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COUPLED FLUX OF HEAT AND MOISTURE ACROSS A SNOWCOVERED LANDSURFACE WITH VEGETATION

BY
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and
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RALPH M. PARSONS LABORATORY
HYDROLOGY AND WATER RESOURCE SYSTEMS

Report Number 271

Part of the research discussed herein was sponsored by the National Science Foundation under Grant Number ATM-8114723, "A Dynamic Landsurface Boundary Condition for Climate Models".

APRIL 1982

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ABSTRACT

A theoretical study of snow accumulation and ablation in vegetated areas is presented. A one-dimensional energy and mass balance mdoel of winter (non-transpiring) vegetation is coupled to an existing one-dimensional energy and mass balance model of the surface snowcover. A sensitivity analysis of the simulated energy fluxes is conducted with respect to canopy density, interception characteristics, and atmospheric temperature.

Results indicate that the presence of vegetation can alter the heat and mass exchange between the snowcovered area and the atmosphere by at least three different processes. At lower canopy densitites, the increased surface area available can increase the magnitude of the energy and mass fluxes. At higher canopy densities, the increased attenuation of ventilating wind tends to limit increases in turbulent transfer. At high canopy densities, the total shielding by the canopy of shortwave radiation and sensible and latent heat transfer to the surface from the atmosphere tends to cool the surface in warm or moderately cold weather. This causes an increase in temperature gradient and sensible heat flux from the canopy to the ventilating air.

The persistence of snow on the canopy has a dramatic effect on the energy balance of the area. A snow-less canopy can become very warm and transmit more heat than a snowcovered canopy.

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A_s	albedo of surface snowcover	(dimensionless)
a	wind attenuation coefficient	(dimensionless)
C	cloud cover in tenths	(dimensionless)
C_H	bulk heat transfer coefficient	(dimensionless)
C_{Hg}	bulk heat transfer coefficient applicable to snow surface	(dimensionless)
C_M	bulk momentum transfer coefficient	(dimensionless)
C_W	bulk water mass transfer coefficient	(dimensionless)
CW_1	maximum allowable lag time	(hr)
CW_2	lag time parameter	(cm^{-1})
CW_3	recession parameter	(hr)
CW_4	attenuation parameter	($cm^3 gm^{-1}$)
	water vapor concentration	(g/cm^3)
c_f	bulk transfer coefficient at foliage	(dimensionless)
C_p	specific heat	(cal/g°K)
D_e	diffusion coefficient	(cm^2/sec)
d_g	measurement depth in soil	(cm)
d_s	depth of snow layer	(cm)
d_s	grain size	(cm)
E	vapor flux at surface	($gm\ cm^{-2} sec^{-1}$)
E_f	vapor flux at vegetation	($gm\ cm^{-2} sec^{-1}$)
e_a	vapor pressure of air	(mb)
e_i	saturated vapor pressure over ice	(mb)
G_c	heat conducted from ground	(cal/cm^2)
g	acceleration of gravity	(m/sec^2)
H	sensible heat flux	(cal/cm^2)

LIST OF SYMBOLS

H	height of canopy (meters)
H_c	heat flux by conduction (cal/cm^2)
H_{cc}	cold content of snow (cal/cm^2)
H_e	evaporative heat flux (cal/cm^2)
H_{sf}	heat flux from vegetation per unit area (cal/cm^2)
I	insolation (2.3.1) (cal/cm^2)
I	depth of intercepted snow, water equivalent (cm)
K_M	coefficient of turbulent flux of momentum ($cm^2 sec^{-1}$)
K_N	coefficient of turbulent flux of heat ($cm^2 sec^{-1}$)
K_w	coefficient of turbulent flux of mass ($cm^2 sec^{-1}$)
k_e	effective thermal conductivity of snow (cal/cm.sec.°K)
k_g	thermal conductivity of soil (cal/cm.sec.°K)
k_i	extinction coefficient for snow (cm^{-1})
k_s	thermal conductivity of snow (cal/cm.sec.°K)
k_s	intrinsic permeability of snow (2.3.4) (mm^2)
k_o	extinction coefficient for clear ice (2.3.1)(cm^{-1})
k_o	Von Karman's constant (dimensionless)
L	Monin-Obukhov length (cm)
L_f	latent heat of fusion (cal/g)
L_M	latent heat of melting (cal/g)
L_s	latent heat of sublimation (cal/g)
L_w	lag time for outflow (hr)
M	empirical melt factor (2.1) (dimensionless)
M	canopy density (dimensionless)
m	net condensation ($g/cm^3 sec$)

LIST OF SYMBOLS

N	leaf area index (dimensionless)
O	rate of outflow (cm/hr)
O_+	rate of outflow (cm/hr)
P	heat advected by rainwater (cal/cm^2)
P_a	atmospheric pressure (mb)
P_f	heat advected by rainwater to foliage (cal/cm^2)
P_x	heat transferred by rainwater to surface cover (cal/cm^2)
Q_f	heat storage of intercepted snow (cal/cm^2)
Q_h	change in heat storage of pack on surface (cal/cm^2)
q	specific humidity (dimensionless)
q_a	specific humidity of air (above vegetation) (dimensionless)
q_{af}	specific humidity of ventilating air (dimensionless)
q_f	specific humidity at foliage or intercepted snow surface (dimensionless)
q_s	specific humidity at surface of surface snowcover (dimensionless)
R_f	Richardson flux number (dimensionless)
$\mathbb{R}_i$	Richardson number (dimensionless)
$\mathbb{R}_{iB}$	bulk Richardson number (dimensionless)
R_{Li}	incoming longwave radiation (cal/cm^2)
R_{Lo}	outgoing longwave radiation (cal/cm^2)
R_s	absorbed shortwave radiation (cal/cm^2)
R_{ss}	insolation received at snow surface (cal/cm^2)
R_w	gas constant for ideal water vapor (cal/gm°K)
S	excess water in storage (cm)
T	temperature (°K)

LIST OF SYMBOLS

T_a	temperature of atmosphere (°K)
T_{af}	temperature of ventilating air (°K)
T_g	temperature of ground (°K)
T_g	temperature of surface snowcover (3.3) (°K)
T_p	temperature of precipitation (°K)
T_s	temperature of snow (°K)
t	time (sec)
u	wind speed (cm/sec)
$\bar{u}$	mean fluid velocity (cm/sec)
U_a	wind speed above vegetation (cm/sec)
U_{af}	ventilating windspeed (cm/sec)
U_*	friction velocity (cm/sec)
V	vapor flux (mm)
W_ℓ	amount of lagged excess water (cm)
Z	measurement height (cm)
Z_o	roughness height (cm)
α	Monin-Obukhov coefficient (dimensionless)
α_s	albedo of snow surface (dimensionless)
α_f	albedo of foliage or intercepted snow (dimensionless)
Γ	adiabatic lapse rate (°K cm^{-1})
ε_f	emmissivity of foliage (dimensionless)
ε_g	emissivity of ground or snow surface (dimensionless)
λ	thermal conductivity of snow (cal/sm.sec °K)
μ_w	viscosity of water (g/cm sec)
ρ	density of water (g/cm^3)

LIST OF SYMBOLS

ρ_a	density of air (g/cm^3)
ρ_i	density of ice (g/cm^3)
ρ_s	density of snow (g/cm^3)
ρ_w	density of water (g/cm^3)
σ	Stefan-Boltzmann constant ($cal/cm^2 {}^\circ K^4 sec$)
τ	shear stress ($dynes/cm^2$)
ϕ	Monin-Obukhov adjustment (dimensionless)

CHAPTER 1

INTRODUCTION

1.1 Introduction

The global-scale circulation of heat and moisture in the atmosphere has been an area of intensive study among hydrologists and meteorologists since the early nineteen fifties, when computer technology made such work feasible. This research is directed toward the development of reliable dynamic climate modeling at all scales on earth, and to a greater scientific understanding of atmospheric dynamics.

In these models, the atmosphere is represented by a multi-layer three-dimensional grid whose boundaries are the terrestrial surface and the top of the upper atmosphere (see Figure 1). A horizontal cross-section of the network is described by grid squares on the order of 10^4 to 10^5 km^2 in area (Figure 2). For each of these grid squares the variables and parameters describing the system are represented by average values. The influx of heat and moisture from the landsurface and of heat from the sun provide the boundary conditions necessary to predict the movement of heat and moisture between and among the atmospheric grids.

The atmospheric boundary conditions at the landsurface are also represented by average values for corresponding grid areas mapped on the earth's surface. Values of the moisture and heat exchange at the interface between the terrestrial and lower atmospheric grids must be calculated in response to the atmospheric excitation of the landsurface which we generally refer to as "the weather". In this manner, the landsurface/ocean and atmospheric models form a coupled system which is a numerical reflection of the hydrologic cycle.

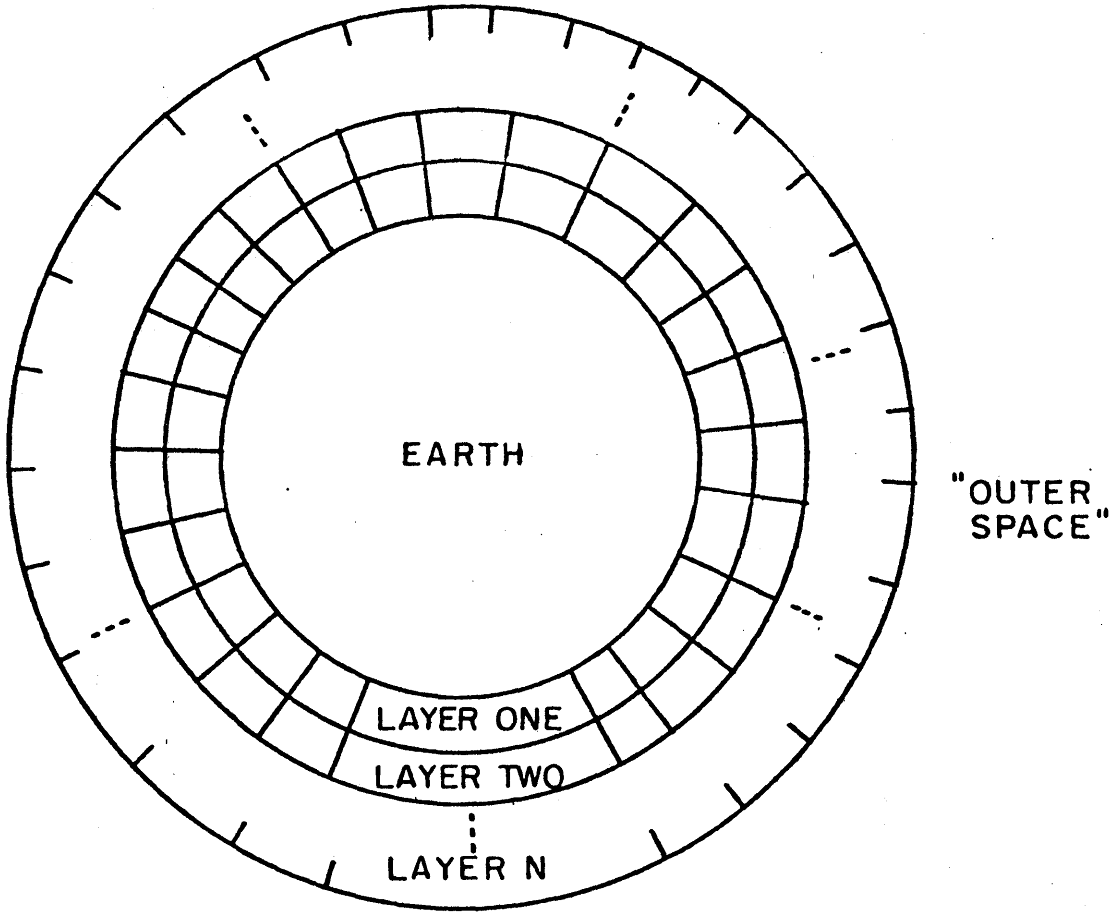


FIGURE 1

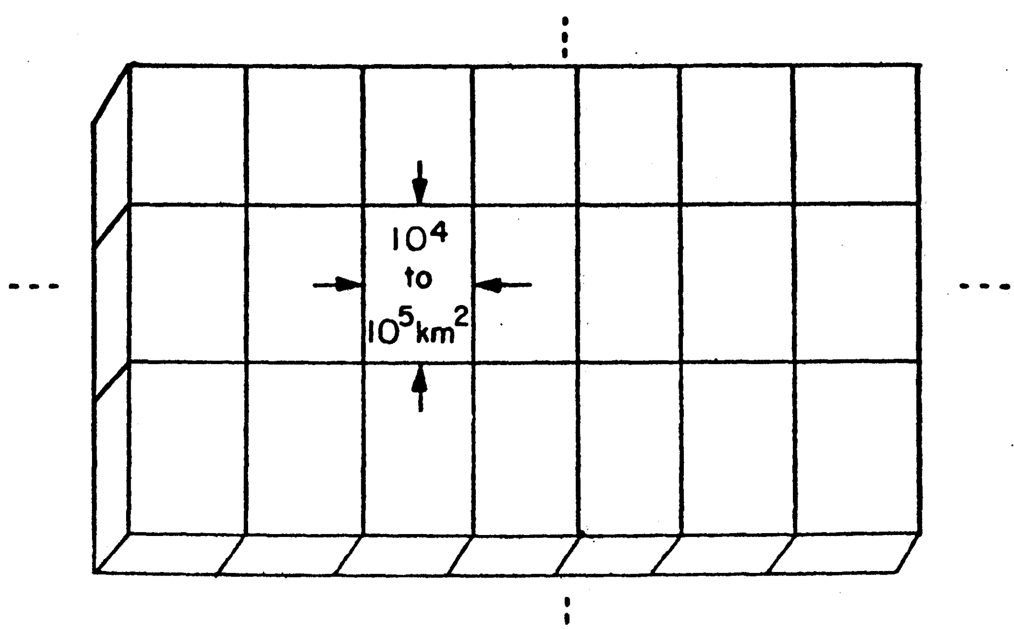


FIGURE 2

Such modeling is of interest to hydrologists for large-scale environmental impact assessment. It may have the capability of responding realistically to perturbations of the system at the landsurface. For instance, it may demonstrate what effect, if any, a massive swamp drainage project (such as is currently proposed in the Sudan) might have on the precipitation patterns of the region.

The work described herein is part of a larger task being conducted at MIT, entitled, "A Dynamic Landsurface Boundary Condition for Climate Modeling". This work has as its purpose the development and improvement of the landsurface boundary conditions, involving heat and moisture fluxes, which are necessary for the numerical modeling of global circulation. The ultimate goal of the study is to provide relatively simple but physically-based parameterizations of the exchanges by examining the relevant processes in one dimension (vertical).

So far, the work has produced a complex model of heat and moisture transport through unfrozen soil, and a separate, simpler model for heat and moisture transpiration from vegetation. The complexity of the soil model is designed to provide a basis with which to compare the performance of future simplifications of the model.

Prior to simplification, however, the complex models must be further developed to accomodate a wider range of climatic conditions. Specifically, these models must have the capability of simulating surface snowcover effects, canopy interception of precipitation, and the hydraulics of frozen soil. Here, the first two of these three processes are modeled individually and are subsequently linked to form a comprehensive model which describes "point" behavior under a variety of climatic conditions.

1.2 Review of Previous Work

The coupled movement of heat and moisture through an isothermal porous medium was studied analytically by J. R. Philip and D. A. DeVries in the late nineteen-fifties (Philip (1957), DeVries (1958), Philip and DeVries (1957)). Their works analyzed moisture transport and heat conduction in soils, accounting for vapor flow and heat storage. Hysteresis in the matric potential-moisture content relationship was omitted from their models, however. Y. Mualem (1973, 1974, 1976) has produced the most notable investigations into the hysteretic phenomenon, wherein the wetting history of the soil becomes a parameter of the matric head in addition to moisture content and soil properties. P. C. D. Milly (1980) incorporated these effects in a finite element model of heat and moisture transport in a soil column, and applied the model to a variety of initial and boundary conditions (e.g., infiltration into very dry soil, saturated soil, unsaturated soil with ponding, etc.) and to a variety of soil types.

The effect of near-freezing temperatures on moisture and heat transfer through soil has been described by Harlan (1973). Harlan draws an analogy between the mechanisms of flow through partially frozen soil and the mechanisms for unsaturated flow. He develops a finite difference scheme which includes freezing and thawing effects on hydraulic conductivity. Most subsequent work in this area follows Harlan's approach.

Guymon and Luthin (1974) present a finite element solution to Harlan's formulation. They also suggest a method for determining the relative distributions of ice and liquid in a soil. Jame and Norum (1980) make an empirical adjustment to the soil water diffusivity as described in Harlan' s article which accounts for the impeding effects of ice lenses on flow paths in the soil. Guymon and others incorporated frost heave effects in later formulations of the problem.

The elemental sourcebook for the vast majority of snowmelt studies is the U. S. Army Corps of Engineers 1956 study, Snow Hydrology. This comprehensive volume embodies a consolidation of all previous major work, plus a wealth of information from field studies.

Kuz'min's (1961) Melting of Snow Cover may be said to be the Russian counterpart to Snow Hydrology, though it relies more heavily on mathematical formulations than on actual field measurements. Melting of Snow Cover includes important work in the area of turbulent heat exchange at the snow surface.

Several articles by S. C. Colbeck (1972, 1973, 1978) study the problem of liquid transmission through snow in great detail. M. R. DeQuervain (1973) further conceptualizes the process of snow metamorphism both before and during the melt season in an attempt to stimulate further studies on this topic.

Eric A. Anderson (1976), in A Point Energy and Mass Balance Model of a Snowcover, compiles a computationally efficient, physically based, point snowmelt model which is a natural extension of his earlier (1964) computer model, which utilized air temperature as the sole snowmelt index. The analysis and testing of Anderson's (1976) model and the research behind it comprise the bulk of the investigations in this report.

Anderson's model does not include provision for a layer of vegetation which affects the accumulation and ablation characteristics of snow, as well as the time dependence of surface albedo. In this work the forest temperature and wind profile studies of Deardorff (1978) and Cionco (1965, 1972) are combined with the studies of snow on and beneath forest canopies by Hendrie and Price (1978), Snow Hydrology (1956), D. H. Miller (1959), A. A. Molchanov (1956), and others, to determine energy and mass fluxes within the atmospheric boundary layer.

The work which follows describes the following separate models of mass and energy fluxes:

1. snow cover on land surface
2. snow cover on vegetal canopy

CHAPTER 2

DYNAMICS OF SURFACE SNOWCOVER

2.1 Theoretical Background

Much of the following description of the theory of snowmelt is condensed from the works of Anderson (1976), Kuz'min (1956), and Snow Hydrology (1956).

The seasonal snowcover has received a great deal of attention from hydrologists because it has a major climatic impact in much of the Northern hemisphere. Twenty-three percent of the globe, including fifty percent of the land, is permanently or temporarily snowcovered during the year. The energy and mass exchange between the forty-two million square miles of seasonal snowcover cause an impact on the local balances of both water and heat. The appearance and disappearance of snow can thus provide an important input to the landsurface boundary conditions for models of global atmospheric circulation.

The research described below aims at a physically-based parameterization of the snow accumulation and "ablation" (decay of pack through melt or sublimation) processes in one dimension, to be used in conjunction with a one-dimensional model of energy and moisture fluxes in the soil.

The one-dimensional approach neglects several important aspects of the problem such as blowing of snow into uneven drifts, slope effects on liquid movement through the pack, discontinuous (patchy) snowcover, water movement around ice layers, etc. Still, a vertical model can provide a wealth of information about the mechanics of the sequence and is fairly accurate in terms of areal averages.

Present approaches to one-dimensional modeling of snowmelt can be placed into seven categories in order of improved accuracy:

1. Air temperature T_a only is used as an index to snowmelt:

$$\text{Melt} = M \cdot (T_a - 32°F) \tag{1}$$

where M = a constant empirical melt factor.

2. The melt factor is permitted to increase with the date of the snow melt season.

3. Presence of liquid water in the pack is included.

4. Refreezing of the pack when the temperature is less than 0° C is accounted for.

5. An allowance for the fact that 0° C is not an absolute boundary between snow and rain forms of precipitation is made.

6. The amount of melt is allowed to depend on temperature and radiation fluxes.

7. All energy and mass fluxes are balanced:

$$\Sigma_i - \Sigma_o = \Sigma_s \tag{2}$$

where

Σ = the considered quantity (heat or mass) and subscripts

i = incoming

o = outgoing

s = stored

The last approach is ideally the most accurate and the most complex to evaluate. The energy and mass balance model may be further subdivided into the "bulk snowpack" and the "layered snowpack" approaches. In a bulk model, fluxes into and out of the snowpack as a whole are considered. The bulk snow model will store heat until the whole pack has reached 0° C, at which time melt may occur. A layered model allows heat and mass transfer between adjacent horizontal layers of the pack in addition to the fluxes to and from outside the pack. In the layered model, melt and runoff may occur in a given layer even when the entire pack has not reached 0° C. This latter approach facilitates an accurate evaluation of density changes by mass redistribution within the pack. Such density changes are important in determining the heat and mass fluxes themselves, as will be shown later.

The heat exchange between a bulk snowpack and its environment may be written:

$$Q_h = R_s + R_{1_i} - R_{1_o} + H_c + H_e + G_c + P \qquad (3)$$

where

Q_h = change in heat storage of the pack

R_s = absorbed shortwave radiation

R_{1_i} = incoming longwave radiation

R_{1_o} = outgoing longwave radiation

H_c = convected sensible heat transfer from atmosphere to snow

H_e = latest heats released by condensation or sublimation

G_c = heat conducted from ground to snowpack

P = heat added by rain water

2.2 External Heat Transfers to Snowpack

2.2.1 Radiation

Over the full melt season, the most influential factors in determining melt are the radiational components of Equation (3). The "shortwave" radiation reaching the ground is in the wavelength range 0.3 μm to 2.2 μm, while the "longwave" is defined as being between 6.8 μm and 100 μm. Other wavelengths comprise a relatively small amount of the total insolation.

The significant sources of shortwave radiation for a snowcover are the direct beam from the sun and diffuse reflections from particles in the atmosphere. The amount of shortwave radiation which is absorbed by the snowpack, R_s, is determined by the snowpack albedo, α_s:

$$R_s = (1 - \alpha_s) R_{ss} \qquad (4)$$

where

R_{ss} = insolation received at snowsurface

= direct beam plus diffuse radiation

There is an interesting feedback between net insolation and surface albedo. It has been found that the albedo of snow is markedly affected by the relative proportions of diffuse to direct radiation (Male and Granger (1981)). This may be due to the wavelength dependence of the albedo. Albedo increases as diffuse radiation increases. Also, for surfaces of high albedo such as snow, the total insolation may increase because of diffuse scattering by the atmosphere of the reflected radiation from the snow pack. This phenomenon is called "multiple reflection" or "multiple scattering".

Bergen (1975) proposed a possible relation of albedo to the density and grain size of natural snowcover under the assumptions that all reflections from snow are diffuse and that the full solar spectrum is received at the surface. The former assumption is invalid for low solar angle, when specular reflection becomes more important, and the latter assumption is violated on cloudy days. Petzold (1977) suggested an adjustment to the clear sky albedo for cloudy days:

$$(\Delta\alpha_s)\% = 0.499 + 0.0097\ (C)^3 \qquad (5)$$

where

$(\Delta\alpha_s)\%$ = % change from clear sky albedo and

C = cloud cover in tenths, varying from one to ten

Equation (5) will provide only a rough estimate of this change, since different types of clouds have different spectral distributions of scattered radiation.

The amount of shortwave radiation that reaches the ground, R_{ss}, is highly dependent on the amounts of clouds, ozone, water vapor, and dust present in the atmosphere, since these will serve to absorb and/or scatter radiation. Solar angle also has a large effect. Dozier (1978) has developed a radiational model which accounts for each of those effects in addition to shielding by forest canopy and reflections from opposing slopes.

Both cloud cover and forest cover tend to block and absorb shortwave radiation but increase the amount of longwave radiation reaching the ground. The longwave emissivity of the snow remains unaffected by cloud and forest cover, however. The net effect of cloud and forest cover is to increase the all-wave radiation flux between atmosphere and snow at the high albedos associated with snow. The effects of forest cover on the underlying snow surface will be discussed in Section 2.2 using Deardorff's (1978) empirical method.

The attenuation of shortwave radiation by clouds has been examined by many researchers [e.g., Angstrom 1924, Davies et. al. 1975, Suckling and Hay 1977]. A good review of the research is this area performed to date may be found in Male and Granger (1981).

2.2.2 Turbulent Heat and Mass Transfer

Turbulence in the atmosphere produces the vertical transfer of sensible heat, water mass, and momentum. In calculating the energy balance of a snowpack, both the sensible heat transfer and the turbulent mass transfer are important. The turbulent transfer of water mass between a snowsurface and the atmosphere involves a phase change, and an accompanying gain or loss of latest heat. The phase change may be from liquid to vapor (or vice versa) over a melting snowcover, or from solid to vapor (or vice versa) when the snow is not melting. The basis for much of modern turbulent flux theory can be found in Prandtl's (1952) work, Essentials of Fluid Dynamics. Prandtl's formulation assumes that the fluid (in our case, the air overlying snow) is in a neutral condition, that is, temperature stratification in the surface boundary layer approaches the equilibrium gradient of ~ 6°/100 meters. The Prandtl equation (1952) may be written:

$$\frac{k_o z}{u_*} \frac{d\bar{u}}{\partial z} = 1 \tag{6}$$

where

k_o = Karman's constant

u_* = friction velocity = $\sqrt{\tau/\rho}$ = constant in layers above ground

τ = shear stress

$\bar{u}$ = mean fluid (wind) velocity

z = height

Integration of this equation leads to a logarithmic wind profile. This agrees with measurements of wind profiles made above a bare landsurface. In the case when snow is on the ground, the assumption of neutral stability is frequently erroneous. This is because the cooling from below causes thermal stratification, with the result being atmospheric inversion or lapse.

Examining first the case of neutral stability, the method of eddy diffusion provides the following equations (Priestley, 1959):

$$E = -\rho K_w \frac{\partial \bar{q}}{\partial z} \tag{7}$$

$$\tau = -\rho K_m \frac{\partial \bar{u}}{\partial z} \tag{8}$$

$$H = -\rho \ C_p \ K_n \ (\frac{\partial T}{\partial z} + \Gamma) \tag{9}$$

where

E = water vapor flux

q = specific humidity

H = sensible heat vertical flux

c_p = specific heat capacity

T = temperature

Γ = adiabatic lapse

K_w, K_m, K_n = coefficients of turbulent flux of water vapor, momentum and heat.

In Pradtl's formulation $K_m = K_w = K_n$ because the flux depends only on the vertical motion field (ignoring buoyancy). For the neutrally stable case, Kuz'min (1961) integrates Equation (6) to obtain:

$$E = \rho_a \ k_o^2 \ \frac{(u_2-u_1) \ (q_2-q_1)}{\ln(z_2/z_1)^2} \tag{10}$$

Temperature stratification is generally indexed by the Richardson number, $\mathbb{R}_i$:

$$\mathbb{R}_i = \frac{g}{T} \frac{(\partial T/\partial z) + \Gamma}{(\partial u/\partial z)^2} \tag{11}$$

where

g = acceleration of gravity

T = temperature

and $\mathbb{R}_i$ = 0 when air is adibatically nertral since $\frac{\partial T}{\partial z} = -\Gamma$

Monin and Obuknov (1954) proposed a simple adjustment to Prantl's logarithmic wind profile Equation (6) to account for non-neutral atmospheric conditions. The adjustment introduces the term $\phi(z/L)$:

$$\frac{k_o z}{u_*} \frac{du}{dz} = \phi(z/L) \tag{12}$$

where $\phi(z/L)$ is a function of altitude (z) and stability length (Prandtl's) L, defined by:

$$L = \rho_a \; C_p \; u_*^3 \; T/k_o gH \tag{13}$$

where

H = flux of sensible heat

C_p = specific heat capacity

It is assumed that $\phi(z/L)$ may be approximated by the relation:

$$(z/L) = 1 + \alpha z/L \tag{14}$$

where

α = determined by experiment

integration of Equation (12), making use of Equation (14), yields:

$$\bar{u} = \frac{u_*}{k_o} \left[\ln\ (z/z_o) + \frac{\alpha}{L}(z-z_o)\right] \tag{15}$$

where z and z_o are limits of integration and the lower limit z_o is the surface roughness length. Equation (15) involves τ (in the form of u_*), H (in the form of L) and α. Sensitive instrumentation has enabled various experimenters to measure τ and H directly and thus to estimate the behavior of α for various values of z/L. The instruments measure small fluctuations around the mean values of vertical and horizontal wind speed and air temperature, and accurately compute τ and H by the equations:

$$\tau = \rho \, \overline{u'w'} \tag{16}$$

$$H = \rho C_p \, \overline{T'w'} \tag{17}$$

where

u' = fluctuation around mean horizontal wind speed

w' = fluctuation around mean vertical wind speed

T' = fluctuation around mean temperature

bar = mean of barred quantity.

Anderson (1976) cites the results of several investigations found in the literature (Webb 1970, Dyer and Hicks 1970, Businger et. al. 1971) and settles on a value of $\alpha = 5$ for stable conditions, and in the case of instability:

$$\phi_M(z/L) = (1 - 16\ z/L)^{-1/4} \tag{18}$$

$$\phi_H(z/L) = \phi_w(z/L) = (1 - 16\ z/L)^{-1/2} \tag{19}$$

In terms of the Richardson number, these become:

for stable conditions

$$\phi_M = \phi_H = \phi_w = (1 - 5\,\mathbb{R}_i)^{-1} \tag{20}$$

for unstable conditions:

$$\phi_M = (1 - 16\,\mathbb{R}_i)^{-1/4} \tag{21}$$

$$\phi_H = \phi_w = (1 - 16\,\mathbb{R}_i)^{-1/2} \tag{22}$$

Calculation of the Richardson number, as given by Equation (11), for the constant flux boundary layer requires measurement of temperature and wind speed at two reference heights. Anderson (1976) describes a method of computation of turbulent transfer using measurements at one level. The stability index he uses is the "bulk" Richardson number ($\mathbb{R}i_B$) defined by:

$$\mathbb{R}i_B = \frac{2gz \frac{(T-T_s)}{}}{(T-T_s)u^2} \tag{23}$$

where

T_s = temperature at snow surface

Deardorff's (1968) study of "bulk transfer coefficients" or drag coefficients for momentum, heat, and moisture flux provides the values of the coefficients under neutral conditions:

$$(C_H)_N = (C_w)_N = (C_M)_N = \frac{k_o 2}{\ln(z/z_o)} \tag{24a}$$

and the ratios of the value of each coefficient under non-neutrally stable conditions, as is common over a snowpack, to its value under neutral conditions:

$$\frac{C_w}{(C_w)_N} = \frac{C_M}{(C_M)_N} = (1 - \alpha \mathbb{R}i_B)^2 \tag{24b}$$

where again $\alpha \cong 5$.

For the unstable case, see Anderson (1976) pp. 16-18, as the equations are too lengthy to be repeated here.

From the equations in this section, the flux of sensible heat may be computed. Another technique for determining turbulent heat transfer uses an empirical evaporation formula of the Dalton form:

$$V = (a + bu_a)\ (e_a - e_o) \tag{25}$$

where

e_a = vapor pressure at level a

u_a = wind speed at level a

a,b = regression coeffieients for site

e_o = vapor pressure at surface

V = evaporation expressed as a depth

The regression coefficients a and (especially) b vary widely depending on the site. Anderson (1976) compiles a table of determinations of a and b found in the literature. Coefficient a is close to zero, and b is of the order 10^{-3}. The latent heat transfer due to sublimation of snow may be expressed as:

H_e = Ls $\rho_w (a + bu_a)\ (e_a - e_o)$

Sensible heat flux may be directly related to latent heat flux by Bowen's ratio, which can be written (Anderson 1976):

$$\frac{H_c}{H_e} = \frac{P_a}{.622} \cdot \frac{C_p}{L_s} \left(\frac{T_a - T_s}{e_a - e_o}\right) \tag{26}$$

Here use has been made of the relation $q = \frac{.622}{P_a} e$

where

q = specific humidty

$$H_c = \frac{\rho_w C_p P_a}{.622} (a+bu_a)\ (T_a - T_o) \tag{27}$$

The turbulent flux theory outlined above is valid for cases in which a logarithmic wind profile exists. Above a bare landsurface where the roughness elements are small in relation to aerodynamic characteristics, the wind profile is logarithmic. However, in the presence of important roughness elements such as foliage, the profile is not logarithmic and Prandtl's equation and the Monin Obukhov ratios are not valid. This is discussed in Section 2.2 where use is made of Deardorff's (1978) research.

2.2.3 Heat Transfer by Rainwater

The heat content of rain is transferred to the snow as the rainwater is cooled to the temperature of the snow. For the bulk pack:

$$P = (T_p - T_s) C_p \, i \tag{28}$$

where

T_p = rain temperature (wet-bulb)

P = heat advected by rainwater (cal/cm^2-sec)

C_p = specific heat of water ($cal/g\,^oK$)

i = precipitation rate ($g/cm^2 sec$)

when rain falls on a subfreezing pack, the water will be frozen in the pack, releasing its latent heat of fusion to the snow. For each cm of rainfall that is refrozen, 80 langleys are given up to the pack in addition to the heat given up by the rain to bring the rainwater temperature to 0° C. Since the specific heat of snow is much smaller than that of water, this can be a large source of melt.

The heat transfer due to rain does not occur directly at the surface since rain can penetrate to some depth before refreezing. This will be discussed in greater detail in the next section.

2.2.2d Heat Transfer from the Ground

G_c, the heat conducted from the ground to the snow, is very small compared to the other instantaneous components of the heat balance, but becomes important over the snowmelt season due to the fact that it does not normally change direction. The soil is generally frozen before snow is accumulated, but due to the insulating properties of the snow and the conduction from warmer, deeper layers of the soil, the soil beneath the snow is usually quickly restored to a warmer temperature. The ground condition will be more important at night relative to other heat exchange processes (particularly insolation).

In Anderson's (1976) model, it is assumed that the snow-soil interface temperature changes slowly enough to be considered in a steady state at any time t. Thus a weighted form of Fourier's heat conduction equation is used:

$$G_c = \frac{2\,K_g \cdot K_s (T_g - T_s)}{k_g \cdot d_g + K_s \cdot d_s} \tag{29}$$

where

- k_g = thermal conductivity of ground
- k_s = thermal conductivity of snow
- T_g = temperature of ground
- T_s = temperature of snow
- d_g = depth below ground of T_g measurement
- d_s = depth of snow layer

2.3 Snowpack Internal Heat and Moisture Transfer

In Section 2.2 above the sources and sinks of heat and mass from outside the snowpack have been discussed. Such heat and mass exchanges are relatively straightforward. The following section examines heat and mass exchange within the layered snowcover which involves complex interrelationships between temperature, liquid water content, density, and grainsize.

In the "bulk" model, as discussed earlier, once the heat stored by the pack has been augmented enough to raise the pack to 0° C, additional heat will contribute to the latent heat necessary to produce melt. Any melt that occurs in the bulk model is either retained as liquid water in the pack, or becomes runoff. However, in the more realistic "layered" models, melt may occur at the surface while layers below are still subfreezing; thus melt may then percolate downward until it refreezes. This liquid flux causes a change in the density of the refrozen layer due to the redistribution of mass. The density in turn partly determines the depth of penetration of melt water.

As the season progresses, more liquid water is present as adsorbed or hygroscopic water in the pack, thus increasing the heat conductivity of the snow pack since liquid is a much better conductor of heat than the snow particles. This serves to create eventually an isothermal pack of zero degree temperature by the time the pack is nearly "ripe" (i.e., when the pack contains all the liquid water it can hold against gravity).

The 0° C pack or pack layer has a capability of storing liquid water up to its maximum storage capacity (Wmax). Any additional influx of heat or liquid will produce runoff from the pack.

Prior to an examination of the equations governing the transport of mass and heat energy within the pack, it may be helpful to describe the approach used in developing the equations. In the layered model, elements of depth, each with their own "cold content" H_{cc}, are considered in the premelt, or "dry" period. The influx of heat to an internal element by radiation penetration, conduction, and latent heat exchange, serves to raise that element's temperature, or, if enough heat is supplied, to melt the element. Meltwater at 0°C (or rainwater at $\geq$ 0°C) then travels to the next layer, giving up its heat by conduction and/or freezing as it is cooled by the frozen porous medium. In the "wet" pack layer, in which the frozen layer has reached 0°C (H_{cc} = "cold content" = 0), infiltrating liquid water 0°C fills each layer's liquid storage capacity in sequence, like "a bathtub filling from the top".

In addition to the heat and liquid mass conduction within the pack, the factors of gravity and vapor flux contribute to the metamorphism of the pack. The gravitational effect is due to the pressure of the overlying snow, compacting the lower layers. The vapor flux is due to the circulation of air in a pack with a temperature gradient. All of these processes combined have the effect of enhancing or producing temperature and density gradients within the pack. The values of temperature and density at any point in the pack determine such snow properties as vapor and liquid diffusivities, permeability, heat conductivity and liquid water holding capacity. Therefore, a temperature and density gradient in a pack may also result in a gradient of each of these properties, which are important in tracking the snowmelt process.

In a multi-layered pack, the energy balance Equation (3), given previously, may be usedto describe the energy balance of the surface layer, if the ground conduction term is excluded. The internal layers of a pack may each be modeled with heat balance equations also. The components of the heat balance equation for an internal snow pack layer are:

1. radiation penetration into snowpack
2. heat conducted from above and below by snow layer temmperature differences

3. heat transmitted by vapor condensation and sublimation due to local temperature gradients.

4. heat transmitted by the cooling of melt or rain water, including latent heat exchange.

2.3.1 Radiation Penetration

Solar radiation penetrates into the snowpack approximately according to an exponential law (Snow Hydrology 1956):

$$I(z) = I_o \exp(-k_i z) \tag{30}$$

where

I_o = insolation received at snowsurface

$I(z)$ = insolation received at depth z below surface

k_i = extinction coefficient for snow

Radiation penetration increases with the density of the snow because the granularity of less dense snow permits a great deal of reflection and scattering of radiation. As it becomes denser, reflection and scattering are decreased until in the limit of clear ice, it is nearly translucent. In a Snow Hydrology (1956) study, the extinction coefficient (in cm^{-1}) was measured to be 0.28 for snow with a density of 0.26 dropping to 0.166 cm^{-1} for a density of 0.45. Anderson (1976) makes use of a theoretical relationship proposed by Bohren and Barkstrom (1976):

$$k_i = 0.45\left(\frac{k_o}{d_s}\right)^{1/2} \frac{\rho_s}{\rho_i} \tag{31}$$

where

k_o = extinction coefficient for clear ice

d_s = grain diameter

ρ_i = density of ice

ρ_s = density of snow.

2.3.2 Conduction

The heat transfer due to conduction between layers may be expressed by Fourier's equation of heat conduction:

$$H_{c_i} = -\lambda_i \left.\frac{\partial T}{\partial z}\right|_{z=z_i} - \lambda_{i-1} \left.\frac{\partial T}{\partial z}\right|_{z=z_i} \tag{32}$$

where

H_{c_i} = heat flux by conduction into i^{th} layer from surface

λ_i = thermal conductivity of snow in the i^{th} layer

Using Fick's second law we may express Equation (32) in terms of the change in temperature of layer i over time t (Yen(1965)):

$$c_i \rho_s \frac{\partial T}{\partial t} = \frac{\partial}{\partial z}(\lambda \frac{\partial T}{\partial z}) \tag{33}$$

Thermal conductivity (λ) has been found to be a strong function of density (Yen (1965)). When the density varies with depth, thermal conductivity also varies with depth, thus Equation (33) becomes:

$$c_i \rho_s \frac{\partial T}{\partial t} = \lambda \frac{\partial^2 T}{\partial z^2} + \frac{\partial \lambda}{\partial z} \frac{\partial T}{\partial z} \tag{34}$$

2.3.3 Water Vapor Diffusion

Because of water vapor diffusion towards cooler parts of the snow pack, some vapor undergoes a phase change. The net amount of condensation, m, where negative indicates a gain of water vapor by the snowpack, is given by Anderson (1976):

$$\frac{\partial c}{\partial t} = \frac{\partial}{\partial z} [D_e \frac{\partial c}{\partial t}] = m \tag{35}$$

where

D_e = water vapor effective diffusion coefficient in snow

C = concentration of water vapor in snow pores.

The amount of heat gained by the snowpack due to water vapor diffusion is then given by (Anderson 1976):

$$Hw_v = L_s \cdot m \tag{36}$$

where

Hw_v = heat flux due to water vapor diffusion

L_s = latent heat of sublimation

If $\frac{\partial c}{\partial z}$ is written as $\frac{\partial c}{\partial T}\frac{\partial T}{\partial z}$, and likewise the second derivative $\frac{\partial^2 c}{\partial z^2}$ is written $\left\{\frac{\partial c}{\partial T}\frac{\partial^2 T}{\partial z^2} + \frac{\partial^2 c}{\partial z^2}\left(\frac{\partial T}{\partial z}\right)^2\right\}$, then the water vapor derivatives in Equation (35) are now all with respect to temperature. According to Anderson (1976), water vapor concentration in a saturated ice medium can be expressed as:

$$c = e_i/R_w \cdot T \tag{37}$$

where R_w = gas constant for water vapor

e_i = saturation vapor pressure over ice

and $$e_i = 3.56 \times 10^{10}\ (\exp(-L_s/R_w T))\text{mb} \tag{38}$$

The partial derivatives of c with respect to temperature may be calculated from these equations.

Alternatively DeQuervain (1973) derives a more empirical relationship for the water vapor diffusion. He expresses m, the net sublimation, as:

$$m = -(f\ D_e/RT)\ AB\ \left[\frac{\partial^2 T}{\partial z^2} + B\left(\frac{\partial T}{\partial Z}\right)^2\right]\ \exp[B(T-273)] \tag{39}$$

where A and B are determined by experiment and f is considered to be unity for snow (DeQuervain (1973)). The heat transferred by this process is again $Hw_v = L_s \cdot m$.

2.3.4 Infiltration Into Snowcover

Rain and meltwater infiltration into snowcover has been studied extensively by Colbeck (1971, 1972, 1973, 1975, 1978a, and 1978b). The process has a large influence on the metamorphism and melt of snow through the advection of sensible and latest heats, and through the advection of water mass.

As water mass percolates through a pack, it is assumed that all thermal and capillary requirements are satisfied in each layer before any liquid travels to the next layer. That is, the flow is assumed slow enough and/or the layer thick enough that there is enough time for the exchange to take place before gravity and pressure force the liquid downward.

As warmer liquid penetrates into a volume layer of snow, the liquid cools, transferring its heat to the subfreezing snow. The heat transferred may or may not be sufficient to raise the snow layer temperature to 0° C. The heat transferred may be written:

$$P_x = (T_p - 0°\ C)\ C_p\ (i\Delta t) \qquad (40)$$

where

P_x = heat transferred by rainwater or meltwater (cal/cm^2)

T_p = temperature of rain or meltwater infiltration (°C)

$i\Delta t$ = mass of precipitation or infiltration in time interval t per unit area

= rate of precipitation or infiltration x time interval (g/cm^2)

c_p = specific heat of water (cal/g°C)

The "cold content" H_{cc} of the layer under consideration is defined as that amount of heat necessary to raise the temperature of the layer to 0° C. Thus quantity may be written:

$$H_{cc} = \rho_s C_p d_s T_s \tag{41}$$

where

d_s = depth of layer (cm)

There are initially three cases to consider: $P_x < H_{cc}$, $P_x = H_{cc}$, or $P_x > H_{cc}$.

Case #1

If $P_x > H_{cc}$, i.e. The snow is still subfreezing after the rain has cooled to 0°C, some or all of the rainwater will freeze, giving its latent heat to snow:

$$P_{Lfmax} = (i\Delta t)_p \tag{42}$$

where

P_{Lfmax} = heat given up if total mass of rain or infiltration in Δt freezes

If $P_x + P_{Lfmax}$ is still less than or equal to H_{cc}, the frozen rain and snow will remain at the average raised temperature ($\leq$ 0°C). If, on the other hand, $P_x + P_{Lfmax}$ is greater than H_{cc}, only enough of the 0°C liquid will freeze to raise the snow to 0°C. Some of the remaining liquid will stay in the layer to contribute to the fractional liquid water holding capacity ($W_{\ell max}$) of the layer. When that layer is full any excess liquid will move on to the next layer. These conditions, which must be satisifed in sequence, may be written:

$$P_x + M_p L_f = H_{cc} \tag{43}$$

and

$$((i\Delta t) - m_p) - (W_{\ell max}\, \rho_\ell\, d_s - W_\ell\, \rho_\ell\, d_s) = O\Delta t \qquad (44)$$

where

m_p = mass of rainwater which freezes to bring snow to 0°C

W_ℓ = amount of fractional liquid water depth held before Δt

O = outflow of liquid water from layer under consideration

ρ_ℓ = density of liquid

d_s = depth of snow

Case #2

If $P_x = H_{cc}$, both the total mass of rainfall in Δt and the snow will have reached 0° C. Therefore, $m_p = 0$ in Equation (44) above, which yields the outflow, if any, of liquid at 0°C from the layer.

Case #3

If $P_x > H_{cc}$, some or all of the snow in the layer may melt:

$$H_m = P_x - H_{cc} \qquad (45)$$

$$M_s = H_m / L_m \qquad (46)$$

where

H_m = heat available for melting snow

M_s = mass of snow which can melt due to H_m

L_m = latent heat of melting

If $M_s \geq d_s\, \rho_s$, all of the snow in the layer will melt and the outflow from the layer will be:

$$O\Delta t = i\Delta t + d_s \rho_s \qquad m \geq d_s \rho_s \tag{47}$$

If $m_s > d_s \rho_s$ the additional heat will be carried with the outflow:

$$P'_x = H_m - L_m d_s \rho_s \tag{48}$$

where

P'_x = heat content of liquid outflow.

This covers all of the possibilities for infiltration and freezing of rainwater in the snowcover under the assumptions previously outlined. The effects of meltwater in periods of no rain are clearly the same. However, as all meltwater is assumed to be initially at 0°C, its heat content P or P' is equal to zero.

The infiltration process is complicated by two major factors: The melting and refreezing processes alter the density of snow; and there is a time lag between introduction of liquid and outflow of liquid which is often significant. The density effects are discussed in the next section. The lag time will now be considered.

In Anderson's (1976) report, the lag time is estimated empirically by:

$$L_w = \frac{L_{wmax} \rho_w}{CW_2 \cdot O\Delta t + 1.0} \tag{49}$$

where

L_w = actual lag time for layer with outflow (hrs)

$O\Delta t$ = outflow from layer (cm)

CW_2 = empirical parameter

L_{wmax} = maximum lag for a snow layer of thickness d_s: (50)

= $CW_1 (1 - \exp(-0.0025 \cdot d_s/\rho_s)$

CW_1 = empirical parameter

A subsequent report by Colbeck (1978) presents a more physically-based method of determing the speed at which a melt- or rain-water wave propagates downward through a snowpack. From earlier experiment (Colbeck 1971) it was determined that the flux U_w and outflow mass (excess liquid) are related by (Colbeck 1978):

$$U_w = \frac{\rho_{wg}}{\mu_w} k_s \left(\frac{0\Delta t}{\rho_w} \cdot \frac{\rho_i}{\rho_i - \rho_s} \right)^3 \qquad (51)$$

where

k_s = intrinsic permeability of snow (mm^2)

μ_w = viscosity of water

U_w = liquid flux = volume liquid/cm^2 sec

ρ_i = density of ice

Then the "melting front" propagates at a speed given by (Colbeck 1978):

$$\frac{d\xi}{\partial t} = \frac{U_w \rho_w d_s}{0\Delta t + m_p} \qquad (52)$$

The intrinsic permeability was postulated to be related to the grain-size and density by the expression (Shimizv, 1970);

$$k_s = 0.77 \, d_s^2 \exp(-7.8 \, \rho_s) \qquad (53)$$

where

d_s = grainsize

Expressions (51) through (53) are complicated by the fact that liquid water percolation through snow tends to increase grainsize in addition to its freezing, melting, and storage effects.

Anderson (1976) ignores this cause of grain growth. In addition to the empirical and somewhat arbitrary expression (49) for lag time, it was found necessary to further attenuate the lagged outflow beyond the attenuation introduced by liquid water storage and freezing of liquid water (Anderson, 1976). He expresses the attenuated outflow:

$$O_+ = \frac{S + W_L}{CW_3 \exp(-CW_4 \cdot W_L \cdot (\rho_s/d_s) + 1.0)} \qquad (54)$$

where

O_+ = snow cover outflow (cm/hr)

S = excess water in storage (cm)

W_L = amount of lagged excess liquid water (cm)

CW_3 and CW_4 = expirical parameters

2.3.5 Density Changes

The most important property of the snow itself in determing the rate of snowmelt is the density of the snow. Snow density may be described as either weight per unit volume, or (in dimensionless form) as weight per unit volume relative to solid ice. Ice density in grams per cubic centimeter is approximately 1.0, so that when snow density is expressed in these units, the dimensional or dimensionless forms are nearly the same.

Other snow properties which affect the rate of snowmelt include the amount of impurities present, and the grainsize. These properties are not discussed here.

Density may be related to the effective thermal conductivity of snow, which governs heat conduction through the snow pack. Yen (1965) experimentally derived the expression:

$$k_e = 0.0077 * \rho_s{}^2 \qquad \text{cal(cm sec°k)} \qquad (55)$$

Anderson (1976), in the model used herein, used the expression.

$$k_e = 0.00005 + C_k \, \rho_s{}^2 \qquad \text{cal/(cm sec°k)} \qquad (56)$$

where C_k is an empirical parameter that may be varied

The radiation extinction coefficient also depends on the density of the snow. This relationship has been given previously in Equation (31).

It has also been suggested that albedo is related to density, although grainsize is the more likely property from which albedo may be determined. However, since grainsize growth and decay is not well understood, attempts have been made to relate albedo to density. Anderson (1976) uses the expression:

$$A_s = (1.0 - 0.206 \, C_\nu \cdot (G_1 + G_2 \, \rho_s^2 + G_3 \, \rho_s{}^4)^{1/2}) \qquad (57)$$

where

ρ_s = density of snow (dimensionless)

A_s = albedo of snow

C_ν, G_1, G_2, G_3 = empirical parameters

Also, the intrinsic permeability of snow, which governs the flow of liquid water through the snow may be related to density through the expression (Anderson 1976):

$$k_s = 7.7 \cdot (G_1 + G_2 \, \rho_s^2 + G_3 \, \rho_s^4)^2 \cdot \exp(-7.8 \, \rho_s) \qquad (55)$$

where

k_s = intrinsic permeability (mm^2)

Density changes occur in a snowcover due to: mass flux by water vapor diffusion or liquid water infiltration with subsequent refreezing; compaction; or destructive metamorphism.

The change in density due to vapor flux has been expressed by DeQuervain (1963) as:

$$\frac{\partial \rho_s}{\partial t} = f \cdot \frac{D_e}{R_w T} (6.42)(0.0857) \left[\frac{\partial^2 T}{\partial z^2} + 0.0857 \left(\frac{\partial T}{\partial z} \right) \right]^2 \cdot \tag{59}$$

$$\exp[0.0857(T-T_o)]$$

where

T_o = 273.16°k

T = absolute snow temperature

R_w = gas constant for water vapor

D_e = water vapor diffusion coefficient

It is seen that this expression depends strongly on the temperature gradient within the pack.

Anderson (1976) derives a similar expression, where D_e varies with depth:

$$\frac{\partial \rho_s}{\partial t} = D_e \cdot \frac{\partial C}{\partial T} \frac{\partial^2 T}{\partial z^2} + \frac{\partial C}{\partial T} \frac{\partial D_e}{\partial z} \frac{\partial T}{\partial z} + D_e \frac{\partial^2 C}{\partial T^2} \left(\frac{\partial T}{\partial z} \right)^2 \tag{60}$$

where

$$\frac{\partial C}{\partial T} = \frac{e_i}{R_w T^2} \left[\frac{L_s}{R_w \cdot T} - 1.0 \right] \tag{61}$$

and

$$\frac{\partial^2 C}{\partial T2} = \frac{e_i}{R_w T^3} \left[\left(\frac{L_s}{R_w T} \right)^2 - \frac{4 L_s}{R_w \cdot T} + 2.0 \right] \tag{61}$$

which represent the first and second derivatives of water vapor concentration with respect to temperature.

e_i = saturation vapor pressure over ice

L_s = latent heat of sublimation

C = water vapor concentration (dimensionless)

= $e_i/R_w T$

Bader (1939, 1963) has studied the compaction of dry snow. Other investigators have included Kojima (1967). These studies are considered by Anderson (1975), who arrives at an expression of the form:

$$\frac{1}{\rho_s}\frac{\partial \rho_s}{\partial t} = C_1 W_s \exp(-C_2 \rho_s) \exp(-0.08(T_o - T)) \quad (63)$$

where

C_1, C_2 = empirical coefficient

W_s = weight of overlying snow interms of water-equivalent (cm)

T_o = 273.16°k

Density may also change in isothermal, steady conditions due to a process called "destructive metamorphism". Destructive metamorphism is the process which converts the pointed ice crystals to spherical ice crystals, changing the relation of the snow particles to each other and thus increasing the density. Anderson (1976) uses the relationship:

$$\frac{1}{\rho_s}\frac{\partial \rho_s}{\partial t} = C_3 \exp(-C_4(T_o - T)) \qquad \rho_s \leq \rho_d \quad (64a)$$

$$\frac{1}{\rho s}\frac{\partial \rho s}{\partial t} = C_3 \exp(-C_4(T_o - T)) \exp(-0.46(\rho_s - \rho_d)) \qquad \rho_s \geq \rho_d \quad (64b)$$

where

ρ_d = experimental limiting density

C_3 = fractional settling rate at 0°C for $\rho_s < \rho_d$

C_4 = empirical constant

2.4 National Weather Service Computer Model

The model used in this study of snowmelt is the National Weather Service Model described in NOAA Technical Report NWS 19, A Point Energy and Mass Balance Model of a Snow Cover, by Eric Anderson (1976). This model utilizes much of the theory outlined above. Where several equations for the same parameter were given in the preceding sections, the one which is referenced as Anderson (1976) is the one used in the computer model. Also, turbulent heat transfer is computed in Anderson's model by Dalton-type expressions such as Equation (25). The computer model allows the option of using an "empirical" wind function which assigns empirical coefficients (e.g., "a" + "b" in Equation (25)) or a "theoretical" wind function which computes the coefficients based on theoretical considerations.

The model is a finite difference model which can be made implicit or explicit. It is recommended that the semi-implicit Crank-Nicolson scheme be used. The Newton-Raphson iteration method is used where iterative procedures are needed.

For a more detailed description of Anderson's model, see Anderson (1976).

The NWS model does not simulate energy and mass transfers for a snowpack overlain by a vegetal canopy. In the next section the theory of vegetation processes over a snowpack is described, and a computer model of the vegetation is derived and linked to Anderson's computer model which may be run with a limited number of options (to be described). Minimal changes are made to Anderson's model and are described in the next section.

CHAPTER 3

DYNAMIC EFFECTS OF VEGETATION

3.1 Introduction

Much of the earth's surface is covered with vegetation, so that computation of the exchange of radiation, sensible heat, latent heat, and moisture between a vegetated surface and the atmosphere is of considerable importance in the study of snowmelt. The presence of forest cover alters these exchanges in several ways. These include interception of snow by the canopy, shielding of the surface from solar radiation and wind by the canopy, and emission of longwave radiation by the canopy. Transpiration may be considered negligible below a temperature of about 40°F (Viessman et. al. 1977), and when snow is present.

3.2 Interception

The amount of snow intercepted by a canopy has been found to be dependent on: canopy density, (defined as that fraction of the surface which is blocked from solar radiation by the canopy); branching angle; foliage type; and height of canopy (Snow Hydrology, 1956). It has also been suggested that precipitation intensity is another determinant of the quantity of inteception (Molkanov, 1955). Hoover and Leaf (1967) observed a subalpine forest for several months and note that the intercepted snow accumulation rate accelerates as the depth of snow on the tree increases. They also observe that snow often remains on the canopy for periods of several weeks or more.

The amount of snow intercepted is commonly expressed as a percentage of total snowfall. Chow (1962) gives interception as a percentage of total snowfall for different tree types in the Northeast (White Mountains), but he fails to specify the canopy densities involved. Both Chow (1962) and

Rowe and Hendrix (1951) note that little difference has been found between rain and snow interception percentages. Regression equations have been developed (Snow Hydrology 1956; Rowe and Hendrix (1951)) which express interception percentage as a function of canopy density. Willen and Shumway (1971) develop a relationship for water equivalent intercepted as a function of canopy density and precipitation. This relationship is (converted to centimeters):

$$I = 0.6893 \cdot P \qquad P \leq 1.27 \text{ cm} \tag{65}$$

$$I = 0.6886 \cdot M + 0.1478 \qquad P > 1.27 \text{ cm} \tag{66}$$

where

I = intercepted water equivalent (cm)

M = canopy density

P = snowfall water equivalent

Willen and Shumway's (1971) equations assume that interception is independent of canopy density until a limiting snow depth is reached. A more easily acceptable conclusion is found by using the relations given in Snow Hydrology (1956). We may choose a representative relationship to provide this equation:

$$I = (0.37M) \cdot P \tag{67}$$

This total intercepted depth is the sum of the depths intercepted, by all the leaves and branches, per unit area. The intercepted snow will be treated as a bulk layer of uniform temperature, density and depth. This bulk layer will, however, be considered to be ventilated by the air within and beneath the canopy. This is consistent with the approach of Deardorff (1978). A definition sketch of the canopy-surface system, which is shown conceptually rather than realistically, is given in Figure (3). The ventilating air promotes heat and moisture fluxes from the surface and from the surface and the canopy.

Snow which is intercepted may leave the treetops either by mechanical removal (i.e., wind) by sublimation, or by melt. Miller (1962) observed that the effects of wind and gravity are opposed by cohesion among snowflakes and their adhesion to the trees. The more cohesive the snow particles, the greater their resistance to mechanical removal. This cohesiveness is largely determined by the liquid water content of the snow. We will assume that the depth of snow stored on the foliage decreases exponentially as the ambient wind speed increases or as the period of wind blowing increases. Thus:

$$I(t_2) = I(t_1) \exp(-k_u U_a \Delta t) \tag{68}$$

where

$I(t_2)$	=	depth of snow at time $t=t_2$	(cm)
$I(t_1)$	=	depth of snow at time $t=t_1$	(cm)
U_a	=	wind speed	(cm/sec)
Δt	=	time interval (t_2-t_1)	(sec)

The unknown coefficient k_u will depend on liquid water content and may be determined from experiments. The snow which is blown off of the canopy is considered to be deposited evenly over the landsurface area for the one dimensional approach herein.

Different types and densities of vegetation have different maximum interception capacities. Molchanov (1956) and Snow Hydrology (1956) have studied these relationships and provide some representative values. The value for a particular site may vary from year to year and from season to season as the canopy density increases or decreases due to growth, thinning, loss, or gain of leaves, etc.

3.3 Canopy Energy and Water Budget

The intercepted snow may collect and store heat similarly to the surface snowcover. To determine heat and moisture transfer from the vegetated fraction of the surface area to the atmosphere above and to the landsurface below, an energy and mass budget must be computed. Deardorff (1978) develops a simple, linear, physically-based model for vegetative heat and moisture transfer which is applied herein. Deardorff (1978) assumes a single layer canopy of negligible heat capacity with meteorological variables as defined in Figure (3). For the case when intercepted snow is present Deardorff's model must be adjusted in two ways: (1) the snow may provide a large heat capacity; and (2) transpiration is negligible.

The environment in Deardoff's model is divided into several components. First, there is the atmosphere above the vegetation which is represented by average values of temperature (T_a), wind speed (U_a) and specific humidity (q_a) just above the canopy. Next, there is the canopy layer itself, which may or may not have snow deposited upon it. This canopy layer is defined by the canopy density (that fraction of unit area landsurface which is shielded from solar radiation), temperature of the foliage or intercepted snow (T_f), specific humidity at the foliage surfaces (q_f), and depth of intercepted water or snow on the canopy (d_f). Beneath the canopy, and ventilating the canopy surfaces, is the in-canopy air which has its own value of temperature (T_{af}), specific humidity (q_{af}) and average wind speed (U_{af}).

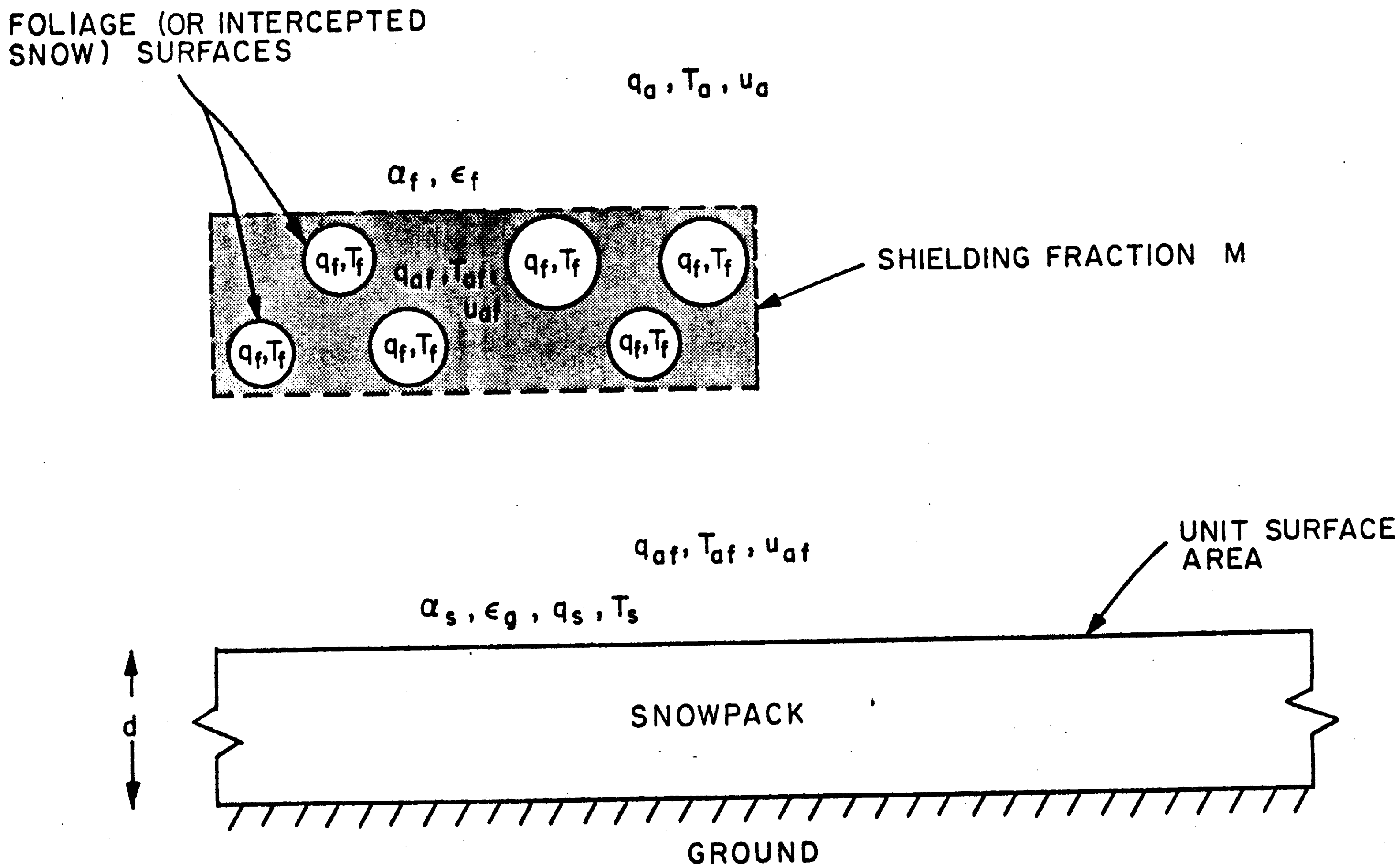


FIGURE 3. CONCEPTUAL DEFINITION SKETCH

The land (snow) surface is defined by the temperature (T_s) and specific humidity (q_s) at the surface.

Deardorff (1978) assumes that the average air temperature and specific humidity which ventilates the canopy and promotes fluxes from the landsurface and canopy is given by a linear interpolation between values of the temperature of the canopy itself, the temperature above the canopy, and the temperature of the ground (snow) surface. The temperature of the canopy is given by the temperature of the intercepted snow when such snow is present. These interpolations are given by:

$$T_{af} = (1-M)T_a + M(.3T_a + .6T_f + .1T_s) \qquad (69)$$

$$q_{af} = (1-M)q_a + M(.3q_a + .6q_a + .1q_s) \qquad (70)$$

The values of q_a and T_a must be known from meteorological data. The values of T_s and T_f must be determined through energy budgets for the surface and canopy, respectively. This requires knowledge of T_{af} and q_{af}, the two in-canopy parameters, which along with U_{af}, will determine the turbulent fluxes of sensible and latent heats. When snow is present, q_s and q_f are equivalent to the saturated specific humidities over ice at the respective temperatures, T_s and T_f. When intercepted liquid-water only is present, q_f is given by the saturated vapor pressure over water at that temperature. When the leaves are dry, q_s is assumed identical to q_{af}. Thus, no evaporative heat transfer would take place in such a case.

When intercepted snow is present, the temperature of the canopy is found by solving an energy budget equation for the canopy:

$$\Delta Q_f = P_f + R_f - H_{sf} - L_s\, E_f \qquad (71)$$

where

R_f = net radiational heat input to foliage

H_{sf} = sensible heat trnasfer from foliage to atmosphere

$L_s\ E_f$ = latent heat transfer from foliage to atmosphere

P_f = heat advected by rainwater to foliage

ΔQ_f = change in heat storage of intercepted snow.

When there is no intercepted snow present, the canopy is assumed to have negligible heat capacity, and so ΔQ_f is set equal to zero. The radiation term in Equation (61) above represents the sum of all shortwave and longwave radiation exchange terms for the canopy. The radiation term is important whether or not intercepted snow is present. Federer (1971) found that a leafless hardwood canopy may absorb roughly 65 percent of the incident radiation, and about 20 percent is reflected, when there is snow upon the ground surface. Deardorff (1978) groups like terms in the radiation budget to reduce the equation for the sum of all radiation terms to the expression:

$$R_f = M\left[(1-\alpha_f)R_s + \varepsilon_f\ R_{\ell i} + \frac{\varepsilon_f\ \varepsilon_g}{\varepsilon_f+\varepsilon_g-\varepsilon_f\ \varepsilon_g}\ \partial\ T_g^{\ 4} - \varepsilon_f\frac{\varepsilon_f+2\varepsilon_g-\varepsilon_f\ \varepsilon_g}{\varepsilon_f+\varepsilon_g-\varepsilon_f\ \varepsilon_g}\ \partial\ T_f^{\ 4}\right] \quad (72)$$

where

α_f = albedo of foliage (or intercepted snow)

$R_{\ell i}$ = incoming longwave radiation

R_s = incoming shortwave radiation

$_f$ = emissivity of foliage (or snow)

$_g$ = emissivity of landsurface (snowsurface)

= Stefan-Boltzmann constant

R_f = radiation exchange from vegetation to atmosphere per unit area.

It should be noted here that the albedo of intercepted snow (α_f) is much lower than that of snowcovered ground (Leonard and Eschner 1968). Whereas, snow on the ground generally has an albedo between 0.70 and 0.90, intercepted snow rarely produces an albedo greater than 0.40 due to the irregularity of the surface and non-uniformity of accumulation.

It will be assumed that intercepted snow will cover the entire leaf or branch uniformly, and therefore, evaporation (sublimation) will proceed at the potential rate. This rate is expressed by (Deardorff 1978):

$$E_f = E_{fpot} = N \rho_a C_f U_{af}[q_{sat}(T_f) - q_{af}] \tag{73}$$

where

N = leaf or branch area index

ρ_a = density of air

$q_{sat}(T_f)$ = q_f = saturated specific humidity at temperature T_f

q_{af} = mean specific humidity of air ventilating and beneath canopy

It has been found that the leaf area index N is approximately equal to 7M in many cases (Allen and Lemon (1972), Monteith et. al. (1965)). Its actual value will vary depending on the type and age of foliage. It is assumed that this ratio of branch area to shielding factor for leafless trees is the same as the ratio of leaf area to shielding factor for leafy trees. This assumption follows from the fact that the number of layers of leaves on a tree corresponds to the number of layers of branches. The shielding factor M will vary from winter to summer for a hardwood forest; however, the relation between M and N will be assumed constant.

The net sensible heat flux, H_{sf}, from the canopy to the surrounding air per unit ground area is approximated by Deardorff (1978):

$$H_{sf} = (1.1)(7M)\rho_a C_p C_f U_{af}(T_f - T_{af}) \quad (74)$$

where

ρ_a = density of air

C_p = specific heat of air at constant pressure

C_f = dimensionless heat transfer coefficient for the foliage

U_{af} = mean wind speed within and beneath the canopy the canopy, which ventilates the canopy and also promotes heat fluxes from the ground

The factor 1.1 in Equation (64) is chosen to account roughly for the effects of stems, stalks, and twigs which exchange heat but have not been included in the branch (leaf) area index relationship (Deardorff 1978).

The dimensionless heat transfer coefficient for the canopy, C_f, is assumed to be given by the product of forced convection and a free convection enhancement (Deardorff 1978):

$$C_f = 0.01[1 + (0.3\ (m/sec)/U_{af})] \quad (75)$$

where the value 0.01 is quoted by Deardorff to be from a study of forced convection over plants by Kumar and Borthakur (1971).

The heat advected by rainwater to intercepted snow may be given by:

$$P_f = [(T_p - 0°C)C_{pw}\ \rho_w\ P] \cdot (1.0 - 0.37M) \quad (76)$$

where

T_p = temperature of precipitation °C

C_{pw} = specific heat of water cal/g°C

ρ_w = density of water g/cm^3

P = depth of precipitation cm water

Note that the last factor in parenthesis in Equation (76) is the relationship previously given in Equation (57) to determine the amount of precipitation intercepted.

The change in heat storage for the intercepted snow may be given by:

$$\Delta Q_f = I \cdot \rho_s \cdot \frac{d[C_i\ T_f]}{\partial t} + L_f\ \rho_w \frac{\partial w}{\partial t} \tag{77}$$

where

I = depth of intercepted snow

C_i = specific heat of ice

ρ_s = density of snow cover (excluding liquid water)

$\partial w/\partial t$ = amount of liquid water that undergoes a phase change in dt, expressed as a depth cm.

The specific heat of ice is once again given by:

$$C_i = 0.0222 + 0.00176 \cdot T_f \tag{78}$$

Liquid water may accumulate in the intercepted snow until the water holding capacity of the snow is exceeded. At that point, runoff from the foliage to the ground will occur. The water holding capacity will be assumed to be the same for intercepted snow as for the surface snowcover. For computational purposes the liquid runoff from the foliage will be considered to be distributed evenly over the unit surface area and treated as rain upon the surface snowcover. This runoff will not advect heat, however, since the runoff is assumed to occur at the freezing temperature. It may refreeze upon reaching the surface cover, whereupon it will release its latent heat of fusion.

The wind profile within and beneath a canopy has been found to be exponential (Cionco 1965, 1970). The average wind speed within and beneath a canopy can thus be expressed:

$$U_{af} = \frac{1}{H} \int_{o}^{H} U_a \exp(a(z/H - 1)) \, dz \tag{79}$$

where

H = height of canopy

a = attenuation coefficient

U_a = wind speed just above canopy

Integrating Equation (65) yields

$$U_{af} = \frac{U_a}{a} (1 - \exp(-a)) \tag{80}$$

The attenuation coefficient, a, depends on the density and rigidity of the vegetation, and has been examined by Cionco (1970).

Because of this non-logarithmic wind profile, the formulation of turbulent energy fluxes from the surface as derived in Section 2.1 based on Prandtl's theory cannot be used. Brutsaert (1979) analytically derived new expressions for the eddy fluxes and transfer coefficients when a canopy is present. However, his solution requires knowledge of boundary conditions which are usually unknown, and is mathematically complex. Deardorff (1978) once again uses a simpler linear interpolation for the bulk transfer coefficient applicable to the surface beneath a canopy:

$$C_{Hg} = (1 - M) \, C_{Ho} + M(C_{Hh}) \tag{81}$$

where

C_{Hg} = bulk transfer coefficient applicable to surface beneath a canopy (dimensionless)

C_{Ho} = bulk transfer coefficient applicable to bare (snow) surface (dimensionless)

C_{Hh} = bulk transfer coefficient applicable to top of dense vegetation.

Deardorff's bulk transfer coefficients can be converted to the form of Anderson's (1976) "theoretical wind function" by comparison of Deardorff's and Anderson's heat transfer equations. The bare surface transfer coefficient is that coefficient determined, as before, with Prandtl's theory and the Monin-Obukhov modifications. The bulk transfer coefficient applicable to the top of the dense vegetation is given in Equation (64) as:

$$C_{Hh} = 7M\ C_f \tag{82}$$

The total sensible heat transfer (H) from the vegetated surface is then given by:

$$H = H_{sf} + H_{sg} \tag{83}$$

where

H_{sg} = sensible heat transfer from surface cover to atmosphere using the transfer coefficient given by Equation (72)

Similarly,

$$LE = LE_f + LE_g \tag{84}$$

where

LE_g = latent heat transfer from surface cover to atmosphere using the transfer coefficient given by Equation (72)

The longwave radiation emitted by the surface beneath a canopy is found using the theory of radiative flux between two parallel surfaces (Deardorff 1978):

$$R_{Ls} = (1-M)\left[\varepsilon_f \; \sigma T_g^{\;4} + (1-\varepsilon_g) R_{Li}\right] + M \left[\frac{\varepsilon_f \; \sigma T_g^{\;4} + (1-\varepsilon_g)\varepsilon_f \; \sigma T_f^{\;4}}{\varepsilon_f + \varepsilon_g - \varepsilon_f \; \varepsilon_g}\right] \tag{85}$$

where

R_{Ls} = longwave radiation emitted by surface beneath canopy. When snow is present, all terms in Equations (73) with subscript g should be considered to represent the snow surface, rather than a bare soil surface.

3.4 Computer Submodel of Vegetation

Anderson's (1976) one-dimensional model provides the core to which a computer model of non-transpiring vegetation has been linked. The principles governing heat and mass exchange when vegetation is present have been described in Section 2.1. The vegetation model consists of several subroutines which are called by Anderson's model, and some adjustments which have been made to Anderson's model itself.

In this section, a description of each newly-introduced subroutine is given. Following the descriptions of the subroutines is a section delineating all of the major changes introduced in Anderson's 1976 program. The names of the vegetation variables as used in the Fortran computer program may be found, with their definitions, in Table 1. A listing of each vegetation subroutine and of the sections of Anderson's (1976) program which have undergone changes is provided in Apendix I. (For a complete listing of the original Anderson (1976) program one must contact the National Weather Service.) In Appendix II a description of input cards necessary to run the program is given, with an explanation of what options are or are not available when simulating vegetation-covered snow surfaces.

Subroutine vbgin

This subroutine initializes the conditions and properties of the vegetation with regard to: total intercepted snow present, density of intercepted snow, canopy density, and empirical factors which determine the fraction of precipitation intercepted, rate of snow blowing from the canopy, and maximum interception.

Subroutine vegvar

This subroutine redefines the meteorological variables as given in Anderson's (1976) subroutine datain as values above the vegetation. These values must have been measured at a height just above the vegetation.

Using the Snow Hydrology (1956) formula, given in Section 2.2 (Equation (57)), the value of the depth of intercepted precipitation at each time step for which precipitation occurs is determined. The depth of the rest of the precipitation, which appears as throughfall to the surface, is also determined.

Average wind speed below the canopy is calculated using Cionco's (1965) formula. Solar radiation is also attenuated as defined by the canopy density.

The other meteorological parameters for the surface and which ventilate the canopy must be determined at each time step in subroutine vegbal. These include air temperature, longwave radiation to the surface, and mean specific humidity (or vapor pressure) within and beneath the canopy.

Subroutine vegobt

This subroutine obtains the values of the meteorological variables which are used for the next time step from the data derived in subroutine vegvar. The form and temperature of precipitation during the time step is also computed. The subroutine follows the form of Anderson's (1976) subroutine obtain fairly closely.

Using information obtained from the previous time step, subroutine vegobt also determines the amount of precipitation actually intercepted during the time period, since the vegetation has a maximum capacity. Any precipitation which would exceed the maximum capacity is added to the land-surface precipitation data for the time step.

Subroutine vegbal

This subroutine performs the energy and mass balance of the vegetal canopy and also determines some meteorological variables which affect the surface cover. In this routine, a constant density, single intercepted snow layer is considered. An estimate is made of the sensible, latent, precipitational, and radiational heat transfer from the vegetation to the atmosphere. It also calculates the depth of intercepted snow, temperature of intercepted snow, depth of snow sublimated, depth of snow blown from vegetation, liquid water content of snow, and melt outflow from intercepted snow. When intercepted snow is not present on the canopy, the temperature of the canopy itself is determined. The subroutine also allows for the possibility of liquid precipitation (rain) interception, on the bare (no-snow) canopy. Such intercepted rain may evaporate and the latent heats released or absorbed would be considered in the energy balance for the vegetation. Intercepted liquid is neglected if snow reappears. Also, intercepted rain is not permitted to be blown to the surface snowcover. However, rain events upon the intercepted snow can be simulated by vegbal.

The procedure used in vegbal is a fully explicit finite difference technique. This often necessitates subdivision of the time step to avoid instabilities. The degree of time step subdivision depends on the degree of instability. The time step may be subdivided for any of three reasons, or their combinations: (1) the intercepted snow has just disappeared; (2) successive iterations during a time step do not yield the same result, within specified error limits; or (3) after even sixty iterations to find the temperature of the vegetation, a solution has not been found.

When snow is not present, a linearization of the energy balance equations is used which determines the temperature of the canopy explicity and without iterative procedures. The linearization is necessary for the radiation term, which involves vegetation temperature to the fourth power, and, when intercepted liquid water is present in the specific humidity term. For these cases, a linearization given in Deardorff (1978) is used:

$$(T^{n+1})^4 = (T^n)^4 + 4(T^n)^3(T^{n+1} - T^n) \tag{86}$$

and

$$q_{sat}(T^{n+1}) = q_{sat}(T^n) + \left.\frac{\partial q_{sat}}{\partial T}\right|_{T^n}(T^{n+1} - T^n) \tag{87}$$

where

T^n = temperature at end of previous time step

T^{n+1} = temperature at end of present time step

The value of $q_{sat}(T^n)$ is found by using a formula given by J. F. Bosen (1960):

$$q_{sat}(T^n) = \frac{.622}{p_a} (33.864) \Big[(0.00738\ T^n + 0.8072)^8 - 0.000019 \left| 1.8T + 48 \right| t\ 0.001316\Big] \tag{88}$$

where T^n is given in ° Celcius. The absolute value bars may be treated as simple parentheses for the temperatures under consideration in natural climates. The value of $\partial q/\partial T$ is found using the Clausius Clapeyron equations of the form (Huang, F. F., (1966)):

$$\frac{\partial q_{sat}}{\partial T} = \frac{.622}{p_a} \frac{h_{fg}}{R_w T^2/p_a} \tag{89}$$

where

h_{fg} = enthalpy of vaporization, evaluated at T_n

R_w = gas constant for water vapor

p_a = atmospheric pressure

T = T^n

The enthalpy of vaporization is given by (Eagleson (1970)):

$$h_{fg} = (597.3 - 0.57\ (T^n - 273.16°K)) \tag{90}$$

where T^n is now given in °K

Subroutine vegbal permits the intercepted snow to disappear within a time step. When this occurs, the method of calculating the canopy temperature switches from the iterative "heat storage" procedure to the linearized, "no heat storage" procedure above. The heat and mass fluxes are determined for both segments of the time step and added together to give the total time step values.

The energy balance components are computed at the end of the time step, using the expressions given in Section 2.2. The average values of latent heat, sensible heat, and radiational heat flux are computed by taking the arithmetic means of their values using T^n and T^{n+1}, where T^n and T^{n+1} are the values of vegetation temperature at the beginning and at the end of the present time step, respectively.

Subroutine vegdat

This subroutine stores the values of the variables computed in subroutine vegbal to be printed out in subroutine statda of Anderson's 1976 model.

Changes to Anderson's Model

Of course, one of the major changes to Anderson's model is that the subroutines named above are called from Anderson's main program when vegetation is present. Subroutine vbgin is always called by the program in order to set a value for the canopy density. The vegetation energy balance is performed first, followed by Anderson's surface snowcover energy and mass balance.

Two calls to subroutine nwsnow are made between the vegetation energy balance and the surface cover energy balance. The first call uses the subroutine, which was designed by Anderson (1976) to determine snow added to the surface by snowfall, to determine the snow added to the surface by the blowing of snow from the canopy. This is done by simply changing three of the subroutine arguments. The second call to nwsnow is left as originally intended by Anderson (1976).

Because of the change in wind profile from logarithmic to exponential due to the presence of vegetation, Anderson's subroutine windf had to be altered. This subroutine computes the "wind function" for the time step. The "wind function" is the coefficient involving wind speed which is present in Dalton-type equations for turbulent transfer. For example, in the case of latent heat flux:

$$Q_e = L_s \cdot \rho_w \cdot f(u) \cdot (e_a - e_o) \qquad (91)$$

In this equation f(u) is the wind function. Anderson's program offers the option of using a "theoretical" or "empirical" wind function. In order to use the vegetation model, the "theoretical wind function" option must be taken. This wind function depends on the wind speed, roughness height of the snow, critical Richardson number (above which turbulent conditions no longer exist) and various constants (e.g., station atmospheric pressure, density of water, etc.).

When vegetation is present, this wind function is altered. The wind function is directly related to the dimensionless, bulk transfer coefficient for the bare snow surface C_{Ho} as defined in Equation (81). In order to determine the new bulk transfer coefficient, Deardorff's (1978) method

(Equation (81)) is used. When converted from dimensionless bulk transfer coefficient to "wind function" coefficient, Equation (81) becomes:

$$f(u)_g = f(u)_o \cdot (1-M) + f(u)_h \cdot (M) \tag{92}$$

where

$f(u)_g$ = wind function applicable to landsurface beneath vegetation

$f(u)_o$ = wind function applicable to bare landsurface

$f(u)_h$ = wind function applicable to top of dense canopy

The "bare" wind function $f(u)_o$ is the same as would be calculated in the absense of vegetation. The "top of canopy" wind function may be expressed as:

$$f(u)_h = N\, C_f\, \rho_a\, U_{af}(.622/P_a) \tag{93}$$

where the variables have been defined in Section 2.2.2.

Other changes to Anderson's computer program are in subroutine statda, wherein the computations performed in the vegetation subroutines are printed as output.

CHAPTER 4

RESULTS AND CONCLUSIONS

4.1 Introduction

In the previous chapter, the effects of a vegetation layer on the heat and moisture transfers between a snowcovered area and its environment (atmosphere and soil) were discussed, and a computer simulation model was described. The computer model requires as inputs; the climatic and soil conditions, vegetation characteristics, and certain coefficients necessary for the equations discussed in Chapters 2 and 3. This model has been tested for its sensitivity to canopy density and vegetation characteristics under several synthetic climate conditions. The results of this sensitivity study are presented below.

4.2 Longwave Radiation Reaching Surface

A vegetal canopy may increase or decrease the amount of longwave radiation reaching the ground or snow surface, depending upon several factors. First, the presence of snow on the canopy causes a natural limit to be placed on the emission of longwave back radiation, since snow can get no warmer than 273.16°K. Thus the total longwave radiation emitted over an extended period depends on the degree of persistence of snow on the canopy. Second, the amount of ambient atmospheric radiation is important. Whether the canopy emits as much longwave radiation as it absorbs depends upon the ambient value. Third, the climatic factors of air temperature, specific humidity, windspeed,and total radiation determine the heat balance of the vegetal and surface snowcover. Therefore, by the Stefan-Boltzmann law, a warmer climate may increase the amount of longwave radiation to the surface.

However, since there is a feedback between the heat balances of the canopy and the surface, as formulated by Deardorff (1978) in Equations (69) through (85), (due to shielding by the canopy, increased surface emission, and other effects) this increase is not always the case.

The persistence of snow on the canopy is governed by several factors. First of all, the capacity of the tree to hold snow is obviously a prerequisite to persistence. The larger the volume of snow deposited on a canopy, the longer it may be expected to persist. Secondly, the exponent in the mechanical removal relation (68) will determine how effective wind is in removing snow from the trees. The smaller the value of k_u, the more likely the snow is to remain on the canopy. In this respect the magnitude of the wind velocity is also important. Thirdly, climatic conditions will determine how much of the intercepted snow is removed by sublimation or melt.

Figure (4), (5), and (6) show the longwave radiation reaching the surface snowcover in a warmer two-day period (average above-canopy air temperature 281°K) as a function of canopy density for three different degrees of intercepted snow persistence.

In Figure (4), the snow is not intercepted at all. The warmth of the air and the ability of the canopy to become warm, even above 273.16°K, enables the canopy to emit a great deat of radiation. It is seen that the radiation reaching the surface steadily increases with canopy density.

In Figure (5), the intercepted snow is permitted to blow from, sublimate, or melt off of the canopy during the two-day test period. As long as the snow remains on the canopy, the back radiation from the canopy to the surface is limited to its value at 273.16°K. Therefore, the longer the snow remains on the canopy, the less radiation it emits. At larger canopy

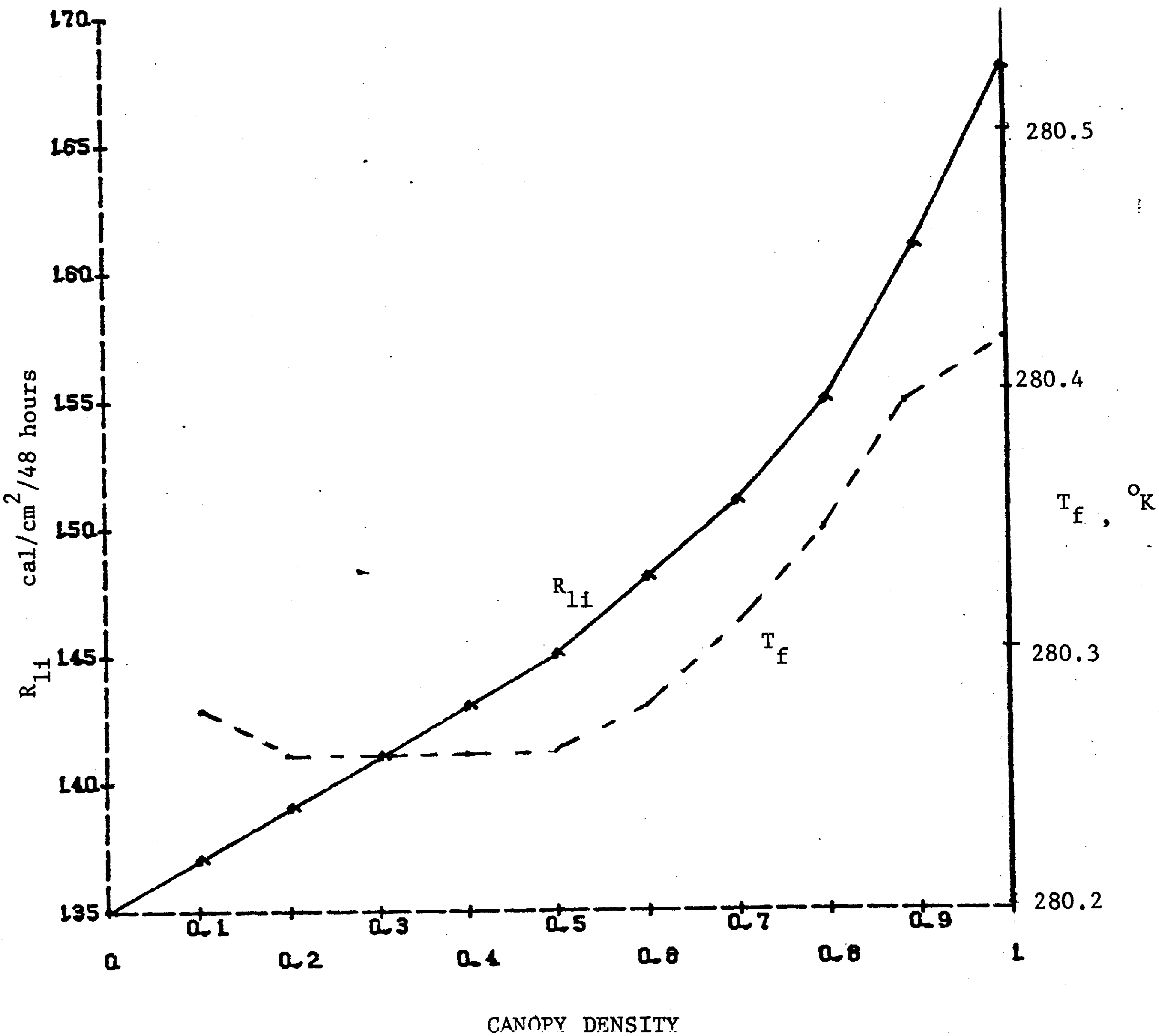


FIGURE 4: Longwave radiation flux to surface cover: case of warm weather, no interception

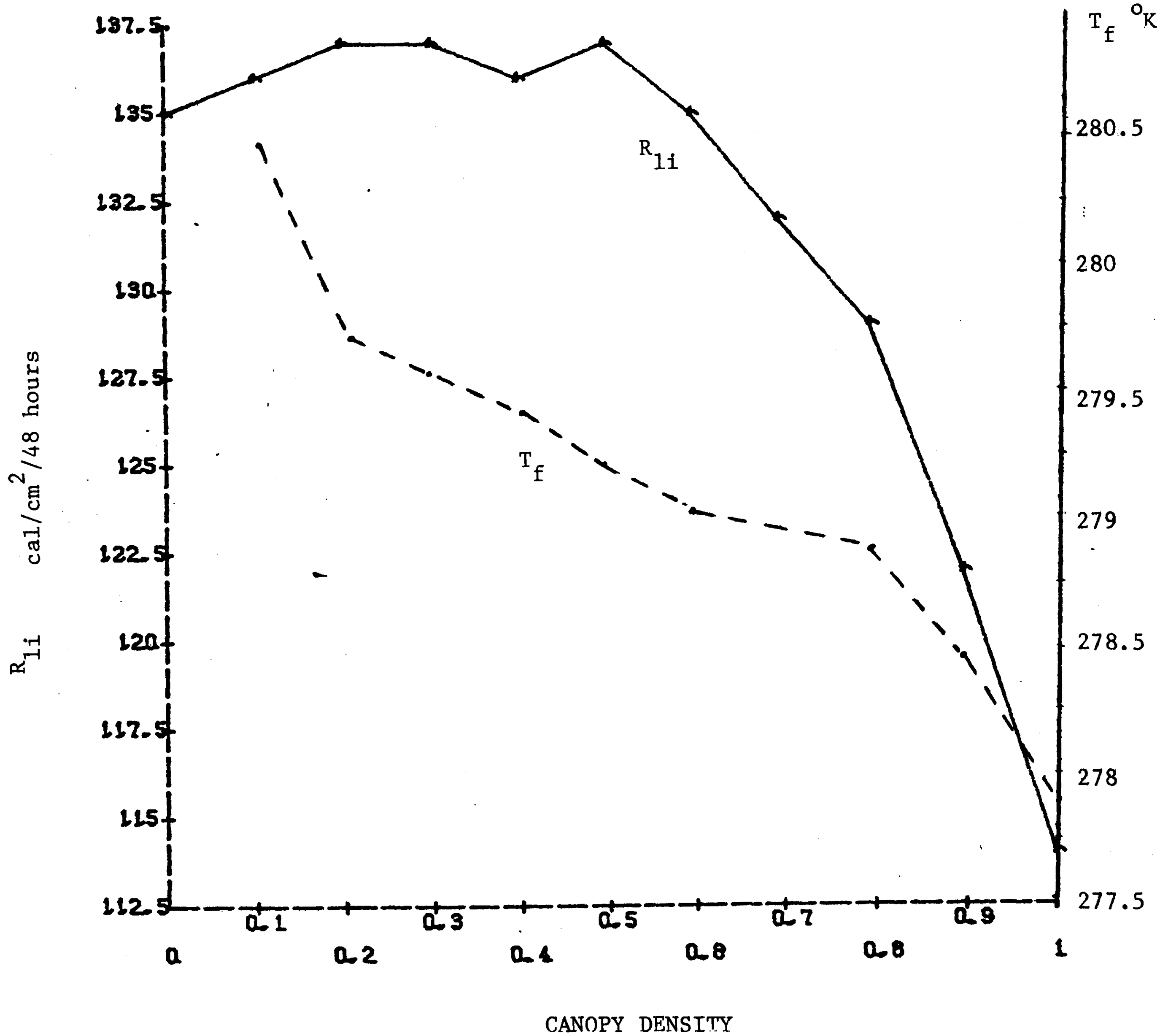


FIGURE 5: Longwave radiation flux to surface cover: case of warm weather, intercepted snow which blows from canopy

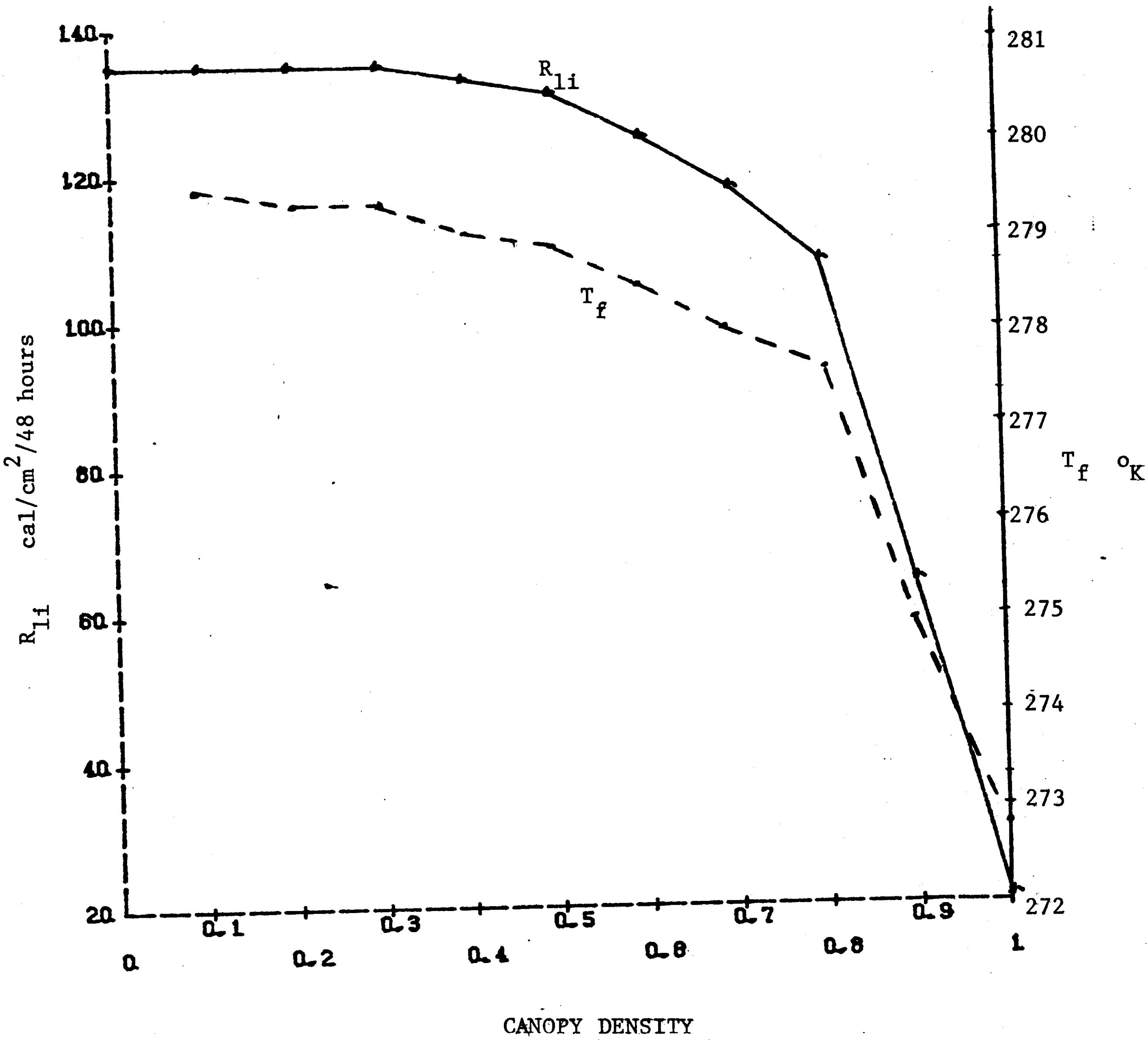


FIGURE 6: Longwave radiation flux to surface cover: case of warm weather, intercepted snow which blows from canopy

densities, a greater depth of snow is intercepted [Equation (67)]. Also, the wind which serves to blow snow from the canopy is decreased [Equation (80)]. The surface snow also will be receiving less solar and ambient atmospheric radiation, and less wind that would promote warming by sensible heat transfer. Therefore, the surface snow temperature may be lowered with increasing canopy density, stifling back radiational flux from the surface to the canopy. The end result is a decrease in net longwave radiation reaching the ground at higher canopy densities. In Figure (5), the longwave radiation reaching the ground appears to fluctuate around a constant value at lower canopy densities before showing the decrease expected at higher densities. At the lower canopy densities the opposing phenomenon of a linear increase in back radiation with increasing M is as strong or stronger than the factors contributing to the decrease. Above a canopy density of 0.5 the factors which contribute to the decrease become dominant. The amount of decrease is only around 15 percent.

In Figure (6), the snow is not blown from the canopy, and therefore, persists on the canopy throughout the two-day period. The longwave radiation steadily drops as canopy density increases, and the decrease is a dramatic 85 percent as full canopy density is reached. Again, it is seen that the factors which contribute to the decrease are most effective at canopy densities above 0.5.

In a colder (average above-canopy densities air temperature = 271°K) climate given the same ambient radiation and humidity conditions, the longwave radiation reaching the surface can be greatly attenuated. In Figure (7), illustrating the case of cold weather and non-intercepted vegetation, the canopy is cooled down and emits less radiation than it blocks. Therefore, the radiation reaching the ground may be expected to decrease with increasing canopy density. However, Figure (7) shows a slight reversal in this trend near M = 0.8, for reasons similar to the previous case. At these higher canopy densities, the wind is greatly attenuated, stifling heat flux from the canopy to the atmosphere, and from the surface to the atmosphere. The surface cover and the canopy thus may tend to stay warmer. The test results showed an increase of about 2°K for both canopy and surface snow cover between M = 0.1 and M = 1.0. In addition, since the albedo of a canopy is less than the albedo of a surface snowcover, the canopy can absorb more solar radiation and thus become warmer. Therefore, at the higher canopy densities the factores which result in increased radiation from the canopy begin to dominate. Figure (8) qualitatively demonstrates this process.

4.3 Turbulent Heat Transfer

Sensible heat transfer to the vegetation from the atmosphere is dependent upon the temperature of the air ventilating the canopy, the temperature of the vegetation, the wind speed, and the Dalton-wind function. The atmospheric temperature, vegetation temperature, and snow surface temperature all control the ventilating canopy air temperature. In Figure (9) we see the sensible heat flux from the atmosphere to the vegetation in warm weather as a function of canopy density for a non- intercepting canopy. It appears as a smooth, steadily decreasing function, showing increased sensible heat flux from the canopy to the atmosphere with increasing vegetal density. The canopy gives off heat in warm weather because the canopy without snow is able to warm up to a temperature greater than the within-canopy air temperature, which is affected by the surface snowcover to some degree. The shape of the graph is the result of the superposition of two effects. First, the direct linear dependence of H_{sf} on M would cause the heat flux to the

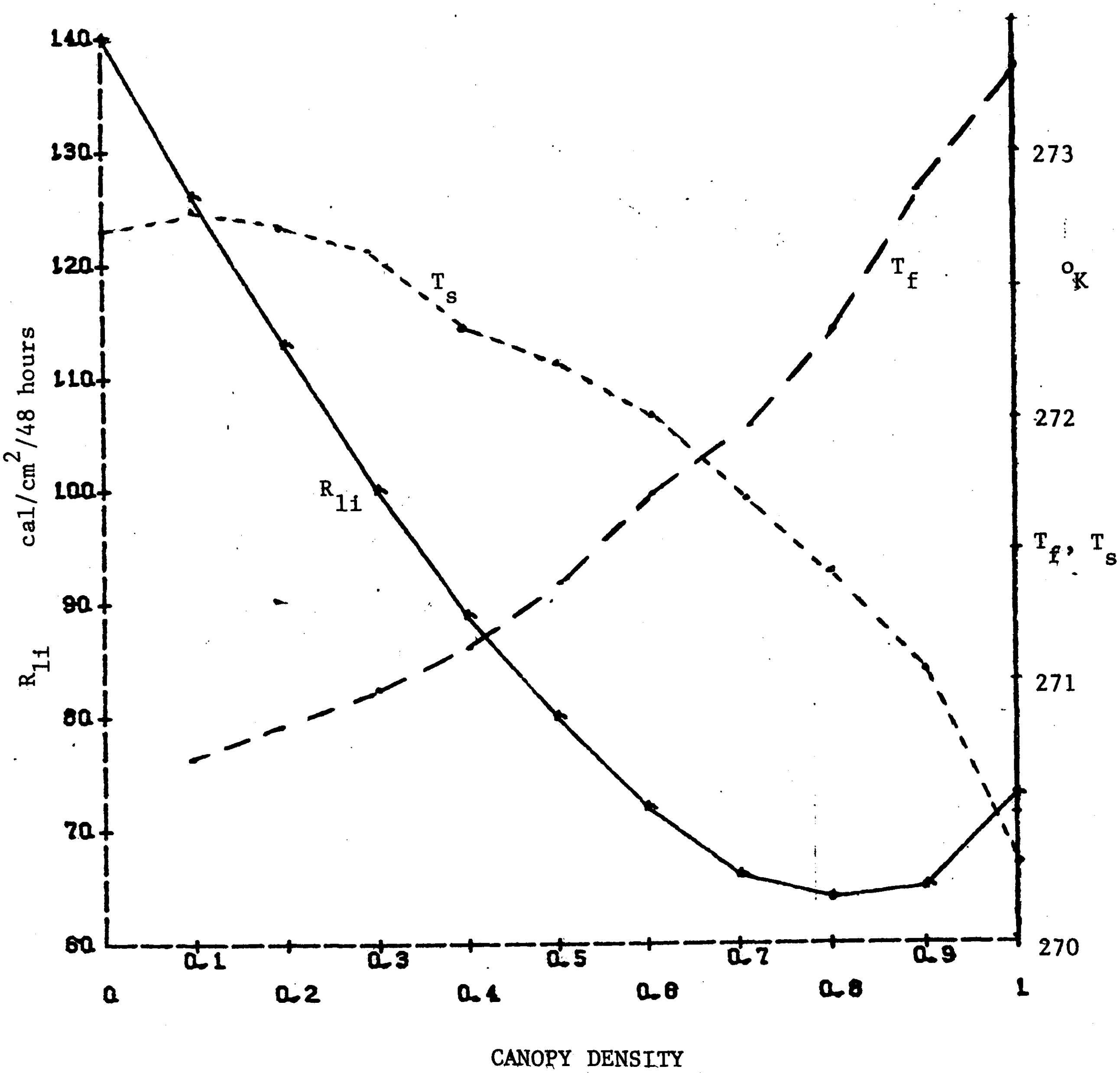


FIGURE 7: Longwave radiation flux to surface cover: case of cold weather, no interception

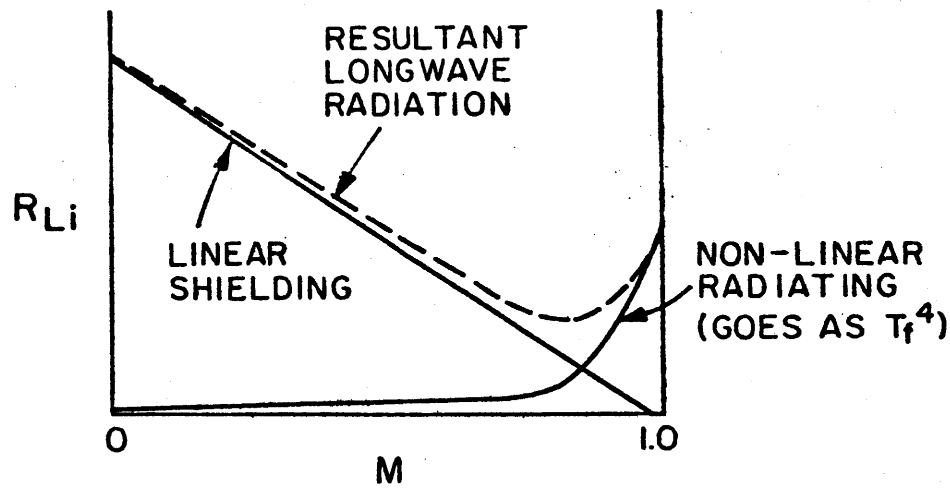


FIGURE 8 SUPERPOSITION OF CANOPY EFFECTS ON R_{Li}

vegetation to decrease linearly with M. Secondly, the value of T_{af} is decreased with increasing M because of the cooling of the surface snowcover, which is non-linear in M and most important at large M. This decrease in T_{af}, combined with a slight increase in T_f, increases the temperature gradient between the foliage surface and the ventilating air, promoting sensible heat flux from the vegetation. It must be noted that this process is in opposition to the wind attenuation process, which would decrease the heat flux from the vegetation as the attenuation coefficient is increased. For the case shown in Figure (9), where the snowless canopy may rise in temperature without bounds, the wind attenuation from U_{af} = 4 m/sec at M = 0.0 to U_{af} = 2.53 m/sec at M = 1.0 is relatively unimportant.

In Figure (10) the case of a canopy with persistent intercepted snow in a colder climate is shown. For the particular climate and vegetation characteristics shown, there are three different regimes that are dominant at different canopy densities. In this colder weather, the canopy with snow gives off heat since the snow is warmer than the air. At small canopy densities the linear dependence of H_{sf} on M is dominant. There is a slight bending upward which takes most notable effect around M = 0.8. This is due to the attenuation of the wind, which is important for the small temperature gradients from canopy snow to air in this case. Meanwhile, at the surface snowcover there is a rapid cooling at higher canopy densitiies, especially between M = 0.9 and M = 1.0. At these high canopy densities, the almost total blockage of solar radiation becomes very important. In a cold climate, this solar radition is usually the major source of heat and the largest

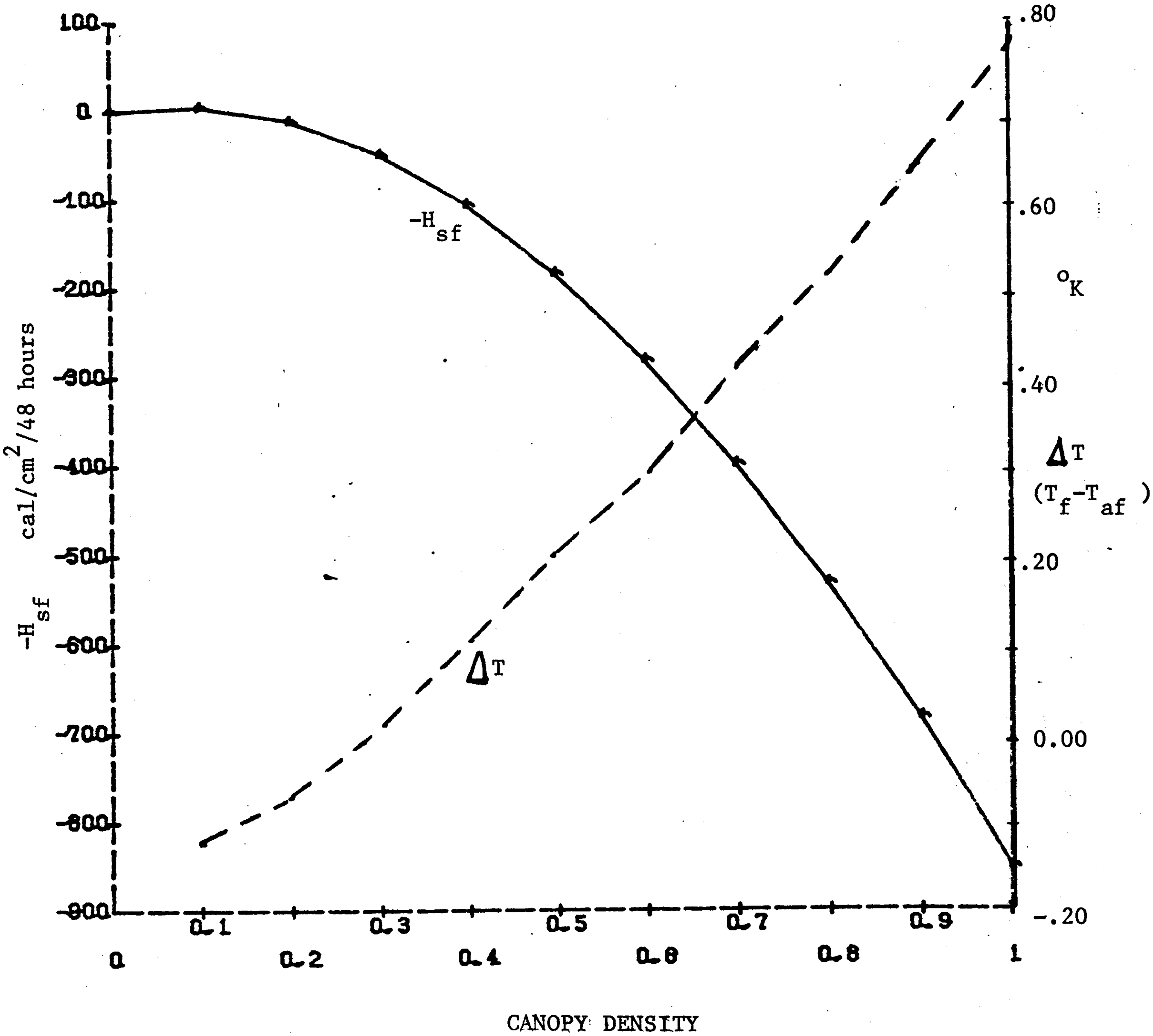


FIGURE 9: Sensible heat flux to vegetation: case of warm weather, no interception

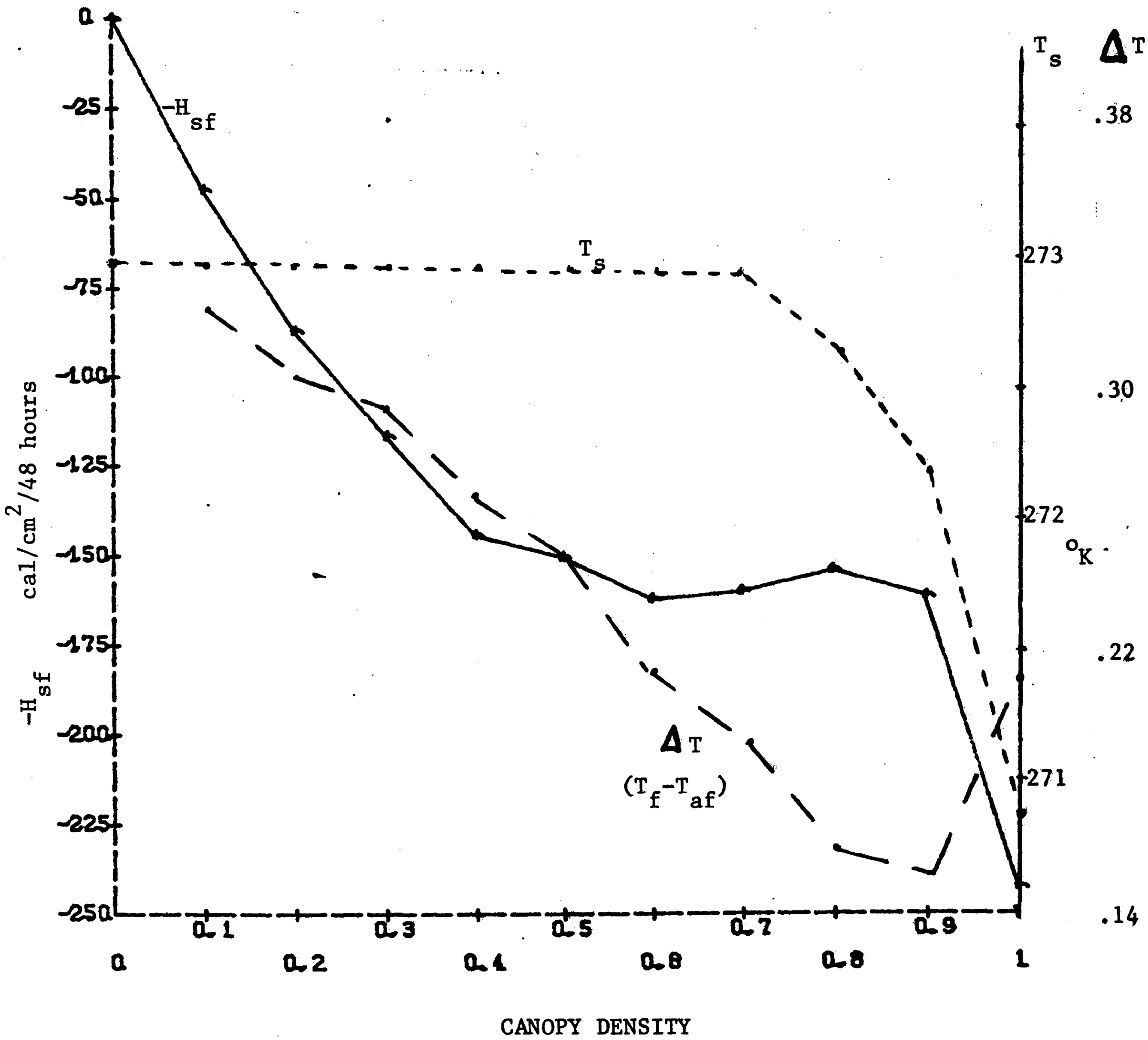
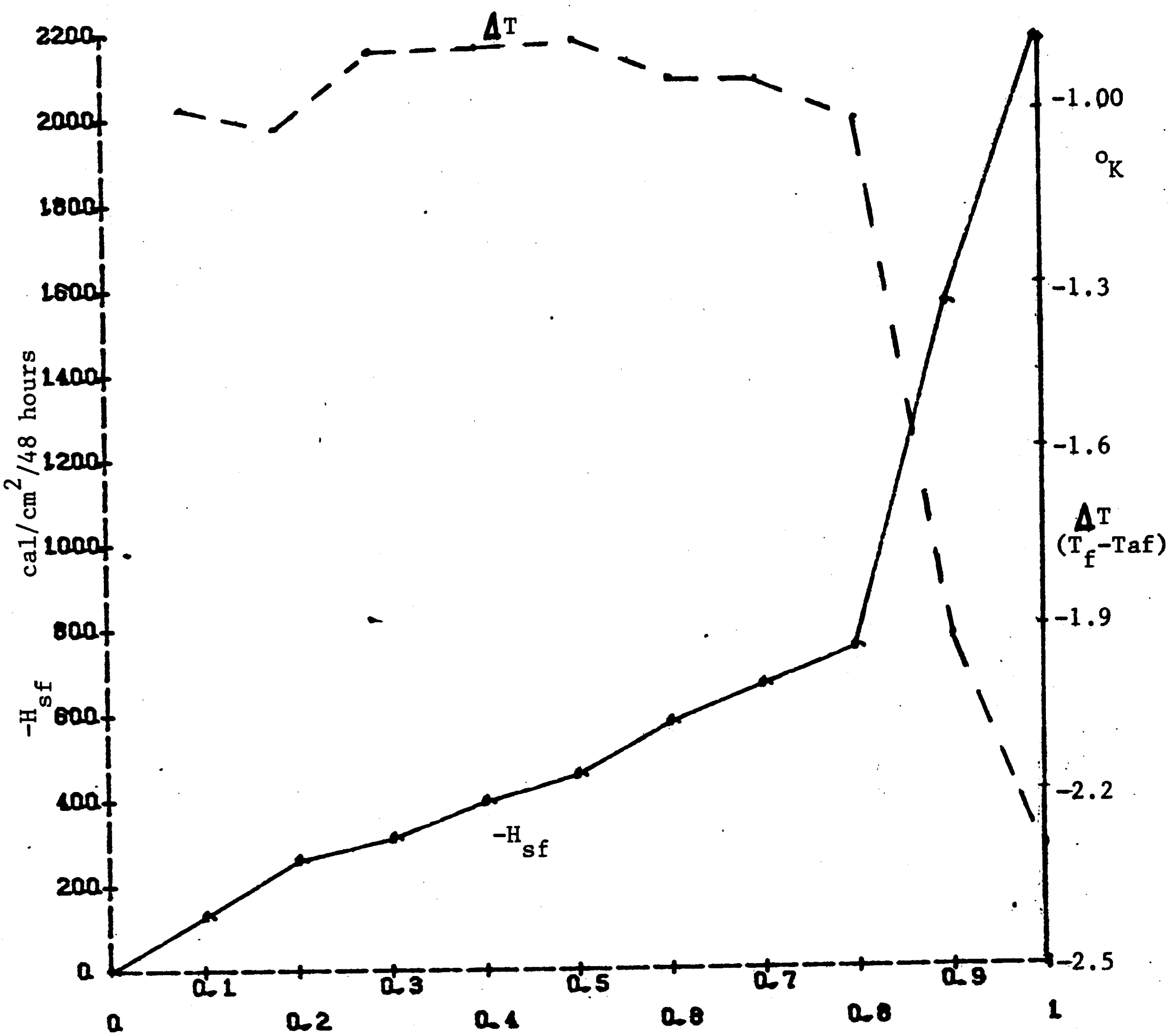


FIGURE 10: Sensible heat flux to vegetation: case of cold weather, intercepted snow, n o blow-off permitted

magnitude component of the heat balance equation. The result of this rapid cooling of the surface cover near M = 1.0 is small drop in T_{af}. Also, at high canopy densities a larger volume of snow is intercepted by the vegetation. In the single layer model, this larger volume of snow provides a larger storage volume for heat. In a cold climate, therefore, the intercepted snow layer will maintain its higher temperature for a longer period. Hence, the sensible heat flux to the vegetation is again decreased, i.e., H_{sf} is increased.

The case discussed above was very sensitive to each aspect of the process mainly because the actual transfers of sensible heat were very small. That is, the various climatic and vegetal parameters used struck a precarious balance which, when magnified, exposed the different pushes and pulls of the various phenomena taking place with full clarity. In Figure (11) we see the case of a non-blown snow-covered canopy in warm weather. In this case, the snow on the canopy absorbs heat. The shape of the plot, when examined closely, is remarkably similar to Figure (10) (inverted), though markedly less dramatic. Once again, at lower canopy densities the heat flux ($-H_{sf}$) is increased linearly, while the opposing force of wind attenuation is limiting the heat flux, and becoming more important with increasing M. However, at the larger canopy densities (M >0.8) the canopy is absorbing much more of the radiation due to its shielding and lower albedo. Also, because the higher canopy densities intercept more snow, we find that even in this warmer weather, at higher canopy densities the snow remains on the canopy longer. The presence of snow on the canopy enables the canopy to absorb more heat. At canopy densities below 0.8, in this warm weather case, the snow melted off of the canopy within the two-day period. Above M = 0.8,



CANOPY DENSITY

FIGURE 11: Sensible Heat flux to vegetation: case of warm weather, intercepted snow, no blow-off permitted

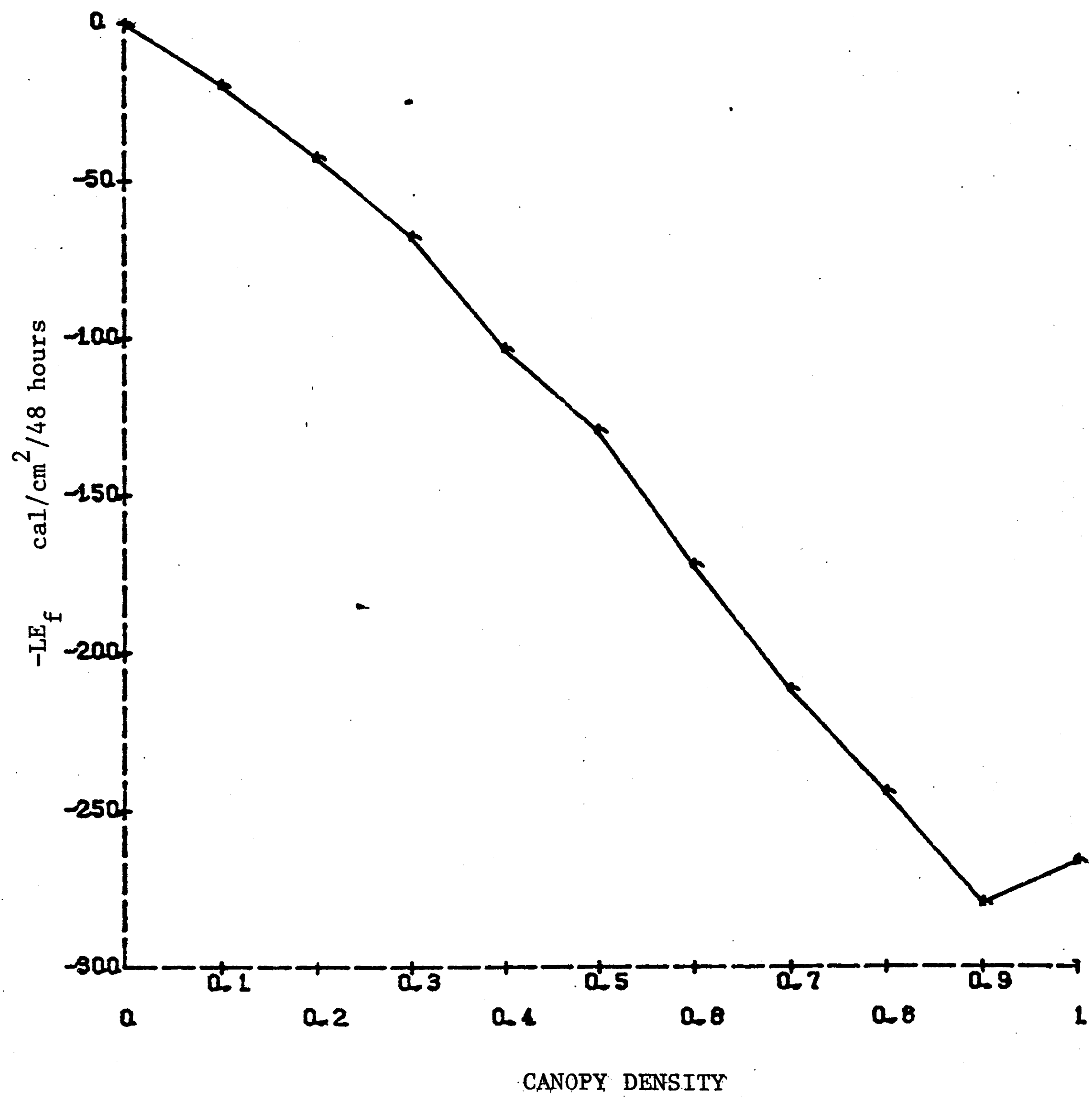


FIGURE 12: Latent heat flux to vegetation: case of cold weather, intercepted snow, no blow-off permitted

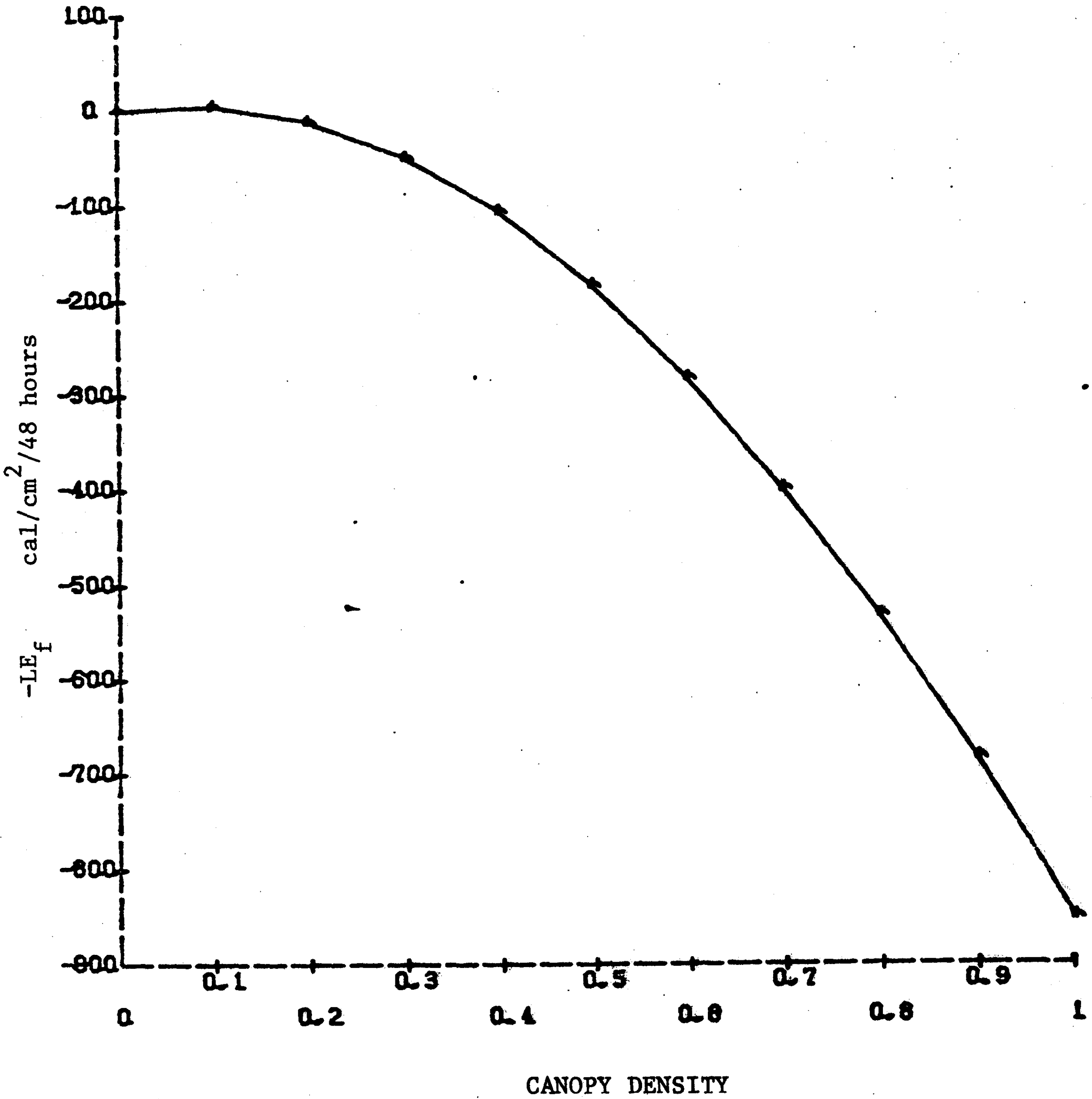


FIGURE 13: Latent heat flux to vegetation: case of warm weather, intercepted snow, no blow-off permitted

the snow remained throughout the period. Hence, we see a dramatic upturn in the canopy's intake of heat above M = 0.8.

Figure (12), corresponding to the case in Figure (10), shows the latent heat flux to the vegetation in a cold weather canopy without snow-blowing from the canopy. Here the latent heat flux ($-L_{Ef}$) decreases in a near-linear fashion with M. There is a slight reversal near M = 1.0 because of the wind attenuation effect, and the small sudden rise in the intercepted snow temperature due to radiation blockage. Figure (13) corresponds to the case shown in Figure (11). Again, we see a near-linear increase in L_{Ef} which is augmented at higher canopy densities by the greater degree of persistence of snow on the canopy. More latent heat flux from the vegetation is possible at the higher canopy densities since more snow is available for sublimation from the vegetation.

4.4 Total Heat Flux From Atmosphere

The total heat flux from the atmosphere to the unit snowcovered, vegetated area is given by the sum of all radiational, sensible, and latent heat fluxes. As shown by Figures (14), (15), (16), (17), and (18). This again is quite dependent on climate and vegetation properties.

In colder weather, with no intercepted snow present, Figure (14) shows that at a canopy density of about 0.5, the heat absorbed from the atmosphere reaches a maximum. Up to this point, the increased surface available for radiation absorption dominates the total heat transfer. Thereafter, with $M \geq 0.5$, the snowless canopy becomes an important emitter of sensible heat. The sensible heat flux from the vegetation at M = 1.0 is almost completely negating the radiation absorption at that point.

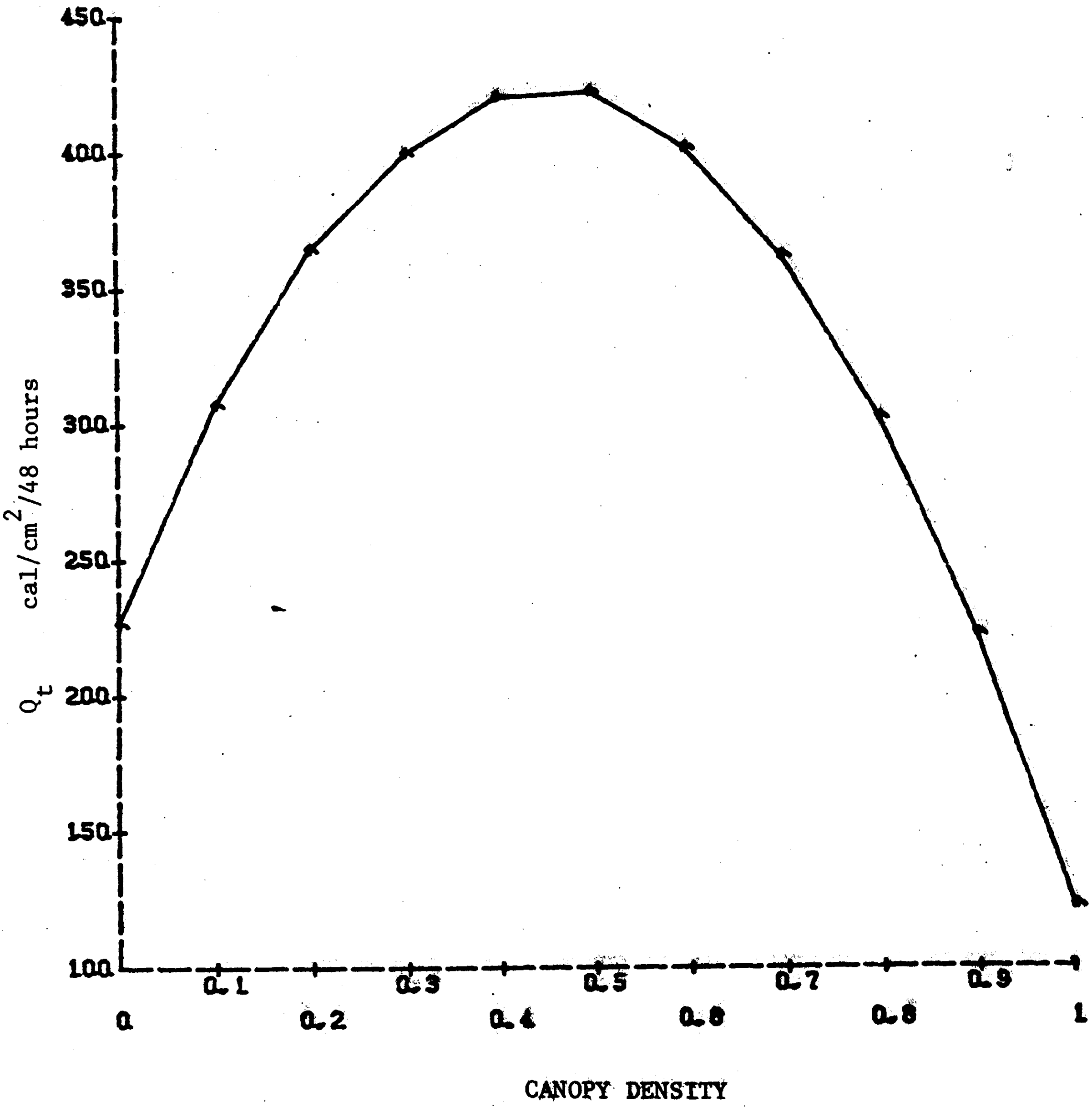


FIGURE 14: Total heat flux from atmosphere to vegetated area: case of cold weather, no interception

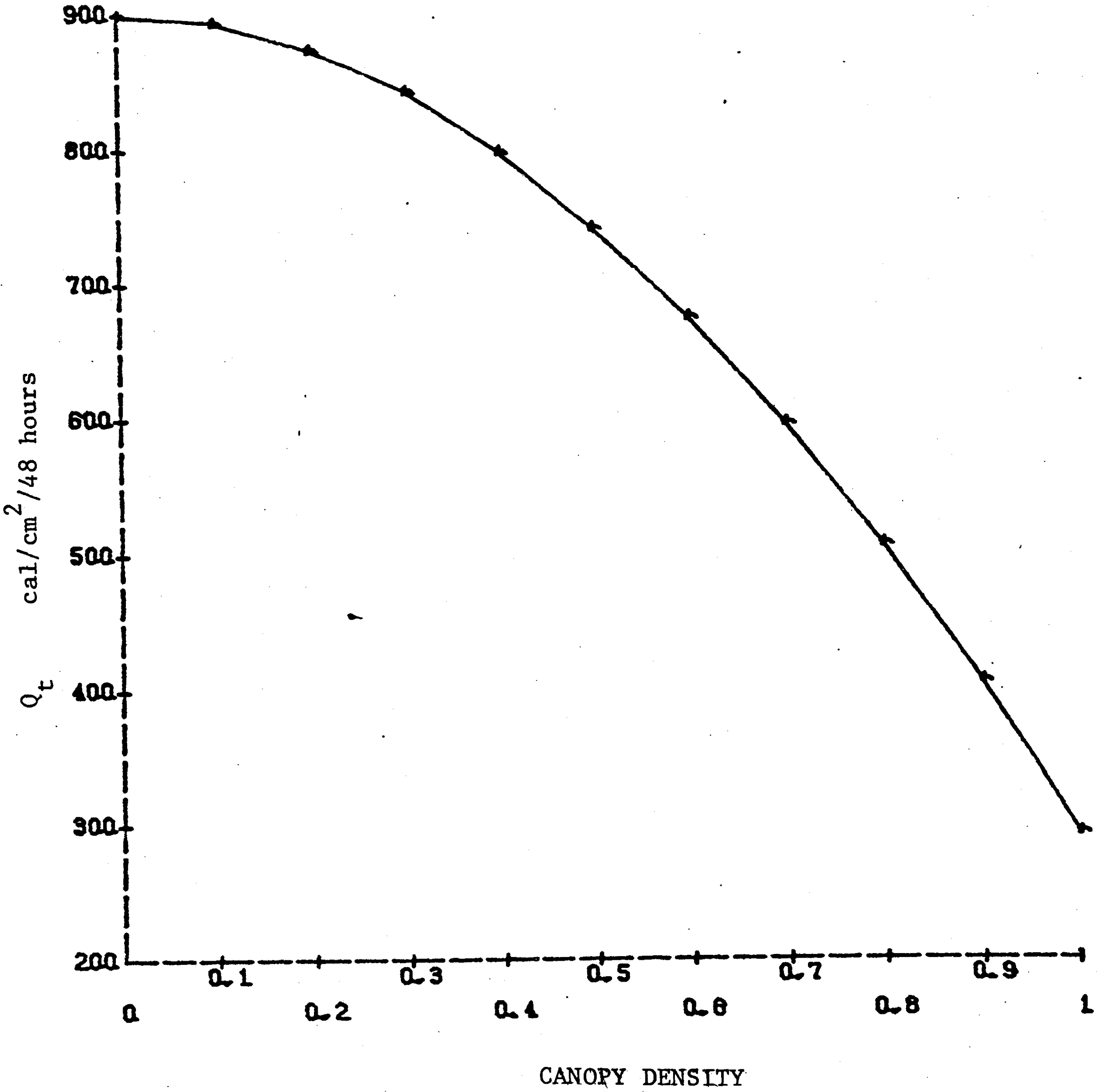


FIGURE 15: Total heat flux from atmosphere to vegetated area: case of warm weather, no interception

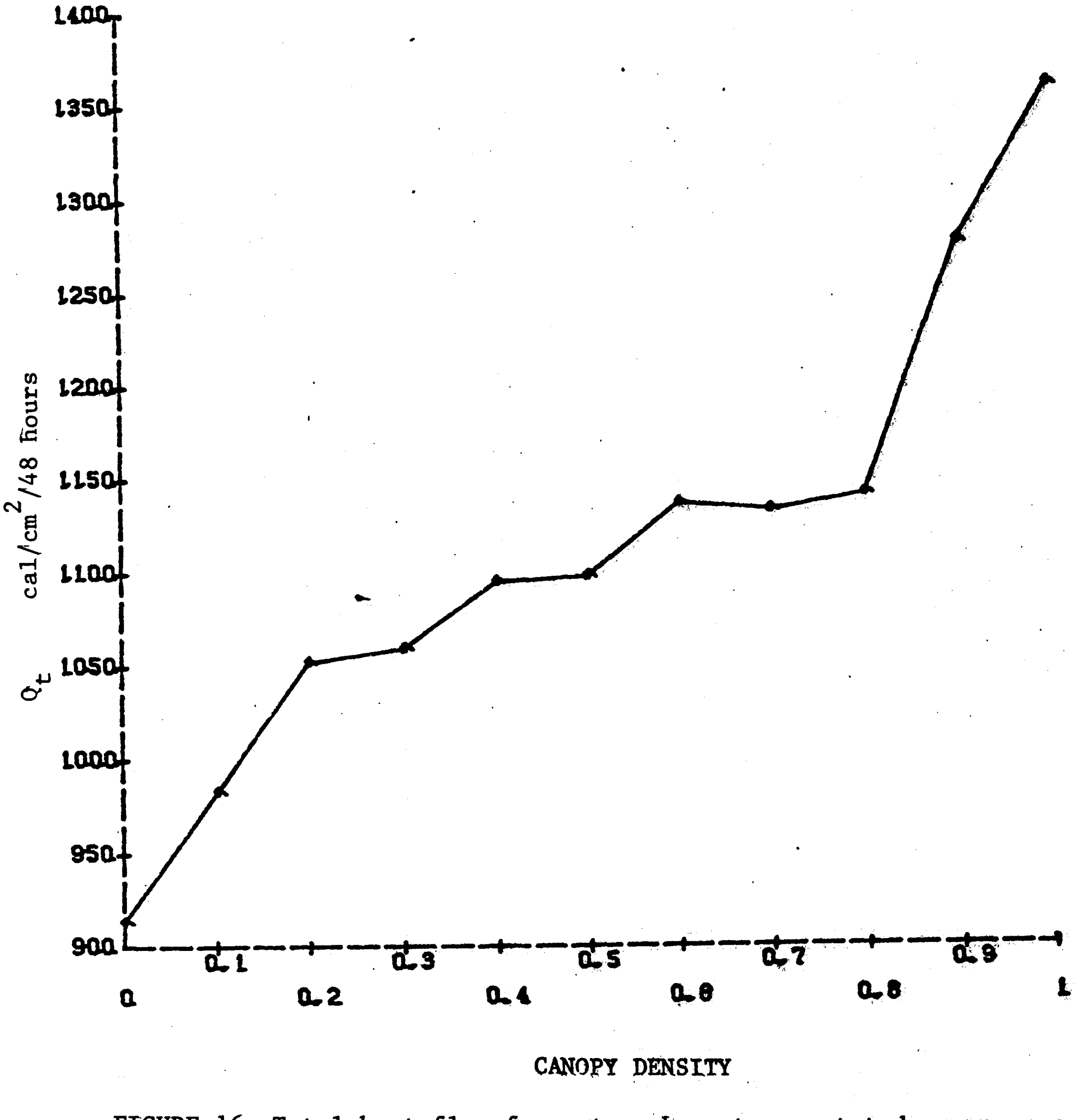


FIGURE 16: Total heat flux from atmosphere to vegetated area: case of cold weather, intercepted snow, no blow-off permitted.

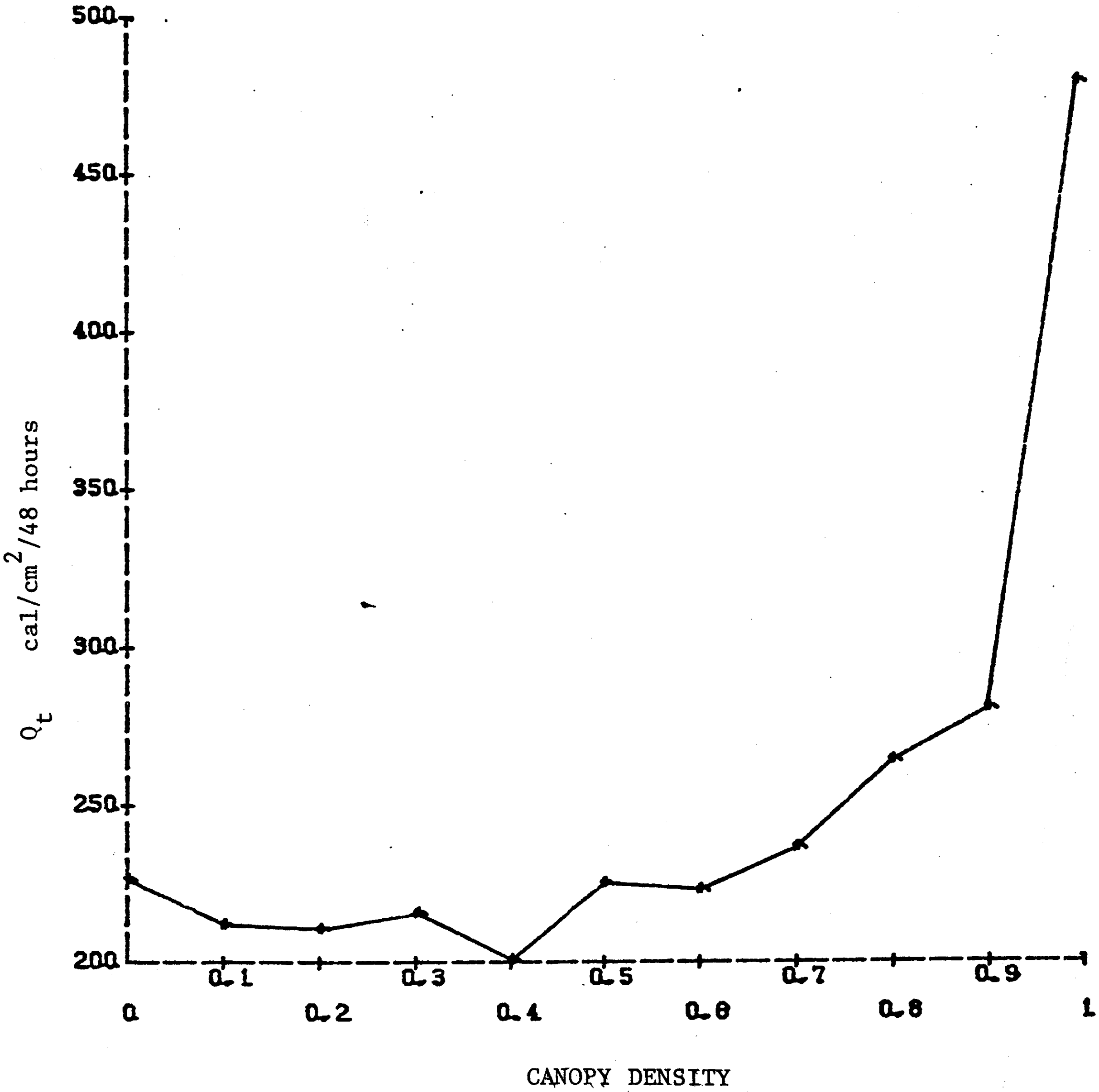


FIGURE 17: Total heat flux from atmosphere to vegetated area: Case of warm weather, intercepted snow, no blow-off permitted

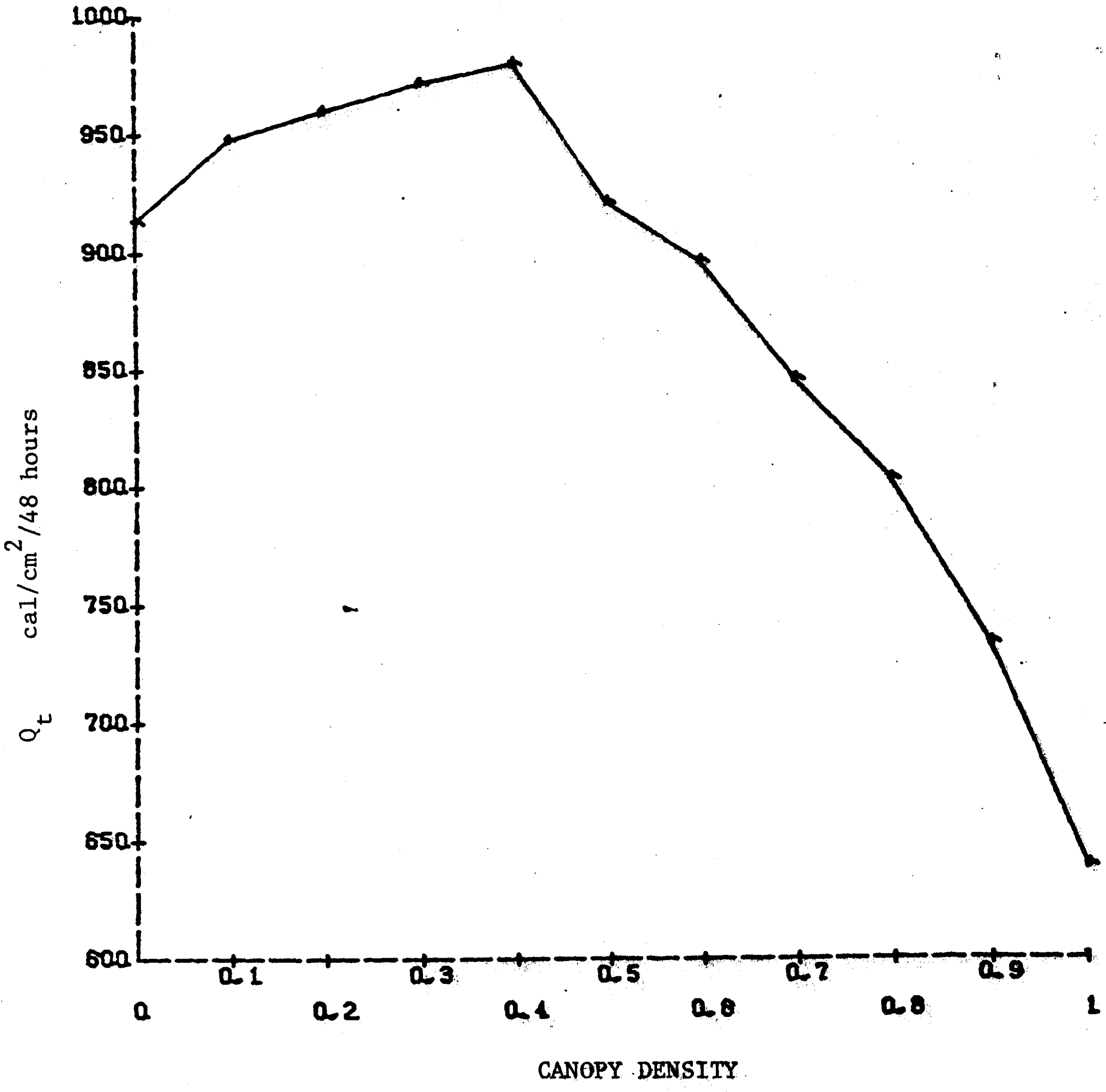


FIGURE 18: Total heat flux from atmosphere to vegetated area: Case of warm weather, intercepted snow which blows from canopy.

In Figure (15), which illustrates the warmer case of a snowless canopy, the canopy is a major emitter of heat through the entire range of canopy densities. Once again the rate of increase of sensible heat flux from the vegetation outstrips the rate of increase of radiation absorption by the vegetation with increasing canopy density.

Figure (16) illustrates the case where a persistently snow-covered canopy area is subject to cold weather. At the lower canopy densities. The turbulent transfers from the surface snowcover and the vegetation attempt to counterbalance the radiation absorption, which increases quickly with increasing canopy density. Eventually the radiation absorption dominates. In Figure (17), for the warm persistent snow case, the direction of sensible heat transfer is reversed, and flows from the atmosphere to the vegetation. Hence, the total heat flux from the atmosphere is generally always increaing, as only the latent heat flux now opposes it. At the intermediate canopy densities of around M = 0.5, the latent heat fluxes which occur from the surface snow and canopy snow to the atmosphere are maximized, and begin to compensate for the radiational and sensible heat transfers. However, the latent heat transfer from the surface cover is driven to 0.0 at increasing M, and the accompanying increase in canopy latent heat transfer does not compensate. Thus, the radiational and sensible heat transfers return to dominance.

In Figure (18), for the case of non-persistent intercepted snow on the canopy in warmer weather, the vegetation becomes increasingly important as a source of latent heat as M increases. This occurs for the reasons previously discussed concerning Figure (13). Meanwhile, above a canopy density of 0.4, the sensible heat transfer to the vegetations begins to decrease due to the growing effect of wind attenuation by the canopy. These combined cause the total heat flux from the atmosphere to decrease in the manner shown in Figure (18).

4.4 Conclusions

The presence of a vegetal canopy can substantially affect the energy and mass fluxes from a unit area to the atmosphere and soil. The same canopy might have opposite effects under different conditions, so it is difficult to make generalizations for use in predicting its effects. A canopy with intercepted snow behaves vastly different from a snowless canopy, the latter being able to emit more radiation and sensible heat. Thus the coefficients governing quantity of snow intercepted and blowing of snow from the canopy must be carefully evaluated at a given site. The coefficient which governs wind attenuation can also be very important in calculating turbulent transfers of heat and mass both from the canopy and from the surface cover.

From the test results, it may be suggested that the direct linear dependence of heat transfer to the vegetation would be dominant at low canopy densities, given a small or average wind attenuation coefficient ($k_u \simeq 10^{-7}$) or a smaller wind. Wind attenuation becomes most important with increasing canopy density due to its exponential dependence on the attenuation coefficient a(M). At the largest canopy densities, the effects of the canopy on the surface cover become most pronounced. Also, the increased depth of snow intercepted increases heat or "cold" storage. Thus there can be said to be at least three possible mass and heat flux regimes, which can be of varying importance depending on canopy properties and the weather:

1. direct linear dependence of H_{sf}, LE_f, and R_f on M
2. wind attenuation, exponential dependence on M, limiting turbulent transfers
3. increased heat or cold storage, and increased effect on surface energy balance at large M.

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LIST OF VARIABLES USED IN VEGETATION SUBROUTINES

a	=	coefficient in wind attenuation equation
blown	=	depth of snow blown from vegetation [cm]
ca	=	specific heat of air [cal/g°K]
caldq	=	combination variable for computation of vegtdt
calem	=	combination variable for computation of vegtdt
calhsf	=	combination variable for computation of vegtdt
callef	=	combination varialbe for computation of vegtdt
calqst	=	combination variable for computation of vegtdt
calrad	=	combination variable for computation of vegtdt
check	=	variable used to check residual between successive interations
checkr	=	residual between successive iterations
ctd	=	specific of ice at tst [cal/g°K]
cw	=	specific heat of water [cal/g°K]
d	=	depth of first surface snow layer [cm]
deltat	=	lenght of time step [hrs]
divdel	=	subdivided length of time step [hrs]
dlw	=	liquid interception [cm]
dlwlos	=	evaporated liquid depth [cm]
dlwmax	=	maximum intercepted water equivalent [cm]
dnc	=	number of time step subdivisions
e	=	emissivity of surface snowcover
ea	=	vapor pressure of air [mb]
fl	=	derivative of residual with respect to vegtdt
factor	=	coefficient in blown-snow equation [cal/cm^2]
fcapl	=	amount of "cold" available in surface layer to freeze liquid [cal/cm^2]

LIST OF VARIABLES USED IN VEGETATION SUBROUTINES

fposs = amount of liquid that can refreeze in intercepted snow [cm]

hsf = specific heat flux from vegetation per unit area to atmosphere [cal/cm^2]

hsf2 = specific heat flux from vegetation per unit area to atmosphere [cal/cm^2]

i = number of iterations

j = counting variable for subdividing time step

k = counting variable in subdivision do loop

lef = latent heat flux from vegetation to atmosphere per unit area

lef2 = latent heat flux from vegetation to atmosphere per unti area [cal/cm^2]

lf = latent heat of fusion [cal/g]

nc = number subdivisions of time step

nci = initial number of subdivisions of time step

ni = counting variable for iterations

nok = noknow (1) from Anderson's program

p = density of first surface cover layer [g/cm^3]

pa = atmospheric pressure [mb]

phsf = sensible heat flux from vegetation for atmosphere/unit area [cal/cm^2]

phsft = sensible heat flux from vegetation for atmosphere/unit area [cal/cm^2]

plef = latent heat flux from vegetation to atmophere per unit area [cal/cm^2]

prbl = radiative heat flux from vegetation to atmosphere per unit area [cal/cm^2]

prblt = radiative heat flux from vegetation to atmosphere per unit area [cal/cm^2]

prmlt = portion of time step before snow disappears [hrs]

pstmlt = portion of time step after snow disappears [hrs]

LIST OF VARIABLESUSED IN VEGETATION SURBROUTINES

pw = density of water [g/cm^3]

qa = longwave radiation reaching ground [cal/cm^2]

radbal = radiative heat flux from vegetation to atmosphere per unit area [cal/cm^2]

radbl2 = radiative heat flux from vegetation to atmsophere per unit area [cal/cm^2]

rainf = depth of rain that freezes [cm]

rainl = depth of rain that remains in liquid form [cm]

rdbal = radiative heat flux from vegetation to atmosphere per unit area [cal/cm^2]

resid = residual in energy calculations [cal/cm^2]

rhoa = density of air [g/cm^3]

sigma = Stefan-Boltzmann constant [cal/cm^3 hour°K^4]

sumwtl = liquid water change [cm]

ta = ventilating air temperature [°K]

tdel = change in intercepted snow temperature [°K]

tdelab = absolute value of tdel [°K]

temdlt = dummy variable for storing time step length [hrs]

temdp = dummy variable for storing snow depth [cm]

temtem = dummy variable for storing snow temp [°K]

tst = temperature of first layer of surface cover [°K]

u = wind speed [cm/sec]

vblwn = depth of snow blown from vegetation [cm]

vegalb = albedo of vegetation

vegcit = specific heat of ice at vegtem [cal/g°K]

vegctd = specific heat of ice at vegtdt [cal/g°K]

vegdel = heat added to vegetation [cal/cm^2]

vegden = density of intercepted snow [g/cm^3]

LIST OF VARIABLES USED IN VEGETATION SUBROUTINE

vegd13 = cold content of intercepted snow to refreeze liquid [cal/cm^2]

vegd14 = amount of cold content used to refreeze liquid [cal/cm^2]

vegdp = previous depth of intercepted snow [cm]

vegdpt = depth of intercepted snow [cm]

vegdpz = depth of intercepted snow [cm]

vege = emissivity of vegetation

vegea = vapor pressure above vegetation [mb]

vegead = vapor pressure above vegetation [mb]

veglos = sublimated loss from canopy [cm]

vegls = sublimated loss from canopy [cm]

veglw = previous value of liquid water content [cm]

veglwc = liquid water content of intercepted snow [cm]

vegm = canopy density

vegout = outflow from canopy [cm]

vegpns = density of new snow or rain [g/cm^3]

vegpx = depth of new snow or rain [cm]

vegpxd = depth of new snow or rain [cm]

vegq = specific humidity above vegetation

vegqa = longwave radiation above vegetation [cal/cm^2]

vegqad = longwave radiation above vegetation [cal/cm^2]

vegqi = shortwave radiation above vegetation [cal/cm^2]

vegqid = incident shortwave radiation [cal/cm^2]

vegqrd - reflected shortwave radiation [cal/cm^2]

vegta = air temperature above vegetation [°K]

vegtad = air temperature above vegetation [°K]

LIST OF VARIABLES USED IN VEGETATION SUBROUTINE

vegtdt	=	temperature of canopy or intercepted snow [°K]
vegtem	=	previous temperature of vegetation [°K]
vegtm	=	previous temperature of vegetation [°K]
vegtw	=	temperature of precipitation [°K]
vegu	=	wind speed [cm/sec]
vegua	=	wind speed [cm/sec]
vegwtl	=	liquid water in snow [cm]
vegxe	=	drop in termperature of intercepted snow [°K]
veqsat	=	saturated specific humidity at vegetation surface
vgden	=	density of intercepted snow [g/cm^3]
vgdela	=	change in heat storage of intercepted snow [cal/cm^2]
vgdpf	=	depth of intercepted snow [cm]
vghsf	=	sensible heat flux form vegetation to atmosphere [cal/cm^2]
vglef	=	latent heat flux from vegetation to atmosphere [cal/cm^2]
vglwc	=	liquid water content of int'd. snow [cm]
vgout	=	outflow from intercepted snow [cm]
vgtdt	=	temperature of intercepted snow [°K]
vnwsn	=	depth of precipitation [cm]
vtsno	=	temperature of precipitation [°K]
wadd	=	dummy variable for liquid water content [cm]
wst	=	liquid water content of first surface cover layer [cm]

APPENDIX I: LISTING OF VEGETATION SUBROUTINES

```
      subroutine vbgin(vegm,vegdpt,vegden,vegtem,veglwc,a,factor,
     &     dlwmax,dlw)
      read(21,3)vegm,vegdpt,vegden,vegtem,veglwc,a,factor,dlwmax
 3    format(f3.2,f5.1,f3.2,f6.2,f6.4,f6.2,f9.7,f6.0)
      dlw=0.0
      if (vegtem.le.273.16)go to 46
      write(6,7)
 7    format(1x,'snow temp on vegetation cannot exceed 0 C')
      stop 7
46    if(vegtem.lt.273.16.and.veglwc.gt.0.0)go to 9
      return
 9    write(6,8)
 8    format(1x,'there can be no liquid in pack below 0 c')
      stop 8
      end
```

```
      subroutine vegvar(vegtad,vegead,vegua,vegqid,vegqrd,vegqad,
     &vegpxd,last,a,factor,vegm)
      common/hrdata/tg(744),ts(744),denns(744),tad(744),ead(744),
     &ua(744),qid(744),qrd(744),qad(744),pod(744),pxd(744)
      dimension vegtad(744),vegead(744),vegua(744),vegqid(744),
     &vegqad(744),vegqrd(744),vegpxd(744)
      lmhr=last*24
      do 7 i=1,lmhr
      vegtad(i)=tad(i)
      vegead(i)=ead(i)
      vegua(i)=ua(i)
      vegqid(i)=qid(i)
      vegqad(i)=qad(i)
      vegpxd(i)=(.37*vegm)*pxd(i)
      ua(i)=(vegua(i)/(a*vegm))*(1.-exp(-a*vegm))
      qid(i)=(1-vegm)*qid(i)
      qrd(i)=(1-vegm)*qrd(i)
      pxd(i)=pxd(i)-vegpxd(i)
7     continue
      return
      end
```

```
      subroutine vegobt(vegta,vegea,vegu,vegqi,vegqr,u,vegqa,rcf,scf,vegpns,vtsno,
     &vegpx,vegtad,vegead,vegua,vegqid,vegqrd,vegqad,vegpxd,iday,ihour,deltat,vegtw,
&     vegdpt,dlw,dlwmax,vegden)
      dimension vegtad(744),vegead(744),vegua(744),vegqid(744),vegqrd(744),
     &vegqad(744),vegpxd(744)
      common/hrdata/tg(744),ts(744),denns(744),tad(744),ead(744),
&     ua(744),qid(744),qrd(744),qad(744),pod(744),pxd(744)
      vegta=0.0
      vegea=0.0
      u=0.0
      vegu=0.0
      vegqi=0.0
      vegqr=0.0
      vegqa=0.0
      vegpx=0.0
      mhr2=(iday-1)*24+ihour
      idt=deltat+0.01
      mhr1=mhr2-idt+1
      lwe=dlw
      if(vegdpt.gt.0.0)lwe=vegdpt*vegden
      do 111 mhr=mhr1,mhr2
      vegta=vegta+vegtad(mhr)
      vegea=vegea+vegead(mhr)
      u=u+ua(mhr)
      vegu=vegu+vegua(mhr)
      vegqi=vegqi+vegqid(mhr)
      vegqr=vegqr+vegqrd(mhr)
      vegqa=vegqa+vegqad(mhr)
      if(lwe.lt.dlwmax)go to 2
      pxd(mhr)=vegpxd(mhr)+pxd(mhr)
      vegpxd(mhr)=0.0
 2    continue
      lwe=lwe+vegpxd(mhr)
      vegpx=vegpx+vegpxd(mhr)
 111  continue
      vegta=vegta/deltat
      vegea=vegea/deltat
      vegu=vegu/deltat
      vegpx=vegpx*.1
      if(vegpx.le.0.0)go to 673
      write(6,674)vegpx
 674  format(1x,'vegpx=',f6.2)
 673  continue
c*************************************************************************
c     DETERMINE DENSITY OF NEW SNOW -- FINAL VALUES.
c             NEGATIVE IF NO PRECIPITATION.
c             =1.0 IF RAIN.
c             .GT.0.0.AND.LT.0.90 IF SNOW.
      vegpns=-1.0
      vtsno=0.0
      if (vegpx.eq.0.0) return
c     COMPUTE PERIOD WET-BULB TEMPERATURE.
      call wtbulb(vegtw,vegta,vegea,pa)
c     CHECK FOR FROZEN PRECIPITATION.
      ice=0
      if (vegtw.le.274.16) ice=1
c     CONVERT FORM IF INPUT VALUES FOR ALL HOURS ARE REVERSED.
      hours=0.1
      do 120 mhr=mhr1,mhr2
```

```
         if (ice.eq.1) go to 121
         if ((denns(mhr).gt.0.0).and.(denns(mhr).ne.1.0)) hours=hours+1.0
         go to 120
  121    if (denns(mhr).eq.1.0) hours=hours+1.0
  120    continue
         if (hours.lt.deltat) go to 125
c        PRECIPIATION FORM IS OPPOSITE TO THAT INDICATED BY THE WET-BULB.
         if (ice.eq.0) index=1
         if (ice.eq.1) index=0
         ice=index
  125    if (ice.eq.1) go to 130
c***********************************************************************
c        PRECIPITATION IS RAIN
         vegpns=1.0
         vtsno=0.0
         vegpx=rcf*vegpx
         return
c***********************************************************************
c        PRECIPITATION IS FROZEN -- COMPUTE DENSITY
  130    vegpns=0.0
         do 131 mhr=mhr1,mhr2
         den=denns(mhr)
         if ((den.gt.0.0).and.(den.lt.0.90)) go to 133
c        COMPUTE DENSITY OF NEW SNOW BASED ON WET BULB TEMPERATURE.
c                ALTA RELATIONSHIP.
         call wtbulb(vegtw,vegtad(mhr),vegead(mhr),pa)
         if (vegtw.le.258.16) go to 132
         den=0.05+0.0017*((vegtw-258.16)**1.5)
         go to 133
  132    den=0.05
  133    vegpns=vegpns+den*vegpxd(mhr)*0.1
         if (vegtw.gt.273.16)vegtw=273.16
         vtsno=vtsno+vegtw*vegpxd(mhr)*0.1
  131    continue
         vegpns=vegpns/vegpx
         vtsno=vtsno/vegpx
         vegpx=scf*vegpx
c***********************************************************************
         return
         end
```

```
      subroutine vegbal(vegtdt,veglwc,vegdpt,vegden,qa,aq,ea,u,hsf,lef,vegu,vegta,
     & radbal,vegtem,vegout,plwhc,tst,alb,vegm,vegea,ta,pa,vegqa,vegqi,deltat,
&     vegdp,vtsno,vegpns,vegpx,vegtw,cw,wst,d,p,blown,veglos,nok,factor,dlw)
c     VEGETATION ENERGY BALANCE- ONE TIME STEP- TEST PROGRAM
c     SOLVES FOR EITHER TEMPERATURE OR LIQUID WATER CONTENT GIVEN MET VARIABLES
      dimension check(35)
      real lef,lf,lef2
      ni=1
      check(ni)=0.0
      hsf=0.0
      lef=0.0
      veglos=0.0
      pxf=0.0
      vegq=vegea*.622/pa
      rainl=0.0
      rainf=0.0
      radbal=0.0
      vegdpt=vegdp
      veglw=veglwc
      if(vegdp.le.0.0)vegalb=0.2
      sigma=60.*.8258*(10.**-10.0)
      i=1
      j=0
      lf=79.7
      cw=1.0
      pw=1.000
      rhoa=.00122
      ca=.24
      prbl=0.0
      plef=0.0
      phsf=0.0
      temdlt=deltat
c     convert wind speed to cm/hr
      u=360000.*u+.0001
      vegu=360000.*vegu +.0001
      vege=.97
      e=.97
      if(vegpx.eq.0.0)go to 50
      if(vegpns.eq.1.0)go to 50
c     add new snowfall to intercepted snow depth and average temperature
      vnwsn=vegpx/vegden
      vegdp=vegdp+vnwsn
      vegtem=((vnwsn/vegdp)*vtsno)+((1.-(vnwsn/vegdp))*vegtem)
      vegpx=0.0
 50   divdel=deltat
      dnc=1.0
      nci=1
      nc=1
      temtem=vegtem
      sumwtl=0.0
c     add rainfall to liquid depth on vegetation
      if(vegpx.gt.0.0)dlw=vegpx+dlw
      if(vegdp.le.0.05)go to 38
      dlw=0.0
 14   temdp=vegdp
      temtem=vegtem
      sumwtl=0.0
      vegwtl=0.0
c*******************************************************************************
c     DO LOOP FOR TIME STEP SUBDIVISION
```

```
C******************************************************************************
         do 9 k=1,nc
         veqsat=(.622/pa)*(3.56*(10.**10.))*exp(-6141.9/vegtem)
         if(vegdp.le.0.05)veqsat=vegq
         vegalb=0.6
         if (vegdp.le.0.0)vegalb=0.2
         vegcit=.0222+(.00176*vegtem)
         vegpxd=vegpx/dnc
         if(vegdp.eq.0.0.and.j.lt.9)go to 9
         if(vegdp.eq.0.0.and.j.ge.9)go to 75
         go to 76
c        SNOW HAS DISAPPEARED DURING TIME STEP
 75      vegtem=temtem
         temtem=273.16
         temdlt=deltat
         pstmlt=(dnc-k)/dnc
         if(pstmlt.le.0.0)go to 76
         prmlt=k/dnc
         deltat=pstmlt*deltat
         divdel=prmlt*temdlt
         veqsat=(.622/pa)*(3.56*(10.**10.))*exp(-6141.9/vegtem)
         plef=(temdp-(sumwtl/vegden))*vegden*677.
         pxf=(vegtw-273.16)*cw*pw*vegpxd*k
         phsf=divdel*(7.7*vegm*rhoa*ca*.01*(1.+(1080./u))*u*(vegtem-ta))
         prbl=(vegm*(((1.-vegalb)*vegqi)+(vege*vegqa)+
        & (((vege*e)/(vege+e-vege*e))*(sigma*tst**4.))
         &-((vege*(vege+2.*e-vege*e)/(vege+e-(vege*e)))*sigma
         &*vegtem**4)))*divdel
         veqsat=(.622/pa)*(3.56*(10.**10.))*exp(-6141.9/temtem)
         phsft=divdel*(7.7*vegm*rhoa*ca*.01*(1.+(1080./u))*u*(temtem-ta))
         prblt=(vegm*(((1.-vegalb)*vegqi)+(vege*vegqa)+
        & (((vege*e)/(vege+e-vege*e))*(sigma*tst**4.))
         &-((vege*(vege+2.*e-vege*e)/(vege+e-(vege*e)))*sigma
         &*temtem**4)))*divdel
         phsf=(phsf+phsft)/2.
         prbl=(prbl+prblt)/2.
         go to 38
c        FIND LATENT HEAT TRANSFER USING PREVIOUS TIME STEP
c        VALUE FOR VEGETATION
 76      lef=divdel*(677.*7.0*vegm*rhoa*.01*(1.+(1080./u))*u*(veqsat-vegq))
         veglos=(lef/677.)/vegden
         vegdpt=vegdp-veglos
         if(vegdpt.gt.0.0) go to 41
c        ALL SNOW HAS EVAPORATED
         lef=(vegdpt*vegden*677.)
         vegdpt=0.0
         veglos=(1./vegden)*(lef/677.)
         go to 38
 41      continue
c        INTERCEPTEDSNOW IS STILL PRESENT.
c        CALCULATE SENSIBLE AND RADIATIVE HEAT TRANSFER FROM PREVIOUS
c        TIME STEPS VALUES.
         hsf=divdel*(7.7*vegm*rhoa*ca*.01*(1.+(1080./u))*u*(vegtem-ta))
         pxf=(vegtw-273.16)*cw*pw*vegpxd
         radbal=(vegm*(((1.-vegalb)*vegqi)+(vege*vegqa)+
        & (((vege*e)/(vege+e-vege*e))*(sigma*tst**4.))
         &-((vege*(vege+2.*e-vege*e)/(vege+e-(vege*e)))*sigma
         &*vegtem**4)))*divdel
         vegdel=radbal-hsf-lef+pxf
         vegtm=vegtem
```

```
      vegtdt=vegtm+(vegdel/(vegcit*vegdp*vegden))
c**************************************************************************************
c     FIND TEMPERATURE OF SNOW BY ITERATION
 7    vegctd=.0222+.00176*vegtdt
      vgdel2=(vegdp*vegden*vegcit*vegtem-vegdpt*vegden*
      &vegtdt*vegctd)-(veglos*677.*vegden)
      resid=vgdel2+vegdel
      if(resid.gt.-0.009.and.resid.lt.0.009)go to 11
c     NEWTON-RAPHSON ITERATION: T=T(guess)  -  f(T(guess))/f'(T(guess))
      f1=(-.0222*vegdpt*vegden)+(-2.*vegdpt*vegden*.00176*
&     vegtdt)
      vegtdt=  vegtdt-(resid/f1)
      i=i+1
      if(i.gt.60)go to 10
      go to 7
c*********************************************************************************
 17   vegtem=vegtdt
      vegdp=vegdpt
      i=0
 9    continue
c*************************************************************************************
c     END OF DO LOOP
c************************************************************************************
      i=0
      vegdp=temdp
      vegtem=temtem
      veqsat=(.622/pa)*(3.56*(10.**10.))*exp(-6141.9/vegtem)
      lef=deltat*(677.*7.0*vegm*rhoa*.01*(1.+(1080./u))*u*(veqsat-vegq))
      pxf=(vegtw-273.16)*cw*pw*vegpx
      hsf=deltat*(7.7*vegm*rhoa*ca*.01*(1.+(1080./u))*u*(vegtem-ta))
      radbal=(vegm*(((1.-vegalb)*vegqi)+(vege*vegqa)+
     & (((vege*e)/(vege+e-vege*e))*(sigma*tst**4.))
      &-((vege*(vege+2.*e-vege*e)/(vege+e-(vege*e)))*sigma
      &*vegtem**4)))*deltat
      veqsat=(.622/pa)*(3.56*(10.**10.))*exp(-6141.9/vegtdt)
      lef2=deltat*(677.*7.0*vegm*rhoa*.01*(1.+(1080./u))*u*(veqsat-vegq))
      hsf2=deltat*(7.7*vegm*rhoa*ca*.01*(1.+(1080./u))*u*(vegtdt-ta))
      radbl2=(vegm*(((1.-vegalb)*vegqi)+(vege*vegqa)+
     & (((vege*e)/(vege+e-vege*e))*(sigma*tst**4.))
      &-((vege*(vege+2.*e-vege*e)/(vege+e-(vege*e)))*sigma
      &*vegtdt**4)))*deltat
      lef=(lef+lef2)/2.
      hsf=(hsf+hsf2)/2.
      radbal=(radbl2+radbal)/2.
      veglos=lef/(677.*vegden)
      tdel=vegtdt-vegtem
      ni=ni+1
      check(ni)=vegtdt
      checkr=abs(check(ni)-check(ni-1))
      tdelab=abs(tdel)
c     MUST SUBDIVIDE TIME STEP WHEN INTERCEPTED SNOW HAS
c     JUST DISAPPEARED. IF THIS IS STILL THE CASE AFTER AT
c     LEAST TEN SUBDIVISIONS, IT WILL BE ACCEPTED AS FACT.
      j=j+1
      if(j.gt.10)go to 70
      if(j.eq.1)go to 83
      if(vegdpt.le.1.0)go to 65
 83   if(tdelab.lt.0.10.and.checkr.lt.0.002)go to 12
      if(checkr.lt.0.001)go to 12
      if(j.lt.10.and.j.gt.1)go to 65
```

```
69          nc=int(tdelab/0.25)
            if(nc.gt.200)nc=200
            if(nc.lt.1)nc=1
            nci=nc
            go to 66
65          nc=nci*j
66          dnc=float(nc)
            if(dnc.lt.1.0)dnc=1.0
            divdel=deltat/dnc
            temtem=vegtem
            if(j.lt.10)go to 14
70          if(checkr.lt.0.001)go to 12
            nc=nc+10
            if(nc.lt.2500)go to 14
            write(6,64)
64          format(1x,'WARNING WARNING IN VEGETATION ENERGY BALANCE')
            write(6,911) checkr, vegdpt, vegtdt
911         format(1x,'RESIDUAL IN SUCCESSIVE DETERMINATIONS=', f6.5,/,1x,
&           'VEGDPT=',f7.2,5x,'VEGTDT=',f6.2)
            go to 12
c*******************************************************************************
11          rainl=rainl+vegpxd
            if(vegtdt.ge.273.16)go to 29
            veglwc=veglwc+vegpxd
            if(veglwc.gt.0.0)go to 39
            vegwtl=0.0
            go to 40
c           A SOLUTION FOR INTERCEPTED SNOW TEMP HAS NOT BEEN FOUND
10          j=j+1
            i=0
            nci=nci*2
            if(j.lt.10)go to 65
            write(6,19)
19          format(1x,'INTERCEPTED SNOW:SOLUTION DID NOT CONVERGE')
            go to 12
c           SOLVE FOR LIQUID WATER CONTENT IN CASE WHERE
c           INTERCEPTED SNOW HAS COOLED OR RAIN HAS OCCURRED
39          if(vegpxd.eq.0.0)go to 44
c           LIQUID WATER IS PRESENT IN OUR SUBFREEZING PACK.
c           SOME OF IT MUST FREEZE UNTIL TEMP IS RAISED TO  O C.
            vegdl3=(273.16-vegtdt)*vegctd*vegden*vegdpt
            fposs=vegdl3/(lf*pw)
            if(fposs.lt.vegpxd)go to 45
c           ALL RAIN IS FROZEN
            vegdl4=vegpxd*lf*pw
            vegtdt=vegtdt+(vegdl4/(vegctd*vegden*vegdpt))
            vegdpt=vegdpt+((pw*vegpxd)/vegden)
            rainf=rainf+vegpxd
            rainl=rainl-vegpxd
            veglwc=veglwc-vegpxd
            if(vegtem.ge.vegtdt)go to 44
            go to 42
c           ONLY SOME OF RAIN HAS FROZEN.
45          vegtdt=273.16
            vegdpt=vegdpt+((pw*fposs)/vegden)
            rainf=rainf+fposs
            rainl=rainl-fposs
            veglwc=veglwc-fposs
            go to 42
44          vegxe=vegtem-vegtdt
```

```
      vegwtl=-(((vegdpt*vegden*vegctd)/(lf*pw))*vegxe)
      vegldt=veglwc+vegwtl
      if(vegldt.ge.0.0)go to 37
c     ALL LIQUID HAS REFROZEN; ADDITIONAL HEAT LOSS
c     DECREASES SNOW TEMPERATURE
      if(-(vegwtl).le.rainl)go to 91
      rainl=0.0
      rainf=rainf+rainl
      go to 92
 91   rainl=rainl+vegwtl
      rainf=rainf-vegwtl
 92   continue
      vegdel=vegldt*lf*pw
      vegwtl=-(veglwc)
      sumwtl=sumwtl+vegwtl
      veglwc=0.0
      vegtdt=273.16
      i=0
      go to 7
c     SOME OF LIQUID WATER HASREFROZEN; IE., TEMPERATURE
c     IS STILL 273.16 K.
 37   vegtdt=273.16
      if(-(vegwtl).le.rainl)go to 93
      rainl=0.0
      rainf=rainf+rainl
      go to 94
 93   rainl=rainl + vegwtl
      rainf=rainf-vegwtl
 94   continue
      veglwc=vegldt
      go to 40
c     TEMPERATURE IS 0 C OR HEAT HAS BEEN ADDED TO 0 C PACK
 29   if(vegtdt.le.vegtem)go to 31
      if(vegtem.lt.273.16.and.vegtdt.ge.273.16)go to 30
c     ALL HEAT ADDED SERVES TO MELT SNOW
 31   vegdel=radbal-hsf-lef+pxf
      vegwtl=(vegdel/(lf*pw))
      vegldt=veglwc
      veglwc=veglwc+vegwtl
      if(veglwc.le.(vegdpt*vegden)+(vegpxd*k)+vegldt)go to 73
      rainf=0.0
      rainl=rainl+rainf
      veglwc=vegdpt*vegden+(vegpxd*k)+vegldt
      vegtdt=273.16
      vegwtl=veglwc-vegldt
      vegdpt=0.0
      go to 40
 73   vegtdt=273.16
      if(vegwtl.le.rainf)go to 95
      rainf=0.0
      rainl=rainl + rainf
      go to 96
 95   rainf=rainf-vegwtl
      rainl=rainl+vegwtl
 96   continue
      if(veglwc.ge.0.0)go to 40
      rainl=0.0
      rainf=rainf+rainl
      vegdel=veglwc*pw*lf
      veglwc=0.0
```

```
          go to 7
c         SOME of HEAT ADDED RAISES SNOW TEMPERATURE; THE REST
c         MELTS SNOW.
 30       vegex=vegtdt-273.16
          vegwtl=(vegdpt*vegden*vegctd/(lf*pw))*vegex
          if(vegwtl.le.rainf)go to 97
          rainf=0.0
          rainl=rainf+rainl
          go to 98
 97       rainf=rainf-vegwtl
          rainl=rainl+vegwtl
 98       continue
          vegldt=veglwc
          veglwc=vegwtl
          vegtdt=273.16
          if(veglwc.le.(vegdpt*vegden)+(vegpxd*k)+vegldt)go to 40
          veglwc=vegdpt*vegden+(vegpxd*k)+vegldt
          sumwtl=veglwc
          vegwtl=0.0
          vegdpt=0.0
          go to 40
 40       continue
          sumwtl=vegwtl+sumwtl
          if(k.lt.nc)go to 17
c         DETERMINE HOW MUCH SNOW REMAINS
          vegmlt=(veglwc-veglw-rainl)/vegden
          if(vegdpt.eq.0.0)go to 74
          vegdpt=vegdpt-vegmlt
          if(vegdpt.gt.0.0)go to 42
c         IF ALL SNOW HAS MELTED, IT aLL BECOMES OUTFLOW.
 74       vegdpt=0.0
          vegout=veglwc
          veglwc=0.0
          go to 17
c         LIQUID WATER INEXCESS OF HOLDING CAPACITY FLOWS OUT
 42       vegwf=(pw*veglwc)/(vegdpt*vegden)
          if(vegwf.gt.plwhc)go to 60
          vegout=0.0
          go to 17
 60       vegout=((vegwf-plwhc)*(vegdpt*vegden))/pw
          veglwc=plwhc*vegdpt*vegden/pw
          go to 17
c         THERE IS NO INTERCEPTED SNOW. USE HEAT BALANCE,
c         LINEARIZED IN VEGTDT, WITH NO LATENT HEAT EXCHANGE
c         TO CALCULATE TEMPERATURE OF VEGETATION ITSELF.
 38       vegtem=temtem
          calrad=(vegm*(((1.-vegalb)*vegqi)+(vege*vegqa)+
        & (((vege*e)/(vege+e-vege*e))*(sigma*tst**4.))
          &-((vege*(vege+2.*e-vege*e)/(vege+e-(vege*e)))*sigma
          &*(-3.*(vegtem**4)))))*deltat
          calhsf=deltat*7.7*rhoa*(.01*(1.+(1080./u)))*ca*u
          calem=deltat*vegm*((vege+(2*e)-(vege*e))/(vege+e-(vege*e)))*vege*
&         sigma*4*(vegtem**3)
          vegtdt=(calrad+(calhsf*ta))/(calhsf+calem)
          vegout=sumwtl
          veglwc=0.0
          vegwtl=0.0
          if(dlw.le.0.0)go to 111
          calqst=(.622/pa)*33.864*(((.00738*(vegtem-273.16)
&         +.8072)**8.)-(.000019*(1.8*((vegtem-273.16)+48.)))
```

```
&          +.001316)
           vegq=(.622/pa)*vegea
           caldq=(597.3-(.57*(vegtem-273.16)))/(.110226*
&          (vegtem**2.))
           callef=calqst-(caldq*calqst*vegtem)
           vegtdt=(calrad+(calhsf*ta)-callef)/(calem+
&          calhsf+(caldq*calqst))
           lef=deltat*((597.3-.57*(vegtem-273.16))*7.0*vegm*rhoa*.01*
&          (1.+(1080./u))*u*(calqst-vegq))
           dlwlos=lef/(597.3-.57*(vegtem-273.16))/pw
           dlw=dlw-dlwlos
           if(dlw.gt.0.0)go to 111
           lef=dlw*(597.3-.57*(vegtem-273.16))*pw
           dlw=0.0
           dlwlos=(1./pw)*(lef/(597.3-.57*(vegtem-273.16)))
 111       continue
           radbal=(vegm*(((1.-vegalb)*vegqi)+(vege*vegqa)+
         & (((vege*e)/(vege+e-vege*e))*(sigma*tst**4.))
           &-((vege*(vege+2.*e-vege*e)/(vege+e-(vege*e)))*sigma
           &*vegtdt**4)))*deltat
           hsf=deltat*(7.7*vegm*rhoa*ca*.01*(1.+(1080./u))*u*(vegtdt-ta))
           radbl2=(vegm*(((1.-vegalb)*vegqi)+(vege*vegqa)+
         & (((vege*e)/(vege+e-vege*e))*(sigma*tst**4.))
           &-((vege*(vege+2.*e-vege*e)/(vege+e-(vege*e)))*sigma
           &*vegtem**4)))*deltat
           hsf2=deltat*(7.7*vegm*rhoa*ca*.01*(1.+(1080./u))*u*(vegtem-ta))
           hsf=(hsf+hsf2)/2.
           radbal=(radbal+radbl2)/2.
           if(sumwtl.le.0.0)go to 12
           hsf=hsf+phsf
           lef=plef+lef
           radbal=radbal+prbl
           go to 12
c****************************************************************************************
c          DETERMINE THE VAPOR PRESSURE, LONGWAVE RAD., AND TEMPERATURE OF
c          THE AIR VENTILATING THE CANOPY AND GROUND
c          SURFACE.
 12        continue
           deltat=temdlt
           if(d.le.0.0)go to 55
           if(tst.lt.273.16)wst=0.0
           if(tst.ge.273.16)wst=wst+vegout
           if(tst.ge.273.16)go to 55
c          MELT OUTFLOW MAY FREEZE IN SURFACE SURFACE LAYER
           ctd=.0222+(.00176*tst)
           fcap1=(273.16-tst)*ctd*d*p
           fposs1=fcap1/(lf*pw)
           wadd=vegout
           if(fposs1.lt.wadd)go to 54
c          ALL OUTFLOW FROZEN IN  SURFACE SURFACE LAYER
           dh1=vegout*lf*pw
           tst=tst+(dh1/(ctd*d*p))
           d=d+((pw*vegout)/p)
           go to 55
c          ONLY SOME OF OUTFLOW FREEZES
 54        tst=273.16
           d=d+((pw*fposs1)/p)
           wst=wst+(vegout-fposs1)
           nok=1
 55        blown=vegdpt-(vegdpt*exp(-factor*u*deltat))
```

```
      vegdpz=vegdpt-blown
      if(vegdpz.ge.0.0)go to 56
      blown=vegdpt
      vegdpt=0.0
      go to 58
56    vegdpt=vegdpz
58    blown=blown*vegden
      q=(.622/pa)*(3.56*(10.**10.))*exp(-6141.9/tst)
      aq=(1.-vegm)*vegq +(vegm*(.3*vegq +(.6*veqsat) +.1*q))
      ea=aq*pa/.622
      qa=((1.-vegm)*vegqa)+(vegm*((vege*sigma*vegtem**4.)
     & +(1.-vege)*e*sigma*tst**4.))/(vege + e -(vege*e))
      ta=((1.-vegm)*(vegta))+(vegm*((.3*vegta)+(.6*vegtdt)+(.1*tst)))
      write(54,437)ta
437   format(1x,'ta=',f12.5)
      vegtem=vegtdt
      vegdp=vegdpt
      return
      end
```

```
      subroutine vegdat(vegtdt,vegout,veglwc,vegdpt,vegden,vgtdt,
     &vgout,vglwc,vgdpt,vgden,ihour,radbal,rdbal,hsf,vghsf,lef,vglef,
     & vblwn,blown,vegls,veglos)
      dimension vgtdt(24),vgout(24),vglwc(24),vgdpt(24),vgden(24),
     & rdbal(24),vghsf(24),vglef(24),vblwn(24),vegls(24)
      real lef
      rdbal(ihour)=radbal
      vgtdt(ihour)=vegtdt
      vgout(ihour)=vegout
      vglwc(ihour)=veglwc
      vgdpt(ihour)=vegdpt
      vghsf(ihour)=hsf
      vglef(ihour)=lef
      vgden(ihour)=vegden
      vblwn(ihour)=blown
      vegls(ihour)=veglos
      return
      end
```

MAJOR CHANGES TO ANDERSON'S (1976) PROGRAM

MAIN PROGRAM

```
c*****************************************************************************
c          INITIALIZE THE SNOW COVER.
           call begin(n,d,p,tt,wt,noknow,its,ti,wei)
c          Matthew initial albedo of surface snowcover!
           a=((g1+(g2*(p(1)**2))+(g3*(p(1)**4)))**.5)*cv*.206
c          Matthew vegetation subroutine
c          Matthew I picked an initial value for ta,ea,qa,qad
           qa=20*idt
           ea=5.0
           aq=0.002
           call vbgin(vegm,vegdp,vegden,vegtem,veglwc,au,factor,
&          dlwmax,dlw)
           ta=vegtem
           surden=p(1)
           sumpx=0.0
           month=imo
           iyear=iyr
           iday=ida
  101      ihour=0
c*****************************************************************************
c          BEGIN MONTHLY LOOP.
           leapyr=0
           if (((iyear/4)*4).eq.iyear)leapyr=1
           last=nd(month)
           if ((month.eq.2).and.(leapyr.eq.1))last=29
c          INPUT HOURLY DATA FOR THE MONTH.
           call datain(last,itopt,its,nobs,tsmax,tt(1),cycle,imo,iyr,month,
&          iyear,iqae)
c                 make these vegetation variables
           if(vegm.le.0.0)go to105
           call vegvar(vegtad,vegead,vegua,vegqid,vegqrd,vegqad,
          &vegpxd,last,au,factor,vegm)
c*****************************************************************************
c          BEGIN DELTA TIME COMPUTATION LOOP.
  105      ihour=ihour+idt
           if (ihour.le.24) go to 110
           iday=iday+1
           ihour=ihour-24
c          GET VALUES FOR INPUT VARIABLES AND VERIFICATION DATA FOR
c             THE TIME PERIOD.
c          Matthew vegetation subroutine
 110       if(vegm.gt.0.0) call vegobt(vegta,vegea,vegu,vegqi,vegqr,u,
          &vegqa,rcf,scf,vegpns,vtsno,vegpx,vegtad,vegead,vegua,vegqid,vegqrd,
          &vegqad,vegpxd,iday,ihour,deltat,vegtw,vegdpt,dlw,dlwmax,vegden)
           if(vegm.gt.0.0)call vegbal(vegtdt,veglwc,vegdpt,vegden,qa,aq,ea,u,
          &hsf,lef,vegu,vegta,radbal,vegtem,vegout,plwhc,tt(1),a,vegm,vegea,ta,pa,
&          vegqa,vegqi,deltat,vegdp,vtsno,vegpns,vegpx,vegtw,cw,wt(1),d(1),p(1),blown,
&          veglos,noknow(1),factor,dlw)
           if(vegm.gt.0.0)call vegdat(vegtdt,vegout,veglwc,vegdpt,
&          vegden,vgtdt,vgout,vglwc,vgdpt,vgden,ihour,radbal,rdbal,
&          hsf,vghsf,lef,vglef,vblwn,blown,vegls,veglos,ta,taf)
           call obtain(vegu,vegm,deltat,iday,ihour,pns,qi,qr,qa,fu,ta,ea,
&          px,tgt,tgtdt,tt(1),ttdt(1),itopt,nobs,to,po,wesc,wepw,depth,
&          stake,pa,tsnow,scf,rcf,noknow(1),adjqa,a)
           call nwsnow(n,tt,wt,ttdt,wtdt,noknow,d,p,rtt,rttdt,blown,vegden,
&          vegtdt,sumpx,thick, itopt)
           if(nobs.ne.1) call nwsnow(n,tt,wt,ttdt,wtdt,noknow,d,p,rtt,rttdt,
&          px,pns,tsnow,sumpx,thick,itopt)
           if (n.eq.0) go to 140
c          GET A FIRST GUESS FOR TEMPERATURE AND LIQUID-WATER
```

```
c             FOR EACH LAYER FOR TIME T+DT.
          call guess(n,tt,ttdt,wt,wtdt,noknow,d,p,itopt,qi,qr,qa,
&         fu,ta,ea,px,do,pa,x,theda,toler,deltat,month,iday,iyear,
&         ihour,rtt,rttdt,igrad,grmax,dtone,texp,ninc,its,tsmax)
c         COMPUTE TEMPERATURE AND LIQUID-WATER FOR EACH LAYER
c             FOR TIME T+DT.
          if (n.gt.1) go to 111
c         SNOW COVER CONSISTS OF A SINGLE LAYER
          iguess=0
          call surfac(vegu,vegm,n,tt(1),ttdt(1),wt(1),wtdt(1),d,p,itopt,qi,qr,
&         qa,fu,ta,ea,px,do,pa,x,dtg,tgt,tgtdt,tcg,dcg,theda,
&         noknow(1),toler,deltat,iter,month,iday,iyear,ihour,iguess)
          go to 112
c         SNOW COVER CONSISTS OF MORE THAN ONE LAYER.
  111     call snowtw(vegu,vegm,n,tt,ttdt,wt,wtdt,d,p,itopt,qi,qr,qa,fu,
&         ta,ea,px,do,pa,x,dtg,tgt,tgtdt,tcg,dcg,theda,noknow,toler,
&         deltat,iter,month,iday,iyear,ihour,ninc)
  112     if (its.eq.0)  call check(n,d,p,tt,ttdt,wt,wtdt,noknow,scout,
&         thick,cthick,we,tdepth,qi,qr,qa,deltat,pa,fu,ta,ea,px,tgt,tgtdt,
&         tcg,dcg,dtg,do,x,itopt)
          if (n.lt.0) go to 130
c         RETAIN TEMPERATURE CHANGE DUE TO HEAT TRANSFER FOR EACH LAYER.
          call retain(n,tt,ttdt,rtt,rttdt)
          if ((do.gt.0.0).or.(its.eq.0)) call vapor(n,p,do,pa,x,tt,ttdt,d,
&         vapour,dtg,dcg,tgt,tgtdt,ea,fu,deltat,wtdt(1),nobs,wtdt(n),
&         noknow(n),soilvt)
          if (nobs.eq.1) go to 113
  130     call water(deltat,n,p,d,wtdt,ttdt,noknow,plwhc,plwmax,plwden,px,
&         scout,month,iday,iyear,ihour,ipunch)
          if (n.lt.0) go to 140
  113     call meta(n,p,d,wtdt,ttdt,plwhc,c1,c2,c3,c4,c5,deltat,dmeta)
c         CHECK IF PRINTER OUTPUT IS WANTED FOR THIS TIME PERIOD.
          mhr=(iday-1)*24+ihour
          if (((mhr/iout)*iout).ne.mhr) go to 125
c         A PRINT IS WANTED FOR THIS TIME PERIOD.
          if (its.lt.2) go to 120
          if (mhr.le.icycle) go to 150
c         OUTPUT FOR SELECTED SNOW SURFACE TEMPERATURE PATTERNS.
          call patern (n,ttdt,wtdt,its,tsmax,cycle,d,p,iday,
&         ihour,iter,do,ti,surden)
          go to 150
c         OUTPUT FOR OTHER CASES
  120     call snowot(n,ttdt,wtdt,d,p,month,iday,iyear,ihour,iter,we,tdepth)
 125      if(nobs.eq.1)go to 150
  140     call statda(tt(1),ttdt(1),to,scout,po,we,wesc,wepw,vapour,tdepth,
&         depth,stake,month,iday,iyear,ihour,deltat,n,soilvt,ta,vegm,vgout,
         &vghsf,vglef,rdbal,vgtdt,vglwc,vgdpt,vblwn,vegls,taf)
c         Matthew store time step veg variables
c         END OF DELTA TIME LOOP, CHECK FOR END OF RUN AND MONTH.
c*********************************************************************
  150     if (ihour.lt.24) go to 105
          if (month.ne.lmo) go to 160
          if ((iyear.eq.lyr).and.(iday.eq.lda)) go to 200
          go to 105
  160     if (iday.lt.last) go to 105
          month=month+1
          iday=1
          if (month.le.12) go to 101
          month=1
          iyear=iyear+1
```

```
          go to 101
c         END OF THE RUN
c*********************************************************************
  200     if (nobs.ne.1) call final(imo,ida,iyr,lmo,lda,lyr,wei,sumpx,we,vegm)
          stop 99
          end
```

CHANGES IN ANDERSON'S (1976) SUBROUTINE OBTAIN

```
c*****************************************************************************
c         COMPUTE INPUT AND VERIFICATION VALUES FROM HOURLY DATA
c                 COMPUTE VARIOUS TOTALS FOR THE PERIOD. NO CHECK MADE FOR
c                 MISSING DATA. USER MUST BE SURE ALL NEEDED DATA ARE
c                 AVAILABLE FOR THE PERIOD BEING USED.
c         Matthew alittle change here
110       if(vegm.gt.0.00)go to 17
          ta=0.0
          ea=0.0
          qa=0.0
 17       u=0.0
          qr=0.0
          qi=0.0
          px=0.0
          po=0.0
          do 111 mhr=mhr1,mhr2
          u=u+ua(mhr)
          qi=qi+qid(mhr)
          px=px+pxd(mhr)
          qr=qr+qrd(mhr)
c         Matthew I avoid these and get them from vegbal if M.gt.0
          if(vegm.gt.0.00)go to 642
          qa=qa+qad(mhr)*adjqa
          ta=ta+tad(mhr)
          ea=ea+ead(mhr)
 642      po=po+pod(mhr)
  111     continue
c         COMPUTE MEANS WHERE NEEDED
          if(vegm.gt.0.00)go to 18
          ta=ta/deltat
          ea=ea/deltat
 18       u=u/deltat
          umsec=u
c*****************************************************************************
```

CHANGES IN ANDERSON'S (1976) SUBROUTINE WINDF

```
      subroutine windf(deltat,ua,ta,tst,tstdt,fucoef,pa,height,ifu,fu,
     & vegu,vegm)
C***********************************************************************
C     COMPUTES THE WIND FUNCTION FOR THE PERIOD--(CM/MB)
C          UA IS AVERAGE WIND--METERS/SEC
C          FUCOEF IS IN--(MM/MB/KM)
C          HEIGHT IS IN METERS
C          ZO IS IN CENTIMETERS.
C***********************************************************************
      common/critri/ricrit,zo
      dimension ri(50),cwr(50),ratio(50),ribulk(50)
      data ifirst/0/
      if (ifirst.gt.0) go to 100
C     MIMIMUM WIND FUNCTION--MOLECULAR CONDUCTION.
      fumin=(0.646*deltat)/(pa*height*100.0)
      ifirst=1
      if(ifu.lt.1)go to 100
C***********************************************************************
C     GENERATE TABLE OF BULK TRANSFER COEFFICIENT RATIO FOR
C        WATER VAPOR AND HEAT TO ITS VALUE FOR NEUTRAL
C        CONDITIONS AS A FUNCTION OF THE BULK RICHARDSON NUMBER.
C     COMPUTE NEUTRAL BULK TRANSFER COEFFICIENT
      cdn=0.16/((alog(height*100.0/zo))**2)
C     GENERATE TABLE
      do 110 j=1,50
      zl=-((j*0.1)**2)
      x=(1.0-16.0*zl)**0.25
      term=1.0-(sqrt(cdn)/0.4)*(alog((1.0+x*x)/2.0)
     &     +2.0*alog((1.0+x)/2.0)-2.0*atan(x)+1.5708)
      cdr=1.0/(term*term)
      chr=sqrt(cdr)/(1.0-5.0*sqrt(cdn)*alog((1+x*x)/2.0))
      rib=(zl*sqrt(cdn)*(cdr**1.5))/(0.4*chr)
      ri(j)=rib
      cwr(j)=chr
  110 continue
C     PUT TABLE INTO COMPUTABLE RI INCREMENTS.
      do 115 j=1,50
      rib=-((j*0.03)**2)
      ribulk(j)=rib
      do 116 jj=1,50
      inc=jj
      if(rib.gt.ri(jj))go to 120
  116 continue
      inc=51
  120 if(inc.gt.1)go to 122
      ratio(j)=((rib/ri(inc))*(cwr(inc)-1.0))+1.0
      go to 115
  122 if(inc.le.50)go to 125
      ratio(j)=(((rib-ri(49))/(ri(50)-ri(49)))*
     &         (cwr(50)-cwr(49)))+cwr(49)
      go to 115
  125 ratio(j)=(((rib-ri(inc-1))/(ri(inc)-ri(inc-1)))*
     &         (cwr(inc)-cwr(inc-1)))+cwr(inc-1)
  115 continue
      ctn=fucoef
C***********************************************************************
C     COMPUTE WIND TRAVEL IN KILOMETERS.
C     Matthew  Different wind function for vegetation
C     Here ua is restored to its value for bare snowcovered soil
C     Matthew note that this value is measured at a heightabove
```

```
c          the height of the vegetation
  100      if(vegm.gt.0.0)uaf=ua
           if(vegm.gt.0.0)ua=vegu/360000.
           ut=3.6*ua*deltat
           if (ifu.eq.1) go to 101
c*********************************************************************
c          USE EMPIRICAL WIND FUNCTION
           fu=0.1*fucoef*ut
           return
c*********************************************************************
c          USE THEORETICAL WIND FUNCTION WITH A STABILITY CORRECTION.
  101      if (ua.gt.0.20) go to 102
c          CALM CONDITIONS
           fu=fumin
           return
c          COMPUTE THE RICHARDSON NUMBER.
  102      ts=0.5*(tst+tstdt)
           rib=(2.0*9.8*(ta-ts)*height)/((ta+ts)*ua*ua)
           if(rib.ge.0.0)go to 103
c          UNSTABLE CONDITIONS.
           j=(sqrt(-rib)/0.03)+1.0
           if((j.eq.1).or.(j.gt.50))go to 106
           r=(((rib-ribulk(j-1))/(ribulk(j)-ribulk(j-1)))
&                  *(ratio(j)-ratio(j-1)))+ratio(j-1)
           go to 107
  106      if(j.gt.50)go to 108
           r=((rib/ribulk(1))*(ratio(1)-1.0))+1.0
           go to 107
  108      r=(((rib-ribulk(49))/(ribulk(50)-ribulk(49)))
&                  *(ratio(50)-ratio(49)))+ratio(49)
  107      ct=r*ctn
           go to 105
c          NEUTRAL OR STABLE CONDITIONS.
  103      if(rib.lt.ricrit)go to 104
           ct=0.0
           go to 105
  104      r=(1.0-(rib/ricrit))**2
           ct=r*ctn
c          COMPUTE WIND FUNCTION.
  105      fu=0.1*ct*ut
           if(vegm.le.0.0)go to126
c          Matthew this changes wind function when vegetation is present
           cf=(.01*(1.+(0.3/uaf)))
           rhoa=.00122
           pw=1.0
           ww=(rhoa*.622)/(pw*pa)
           fu=((1-vegm)*fu)+(vegm*(7.0*vegm*cf*uaf*ww*deltat*0.1))
 126       continue
c          Matthew change ua back to its original value
           if(vegm.gt.0.0)ua=uaf
           if(fu.lt.fumin)fu=fumin
c*********************************************************************
           return
           end
```

CHANGES IN ANDERSON's (1976) SUBROUTINE STATDA

```
      subroutine statda(tst,tstdt,to,scout,po,we,wesc,wepw,vapour,
&     tdepth,depth,stake,month,iday,iyear,ihour,deltat,n,soilvt,ta,
     &vegm,vgout,vghsf,vglef,rdbal,vgtdt,vglwc,vgdpt,vblwn,vegls,taf)
C*********************************************************************
c     THIS SUBROUTINE COMPARES VERIFICATION DATA BY.
c          1. STORING SURFACE TEMPERATURE,OUTFLOW AND VAPOR TRANSFER
c                FOR DISPLAY ONCE PER DAY.
c          2. COMPUTING RUN STATISTICS BETWEEN OBSERVED AND COMPUTED
c                SURFACE TEMPERATURE AND OUTFLOW.
c          3. PRINTING A DAILY SUMMARY OF WATER-EQUIVALENT,
c                AND DEPTH WHEN IHOUR=24.
C*********************************************************************
      common/stats/ncpo,opo,spo,spo2,ospo,ospo2,podiff,ncto,oto,
&     sto,sto2,osto,osto2,todiff,vapor,water,vsoil,opo2,oto2,
&      tqs,tqlw,tqh,tqe,tqg,tqpx,irf(10),itf(10)
      common/ebal/qs,qlw,qh,qe,qg,qpx,dq
      common/wspeed/ifu,umsec,fucoef,height
      common/stemp/tosim
C*********************************************************************
      dimension spoa(24),poa(24),stoa(24),toa(24),vt(24)
      dimension qsa(24),qlwa(24),qha(24),qea(24),qga(24),qpxa(24),
&     dqa(24),ri(24),rl(9),tl(9)
      dimension vgout(24),vghsf(24),vglef(24),rdbal(24),vgtdt(24),vglwc(24),
&     vgdpt(24),vblwn(24),vegls(24),taf(24)
      data ifirst/0/
      data rl/-0.2,-0.02,0.0,0.01,0.02,0.05,0.1,0.2,0.4/
      data tl/-4.0,-2.0,0.0,2.0,4.0,6.0,8.0,10.0,12.0/
      tvout=0.0
      tvlef=0.0
      tvhsf=0.0
      trbl=0.0
      do 8 i=1,24
      tvout=tvout+vgout(i)
      tvhsf=tvhsf+vghsf(i)
      tvlef=tvlef+vglef(i)
      trbl=trbl+rdbal(i)
 8    continue
      if(ifirst.gt.0)go to 100
c     INITIALIZE STAT VALUES.
      taft=0.0
      ncpo=0
      opo=0.0
      spo=0.0
      spo2=0.0
      opo2=0.0
      ospo=0.0
      ospo2=0.0
      podiff=0.0
      ncto=0
      oto=0.0
      sto=0.0
      sto2=0.0
      oto2=0.0
      osto=0.0
      osto2=0.0
      todiff=0.0
      vapor=0.0
      vsoil=0.0
      water=0.0
      tqs=0.0
```

```
      tqlw=0.0
      tqh=0.0
      tqe=0.0
      tqg=0.0
      tqpx=0.0
      np=0
      do 116 j=1,10
      irf(j)=0
116   itf(j)=0
      ifirst=1
100   idt=deltat+0.01
c***********************************************************************
c     CHECK FOR THE FIRST PERIOD OF A DAY -- INITIALIZE SUMS.
      if (ihour.gt.idt) go to 112
      if (n.eq.0) return
      sumvt=0.0
      sumspo=0.0
      sumpo=0.0
      sumqs=0.0
      sumqlw=0.0
      sumqh=0.0
      sumqe=0.0
      sumqg=0.0
      sumqpx=0.0
      sumdq=0.0
      mpo=0
      mto=0
      go to 104
c     DO NOT COMPUTE VALUES IF A SNOW COVER DID NOT EXIST ON THE
c          FIRST RERIOD OF THE DAY.  (RUN TOTALS MUST BE COMPUTED.)
112   if ((np.eq.0).and.(n.eq.0)) return
      if (np.eq.0) go to 105
      if (n.eq.0) go to 115
c***********************************************************************
c     STORE VALUES FOR PRINTING AT END OF DAY.
104   i=ihour/idt
      if (n.lt.0) n=0
      np=np+1
      if (np.eq.i) go to 130
c     FILL IN INTERMEDIATE VALUES ON DAY WHEN SNOW DISAPPEARS AND REAPPE
      jj=i-1
      do 131 j=np,jj
      spoa(j)=999.9
      poa(j)=999.9
      stoa(j)=9999.
      toa(j)=9999.
      vt(j)=99.99
      qsa(j)=999.9
      qlwa(j)=999.9
      qha(j)=999.9
      qea(j)=999.9
      qga(j)=999.9
      qpxa(j)=999.9
      ri(j)=99.99
131   dqa(j)=999.9
      np=i
130   spoa(i)=scout
      poa(i)=po
      ts=(tst+tstdt)*0.5
      if (n.gt.1) ts=tosim
```

```
c          Matthew COMPUTE AVERAGE VENTILATING TEMPERATURE
           if(vegm.le.0.0)go to 777
           if(ihour.lt.24)go to 777
           taft=0.0
           do 419 mm=idt,ihour,idt
           taft=taf(mm)+taft
 419       continue
           tafav=taft/(ihour/idt)
 777       continue
c          COMPUTE THE RICHARDSON NUMBER.
           ua=umsec
           if (ua.lt.0.01) ua=0.01
           rin=(2.0*9.8*((ta-ts)/height))/((ta+ts)*((ua/height)**2))
           if (rin.lt.-2.0) rin=-2.0
           if (rin.gt.2.0) rin=2.0
           ri(i)=rin
c          FILL FREQUENCY TABLES.
c               RICHARDSON NUMBER.
           do 132 j=1,9
           if (rin.gt.rl(j)) go to 132
           irf(j)=irf(j)+1
           go to 135
  132      continue
           irf(10)=irf(10)+1
c          AIR TEMP. MINUS SNOW SURFACE TEMP.
  135      tdiff=ta-ts
           if (ta.gt.273.16) go to 140
           do 136 j=1,9
           if(tdiff.gt.tl(j)) go to 136
           itf(j)=itf(j)+1
           go to 140
  136      continue
           itf(10)=itf(10)+1
  140      continue
           ts=ts-273.16
           if (n.eq.0) ts=0.0
           stoa(i)=ts
           toa(i)=to
           vt(i)=vapour*10.0
           if(n.gt.0)go to 120
           vapour=0.0
           vt(i)=0.0
           qsa(i)=0.0
           qlwa(i)=0.0
           qha(i)=0.0
           qea(i)=0.0
           qga(i)=0.0
           qpxa(i)=0.0
           dqa(i)=0.0
           qs=0.0
           qlw=0.0
           qh=0.0
           qe=0.0
           qg=0.0
           qpx=0.0
           dq=0.0
           go to 121
  120      qsa(i)=qs
           qlwa(i)=qlw
           qha(i)=qh
```

```
      qea(i)=qe
      qga(i)=qg
      qpxa(i)=qpx
      dqa(i)=dq
121   sumqs=sumqs+qs
      sumqlw=sumqlw+qlw
      sumqh=sumqh+qh
      sumqe=sumqe+qe
      sumqg=sumqg+qg
      sumqpx=sumqpx+qpx
      sumdq=sumdq+dq
c*********************************************************************
c     COMPUTE TOTALS AND ADD TO STATISTICS.
      sumvt=sumvt+vt(i)
      sumspo=sumspo+scout
      if (po.lt.9000.0) go to 101
      mpo=mpo+1
      go to 102
c     SNOW COVER OUTFLOW STATISTICS
101   sumpo=sumpo+po
      ncpo=ncpo+1
      opo=opo+po
      spo=spo+scout
      spo2=spo2+scout*scout
      opo2=opo2+po*po
      ospo=ospo+po*scout
      ospo2=ospo2+(po-scout)*(po-scout)
      podiff=podiff+abs(po-scout)
102   if (to.lt.9000.0) go to 103
      mto=mto+1
      go to 105
c     SNOW SURFACE TEMPERATURE STATISTICS.
103   ncto=ncto+1
      oto=oto+to
      sto=sto+ts
      sto2=sto2+ts*ts
      oto2=oto2+to*to
      osto=osto+to*ts
      osto2=osto2+(to-ts)*(to-ts)
      todiff=todiff+abs(to-ts)
c     RUN TOTALS
105   vapor=vapor+vapour
      water=water+scout*0.1
      vsoil=vsoil+soilvt
      tqs=tqs+qs
      tqlw=tqlw+qlw
      tqh=tqh+qh
      tqe=tqe+qe
      tqg=tqg+qg
      tqpx=tqpx+qpx
      if (np.eq.0) return
115   if (ihour.lt.24)return
c*********************************************************************
c     END OF DAY SUMMARY.
```

```
        if (n.eq.0) print 926
  926   format (1h1)
        write(6,900)
  900   format (1h0,120h*****************************************************
       &****************************************************************
       &****)
        write(6,901)month,iday,iyear
  901   format (1h0,15hDAILY SUMMARY--,i2,1h/,i2,1h/,i4,10x,
     &  34h(ALL NINES INDICATES MISSING DATA))
        write(6,902)(i,i=idt,ihour,idt)
  902   format (1h0,4hHOUR,1x,24i5)
        write(6,903)
c       Matthew printing veg output
        if(vegm.le.0.0)go to 4
        write(6,806)
 806    format(1x,'INTERCEPTED OUTFLOW  (CM WATER)')
        write(6,807)(vgout(i),i=idt,ihour,idt)
 807    format(1x,'sim.',1x,24f5.2)
        write(6,90)tvout
 90     format(123x,f8.1)
        write(6,797)
 797    format(1x, 'VENTILATING AIR TEMPERATURE')
        write(6,796)taf
 796    format(6x,24f5.1)
        write(6,795)tafav
 795    format(1x,'AVERAGE VENTILATING AIR TEMP =',f9.3)
        write(6,808)
 808    format(1x,'ENERGYBALANCE COMPONENTS ABOVE VEGETATION')
        write(6,809)
 809    format(1x,'SENSIBLE HEAT TRANSFER FROM VEG TO ATM (CAL/CM2)')
        write(6,810)(vghsf(i),i=idt,ihour,idt)
 810    format(1x,'sim.',1x,24f5.1)
        write(6,91)tvhsf
 91     format(123x,f8.1)
        write(6,811)
 811    format(1x,'LATENT HEAT TRANSFER FROM VEG TO ATM (CAL/CM2)')
        write(6,812)(vglef(i),i=idt,ihour,idt)
 812    format(1x,'sim.',24f5.1)
        write(6,92)tvlef
 92     format(123x,f8.1)
        write(6,813)
813     format(1x,'RADIATION BALANCE FROM ATM TO VEG (CAL/CM2)')
        write(6,814)(rdbal(i),i=idt,ihour,idt)
 814    format(1x,'sim.',1x,24f5.1)
        write(6,93)trbl
 93     format(123x,f8.1)
        write(6,820)
 820    format(1x,'sim',1x,'TEMPERATURE OF INTERCEPTED SNOW OR OF CANOPY')
        tfav=0.0
        do 1200 nm=idt,ihour,idt
        tfav=tfav+vgtdt(nm)
 1200   continue
        tfav=tfav/(ihour/idt)
        write(6,815)(vgtdt(i),i=idt,ihour,idt)
 815    format(1x,'sim',1x,24f5.0)
        write(6,1201) tfav
 1201   format(124x,f7.3)
        write(6,816)
 816    format(1x,'LIQUID WATER CONTENT OF INTERCEPTED SNOW')
        write(6,817)(vglwc(i),i=idt,ihour,idt)
```

```
817     format(1x,'sim',1x,24f5.2)
        write(6,818)
818     format(1x,'DEPTH OF SNOW ON VEGETATION')
        write(6,819)(vgdpt(i),i=idt,ihour,idt)
819     format(1x,'sim',1x,24f5.1)
 903    format (1h+,125x,2x,5hTOTAL)
        write(6,825)
825     format(1x,'DEPTH OF SNOW SUBLIMATED FROM VEGETATION (CM SNOW)')
        write(6,821)(vegls(i),i=idt,ihour,idt)
821     format(1x,'sim',1x,24f5.1)
        write(6,822)
822     format(1x,'SNOW BLOWN FROM VEGETATION TO SURFACE (CM WATER EQUIVaIENT)')
        write(6,823)(vblwn(i),i=idt,ihour,idt)
823     format(1x,'sim',1x,24f5.1)
c       SNOW COVER OUTFLOW.
 4      write(6,904)
  904   format (1h ,22hSNOW COVER OUTFLOW-MM.)
        write(6,905)(spoa(i),i=1,np)
  905   format(1h ,4hSIM.,1x,24f5.1)
        write(6,906)sumspo
  906   format (1h+,125x,f7.1)
        if (mpo.ge.np) go to 106
        if (mpo.gt.0)sumpo=9999.9
        do 113 i=1,np
        if (poa(i).lt.9000.0) go to 113
        poa(i)=999.9
  113   continue
        write(6,907)(poa(i),i=1,np)
  907   format (1h ,4hOBS.,1x,24f5.1)
        write(6,906)sumpo
c       SNOW SURFACE TEMPERATURE.
  106   write(6,908)
  908   format (1h ,31hSNOW SURFACE TEMPERATURE-DEG.C.)
        write(6,915)(stoa(i),i=1,np)
  915   format(1h ,4hSIM.,1x,24f5.0)
        if (mto.ge.np) go to 107
        write(6,916)(toa(i),i=1,np)
  916   format (1h ,4hOBS.,1x,24f5.0)
c       VAPOR TRANSFER.
  107   write(6,909)
  909   format (1h ,27hAIR-SNOW VAPOR TRANSFER-MM.)
        write(6,910)(vt(i),i=1,np)
  910   format (1h ,4hSIM.,1x,24f5.2)
        write(6,911)sumvt
  911   format (1h+,125x,f7.2)
        write(6,917)
  917   format(1h ,34hENERGY BALANCE COMPONENTS-CAL/CM2.)
        write(6,918)
  918   format(1h ,2hQS)
        write(6,919)(qsa(i),i=1,np)
  919   format(1h+,5x,24f5.1)
        write(6,906)sumqs
        write(6,920)
  920   format(1h ,3hQLW)
        write(6,919)(qlwa(i),i=1,np)
        write(6,906)sumqlw
        write(6,921)
  921   format(1h ,2hQH)
        write(6,919)(qha(i),i=1,np)
        write(6,906)sumqh
```

```
        write(6,922)
  922   format(1h ,2hQE)
        write(6,919)(qea(i),i=1,np)
        write(6,906)sumqe
        write(6,923)
  923   format(1h ,2hQG)
        write(6,919)(qga(i),i=1,np)
        write(6,906)sumqg
        write(6,924)
  924   format(1h ,3hQPX)
        write(6,919)(qpxa(i),i=1,np)
        write(6,906)sumqpx
        write(6,925)
  925   format(1h ,2hDQ)
        write(6,919)(dqa(i),i=1,np)
        write(6,906)sumdq
        write(6,927)
  927   format (1h ,2hRI)
        write(6,928)(ri(i),i=1,np)
  928   format (1h+,5x,24f5.2)
c       DEPTH AND WATER-EQUIVALENT.
        if (tdepth.eq.0.0) go to 109
        sden=(we*0.1)/tdepth
        go to 111
  109   sden=0.0
  111   if (wesc.gt.9000.0) go to 108
        if (depth.lt.0.001) go to 108
        oden=(wesc*0.1)/depth
        go to 110
  108   oden=99.99
  110   write(6,912)
  912   format (1h0,7x,15hSIMULATED(2400),11x,24hSNOW COURSE(MID-MORNING),
&       6x,12hPILLOW(2400),3x,23hSNOW STAKE(MID-MORNING))
        write(6,913)
  913   format (1h ,4x,6hWE(MM),1x,9hDEPTH(CM),3x,7hDENSITY,
&       4x,6hWE(MM),1x,9hDEPTH(CM),3x,7hDENSITY,9x,6hWE(MM),
&       6x,9hDEPTH(CM))
        print 914,we,tdepth,sden,wesc,depth,oden,wepw,stake
  914   format (1h ,2f10.0,f10.3,2f10.0,f10.3,2f15.0)
 709    np=0
c***********************************************************************
        return
        end
```

APPENDIX II

Sample Input

Input cards are as in Andersons's (1976) program, with the exception of the eighth card, which is read by subroutine vbgin. The values that should be, or were, used are shown:

Card No.	Format	Name	Remarks
1	I5	itopt	= 0, solve for snow surface temp.
	I5	its	blank
	I5	ipunch	= 1, punch excess liquid water
2	5X,15A4	title	general information
	I5	imo	initial month used
	I5	ida	initial day
	I5	lyr	initial year
	I5	1 mo	last month to be used
	I5	ida	last day
	I5	lyr	last year
3	F5.0	deltat	= 1, time step length in hours
	F5.0	dtout	= 24, output time interval in hours
	F5.1	theda	= 0.5, semi-implicit finite defference
	F5.2	toler	= 0.01, tolerance for iteration
	I5	igrad	= 0, use regular guess method
	F5.0	grmax	= 2.0, maximum gradient ratio
	F5.0	dtone	= 99.0, time step subdivision parameter
	F5.1	texp	= .5, exponent in subdividing equation
	I5	iqae	= 0, use observed longwave radiation
	F5.1	thick	= 10.0, desired thickness of surface layers
	F5.2	cthick	= 0.05, increases thickness with depth.
	F5.2	adjqa	= 1.00, longwave adjustment factor
	I5	ifu	= 1, theoretical wind function (important)
	F5.1	height	= 1, height of meteorological measurements in meters. Must be above height of vegetation. In this case, vegetation is bushes.

Card No.	Format	Name	Remarks
4	F5.2	do	=.90, diffusion coefficient for water vapor in snow at 0°(cm^2/sec)
	F5.0	pa	=950, average station pressure, mb.
	F5.1	x	= 14 exponent in diffusion equation
	F5.0	dtg	= 5, depth of soil temperature measure, cal/cm/sec/°C
	F5.3	tcg	=.001, thermal conductivity of soil
	F5.2	dcg	= .20, diffusion coefficient for water vapor in soil, cm^2/sec
	F5.2	scf	= 1.00, snow multiplication factor
	F5.2	rcf	= 1.00, rain multiplicaton factor
	F5.2	plwhc	= .03, basic liquid water holding capacity of snow, (decimal)
	F5.2	plwmax	= .10, maximum liquid water holding capacity (decimal)
	F5.2	plwden	= .20, density above which plwmax is used
	F5.2	fucoef	blank used for empirical f(u) only
	F5.4	coefke	= .0060, effective thermal conductivity equation parameter
	F5.2	ricrit	= .20, critical richardson number
	F5.2	20	= .15, roughness height, cm
5	F5.4	C1	= .01, compaction equation parameter
	F5.1	C2	= 21., compaction equation parameter
	F5.3	C3	= .01, destructive metamorphism parameter
	F5.2	C4	= .04, destructive metamorphism parameter
	F5.2	C5	= .15, destructive metamorphism parameter
	F5.1	dmeta	= 2.0, destructive metamorphism parameter
	F5.1	CW1	= 10., transmission parameter
	F10.5	CW2	= 1., transmission parameter
	2F5.1	CW3, CW4	= 5., 450., transmission parameter
	3F5.1	g1, g2, g3	= .16, 0., 110., grainsize aprameter
6	I5	n	= 0, initital number of snow layers
	I5	mm	= 1, units of liquid water on card 7 1 = millimeters

Card No.	Format	Name	Remarks
7	I5	nn	= 0, layer number
(read by begin)	F5.0	thick	= 10., thickness of layer in cm
	F5.2	den	= .25, density of layer (decimal)
	F5.0	temp	= -1., mean temperature of layer, °C
	F5.2	water	= 0., liquid water content of layer
8	F3.2	vegm	= canopy density (decimal)
(read by vbgin)	F5.1	vegdpt	= 0., initial depth of snow on vegetation, cm
	F3.2	vegden	= .35, density of snow on canopy
	F6.2	vegtem	= 268., temperature of snow on canopy °K
	F6.4	veglwc	= 0., liquid water content of snow or canopy cm
	F6.2	a	= 1.00, wind attenuation coefficient
	F9.7	factor	= blown snow coefficient (see below)
	F6.0	dlwmax	= max intercepted snow capacity, cm (see below)
9	I2	mo	= month number
	IX, I2	jyr	= year (digits)
		Soil temperature data. Must be repeated as a group five times. Needed for last hour of month and all other hours when changes occur.	
	I5	id(1-5)	1, 10, 15, 20 31, = day
	I5	ih(1-5)	1, 1, 1, 1, 24 = hour
	F5.0	gt(1-5)	16., 32, 32., 32, 32 = temperature
		(Sample runs on two day period of March)	
10	I2	mo	= 3, month number
	I3	jyr	= 81, year number
	I1	iend	= 1, this is the last new snow density card

Note:

For cases of no blown snow in Chapter four, factor = 0. Other cases, factor = 0.0000001. For case of no interception, dlw = 0. Other cases, dlw was set to 100.

Meteorological Variables Used:

Initial values:

atmospheric radiation = qa = 20 cal/cm^2/hr

vapor pressure = ae = 5 mb

specific humidity of air = aq = .002

air temperature = ta = vegtem (268°K)

Subsequent values:

vapor pressure	ead(i), i=i, 48:	5 mb
wind speed	ua(i), i=i, 48:	4 mps
insolation	qid(i), i=i, 48:	28 cal cm^{-2}
reflected insolation	qrd(i), i=i, 48:	30 cal cm^{-2}
atmospheric radiation	qad(i), i=i, 48:	
precipitation	pxd(i), = 300 mm	
	pxd(i), i=2, 48 = 0	
precipitation density	denns(i),= .25 g/cm^3	
	denns (i),i=2, 48: 0	
Air temperature	tad (i), i=1, 48: read from "file 23". Given a sine fluctuation:	

For cold weather cases: 276.5

$$t(i) = 266.5 + (7 \, . \, (\sin^2(i/24)))$$
$$i = 1,24$$

$$t(i) = 267.5 + (7(\sin^2((i-24)/24)))$$
$$i = 25, 48:$$

For warm weather cases:

$$t(i) = 276.5 + (7(\sin^2(i/24))$$
$$i = 1,24$$

$$t(i) = 277.5 + (7(\sin^2((i-24))/24)))$$

All "verification" variables for Anderson's (1976) model were set to the default (= not known) value, 9999.9.