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Mathematical Modeling and Microphysics of Snow Distribution

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Abstract

Snow cover plays a significant role in climate, hydrology, and ecological systems through its influence on the surface energy balance. Due to its excellent radiation reflection capacity, snow immediately affects the equilibrium between the energy absorbed by the Earth and the energy reflected in the atmosphere. The planet cools down as 80% *and* 90% *of the total sunlight that strikes its surface is reflected in the atmosphere. In addition, snow serves as many regions of the world's primary water source. The world's one-sixth of the population depends on snowmelt water for survival. Due to global warming, the traditional pattern of snowfall and snowmelt has changed. To better understand how to utilize the snow, it is necessary to simulate how this snow is dispersed precisely. In this study, we survey articles that have touched extensively on the snow distribution model, why we model snow distribution, and some of the tools to implement this model. Also, current gaps in the modeling of snow distribution are discussed, and some open questions are presented.*

1 Introduction

The cryosphere comprises areas of the Earth where snow and ice cover varies with latitude, altitude, and season. It comprises regions with seasonal snow cover, oceans, rivers, glaciers, sea ice, and permafrost. The cryosphere covers a mean total area of $70 \times 10^6 km^2$ about 14% of the Earth's surface area Sharma et al. (2021).
Average snowfall and global ice cover have decreased in many regions over the last century due to rising average temperatures caused by the emission of greenhouse gases as a consequence of human activities. This decrease in snow and ice cover results in changes that affect our ecosystems and climate due to various Earth system processes, the extent of snow and ice cover has also varied dramatically in the past, from tens of thousands to millions of years ago.
Due to its strong ability to reflect radiation (high albedo) and its thermal characteristics, snow is a crucial part of the Earth's climate system because it directly affects the equilibrium between the energy the Earth absorbs and the energy it reflects into the atmosphere. Snow reflects about 80% to 90% of the total sunlight hitting its surface, which regulates the planet's temperature. Also, snow acts as an insulating blanket to the ground beneath it by preventing heat and moisture from escaping into the atmosphere, and it reduces the risk of wildfire.
The term "snow cover" describes the blanket of snow that covers the ground and incorporates the ideas of snow depth and area. Seasonal snow cover is the cryosphere's most significant single component in terms of area, covering an average of $46 \times 10^6 km^2$ of Earth's surface (31% of the land area) annually. It is, therefore, an integral piece and driver of the Earth's climate through its involvement in changing energy and moisture fluxes between the surface and the atmosphere and its function as a water reserve in hydrological systems Brown and Robinson (2005). The intense interaction between the atmospheric boundary layer and land-surface via energy and mass exchange processes makes the distribution of snow a vital element of the Earth system Roesch et al. (2001) with significant consequences for the hydrological cycle and the climate of cold regions Beniston et al. (2018). The mass and energy flow between places with and without snow show a striking disparity due to snow's highly distinctive material characteristics, which include: albedo, thermal conductivity, latent heat of phase change, and mechanical roughness Sharma et al. (2021).

For a continuous snow cover, Snowpack energy balance can be written as

$$-\frac{dH}{dt} = Q_S + Q_L + Q_h + Q_e + Q_a + Q_G \tag{1}$$

where:

- $-\frac{dH}{dt}$ is the net change rate of the snowpack internal energy per unit area.
- Q_S is net shortwave radiation (incoming minus reflected shortwave radiation);
- Q_L is the net longwave radiation (downward and upward component of longwave radiation);
- Q_h is the turbulent flux of sensible heat exchanged at the surface due to the temperature gradient between snow surface and atmosphere;
- Q_e is the turbulent flux of latent heat exchanged between the surface and the overlying atmosphere due to water vapor pressure differences; it represents the sublimation and evaporation from and the condensation to the snow surface and is thus directly connected with the mass balance of the snow cover;
- Q_a is the flux of energy advected via precipitation or blowing snow.

- Finally, Q_G is the ground heat flux due to conduction.

Armstrong and Brun (2008) in their work gave the equation for the mass balance of the snow cover, including blowing snow as

$$\frac{dM}{dt} = P - \nabla \cdot D_{bs} - E_{bs} \pm E - R \tag{2}$$

where,

- $\frac{dM}{dt}$ is the snowpack mass change rate;
- P is the precipitation rate.
- D_{bs} is the horizontal blowing snow transport due to redistribution of surface snow (mass per unit length per unit time)
- E_{bs} the rate of sublimation of blowing snow.
- E is the sum of sublimation or evaporation (loss of mass) or condensation (gain of mass) rates at the surface.
- R is the runoff rate (liquid water leaving the bottom of the snowpack) and negatively contributes to the mass balance.

In 1995, Matthew Sturm and Jon Holmgren proposed a new classification system for seasonal snow covers. Each of these classes was defined by a unique ensemble of textural and compositional characteristics, such as the sequence of snow layers, their thickness, density, crystal morphology, and grain characteristics within each layer. These classes include tundra, taiga, alpine, maritime, prairie, and ephemeral. Sturm et al. (1995).
Snowfall is an essential component of solid precipitation in the alpine region and a vital part of mountain hydrological processes during the cold season. It is also regarded as a climatic change indicator due to its high sensitivity to temperature and precipitation changes. Schirmer et al. (2011) in their work described that snow distribution in the alpine terrain is strongly variable and depends on the mountain's slope elevation, steepness, slope direction, and wind exposure. This strong variability results from processes occurring during snow accumulation and melting. It has consequences on the evolution of local avalanche danger, the hydrological response of mountainous area, and alpine ecosystem developments Vionnet et al. (2017). Liston et al. (2007) in their work found that at regional scales, the spatial variability during snow accumulation is governed by three main processes:

- Orographic snowfall,
- Preferential deposition of falling snow and
- Wind-induced snow transport of snow on the ground during or after snowfall.

Wind-induced snow transport occurs in regions seasonally or permanently covered by snow when the wind speed exceeds a threshold value that depends on the snow type at the surface Guyomarc'h and Mérindol (1998). In the alpine terrain, snow transport creates in-homogeneous snow depth distribution, strongly influenced by the local topography Durand et al. (2005); Mott et al. (2013). Snow is eroded in areas exposed to strong wind (crest, for example) and is deposited in areas sheltered from the wind.
In 1992, Kosugi et al, described the three main transport modes for snow as the following:

- Reptation (creep) corresponds to the rolling of particles over the snowpack's surface, and its contribution is negligible compared to the other processes. It is commonly neglected in blowing snow models.

- Saltation - this corresponds to the transport of snow crystals in a layer close to the ground (typical thickness: 5 – 10 cm). In saltation, particles follow ballistic trajectories in a shallow layer close to the ground. They may rebound and eject new grains when returning to the surface Kosugi et al. (1992). Aerodynamic and splash entrainment were the two primary routes for transporting snow particles in the saltation layer in the 2000s Nishimura and Hunt (2000). Aerodynamic entrainment happens when the wind has enough momentum to carry snow particles from the surface. Splash entrainment is the movement of arriving grains that are already in motion and contact the snow surface before rebounding or projecting other grains into the saltation layer Nishimura and Hunt (2000).
- Turbulent suspension occurs in the atmosphere above the saltation layer, where turbulent eddies transport snow grains without contact with the surface. Distances of transport are limited by the sedimentation and sublimation of snow grains. The latter process modifies the vertical temperature and humidity profiles in the surface boundary layer Schmidt (1982); Déry and Yau (2001).

In the paper written by Sharma et al. (2021), they referred to the combination of saltation and suspension as the blowing snow. This blowing snow was a significant ingredient in forming the blowing snow model used to couple WRF and SNOWPACK to create CRYOWRF, one of the newest numerical models for snow. This blowing snow scheme used lately in CRYOWRF follows closely with what was implemented as the blowing snow dynamics in Meso-NH by Lafore et al. (1998) and Déry and Yau (2001), with the difference in this implementation in the details of calculating terminal fall velocities and threshold friction velocity for snow transport.

The blowing snow model implemented in CRYOWRF is a double-moment scheme that solves predictive equations for the mass and number mixing ratios of blowing snow particles as seen in equation 3 and equation 4 respectively. These equations are essentially Eulerian advection-diffusion type equations along with sublimation and vapor deposition phase changes.

The equations in Einstein notation for q_{bs} and N_{bs} respectively are

$$\frac{\partial q_{bs}}{\partial t} + u_i \frac{\partial q_{bs}}{\partial x_i} = -K_{bs} \frac{\partial^2 q_{bs}}{\partial x_i^2} \delta_{i3} + \frac{\partial}{\partial x_i}(q_{bs} V_q \delta_{i3}) + S_q, \tag{3}$$

$$\frac{\partial N_{bs}}{\partial t} + u_i \frac{\partial N_{bs}}{\partial x_i} = -K_{bs} \frac{\partial^2 N_{bs}}{\partial x_i^2} \delta_{i3} + \frac{\partial}{\partial x_i}(N_{bs} V_N \delta_{i3}) + S_N, \tag{4}$$

where $u_i = 1,2,3 = (u, v, w)$ are the zonal, meridional and vertical velocity components of the air respectively. K_{bs} is the turbulent diffusion coefficient for blowing snow particles, V_q and V_N are the mass and number-weighted terminal fall velocities of blowing snow particles and the S_q and S_N are sink (source) terms to account for sublimation (deposition) of blowing snow particle.

With this study, our goal is to survey and review previous articles which have simulated and modeled snow cover, the microphysics involved, and how the wind is necessary for snow transport. To answer the question of what modeling snow distribution entails, why we need to model snow distribution, and what are the possible outcomes and challenges in snow modeling and distribution.

2 Snow Distribution scales

In 2018, Mott. et al. discussed wind-driven snow processes by shaping the snow accumulation patterns at different scales of a mountainous area. The snow process scales included the following, as shown in the figure 1:

- The mountain range scale,
- The mountain-slope scale, and
- The mountain-ridge scale.

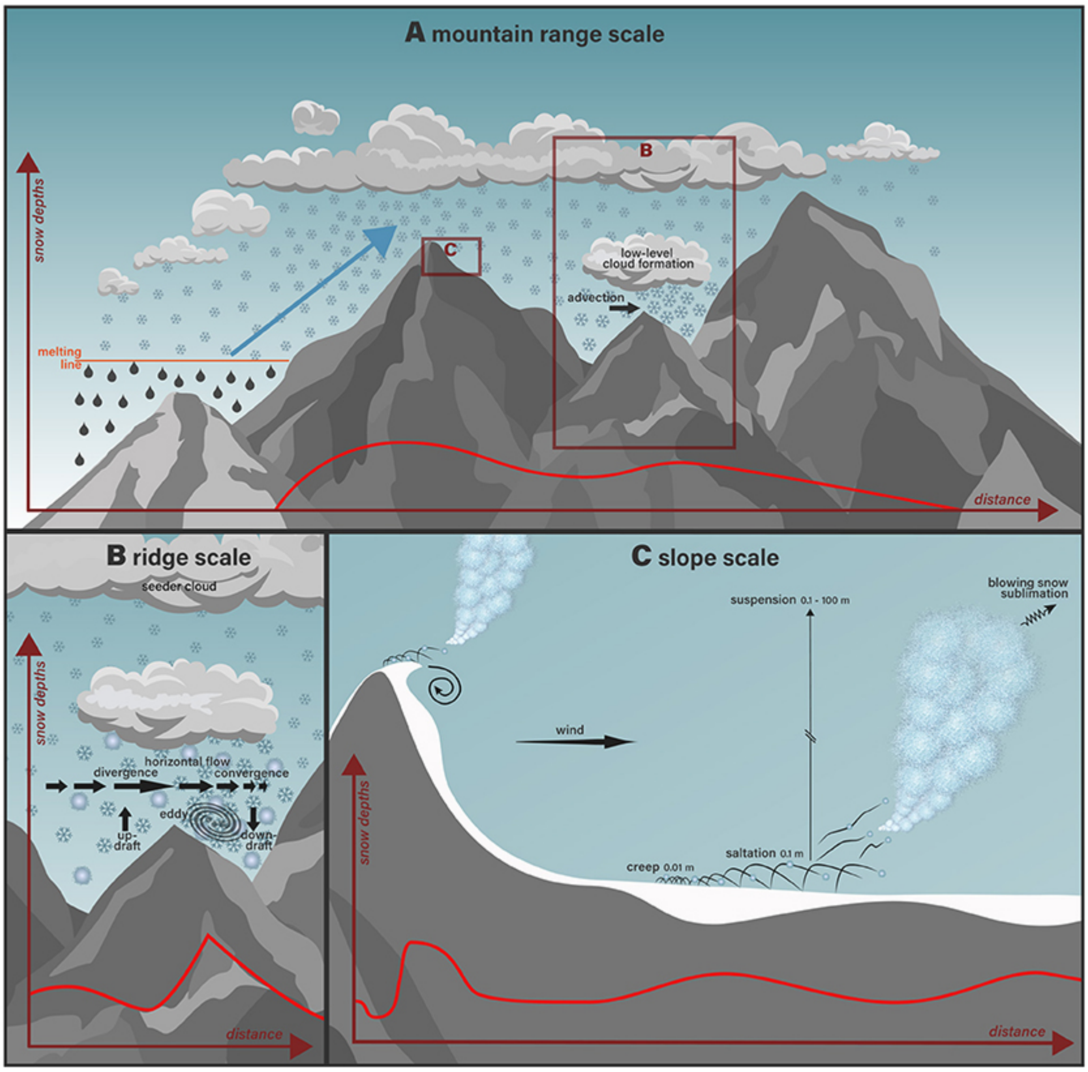

Figure 1: Source Mott et al. (2018). Schematic description of snow accumulation processes acting on different scales: the Mountain range scale (A), the ridge scale (B), and the slope scale (C).

At the mountain-range scale in figure 1A, snow accumulation depends on climate, elevation, and vegetation. Mountain-range scale precipitation patterns are primarily influenced by hilly precipitation, which captures all processes associated in mountainous terrain with regional precipitation patterns where the interaction of the atmospheric flow with the underlying topography results in regions of enhanced or reduced snowfall Colle et al. (2013).

The primary driving process in this scale is the forced dynamical lifting of air masses leading to cooling of the air column and resulting in condensation and precipitation and a phase change from rain to snow above the zero-degree elevation band. Dynamical and cloud micro-physical trends tend to make leeward slopes drier than windward slopes. Such regional trends of decreasing precipitation are typically aligned with the prevailing synoptic wind direction Gerber et al. (2018).

Contrary to what was observed for large mountain ranges, ridge scale snow accumulation patterns with a height range from (hundreds to thousands of meters) as seen in figure 1B, typically reveal much larger spatial variability of snow and enhanced snow deposition over leeward slopes of single mountain peaks Lehning and Fierz (2008);Mott et al. (2010);Vionnet et al. (2017). At higher model resolutions (20m and less), Mott et al. (2010) and Gerber et al. (2018) suggested that preferential deposition of snowfall mainly drives spatial variability of snow depths at the ridge scale of about (hundreds to thousands of meters), causing snow loading on leeward slopes of mountain ridges and reduced snow deposition on the windward slopes as seen in the slope scale in figure 1C. Snow accumulation patterns at the ridge scale figure 1B, driven by preferential deposition, are superimposed by snow drift processes (acting on the scales of a few to hundreds of meters) and snow avalanches.

Snow redistribution methods, such as saltation and turbulent suspension, are dominant drivers of snow accumulation patterns at the slope scale and the lower range of the ridge scale, shaping snow deposition patterns across a wide scale range of a few meters to hundreds of meters in various environments for example in the alpine, arctic, and prairies. Snow loading on leeward slopes results in more homogeneous snow depth distributions.Mott et al. (2010). Snow erosion by wind is most commonly observed in flat areas, such as the prairies or over frozen lakes, where a good fetch exists to establish wind erosion or in wind-exposed areas of varying sizes (from local bumps to large ridge crests). Snow preferentially tends to fall in the lee of topographic disturbances such as ridges or local depressions, resulting in snow deposition features such as snow dunes, drifts, cornices, and trough filling as seen in figure1C.

Mott et al. (2018) addressed the current state of knowledge on wind-driven snow transport, interactions between snowfall and atmospheric flow, snow-mass loss and feedbacks on the atmosphere due to sublimation (surface and blowing snow), and heat exchange processes over continuous and patchy snow covers. They pointed to the strong scale dependency of snow-depth variability and dominant wind-driven processes affecting snow accumulation. Comparing the scales used in their paper, they discovered that, on the mountain-range scale, snow depth patterns might mostly be attributable to orographic precipitation patterns, with elevation as the most crucial influence. Conversely, snow patterns on the ridge scale are primarily influenced by the interaction of the local wind with snowfall and the snow surface.

Small-scale updrafts produce local cloud-formation processes, typically forming distinct patterns of snowfall enhancement in the summit region. Snowfall is also preferentially deposited over leeward mountain slopes due to downstream advection of snow particles. Streamwise flow convergence and increased snow deposition velocities lead to increased snow-deposition rates. Finally, local solid winds drive different snow erosion and deposition patterns that shape the snow cover at the ridge and slope scales. These strong winds promote the redistribution of snow by saltation and suspension.

The physical characteristics of the surface snow, as well as the roughness of its surface, are significantly influenced by wind-induced snow transport. The physical characteristics of the surface snow, as well as the roughness of its surface, are significantly influenced by wind-induced snow transport. Various attempts have been made to predict snow accumulation at various scales. Most precipitation studies use horizontal grid resolutions of 1 km and coarser to simulate precipitation patterns at the regional scale (mountain-range scale). However, based on research, ridge-scale precipitation and accumulation patterns require better model resolutions of 50 m or less to illustrate how local flow fields affect precipitation patterns. Even higher model resolutions of 5 m and less are needed to resolve wind-induced snow-redistribution processes. Various model approaches can be categorized into two main groups to simulate snow-depth fluctuation at the ridge and slope scales. The categories include dependence on vertically integrated snow -models and transport rates for 3D turbulent blown snow particle diffusion equations in the atmosphere Mott et al. (2018).

Bernhardt et al. (2010) showed that the final pattern of snow accumulation on the ground is strongly influenced by post-depositional processes, which include wind-induced snow transport during drifting and blowing snow events and redistribution by avalanches that may occur on steep slopes.

3 Microphysics

Despite random "inconsistent" forecasts, numerical weather prediction of winter precipitation has been steadily improving over the past two decades. Advancing technology allows increasingly sophisticated physics to be incorporated into numerical models of increasing resolution, either by an explicit bin-resolving cloud model or bulk microphysical parameterizations for modeling cloud microstructure Thompson et al. (2004). Bin models are costly in terms of computer time and memory since they forecast many variables for particular intervals of the size spectrum of each hydro meteor species. Because of this, bin models are not yet practical for real-time numerical weather prediction attempts, though this is likely to change due to ongoing technological advancements. Alternatively, models designed for real-time applications utilize bulk microphysical parameterizations. They reduce the number of prognostic variables by assuming hydro-meteor size spectra follow a prescribed exponential or gamma distribution.

3.1 Bulk microphysics parameterization

The bulk microphysical parameterization (BMP) used by Thompson et al. (2008) in his work was created to work with WRF (Weather Research and Forecasting) model or other mesoscale models. In comparison to previous single-moment BMPs, the new scheme incorporated a large number of improvements to both physical processes and computer coding, employing many techniques in far more sophisticated spectral/bin schemes using lookup tables. In contrast to any other BMP, the assumed snow size distribution depended on ice water content and temperature and is represented as a sum of exponential and gamma distributions. Furthermore, as observed, snow has a nonspherical shape with a bulk density that varies inversely with diameter, in contrast to nearly all other BMPs that assume spherical snow with constant density. This scheme's snow category was readily modified to match previous research in sensitivity experiments designed to test the sphericity and distribution shape characteristics. From the analysis of four idealized sensitivity experiments, it was determined that the sphericity and constant density assumptions play a major role in producing supercooled liquid water, whereas the assumed distribution shape plays a lesser, but nonnegligible role.
When deciding which representation of snow to use in a microphysical parameterization, there are three major points to consider according to Thompson et al. (2008), and they are:

- Snow geometry
  In previous studies by Reisner et al. (1998); Gilmore et al. (2004) spherical snow with a constant density of $100kgm^3$ has been used in many models. This is a good approximation for snow with a 1.5 mm diameter, but it is not accurate for snow with a bigger or smaller diameter. Theoretical work by Westbrook et al. (2004) and empirical observations from Heymsfield et al. (2007) suggest that the mass of snow is proportional to D^2, or As an alternative, snow's bulk density is inversely

related to its size. As a result, adopting a mass-dimension relation of the form is more realistic $m \propto D^2$.

- Particle Size Distribution (PSD) diagnostic method
  Only a few predicted factors can be employed in a bulk microphysical approach to identify snow PSD qualities. Based on the work of Houze Jr et al. (1979), the most widely used technique is to utilize temperature to determine the y intercept of an exponential distribution. Smith (1984), also used another approach to relate the intercept parameter to snow content different from Houze Jr et al. (1979). Thompson et al. (2008) in their work, defined the PSD by employing both temperature and snow content using orders of magnitude more observed PSDs than is found in either of the other references. In this way, it was better for them to be able to reproduce more of the variability associated with the observed snow PSDs.

- PSD shape
  Even when the shape is represented mathematically as a gamma distribution, the dominant shape assumption for the snow size distribution is exponential in practice. An extensive database of measured PSDs was used to estimate the Control PSD shape in this case, and the underlying form was then revealed through rescaling. The usefulness of the Control PSD shape appears to be supported by two research. Doherty et al. (2007) used model-predicted ice water content, with assumed PSD shape and mass-diameter ratios, as input into a radiative transfer model to replicate satellite brightness temperatures 183 GHz. They found that of the parameterizations examined, the combination of the Field et al. (2005) PSD and m= $0.069D^2$ mass–diameter relationship resulted in the best agreement versus observations for a range of measures. In the second study by Kim et al. (2007), radar measurements were combined with the Field et al. (2005) PSD representation to compute microwave brightness temperatures. Additionally, they discovered good consistency with satellite observations. Therefore, we think consistent microphysics and assimilation suite might be built on the foundation of this PSD representation.

3.2 Suspension Modeling

The atmospheric boundary layer is strongly related to blowing and drifting snow. It is widely accepted that the primary transport mechanisms for erosion, transport, and deposition of snow are saltation and suspension.Pomeroy et al. (1997) found out in their work that saltation is generally restricted to a vertical extension of approximately $0.001 - 0.1$m and drift densities range between 0.1 to $1kgm^{-3}$. Regardless of whether the transition from saltation to suspension appears to be continuous, noticeable suspension starts at moderate wind speeds ($M_{10} = 7 - 11m^{-2}$) depending on the particle size and flow turbulence. The airflow within the surface layer is the driving force for wind-drifting snow.The equation of state can be used to describe this flow in equation 5 , and the conservation equations for mass in equation 6, momentum in equation 7, moisture in equation 8.
The Einstein summation notation will be used for the following equations (6 - 8), similar to what we have in the equations 3 and 4 and all unprimed quantities should be interpreted as Reynolds averaged quantities.

$$\frac{p}{R_a} = \rho_a T_v \tag{5}$$

$$\frac{\partial u_j}{\partial x_j} = 0 \tag{6}$$

$$\frac{\partial u_i}{\partial t} + u_j \frac{\partial u_i}{\partial x_j} = -\frac{1}{\rho_a}\frac{\partial p}{\partial x_i} + \nu \frac{\partial^2 u_i}{\partial x_i \partial x_j} - \frac{\overline{\partial u_i^{'} u_j^{'}}}{\partial x_j} - g\delta_{i3} \tag{7}$$

$$\frac{\partial q_T}{\partial t} + u_j \frac{\partial q_T}{\partial x_j} = -\frac{\overline{\partial q_T^{'} u_j^{'}}}{\partial x_j} + S_{q_T}. \tag{8}$$

In equation 8, q_T denotes the total specific humidity of air measured by $kg_{water}kg_{air}^{-1}$, which can be split to the vapor and non-vapor part, using $q_T = q + q_L$.

- M_{10} is the wind speed at $10m$ surface layer in ms^{-1}
- q_L is the humidity (non-vapor part), measured by $kg_{water}kg_{air}^{-1}$
- p is pressure in (Pa),
- ρ_a is the density of air in (kgm^{-3}),
- R_a is the ideal gas constant of dry air with unit $(JK^{-1}kg^{-1})$,
- T_v is the virtual temperature (K),
- S_{qT} is the moisture source term measured by $kg_{water}kg_{air}^{-1}s^{-1}$.

We assumed that snow grains have the same size and they travel with a velocity $U_{pi} = u_i - W_f\delta_{i3}$, W_f is the absolute value of the free-fall velocity of a grain. The equation 8 can be replaced by the balance equation for only the solid part,

$$\frac{\partial q_L}{\partial t} + u_j \frac{\partial q_L}{\partial x_j} = -\left(\overline{q_L^{'} u_j^{'}} - q_L W_f \delta_{j3}\right), \tag{9}$$

to define precipitation as well as suspension transport. The solid part of the moisture can be written as $q_L = c\rho p/\rho_a$,

- where c is the volumetric concentration of snow,
- ρp is the intrinsic density of ice (kgm^{-3}),
- ρ_a is the density of air (kgm^{-3}),
- W_f is the absolute value of free-fall velocity of a particle expressed (ms^{-1}),
- δ_{ij} is Kronecker delta,
- g is the acceleration due to gravity ms^{-2},
- ν_T is the turbulent kinematic viscosity (m^2s^{-1}),
- e is the instantaneous turbulence kinetic energy (m^2s^{-2}),
- ϵ is the dissipation rate (m^2s^{-3}),
- u_i, u_j is the vector of wind velocity (u, v, w) (ms^{-1}).

On the right-hand side of Equation 9, a sink/source term for the phase change due to sublimation is neglected.
For turbulent equation closure we use the well-known $e \sim \epsilon$ equation as used by Rodi (2017). Here, Reynolds flux and Reynolds stress are approximated by

$$-\overline{u_i^{'} u_j^{'}} = \nu_T \left(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right) - \frac{2}{3} e \delta_{ij} \tag{10}$$

and

$$-\overline{u_i^{'} q_L^{'}} = \frac{\nu_T}{\sigma_T} \frac{\partial q_L}{\partial x_i}, \tag{11}$$

where the turbulent viscosity is parameterized by

$$\nu_T = c_\mu \frac{e^2}{\epsilon}. \tag{12}$$

Separate balance equations are formulated for the dissipation rate, ϵ, and the turbulent kinetic energy, e, i.e

$$\frac{\partial e}{\partial t} + u_j \frac{\partial e}{\partial x_j} = \frac{\partial}{\partial x_j} \left[\left(\nu + \frac{\nu_T}{\sigma_e} \right) \frac{\partial e}{\partial x_j} \right] - \overline{u_i^{'} u_j^{'}} \frac{\partial u_i}{\partial x_j} - g \overline{q_L^{'} u_j^{'}} \delta_{j3} - \epsilon \tag{13}$$

and

$$\frac{\partial \epsilon}{\partial t} + u_j \frac{\partial \epsilon}{\partial x_j} = \frac{\partial}{\partial x_j} \left[\left(\nu + \frac{\nu_T}{\sigma_\epsilon} \right) \frac{\partial \epsilon}{\partial x_j} \right] + c_{1\epsilon} \frac{\epsilon}{e} \left(\overline{u_i^{'} u_j^{'}} \frac{\partial u_i}{\partial x_j} - g \overline{q_L^{'} u_j^{'}} \delta_{j3} \right) - c_{2\epsilon} \frac{\epsilon^2}{e}. \tag{14}$$

The empirical constants are set to $c_\mu = 0.09, c_{1\epsilon} = 1.44, c_{2\epsilon} = 1.92, \mu_T = 0.9, \mu_e = 1.0, \sigma_\epsilon = 1.3$ Rodi (2017).

4 Snow Model Description

Snow models compose most of climate models, which until recently included a very crude representation of snow. The overall aim of snow model construction is to develop a model of surface processes for specific applications, thereby emphasizing a detailed and high-resolution description of the processes and their spatial and temporal variability Lehning et al. (2006).

The primary task of the numerical model is to compute the time and space evolution of prognostic state variables characterizing the snow conditions and their interactions with the environment (atmosphere, ground, vegetation) based on the time integration of physical laws expressed in controlled mathematical equations. The model needs to account for the following items:

- Physical processes operating within the snowpack: processes are represented by a set of equations responsible for the time evolution of the physical properties of snow under the influence of boundary conditions and the intrinsic snow properties.
- Initial and Boundary conditions: this concerns mainly the energy and mass balance at the interface between the snowpack and the overlying atmosphere and underlying ground, which depends on atmospheric and ground characteristics but also processes taking place within the snowpack.

4.1 Snow Model Components

Durand et al. (2005); Liston et al. (2007); Lehning and Fierz (2008); Schneiderbauer and Prokop (2011) in their works, developed several models to simulate wind-induced snow distribution in alpine terrain and resulting snow-depth patterns. These models typically consist of two parts:

- a snowpack component to calculate the erodible snow mass and the threshold wind speed for snow transport,
- a component of the atmosphere to correctly reproduce the spatial and temporal evolution of the wind field and the subsequent snow transport. Whether they are focused on simulating a single blowing snow event or the entire snow season, they range in complexity from simple to quite complex Liston et al. (2007).

In order to accurately simulate wind-induced snow transport in alpine terrain, one must have a solid understanding of the high-resolution wind field over complex topography. Mott et al. (2010) described from his work that Crest speed-up, flow channeling, and the re-circulation zone formation are the driving mechanisms behind the in-homogeneous snow distribution resulting from blowing snow events. They found that more sophisticated snow transport models use three-dimensional wind fields to replicate these properties. They are calculated using computational fluid dynamics models as used by Gauer (1999); Schneiderbauer and Prokop (2011) or atmospheric models run in the Large Eddy Simulation (LES) mode.

Xue et al. (2000); Xue et al. (2001) described an atmospheric model called the Advanced Regional Prediction (ARPS), which offers wind fields with a $5m$ horizontal grid spacing Mott et al. (2010) to drive the snow transport module of Alpine3D Lehning and Fierz (2008). The previous atmospheric models are utilized to drive snow transport models in alpine terrain; however, they do not replicate drifting and blowing snow in a coupled mode. However, previous works from Michioka and Chow (2008) have shown that atmospheric models can be run at high resolution in complex terrain to simulate in coupled mode meteorological situations or scalar dispersion. This study was successful in capturing the flow structures in complex terrain. Consequently, atmospheric models can be applied to the coupled simulation of blowing snow events in alpine terrain.

4.2 Modeling Approaches of Snow Accumulation Processes

A large variety of models have been developed to simulate and better understand snow accumulation processes in different environments seasonally covered by snow (e.g., alpine, arctic, prairies). Accounting for wind-induced snow transport is required to capture the small-scale pattern of snow accumulation. Table 1 gives the main characteristics of models capable of simulating snow variability influenced by wind redistribution. There are two broad groups into which these models fall: (i) models based on semi-empirical parameterizations of the physics of snow transport and (ii) models resolving the 3D turbulent- diffusion equation for blown snow particles in the atmosphere.

Model	Main references	Resolution	Wind input	Saltation	Suspension	Preferential deposition	Cloud scheme
Meso-NH /Crocus	Vionnet et al., 2014, 2017	50 m	Dynamic wind field from an atmospheric model in LES mode	Semi-empirical parameterization[a]	Turbulent-diffusion equation (non-uniform PSD)	Advection of solid hydrometeors (non-uniform PSD)	ICE3[h]
Alpine-3D	Lehning et al., 2008; Mott et al., 2010	5–50 m	Library of wind fields from an atmospheric model in LES mode	Saltation model[b]	Turbulent-diffusion equation (uniform PSD)	Turbulent-diffusion equation (uniform PSD)	None
Wang and Huang, 2017		25 m	Dynamic wind field from an atmospheric model in LES mode	Vertically-integrated total transport rate[c] (saltation +suspension)		Lagrangian tracking of falling snow particle (non-uniform PSD)	None
Sharma et al., 2018		0.8 m	LES of turbulent flows with a specific model	Saltation model[d]	Lagrangian stochastic modeling (non-uniform PSD)	Lagrangian stochastic modeling (non-uniform PSD)	None
PBSM	Pomeroy et al., 1993	Hydrological response units	Wind flow models of varying complexity[e]	Semi-empirical parameterization[f]	Vertically-integrated transport rate	None	None
SnowTran3D	Liston and Sturm, 1998; Liston et al., 2007	5–30 m	Station data with topographic corrections or library of wind fields from an atmospheric model[g]	Semi-empirical parameterization[f]	Vertically-integrated transport rate	None	None
Sytron 3	Durand et al., 2005	45 m	2D wind model	Semi-empirical parameterization[f]	Same as SnowTran3D	None	None
Gauer, 2001		2.5–20 m	CFD model	Saltation model	Turbulent-diffusion equation (uniform PSD)	Turbulent-diffusion equation (uniform PSD)	None
SnowDrift 3D	Schneiderbauer and Prokop, 2011	2–10 m	CFD model	Saltation model	Turbulent-diffusion equation (uniform PSD)	Turbulent-diffusion equation (uniform PSD)	None
NEMO	Naaim et al., 1998; Michaux et al., 2001	1–10 m	CFD model	Semi-empirical parameterization[e]	Turbulent-diffusion equation (uniform PSD)	None	None

PSD stands for Particle Size Distribution, CFD for Computational Fluid Dynamics and LES for Large Eddy Simulations. [a]Adapted from Sørensen (2004); [b]Doorschot and Lehning (2002); [c]Essery et al. (1999); [d]Comola and Lehning (2017); [e]Musselman et al. (2015); [f]Pomeroy and Gray (1990); [g]Bernhardt et al. (2009); [h]Pinty and Jabouille (1998).

Table 1: Source Mott et al. (2018). Main characteristics of the different numerical models capable of simulating snow variability influenced by wind redistribution.

Models that belonged to the first category based on the works of (Pomeroy et al. (1997);Durand et al. (2005);Liston et al. (2007)) rely on saltation and suspension layer transport rates that have been vertically integrated. Due to their comparatively low computational demands, these models successfully simulated the whole snow seasons in the Canadian Prairies and the Arctic or mountainous regions. A better understanding of atmospheric turbulence is necessary for snow accumulation modeling.

From the works of Naaim et al. (1998); Gauer (1999); Schneiderbauer and Prokop (2011), they found the second category of models that used the CFD (Computational Fluid Dynamics), these atmospheric models are either fully coupled to a snow cover process model as in the case of Vionnet et al. (2014) or operate in Large Eddy Simulations (LES) mode, providing libraries of flow fields as input for snow cover process models as seen in (Lehning and Fierz (2008); Mott et al. (2010)).

Due to their complexity, these models typically concentrate on a single blowing snow event, but they can occasionally be applied to a whole snow season Groot Zwaaftink et al. (2013). With various assumptions regarding the particle size distributions, the 3D turbulent-diffusion equation for snow particles in the suspension layer is solved (fixed or non-uniform). The description of the saltation layer varies from semi-empirical relationships Sørensen (2004) to more advanced models representing the essential characteristics of saltating snow particles (aerodynamic entrainment, splash) JDoorschot et al. (2004). In these models, the 3D wind field is obtained from a library of pre-computed situations Raderschall et al. (2008); Mott et al. (2010) or downscaled from meteorological analysis or forecast using a grid nesting approach Vionnet et al. (2017); therefore preferential deposition can also be explicitly simulated.

A different method for investigating terrain-flow-particle interactions in greater depth is lagrangian tracking of falling snow particles, as Wang and Huang did in (2017). In addition, the coupled snow-atmosphere modeling approach proposed by Vionnet et al. (2014) explicitly allows the simulation of local cloud dynamical effect to debate the relative importance of the various processes influencing snow accumulation variability in alpine terrain Vionnet et al. (2014). Due to processing costs and numerical stability on steep terrain, this approach is currently limited to intermediate resolutions ($50m$), and it cannot be utilized to examine patterns of snow accumulation at very high resolution Mott et al. (2010). Modeling snow accumulation processes at different times and spatial scales remains a significant challenge, and existing models could be improved in various ways. In particular, the physical parameterizations used in numerical modeling of blowing snow do not include the latest findings by Aksamit and Pomeroy (2018); Paterna et al. (2016) in the complex coupling between turbulence and snow transport.

Real-world simulations of wind-induced snow transport in alpine environments will call for a modeling strategy that combines atmospheric LESs as seen in Vionnet et al. (2017); Wang and Huang (2017) to capture the complexity of the atmospheric flow with cutting-edge particle motion models from Nemoto and Nishimura (2004);Comola and Lehning (2017) to represent the interactions between turbulence, grain dynamics, and snow surface. Additionally, meteorological models in LES mode offer a significant opportunity to research regional snowfall mechanisms and how they affect snow accumulation variability Vionnet et al. (2017); Gerber et al. (2018). The development of a model, or a set of models, that can simulate the inherently turbulent nature of the various mechanisms causing the spatial heterogeneity of snow accumulation across a wide range of scales would, in general, pose the most significant difficulty in the future.

4.3 Some Examples of Coupled Snow Models

SNOWPACK is a well-established high-complexity snow cover model developed at the Swiss Federal Institute for Snow and Avalanche Research (SLF), Davos, Switzerland, originally developed for avalanche warning and forecasting Lehning et al. (1999). SNOWPACK has continually developed over 20 years into a multi-purpose model for cryospheric snow-atmosphere interactions Haberkorn et al. (2017) and has been applied in different topics of cryospheric research such as snow hydrology by (Wever et al. (2016a), Wever et al. (2015)), climate-change-induced impact assessments on snow and snow hydrology by (Bavay et al. (2013), Marty et al. (2017)), ice-sheets mass balance and thermodynamics by Keenan et al. (2021). The model is used for predicting snow development with fine stratigraphic details, enabling snow stability estimations. The core SNOWPACK module solves the heat equation on a dynamic finite element mesh, which evolves with mass changes of ice and snow to preserve the identity of layers. Using transient snow microstructure to determine snow properties such as viscosity, thermal conductivity, or albedo, a very detailed representation of snow processes from settling to phase changes and water transport results Wever et al. (2014). The finite element implementation is particularly valuable in describing the varying snow stratigraphy, and diverse soil compositions Lehning et al. (2006), upper boundary conditions for mass and heat for initial snow density Groot Zwaaftink et al. (2013), for snow erosion Lehning and Fierz (2008), or for stability corrections Schlögl et al. (2017).

In 2006,Lehning et al. (2006) formed a detailed model of mountain surface processes and its application to snow hydrology known as ALPINE3D; a model for the high-resolution simulation of alpine surface processes, in particular snow processes. It was discovered to be a valuable tool for investigating mountain surface dynamics. It is currently used to investigate snow cover dynamics for avalanche warnings, permafrost development, and vegetation changes under climate change scenarios.
The ALPINE3D system consists of two structural units: the input/output (I/O) part and the computation part. The I/O unit has various components: data on topography, initial vegetation, snow cover, soil conditions, the time series of meteorological conditions, simulation parameters, and a choice of output options.

The computation consists of three core modules: SNOWPACK, SnowDrift, and energy balance, driven by the control module. The runoff module is a computationally inexpensive additional module, which is invoked after the simulation of each time step. The control module can turn a computation module on or off depending on the simulation parameters from the I/O part, control the simulation time, synthesize the results, and output them to the I/O part. The three computation modules will exchange data after each simulation time step Lehning et al. (2006).

The one-dimensional SNOWPACK model was previously coupled (in an offline setting) with a three-dimensional solver in the atmosphere specifically for snow transport, and a spatial description of snow-atmosphere interaction Raderschall et al. (2008) to form ALPINE3D above.
While in an online approach, SNOWPACK was used in the coupling with WRF by Sharma et al. (2021). A unique coupling library was implemented to link SNOWPACK and WRF to form a coupled model called CRYOWRF. CRYOWRF is a fully coupled atmosphere-snowpack solver suitable for simulations in alpine and polar regions. CRYOWRF brings together two state-of-the-art models, with WRF (the Weather Research and Forecasting model)being the atmospheric core of the model while SNOWPACK, a high complexity snow model, acts as the land-surface model (LSM) Sharma et al. (2021).

This development combines several benefits: Due to its non-hydrostatic dynamical core and domain nesting, WRF is a commonly used community model with multi-scale capabilities. This enables the use of the same software package for simulations at scales ranging from the global (with grid sizes of (50–100km) to the turbulent scales (with grid sizes of (10–25m).

The SNOWPACK model allows for the intelligent merging (splitting) of vertical layers based on their similarity (strong gradients), apart from the physical processes of snow accumulation, melting, and settling. The minimum thickness of a snow layer is set at $0.001m$. This is quite a small value and was chosen to demonstrate CRYOWRF's ability to model snowpacks at such extreme resolutions. The blowing snow model was active with an eight-layer fine mesh between the surface and the first mass point of WRF.

A notable feature of CRYOWRF is its ability to run atmospheric simulations with highly (vertically) resolved snowpacks. This allows accurately representing of thin subsurface ice layers of thickness of the order of a few millimeters that have a significant impact on snow hydrology at larger scales Wever et al. (2016b).
The significant difference between these coupling methods and SNOWPACk is that there is a transition from an offline approach to an online one. Various feedback processes that the offline approach cannot describe were well described here.

In order to simulate blowing snow events in alpine terrain and the ensuing snow redistribution, Vionnet et al. (2014) incorporated a new coupled model. For the atmospheric component, the Meso-NH Lafore et al. (1998) atmospheric model is used to simulate the evolution of meteorological conditions and the ensuing transport of snow. At the bottom of the atmosphere, the latest version of the detailed snowpack model Crocus Vionnet et al. (2012) describes snowpack properties. Fully coupled snowpack/atmosphere simulations of blowing snow occurrences in alpine terrain can be used to understand better a variety of issues, such as

- the significance of blowing snow sublimation and its feedback on the atmospheric boundary layer.
- the application of grid nesting methods to provide realistic boundary conditions to local atmospheric simulation (resolution $\approx 50m$) Talbot et al. (2012).
- the relative contribution of preferential deposition of snowfall and wind-induced snow transport

to the spatial variability of the mountainous snow cover Mott et al. (2010).

In contrast with earlier-developed models, MesoNH/Crocus includes the essential feedback mechanisms between snow and atmospheric dynamics, such as the temperature and humidity effects of blowing snow sublimation.

5 Limitations of Snow Models

The papers reviewed and analyzed in this study about snow modeling and microphysics have several limitations, which include;

- Identification of the right choice of topography for the model estimation: In some of the papers reviewed the topographic limitations encountered were in terms of the slope with the questions should the slope be continuous or not? Would the model work for discontinuous slopes, especially slopes with cliff-type behaviors if continuous? How to account for the errors and map out such topography?

- Lack of details for reproducible results: while all authors showed keen interest in modeling snow at different terrains under different conditions using various snow model components, the lack of detailed explanation made most of the results difficult.

  It was noted in some of the papers reviewed that specific models could not replicate specific structures because certain heights in the terrains used in their research could not accurately represent the local topographic features that affect the atmospheric flow and cause snow erosion and deposition. They do not discuss in detail why some of the scientific or physical basis of the choices were made and the model's comparison with other previously published models.
  In some papers, the lack of data to evaluate the spatial variability of some simulated snowfall and snow accumulation limits the ability to analyze model results more deeply.

  Mott et al. (2010) encountered something related to that in their work at the Gaudergrat ridge (Swiss Alps), where their model at a horizontal resolution of $50m$ reproduced snow deposition patterns at the ridge scale but missed smaller-scale deposition patterns. These patterns were partially produced using horizontal resolutions of 10 and 5 m.

- Some of the models used are computationally expensive and time-consuming. There is undoubtedly a further need for improvements in coupling advanced snowpack models with atmospheric models and dynamical downscaling using multi-scale atmospheric models, including snow drift resolving scales. But to produce such models, they strongly rely on advances in computational power, allowing for more complex simulations also for more extended periods, larger areas, and higher resolution, which most of the researchers do not have.
  Sauter and Galos (2016) in his work found a better approach to use LES to resolve boundary layer advection problem of sensible heat. Still, his model approach was restricted to intermediate resolutions (50 m) due to computational costs and numerical stability in steep terrain. It cannot be used to explore snow accumulation patterns at a very high resolution.

- There are many assumptions about the size and shape of the snow particles based on intuition or just random guess by the previous researcher, causing a fluctuation in the efficiency of the mass and energy exchange in the snow field. An example was from the results of Thorpe and Mason (1966), Wever et al. (2009) they estimated the mass's efficiency. They found that energy exchange increased by 2.5 to 5 for dendritic shapes compared to hexagonal plates. A result of this leads to

the imbalance and underestimation of the sublimation rate used for the model.
Also, Sharma et al. (2021) in their work assumed values for some constants and variables needed to formulate a model as of the case of K_{bs}, the turbulent diffusion coefficient for blowing snow particles.

6 Open Research Questions

We observed that most models contain some questions, which may be categorized as the following while reading the articles for this synthesis.

- How do we know the correct values of parameters to be used? Does random guess and assumption of some snow parameters affect the results obtained from the modeling of specific snow variables. Also, is it reasonable to use these supposed parametric values and still have minimal error? Furthermore, under what conditions will they hold?
- How do we account for errors in snow modeling research? Apart from the measure of dispersion, are there possible ways to test if a model is correct or not just by looking at it, even when the results look convincing enough?
- Can we choose a specific snow model based on the amount of snow we get in the study region? How do we account for extreme situations in the snow models input? Because climate change also affects the mountainous regions. In general, how can global warming affect the snow models in terms of their accuracy?

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