Seasonal precipitation explains CH_4 sensitivity to soil temperature in northern peatland

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Key Points:

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1 Materials and Methods

1.1 Daily surface water inputs

We are interested in surface water available to peat soil microbes, but snow will not become available until it melts. To account for this in our surface water budget, we look at daily contributions from rainfall and snowmelt, including delayed responses in rain water contributions in the event of significant snowpack or icepack. Since we do not have data on snowfall or snow accumulation at this site, we need to model these winter dynamics. We use an energy-balance approach, similar to that outlined in Equation X in (?, ?). Consider some fixed ground area with snowpack m_S (kJ/m^2). Melting will primarily be driven by solar radiation, so we expect melt to first occur near the surface. Some proportion of snowmelt will travel through the snowpack and enter the soil column as surface water input, but the rest will refreeze before it reaches this destination. Similarly, some proportion of rainfall will refreeze before reaching the soil surface, depending on climate conditions, and the snowpack depth and density. This refrozen liquid water will contribute to an icepack of mass m_I. Together, we'll refer to the snowpack-icepack as the frozen-pack, where $m_F = m_S + m_I$. On a daily time scale, we ought to also consider the amount of "trapped" water m_W, that is rain water and frozen-pack melt that does not immediately contribute to surface water input because it must first travel through snow and ice. In the following, we outline an energy balance capturing changes in the state of some system, where the system here consists of the frozen-pack and trapped water, and state changes can be mass gained or lost, as well as changes in the system's physical properties (e.g. temperature, density, relative ratio of solids to liquids). We define,

$$\mathcal{E}_P + \mathcal{E}_T = R_n + H + LE + G + P, \tag{1}$$

where $\mathcal{E}_M$ corresponds to the energy absorbed or lost when water in the system undergoes a phase change (either melting or freezing; liquid-vapor phase changes are captured by LE), and $\mathcal{E}_W$ is the energy associated with temperature changes (both in kJ/m^2). On the right-hand side are the energy input fluxes (in kJ/m^2), where R_n is the net incoming solar radiation, H is the sensible heat exchange, LE is latent energy flux associated with latent heats of vaporization and condensation at the surface, G is the ground

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heat conduction, and P is the heat added by rainfall events. We'll refer to the sum of the right-hand side energy inputs as the net input energy, denoted E_{net}.

For some fixed day t, we assume snowfall and rainfall contribute to snowpack or trapped water, before the system undergoes any changes in phase or temperature. If net input energy is negative $E_{net} < 0$, we assume energy first goes to cooling trapped water to 0°C, and then freezing it *before* snowpack or icepack temperatures change. If $E_{net} > 0$, we assume energy goes towards warming snowpack and icepack to 0°C before any melting occurs. We arbitrarily allow snowpack warming before icepack warming, and snowpack melting before icepack melting, though such a distinction ultimately makes no difference when we're only concerned with surface water input contributions (e.g. mass of liquid water that leaves the system and becomes available to soil microbes).

Energy associated with warming or cooling, $\mathcal{E}_W$. On some fixed day t, we define initial snowpack mass m_S^*, and temperature T_S^* as,

$$m_S^*(t) = m_S(t-1) + X_S(t), \text{ and } T_S^*(t) = \frac{T_S(t-1)m_S(t-1) + T_{X_S}(t)X_S(t)}{m_S^*(t)}, \tag{2}$$

where $m_S(t-1)$ and $T_S(t-1)$ are respectively the snowpack mass and temperature at the end of the previous day. $X_S(t)$ and $T_{X_S}(t)$ denote the mass and temperature, respectively, of any snowfall that fell on day t. Analogous definitions follow for initial mass and temperature of trapped water (m_W^*, T_W^*), where rainfall temperature is always set to 0°C (rainfall temperature is accounted for in the energy flux P). Icepack mass and temperature (m_I^*, T_I^*) can similarly be defined, though precipitation events will not alter these from the end of the previous day.

The energy required to warm (or cool) snowpack temperature to T_S is given by,

$$w_S = m_S^* C_S (T_S - T_S^*), \tag{3}$$

where C_S is the heat capacity for snow ($C_S = 2.09\times10^{-1}$ kJ(kg°C)$^{-1}$). We can similarly define this for icepack ($C_I = 2.093\times10^{-1}$ kJ(kg°C)$^{-1}$) and trapped water ($C_W = 4.182\times10^{-1}$ kJ(kg°C)$^{-1}$), keeping in mind that snowpack and icepack temperature can only warm to 0°C before undergoing melt, and trapped water can only cool to 0°C before refreezing occurs. In total, $\mathcal{E}_W = w_S + w_I + w_W$.

Energy associated with melting or freezing, $\mathcal{E}_P$. We estimate the energy required to melt a mass δ_S of snow as,

$$e_S = \lambda\delta_S, \tag{4}$$

where $\lambda = 334$ kJ/kg is the latent heat of fusion for water. Similarly the energies needed to melt a mass δ_I of icepack, or freeze a mass δ_W of trapped water can be calculated as $e_I = \lambda\delta_I$, and $e_W = -\lambda\delta_W$, respectively. In total, $\mathcal{E}_P = e_S + e_I + e_W$, and the final masses on some day t

Surface water input due to rainfall and snowmelt. The final icepack mass on day t is given by $m_I(t) = m_I^*(t) - \delta_I + \delta_W$. For snowpack, $m_S(t) = m_S^*(t) - \delta_S$. In the absence of frozen-pack (e.g. $m_S^*(t) + m_I^*(t) = 0$), we assume all rainfall X_R contributes to surface water input I. However, if frozen-pack is present, we use the gravity-flow theory of water percolation through snow (?, ?), namely,

$$n(\alpha k)^{1/n} u^{(n-1)/n} \frac{\partial u}{\partial z} + \phi(1 - S_{wi})\frac{\partial u}{\partial \iota} = 0, \tag{5}$$

where $n = 3$ is the power-law exponent, and $\alpha = 5.47\times10^6$ (ms)$^{-1}$ is a constant based on the density ρ and viscosity of water μ, as well as gravitational acceleration g ($\alpha = \rho g \mu^{-1}$). $k = k(z)$ is the depth-dependent permeability (m^2), u is the volumetric flux (m^3/(m^{2}s)), z is the depth below the snow surface (m), ι is the time (s), and $S_{wi} = 0.07$ is the irreducible water saturation. To derive the drainage over a 24 hour period, we take an approach similar to that in the Appendix of (?, ?). We have $\phi(z) = \phi_0 - cz$, where ϕ_0 is the porosity at the snow surface, and c is the porosity gradient. We fix $\phi_0 = 7.41 \times 10^{-3}$, and $c = 12.6 \times 10^{-6}$—these values were derived from intermittent snow measurements taken at several nearby locations. Ultimately, we estimate the mass of drainage ς passing through a depth z over the time period from ι to $\iota + d\iota$ as,

$$\varsigma(z, \iota) = \rho_S u(z) d\iota, \tag{6}$$

where ρ_S is the density of the snowpack. Snowpack density is reevaluated at each snowfall event. For each snowfall event, we take $\rho^* = 50 + 3.4 * (T_a + 15)$, where snowfall density $\rho_{Sf} = \rho^*$ if $\rho^* < 50$ and $\rho_{Sf} = 50$ otherwise, and T_a is the air temperature that day (?, ?).

We take $u = \alpha k S_*^n$, where $S_* = (S_w - S_{wi})/(1 - S_{wi})$ is the effective water saturation, and S_w is the water saturation (water volume/pore volume). From (**??**), we obtain the characteristic of the differential equation $\frac{dz}{d\iota}\Big|_u$, which we integrate to obtain $d\iota$ at some depth z. By this formulation, we obtain the potential drainage on day t,

$$\widetilde{m_d}(t) = \int_{z=0}^{z_D} \int_{\iota=0}^{24(3600)} \varsigma(z, \iota) d\iota dz, \tag{7}$$

where z_D is the total depth of the frozen-pack on day t. This yields the actual drainage,

$$m_d(t) = \begin{cases} \widetilde{m_d}(t) & \widetilde{m_d}(t) < m_W^*(t) - \delta_W \\ m_W^*(t) - \delta_W & \text{else,} \end{cases} \tag{8}$$

Overall, we obtain the surface water input I on day t,

$$I(t) = \begin{cases} X_R(t) & m_S^*(t) + m_I^*(t) = 0 \\ m_d(t) & \text{else,} \end{cases} \tag{9}$$

and the new mass of trapped water,

$$m_W(t) = \begin{cases} 0 & m_S^*(t) + m_I^*(t) = 0 \\ m_W^*(t) - \delta_W + X_R(t) + \delta_S + \delta_I - m_d(t) & \text{else.} \end{cases} \tag{10}$$

Energy input fluxes. For our site, we have daily eddy flux tower measurements for R_n, H and LE. We use these energy measurements directly in our model.

We take G to be dependent on surface temperature. For this specific bog, G was found to be between 3 and 8% of net radiation R_n. For all years, we have daily soil temperature 10 cm below the soil surface, $T_{soil,10}$. With the absolute minimum and maximum values (taken over the entire dataset, years 2009-2018), we set the ordered pairs $(\min T_{soil,10}, 3\%)$ and $(\max T_{soil,10}, 8\%)$, and found a simple linear model for estimating G as a percent p of R_n, where p depends on soil temperature. This approach is based on work linking high ground surface fluxes to high ground surface temperatures (?, ?), and an assumption that high soil temperatures 10 cm below the soil surface will correspond to high surface temperatures. When significant snowpack is present, we use a constant $G = 173$ kJ/m^2, based on US Army Corps of Engineers, 1960 melt estimates.

Warm rainfall events can induce melting, so it's important we capture the heat input from rainfall events. For a rainfall (liquid water) event of mass X_R, we estimate the heat energy it imparts on the system as $P = X_R C_W T_w$, where T_w is the wet-bulb temperature. This approach parallels that taken in Eq X in (?, ?), with the notable modification of taking wet-bulb temperature, rather than air temperature, to be the temperature of the rainwater. We make this change as wet-bulb temperature is a more accurate predictor of actual rainfall temperature.

Differentiating snowfall and rainfall events. Climatological conditions dictate whether a precipitation event will contain snowfall. For any given precipitation event, we define the proportion of snowfall S_{Pr} by, (?, ?)

$$S_{Pr} = \begin{cases} 1 - 0.5\exp(-2.2(1.1 - T_w)^{1.3} & T_w < 1.1 \\ 0.5\exp(-2.2(T_w - 1.1)^{1.3} & T_w \geq 1.1, \end{cases} \tag{11}$$

where T_w is once again wet-bulb temperature (°C). This method is derived in (?, ?), and replicated in (?, ?) for a longterm, global dataset. In many models, air temperature T_a is used to determine if precipitation falls as rain or snow, but models often use wet-bulb temperature T_w instead, since snowfall can be observed at temperatures above 0°C (?, ?) in low humidity conditions.

While we don't have measured values for T_w, we have daily measurements of air temperature T_a (°C), and relative humidity RH (%), so we can estimate T_w based on these later two climatological inputs. We use the empirically-derived formula from (?, ?), that is,

$$T_w = T_a \tan^{-1}\left[a_1(RH + a_2)^{1/2}\right] + \tan^{-1}(T_a + RH) - \tan^{-1}(RH - a_3) + a_4(RH)^{3/2}\tan^{-1}(a_5 RH) - a_6, \tag{12}$$

where a_i for $i = 1, ...6$ are empirically-derived coefficients, given in Table **??**.

a_1	a_2	a_3	a_4	a_5	a_6
0.151977	8.313659	1.676331	0.00391838	0.023101	4.686035

Table 1: Empirical coefficients for Equation (**??**) for wet-bulb temperature, based on air temperature and relative humidity.

Appendix

Net radiation

Approach based on (?, ?).

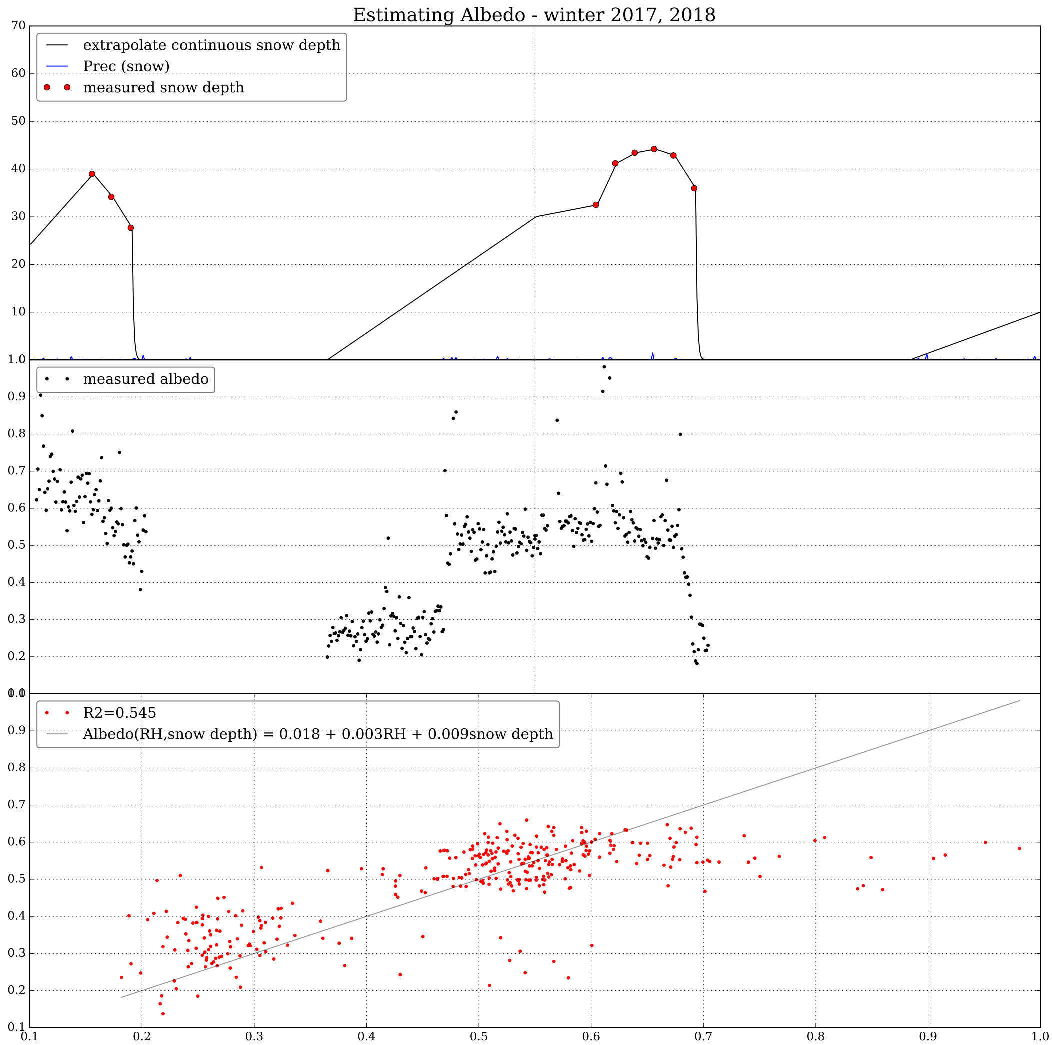

Figure 1

$$R_{net} = R_{short}(\min T_{air}, Albedo) + R_{long}(T_{surface}, T_{air}) \tag{13}$$

Acknowledgments

Enter acknowledgments, including your data availability statement, here.

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