\Delta t}{c_{v,n} \Delta z_n} (F_{in} - F_{out}).$$

(43)

In eq 43, $T$ is the temperature [K], $\Delta t$ is the time step [s], $c_{v,n}$ is the heat capacity of the $n^{th}$ layer [J/m$^3$/K], $\Delta z_n$ is the thickness of soil layer $n$ [m], and $F_{in}$ and $F_{out}$ are, respectively, the fluxes of energy into the top of and out of the bottom of the $n^{th}$ soil layer [W/m$^2$].
At each soil layer, the fluxes into and out of each soil layer are given by:

\[
F_n = \begin{cases} 
R_{in} - (LW^+ + LH + SH) & n = 0 \text{ (top)} \\
-k_{x,n} \frac{(T_n - T_{n+1})}{(z_n - z_{n+1})} & n = 1 : (N - 1) \\
0 & n = N \text{ (zero-flux bottom)}.
\end{cases}
\]  

(44)

where \(LW^+, LH,\) and \(SH\) are the linearized surface fluxes.

Representing each soil layer with the fluxes of energy into and out of that soil layer results in a tri-diagonal matrix which we solve using the Thomas Algorithm. We start at the bottom of the soil column and sweep up the matrix to solve for an initial estimate of surface temperature \(T_s\). If there is no snow on the ground, or if there is snow on the ground, but \(T_s\) is below freezing, that \(T_s\) is used to complete the downwards sweep of the matrix and calculate the remaining soil temperatures. If the estimated surface temperature is above freezing and there is snow on the ground, the surface temperature is set to 0°C, and the difference between the predicted surface temperature and 0°C is used to melt snow. If there is still snow left after all the energy from the temperature difference is used, the surface temperature is kept at 0°C, and the downwards sweep of the matrix is used to calculate the temperature of the remaining soil layers. If there is enough energy associated with the difference in the predicted surface temperature and 0°C, all the snow is melted and the remaining energy is converted back to a temperature to calculate a modified \(T_s\), which is then used to solve for the remaining soil temperatures. This representation of snow melt is comparable to that used in the LM2 model [Anderson et al., 2004]. The energy used to melt snow is saved as \(G_{snow} = \text{snowmelt} \times h_{fus}\). A similar procedure is used to calculate the temperature profile of glaciated gridcells, but using the thermal properties of ice rather than soil.

After the soil temperatures have been calculated, we set the temperature of the top soil layer to be the surface temperature \(T_s\) (the topmost soil layer is very thin).
1.4.4 Water Accounting

Water enters the bucket via either rain (liquid precipitation) or snow melt. The bucket has a prescribed capacity; by default, the bucket capacity is 200 kg/m$^2$ (as in the LM2 code [Anderson et al., 2004]), but this can be modified to vary spatially by the user. Water can leave the bucket through evaporation (latent heat flux) or runoff (if the bucket exceeds capacity).

Snow accumulation is unlimited. Snow is added to the snow ‘bucket’ via snowfall (frozen precipitation) from the atmosphere. Snow can leave the snow bucket via either sublimation (directly to the atmosphere) or snow melt (to the water bucket on the land). Because snow accumulation is not limited by any ‘capacity’, this has the consequence that over glaciated regions, snow can accumulate indefinitely. Because the land/atmosphere/slab-ocean system does not conserve water, this is not a problem (the atmosphere doesn’t see any physical height to the snow), but if a dynamic ocean were used, a calving-scheme would need to be implemented to deposit ice into the ocean at high latitudes. This is similar to the implementation of snow in LM2 [Anderson et al., 2004].

1.5 Model behavior comparison with CLM

To demonstrate the general behavior of SLIM, we present a comparison of SLIM with CLM5 [Lawrence et al., 2018], forced with GSWP3 reanalysis data, repeating the data from year 2001-2010 for 50 years. Results shown are the average of the last 30 years of the simulations (allowing 20 years of model spin-up). We also compare the last 30 years of coupled simulations with SLIM and CLM5 coupled to the Community Atmosphere Model v5 (CAM5) [Neale et al., 2012], a slab ocean model (SOM) [Neale et al., 2012], and the Los Alamos Sea Ice Model for interactive sea ice (CICE5) [Hunke et al., 2013, LANS, 2017]. CLM is run in bgc mode (interactive biogeochemistry) with an 1850 map of vegetation. The pattern of vegetation height for SLIM is derived from the last 30 years of the CLM5 simulation. The pattern of evaporative resistance is derived from the stomatal conductance, saved from the CLM5 simulation. The stomatal conductance in CLM5 is calculated separately for sunlit and shaded leaves; we weight the conductance by the leaf area of sunlit and shaded leaves then convert to units of resistance. The four streams of radiation impacted...
by surface albedo (visible/near-infrared direct/diffuse) are calculated from the summertime surface albedo of CLM5 (to avoid imposing any pattern of seasonal snow cover). Gridcells which have over 50% glacier cover in CLM5 are defined as glaciers in SLIM, and thus use the thermal properties of ice and albedo of snow. Unless explicitly stated, we compare the results of the offline simulations rather than the coupled land-atmosphere simulations.

Only summertime conductance and albedo values are used for each hemisphere (June, July, August in the Northern Hemisphere, and December, January, February in the Southern Hemisphere), but the resulting maps of surface conductance and albedo are fixed throughout the year in the SLIM simulations, while the CLM albedo can vary as leaf area and soil moisture change. Snow cover can modify this base-line albedo throughout the year in both SLIM and CLM5, but the snow-free albedo in the SLIM simulations has no seasonal cycle, nor does the evaporative resistance. The snow-masking depth is fixed to 50 kg/m$^2$ everywhere in this SLIM simulation (and is not a function of vegetation height, as it is in CLM).

As such, we do not expect SLIM to produce a surface climate identical to that of CLM; rather, we demonstrate that even with this fairly crude approximation of the vegetation patterns of CLM, SLIM can still produce surface temperatures and fluxes comparable to those from the much more complex CLM.

The annual mean temperature of SLIM is comparable to that of CLM in most regions (figure 2). Portions of the northern high latitudes are over 1K cooler in SLIM than CLM, largely due to SLIM having a much brighter snow albedo over non-glaciated regions. Over the interior of Antarctica, sensible heat fluxes are slightly (10 W/m$^2$) too high and longwave fluxes are too low (figure 3). Albedo differences along the Antarctic coast (non-glaciated regions, where albedo is calculated as a combination of ground albedo and snow) additionally contribute to differences in surface temperature and surface energy fluxes.

Parts of the tropics and mid-latitudes are too hot (notably the Saraha/Sahel region of Africa, and the Tibetan plateau; figure 2). Over the Tibetan plateau, SLIM has a lower albedo (is darker), contributing to the warmer surface temperatures (figure 4). Over the Sahara, sensible heat fluxes
are too low (perhaps because of surface roughness differences) resulting in warmer surface temperatures. In sub-Saharan Africa, indeed in most of the tropics, latent heat fluxes are much lower than in CLM, which are roughly compensated for by sensible heat fluxes which are higher than in CLM (figure 3). That is, with the maps of surface properties used in this simulation of SLIM, the evaporative fraction is much lower than that of CLM.

In the coupled land-atmosphere simulations, the temperature anomalies between SLIM and CLIM increase substantially. In particular, parts of the tropics and mid-latitudes in the SLIM-CAM5 simulations are up to 3K warmer than the CLM5-CAM5 simulations, while parts of the Arctic are 1-3K cooler in SLIM-CAM5 than CLM5-CAM5. However, the temperature difference over other areas, specifically Antarctica, improve in the coupled simulations.

Seasonal cycles are shown for four locations with very different climates: the Sahara, the Amazon, Siberia, and the Great Plains (figure 5). The seasonal cycle of temperature is very similar between SLIM and CLM (figure 6), as the seasonal temperature differences driven by the atmospheric forcing data are much larger than the difference in temperature produced by the land models themselves. The differences between SLIM and CLM in individual terms of the surface energy budget are much larger than the differences in temperature, mostly coming from a difference in evaporative fraction: when latent heat flux in SLIM is lower than in CLM, sensible heat flux tends to be higher (figure 7).

The seasonal cycle of soil temperatures is physically consistent with our intuition (figure 8). In all areas, the peak in surface soil temperatures occurs in summer, with the peak in deep soil temperatures lagging. Deep soils are cooler than surface soils in summer, and warmer than surface soils in winter, as we would expect, with the ground taking up heat during warm summer months and releasing heat during cold winter months. The soil properties of all non-glaciated land areas in SLIM are identical in this simulation. The diurnal temperature profile of soil temperatures is also consistent with our physical expectation, with surface soil temperatures having a large diurnal temperature cycle and deep soils having no diurnal temperature cycle, and surface soil temperatures peak a few hours after local noon (figure 9).
Figure 2: Annual mean surface radiative skin temperature (left) and 2m air temperature (right) in the SLIM model run offline (top row), the difference between offline SLIM and CLM5 (middle row), and the difference between SLIM and CLM5 when coupled to CAM5.
Figure 3: Annual mean surface energy budget: net flux of shortwave radiation (row 1), net flux of longwave radiation (row 2), sensible heat flux (row 3), and latent heat flux (row 4) for the offline SLIM model (left column), and the difference between offline SLIM and CLM5 (right column).
Figure 4: Annual mean land albedo for visible direct radiation (row 1), visible diffuse radiation (row 2), near-infrared direct radiation (row 3), and near-infrared diffuse radiation (row 4) for the offline SLIM model (left column), and the difference between offline SLIM and CLM5 (right column).
Figure 5: Locations used for seasonal cycle plots: Sahara: 23.7°N, 12.5°E (orange); Siberia: 65.4°N, 150°E (blue); Great Plains: 42.6°N, 92.5°W (pink); Amazon: 4.7°S, 65°W (green).

Figure 6: Seasonal cycle of 2m air temperature [K] over 4 locations in SLIM (solid lines) and CLM5 (dash-dot lines). The climatological cycle is shown in black lines, while individual years are shown in gray.
Climatological surface energy fluxes in SLIM and CLIM5

Figure 7: Seasonal cycle of the individual terms of the surface energy budget [W/m$^2$] over 4 locations in SLIM (solid lines) and CLM5 (dash-dot lines). The net flux of shortwave radiation (absorbed shortwave) is shown in yellow; net longwave radiation (positive upwards) is shown in purple; sensible heat (positive upwards) is shown in red; latent heat (positive upwards) is shown in blue; ground heat flux (heat uptake by soil or snow) is shown in brown.
Figure 8: Seasonal cycle of soil temperature over 4 locations in SLIM, as a function of soil depth (darker lines are closer to the surface, lighter lines are deeper in the soil).

SLIM executes over 98% faster than CLM.
Figure 9: Diurnal cycle of soil temperature (averaged over all July days in a single year) over 4 locations in SLIM, as a function of soil depth (darker lines are closer to the surface, lighter lines are deeper in the soil). Local noon is indicated by the vertical dashed gray line.
2 Non-linear response to surface roughness

Initial simulations exploring the response of surface fluxes to vegetation height (using $h_c = 0.1$, 1.0, and 10.0 m) showed a distinctly non-linear response of surface temperature and surface energy fluxes to changes in vegetation height. This is in contrast to the linear response of surface temperatures and energy fluxes to incremental changes in albedo and evaporative resistance. To explore this further, we performed additional experiments with vegetation heights of $h_c = 2.0, 5.0,$ and 20.0 m, and found that the response is qualitatively similar to a negative exponential (supplemental figure 15). To proceed with our linear approximation of the response, we separate the response to vegetation height into two distinct regimes - that of ‘short’ vegetation ($\leq 2m$) and that of ‘tall’ vegetation ($\geq 2m$); this roughly corresponds to one relationship for grasses to shrubs and small trees, and second relationship for tall trees. We chose 2m as the separation point by calculating vegetation height associated with the maximum change in the slope of the change in surface temperature between consecutive vegetation height experiments for each non-glaciated land point, then taking the mean of the resulting vegetation heights. The resulting vegetation height with on average the largest change in the slope of the temperature response to changing vegetation height was approximately 1.5 m. So, we calculate one slope for the three experiments with $h_c = 0.1, 1.0,$ and 2.0 m, and a separate slope for the four experiments with $h_c = 2.0, 5.0, 10.0, \text{ and } 20.0 \text{ m.}$

The response of surface temperatures to incremental changes in short vegetation is much stronger than the response of surface temperatures to incremental changes in tall vegetation. The scaling factors applied to the slope of the temperature response (ie, the scaling factor that leads to roughly 1K maximum warming per incremental vegetation height change) are a 0.5 m decrease in vegetation height for the short vegetation regime, and a 10.0 m decrease in vegetation height for the tall vegetation regime. However, the overall patterns (though not magnitudes) of surface temperature response to decreased vegetation height are similar both between the short and tall response regime, and between the coupled and offline simulations (supplemental figure 16).
Figure 10: Surface temperature response to changing surface properties, but with a smaller range to better show spatial pattern of temperature response in offline simulations only. Annual mean scaled surface temperature $T_s$ response [K] for the offline simulations, per 0.04 darkening of the surface albedo (a), 50 s/m increase in evaporative resistance (b), and 5.0 m decrease in vegetation height (c). Cyan regions ($\Delta T_s < 0.1$) indicate regions where the temperature cooled substantially in response to the prescribed surface change.
Figure 11: Annual mean downwelling shortwave radiation at the surface [W/m²] in the ‘baseline’ idealized simulation (albedo = 0.2, evaporative resistance = 50 s/m, vegetation height = 0.1 m), with SLIM coupled to CAM5.
Figure 12: Change in shortwave cloud forcing (left) and longwave cloud forcing (right) per 0.04 decrease in albedo (a,d), 50 s/m increase in evaporative resistance (b,e), and 5 m decrease in vegetation height (c,f). Stippling indicates statistically insignificant regions ($p > 0.05$). The shortwave and longwave cloud forcing are calculated by the model, and are equal to the difference in radiation reaching the surface between a sky that includes the radiative effects of clouds, and a ‘clear’ (cloud-free) sky.
Figure 13: Change September ice fraction per (a) 0.04 decrease in land albedo, (b) 50 s/m increase in land surface evaporative resistance, and (c) 5m decrease in land surface vegetation height. Stippling indicates regions which are not significant ($p > 0.05$).
Figure 14: Annual mean change in cloud fraction per 0.04 decrease in surface albedo for (a) high (400 hpa - top of model), (b) medium (700-400 hpa) and (c) low (surface - 700 hpa) clouds per 0.4 decrease in surface albedo. Stippling indicates insignificant changes with p > 0.05. Horizontal blue lines show the region where subsidence was analyzed (not shown).
Figure 15: The annual mean surface temperature at select locations across the range of vegetation height experiments, with $h_c = 0.1, 1.0, 2.0, 5.0, 10.0, \text{ and } 20.0 \text{ m}$. Coupled simulations are shown in the left column, while offline simulations are shown in the right column. Mid and low latitude locations are shown in the top row, while high latitude locations are shown in the bottom row (not differing y-axis ranges). The latitude (positive for Northern hemisphere, negative for Southern hemisphere) and longitude locations (increasing Eastward from 0 to 360) are given for each location in the legend.
Tall regime: \( \Delta T_s \) [K] per ↓10.0 m vegetation height

Short regime: \( \Delta T_s \) [K] per ↓0.5 m vegetation height

Figure 16: Change in surface temperature in the coupled (left) and offline (right) simulations for the short (top row) and tall (bottom row) vegetation height regimes. The short regime is scaled by a 0.5 m decrease in vegetation height, while the tall regime is scaled by a 10.0 m decrease in vegetation height. Stippled regions do not pass a t-test with \( p=0.05 \).
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