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REVIEW PAPER

Snowcover Simulation Model – A Review

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ABSTRACT

Numerical simulation of seasonal snowcover has attracted the interest of many scientists in the recent past. The present paper summarises chronologically developments in the understanding of snow properties and discusses various modelling approaches towards simulating the snowpack numerically. The authors describe the evolution of snowcover and the intricate relationship between the evolving snowpack and the atmosphere. The governing equations that describe the evolution of snowcover have been discussed. The merits and limitations of each equation describing a single process have been explained. Modelling strategies adopted by various workers have been analysed, and lastly the requirements of a perfect model have been brought out. In the absence of complete answers to many other processes, a strategy for the development of an operational snowcover model has been discussed.

NOMENCLATURE

C_h, C_e	Turbulent heat transfer coefficients
C_{pa},C_{ps},C_{pw}	Specific heat capacity of air, snow and water (cal g^{-1} K^{-1})
$\dot{\varepsilon}$	Strain rate (s^{-1})
e	Water vapour pressure (mb)
$E_i(T)$	Saturated vapour pressure above a flat ice surface at temperature T (K)
P_a	Atmospheric pressure (mb)
P_c	Capillary pressure (mb)
T_{air}, T_{surf}	Temperature of air and snow surface (K)
T_0	Snow melting temperature (K)
U_a	Wind velocity (m s^{-1})
σ_s	Vertical stress (g cm^{-2})
η	Viscosity coefficient
σ	Stephan-Boltzmann constant (1.355×10^{-12} cal cm^{-2} K^{-4} s^{-1})
ε_a	Atmospheric emissivity
ε_s	Snow emissivity (0.97–1.0)
ρ_a, ρ_w, ρ_s	Density of air, water and snow (g cm^{-3})
β	Extinction coefficient (cm)

1. INTRODUCTION

Of the three main ingredients, viz., terrain, snow and weather, that are essential for the formation and release of avalanches, snowpack structure on a mountain slope has an important role to play. One of the most promising ways to understand the structure and properties of snowcover is the numerical parameterisation of all the involved processes. An understanding of the evolution of snowcover has urged many a scientist to simulate a snowpack using values of easily measurable snow-meteorological parameters collected in the field. The necessity for simulation

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was first felt by field scientists who wanted to know the structure of the snowpack without having to dig a pit for it. Snow scientists, realising the potential application of the simulated snowpack to various snow-related problems, e.g., avalanche-prediction, hydrological aspects, agricultural aspects, etc., attempted to develop models using the physical laws.

2. EARLY DEVELOPMENTS

The early attempts by Bader[1,2], Kojima[3,4], Keeler[5], Mellor[6,7] and Yoshida[8], to name a few, represented a beginning in this direction by defining various properties of snow. Snow densification attracted the scientists and various constitutive laws were put forward to define the behaviour of snow. Following these efforts, Anderson[9] made the first attempt to present a point energy and mass balance model of a snowcover. However, this model and many others (e.g., those of Obled[10], Morris and Godfrey[11] and Navarre[12]) were principally meant for hydrological applications. While efforts were continued to understand the basic properties of snow and to formulate expressions to define its behaviour, avalanche forecasters tried to use these expressions for deducing the behaviour of snowpack on a slope.

Brun[13], *et al.* made the first attempt at consolidating the information available in seventies and eighties to devise an energy and mass model of snowcover for operational avalanche forecasting. The model, named CROCUS, was tested at a well-instrumented site, *Col de Porte*. It was found to be a useful tool for operational avalanche forecasting. Following this, a series of attempts were made by other scientists in different parts of the world to produce snowcover simulation models. Motoyama[14] attempted simulation of seasonal snowcover based on air temperature and precipitation. Bader[15], *et al.* following the mixture theory for a phase changing snowpack (Morland[16]), modelled temperature distribution, and energy and mass flow in a snowpack. The model was named SNOWPACK. Later, many versions of SNOWPACK were reported. Ganju[17], *et al.* developed a snowcover model and realising the limitations of an exact simulation suggested a semi-quantitative model for operational purposes. Lately, Brun[18], *et al.* incorporated laws of metamorphism into CROCUS for more effective usage of the model. Yamazaki[19], *et al.* using certain basic variables, attempted to calculate albedo, density profile, etc. for one-dimensional simulation of snowpack. Since then, more emphasis is being laid on understanding the metamorphism aspect of the model so as to be in a position to draw a realistic picture of the strength characteristic of different types of snow layers.

The difficulty in using the models for a particular region was realised by many workers, primarily because the existing models could be used to simulate the snowcover only at a location where all the relevant parameters required for modelling were collected by installing adequate instrumentation. Realising the lacuna, Durand[19], *et al.* developed yet another model, viz., SAFRAN, that could deduct relevant parameters needed by the model at numerous points of concern by using the data available around these points.

Notwithstanding the progress made, in the simulation of snowpack, aspects like formation of weak layers, strength of various constituent layers, true representation of metamorphosed snow, melt-water percolation, etc. have not yet been taken into account.

3. SNOWCOVER EVOLUTION

Snowcover evolves layer by layer as the seasonal snowcover progresses. These layers, formed naturally from each snowstorm, undergo the process of densification under their own weight and the overburden pressure. Subsequently, more layers are formed by the process of metamorphism, which further enhances the distinct layering structure of the snowpack. The seasonal snowcover may evolve on a barren ground surface, or on an ice bed of a glacier or on the remnant snow bed of previous years that did not melt completely. On all these surfaces, snowcover evolves differently, depending on the temperature regime and the stress conditions it has been subjected to. Generally speaking, a snowcover in a climatic zone should have similar characteristics everywhere. However, large variability has been observed, which is totally governed by the local meteorological conditions of a specific area or location or slope that interact with the snowpack.

A snowcover, as it evolves, behaves like a dynamic reservoir of mass and energy, the evolution of which is governed by the prevailing boundary conditions. While at one instant it may be a sink, at another instant it may act as a source of energy. Being a porous material, it conducts very little energy to its underlying layers. As a result, a strong temperature gradient sets in a relatively shallow snowpack. This temperature gradient causes the snowpack to metamorphose into a different crystalline structure. The dynamic interaction of the snowpack with the atmosphere causes it to progress either in constructive or in destructive metamorphism regime at a time. Constructive metamorphism regime weakens the snowpack with the growth of depth hoar crystals. Destructive metamorphism regime strengthens the snowpack with the formation of necks between the crystals. Depending on the prevailing metamorphism regime, the snowpack can either gain or loose strength by the process of sintering or formation of weak layers, respectively. The evolution of different types of weak layers can be due to either of the following processes:

- Precipitation of needle, graupel crystals. These crystals retain their original form for a long period and have low resistance to shear.
- Formation of surface hoar crystals on the snow surface, which once buried under subsequent snow spells form a thin weak layer that also has low resistance to shear.
- Formation of crust layers on the snow surface as a result of melt-freeze cycle that the top layer undergoes during clear weather days.
- On clear cold days, radiation recrystallisation process sets in a strong temperature gradient in the top few centimeters. This results in the formation of a recrystallised crystal layer, which can fail in shear due to subsequent loading on it.
- Melt-water in the winter percolates at times to a certain depth during daytime, which upon freezing during night produces an ice layer or lenses within the snowpack. Melt-water percolation through preferred channels can produce varied types of ice layers within the snowpack.

Thus, the evolving snowpack develops a varied structure through its interaction with the atmosphere. While the evolution of some layered structures is well understood, it is not known for many others, e.g., layers with radiation recrystallised crystals or with ice lenses.

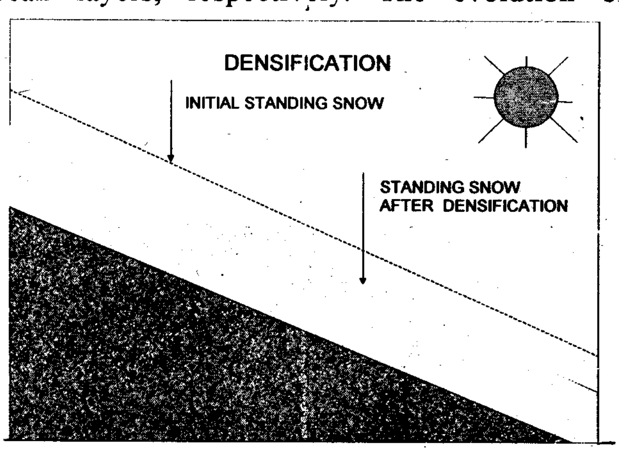

Figure 1. Process of densification

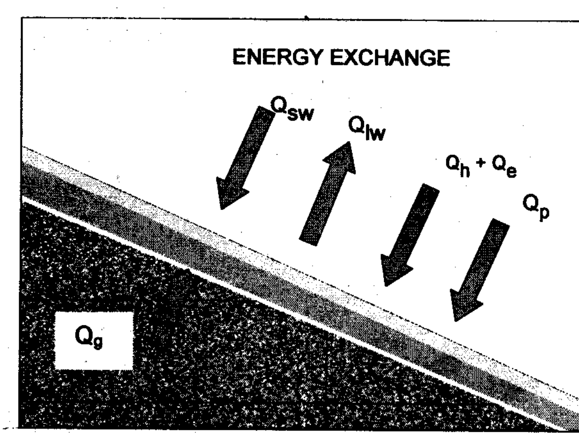

Figure 2. Energy exchange between snowcover and surroundings.

It is not only the evolution of snowpack structure with which the scientists and forecasters are concerned, the determination of strength characteristics of each layer is also very important. Thus, the modelling of evolution of a snowpack requires understanding of the main processes pictorially represented in Figs 1-6. Details of each of the processes follow in the subsequent paragraphs.

- *Densification*: Ideally, each snow layer follows its own densification law.
- *Energy exchange*: Precise determination of turbulent heat exchange coefficients between snow and atmosphere is essential. Heat conduction and diffusion through the snowpack must be evaluated correctly.
- *Temperature regime*: Needs to be modelled by incorporating sink and source terms.
- *Metamorphism*: A precise model of different types of snow grains and their sizes need to be developed.
- *Weak layer*: Constructive metamorphism, interaction of surface layer with the atmosphere, and precipitation to needle and graupel all contribute to the formation of weak layers.
- *Melt-water percolation*: Melting generally proceeds from top. However, some melting at times starts at the bottom of the snowpack. Melt-water follows preferential flow paths through the snowpack during its descent. Flow of melt-water through the snowpack with phase-change phenomenon involves a complex set of mathematical equations for analysis.

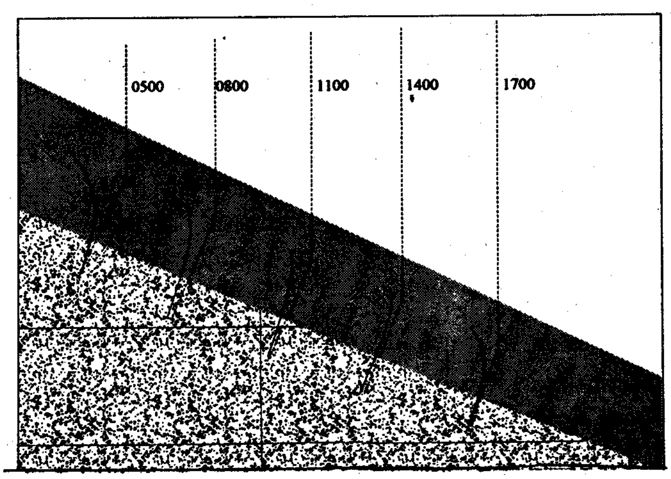

Figure 3. Diurnal variation in temperature profile of the snowcover.

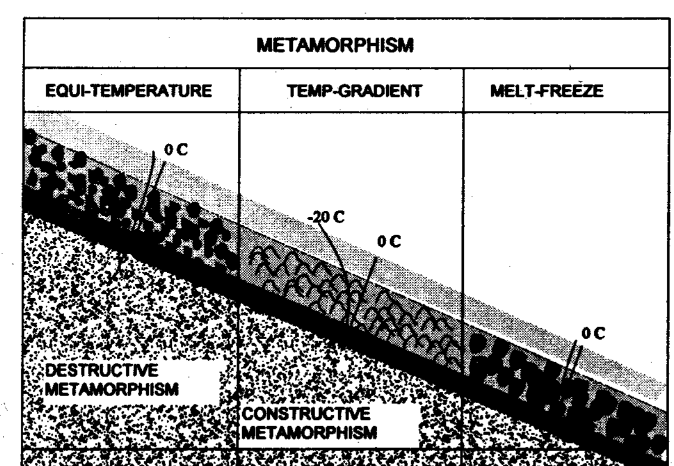

Figure 4. Process of metamorphism occurring within the snowcover.

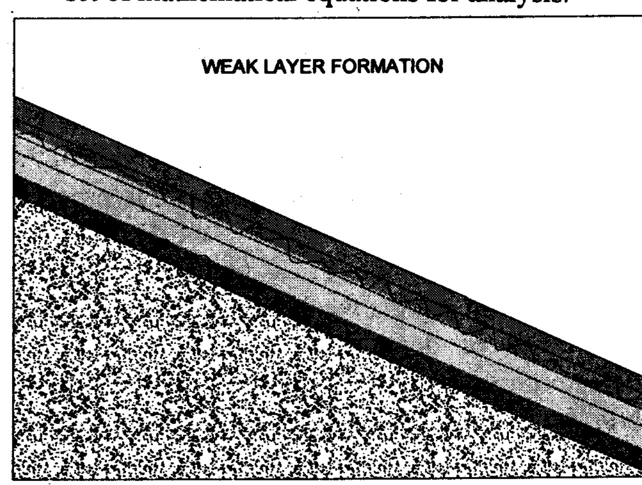

Figure 5. Weak layer formation within the snowcover

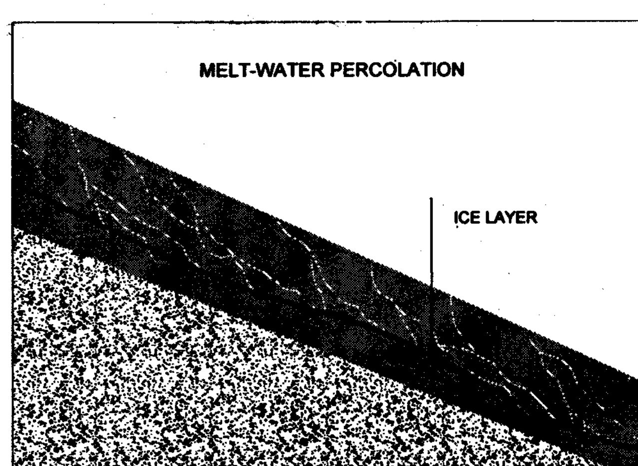

Figure 6. Melt-water percolation

4. GOVERNING EQUATIONS

Various equations explaining the process of snowcover evolution have been developed at different stages. Some equations have sound theoretical explanations, while others are empirical. An analysis of each process and the governing equation is given here.

4.1 Densification

Bader and Landauer[21] first explained the densification of snow. From the theoretical proposal of creep for metals, strain rate can be related to stress through a hyperbolic sine function:

$$\dot{\varepsilon}=\frac{\sigma_o}{\eta}\sinh\left(\frac{\sigma_s}{\sigma_o}\right) \tag{1}$$

where σ_o is the empirical constant.

Investigations revealed that snow behaves as a Newtonian fluid under state of low stresses. Following the reasoning extended by Mellor[22] that the snow lying on a slope, when stressed beyond limits of linear viscosity, either fails or densifies rapidly to behave once again like a Newtonian body. The seasonal snowpack has been assumed to behave like a Newtonian body in all models developed for simulating a snowpack.

Snow, in fact, demonstrates many features of a brittle solid, an elastic rubber or a viscous liquid, depending on its temperature and time scale of measurement. At the lower limit of the stress range, there is a linear relationship between strain rate and stress. At higher stresses, strain rate tends to become nonlinear. While studying the effect of density on the deformation resistance of snow, Kojima[23] and Yoshida[24] represented the results of their observation on the natural settlement of seasonal snow ($0.1 < \rho_s < 0.5$ gm/cc) in the form of a simple experimental equation:

$$\eta=A\varepsilon^{a\rho_s} \tag{2}$$

where A is a constant (varies with snow type and temperature), and a, an exponent constant (which is apparently about 21 cm^3/g).

Following the above, many other researchers described settlement law of snow. Navarre[12] established the settlement law of snow using mean viscosity, η, as follows:

$$\frac{dl}{l}=\frac{-\sigma_s}{\eta}dt \tag{3}$$

with

$$\eta=\frac{6\times10^4}{1-f(d)}e^{[23\rho_s-0.1(T_n-T_o)]} \tag{4}$$

where l is the layer thickness; σ_s, vertical stress; dt, the time interval; and $f(d)$, the function of snow type. T_n is the snow temperature and T_o, the melting temperature of snow. Brun[13], *et al.* used the same expression with a constant value of $f(d)$ = 0.4 to simulate the process of densification of seasonal snow in CROCUS model.

While approximation of linear deformation under conditions of constant stress with strain remaining limited holds good to a large extent, the response no longer remains linear as the stress exceeds a critical value. Transition from linear to nonlinear response depends on several factors, e.g., stress, strain rate and duration of stress application. Many workers demonstrated a nonlinear viscoelastic behaviour of the material. Working on these lines, Agrawal[25] attempted to formulate the power law using microstructural parameters of snow to describe the behaviour of snow.

Despite the efforts mentioned above, in their snowpack simulation attempts, various workers have adopted simple linear densification law either through empirical constants derived from field data (Ganju[17], *et al.*) or by using experimental transfer functions to transform grain type to viscosity. Since snowpack evolves nonhomogeneously, the effect of different types of crystals, their intergranular bonding, etc. on viscosity of the snowpack needs to be investigated.

For modelling purposes, while snowpack deformation is assumed to follow linear viscous behaviour, separate deformation laws have to be evolved for different layers (of uniform composition) of snowpack. Yamazaki[20] was the first to formulate compactive viscosity coefficient for wet snow by using a multiplicative factor that describes the decrease in compactive viscosity coefficient due to liquid water content. The modified viscosity expression is:

$$\eta = A(w)\eta_o \exp(k\rho_{dry} - \alpha_s T_s) \qquad (5)$$

where

ρ_{dry} is dry snow density and $A(w)$ is assumed as

$$A(w) = \frac{\exp(-\beta_s w) - \exp(-\beta_s)}{1 - \exp(-\beta_s)} \qquad (6)$$

where w is the water content; η_o (=6.9×10^5 kg s m^{-2}); k (=2.1×10^{-3} m^3 kg^{-1}); α_s (=9.58×10^{-2} °C^{-1}); and β_s (=18) are constants and T_s is the snow temperature.

Comparing different viscosity laws, Lehning[26], *et al.* formulated the expression of viscosity for calculating settlement of snow in the SNOWPACK model as

$$\eta = 7.0 \times 10^{-3} \rho_s^{(5-0.025T)} \qquad (7)$$

where η is compactive viscosity (kg m^{-1}s^{-1}), and T, the temperature of snow (°C).

4.2 Energy Exchange, Temperature Profile & Mass Conservation

Snowpack, a dynamic reservoir of heat and mass, comprises a complex assembly of ice, water, water vapour and air. Its modelling did not progress satisfactorily for quite some time, since many processes were not understood completely for a long time. In addition, the four coexisting constituents of the snowpack have to be so treated as to satisfy the conservation of mass, momentum and energy.

Snowpack, during its evolution, is subjected to heat exchange processes that affect its temperature profile. Various workers attempted to determine the temperature profile by considering all the energy flux exchanges. Since Rockwood[27], who made the earliest attempt at determining the temperature profile using ambient temperature as the sole index, to the more recent work of Singh[28], many researchers have attempted to determine the precise temperature profile of the snowpack. While doing so, it was considered important to have more or less exact values of thermal conductivity, diffusion coefficient, etc. and a good modelling approach. Quick[29], Arons[30] and many others suggested expressions for the determination of thermal conductivity. Anderson[9] fina y came up with more practical solutions with his point energy and mass balance model. The phenomenological approach of snowcover simulation propounded by Bader[15], *et al.* discusses all aspects of energy exchange process within the snowcover.

The energy budget for a snowpack, with the surface-atmosphere interface temperature and the snow-ground interface temperature as boundary conditions, can be estimated from the following equation:

$$\Delta Q = Q_{sw} + Q_{lw} + Q_h + Q_e + Q_g + Q_p + Q_{wp} - dU/dt \qquad (8)$$

where

ΔQ	Net balanced energy flux
Q_{sw}	Net short wave radiation flux absorbed by snow
Q_{lw}	Net long wave radiation flux at snow-air interface
Q_h	Convective or sensible heat flux exchange between air and snow surface
Q_e	Latent heat flux (evaporation, sublimation, condensation) at the snow-air interface
Q_g	Flux of heat supplied by ground
Q_p	Flux of heat supplied by precipitation (snow/rainfall)
Q_{wp}	Flux of heat supplied by wind pumping
dU/dt	Rate of change of internal energy per unit area of snowpack

4.2.1 *Energy Exchange at the Surface-Atmosphere Interface*

4.2.1.1 *Net Long Wave Radiation*

The exchange of long wave radiation (5-40 μm) is confined to snow surface only. The net long wave radiation at snow-air interface is the resultant of incident long wave radiation ($Q\downarrow$) minus the emitted long wave radiation ($Q\uparrow$). Since mostly $Q\downarrow$ is smaller than $Q\uparrow$, there is generally a loss of energy from the snowpack in long wave part of the electromagnetic spectrum.

The main sources of incident atmospheric long wave radiation are the water vapour present in the atmosphere, which emit about 80 per cent of the total long wave radiation. Various gaseous molecules (mainly carbon dioxide and ozone) are also present. There is a direct correlation between temperature and concentration of water vapour and the intensity of radiation. The widely accepted relation in this regard proposed by Brunt[31] is:

$$Q\downarrow=\sigma . T_a^4 (a+b\sqrt{e}) \tag{9}$$

where σ is Stephan-Boltzmann constant; T_a and e are the absolute temperature and vapour pressure (Pa) measured at 1.5–2.0 m, respectively; and a and b are empirical constants whose values change with location and time of the year.

Snow is assumed as a nearly perfect black body in the long wave portion of the spectrum and the long wave radiation emitted by snow can be calculated as

$$Q\uparrow=\varepsilon_s \sigma (T_{surf})^4 \tag{10}$$

The ε_s depends on the wavelength and the angle of incidence and for snow its value is assumed to lie between 0.97 and 1.

Clouds affect the net long wave radiation significantly. In general, the incoming long wave radiation in the presence of clouds may be assumed as the black body radiation emitted at the cloud base[32] temperature (T_c). Hence, the net long wave radiation exchange between the atmosphere and the snow can be approximated as

$$Q_{lw}=\sigma(\varepsilon_a T_c^4 - \varepsilon_s T_{surf}{}^4) \tag{11}$$

During the scattered cloud cover, the situation is more intricate. Over snow, the US Army Corps of Engineers[33] suggests the following equation to calculate the net long wave flux under partial cloud cover:

$$Q_{lw}=\sigma(\varepsilon_a T_c^4 - \varepsilon_s T_{surf}{}^4)(1-kN) \tag{12}$$

where k is a coefficient (with value range 0.25-0.90) depending on the cloud type and height of the cloud base, and N, the fraction of sky covered by clouds.

4.2.1.2 *Turbulent Heat Exchanges with Atmosphere*

Both sensible and latent heat exchanges play very important roles in the evolution of snowcover, especially in determining the rate of melt. These are governed by the intricate turbulent heat exchange process occurring in 2-3 m layer just above the snow surface. An independent estimation of these two exchanges is difficult and at present no satisfactory operational method for their estimation is known. Gray and Male[32] described in detail the theoretical approaches and their practical limitations, as suggested by various scientists for estimating the two fluxes. For relatively simplified expressions, one may refer to Brun[13], *et al.* and Deardorff[34], who suggested that

$$Q_h=\rho_a C_{pa} C_h (a+bU_a)(T_{air}-T_{surf}) \tag{13}$$

$$Q_e=\frac{L_s \rho_a}{P_a}\frac{M_v}{M_a} C_e (a+bU_a)(T_{air}-T_{surf})E_i \tag{14}$$

where, L_s, latent heat of sublimation; P_a, atmospheric pressure; and M_v/M_a is the ratio between water vapour and dry air molecular weight. C_h and C_e are generally assumed equal under the condition of atmospheric stability. Values of coefficients a and b are adjusted according to the test site. However, their typical values are a = 0.8 and b = 1.8-3.0.

4.2.1.3 *Heat Exchange due to Precipitation of Snow or Rain*

Rainfall on a snowcover is a common phenomenon. There may be two possibilities in the case of rainfall over a snowcover:

- Rainfall over a melting snowcover (isothermal snowpack at 0 °C), where rain does not freeze.
- Rainfall on a snowcover whose temperature is below 0 °C and where rain freezes after attaining equilibrium with the snowpack.

Whereas for the first case, calculations can be made for energy supplied to the snowpack by conduction, the second situation is more complicated. Some of the rain freezes and thereby releases heat by fusion, which may significantly change the thermal regime of the snowpack, since latent heat of fusion for water is much higher than the specific heat of ice. Bader[15], *et al.* attempted to incorporate this complexity and suggested the following equations:

$$Q_{sn}=\rho_s C_{ps} r_s [T_{snow}-T_{surf}] \tag{15}$$

$$Q_m=\rho_w r_r \begin{Bmatrix} C_{pw}[T_{rain}-T_{surf}], T_{surf}=T_o \\ C_{pw}T_{rain}-C_{ps}T_{surf} \\ +(C_{ps}-C_{pw}T_o), T_{surf}<T_o \end{Bmatrix} \tag{16}$$

where, r_s, r_r are the accumulation rates of snow/rain, respectively.

According to Brun[13], *et al.* generally snowfall takes place at snow surface temperature and rain is supposed to fall at the wet bulb temperature. When rainfall takes place, liquid water is introduced within the upper snow layer at temperature T_o and the energy $Q_r=\rho_w r_r C_{pw}(T_{air}-T_o)$ is supplied to the upper layer.

4.2.2 *Energy Absorption Inside the Snowcover*

4.2.2.1 *Net Short Wave Radiation*

Most of the incoming solar radiation is reflected back by snow. The remaining part penetrates into the snowpack and is absorbed by snow. Reflectance and absorption depend on the wavelength of the incoming radiation, grain size of the snow layers at the surface, age of the snowpack and the presence of impurities in its top layer. Penetration of solar radiation with depth is an exponential function $e^{-\beta z}$, with β (known as extinction coefficient) strongly varying with wavelength.

At a depth z below the snowpack surface, the solar flux, Q_s, is modified as follows (Brun[13], *et al*[13]):

$$Q_s(z)=\int\left(1-\alpha(\lambda)\right)R_s(\lambda)e^{-\beta(\lambda).z}\,d\lambda \tag{17}$$

where $R_s(\lambda)$ and $\alpha(\lambda)$ are the incoming solar radiation and the surface reflectance, respectively corresponding to the wavelength λ. The radiation absorbed by a layer of thickness Δz at a depth z is equal to:

$$\frac{dQ_s(z)}{dz}\Delta z \tag{18}$$

4.2.2.2 *Wind Pumping*

Air movement through snow due to wind action can alter the thermal regime of the snowpack by transferring heat and mass (vapour) within it. This forced thermal convection has important effects on various physical processes occurring in the snowpack like grain growth and crust formation. Heat and mass transfer in the snowpack without air movement are limited by diffusion, which is very slow as compared to that associated with air movement. According to Colbeck[35], however, thermal convection is not common in snow, but variable nature of wind and atmospheric pressure may cause forced air movement for the following reasons. First, change in normal atmospheric pressure may induce air movement in response to surface pressure changes. Secondly, turbulent winds over a snow surface produce rapid pressure changes due to deceleration of downward air currents at the surface. Third and the most important reason is the geometry of the snow surface, which induces a pressure distribution on the surface that causes movement through the permeable snow. In general, flow of air through snow can be treated as Darcian flow through a porous medium. Colbeck suggested the following expression for heat flux due to wind pumping at a depth z below the surface:

$$Q_{wp}(z)=11.6\pi^2 \frac{\rho_a^2 C_{pa} U_a^2}{\mu_a \lambda^3} H . k_a [T_a - T(z)] \tag{19}$$

where λ and H are the wavelength and height of the surface ripples, respectively. μ_a is the kinematic viscosity; k_a, permeability of snow, and $T(z)$, the temperature of snowpack at depth z.

4.3 Metamorphism of Snow & Weak Layers

The process of metamorphism ultimately governs the structure of the snowpack which has direct relevance to the formation of avalanches. Metamorphism of snow can progress in many directions in a snowpack simultaneously. All the parameters like temperature, temperature profile, snow density and presence of melt-water have important roles to play in the metamorphism of snow.

Giddings and LaChapelle[36] were the first to deduce the growth rate of snow grains exposed to high temperature gradients. Colbeck[37-39] developed the theory of metamorphism of wet and dry snow describing the effect of liquid-water content, capillary pressure, radius of curvature and temperature gradient. Experiments by Akitaya[40], Marbouty[41], Pahaut and Marbouty[42] experimentally confirmed these findings. Christon[43], *et al.* discussed the geometric effects on metamorphism using two-dimensional temperature gradient model.

Satyawali[44] developed a more refined equation for calculating the grain size. He (in this issue) has brought to notice the effect of bond radius on the role of metamorphism.

It was finally the work of Brun[18], *et al.* which integrated metamorphism in a numerical snowcover model. They described snow evolution by introducing a new formulation that described snow grains by their dendricity (1 to 0 range), sphericity (0 to 1 range) and grain size.

The effect of melt-water percolation, formation of ice lenses and, most important, the effect of metamorphism on the properties of snow layers has not been modelled yet.

Weak layer studies by Fohn[45] have yet to be modelled in the snowpack. Fierz[46] showed that the evolution of weak layers, the most critical part of avalanche prediction, is still unsatisfactorily represented[18] in CROCUS. This can be attributed to the fact that investigations of snow metamorphism in general (Bader[47], *et al.*, Colbeck[48], *et al.* and McClung and Shaerer[49]) as well as investigations of depth hoar development (Akitaya[40], Marbouty[41], Colbeck[35], Fokuzawa and Akitaya[50]) concentrated on single snow grains. Research work, taking the grain connection into account, is scanty and incomplete (Colbeck[38], Gubler[51]). Detailed description of the formation and texture of near-surface depth hoar in relation to the environmental conditions is also not available.

Although a comprehensive modelling approach to snow metamorphism is missing at present, many particular problems on snow microstructure and metamorphism have been addressed recently by Brown[52], *et al.*, Edens and Brown[53]. Further research on these aspects must be taken up and incorporated into a complete description of snow properties and metamorphism that can then be used in a numerical model.

4.4 Melt-Water Percolation

The earliest attempts to study the phenomenon of melt-water percolation within a snowpack date as far back as 1915, when Horton[54] made the first comparison of wet snow with other porous media. However, snow was later found to be different, since the snowpack, unlike many other porous materials, changes its physical properties with changing temperature and melt-water content. As was later explained by Gerdel[55], the application of simulated rain (Himmel[56]) to the snow surface may accelerate morphological processes conducive to marked changes in permeability. The emphasis at that time was on developing some empirical techniques to predict snow melt from an isothermal[33] snowpack. Studies have continued since then, focusing on the processes that govern the flow of melt-water within the snowpack. Colbeck[57] proposed a gravity flow theory for melt-water percolation into an unsaturated homogeneous snowpack using Darcy's Law (Colbeck and Davidson[58]). Later, he attempted to add complications like water pressure, effect of ice layers, grain size, etc. to it. Colbeck[59] made the first review on water movement in the snow. Subsequently, Wankiewicz[60] made another review of the progress made between 1976-78. Some of the ideas, concepts and experiments that were tried and tested on sand, soil and other porous materials were tested on snow which helped in understanding the wet snow phenomenon better. The important contributions to soil science by Hill[61], *et al.*, Parlange[62] and Raats[63] on wetting front instability, and by Marshall[64], and Millington and Quick[65] on water retention curves helped in understanding the process in snow too. Since then, continued research has significantly increased understanding of the wet snow phenomenon. Marsh[66] in his review paper brought out some important gaps where research is still required.

According to Colbeck[67], water flow starts when water saturation exceeds irreducible water saturation (capillary force). The capillary force may be checked on the basis of the following empirical expression:

$$P_c = 43(S^*)^{-1} + 380 \tag{20}$$

where S^* is a function of water saturation of the snowpack. When P_c falls below 2000 N/m^2, the capillary retentivity requirements (9-10 per cent per mass) are supposed to have been met by Ganju[17], *et al.* The rate of movement of the wetting front can then be calculated[68] on the basis of the following expression:

$$\frac{d\xi}{dt} = \alpha . K . S^{*3} . \phi^{-1} . S_w^{-1} \tag{21}$$

where $d\xi/dt$ is wetting front speed; α, $\rho_w g/\mu_w$ (=54700 cm^{-1}s^{-1} at 0 °C); K, the intrinsic permeability (cm^2) just above the front; ϕ, porosity (pore volume/total volume); and S_w, water saturation (water vol./pore vol). Shimizu[69] suggested:

$$S^* = \frac{S_w - S_{wi}}{1 - S_{wi}} \tag{22}$$

If the wet front happens to reach the bottom, the entire pack is assumed to have acquired the saturation level exceeding the irreducible water saturation level. The calculation for wet front speed is now replaced with the wave front speed, suggested by Colbeck and Davidson[58] as follows:

$$\left[\frac{dz}{dt}\right]_{U_w} = n(\alpha K)^{1/n} \phi_e^{-1} U_w^{(n-1)/n} \tag{23}$$

where $[dz/dt]_{U_w}$ is the speed of propagation (cm/s) of a wave of constant volume flux (U_w) and n is equal to 3. The above equation helps in determining the rate at which the melt-water is allowed to be drained off from the snowpack.

5. MODELLING APPROACH

Anderson[9] first presented a comprehensive approach to computer simulation of a snowpack. The basic modelling was around two main equations. These equations are energy exchange equation and equation for heat transfer within the snowcover. In order to combine and solve these equations, the snowcover was supposed to be made of layers of finite thickness, parallel to the ground. The energy balance equation for each layer was expressed in an implicit finite difference form. Newton-Raphson iteration technique was used to solve these equations. Besides the computation of energy exchange, the model included mathematical representation of the other processes also which affect the energy and mass balance. These processes are mainly the addition of fresh snow to the snowpack, densification and retention and transmission of liquid water.

To determine if the precipitation is in the form of rain or snow, wet-bulb temperature was used as an indicator. For wet-bulb temperatures less than 1 °C, precipitation is assumed to be snow. The density of snow was based on a plot of fresh snow density versus temperature for Alta, Utah[70]. The mathematical expression that best describes the data is:

$$\rho_{ns} = 0.05 + 0.0017(T_w - 258.16)^{1.5} \tag{24}$$

where ρ_{ns} is the density of fresh snow; and T_w, wet-bulb temperature (assumed to be temperature of precipitation)(K)

The model used empirical relations for thermal conductivity and extinction coefficient of snow, using density as the sole index.

In the approach adopted by Brun[13], *et al.*, to calculate change in temperature, density, liquid water and grain parameters of the snowpack with time, the snowpack is divided into layers parallel to the slope. Energy transfers are projected perpendicular to the slope. Since the greatest changes in the thickness occur in fresh snow, the initial thickness of the fresh snow layers are chosen very small, as compared to the densified bottom layers.

All the possible processes involved in the evolution of a snowcover, like change in internal energy, phase change, metamorphism, melt-water percolation and densification occur simultaneously in nature, but numerically these are processed one after the other due to their interdependence. Knowing the values of various parameters at time t_n, the aim is to simulate the values of these parameters at time t_{n+1}. Brun[13], *et al.* in their one-dimensional model CROCUS, decided to calculate first the temperature profile at time t_{n+1} using the energy balance equations. To calculate the temperature of individual layers, the classical resolution technique of Cranck and Nicolson was used, which is an implicit method using finite differences equivalent of derivatives. Afterwards, the model makes the necessary phase changes corresponding to the possible freezing of wet layers or their possible melting. The calculation for melting is made only when the temperature is equal to the melting point. Computation for water percolation is then made after introducing possible rainwater. To avoid any numerical instability, the percolation equation has been linearised. Settling is then taken into account and the possible decrease in thickness of each layer is calculated corresponding to the increase in density. Possible new snow layers are then added to the snowpack.

All these phenomena have been derived using a time increment of 15 min, except for settlement, which is derived at an interval of 1 hr. The derivations for metamorphism are calculated just before the computation of settlement. The evolution of snow type and grain size of each layer of the snowcover is derived for an hour using the temperature, density and liquid water content profiles simulated during the previous hour.

Bader[15], *et al.* considered snow as a continuum of mixture of the three phases of water and air, taking the average values of parameters over small volumes rather than taking them at specific space points. Under this assumption, the governing equations are the conservative laws for mass, momentum and energy equations of state. Based on the work of Morland[16], *et al.* who studied these equations in detail, Bader[15] proposed a reduced one-dimensional model, making the following approximations:

- Slope (slope angle ψ) with z-axis perpendicular to the slope upward from the ground and translational symmetry in x- and y-directions.
- Energy contribution due to air and water vapour taken into account only at the surface (sensible and latent heat exchanges) or via an absorption function inside the snowcover (wind pumping).

The set of governing equations is as follows:

$$\frac{\partial \rho^{(\omega)}}{\partial t} + \frac{\partial}{\partial z}(\rho^{(\omega)} \upsilon_z^{(\omega)}) = m^{(\omega)}, \omega = 1,2 \tag{25}$$

$$\frac{\partial \sigma_{iz}^{(1)}}{\partial z} + \rho g_i = 0 \quad , i=y,z \tag{26}$$

$$\phi^{(2)}\upsilon^{(2)} = \phi^{(2)}\upsilon^{(1)} - \frac{k^{(2)}}{\mu^{(2)}}\left[\frac{\partial p^{(2)}}{\partial z} e_z + \rho_w g\right] \tag{27}$$

$$\rho C_{eff}\frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left(\lambda_{eff}\frac{\partial T}{\partial z}\right) + r_{tot} + m^{12} L_{12} - \left(\rho^{(1)} C_p^{(1)} \upsilon_z^{(1)} + \rho^{(2)} C_p^{(2)} \upsilon_z^{(2)}\right)\frac{\partial T}{\partial z} \tag{28}$$

Here, the superscript ω = 1,2 denotes the parameter associated with ice and water, respectively. Eqn (25) describes the density change due to settling, phase change and percolation. Settling is described by Eqn (26). The constitutive equation used here is the linear viscous law:

$$\sigma_{ij}^{(1)} = 2\mu^{(1)}\left[D_{ij}^{(1)} + \frac{1}{m-2}\delta_{ij} D_{kk}^{(1)}\right] \tag{29}$$

where m is the inverse of the viscous analogue of Poisson's ratio; δ, Kronecker symbol and $D_{ij}^{(1)} = (1/2)[v_{i,j}^{(1)} + v_{j,i}^{(1)}]$, the strain rate tensor. Eqn (27) describes the melt-water percolation and the heat conduction with source. The term e_z is the unit vector in z direction. The melting term is described by Eqn (28). Mathematically, Eqns (25-28) represent a set of initial and boundary problems. To take into account the interaction of snowcover with the atmosphere, the boundary conditions and source terms can be derived in two different ways:

- By using measured values from a test site
- By using phenomenological principles

Under the second approach, it is assumed that a snowpack can exchange energy and mass via (i) the surface/ground, or (ii) internal exchanges. The external energy absorbed at the surface is conducted into the snowcover. This energy is different from a direct absorption inside the snowcover. The components of the first process are long wave

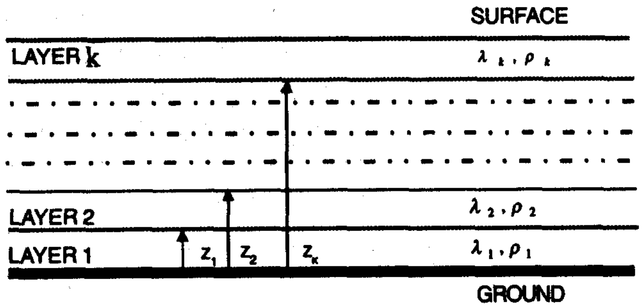

Figure 7. Layered snowpack [z_i, ρ_i, λ_i are the height, density and the conductivity, respectively of layer i].

radiation, turbulent exchanges with air and precipitation. The second process comprises short wave radiation and wind pumping.

Knowing the values of all the parameters like temperature, density, thickness, etc. at time t_n, the corresponding values are calculated at time t_{n+1}. Layered snowpack, as used in this approach, is shown in Fig. 7.

To numerically solve the governing equations for each layer, Bader[15], *et al.* also applied finite difference schemes, using implicit Cranck and Nicolson method, which is unconditionally stable for moving boundaries.

Using the easily measurable snow-meteorological parameters—amount of fresh snow, fresh snow density, insolation, albedo, air temperature, humidity and wind speed recorded at a regular interval of three hours—Ganju[17], *et al.* proposed a relatively simpler approach. Three-hourly snowfall records after averaging into an equal amount of one-hourly snowfall records has been densified by half of its own weight and overburden. The process of densification of each hourly layer is followed separately till the end of the season. Since the rate of densification reduces considerably after it attains a density value of 0.45 g/cc or more, layers with density value greater than 0.45 g/cc have been merged with the adjacent layers. This has been done to reduce the computational complexity. However, layers from two different storms have not been merged due to possible differences in their grain properties. The density of top snow layers where melt-water has percolated has been modified to accommodate additional water and the frozen layer is treated as a single layer for further densification. The concept of satisfying cold content of the top 5 cm layer has been proposed, in order to calculate the melting or freezing depth. In case energy exchange at the snow-air interface is found positive and snow surface temperature has reached 0 °C, melt has been affected after satisfying the cold content of the top 5 cm layer of the snowpack. Similarly, when energy exchange at the snow-air interface has been found negative, the re-freezing depth has been calculated, provided the snow surface temperature has reached 0 °C and has effected melt during the preceding hour.

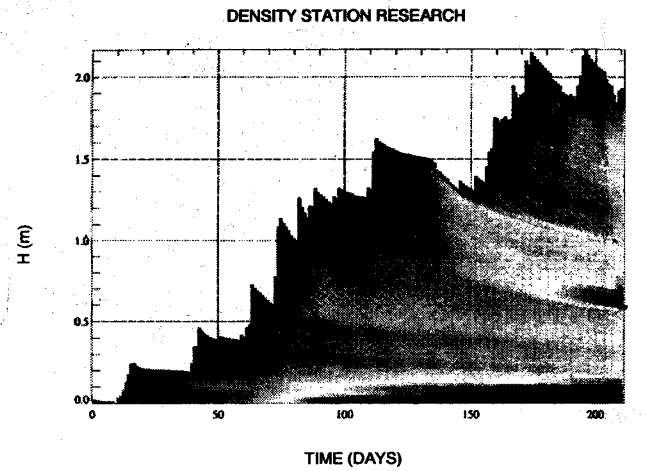

Figure 9. Density profile evolution at a Swiss station, as simulated by the model developed by Lehning[26] at SLF.

The computation for melt-water percolation commences the moment melt-water surfaces. Melt-water produced at the snow surface is allowed to percolate down into the snowpack either as a wetting front or as a wave front, depending on the saturation level the snowpack has attained. After effecting melt, the top snow layer, equivalent to snowmelt, has been removed and the density of the underlying layers into which melt-water has percolated is modified. Once the irreducible water saturation limit has been satisfied, the melt-water is allowed to percolate as the wetting front. The moment the wetting front reaches the bottom of the snowpack, melt-water percolation within the snowpack progresses as wave front. Change in the size of melt-freeze grains is computed every day once the propagation of melt-water within the snowpack commences.

The models developed by Brun[18], *et al.* (CROCUS), Lehning[15], *et al.* (SNOWPACK) and Ganju[17], *et al.* (SCM1) are shown in Figs 8, 9 and 10, respectively.

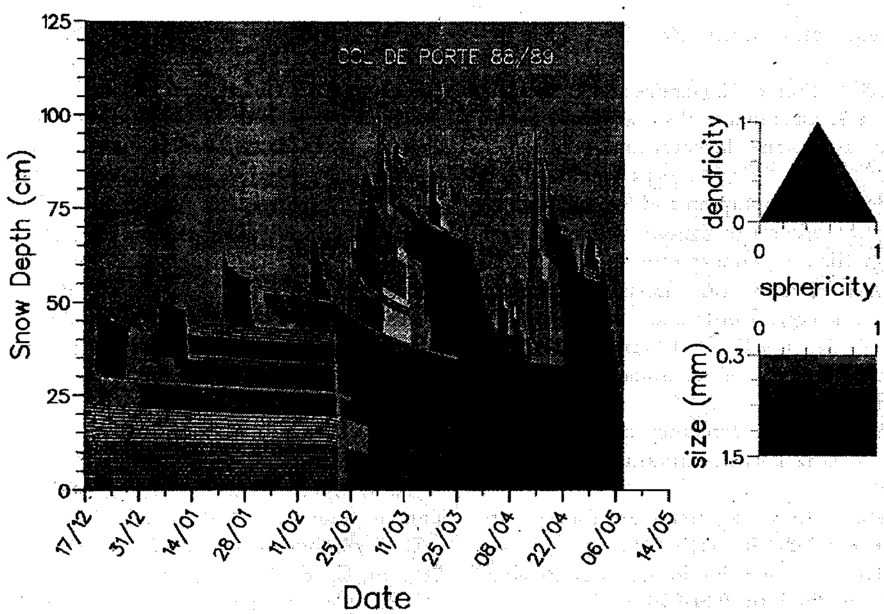

Figure 8. Continuous evolution of stratigraphy on an experiment site simulated by CROCUS throughout the winter season of 1988-89.

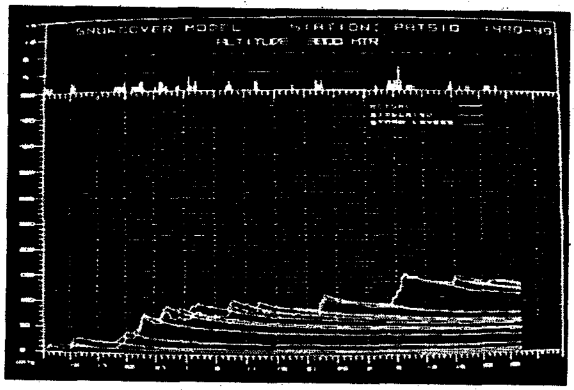

Figure 10. Evolution of different storm layers at a SASE station in Himalayas, as simulated by the model developed by Ganju, *et al.*

6. GREY AREAS

The snowcover simulation model has an important role to play in the prediction of avalanches, snow melt run off, mountain hydrology, agriculture and sub-soil characteristics. Considering its importance, it is all the more important to take a fresh look at the modelling approach adopted by various workers. While considering the complexity of various processes involved in snowcover evolution, a more realistic model with less mathematical jargon needs to be developed. The present-day focus on snowcover simulation has to be addressed on two fronts:

- Experimental work
- Adoption of a simplified modelling strategy that can be used for various operational works.

A thorough experimentation programme has to be chalked out to establish expressions that can be used in snow simulation. They are:

(a) Viscous deformation law for different crystal types

(b) Strength characteristics of snow layers having different crystal morphologies

(c) Melt-water percolation and preferential flow paths

(d) Weak layer formation and its incorporation in the model.

While the above task is being attempted, a modelling strategy has to be evolved so as to be able to run the model in the absence of complete answers to all the queries. As and when more answers are available, the model should be upgraded on the basis of the latest findings.

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