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METAMORPHISM AND REMOTE SENSING PHYSICS OF SNOW

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SUMMARY

Snow and ice are significant elements of the world hydrological system, which occur subject to tremendous variations in space.

After the snow is deposited the particle shapes are modified by a process known as metamorphism. Thus dendritic crystals decompose into fragments and the larger fragments grow at the expense of the smaller ones. This process continues until the fragments have been reduced to more or less rounded grains of ice or until a significant temperature gradient develops within the pack. Snowflakes undergo a rapid metamorphism that reduces their surface area and brings them to a more stable thermodynamic state.

Since vapour pressure is temperature dependent, temperature gradients produce associated vapour pressure gradients, which cause water vapour to diffuse from warmer to colder parts of the snowpack. Metamorphism can also result from compaction caused by the pressure of overlying layers of snow. This process is responsible for transforming snow into glacial ice whose crystals sometimes attain sizes of the order of 10 cm. (Langham, 1981).

In case of snow, crystal size, temperature, liquid water, density, thickness vary within a short distance to a great extent. So the field measurements about snow which are point data may be misleading. Besides, the snow bound areas are remote and hazardous to access. Remote sensing

techniques offer an excellent synoptic view in various spectral channels of electromagnetic spectrum which serve as a spatial data base for snow related studies.

Spectral reflectivity of snow is dependent on snow parameters such as grain size, and shape, impurity content, near surface liquid water content, depth and surface roughness, as well as solar elevation. Freshly fallen snow has a very high reflectance in the visible wavelength, but decreases as it ages. Thermal-infrared portion of the electromegnetic spectrum enables to the observations in the night-time passes of the satellite.

Microwave measurements have the capability to penetrate the snow and respond to variations in sub-surface properties. It permits remote observations of snow under nearly all weather and lighting conditions.

The objective of the report is to apply remote sensing studies on the metamorphism of snow. The report highlights the various remote sensing techniques available for the study of snow parameters. The reflectance goes down as density increases, as snow undergoes metamorphism.

1.0 INTROUCTION

Snow, its formation, deposition and melting, is an integral phenomenon of the hydrologic cycle. It is a variable source, after being deposited on the Earth's land surface, subsequently serves as a source of water vapour input to the atmosphere through the process of sublimation and evaporation and as a source of water to the soil and river systems when melting occurs. Though it is affected by weather, it affects the weather and zonal conditions.

The various forms of snow crystals, that constitute newly deposited snow, results from the variation in the temperature and humidities of the atmosphere at the time of their formation and the action of the wind during their descent to the ground. Once on the ground, snowflakes undergo a rapid metamorphism that reduces their surface area and brings them to a more stable thermo-dynamic state. This process is accompanied by an increase in the strength of bonds between grains. The introduction of water into the snowcover, in the form of either rain or snowmelt, causes a rapid metamorphism that eliminates the smaller particles and reduces the larger particles to a more or less spheroid shape.

Remote sensing has become a valuable tool in snow and ice studies because of its unique capability for aquiring measurements of glaciological conditions over large areas.

The reflectance of snow depends not only on the intrinsic properties of snow itself relative to such factor such as grain size and liquid water content but also on surface contaminants.

Generally snow reflectance decreases, at all wavelengths, with impurities and with increasing grain size. Consequently, as snow ages, reflectance decreases, melting coditions als tend to decrease snow reflectance.

Sensor band in visible and near infrared range is very useful for albedo measurements, thermal infrared is useful to snow temperatures.

The remote sensing interest is exploiting microwaves stems from the fact that the influence of the atmosphere in general is very small so that an essentially all-weather system can be implemented in contrast to visual and infrared techniques which are influenced by cloud cover, for instance. Also microwaves give a larger penetration into the surface observed than infrared so that sub-surface features may be studied. This technique is very useful for monitoring ice and snow and related dynamic processes.

The reflectance for snow changes especially when meltwater appears and snow crystals metamorphose to coarse grains and density increases. The reflectance goes down with increase in density of snow.

In an experimental phase extensive ground measurements and field observations should be carried out in addition to remote sensing studies.

2.0 SNOW METAMORPHISM

2.1 Snow Deposition:

The purpose is to describe the physical properties of snow on the ground and the processes that take place within the snow cover to change these properties.

Ice particles that form in the atmosphere have a large variety of crystal habits and sizes. By the time they reach the ground they have already undergone a number of transformations resulting from growth, disintegration and agglomeration.

Wind speed near the surface also determines how crystals arriving at the surface are packed. At high speeds, crystals are broken first in the extremely turbulent layer in the lowest few meters above the surface, further contamination occurs as the crystals are bounced and dragged over the snow surface (saltation) either during a snow storm or afterwards as blowing snow. After being reduced in size and shape more symmetrically the crystal can be packed much more closely to produce much denser surface layer than would otherwise occur (Langham, 1981).

2.2 Evolution of Snow-Cover:

After the snow is deposited the particles shapes are modified by a process known as metamorphism. Thus dendritic crystals decompose into fragments and the larger fragments have been reduced to more or less rounded grains (A grain is a single crystal, in that all its molecules lie in the same three dimensional array, but whose surface is irregular in shape and does not display a habit, which is a set of plane outer surfaces whose orientation is related to the symmetry of the molecular array) of ice or until a significant temperature gradient develops within the pack.

Thermodynamically, the snow crystals are moving to an equilibrium state and the thermodynamic property which determines this state is the free energy which for snow crystals implies minimizing the ratio of surface area volume.

2.3 The Role of Free Energy in Metamorphism of Snow:

The principles of thermodynamics are often used to explain observed changes from the metamorphism of snow crystals. If a volume containing several snow crystals is assumed to be a thermodynamic system, then the differential changes in their state may be described by the first law of thermodynamics (or conservation of energy) according to the expression:

$$dU = \delta Q - \delta W \qquad .....(1)$$

where:

dU = the change in internal energy, U

δQ = heat added to the system, and

δW = work done by the system.

If it is assumed that work done on the system of snow crystals will only results in changes in volume or surface area of the ice then:

$$W = pdV + \xi dA \qquad ....(2)$$

where:

p = pressure,

V = volume,

ξ = surface energy/unit area, and

A = surface area

The second law of thermodynamics for a system of fixed mass is:

$$ds > \delta Q/T \qquad \text{.... (3)}$$

where;

dS = change in entropy

T = temperature of the system at the point of heat transfer

If the heat transfer takes place reversibly and temperature differences are kept infinitely small or if the system undergoes in the addition or removal of heat while being maintained at a constant temperature then:

$$dS = \delta Q/T \qquad \text{.... (4)}$$

For a completely isolated system the second law states that all changes occur such that

$$\Sigma dS_i > 0, \qquad \text{.... (5)}$$

where, the summation extends over the i phases of the system. For a given mass of snow implies that the summation should extend over all crystals plus the air or water vapour likely to influence the system during the period when the differential change of state takes place.

The equilibrium criterion maybe stated in terms of the Gibbs free Energy, as;

$$dG = \Sigma dG_i = 0, \qquad \text{.... (6)}$$

where, $G = U\text{-}TS + pV + \xi A$ .... (7)

Expanding (6) in terms of (7) gives

$$dG = dU - TdS - SdT + pdV + \xi dA + Ad\xi = 0 \quad .... (8)$$

Under conditions of constant temperature, pressure and surface energy the Gibb's free energy may decrease because of decrease in surface area or volume of one of the phases. Thus, when their temperatures are uniform the snow crystals tend to change their shape so that the ratio of surface area to volume approaches a minimum. Equilibrium is attained when changes produce no further decrease in the Gibb's free energy (Equation 6, i.e. $dG = 0$). While surface effects are important in a consideration of snow crystals some of the better known properties of the snow can be explained without reference to surface conditions. In this case it is convinient to consider a thermodynamic system consisting of pure water in the bulk state. Equations 1,2, & 3 still apply to the bulk state although equation 2 and equation 4 take correspondingly simpler forms;

$$\delta W = pdV \quad ....(9)$$

and

$$dG = dU - TdS - SdT + pdV + Vdp \quad ....(10)$$

if equ. (9) and equ. (4) are substituted into equ. (1)

$$dU = \delta Q - \delta W, \quad ds = \delta Q/_T, \delta W = pdV$$

$$dU = TdS - pdV$$

or;

$$dG = -S\,dT + Vdp \quad ....(11)$$

For a system with different phases (solid, liquid, vapour) a condition of equilibrium requires that they have the same free energy per unit mass. (Watsopoulos and Keenan, 1965). It follows that changes in free energy per unit mass of each phase must also be equal if they are to remain in equilibrium. Thus,

$$dG_v = dG_1 = dG_s \qquad \ldots. (12)$$

where,

$$dG_v = -S_v\,dT + V_v\,dp \text{ (vapour)}, \qquad - 12a$$

$$dG_1 = -S_1\,dT + V_1\,dp \text{ (liquid), and} \qquad - 12b$$

$$dG_s = -S_s\,dT + V_s\,dp \text{ (Solid)} \qquad - 12c$$

Eliminating dG from the first two of the equations 12 a,b,c,;

$$- S_v\,dT + V_v\,dp = -S_1\,dT + V_1\,dp$$

or, $$(S_1 - S_v)\,dT = (V_1 - V_v)\,dp$$

or, $$\frac{dp}{dT} = \frac{S_1 - S_v}{V_1 - V_v}$$

or, $$dp/dt = (S_1 - S_2)/(V_1 - V_2),$$

in which the subscripts 1 and 2 refer to any two phases. At the point of equilibrium where $G_1 = G_2$ the transition between the two phases is reversible so that equation (2) applies. Therefore,

$$\Delta S = S_1 - S_2 = Q/T = L_{12}/T, \qquad \ldots. (13)$$

where L_{12} is the latent heat for the phase transition.

Hence:

$$dp/dT = L_{12}/(T\,\Delta V_{12}) \qquad \ldots. (14)$$

where V_{12} is the increase in volume during the phase change. This equation known as the clausius - Clapeyron equation, determines the pressure increase dp needed to maintain equilibrium for a temperature increase dT. The derivation of the three equations of this type does not depend on any assumptions about the phases, although here the discussion refers to a single-component system (i.e. pure water). The graphical representation of equation (14) is called a phase diagram, eg. fig.(1) for pure water.

When all the three phases are present in a system there is a unique solution of equation (14) which is the meeting point, of the three boundary curves at the triple point (+ 0.0099^{o}C, 0.615 kPa, for pure water).

In the presence of air at atmospheric pressure the position of these curves are modified very slightly so that they intersect at a new point, known as the ice point (0^{o}C and 0.6095 kPa). The free energy of ice crystals in air depends on the air pressure. Since this effect is extremly small and is same for all crystals.

In a dry snowpack deposited over a period of several days, temperature variations with depth are quite common. Once a significant temperature difference is established across any layer within the snow cover (Snowcover is used in the most generally context; and snowpack in reference to deep accumulations) the process of metamorphism is completely altered. The resulting temperature differences are associated with vapour pressure differences which can be calculated using an integrated form of eq. 14 (Denfigh 1966).

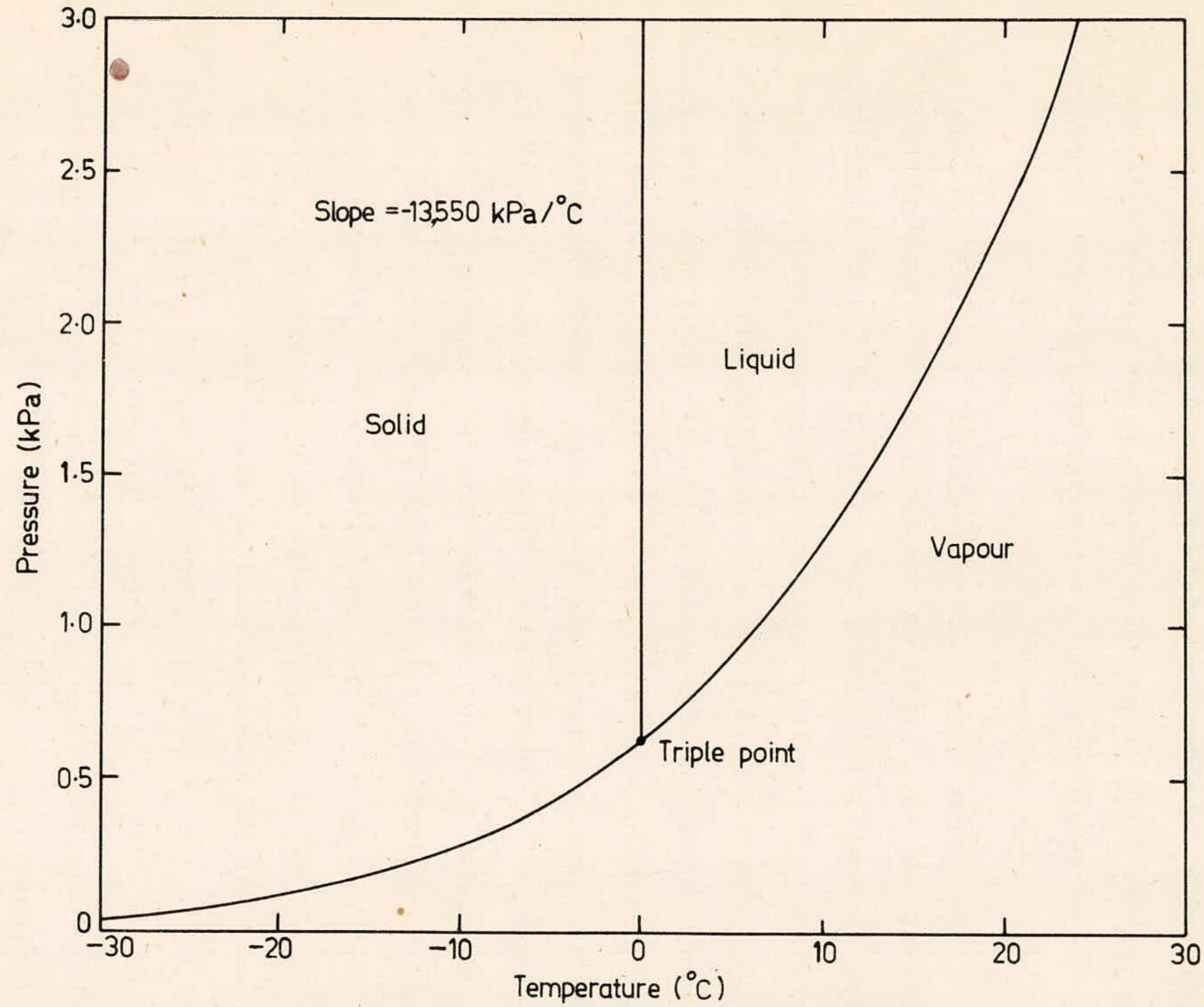

FIG. 1 – PHASE DIAGRAM FOR PURE WATER

Since vapour diffuses towards lower pressures, there is a net transfer of material from the warmerpart of the snow pack where the snow crystals have a higher vapour pressure to the colder part where they have a lower vapour pressure. This process is referred to as temperature gradient metamorphism.

The water vapour is transfered from one ice grain compositing the snow layer to the next in a "hand-to-hand" transfer process. Thus there is a series of transfer involving movement from the solid phase to the vapour phase and back to solid; a significant portion of the snow will pass through the vapour phase and be deposited as per of a new crystal. The newly formed crystals (known as depth hoar) take many shapes but have a characteristic layer structure resulting in a stepped or ribbed surface facets(Roshida et al, 1955). These crystals are only weakly bounded together. These crystals frequently grow in the lowest layers or a cold snowpack as a result of diffusion of vapour from the relatively warmer soil below. They may also occur elsewhere in a snowpack, i.e., after cold new snow is deposited on a relatively warmer pack. Frost on the upper surface may also consist of crystals with similar shapes, but they form because of noctural radiative cooling of snow surface and are known as hoar frost.

Metamorphism can also result from compaction caused by the pressure of overlying layers of snow. This process is responsible for transforming snow into glacial ice whose crystals sometimes attain sizes of the order of 10 cm.During early stages, the refreezing of melt water can accelerate the densitification process.

Metamorphism of snow crystals takes place even in the absence of a temperature gradient and its corrosponding vapour pressure differences. During the first few hours after snow is deposited (before a significant temperature gradient can be established within it), destructive metamorphism is the dominant process. Its first effect is to reduce snow consisting of dendritic or needle-like crystals to relatively round grains. Mass transfer via the vapour phase is known to play an important role in this process but the mechanism is not clearly understood.

On the otherhand, the vapour pressure can vary over an irregular surface such as that of a snow crystal. Adam (1941) showed that the pressure differences across the curved interface between two phases may be expressed as:

$$dp_s - dp_v = \xi\left(\frac{1}{r_1} + \frac{1}{r_2}\right) \qquad \text{.... (15)}$$

where;

dp_s = difference between the vapour pressure of the solid at a curved surface and that of an infinite plane

dp_v = corrosponding pressure difference for the vapour phase

ξ = surface energy per unit area, and

r_1, r_2 = principle radii of curvature of the ice surface

In general equation (15) suggests that curvex surfaces have a higher vapour pressure and concave surfaces a lower vapour pressure than plane surfaces.

In case where the surface is anticlastic a net increase in vapour pressure may still occur. For example, although the radius of curvature of the narrowest part of the crystal may be infinite or even large and negative in one direction, its equilibrium vapour pressure may still be greater than that of a flat surface because the radius of curvature in the plane perpendicular to its length is small. A relatively flat, plate-shaped crystal where both radii of curvature are large produces a very small increase in the vapour pressure. If such high and low vapour pressure areas exist near each other then diffusion towards lower pressure can transfer ice by sublimation from one area to the other.

When the ice grains are subject to local stress such that dp_s is positive, then, for the same radius of curvature dp_s increases. Although, dp is defined herein as a compressive stress, and shear stress, or combined stresses such as those for bending, will also increase the free energy and thus also the vapour pressure. The experiments of Yoshida and others (1955) support this theory.

Once the dendritic or needle-like crystals have been reduced to more or less round grains equi-temperature metamorphism can still continue through sublimation from the small grains to the larger ones. This reduces the net surface area and consequently the free energy.

Sintering also reduces the air/ice surface area and by increasing the area of contact between the larger crystals; Vapour transport is the dominant mechanism by which the areas of contact or neck between the sintering spheres grows. The gradient that causes vapour movement results from the difference in curvatures of the ice surface at the neck and the adjoing ice particles. This produces densification and increasing mechanical continuity which is responsible for the increased strength and hardness of the snow.

Any water in the snowpack collects at points of contact between grains. Colbeck (1973) distinguishes two saturation regimes in wet snow; the pendular regime-saturation < 14% of the pore volume, with air in more or less continuous paths throughout the snow ;the funicular regime - saturations > 14% of the pore volume with air in distinct bubbles. According to colbeck (1973) the metamorphism of wet snow is best considered by looking at local temperature differences in the vicinity of the grain boundaries. These differences exist as a result of the radii of curvature of the solid-liquid, liquid-vapour or solid-vapour interfaces. He showed that in the funicular regime heat flows from the larger to the smaller grains causing them to dissipate while the larger grains increase in size. In the pendular regime the radii of curvature of the ice grains have less effect on the transfer process so that the capillary effects dominate, resulting in a reduction of temperature differences between particles. Also, the amount of liquid through which heat may flow is reduced. Hence much lower rates of grain growth are observed. Colback states that the snow strength in the pendular regime is quite high, since no melting occurs at the grain contacts, whereas in the funicular regime it is relatively low because very little bond-to-bond strength exists. On refreezing, the snowpack becomes very strong beacuse of the continuity between grains. This process is melt-freeze metamorphism, also referred to as firnification in perennial snowpack.

Recrystallization, is somewhat peripheral to snow studies, since it relates mainly to crystal growth in polycrystalline ice, it is also partly responsible for transforming snow into ice. Variations in stress, and consequently free energy, occur in the matrix of crystals, within the layer of deposited snow. Such variations act to transfer material between crystals at snowpack densities exceeding 580 kg./m^3, corresponding to close random packing of loose grains. The transfer mechanism at these densities is volume diffusion through the ice lattice to the highly stressed zone at the area of contact of crystals. The rate of compaction is independent of pressure at least four times upto 15 hours (Hobbs, et.al, 1967). However, as the pressure of the overlying snow,increases, viscoplastic flow may occur and dominate the densification process.

During viscoplastic flow, the crystals are permanently deformed. To explain the relationship between this flow and grain growth the crystal structure of the ice must be considered. The oxygen atmos, and hence molecules of ice, are arranged on the regular three-dimensional lattice. In the direction of the c-axis the lattice has hexagonal symmetry, i.e. a rotation of 60° (and a displacement in the C-direction) brings the lattice into coincidence with the original molecular arrangement. This symmetry is responsible for the hexagonal crystal habits of plates, stars and prism of snow. In an ice crystal lattice the molecules lie in planes whose separations are greatest perpendicular to the C-direction. Therefore, the molecular bonds between such planes are the weakest so that the resistance of ice to shear is least for stresses parallel to them. This plane of least resistance is called the basal plane.

In a snow subject to high stresses resulting from the weight of the overlying layers or the slope of the snow-cover, permanent deformation takes place in those crystal lattices oriented so that the shear stress is parallel to the basal planes. Other crystals, withstand higher stresses before deforming permanently, but gain free energy. This increase acts so as to transfer material to those crystals which have undergone plastic deformation.

Also, the number of defects in one crystal lattice affect the free energy, and hence the growth of crystal. The lattice arrangement is usually not perfect, and can have various kinds of dislocation in the regular array of molecules. These constitute a form of internal free energy and are more numerous in damaged crystals. Dislocations move most easily in the basal plane, that further reducing resistance to shear. In other directions the stresses tend to increase the concentration of dislocations and therefore their free energy. Thus in addition to the elastic energy, the dislocation free energy caused by inelastic deformation is conductive to preferential grain growth.

Because the metamorphism is controlled by and in-turn influences, heat and vapour transfer within snow, there is a close alignment between the snow hydrology community and the snow physics community. Climate interactions are affected by the physical properties of the snow pack, and the metamorphism of wet snow clearly influences of the flow of water through snow (Berg, 1982, Denoth, 1982, Colbeck, 1982). Important recent advances in the study of snow metamorphism include improved specification of dry snow metamorphism, laboratory experiments on metamorphism, are better methods to measure snow characteristics.

Adams and Brown (1982), Colbeck (1983), Sommerfeld (1983) and Gubler (1988) have considered the details of vapour transfer and crystal growth within the snow cover. The major differences in their formulations lie in the effects of local temperature differences vs. local geometric effects. One current problem is that the terminology currently in use temperature gradient metamorphism and equitemperature metamorphism (Colbeck, 1986). The rounded forms we associate with so called equitemperature metamorphism apparently require a small temperature gradient to occur; under truly equi-temperature conditions metamorphism is very slow (Perla and Ommanney, 1985).

Research on snow metamorphism also depends on improved methods of defining and measuring the properties of snow. We understand the mechanism of the transfer processes of heat and water vapour in the snow pack, but we do not know the geometry of the grains and bonds (Colbeck, 1986). Although important qualitative information about the morphology of snow grains and estimates of mean and maximum grain size can be obtained from analysis of disaggregated crystals, serious disadvantages prevent easy acquisition of other quantitative data about the snow-ice phase. Disaggregation destroys an unknown number of necks and connections between grains as well as their orientation.

All the phenomena mentioned above are responsible for the evolution of snowcover. However, it should be clear that the variation of snow cover structure, because of its mode of deposition and subsequent evolution, produces an extraordinarilly complex medium changes in the snowcover are closely related to the metamorphic processes.

During the decade 1960-1970, interest in physical processess occuring in the snowpack resulted in an emphasis on classification based on the state of metamorphism. The best classification of this type was presented by Sommerfeld and La Chapell in 1970 (Table 1). This classification is based on a more complete physical knowledge.

Table - 1

CLASSIFICATION OF SNOW BY METAMORPHIC STATE

(Sommerfeld and La Chapelle, 1970)

1. Unmetamorphosed Snow

   A. No wind action: Many fragile snow crystal forms easily distinguishable, little difference from snow in air.

B. Wind blown: Shards and splinters of original snow crystal; parts of original forms may be recognizable but whole forms very uncommon.

C. Surface hoar.

2. Equi-Temperature metamorphism

A. Decreasing grain size

i) Beginning: Original snow crystal shapes recognizable, but corners show rounding and fine structure has disappeared.

ii) Advanced: Very few indistinct plates or fragments recognizable; grains show distinct rounding.

B. Increasing grain size

i) Beginning: No original crystal shapes recognizable; grains show a distinct equi-dimentional tendency, a few indistinct facets may be visible.

ii) Advanced: Larger equidimensional grains present; a strong tendency toward uniform grain size; faceting generally absent.

3. Temperature-gradient metamorphism

A. Early: the result of a strong thermal gradient on new-fallen snow; associated with the first snowfalls of the season.

(i) Beginning: angular or faceted grains common; stepped surfaces not visible.

(ii) Partial: medium-sized angular grains predominate; poorly formed steps visible.

(iii) Advanced: medium to large angular grains predominate; well-developed facets and steps visible; a few filled or hollow cups may be found.

B. Late: The result of a strong thermal gradient acting on snow in the later stages of equi-temperature metamorphism.

(i) Beginning: medium to large angular or faceted grain predominate; some stepped surfaces visible.

(ii) Advanced: large grains predominate; many very fragile hollow cups or lattice grains; very deep steps.

4. Firnication

A. Melt- freeze metamorphism

(i) Limited: single thaw-freeze cycle and limited gain in ice density.

(ii) Advanced: repeated thaw-freeze cycles and appreciable gain in density and mechanical strength; density range 600 to 700 kg./m³.

B. Pressure metamorphism

(i) Beginning: grains deformed and rearranged by pressure; density reange 700 to 800 kg/m^3.

(ii) Advanced: pore spaces become non-communicating; Permiability Zero; density range 800 to 830 kg/m^3.

3.0 SPECTRAL REFLECTANCE CHARACTERISTICS OF SNOW

Snow is not a simple substance. So defining its optical properties can be quite complex. Spectral reflectance properties of snow are dependent upon the grain size, liquid water content, shape, depth and surface roughness of snow pack (Dozier et al. 1981). Freshly fallen snow has high reflectance in the visible spectral band but as it ages the reflectivity of snow decreases particularly in the infrared wavelengths (O' Brien and Munis, 1975). Dozier (1985) has carried out extensive studies to show the dependence of IR reflectance on the snow grain size.

Field Studies of reflectance characteristics have been studied by Dhanju in the snow fields of the Himalayas.

3.1 Field radiometer

The field radiometer used to collect appropriate data of the target areas has seven bands in the following range.

Band No.	Range (Micrometers)
1	.433 to .453
2	.510 to .580
3	.500 to .640
4	.648 to .692
5	.700 to .800
6	.818 to .883
7	.960 to 1.040

A barium sulphate coated white plate was prepared using standard Kodak paint which was specially meant for making non-reflecting white surface. The observations taken from the white plate with the radiometer were used to normalize the radiometric values of the target areas.

3.2 Data Collection Site

From March 19 to 26, 1987 spectral observations were carried out below Rohtang pass, Manali (HP) at the height of about 2600 meters for snow. Simultaneously observations were also carried out for density and grain size. A snow pit was dug and spectral reflectance from the various layers together with the grain size and density of each of the layers were recorded.

3.3 Spectral reflectance characteristics of snow

Freshly fallen snow in general shows a density value 0.1 gm/cc. But immidiately after it has fallen metamorphic processes start depending upon the environmental parameters such as ambient temperature, wind velocity etc. resulting in the increase in density and grain size. Compaction may start of the different layers of snow if further snowfall accumulates over the top of the layer. The observations were confined in the area immidiately above the snowline and so mostly wet snow was encountered. A snow pit was dug out and sequentially its four layers were exposed for observations. The various parameters of each layer are as follows:

1.	First layer:	Snow density	:	0.4 gm/cm^3
	(Top)	Grain size	:	0.5 mm^2
		Temperature	:	-1.5^{o} C
		Depth of layer	:	16 cm.

2.	Second layer :	Snow density	: 0.41 gm/cm^3
		Grain size	: 0.5 mm^2
		Temperature	: $-2.1^{o}C$
		Depth of layer	: 10cm.
3.	Third layer :	Snow density	: 0.44 gm/cm^3
		Grain size	: 1 mm^2
		Temperature	: $-2.5^{o}m$
		Depth of layer	: 13 cm
4.	Fourth layer :	Snow density	: 0.48 g/cm^3
		Grain size	: 1.5 mm^2
		Temperature	: $-2.5^{o}C$
		Depth of layer	: 18 cm.

Each of the layer was exposed and spectral reflectance observations were carried out. These values were normalized using Barium sulphate white plate.

The normalized values are plotted in Figure 2 showing spectral band versus percent reflectance for different snow densities.

3.4 Snow Density and Spectral Reflectance

Spectral reflectance of snow is dependent upon parameters such as grain size, liquid water content, surface conditions and depth of snow cover.

Figure 3 shows the plot of the spectral reflectance percentage and snow density as observed in band 1 and band 7. There seems to a distinct relationship between the density of snow and the spectral reflectance, i.e. higher the density lower is the reflectance value. The observations were limited between the density values of 0.4 to 0.5 gm/cm^3.

Freshly fallen snow having density value in the neighborhood of 0.1 gm/cm^3 shows high reflectance. But as the snow field goes through one or two diurnal temperature cycle, compactation starts taking place resulting in the increase in density.

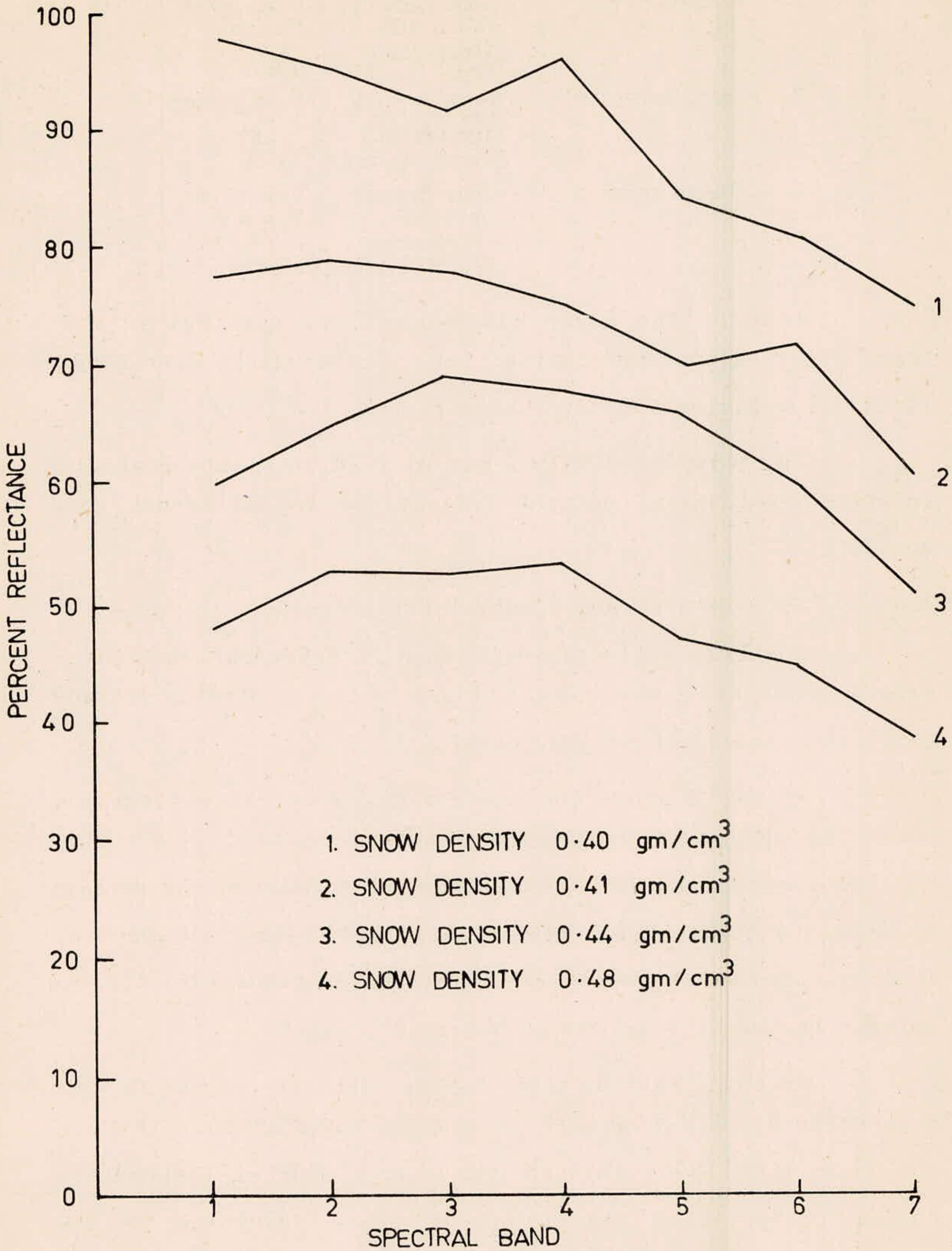

FIG.2-SPECTRAL REFLECTANCE OF SNOW

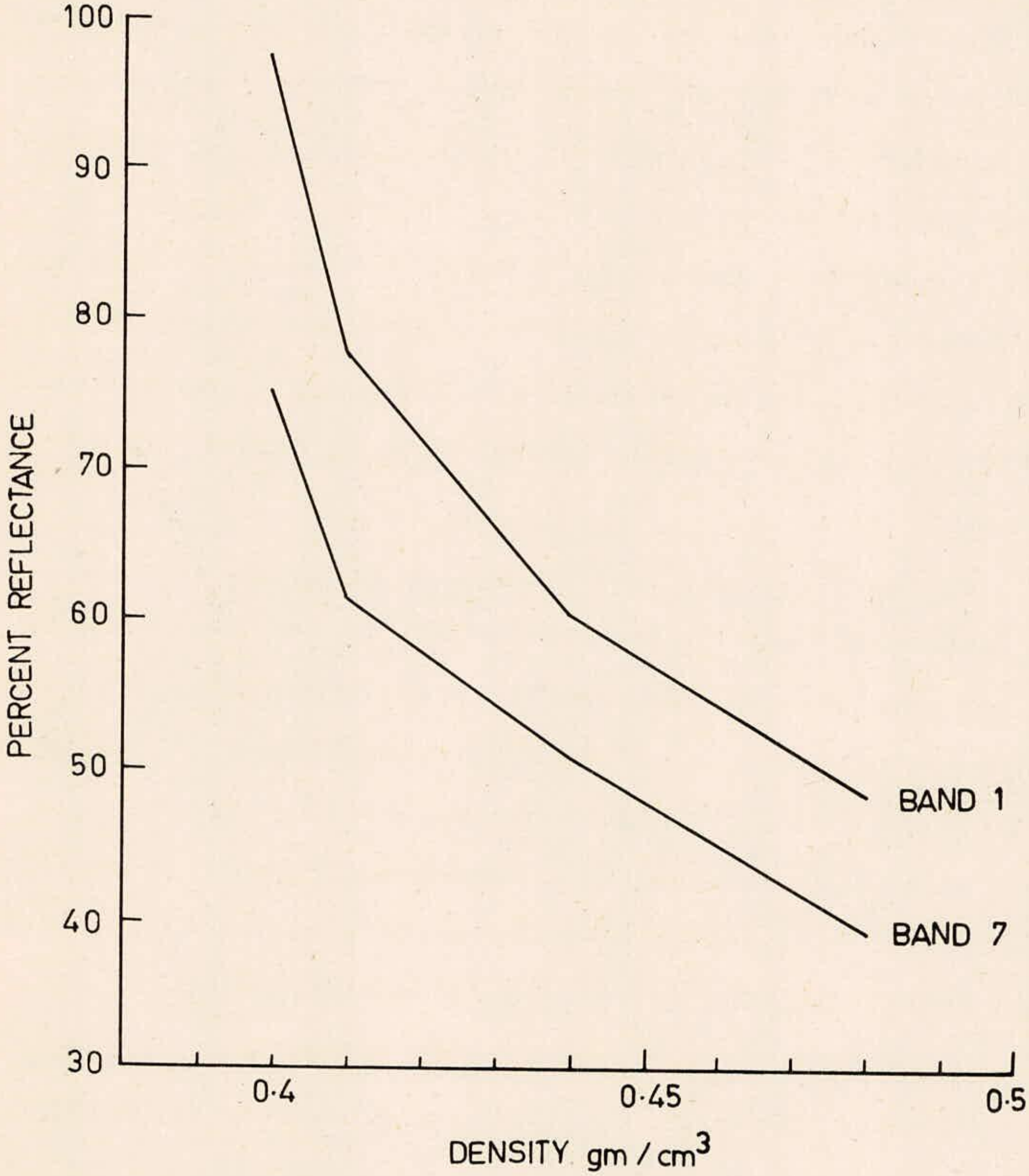

FIG.3 - SNOW DENSITY AND SPECTRAL REFLECTANCE

4.0 REMOTE SENSING OF SNOW COVER

4.1 Introduction

Ground-based measurement techniques will not always detect significant areal snowcover variations. In mountaneous regions large variations in snow depth and water equivalent occur because of variations in slope, aspect, elevation, exposure and surface cover. In specific areas, accessibility can limit both the number and representativeness of ground measurements. In plain regions, variations in snowcover are dominated by local land use and topographic variations, in addition to the spatial variations in initial snowfall distribution.

For many studies the percentage snowcover in a basin is an important areal parameter which ground measurements alone may not provide with sufficient accuracy, especially in sparsely-instrumented regions. The development of remote-sensing techniques for snowcover applications has provided new methods for obseravation, measurement and analysis. The successful application of these methods relies on accurate "ground truth" data for calibration and verification.

Satellite remote sensing has provided a promising alternative for the areal analysis of snowcover over large regions.

Freshwater in high mountain environments derives primarely or to a substantial part from the melting of the seasonal snowpack. Therefore, snow is a very important natural resource which needs continuous monitoring. This may be undertaken best by remote sensing techniques, especially by high resolution satellite systems.

Snow mapping with remote sensing techniques has several advantages:

- Large areas can be observed simultaneously, allowing a truely regional comparision (which is impossible on the ground).
- Areas which are remote and inaccessible can be surveyed.
- The reflectance characteristics of snow are distinct and - in general-clearly separable from other features.
- Only a single feature - snow - has to be classified and separated from all other objects (whilst for other purposes, eg. land use, there are always numerous catagories to be distinguished).
- Satellites yeild spatially continuous data, contrasing strongly with those obtained from net-work of surface stations.

4.2 Gamma Radiation Studies

Detection of natural gamma radiation emissions from the Earth has been used to measure the water equivalent of snow, Gamma ray detection must be carried out from aircraft at low attitudes (about 150 m) because of the significant atmopheric attenuation of the radiation. Background gamma radiation of the soil is obtained before the snow falls, and subsequent flights are flown to measure the gamma radiation through the attenuating snow cover. The degree of attenuation is related to the snow water equivalent through various calibration curves. Deep mountaneous snowpacks may drastically attenuate the gamma radiation, and mountaneous terrain presents obvious safety problems for aircraft. In addition, interpretation of the data becomes more confusing when the soil moisture level changes significantly between the calibration flight made before the snow and the snow seasons flight. Because of the low altitudes of the flights and narrow width of the data swath that is obtained, the gamma radiation method is confined to measurements of limited index (Rango 1985).

The gamma radiation flux near the ground originates primarily from the natural potassium (40_K), uranium (238_U) and thallium (208_{Tl}) radioisotopes in the soil. In a typical soil, 96% of the gamma radiation is emitted from the top 20cm. After the background (i.e. no snowcover) radiation and soil moisture are measured over a specific flight line, the attenuation of the radiation signal due to the snow pack overburden is used to calculate the amount of water in the snow. The snow water equivalent values are calculated by measuring the attenuation of the gamma radiation flux with data from the K window (1.36 - 1.56 MeV), the T1 window (2.41 - 2.81 MeV). The potassium "photopeak", the K window, is consistently the strongest in the energy spectrum.

4.3 Optical Studies

Albedo, which is the ratio of the reflected to the incoming solar radiation, is a property of snow that is especially suitable for remote sensing target. It amounts to 90% or even more for a freshly fallen snow cover and drops below 40% of the snow surface is weathered and dirty. Consequently, fresh snow is easily recognized on satellite imagery, while old snow cover may be less reflective than some snow-free surfaces.

Spectral reflectivity of snow is dependent on snow parameters such as grain size and shape, impurity content, near-surface liquid water content, depth, and surface roughness, as well as solar elevation. Freshly fallen snow has a very high reflectance in the visible wavelengths. As it ages, the reflectivity of snow decreases in the visible wavelengths and even more in the longer (near-infrared) wavelengths. (Figure 4).

This greater decrease in the near-infrared wavelengths is largely due to grain size increases that are caused by melting and refreezing within the surface layers and also to the natural addition of impurities. Melting of snow also increases the mean grain size and density by melting the smaller particles.

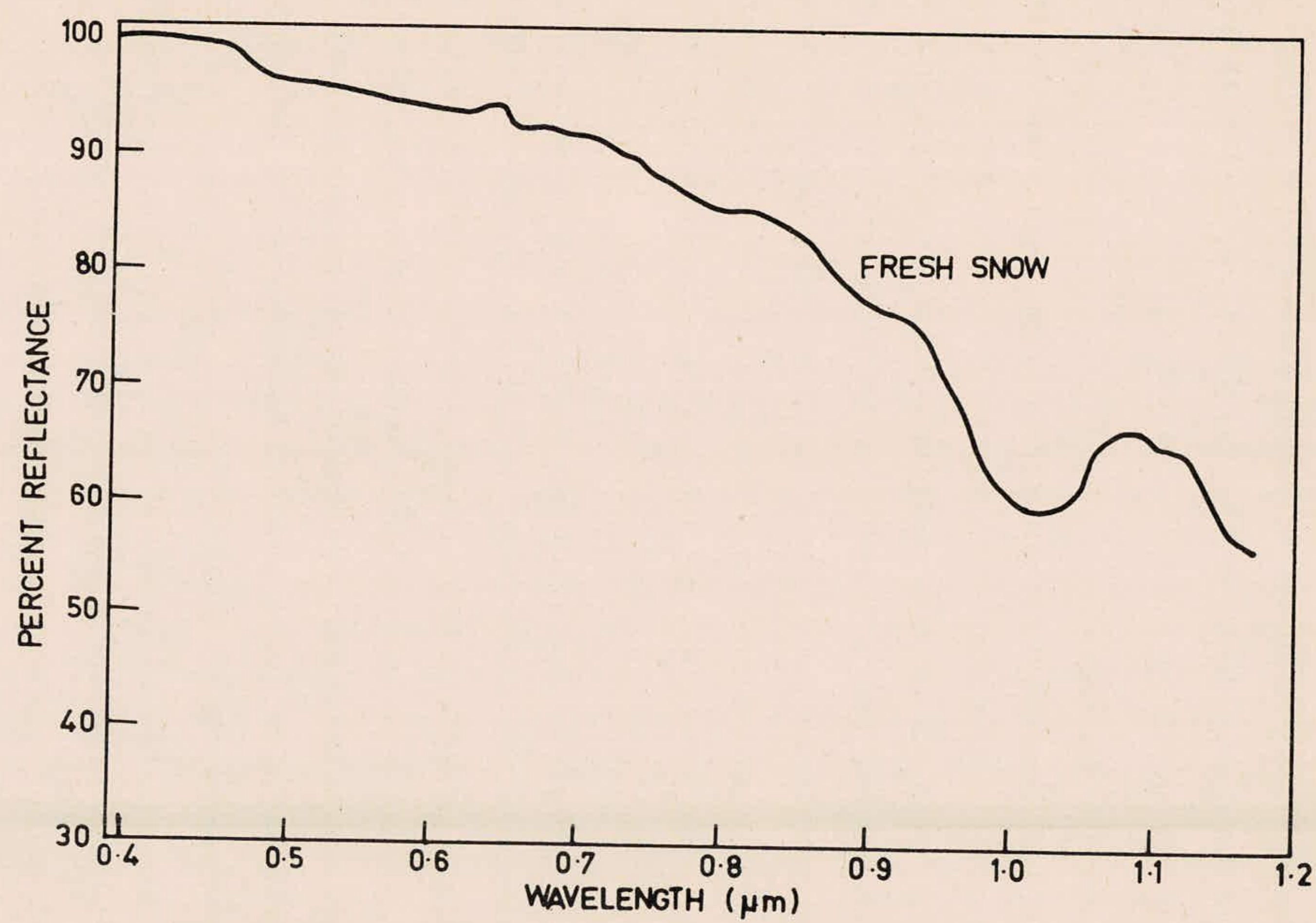

FIG. 4 — SPECTRAL REFLECTANCE CURVE OF SNOW

In the visible (0.4 - 0.7 µ m) bands snow reflectance is not sensitive to grain size, so measurements in these wavelengths will show the extent to which snow albedo is degraded by contamination from atmospheric aerosols, dust, pine pollen, etc.

Dozier has shown that, in the red and near-infrared (0.7-1.0 µ m) snow reflectance is sensitive to grain size but not to contaminants, so grain size estimates in these wavelengths can be used to extend albedo measurements spect-rally.

The reason that snow reflectance in the 0.4 - 0.7 µ m range is not sensitive to grainsize is that ice is so transperent in these wavelengths that increasing the size of a snow crystal does not significantly change the probability that a photon impinging on the crystal will be observed. Impurities are much more absorptive than ice in these wavelengths, however, so small amounts of contaminants will affect reflectance. In the near-infrared, ice is shightly absorptive, so an incident photon is more likely to be absorbed if the crystal is larger, and snow reflectance is therefore more sensitive to grain size. Impurities are not so important at these wavelengths because their absorption coefficients are much larger than those of ice.

In the "shortwave infrared" (0.4 - 0.7/µm), snow is much darker than clouds, and water clouds are brighter than ice clouds. In these bands, ice is highly absorptive, and snow reflectance is low and sensitive to grain size for small sizes, which explains the higher reflectance of ice clouds when compared to snow. Addtionally, water is less absorptive than ice in this band, so water clouds are more reflective than ice clouds.

Snow albedo is increased at all wavelengths as the solar Zenith angle increases but is most sensitive around $\lambda = 1\,\mu m$. Cloud cover affects snow albedo both by converting direct radiation into diffuse radiation and by altering the spectral distribution of the radiation. Warren has showed, that the cloud cover also normally causes an increase in spectrally integrated albedo.

The albedo is just the "upflux" devided by the "downflux" at a particular wavelength, which is usually measured just above the snow surface. In general, the albedo depends on the distribution of incident radiation with angle, but it also depends on snow properties and the atmosphric composition (water vapour content, cloud thickness etc.), which affect the spectral distribution of the sunlight. Since snow is grainy, the individual grains play a role in reflecting light from both the surface and the interior of the pack.

Landsat Thematic Mapper (TM) data are currently being analyzed interms of their ability to provide reflectances of glaciers and ice sheets in various areas of the world. Because the TM spectral data are in descrete bands (Fig. 5), these data are useful for comparisons with the hand-held spectrometer data to facilitate the development of an algorithm to calculate snow albedo.

The landsat TM sensor aquires data in seven spectral bands. TM bands 1 through 5 and 7 are in the visible, near-

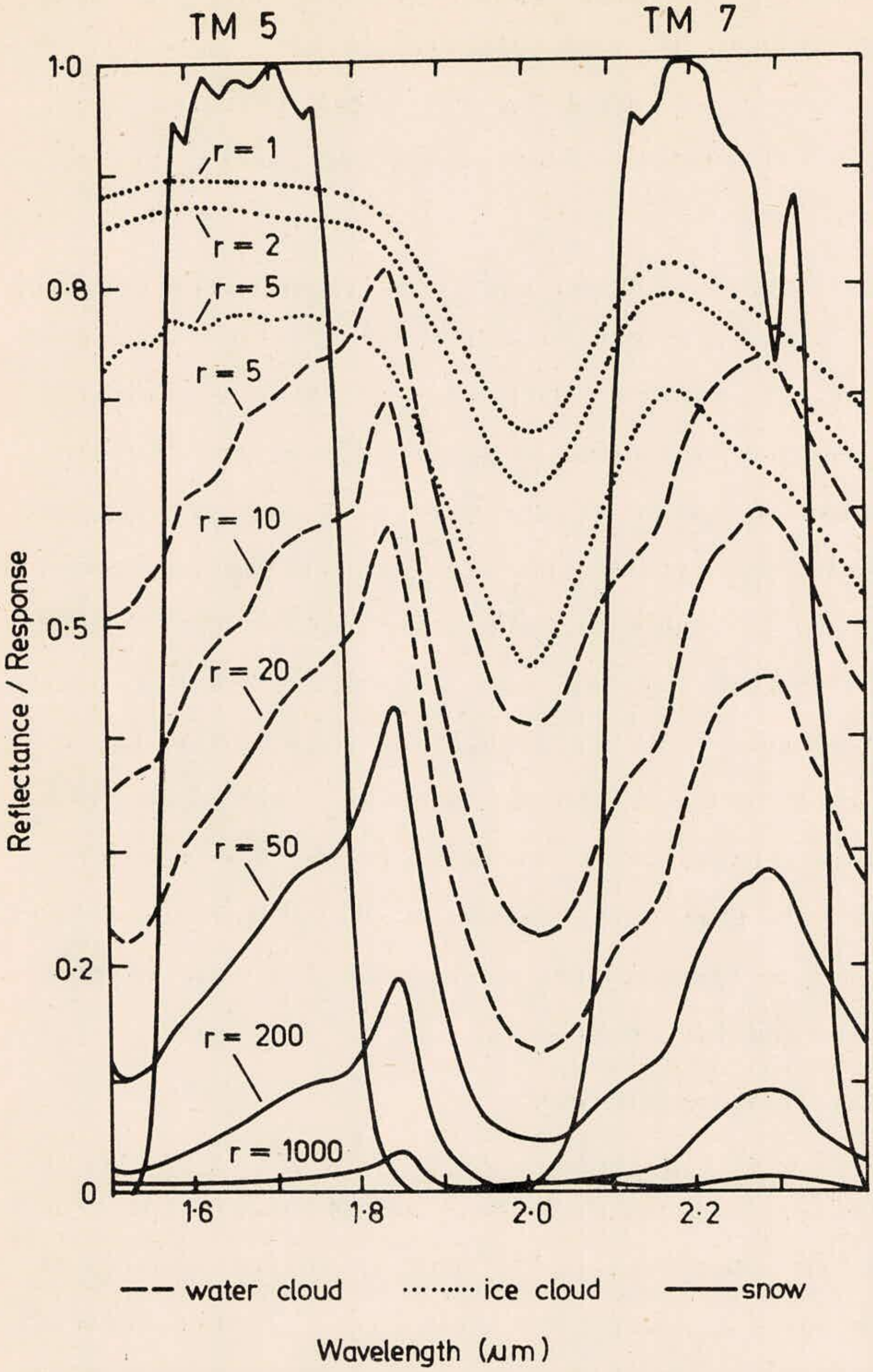

FIG. 5-SENSOR RESPONSE FUNCTIONS AND SNOW REFLACTANCES AT ILLUMINATION ANGLE OF 60 AND GRAIN RADII FROM 1 TO 1000 µm FOR THE THEMATIC MAPPER 5 AND 7 LANDSAT 4.

infrared and middle infrared wavelength regions and have a spatial resolution of each picture element (pixel) of approximately 30 m. TM band 6, a thermal infrared band, is sensitive to infrared surface temperature and has a resolution of 120 m.

TM band 4 (0.76-0.90μm) was found to show the greatest variability of the 6 reflective bands, in spectral response in the glacierized areas. Much of the TM 4 variability in spectral response is caused by snow grain size difference in the accumulation araea of the glaciers, and melting or refrozen, previously melted snow. The spectral response pattern of TM (0.52-0.60 μ m) generally follows that of TM 4 but detector saturation is more common in band 2 over snow covered areas. TM 5(1.55-1.75 μm) is quite useful for distinguishing between clouds and snow, and also shows suitable surface reflectivity differences on the glaciers. TM 6(10.4-12.5 μm), the thermal band, is useful for measuring radiometric surface temperature and detecting high cirrus clouds over snow and ice (Hall et al 1987).

4.4 Thermal Infrared Studies

One of the significant advantages of working in the thermal infrared portion of the electromagnetic spectrum for snow and ice observations is the increased chance of observation because of night-time overpasses. Though thermal infrared imaging systems are not capable of imaging through cloud-cover, their use during cloudy periods, owing to the addition of night time pass, essentially doubles the possible number of observations. This increases the chances of obtaining an observation during a cloud-free period observations of the snow-pack in the thermal band can be used to delineate the area covered by snow because of the temperature contrast with the snow-free areas, especially in the daytime during spring snowmelt. In addition, the temperature of the snowpack can never rise above 0^{o} C so that observation of the surface temperature can be used, first, to identify when the snowpack surface reaches 0^{o} C during the day and could possibly be melting and, second, to observe when the 0^{o} C temperature is maintained diurnally indicating a possible ripe snowpack.

When consideration is given to the use of satellite thermal infrared data for snowpack monitoring it is found that the spatial resolution is generally poorer than that of the visible and near-infrared, thus resulting in a less detailed monitoring capability.

4.5 Microwave Studies

Measurement of some properties of a snowpack without digging a pit are possible with small microwave radar system, which can either be buried in the ground looking upward into the snowcover or towed on Skis looking downward into the snow (Gubler and Hiller, 1984). Geometrical layering, density, water equivalence, total height, and liquid water content can be estimated for flowing avalanches, an upward looking system can estimate the height of dense flow.

Microwave measurements have the capability to penetrate the snow and respond to variations in subsurface properties. Working in the Microwave region also permits remote observation of snow under nearly all weather and lighting conditions. Often the areas that are most dynamic are also the most cloudy, such as the boundaries of sea ice packs.

The equivalent temperature of the microwave radiation thermally emitted by an object is called the brightness temperature, T_B. It is expressed in units of temperature (Kelvin) because for wavelengths in the Microwave range, the radiation emitted from a perfect emitter is propertional to its physical temperature T. However, most real objects emit only a fraction of the radiation, while a perfect emitter would emit at its physical temperature. This fraction defines the emissivity e of the object. In the microwave region, $e = T_{B/T}$, which is a basic equation of passive microwave radiometry.

As an electromagnetic wave emitted from the underlying earth surface propagates through the snowpack, it is scattered by the randomly spaced snow particles into all directions. Consequently, when the wave emerges at the snow/air interface, its amplitude has been attenuated. The dry snow absorbs very little energy from the wave, and therefore it also contributes very little in the form of self emission. When the snowpack grows deeper, the wave suffers more scattering loss, and the emission from the snowpack is further reduced.

Microwave radometry is useful as a remote sensing tool because the emissivity of an object depends very much on its composition and physical structure. Thus determination of emissivity provides information on the physical properties and conditions of the emitting medium. The emissivity is determined bymeasuring the brightness temperature radiometrically and by measuring the physical temperature in some manner (Foster et.al., 1984). The microwave radiometer is an ultrasensitive electromagnetic receiver which responds to thermal radiation at microwave frequencies.

The intensity of the microwave radiation that is emitted from a snowpack depends on the physical temperature, the grain size, the density, and the underlying surface condition of the snowpack. By knowing these parameters, the radiation that emerges from a snowpack can be derived by solving the radiative transfer equation, which usually serves as the starting point for electromagnetic modelling of snow.

The snow grains scatter the electromagnetic radiation incoherently and are assumed to be spherical and randomly spaced within the snowpack. Although snow particles are generally not spherical in shape, their optical properties (both the scattering and the extinction effects) can be simulated as spheres by utilizing Mie theory. The scattering effect is more pronounced at the shorter wavelengths and for larger particle sizes and drier snow. If a monodispherse distribution of snow particles and a snow density that is representative of moderately packed snow are assumed, the extinction coefficient is larger at longer wavelengths, and there is a drastic increase in extinction values for wet snow in comparision to dry snow conditions. By using the radiative transfer equation, the emerging brightness temperatures can be calculated with different physical parameters.

The large contrast between dielectric properties of water and those of most solids is the factor that makes the use of microwave radiometric techniques important for problems related to water resources. The dielectric properties of a material are characterized by its dielectric constant ε, which is a measure of the response of the material to an applied electric field, such as an electromagnetic wave. This response can be split into two factors: the first determines the propagation characteristics of the wave into material, i.e., velocity and wavelength, and the second is a measure of the energy losses in the media. The two factors are represented by the real (ε') and imaginary (ε'') parts of a complex dielectric constant. Thus $\varepsilon = \varepsilon' - j\varepsilon''$, where $j = (-1)^{\frac{1}{2}}$. The ratio $\varepsilon''/\varepsilon'$ is called the loss tangent for the material. In general, the values of ε' and ε'' will be functions of temperature and frequency. The dielectric constants of water ($\approx$ 1.0), ice ($\approx$ 3.2) and snow ($\approx$ 2.0) are different enough so that even a little melting causes a strong microwave response. In addition, the dielectric constant for snow is usually lower than that of dry soil, and since scattering further reduces the brightness temperature, there is sufficient contrast in the brightness temperature range for snow field monitoring.

Because of the complexities in the microwave, significantly more ground information is needed for microwave snow studies than for comparable visible, near-infrared, and thermal infrared studies.

The nimbus series of spacecraft has provided passive microwave snow cover observations since the early 1970s. From a global change prespective, the most attractive passive microwave data set is that provided by the Nimbus 7 Scanning Multichannel Microwave Radiometer (SMMR). This data set extends from 1979 to the present and includes the frequencies provided by the sensors on the Nimbus 5 and 6 satellites (19.35 and 37 GHz, respectively), in order to lower-frequency observations. These passive microwave observations can be used to map snowcover extent on a continental or hemispheric basis. The resolution of the SMMR ($\sim$ 25 km² at 37 GHz) limits the precision of the snowline on regional maps, but on hemisphsic maps that are at a scale of $1/2^{o}$ latitude by $1/2^{o}$ longitude, the resolution does not adversely affect the determination of the snowline.

Currently, several algorithms are available to evaluate and retrieve snowcover and snow depth parameters for specific regions and specific seasonal conditions. A straight forward method to relate microwave radiometric data to snow cover and snow depth is to examine the difference between the brightness temperature observed for snowcovered ground and that for snow free gorund. The general form of a snowcover algorithm is:

$$\delta T_{SC} = F_{SC} - F_{SC=0}$$

where,

δT = change in brightness temperature

SC = snow covered terrain

F_{SC} = observed radiometric value for snowcovered terrain

$F_{SC=0}$ = observed radiometric value for snowfree terrain

where, F may be either the brightness temperature at a single frequency or a more complicated expression involving the brightness temperature at several frequencies or polarizations.

Chang et al. have developed an algorithm that assumes a snow density of 0.30 and a snow grain size of 0.35 mm for the entire snowpack. The difference between the SMMR 37-GHz and 18-GHz channels is used to derive a snow depth brightness temperature relationship for a uniform snowfield.

This is expressed as follows:

$$SD = 1.59 * (T_{B18H} - T_{B37H})$$

where SD is snow depth in centimeters, H is horizontal polarization, and 1.59 is a constant derived by using the linear portion of the 37 GHz and 18 GHz responses to obtain a linear fit of the difference between the 18-GHz and 37 - GHz frequencies. If the 18 - GHz T_B is less than the 37 - GHz T_B, the snow depth is zero, and no snow cover is assumed.

An evaluation of the various algorithms that have been derived shows that only algorithms including the 37-

GHz channel provide adequate agreement with the manually measured snow depth and snow water equivalent values. It may also be noted the (T_{18H} - T_{37H}) often gives better results than the 37-GHz channel alone. Use of the 18-GHz channel helps to partly eliminate the effects of the snow and ground temperature and the atmospheric quantities (integrated water vapour and clouds) on changes in T_B.

The algorithms that have been developed to model microwave emission can be inverted so that the snowcover and snowdepth may be calculated from the microwave T_B. This appears to be possible in areas for which the relevant properties of the snowpack are well established.

At present, radiometers at 0.8 - cm wavelength (37-GHz frequency) are the most widely used sensors for snowpack monitoring, but during the 1983 Marginal Ice Zone Experiment (MIZEX) over Alaska, a 92-GHz sensor was employed aboard a NASA CV-990 aircraft. It was found that the 92-GHz data are even more sensitive to snow crystal scattering then the 37-GHz data but that they are also more sensitive to atmospheric constituents (Figure 6). Chang et al. have shown that the microwave brightness temperature patterns in the Alaskan study area were found to be similar to the T_B patterns from the 37-GHz (0.8 cm) data.

In the microwave region, the spatial resolution is basically limited by the physical size of the antenna. Thus for the same size of antenna, radiometers operating at shorter wavelengths (eg., 0.3 cm) will provide a better spatial resolution might outweight this disadvantage for

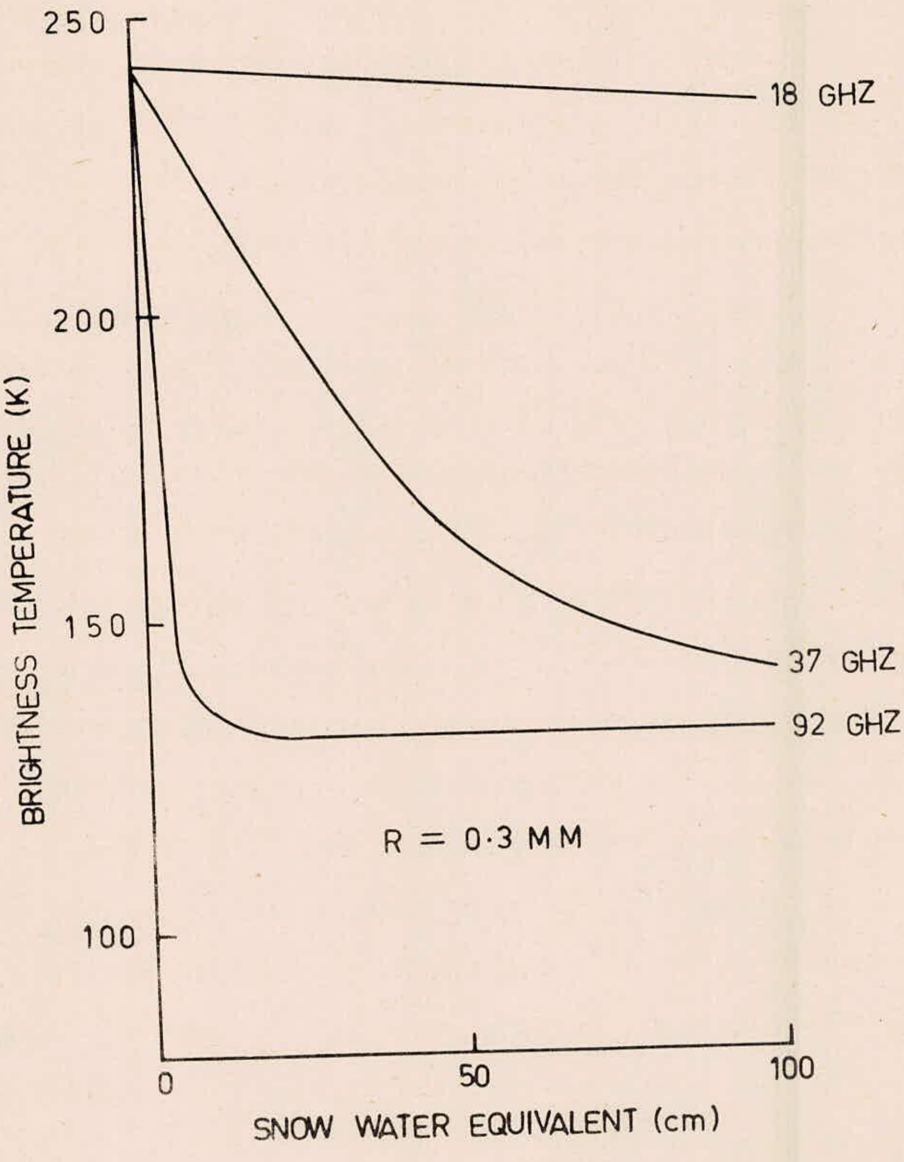

FIG. 6-SNOW WATER EQUIVALENT VERSUS MICROWAVE BRIGHTNESS TEMPERATURE FOR 18.37 AND 92 GHZ FREQUENCIES.

snowpack monitoring. Additionally, at 0.3-cm wavelength, scattering of the radiation by snow crystals will be stronger than that at 0.8 cm, and thus a stronger signal will be present.

The potential for monitoring snow water equivalent using passive microwave sensors on baord trucks, aircraft and satellites has been demonstrated by many authors. Results and findings of numerous studies and reports are summarised below:

1. At microwave frequencies the dielecatric constant of ice and water differs profoundly, rendering the signatures of dry and wet snow stricking differently (Sweeney and Colbeck, 1974; Ambach and Denoth, 1975).

2. Because of the dielectric differences between ice and water, microwave attenuation increases as the liquid water content of the snow increases (Stiles and Ulaby, 1981; Linor, 1980).

3. Snow depth and/or water equivalent is correlated with the observed brightness temperature. Brightness temperature tends to decrease as snow depth increases (Edgerton et al 1971).

4. Microwave penetration depth in snow decreases with increasing frequency. Shorter wavelengths (0.8 and 1.7 cm; 37 and 18 GHz) seem more responsive to snow depth than longer wavelengths (NASA, 1982).

5. Radiation from dry snow is strongly influenced by crystal or grain size because the radiation emitted by the snowpack is scattered on its way to the surface by snow grains. Since most of the radiation emitted at short wavelengths comes from depths of the order of

meters (Zwally and Gloersen, 1977), the depth of snow (and thus the number of potential scatterers) also affects the brightness temperature (Rango et al, 1979).

6. Short-wavelength radiation is scattered by snow crystals and grains (1 mm in size) which are comparable in size to the wavelength. Longer wavelength radiation (>~ 10 cm) is affected by verylarge crystals, lenses, and layers within the snow (Shiue et al, 1978).

7. Larger grain sizes cause more scattering than smaller grains, resulting in a lower and therefore a lower brightness temperature (stiles and Ulaby, 1980).

8. A beam incidence angle of 50 or more results in a lower brightness temperature and a higher correlation between snow depth and brightness temperature than does a scan angle of 0 or 30 . The greater incidence angle allows the radiometer to sense more of the snow-pack and less of the underlying soil (Chang and Shine, 1980).

9. Knowledge of the condition of the ground underlying the snow is important for the interpretation of observed brightness temperatures and can often be determined from observations using long microwave wavelengths (Hall et al., 1978). The state (frozen or thawed) of the underlying soil can sometimes be deduced from the polarisation ratio derived from micarowave at 10.7 GHz, which are long enough to penetrate through the pack (Chang et al 1982). Dry or frozen ground has

a high emissivity (0.90-0.95), whereas wet ground has a much lower emissivity (0.70) with correspondingly lower brightness temperature. The presence of vegetation causes an increase in emissivity, especially at shorter wavelength.

10. Normally, for spring snow there occurs a warmming each day which results in melting and formation of liquid water suspended between the snow crystals, while in the evenings, freezing restores the snow cover to a typical morning state. The repeated freezing and melting results in larger grain sizes and even lower brightness temperature. Volume scattering by the ice crystals at all wavelengths caused a more black body like behaviour at the lower frequencies with their much deeper penetration, while the higher frequencies are scattered more efficiently by each crystal, thus reflecting more of the cold sky (Schanda and Hofer, 1977).

11. In areas where the snowpack undergoes considerable freezing and thawing, it appears that the horizontal polarization is better studied than the vertical polarization for detecting variations in snow depth. The ice lenses, layers and surface crust are quite transparent to the horizontally polarization (D.K.Hall et al.).

12. Vegetation over a snowpack increases the microwave brightness temperature of snow: the denser the cover,

the greater the increase. The emissivity of the vegetation tends to overwhelm the scattering effect of the underlying snowcover. Even considering the different types of surface covers, remote sensing of the water equivalent of dry snow still appears possible (Tiuri and Hallikainen, 1981).

13. Generally, at the 21 - GHz frequency relationship between the snow depth or snow water equivalent and brightness temperature is not as good as it is at the 37-GHz or 18- GHz frequences. This is due to the proximity of the 21-GHz frequency to an absorption line for water vapour (at 22-GHz) for which the transmission is less than about 80% (Ulaby et al., 1981). Therefore the amount of radiation received by a radiometer at 21-GHz from the snowpack will be attenuated by the presence of moisture in the atmosphere.

14. In an area that produces snow grains of similar size from year to year because of similar winter climatic conditions, the microwave emissivity should provide an indication of the relative snow accumulation in that location (Rango et al., 1979).

15. Because different geographic areas are likely to have different snowpack conditions, ground cover, underlying soil conditions, and surface temperatures, it is difficult to extrapolate empirical relationship between microwave brightness temperature and snow depth from one area to another. But specific relation-

ship can be derived for individual areas and should be useful in improving assessment of snow pack conditions (Rango et al. 1979).

16. Frequent monitoring of dry snow pack during the accumulation period until the onset of snowmelt is useful in determining the snow water equivalent. The water equivalent determined in this way should be close to the maximum seasonal value (Rango, 1983).

4.6 Major Sensor Bands

The major sensor bands for making snow observations from space are listed in Table 2 along with important snowpack properties that influence the electromagnetic signature. The sensor responses relative to snowpack properties in Table 2 indicate that the microwave region seems to be invested with a wealth of information about snow followed by the visible/near infrared and thermal regions.

Table 2: Sensor Band Response Relative to Various Snow Pack properties (Rango, 1983)

Property	Sensor band: visible/near infrared	Thermal infrared	Microwaves
Snow-covered area	Yes	Yes	Yes
Depth	If very shallow	weak	moderate
Snow water equivalent	If very shallow	weak	moderate
Stratigraphy	No	weak	strong
Albedo	Strong	No	No
Liquid water content	weak	weak	strong
Temperature	No	Strong	weak
Snow/Soil boundary	No	No	weak (high frequency) to strong (low frequency)
All weather capability	NO	No	Yes
Current best spatial resolution from space platform	10's of meters	100's of meters	Passive: 30 Km (high frequency) to 150 km (low frequency) Active: 10's of mts.

4.7 SNOW PARAMETERS AFFECTING SENSOR RESPONSES

The snow parameters affecting the sensor responses are given in Table 3.

Table 3

A. Gamma Ray Sensor

(i) Snow depth

(ii) Background radiation

B. Visible/Near-Infrared (Albedo) Sensors

(i) Crystal size

(ii) Contaminants

(iii) Snow depth (only for shallow snow, i.e., up to a few centimeters)

(iv) Liquid water

(v) Surface roughness

C. Thermal Infraraed Sensors

(i) Temperature

(ii) Crystal size

(iii) Liquid water

D. Microwave Sensors

(i) Liquid water

(ii) Crystal size

(iii) Water equivalent depth

(iv) Stratification

(v) Snow surface roughness

(vi) Density

(vii) Temperature

(viii) Soil condition

5.0 SATELLITE SYSTEMS

5.1 Existing and Future Satellite Systems for Hydrological Applications

Many parameters of hydrological interest can currently be monitored and studied using satellite remote sensing platform. Polar-orbiting and geostationary satellites data are used to monitor parameters including rainfall, snow cover, ice jam, floods, water quality, etc. Currently, Indian Space Research Organisation (ISRO), India operates the polar orbiting satellite system (IRS series), the National Environmental Satellite, Data and Information Service (NESDIS) operates two polar-orbiting satellite systems (Landsat and NOAA series). Frensh space agency operates the polar orbiting satellite system System Probetoir d' Observation de la Terra (SPOT). Some specifications for these systems are listed in Table 4.

The polar-orbiting satellites are useful for hydrological applications because of their multispectral capability, and their global coverage.

Table 4

(Satellites, spectral ranges and resolution)

Satellite	Spectral Range (microns)	Resolution (m)
IRS-N	LISS-I and LISS-II*	
	0.45-0.52	72.5 (LISS-I)
	0.52-0.59	36.25(LISS-II)
	0.62-0.68	
	0.77-0.86	
Landsat-D	Multispectral Scanner(MSS)	
	0.50-0.60	80
	0.60-0.70	80
	0.70-0.80	80
	0.80-1.10	80
	Thematic Mapper(TM)	30
	0.45-0.52	30
	0.52-0.60	30
	0.63-0.69	30
	0.76-0.90	30
	1.55-1.75	30
	2.08-2.35	30
	10.40-12.50	120
NOAA Series	Advanced very high resolution radiometer (AVHRR)	
	0.58-0.68	1100
	0.725-1.10	1100
	3.55-3.93	1100
	10.30-11.30	1100
	11.50-12.50	1100
SPOT	Multispectral colour band	
	0.50-0.59	20
	0.61-0.68	20
	0.79-0.87	20
	Panchromatic	
	0.51-0.73	10

* Linear Imaging Self Scanning Sensor

5.2 Indian Remote Sensing Satellite

The primary objective of IRS is to provide for the systematiac repetivity acquisition of high resolution multi-spectral data of the earth's surface under nearly constant illumination conditions. IRS operates in near polar-orbit at an inclination of 99.02 degree at an altitude of approximately 904 km., descending mode. The satellite circles the earth every 103.2 minutes, completing 14 orbits per day. Entire earth is covered by 307 orbits during a 22 day cycle.

The IRS payloads (LISS-I and LISS-II cameras) obtain data over a swath along the ground track. The swath of LISS-I is 148.48 Km and that of LISS-II is 74.24 km at normal altitudes of 904 km.

It is basically the ability of the sensor to discriminate between two objects radiometrically. This depends upon the quantization (grey) levels generally available for orbiting satellites. In IRS 128 levels have been provided which will be sufficient for most of the applications.

The IRS has four spectral bands with spatial resolution of 36.25 m for LISS-II data and 72.5 m for LISS-I data. The capability of these bands in various appllications are as follows:

Table 5

SPECTRAL RATING AND APPLICATION OF IRS DATA

Band	Spectral Range	Application
1.	0.45-0.52	i) Coast environment studies (coastal morphology and sedimentation studies) ii) Soil/vegetation differentation iii) Coniferous/Deciduous vegetation discrimination
2.	0.52-0.59	i) Vegetation vigour ii) Rock/Soil discrimination iii) Turbidity and bathymetry in shallow waters.
3.	0.62-0.68	i) Strong chlorophyl sbsorption leads to discrimination of plant species.
4.	0.77-0.86	i) Delineation of water features. ii) Landform/geomorphic studies.

Optimum utilisation of water resources calls for an efficient management system for its inventory, utilisation and conservation both in time and place. Information is required on the volume and availability of water and snow. The characteristic factor in remote sensing which has played an important role for water resources management is the spectral band in the near infrared region where water can be discrimination from all other surface features and objects, due to its very low reflectance in this band.

The important sources of water are snow, ice and glaciers. Snowmelt contributes significantly to many of the Indian rivers originating in the Himalayas. Moreover this contribution occurs mainly in lean period and becomes very significant for its optimum use. The snow-melt run-off forecasting depends upon a number of factors such as the accumulation/depletion, the areal extent, thickness, temperature etc., of the snow, ice cover. The collection of such information by conventional methods is difficult because it requires continuous monitoring in adverse weather conditions and in difficult and hazardous terrain. The satellite data has been used for mapping snow cover, monitoring and temporal changes in snow cover (accumulation and depletion of surface extent) and glaciological studies.

6.0 REMARKS

After the snow is deposited, the particle shapes are modified by a process known as metamorphism which reduces the area of the snowflakes accompanied by an increase in the strength of bonds between grains. The introuction of water into the snow cover either in the form of rain or snow melt, causes a rapid metamorphism.

Metamorphism leads to increase the density, which in turn affect the reflectance characteristics of snow.

Figure 2 includes some typical characteristics reflection curves for snow.

It is evident that reflectance for snow will change especially when meltwater appears and snow crystals metamorphose to coarse grains and density increases. The short wave reflectance of low-density is high, but decreases once the snowcover begins to melt. Reflectance has been measured on the snow having increasing densities from 0.40 to 0.48 g/cm^3. The reflectance goes down with increase in density of snow from 0.40 to 0.48 g/cm^3. But the pattern of reflectance does not behave in a very systematic manner in band range 1 to 6, while in band range 6 to 7, reflectance shows systematic behaviour of sharp fall.

So, Remote sensing can be an efficient tool for the study of metamorphism in snow.

In an experimental phase extensive ground truth measurements and field observations should be carried out in addition to remote sensing studies for a better understanding of the characteristics of importance of the various snow parameters and its influence on the spectral signatures.

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