I. Model description

(1) The remote sensing driven permafrost model (RS-PM)

The remote sensing based permafrost model, as described in Yi et al. (2018; 2019), uses a numerical approach to simulate soil temperature and soil freeze/thaw (F/T) process (and changes in soil liquid water content) along the 60 m soil profile using 23 soil layers, with finer vertical resolution at the surface and increasing layer thickness at depth. The soil nodes of 0-1m are distributed at 0.01, 0.03, 0.08, 0.13, 0.23 ,0.33, 0.45, 0.55, 0.70, 1.05 m. Multiple snow layers are used to simulate the snow insulation effects accounting for changes in snow density and thermal properties due to seasonal snow cover evolution. The snow thermal properties including heat capacity and thermal conductivity are empirically estimated from snow density (Calonne et al., 2011). External model inputs include the upper boundary temperature conditions, total soil moisture content, and snow depth and density. Model outputs include soil temperature and unfrozen liquid water fraction along the soil profile, which is also used to define the soil F/T state.

The following 1-D heat transfer equation with phase change was used to simulate the snow and ground thermal dynamics:

$$C\frac{\partial}{\partial t}T(z,t)+L\zeta\frac{\partial}{\partial t}\theta(T,z)=\frac{\partial}{\partial z}\left(\lambda\frac{\partial}{\partial z}T(z,t)\right), \quad z\in[z_s,z_b] \tag{S1}$$

where $T(z,t)$ is the temperature (°C) at a specific soil depth (z) and time step (t), L is the latent heat of fusion of water (J m^{-3}), ζ is the total soil water content (m^3 m^{-3}), and θ is the unfrozen liquid water fraction (%). C and λ are the volumetric heat capacity (J m^{-3} K^{-1}) and thermal conductivity (W m^{-1} K^{-1}) of soil respectively, varying with depth, soil moisture and F/T state. The upper boundary condition is set as the surface temperature at the snow/ground surface (z_s), while a heat flux characterizing the geothermal gradient is applied at the lower boundary ($z_b = 60\ m$). The soil thermal properties including the soil heat capacity and thermal conductivity are function of thermal properties of mineral and organic soil solid and liquid water, and ice components, weighted by their volumetric fraction.

The thermal conductivity λ is estimated as a normalized thermal conductivity of the dry (λ_{dry}) and saturated (λ_{sat}) soil thermal conductivity weighted by soil saturation:

$$\lambda = K_e\lambda_{sat} + (1 - K_e)\lambda_{dry} \tag{S2}$$

where the Kersten number (K_e) is a function of the soil saturation degree, which uses a logarithm form for unfrozen soils and linear form for frozen soils (Farouki, 1981; Lawrence and Slate, 2008). λ_{dry} is estimated from the soil bulk density. λ_{sat} is estimated as a geometric mean of the thermal conductivity of different soil components (Farouki 1981), including mineral and organic soil solid, liquid water and ice, which can vary several-fold from highly organic soil (~0.5 W m^{-1} K^{-1}) to mineral soils (1.5 ~ 3 W m^{-1} K^{-1}).

Soil water usually freezes at a sub-zero temperature depending on solute concentration and other factors, and the model uses the following empirical function to estimate the unfrozen liquid water fraction (θ):

$$\theta = \begin{cases} 1 & T \geq T_* \\ |T_*|^b |T|^{-b} & T < T_* \end{cases} \tag{S3}$$

The constant T_* represents the freezing point depression, with values generally above -1°C (Woo, 2012). b is a dimensionless parameter determined by fitting the unfrozen water curve, which can vary significantly depending on soil type (Schaefer and Jafarov, 2016).

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