COSIPY v1.2 - An open-source coupled snowpack and ice surface energy and mass balance model

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Abstract. Glacial changes play a key role both from a socio-economical and political, and scientific point of view. The identification and the understanding of the nature of these changes still poses fundamental challenges for climate, glacier and water research. Many studies aim to identify the climatic drivers behind the observed glacial changes using distributed surface mass and energy balance models. Distributed surface mass balance models, which translate the meteorological conditions on

5 glaciers into local melting rates, thus offer the possibility to attribute and detect glacier mass and volume responses to changes in the climatic forcings. A well calibrated model is a suitable test-bed for sensitivity, detection and attribution analyses for many scientific applications and often serves as a tool for quantifying the inherent uncertainties. Here we present the open-source coupled snowpack and ice surface energy and mass balance model in Python COSIPY, which provides a lean, flexible and user-friendly framework for modelling distributed snow and glacier mass changes. The model has a modular structure so that

10 the exchange of routines or parameterizations of physical processes is possible with little effort for the user. The framework consists of a computational kernel, which forms the runtime environment and takes care of the initialization, the input-output routines, the parallelization as well as the grid and data structures. This structure offers maximum flexibility without having to worry about the internal numerical flow. The adaptive sub-surface scheme allows an efficient and fast calculation of the otherwise computationally demanding fundamental equations. The surface energy-balance scheme uses established standard

15 parameterizations for radiation as well as for the energy exchange between atmosphere and surface. The schemes are coupled by solving both surface energy balance and subsurface fluxes iteratively in such that consistent surface skin temperature is returned at the interface. COSIPY uses a one-dimensional approach limited to the vertical fluxes of energy and matter but neglects any lateral processes. Accordingly, the model can be easily set up in parallel computational environments for calculating both energy balance and climatic surface mass balance of glacier surfaces based on flexible horizontal grids and with varying

20 temporal resolution. The model is made available on a freely accessible site and can be used for non-profit purposes. Scientists are encouraged to actively participate in the extension and improvement of the model code.

1 Introduction

Glacier variations are of great interest and relevance in many scientific aspects and application such as climate sciences, water resources management and tourism. In order to identify the climatic drivers for past, current and future changes, process understanding, observations and models of glacier mass change need to be combined appropriately. Schemes that relate the

5 surface mass balance of snow and ice bodies to meteorological forcing data have been set up and applied since many decades (e.g. Anderson, 1968; Kraus, 1975; Anderson, 1976; Kuhn, 1979; Male and Granger, 1981; Kuhn, 1987; Siemer, 1988; Morris, 1989, 1991; Munro, 1991). Studies have shown that the synthesis of these information provides a consistent understanding of the relevant mass and energy fluxes at the glacier-atmosphere interface, which in turn provides the necessary physical foundations to translate micro-meteorological conditions on glaciers into local melt rates (e.g. Sauter and Galos, 2016; Wagnon

10 et al., 1999; Oerlemans, 2001; Mölg and Hardy, 2004; Obleitner and Lehning, 2004; Van Den Broeke et al., 2006; Reijmer and Hock, 2008; Mölg et al., 2008; Nicholson et al., 2013).

Distributed mass balance models combine the local melt information to a glacier-wide surface mass change information and thus offer the possibility to attribute and detect glacier mass and volume responses to changes in the climatic forcings (e.g. Klok and Oerlemans, 2002; Hock and Holmgren, 2005; Mölg et al., 2009; Sicart et al., 2011; Cogley et al., 2011). Although

15 the accumulation and redistribution of snow are still deficient (e.g. Sauter et al., 2013), when coupled with atmospheric models such models have the potential to simulate present and future glacier evolution or to serve as a useful tool for monitoring climatic glacier mass change (Machguth et al., 2006). A well calibrated model is a suitable platform for sensitivity, detection and attribution analyses as well as a tool for the quantification of inherent uncertainties (e.g. Mölg et al., 2014; Maussion et al., 2015; Rye et al., 2012; Mölg et al., 2012; Sauter and Obleitner, 2015; Galos et al., 2017).

20 Over recent decades, several distributed mass balance models of varying complexity have been developed and successfully applied to different glacier systems and climate regimes. The models range from simple degree-day models (e.g. Radić and Hock, 2006; Schuler et al., 2005) to intermediate models (e.g. Machguth et al., 2009) and complex snow cover and glacier resolving physical models (e.g. Bartelt and Lehning, 2002; Vionnet et al., 2012; Hock and Holmgren, 2005; Klok and Oerle-mans, 2002; Sicart et al., 2011; Weidemann et al., 2018; Huintjes et al., 2015; Mölg et al., 2009; Michlmayr et al., 2008). The

25 latter model class is usually based on the same fundamental physical principles but differ in the parameterisation schemes and implementation techniques. Different research groups have their own in-house solutions which are often extended and modified for specific scientific questions and studies. The fact that often several sub-versions of the same model exist, with some of them being not freely available, makes it difficult for users to having access to up-to-date software. Ideally, a platform should (i) be continuously maintained, (ii) provide newly developed parameterisations, (iii) compile different model subversions developed

30 for specific research needs, (iv) be easily extensible and (v) be well documented and readable.

Here we present an open-source coupled snowpack and ice surface energy and mass balance model in Python (COSIPY) designed to meet these requirements. The structure is based on the predecessor model COSIMA (COupled Snowpack and Ice surface energy and MAss balance model, Huintjes et al., 2015). COSIPY provides a lean, flexible and user-friendly framework for modelling distributed snow and glacier mass changes. The framework consists of a computational core that forms the

runtime environment and handles initialization, input-output (IO) routines, parallelization, and the grid and data structures. In most cases, the runtime environment does not require any changes by the user. Physical processes and parameterisations are handled separately by modules. The modules can be easily modified or extended to meet the needs of the end user. This structure provides maximum flexibility without worrying about internal numerical issues.The model is provided on a freely

5 accessible git repository (https://github.com/cryotools/cosipy) and can be used for non-profit purpose. Scientists can actively participate in extending and improving the model code.

In this work, we describe the physical basics, parameterisations and outline the numerical implementation of the model version COSIPY v1.2 (https://doi.org/10.5281/zenodo.3613921). Section 2 gives an overview of the model concept, followed by the description of the modules (Section 3). The model architecture and the input/output are explained in Section 4). Section

10 5 shows an application of the model. The last section (Section 7) documents the code availability and software requirements.

2 Model concept

COSIPY is a multi-layered process resolving energy and mass balance model for the simulation of past, current and future glacier changes. The model is based on the concept of energy and mass conservation. The snow/ice layers are described by the volumetric fraction of ice θ_i, liquid water content θ_w and air porosity θ_a. Continuity constraints require that

15 $$\theta_i + \theta_w + \theta_a = 1. \tag{1}$$

The inherent properties, such as snow density ρ_s or specific heat of snow c_s, follow from the volumetrically weighted properties of the constitutes. For example, snow density is related by

$$\rho_s = \theta_i \cdot \rho_i + \theta_w \cdot \rho_w + \theta_a \cdot \rho_a, \tag{2}$$

where ρ_i is the ice density, ρ_w the water density, and ρ_a the air density (Bartelt and Lehning, 2002). Exchange processes

20 at the surface, the energy release and consumption through phase changes, control the vertical temperature distribution within the snow and ice layers. The energy balance also includes incoming shortwave radiation absorption and the sublimation or deposition of water vapour. Assuming the vertical temperature profile is given by $T_s(z,t)$, where z is the depth, the energy conservation can be represented by

$$\rho_s(z,t)c_s(\theta,z,t)\frac{\partial T_s(z,t)}{\partial t} - k_s(\theta,z,t)\frac{\partial^2 T_s(z,t)}{\partial z^2} = Q_p(z,t) + Q_r(z,t) \tag{3}$$

25 where $c_s = \rho_i c_i \theta_i + \rho_w c_w \theta_w + \rho_a c_p \theta_a$ and $k_s = k_i \theta_i + k_w \theta_w + k_a \theta_a$ are the volume-specific heat capacity and thermal conductivity of the snow cover (Bartelt and Lehning, 2002). The first term on the right-hand side (Q_p) is the volumetric energy sink or source by melting and meltwater refreezing. The second term (Q_r) is the volumetric energy surplus by the absorption of shortwave radiation (see Eq. 13).

The exchange processes at the snow/ice-atmosphere interface control the surface temperature $T_s(z=0,t)$ at an infinitesimal skin layer. From the energy conservation follows

$$k_s(\theta, z=0, t)\frac{\partial T_s(z=0,t)}{\partial z} = q_{sw} + q_{lw} + q_{sh} + q_{lh} + q_{rr}, \tag{4}$$

where q_{sw} is the net-shortwave radiation energy, q_{lw} is the net-longwave radiation energy, q_{sh} is the sensible heat flux, q_{lh}
5 is the latent heat flux, and q_{rr} is the heat flux from rain. To solve Eq. (4) for $T_s(z=0,t)$, the fluxes on the right-hand side must be parameterized (see Section 3). The parameterization results in a nonlinear equation which is solved iteratively. The left side of Eq. (4) provides the upper Neumann boundary condition (prescribed gradient) for Eq. (3). At the base of the glacier, the temperature must be specified (Dirichlet boundary condition) by the user. The melting rates in the snow cover and glacier ice are derived diagnostically from the energy conservation by ensuring that the temperature does not exceed the melting point
10 temperature T_m.

To solve the underlying differential equations, the computing domain must be discretized. Since extreme gradients in temperature, density and liquid water content can develop in the snowpack, COSIPY uses a dynamic, non-equidistant mesh. The mesh consists of so-called nodes that store the properties of the layers (e.g. temperature, density, and liquid water content), and is continuously adjusted during run-time by a re-meshing algorithm, i.e. the number and height of the individual layers
15 vary with time. Currently, two algorithms are implemented: (i) A logarithmic approach, where the layer thicknesses gradually increase with depth by a constant stretching factor. Re-meshing is performed at each time step. This is a fast method, but does not resolve sharp layering transitions, as these are smoothed by the algorithm. This approach is only recommended if a detailed resolution of the snow and ice cover is not required. (ii) An adaptive algorithm that assembles layers according to user-defined criteria. It uses density and temperature thresholds to determine when two successive layers are considered similar and when
20 they are not. When both criteria are met, these layers are merged. The liquid water content and the heights of the two layers are added and the new density is calculated from the volumetrically weighted densities of the two layers. To ensure energy conservation, the total energy is determined from the internal energies and converted into the new temperature. The adaptive re-meshing proves to be useful feature, but slightly increases both the computing time and the data volume.

Eq. (4) is solved using a Limited Memory Broyden–Fletcher–Goldfarb–Shanno (BFGS) algorithm (Quasi-Netwon method)
25 for bound constrained minimisation (Fletcher, 2000). Eq. (3) is then integrated with an implicit second-order central difference scheme (Ferziger and Perić, 2002). The heat sources can warm the snowpack and lead to internal melt processes. In case the liquid water content of a layer exceeds its retention capacity (Coléou and Lesaffre, 1998),

$$\theta_e = \begin{cases} 0.0264 + 0.0099\,\dfrac{(1-\theta_i)}{\theta_i}, & \text{if } \theta_i \le 0.23 \\ 0.08 - 0.1023\,(\theta_i - 0.03), & \text{if } 0.23 < \theta_i \le 0.812\,, \\ 0, & \text{if } \theta_i > 0.812 \end{cases} \tag{5}$$

the excess water is drained into the subsequent layer (bucket approach). The liquid water is passed on until it reaches the glacier surface where it is considered to be runoff. Meltwater refreezing and subsurface melting during percolation change the volumetric ice and water contents. Subsurface melt occurs when energy fluxes, e.g. penetrating shortwave radiation, warms the layer to physically inconsistent temperatures of $T_s > T_m$. Since physical constraints require that $T_s \leq T_m$, the energy surplus

5   is used to melt the ice matrix. Melt takes place when $T_s > T_m$ and the liquid water content increases by

$$\Delta\theta_w(z,t) = \frac{c_i(z,t)\theta_i(z,t)\rho_i(z,t)(T_s(z,t) - T_m)}{L_f \rho_w},$$ (6)

where $L_f = 3.34 \times 10^5 JKg^{-1}$ is the latent heat of fusion (Bartelt and Lehning, 2002). Mass conservation requires that the mass gain of liquid water content equals the mass loss of the volumetric ice content, so that

$$\Delta\theta_i(z,t) = \frac{\rho_w \Delta\theta_w(z,t)}{\rho_i}.$$ (7)

10   The latent energy needed by the phase change is

$$Q_p(z,t) = L_f \Delta\theta_i(z,t)\rho_i,$$ (8)

which is an heat sink because $\Delta\theta_i(z,t)$ is positive at melting. The energy used for melting ensures that $T_s(z,t)$ does not rise above T_m. In case $\theta_w > 0$ and $T_s < T_m$, refreezing can take place. Changes in volumetric fractions and the release of latent energy due to phase changes are treated equally. As the temperature difference must be negative due to the given constraints, it

15   follows from Eq. (6), Eq. (7), and Eq. (8) that Q_p becomes positive and latent heat release warms the layer.

Many of the quantities and fluxes in Eq. (3) and Eq. (4) are not measured directly and have to be derived via corresponding parameterizations. The next section describes the parameterizations implemented in COSIPY v1.2.

3   Model physics and modules

3.1   Snowfall and precipitation

20   When snowfall is given, it is assumed that the data represents the effective accumulation since snowdrift and snow particle sublimation are not explicitly treated in the model. Otherwise, snowfall is derived from the precipitation data using a logistic transfer function. The proportion of solid precipitation smoothly scales between 100 % (0 °C) and 0 % (2 °C), as suggested by Hantel et al. (2000). The fresh snow density for the conversion into snow depth is a function of air temperature and wind velocity

25    $$\rho_s(z=0,t) \ = \ \max\left[a_f + b_f(T_{z_t} - 273.16) + c_f u_{z_v}{}^{1/2}, \rho_{min}\right],$$ (9)

with the empirical parameters $a_f = 109 \, \mathrm{kg\,m^{-3}}$, $b_f = 6 \, \mathrm{kg\,m^{-3}\,K^{-1}}$, $c_f = 26 \, \mathrm{kg\,m^{-7/2}\,s^{1/2}}$, and $\rho_{min} = 50 \, \mathrm{kg\,m^{-3}}$ (Vionnet et al., 2012). In both cases fresh snow is only added when the height exceeds a certain user-defined threshold.

3.2 Albedo

The approach suggested by Oerlemans and Knap (1998) parametrizes the evolution of the broadband albedo. The decay of the
5 snow albedo at a specific day depends on the snow age at the surface and is given by

$$\alpha_{snow} = \alpha_f + (\alpha_s - \alpha_f) \exp\left(\frac{s}{\tau^*}\right), \tag{10}$$

where α_s is the fresh snow albedo and α_f the firn albedo. The albedo time scale τ^* specifies how fast the snow albedo drops from fresh snow to firn. The number of days after the last snowfall is given by parameter s. Besides the temporal change, the overall snowpack thickness impacts the albedo. If the thickness of the snowpack d is thin, the albedo must tend towards the
10 albedo of ice α_i. If one introduces a characteristic snow depth scale d^* (e-folding) the full albedo can be written as

$$\alpha = \alpha_{snow} + (\alpha_i - \alpha_{snow}) \exp\left(\frac{-d}{d^*}\right). \tag{11}$$

The model resets the albedo to fresh snow, if the snow accumulation exceeds a certain threshold (default value is 0.01 m) within one time step. This approach neglects sudden short-term jumps in albedo, which can occur when thin fresh snow layers quickly melt away. To account for this effect, the age of the underlying snow is also tracked. If the fresh snow layer melts faster
15 than τ^*, the age of the snow cover is reset to the value of the underlying snow (Gurgiser et al., 2013).

3.3 Radiation fluxes

The net-shortwave radiation in the energy conservation equation Eq. (3) is defined as

$$q_{sw} = (1 - \alpha) \cdot q_G, \tag{12}$$

where q_G is the incoming shortwave radiation, and α the snow/ice albedo. A proportion of the net shortwave radiation q_{sw}
20 can penetrate into the uppermost centimetres of the snow or ice (Bintanja and Van Den Broeke, 1995). The resulting absorbed radiation at depth z is calculated with

$$Q_r(z,t) = \lambda_r \, q_{sw} \, \exp(-z\beta), \tag{13}$$

where λ_r is the fraction of absorbed radiation (0.8 for ice; 0.9 for snow), and β the extinction coefficient (2.5 for ice; 17.1 for snow). Physical constraints require that $T_s \leq T_m$ so that the energy surplus is used to melt the ice matrix (see Section 2).

In case the incoming longwave radiation $q_{lw_{in}}$ is observed, the net-longwave radiation is obtained by

$$q_{lw} = q_{lw_{in}} - \epsilon_s \sigma T_0{}^4, \tag{14}$$

where ϵ_s is the surface emissivity which is set to a constant close to or equal to 1. In the absence of $q_{lw_{in}}$, the flux is parametrized by means of air temperature T_{z_t} and atmospheric emissivity,

$$5 \quad \epsilon_a = \epsilon_{cs}(1 - N^2) + \epsilon_{cl}N^2, \tag{15}$$

using the Stefan-Boltzmann law. Here, N is the cloud cover fraction, ϵ_{cl} the emissivity of clouds which is set to 0.984 (Klok and Oerlemans, 2002), and ϵ_{cs} the clear sky emissivity. The latter is given by

$$\epsilon_{cs} = 0.23 + 0.433 \, (e_{z_t}/T_{z_t})^{1/8}, \tag{16}$$

where e_{z_t} is the water vapor pressure (Klok and Oerlemans, 2002).

3.4 Turbulent fluxes

The turbulent fluxes, q_{sh} and q_{lh}, in Eq. (4) are parametrized based on the flux-gradient similarity which assumes that the fluxes are proportional to the vertical gradient of state parameters. However, since meteorological parameters are only considered from one height in the model a bulk approach is used whereby the mean property between the measurement height and the surface is considered (e.g. Foken, 2008; Stull, 1988). Assuming that fluxes in the Prandtl layer are constant, dimensionless transport 15 coefficients C_H (Stanton number) and C_E (Dalton number) can be introduced by vertically integrating the turbulent diffusion coefficients (Foken, 2008; Stull, 1988) so that the turbulent vertical flux densities can be written as

$$q_{sh} = \rho_a \, c_p \, \frac{1}{Pr} \, C_H \, u_{z_v} \, (T_{z_t} - T_0) \tag{17}$$

$$q_{lh} = \rho_a \, L_v \, \frac{1}{Pr} \, C_E \, u_{z_v} \, (q_{z_q} - q_0), \tag{18}$$

where ρ_a is the air density (derived from the ideal gas law), c_p is the specific heat of air for constant pressure, L_v is the latent 20 heat of vaporisation which is replaced by the latent heat of sublimation L_s when $T_0 < T_m$, Pr (0.8) is the turbulent Prandtl number, u_{z_v} is the wind velocity at height z_t, T_{z_t} and q_{z_q} are the temperature and mixing ratio at height z_t (assuming $z_t = z_q$), respectively, and q_0 is the mixing ratio at the surface where it is assumed that the infinite skin layer is saturated. Unlike the turbulent diffusion coefficients, the bulk coefficients are independent of the wind speed and only depend on the stability of the atmospheric stratification and the roughness of the surface. The aerodynamic roughness length z_{0_v} is simply a function of time 25 and increases linearly for snowpacks from fresh snow to firn (Mölg et al., 2012). For glaciers, z_{0_v} is set to a constant value.

According to the renewal theory for turbulent flow, z_{0_q} and z_{0_t} are assumed to be one and two orders of magnitude smaller than z_{0_v}, respectively (Smeets and van den Broeke, 2008; Conway and Cullen, 2013).

COSIPY provides two options to correct the flux-profile relationship for non-neutral stratified surface layers, by adding a stability correction using the (1) bulk Richardson-Number, and (2) Monin-Obukhov similarity theory (e.g. Conway and

5   Cullen, 2013; Radić et al., 2017; Stull, 1988; Foken, 2008; Munro, 1989). Using the bulk Richardson number the dimensionless transport coefficients can be written in the form

$$C_H = \frac{\kappa^2}{\ln\left(\dfrac{z}{z_{0_v}}\right)\ln\left(\dfrac{z}{z_{0_t}}\right)}\Psi_{Ri}(Ri_b) \tag{19}$$

$$C_E = \frac{\kappa^2}{\ln\left(\dfrac{z}{z_{0_v}}\right)\ln\left(\dfrac{z}{z_{0_q}}\right)}\Psi_{Ri}(Ri_b), \tag{20}$$

whereas the stability function

10   $$\Psi_{Ri}(Ri_b) = \begin{cases} 1, & \text{if } Ri_b < 0.01 \\ (1 - 5\,Ri_b)^2, & \text{if } 0.01 \leq Ri_b \leq 0.2 \\ 0, & \text{if } Ri_b > 0.2 \end{cases} \tag{21}$$

accounts for reduction of the vertical fluxes by thermal stratification and is a function of the Richardson number. The Richardson number

$$Ri_b = \frac{g}{T_{z_t}} \cdot \frac{(T_{z_t} - T_0)(z_t - z_{0_t})}{(u_{z_v})^2}, \tag{22}$$

follows from the turbulent kinetic energy equation and relates the generation of turbulence by shear and damping by buoy-

15   ancy (Stull, 1988). In the stable case ($0.01 \leq Ri_b \leq 0.2$), the function describes the transition from turbulent flow to a quasi-laminar non-turbulent flow, and hence, reduces the vertical fluxes. Once Ri_b exceeds the critical value $Ri_b = 0.2$, turbulence eventually extinguishes, and the vertical exchange is suppressed.

According to the Monin-Obukhov similarity theory, atmospheric stratification can be characterised by the dimensionless parameter

20   $$\zeta = z/L. \tag{23}$$

where

$$L = \frac{u_*^3}{\kappa \dfrac{g}{T_{z_t}} \dfrac{q_{sh}}{\rho_a \cdot c_p}} \tag{24}$$

is the so-called Obukhov length with u_* is the friction velocity and κ (0.41) the Von Karman constant (Stull, 1988; Foken, 2008). The length scale relates dynamic, thermal and buoyancy processes and is proportional to the height of the dynamic
5   sub-layer. The bulk aerodynamic coefficients for drag, heat and moisture for non-neutral conditions

$$C_D = \frac{\kappa^2}{\left[\ln\left(\dfrac{z}{z_{0_v}}\right) - \Psi_m(\zeta) - \Psi_m\left(\dfrac{z_{0_v}}{L}\right)\right]^2} \tag{25}$$

$$C_H = \frac{\kappa C_D^{1/2}}{\left[\ln\left(\dfrac{z}{z_{0_t}}\right) - \Psi_t(\zeta) - \Psi_t\left(\dfrac{z_{0_t}}{L}\right)\right]} \tag{26}$$

$$C_E = \frac{\kappa C_D^{1/2}}{\left[\ln\left(\dfrac{z}{z_{0_q}}\right) - \Psi_q(\zeta) - \Psi_q\left(\dfrac{z_{0_q}}{L}\right)\right]} \tag{27}$$

$$\tag{28}$$

10   can be derived by integrating the universal functions (Businger et al., 1971; Dyer, 1974) where

$$\Psi_m(\zeta) = \begin{cases} 2\ln\left(\dfrac{1+\chi}{2}\right) + \ln\left(\dfrac{1+\chi^2}{2}\right) - 2\tan^{-1}\chi + \dfrac{\pi}{2} & \zeta < 0 \\ -b\zeta & 0 \leq \zeta \leq 1 \\ (1-b)(1+\ln\zeta) - \zeta & \zeta > 1 \end{cases}, \tag{29}$$

$$\Psi_t(\zeta) = \Psi_q(\zeta) = \begin{cases} \ln\left(\dfrac{1+\chi^2}{2}\right) & \zeta < 0 \\ -b\zeta & 0 \leq \zeta \leq 1 \\ (1-b)(1+\ln\zeta) - \zeta & \zeta > 1 \end{cases}, \tag{30}$$

with $\chi = (1 - a\zeta)^{1/4}$, $a = 16$ and $b = 5$ are the stability-dependent correction functions. The computation of the stability functions requires an a priori assumption (Munro, 1989) about the L which in turn depends on q_{sh} and the friction velocity

15   $$u_* = \frac{\kappa u_{z_v}}{\ln\left(\dfrac{z}{z_{0_v}}\right) - \Psi_m(\zeta)}. \tag{31}$$

COSIPY uses an iterative approach to resolve the dependency of these variables. At the beginning of the first iteration u_* (Eq. 31) and q_{sh} (Eq. 17) are approximated assuming a neutral stratification ($\zeta = 0$). These quantities are then used to calculate L (Eq. 24). In the next iteration, the updated L is then used to correct u_* and q_{sh} . The iteration is repeated until either the changes in q_{sh} are less than $1\,e{-}2$ or a maximum number of 10 iterations is reached. As already shown by other studies, the

5   algorithm usually converges in less than 10 time steps (Munro, 1989).

3.5 Snow densification

Snow grain settling during metamorphism and compaction under the weight of the overlying snowpack generally increases the snow density over time (Anderson, 1976; Boone, 2004). The snow density is a key characteristic of the snowpack, which is used by COSIPY to derive important snow properties such as thermal conductivity and liquid water content. Assuming that a

10   rapid settlement of fresh snow occurs simultaneously with slow compaction by the load resisted by the viscosity, the rate of density change of each snow layer becomes

$$\frac{1}{\rho_s(z,t)}\,\frac{d\rho_s(z,t)}{dt} \;=\; \frac{M_s(z,t)\,g}{\eta(z,t)} \;+\; c_1\,\exp\left[-c_2(T_m - T_s) - c_3\max\left(0, \rho_s(z,t) - \rho_0\right)\right], \tag{32}$$

with M_s is the overlying snow mass, $c_1 = 2.8 \times 10^{-6}\,\mathrm{s}^{-1}$, $c_2 = 0.042\,\mathrm{K}^{-1}$, $c_3 = 0.046\,\mathrm{m}^3\,\mathrm{kg}^{-1}$, and the viscosity

$$\eta(z,t) = \eta_0 \exp\left[c_4(T_m - T_s) + c_5\rho_s\right] \tag{33}$$

15   where $\eta_0 = 3.7 \times 10^7\,\mathrm{kg\,m^{-1}\,s^{-1}}$, $c_4 = 0.081\,\mathrm{K}^{-1}$, and $c_5 = 0.018\,\mathrm{m}^3\,\mathrm{kg}^{-1}$ (Anderson, 1976; Boone, 2004).

3.6 Mass changes

The total mass changes may be written as the integral expression

$$\frac{\partial}{\partial t}\int_0^d \rho_s\,dz = \frac{\partial}{\partial t}\int_0^d \theta_i(z,t)\rho_i\,dz + \frac{\partial}{\partial t}\int_0^d \theta_w(z,t)\rho_w\,dz + \frac{\partial}{\partial t}\int_0^d \theta_a(z,t)\rho_a\,dz, \tag{34}$$

which follows directly from Eq. (2). So any net mass change must be accompanied by changes in ice fraction, liquid water

20   content, and porosity within the snow/ice column of height d. The continuity equation for ice fraction (first term on the right side) may be written as

$$\frac{\partial}{\partial t}\int_0^d \theta_i(z,t)\rho_i\,dz = \frac{\partial}{\partial t}\int_0^d \Delta\theta_i(z,t)\rho_i\,dz + SF - \frac{q_m}{L_f} + \frac{q_{lh}}{L_s} + \frac{q_{lh}}{L_v}, \tag{35}$$

where the integral on the right side describes the internal mass changes by melt and refreezing, SF the mass gain by snowfall, and q_m/L_f is the mass loss by melt. The last two terms of this equation, q_{lh}/L_s and q_{lh}/L_v, are the sublimation/deposition and

evaporation/condensation fluxes at the surface, respectively, depending on the sign of q_{lh} and $T_s(z=0,t)$. Melt energy q_m is the energy surplus at the surface which is available for melt, and follows from Eq. (4). Similarly, we can extend the continuity equation for the liquid water content which reads as

$$\frac{\partial}{\partial t}\int_0^d \theta_w \rho_w(z,t)\,dz = \frac{\partial}{\partial t}\int_0^d \Delta\theta_w(z,t)\rho_w\,dz + R_f + \frac{q_m}{L_f} + \frac{q_{lh}}{L_v} - Q \qquad (36)$$

5   with the integral on the right side describing the internal mass changes of liquid water by melt and refreezing, R_f the mass gain by liquid precipitation, and Q the runoff at the bottom of the snowpack. COSIPY calculates all terms and writes them to the output file.

4 Model architecture

Basically, the COSIPY model consists of a model kernel which is extended by modules. The model kernel forms the underlying
10   model structure and provides the IO routines, takes over the discretisation of the computational mesh, parallelizes the simulations, and solves the fundamental mass- and energy conservation equations Eq. (2) and Eq. (3). These tasks are independent of the implementations of the parametrization and usually, do not require any modification by the end-user.

COSIPY is a one-dimensional model that resolves vertical processes at a specific point on the glacier. For spatially distributed simulations, the point model is integrated independently at each point of the glacier domain, neglecting the lateral mass and
15   energy fluxes. The independency of the point models simplifies scaling for larger computer architectures, which led to the COSIPY model architecture being designed for both local workstations and High-Performance Computing Cluster (HPCC). So far, the model has been successfully tested on Slurm Workload Manager (https://slurm.schedmd.com) and PBS job scheduling systems (https://www.pbspro.org). Regardless of whether the distributed simulations are integrated on a single-core or multi-core computing environment, the point model sequence is always the same. During initialisation, the atmospheric input data
20   is read in, and the dynamic mesh is generated. With distributed spatial simulations, the data is distributed across the available cores, and one-dimensional calculations are performed for each grid point.

At the beginning of each time step, it is checked whether snowfall occurs and must be added to the existing snow cover. Subsequently, the albedo (Eq. 11) and the roughness length are updated. Afterwards, the computational mesh is re-meshed to ensure numerical stability. Once these steps have been performed, the heating and melting of snow by penetrating short-
25   wave radiation (Eq. 13, 6, and 7) is determined and the surface energy fluxes and surface temperature (Eq. 4) are derived. The resulting meltwater, both from surface and subsurface melt, is then percolated through the layers (bucket approach). Next the heat equation (Eq. 3) is solved after all terms on the right side have been determined.

4.1 Input and Output

The model is driven by meteorological data that must be provided in a corresponding NetCDF file (see https://cosipy.readthedocs.io/en/latest/Ressources.html). Input parameters include atmospheric pressure, air temperature, cloud cover fraction, relative humidity, incoming shortwave radiation, total precipitation and wind velocity. Optional snowfall and incoming longwave radi-
5 ation can be used as forcing parameters. In addition to meteorological parameters, COSIPY requires static information such as topographic parameters and a glacier mask. Various tools are available to create and convert the static and meteorological data into a corresponding input file. An example workflow for creating the required NetCDF input is included in the source code. Besides the standard output variables, there is also the possibility to store vertical snow profile information (not recommended for distributed simulations). To reduce the amount of data, the users can specify which of the output variables will be stored.

10 ## 5   Example - Zhadang glacier, High mountain Asia

To illustrate a model application, we show a mass- and energy balance simulation of the Zhadhang glacier, which is located on the north-eastern slope of the Nyainqentanglha Mountains (30°28.2'N, 90°37.8'E) on the Central Tibetan Plateau. We use hourly data from May 2009 to June 2012 from an automatic weather station (AWS) on the Zhadhang Glacier (Huintjes et al., 2015). The relevant variables air pressure p_{z_t}, air temperature T_{z_t}, relative humidity rH_{z_t}, incident short-wave radiation q_G
15 and wind speed u_{z_v} were measured by the AWS. Due to the harsh and remote environment, the time series show gaps that were filled with the High Asia Refined Analysis (HAR; Maussion et al., 2014) product. The cloud cover fraction N was provided by ERA5 (Hersbach and Dee, 2016) data. We compare the simulated snow temperature T_s and surface height change ΔH with the AWS measurements. Furthermore, ablation stakes were installed on the glacier to determine the loss of mass at various locations on the glacier. A detailed description of the data, the AWS sensors used, the post-processing procedure and the
20 discussion can be found in Huintjes et al. (2015) and Huintjes (2014).

Simulation. Figure 1a and 1b show the glacier surface temperatures for two periods where in-situ measurements were available. The model represents both the daily variability (R^2 0.83, p-value < 0.001) and the magnitude of the observed surface temperature. The root mean square error is 2.9 K and 2.7 K for the two periods, respectively, and is thus within the typical uncertainty range of long-wave radiation measurements. The modelled cumulative mass balance over the entire period
25 from October 2009 to May 2012 is -6.71 m w.e. The most negative ablation stake shows a negative surface mass balance of -6.4 m w.e, and thus the modelleld MB is slightly more negative over the entire study period. Figure 1c and 1d show the modelled and measured ΔH for the two periods October 2009 to June 2010 and October 2011 to May 2012 where measurements are available. The daily and seasonal variability is well captured by the model ($R^2 = 0.78$, p-value < 0.001 and $R^2 = 0.61$, p-value < 0.001), even if snowfall seems to be too low during the winter months. Nevertheless, overall the differences are consistently
30 small with RMSE of 0.07 m and 0.11 m. The glacier-wide cumulative surface mass balance for the decade 2009 to 2018 is presented in Figure 2. The model was driven by ERA5 data instead of in-situ observations. The computational domain consisted of 1837 grid cells with a spatial resolution of approximately 30 m (1 arcsecond) (see Fig. 2). The modelled total mass loss during this period was $-13.9\ m\ w.e.$. The distributed model shows a lower mean annual mass loss ($-0.7\ m\ w.e.\ a^{-1}$, only

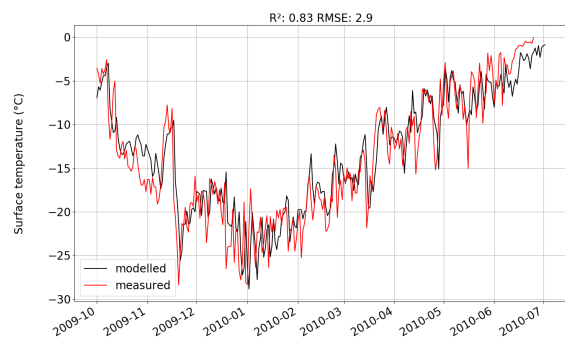
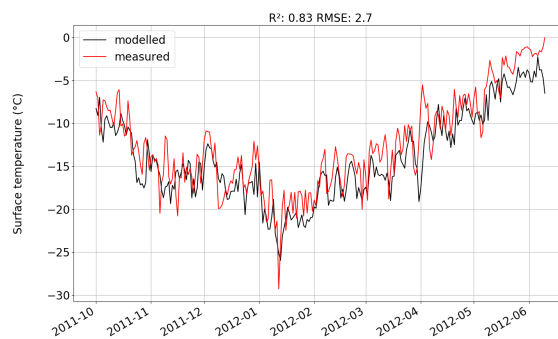
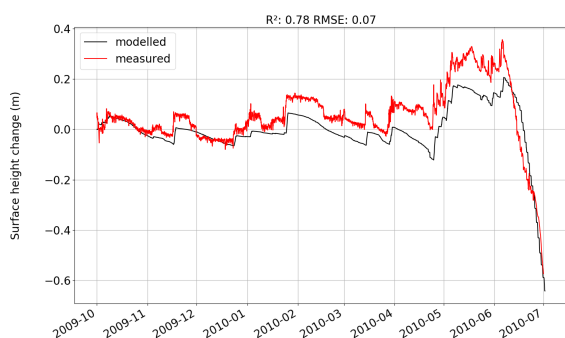
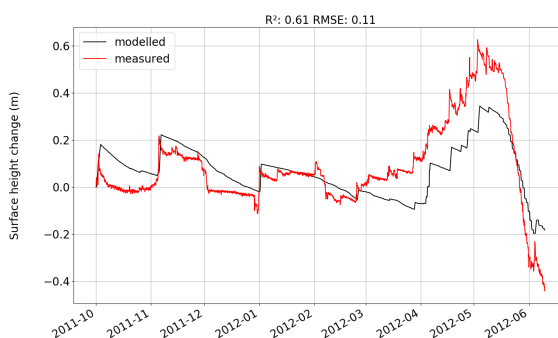
May 2009 to June 2012 considered) than the point model ($-2.6 \ m \ w.e. \ a^{-1}$) driven by the in-situ observations from May 2009 to June 2012 at the position of the AWS. The results are in line with the analysis of Qu et al. (2014) who reported negative mass balances of $-1.9, -2.0, -0.8$ and $-2.7 \ m \ w.e$ for the years 2009 to 2012.

(a) Daily surface temperature October 2009 until June 2010

(b) Daily surface temperature October 2011 until May 2012

(c) Hourly surface height change October 2009 until June 2010

(d) Hourly surface height change October 2011 until May 2012

Figure 1. Modelled and measured surface temperatures and height changes at the location of the automatic weather station at the Zhadang glacier.

Scalability. A big challenge for large applications is usually the computational cost. To achieve the required performance,
5 models should be scalable on parallel high-performance computing environments. For the model performance analysis, we use a cluster with identical nodes, each consisting of two Intel Xeon(R) E5-2640 v4 CPUs operating at 2.4 GHz and connected via InfiniBand. Each processor has ten cores, 32 GB memory and a memory bandwidth of 68.3 GB/s. To test the performance of the parallelized COSIPY version, we performed a spatial simulation of the Zhadhang glacier. We used a 3 arcsecond ($\sim$ 90 m) Shuttle Radar Topography Mission (SRTM) terrain model so that the computational grid consists of 206 points. The
10 performance of the parallel version was then compared to the single-core solution by measuring the required execution time for different core setups (1-220 cores). Figure 3 shows the speedup compared to the single-core version, i.e. the ratio of the original execution time with the execution time of the corresponding node test. If the model is executed with 20 cores, the speedup is ~ 2. With 120 cores a speedup of ~ 10 is reached, i.e. each core has to calculate a maximum of two grid points. A

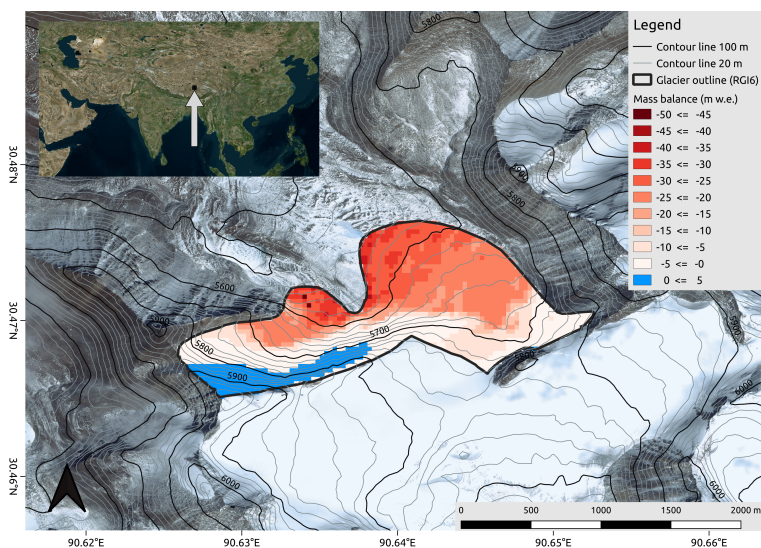



Figure 2. Cumulative climatic surface mass balance from 2009 to 2018 Zhadang glacier with 1837 grid points, contour lines (SRTM), glacier outline from Randolph Glacier Inventory 6.0 and a topographic map from Bing Maps (© Microsoft, 2020).

speedup of more than ∼ 16 is not possible with this system and is achieved with a number of 220 cores (more cores than grid points). The computation time is less than 35 minutes for a ten year period (hourly resolution) when using 220 cores. At this point, it should be mentioned that the performance can vary significantly on other HPCC systems and simulation conditions and should always be checked before submitting larger simulations to the cluster.

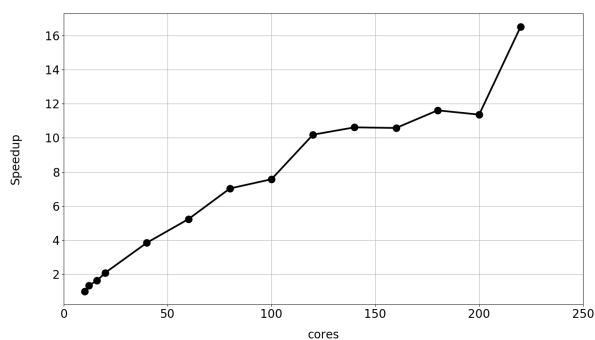



Figure 3. Speedup for computing a 10-year distributed COSIPY run on Zhadang glacier with 206 grid points.

6 Conclusions

Model frameworks for calculating the climatic surface mass balance of glaciers based on energy exchange at the surface of snow and ice have been available since the last decades of the 20th Century. The approaches commonly use either meteorological observations or modelled atmospheric data. The energy available for melting snow and ice mostly results as the residual term of the energy balance equation. The energy balance at the surface and the energy fluxes within the snow or ice volume are coupled via the surface properties, in particular the surface skin temperature. Since snow and ice surfaces cannot exceed the melting point temperature, any energy excess must be transferred to the phase change from solid to liquid water, respectively snow or ice melt. COSIPY is an open-source and fully documented community model of its kind. The model has a modular structure facilitating the exchange of routines or parametrizations for individual processes with only little effort. The open design of COSIPY, its well-documented and its modular approach allows for joint community-driven advancement of the model in the future.

COSIPY may serve to analyse inter-annual and intra-annual variations of energy and surface mass balance of glaciers. Further, it allows identifying sensitivities, non-linearities, co-variances, and tipping points in the components of glacier surface energy and mass fluxes resulting from the variability of atmospheric forcing. Since it is notoriously difficult to obtain sufficiently reliable atmospheric forcing data, the uncertainties in climatic mass balance estimates derived only from modelling can be quite high. Therefore, it is recommended to use observations of mass balance either from fieldwork or from remote sensing data analysis to benchmark model results whenever possible.

Future improvements of COSIPY are expected by applying the model in different climates and varying topographical settings. Additional processes affecting the climatic mass balance of glaciers such as debris cover, snowdrift and avalanches can be considered in further developments of the model.

7 Code availability, documentation, and software requirements

COSIPY is based on the Python 3 language and is provided on a freely accessible git repository (https://github.com/cryotools/cosipy, last access: January 20, 2020). COSIPY can be used for non-profit purposes under the GPLv3 license (http://www.gnu.org/licenses/gpl-3.0.html). Scientists can actively participate in model development. A documentation with a sample workflow, information about input/output formats and the code structure is available under 'Read the Docs' (https://cosipy.readthedocs.io/en/latest/index.html, last access: January 20, 2020). As a community platform and user support, we use the groupware Slack (https://cosipy.slack.com, last accessed: January 20, 2020). The various official model releases will be registered with a unique DOI on Zenodo (https://doi.org/10.5281/zenodo.2579668, last access: January 20, 2020). For the result of this publication the version v1.2 (https://doi.org/10.5281/zenodo.3613921) was used. Each commit will be automatically tested with different Python 3 releases on Travis (https://travis-ci.org/cryotools/cosipy, last accessed January 20, 2020). The tested code coverage is tracked on CodeCov (https://codecov.io/github/cryotools/cosipy/, last accessed January 20, 2020). Since we have just started writing the tests, code coverage of 35 % is still low but will be increased in the near future. With the exception of the pre-processor for creating the static file (currently not working on Windows systems) the model should work on any operating

system with Python 3 installed. However, support for operating systems other than Linux-based systems is limited because we develop and run COSIPY exclusively on Linux-based systems. COSIPY is built on the following open-source libraries: numpy (van der Walt et al., 2011), scipy (Virtanen et al., 2019), xarray (Hoyer and Hamman, 2017), distribued, dask_jobqueue (Dask Development Team, 2016), and netcdf4 (https://doi.org/10.5281/zenodo.2669496).

Appendix A: List of symbols

Constant	Description	Unit	Default value
c_p	specific heat of air	$J\ kg^{-1}\ K^{-1}$	1004.67
c_i	specific heat of ice	$J\ kg^{-1}\ K^{-1}$	2050.0
c_w	specific heat of water	$J\ kg^{-1}\ K^{-1}$	4217.0
d^*	albedo depth scale	cm	3.0
g	gravitational acceleration	$m\ s^{-2}$	9.81
k_a	thermal conductivity of air	$W\ m^{-1}\ K^{-1}$	0.026
k_i	thermal conductivity of ice	$W\ m^{-1}\ K^{-1}$	2.25
k_w	thermal conductivity of water	$W\ m^{-1}\ K^{-1}$	0.6089
Pr	turbulent Prandtl number	$-$	0.8
T_m	melting point temperature	K	273.16
α_s	fresh snow albedo	$-$	0.9
α_f	firn albedo	$-$	0.55
α_i	ice albedo	$-$	0.3
ϵ_s	surface emissivity	$-$	0.99
η_0	snow viscosity	$kg\ m^{-1}\ s^{-1}$	3.7×10^7
κ	Von Karman constant	$-$	0.41
ρ_a	air density	$kg\ m^{-3}$	1.1
ρ_w	water density	$kg\ m^{-3}$	1000.0
ρ_i	ice density	$kg\ m^{-3}$	917.0
ρ_0	snow compaction parameter	$kg\ m^{-3}$	150.0
σ	Stefan–Boltzmann constant	$W\ m^{-2}\ K^{-4}$	5.67×10^8
τ^*	albedo time scale	$days$	22

Variable	Description	Unit
c_s	specific heat of snow	$J\ kg^{-1}\ K^{-1}$
C_D	bulk transfer coefficient for momentum	$-$
C_E	bulk transfer coefficient for latent heat	$-$
C_H	bulk transfer coefficient for sensible heat	$-$
e_{z_t}	water vapour pressure at height z_t	Pa
Ew_{z_t}	saturation water vapour at height z_t	Pa
$Ew_{z_{0_t}}$	saturation water vapour at the surface	Pa

k_s	thermal conductivity of snow	$W\ m^{-1}\ K^{-1}$
L	Obukhov length	m
L_s	latent heat of sublimation	$J\ kg^{-1}$
L_f	latent heat of fusion	$J\ kg^{-1}$
L_v	latent heat of vaporisation	$J\ kg^{-1}$
ME	available melt energy	$W\ m^{-2}$
M_s	overlying mass	kg
N	cloud cover fraction	$-$
q_{lw}	net longwave radiation	$W\ m^{-2}$
$q_{lw_{in}}$	incoming longwave radiation	$W\ m^{-2}$
$q_{lw_{out}}$	outgoing longwave radiation	$W\ m^{-2}$
q_{sw}	net shortwave radiation	$W\ m^{-2}$
q_{sh}	sensible heat flux	$W\ m^{-2}$
q_{lh}	latent heat flux	$W\ m^{-2}$
q_{rr}	heat flux from rain	$W\ m^{-2}$
q_m	melt energy	$W\ m^{-2}$
q_0	mixing ratio at the surface	$kg\ kg^{-1}$
q_{z_q}	mixing ratio at height z_q	$kg\ kg^{-1}$
p_{z_t}	air pressure at height z_t	hPa
Q	runoff	$mw.e.$
Q_p	volumetric energy sink/source by melting and refreezing	$W\ m^{-3}$
Q_r	volumetric energy surplus by absorption of shortwave radiation	$W\ m^{-3}$
rH_{z_t}	relative humidity at height z_v	$\%$
Ri_b	Bulk Richardson number	$-$
SF	snowfall	m
T_s	snow temperature	K
T_v	virtual air temperature	K
T_{z_t}	air temperature at height z_t	K
T_0	surface temperature	K
u_{z_v}	wind speed at height z_v	$m\ s^{-1}$
u_*	Friction velocity	$m\ s^{-1}$
z_t	measurement height of temperature	m
z_q	measurement height of humidity	m
z_t	measurement height of wind velocity	m
z_{0_v}	aerodynamic roughness length	m

z_{0_t}	roughness length for temperature	m
z_{0_q}	roughness length for specific humidity	m
α	snow/ice albedo	$-$
ϵ_{cl}	emissivity of clouds	$-$
ϵ_{cs}	clear sky emissivity	$-$
ϵ_a	total atmospheric emissivity	$-$
η	snow viscosity	$kg\ m^{-1}\ s^{-1}$
θ_w	liquid water content	$-$
θ_a	air porosity	$-$
θ_i	volumetric ice fraction	$-$
θ_e	irreducible water content	$-$
Θ	local slope	$^\circ$
λ_r	fraction of absorbed radiation	$-$
ρ_s	snow density	$kg\ m^{-3}$
Ψ_{Ri}	stability function based on the Richardson-Number	$-$
Ψ_m	stability function for momentum based on the Monin-Obukhov similarity theory	$-$
Ψ_t	stability function for heat based on the Obukhov length	$-$
Ψ_q	stability function for moisture based on the Obukhov length	$-$

Author contributions. TS and CS are the initiators of the model. TS developed the model design and wrote most of the sections on the physical and numerical principles of the model. AA developed parts of the parameterizations and core classes, applied the model to the Zhadang glacier and wrote the sections about the application and code availability. AA and TS were equally involved in the development of the documentation and maintenance of the Community Platform (Slack). During the entire development process, all authors discussed the
5 individual steps of model development, the results and the structure of the manuscript.

Competing interests. The authors declare that they have no conflict of interest.

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