SNOW DEPTH VARIATIONS, SCHEFFERVILLE P.Q.

SNOW DEPTH VARIATIONS IN A FOREST-TUNDRA ENVIRONMENT

SCHEFFERVILLE, P.Q. WINTER 1968-69

by

Hardy B. Granberg

ABSTRACT

Snow depth data from 147 sampling points were collected through winter. Analysis shows that three periods of characteristically different patterns of depth increase are produced by sequential changes in surface roughness. Six different types of depth growth curves are identified. Two methods of indirect mapping of snow depths are described. One method is based on sequence aerial photographs during melt. The second method employs a stepwise multiple regression model to select variables relevant to the particular pattern of snow accumulation. A set of topographic variables are generated from a matrix of altitudes obtained from a map. Four variables in a multiple regression equation explain 75 per cent of the original depth variations. A computer mapping program is appended.

M.Sc. Thesis
Department of Geography
McGill University
Montreal 110, P.Q.

March, 1972

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A thesis submitted to the Faculty of Graduate Studies
and Research in partial fulfilment of the requirements
for the degree of Master of Science

Department of Geography
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ACKNOWLEDGEMENTS

I would like to thank Dr. Bruce G. Thom who supervised the initial stages of the study and also suggested the general topic. Secondly I am grateful to Dr. Frank H. Nicholson who supervised the final stages of the thesis. The project was greatly helped by the participation of Mr. Lee Nichols. The study was jointly sponsored by the Iron Ore Company of Canada, the National Research Council of Canada and the McGill Sub-Arctic Research Laboratory in Schefferville. I would also like to thank previous members of the staff of the McGill Sub-Arctic Research Laboratory for their assistance in the field work and Mr. Atsumu Ohmura for help in the initial stages of the computational procedure. Lastly I would like to thank Mrs. Margaret Nicholson and Mrs. Nuala Orth, the former for help in making final drafts of some of the diagrams and the latter for typing the thesis.

TABLE OF CONTENTS

TABLE OF CONTENTS (Continued)

TABLE OF CONTENTS (Continued)

LIST OF FIGURES

LIST OF FIGURES (Cont.)

LIST OF TABLES

CHAPTER I

INTRODUCTION

1. General

The aim of the study is to investigate the spatial and temporal variations in the accumulation of the seasonal snow cover. The study concentrates on the short distance depth variations produced by the redistribution of snow by wind. The object of the study is to define a relationship between terrain roughness factors and the patterns of snow accumulation in terrain.

The term "forest tundra" was used by Hare (1950) to describe an area which is predominantly tundra but where forest occurs in sheltered valleys and depressions. In Labrador - Ungava this type of terrain is common in the more elevated central parts and at lower levels farther north. It is an environment where wind is a very important factor.

The accumulation of snow in this type of environment is extremely irregular due to strong redistribution of snow by wind. In the Schefferville area the number of days of blowing snow may be as high as 75 in one winter. Some parts of the terrain remain snow-free throughout winter whereas occasional localities may have accumulations sufficiently deep to persist through summer in some years, forming semi-permanent snow patches (Ives, 1960). The redistribution and irregular accumulation of snow affects ecology, hydrology, geomorphology and human activity. The variation in accumulation creates large variations in ground insulation and leaves large parts of the terrain fully exposed to heat losses throughout winter. This is probably the main factor controlling permafrost distribution in the Schefferville area (Annersten, 1963).

The first objective of this study is to characterise the sequential development of the seasonal snowcover. The working hypothesis is that since wind variations are controlled by terrain roughness, the accumulation of snow follows a sequence which is determined by snowfall and terrain roughness. The second objective of this study is to develop means of indirect mapping of the snowcover at different stages of winter. The general approach to the problem has been suggested by previous research by Black and Budd (1964), Mock (1968) and Young (1970) who have shown that it is possible to obtain a statistical approximation of snow accumulation from variables depicting terrain geometry. These studies, however, have so far only been successful in predicting accumulation of snow on glaciers, where terrain roughness is less pronounced and where snow accumulation has reached close to an equilibrium. The study forms an integrated part of the permafrost research program at Schefferville.

2. Outline of Thesis

The thesis has been structured in the following way. In the subsequent section of the introduction the field area, Timmins 4, is described. The climatic characteristics of the Schefferville area are briefly outlined. This is followed by an account of some of the work on snow that has been previously undertaken at the McGill Sub-Arctic Research Laboratory in Schefferville.

Chapter II outlines the salient aspects of the processes involved in the redistribution of snow by wind and the interrelationship between snow drift transport and terrain roughness.

Field observations are described in Chapter III and the patterns of accumulation are investigated by analysis of snow depth data collected on a 147 point grid through winter. The influence of surface roughness is discussed with respect to the overall patterns of accumulation through winter and with respect to individual points in the terrain. A comparison is made between the accumulation at the Timmins 4 ridge area and lower lying, less exposed areas.

A statistical model of snow accumulation is described in Chapter IV. Stepwise multiple regression is used to obtain equations describing the state of the snowcover at different times during winter in terms of a series of terrain variables. These variables are intended to account for terrain effects at different stages of winter and the varying influence of winds from different directions. This chapter also describes the collection and calculation of terrain variables. The topographic variables were produced from map analysis. Aerial photographs in black and white, and color infrared, were used for vegetation variables. One vegetation variable was measured in the field.

Chapter V is an analysis of the efficiency of the different terrain variables as predictors of snow depth. Simple correlation is used to indicate the efficiency of single variables. Multiple regression equations are analysed. A snow depth map is produced for March 13 using one of the equations and the map is compared to a snow depth map produced from sequence aerial photographs.

Conclusions and suggestions for further work are given in Chapter VI.

3. Description of the field area

The area selected for the study is Timmins 4, a site of permafrost research 20 km northwest of the Schefferville townsite (Figure 1). A plot approximately 1200 x 400 m was outlined. Within this area the short distance variations in snow accumualtion were monitored after each major snow event through the winter by means of snow depth sampling at 147 points. The experimental site is located at an altitude of 700 m A.S.L. on the northwest-facing slope of a ridge and has a range in altitude of about 50 m.

The experimental site (Figure 2) may be subdivided into three parts. The southeastern third is characterised by a large, relatively flat area and a wide valley. In the centre part the general slope towards northwest is steepest. The microrelief is more pronounced in this section and four drainage channels converge towards the northwestern part of this section, where several shallow, ephemeral ponds are found. The northwestern part of the experimental area has a general slope towards northeast and only gently undulating topography.

The vegetation distribution of Timmins 4 shows an apparent relationship to the topographic trend (Figure 3). In the map which was produced from infrared color aerial photographs (see Chapter III) five different classes of low-grown vegetation are distinguished:

1. Bare rock or soil
2. Discontinuous vegetation cover (mainly lichen, _Vaccinium_ spp. and low grown _Betula_ spp.)
3. Continuous lichen mat with scattered, low grown woody plants (mainly _Betula_ spp. and _Ledum Groenlandicum_)
4. Continuous cover of scrub (mainly _Betula_ spp.)
5. Sphagnum, mosses and sedge; scrub scattered or absent

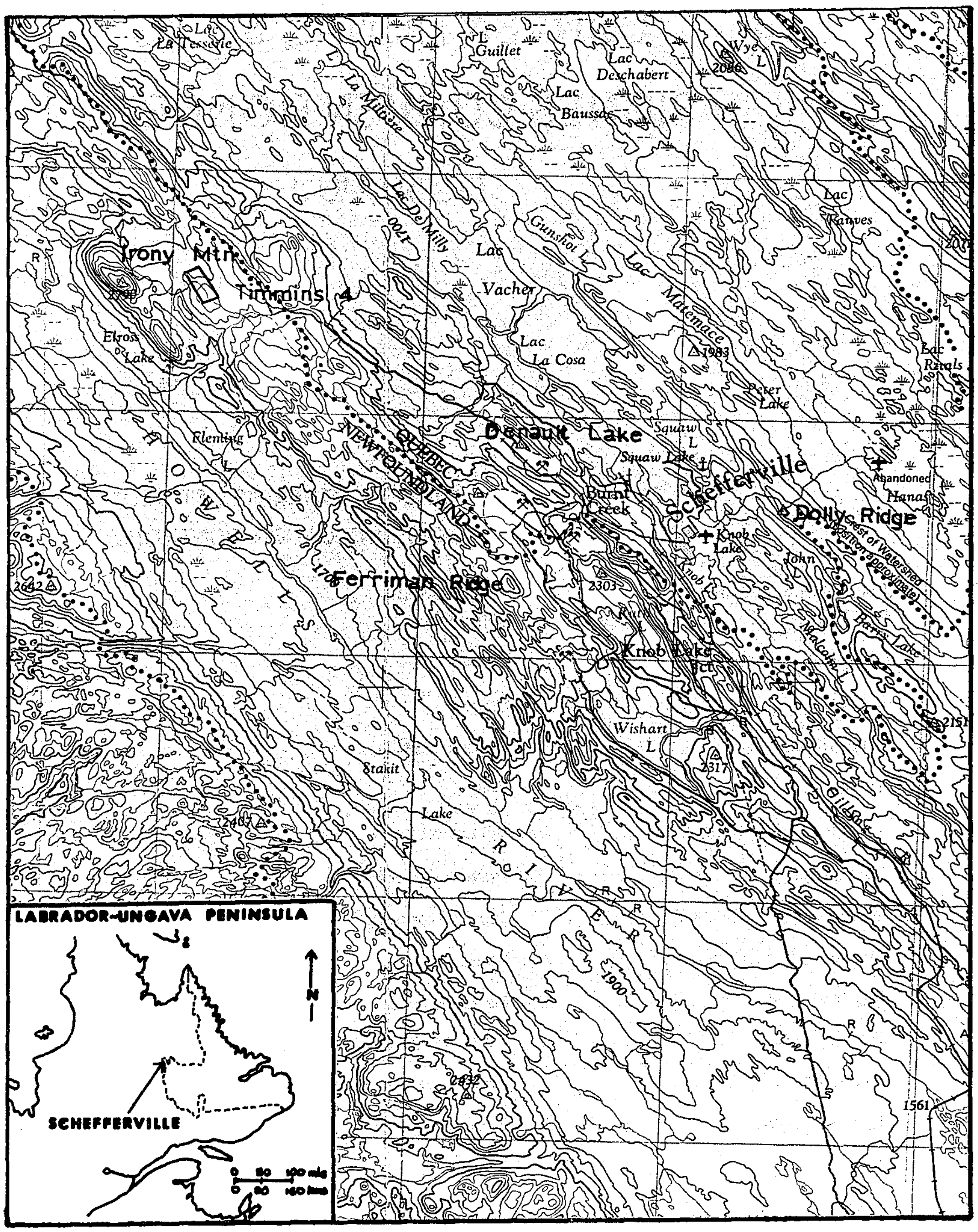


Figure 1. Location Map

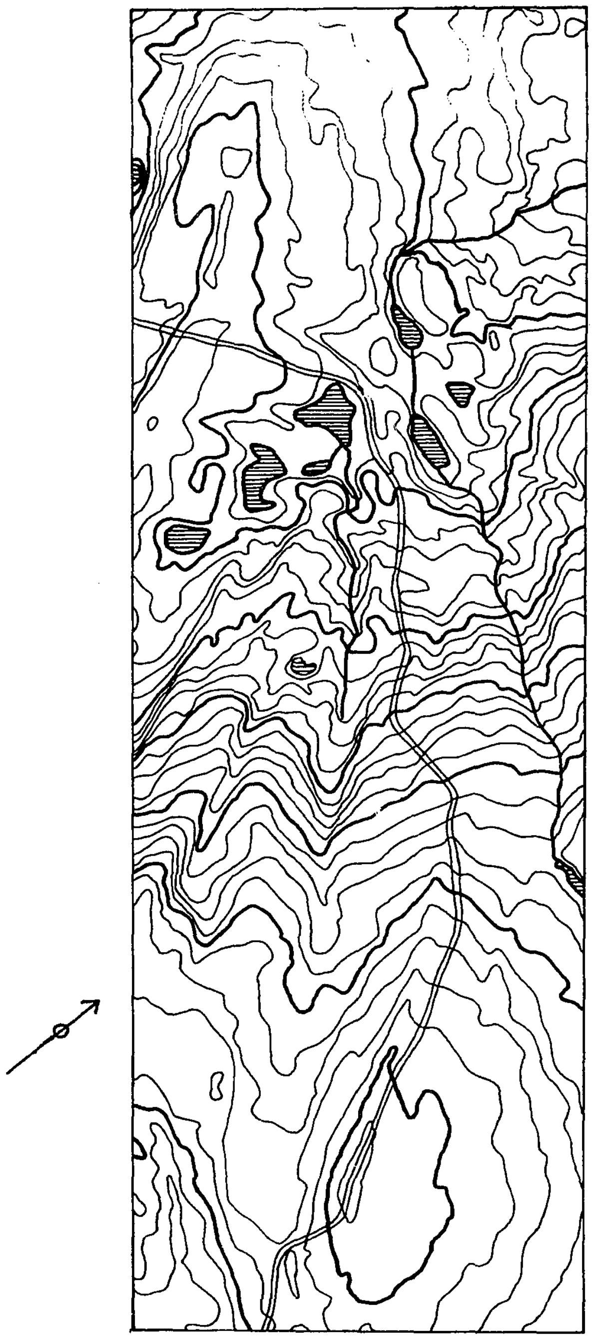


Figure 2 Topography Timmins 4

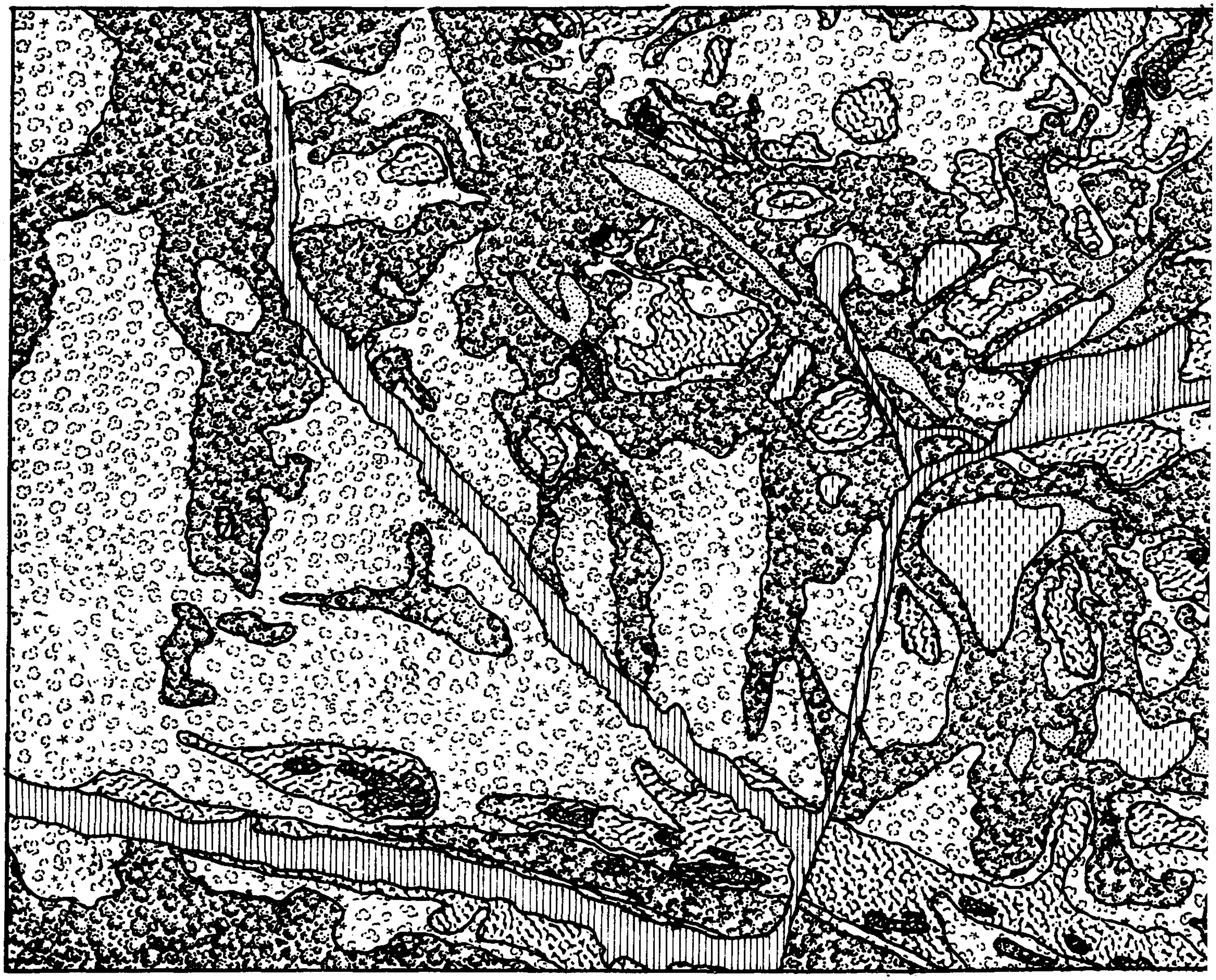
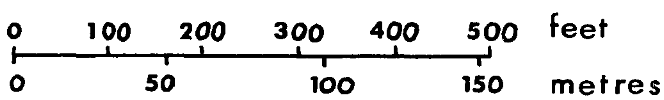
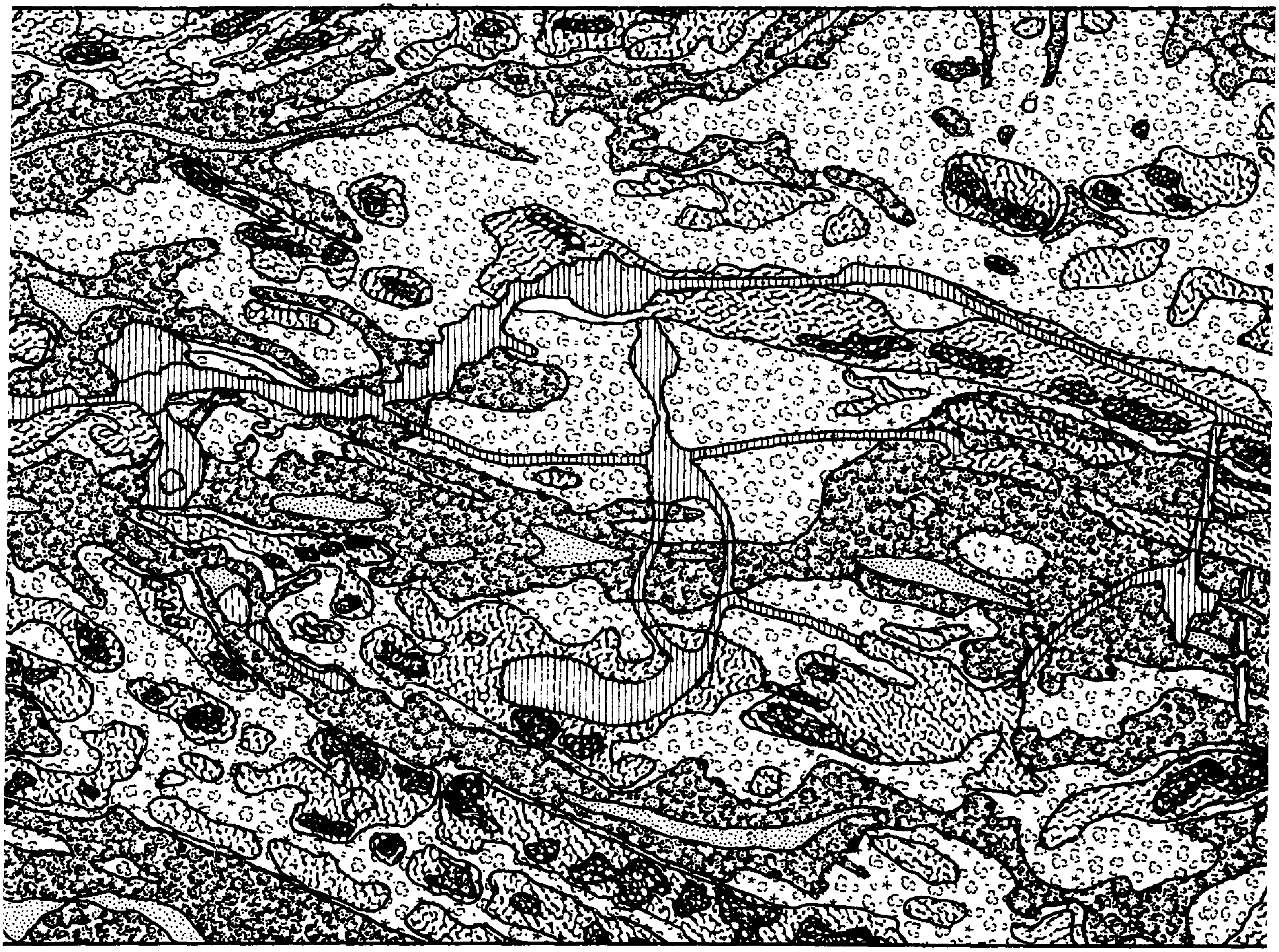


FIGURE 3 VEGETATION TIMMINS 4





LEGEND

BARE ROCK OR SOIL

DISCONTINUOUS VEGETATION

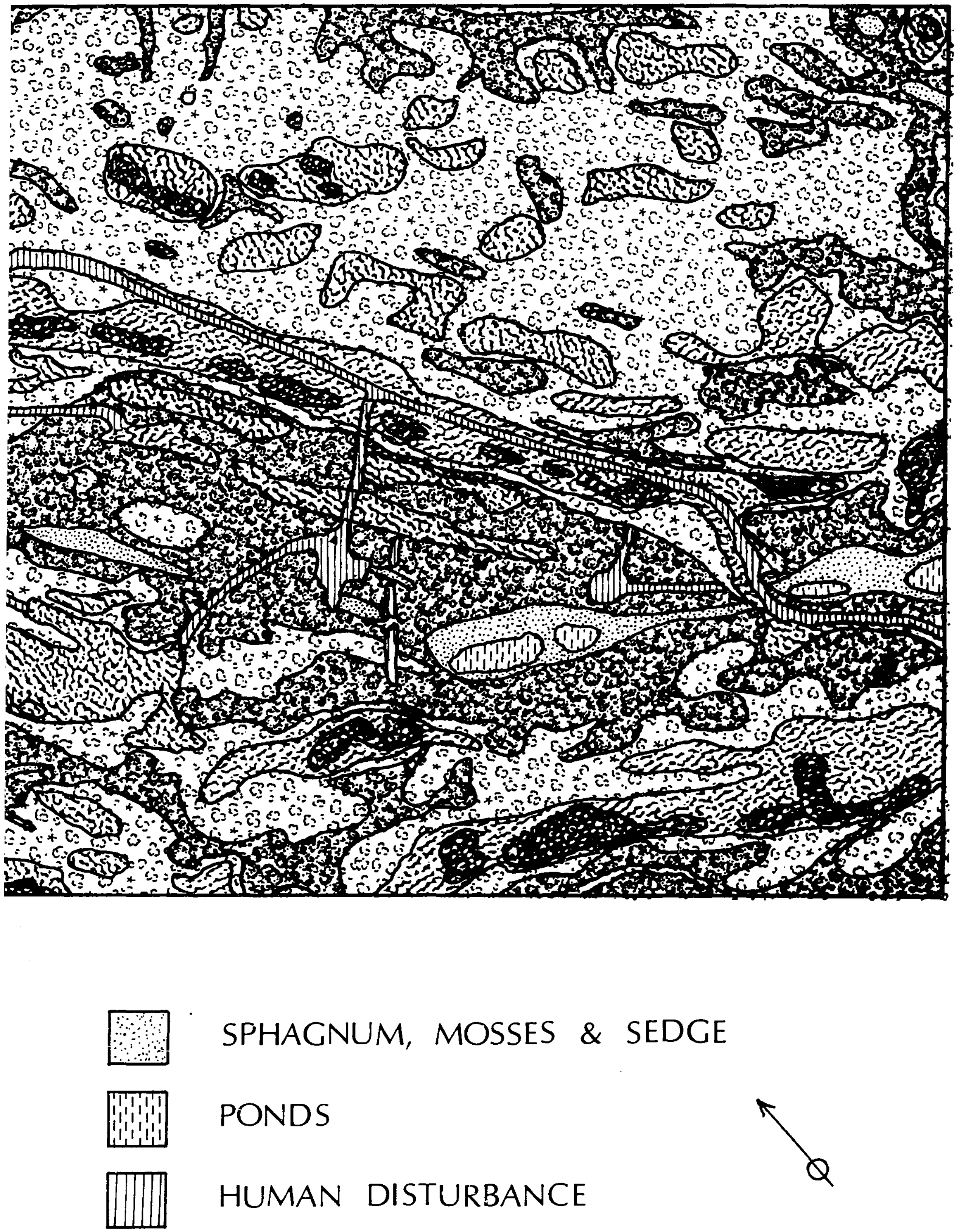
LICHEN WITH SCATTERED WOODY PLANTS

CONTINUOUS COVER OF SCRUB

SPHAGNUM,

PONDS

HUMAN DIS


SPHAGNUM, MOSSES & SEDGE
PONDS
HUMAN DISTURBANCE

The general occurrence of this sequence of vegetation classes is frequently seen in a transect from a ridge crest to a valley bottom. Trees are almost totally absent in the more elevated parts of the study area but become more common in the northwestern portion. The area is located at the local timber line and the trees within the area are stunted and procumbent on the more exposed parts. Wind gaps (Hustich, 1954; Fraser, 1956) are common and the brush vegetation near ridge crests shows a very uniform height in contrast to the brush near valley bottoms. Both of these features are possibly associated with the patterns of snow accumulation.

4. Climate

Hare (1950) provides an excellent analysis of the general characteristics of the climate of the Labrador - Ungava peninsula. It is still the main source of information available and forms the background for subsequent analyses of regional (Barry, 1959) and local (Tout, 1964) climate of the Labrador - Ungava peninsula and the Schefferville area.

The circulation patterns of the Labrador - Ungava peninsula are determined by strong pressure gradients which, especially in winter, are controlled by the Icelandic low and the high pressures over Greenland and northwest Canada. The area is affected by depressions in all seasons. These generally move in over the peninsula from southwest or south along the polar front, and can be of Central American (Gulf), Pacific or American East Coast origin. These depressions generally pass south of Schefferville in winter and move northeast. This gives a pattern of snowfall with winds of easterly components and strong winds from northwest behind the depressions. Frequently depressions slow down after entering the peninsula and tend to remain over eastern Labrador or along the east coast for considerable periods of time, whilst slowly filling. Another relatively frequent depression track is from west or northwest. These lows are generally not very deep and only give light precipitation because they are not of polar front origin. The winds are generally from southwest or west.

The climate of the Central Labrador - Ungava region is characteristically cloudy and windy with an average annual precipitation of just over 700 mm of which approximately 45 per cent is snow. The summers are short and cool, with few sunshine hours despite the latitude. Two summer months, July and August, show an average temperature above 10° C. Mean monthly temperatures from October to May are below freezing, and during those months the days with average temperatures above 0° C are few. Commonly mean monthly temperatures from December to March are below -15° C. The average annual temperature at Schefferville is close to -4.5° C.

Wind speeds average close to 5 meters per second for every month, although in September - October the averages tend to be about 10 per cent higher. In gusts, winds of more than 30 meters per second have been recorded, and hourly means of 15 - 20 meters per second or more occur fairly often in connection with deep lows. There is a strong prevalence of northwest winds, most marked in winter. However, the passages of frontal systems are often accompanied by short periods of strong winds from other directions. Wind patterns may vary considerably from year to year depending on variations in the large-scale circumpolar circulation patterns, which determine the low-pressure tracks. This variation in wind patterns from winter to winter has a significant influence on the snow accumulation patterns for the particular year (Matthews, 1962; Rogerson, 1967). Similarly, the actual snowfall amount is influenced by variations in the general circulation (Hare, 1951).

The snow falling during winter is generally dry, since the temperature is well below freezing from November to March. The snow is therefore highly susceptible to erosion and transport by the strong winds accompanying the depressions. The number of days when blowing snow has been recorded in Schefferville is high, and ranges from 26 days in 1965-66 to 75 days in 1958-59. Freezing rain or drizzle can occur in combination with deep lows centered along the east coast of the peninsula, but in general liquid precipitation during winter is not common.

Not much is known about variations in the microclimate of the Schefferville area. Davies (1962) compared winds and temperatures at Ferriman Ridge and Garage (Figure 1) with those of the weather station in Schefferville and found that the Ferriman site had about 2 meters per second higher wind speeds, and about 2^{o} C lower temperatures, and also temperature variations showed a lesser amplitude. Dyke (1967) made periodic measurements of wind speeds in different ridge locations and found about the same difference as was recorded by Davies. Recent wind data from mid February 1971 to the end of May 1971, from an anemometer at 2 m height at Timmins 4 and an anemometer at the same height at Schefferville indicate that the wind velocity at Timmins 4 is 80 per cent higher. Respective average windspeeds for the period of measurement were 5.7 meters per second at Timmins 4 and 3.2 at

Schefferville (J.L. Franks, pers. comm.). However, wind velocities measured at low level are strongly influenced by the surrounding terrain and are therefore not necessarily comparable.

Tout (1964) suggests that the elongated ridges would have a funneling effect on the wind. It is possible that this would produce a less pronounced prevalence of northwest winds in ridge locations, but there are no measurements of wind directions available to test this hypothesis.

There are significant differences in microclimate between the experimental site and the Schefferville townsite. The most significant differences with respect to the present study are the differences in wind speed and direction. The greater wind velocities increase the amount of redistribution of snow by wind. Lower average temperatures at Timmins would tend to make the snow more susceptible to redistribution and could also influence the early winter snowfall amounts, in that rainfall at near 0^{o} C in Schefferville may be solid precipitation at Timmins. However, this study does not attempt to link snow distribution patterns to weather variables in more than a very general way, and for this purpose the records from the Schefferville weather station are considered sufficient.

5. Weather in 1968-69

Figure 4 shows the variations of the most important weather variables through winter. The association between snowfall and high winds is apparent and shows that most of the redistribution took place during and immediately after snowfall when loose snow is available. Blowing snow was recorded on 39 days during the winter. The average

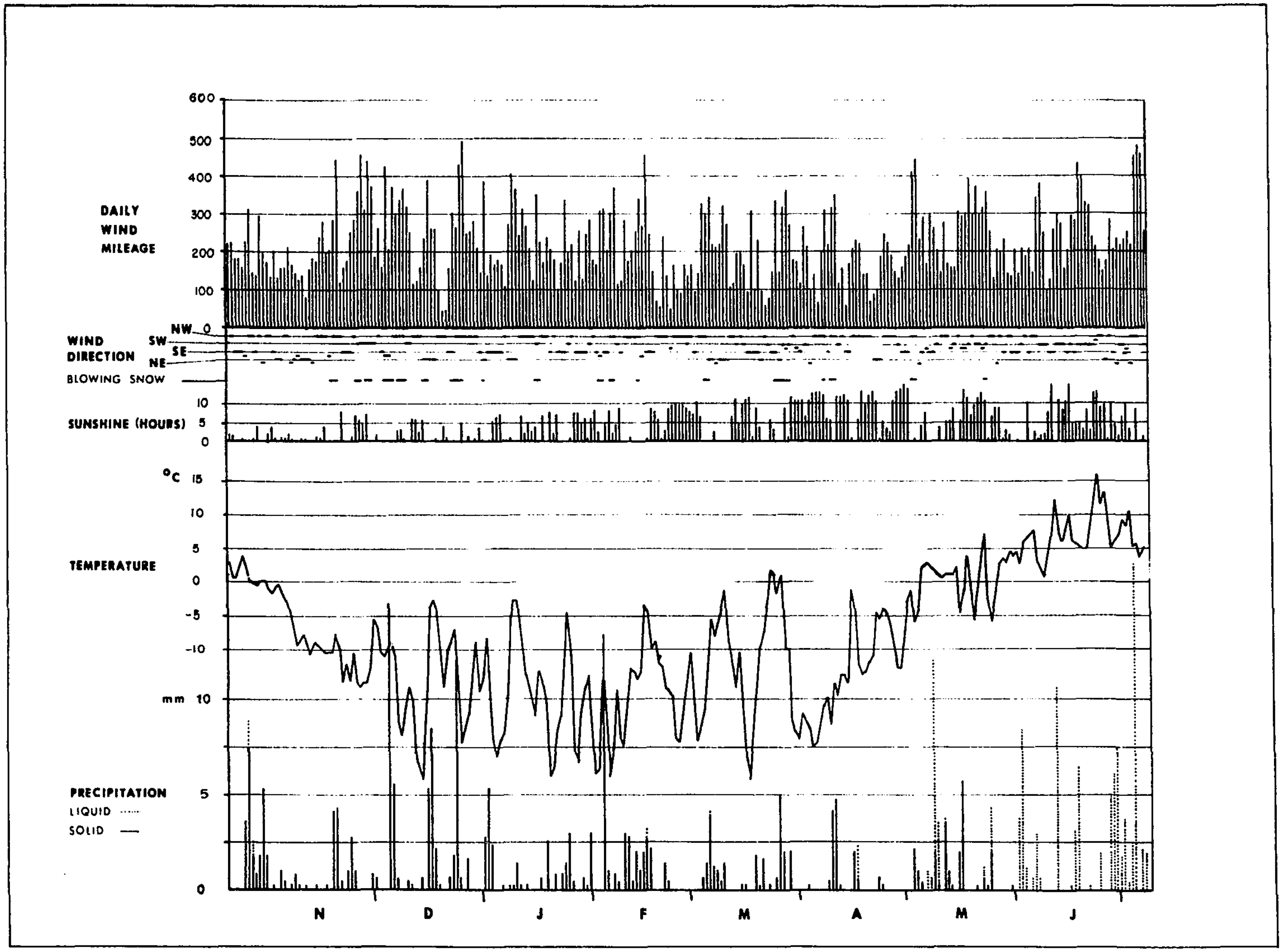


Figure 4 Weather in 1968 - 1969

daily temperature remained below freezing up to late March, and although freezing precipitation was encountered on several occasions the maximum amount for any one occasion was only 2 mm which is not sufficient to form wind resistant crusts. The number of sunshine hours was very low up to mid February when a period of cold, clear weather occurred. The sunniest month was April, whereas the melt period was relatively cloudy.

6. Research on the spatial distribution of snow near Schefferville

Considerable effort has been put into snow studies in the vicinity of Schefferville since the start of the operations of the McGill Sub-Arctic Research Laboratory. The studies undertaken up to 1964 have been reviewed by Gardner (1966).

The influence of wind on the accumulation of snow in a muskeg area was studied by Matthews (1962) who found that different patterns of accumulation were produced by different winds. Variations in snow accumulation in ridge terrain as compared to the accumulation in the more sheltered woodland in the Schefferville vale were investigated by Gardner (1964) who found greater variations of the snowcover on ridge areas. In 1962-63 the present Schefferville snow course was established which provides a continuous, comparable record of snow data up to the present date (Harrison, 1963). Comparable snow maps for the vicinity of this snow course were made in March 1965 and 1966 (Cowan, 1966; Adams and Findlay, 1966; Adams et al, 1966; Rogerson, 1967). These maps show a similar general pattern of snow depth and water equivalents from year to year although the absolute values differ. There appears to be a relationship between cover type and snow accumulation, and both snow depth and water equivalents differ significantly from one cover type to another (Adams et al, 1966).

Inventories of the snow cover within the Knob Lake drainage basin were initiated in 1964-65 (Adams and Findlay, 1966). Snow depths and water equivalents were related to cover type to produce a basin estimate of the water content of the snowpack. The survey was repeated in 1965-66 (Rogerson, 1967). Subsequently similar surveys (unpublished) have been made in 1968-69 by Granberg, 1969-70 by Price and Petch, and 1970-71 by Nicholson (M.S.A.R.L. files).

The snow maps of the snowcourse area and the basin snow surveys have provided valuable information about the catch efficiency of the Nipher shielded snow gauge, which is standard equipment for Department of Transport meteorological stations throughout Canada (Adams and Findlay, 1966). Figures of up to 45 per cent underestimate by the gauge have been given (Rogerson, 1967, p.91). There is considerable difference in the variability of the snowpack from year to year depending mainly on the influence of wind. In windy years the depth variations are greater than in years of greater snowfall but less wind (ibid.).

The patterns of snow accumulation on lakes are of great importance in the formation of white and black ice. Snow surveys on lakes in conjunction with basin snow surveys and ice studies show a greater accumulation on the southeast parts of the lakes in the Schefferville area (Archer, 1966; Adams et al, 1966).

Mapping surveys of depth variations in ridge terrain have been carried out in conjunction with studies of periglacial phenomena (Roy, 1963) and permafrost research (Barnett, 1963). Roy's survey was carried out in the Ferriman area in 1959-60 and Barnett repeated the study in

1961-62. In addition, Barnett similarly investigated the Denault Lake area. Comparing the two Ferriman surveys of 1960 and 1962 he found little correlation between the two maps, but attributed this to errors arising from varying density of depth readings between the two surveys and differences in actual snow accumulation due to greater snowfall in 1959-60. In his analysis of the differences in accumulation patterns between the Ferriman and Denault areas, he found that in the more exposed Ferriman area, topography has a greater influence on snow accumulation patterns than in the Denault area, where the more abundant vegetation accounts for most of the variations in snow depth.

Throughout the literature produced at Schefferville the strong redistribution of snow by wind has been emphasized. Obstructions to wind erosion in the form of ice layers have been reported in some years (Cowan, 1966; Gardner, 1966, p. 18) but are not common to all winters (Adams et al, 1966; Rogerson, 1967). Time profile studies indicate a rapid initial decrease in snow depth after snowfall due to redistribution of snow by wind which compacts the snow. The subsequent depth decrease, however is very small due to the low temperatures in the Schefferville area (Adams et al, 1966).

Two previous reports have been written on snow studies at Timmins 4 (Granberg and Thom, 1970; Thom and Granberg, 1970).

CHAPTER II

A DISCUSSION OF REDISTRIBUTION OF SNOW BY WIND

1. Some important properties of the snow cover

Snow accumulation is generally measured in terms of the water equivalent of the snowpack. It may be represented by the following balance (Mellor, 1964):

Net accumulation = precipitation + snow blown in + condensation - snow blown away - evaporation

The bulk or average density of a snowpack relates the depth of the snow to the water equivalent. The initial density of snow is firstly a result of the characteristics of the snow crystals, which may vary between different storms (Nakaya, 1954; Mason, 1958). Secondly it is a result of the fragmentation of snow crystals due to repeated impacts with the snow surface during drifting (Mellor, 1964; Kotlyakov, 1961). These fragments pack more closely together than the original crystals. Once deposited the density is further increased by metamorphic processes (Bader et al, 1939), melt, liquid precipitation and compaction of the snow under its own weight (Bader, 1953; Anderson and Benson, 1963). The densification of snow is temperature dependent in two ways. Metamorphic processes act more rapidly at temperatures near the freezing point and the densification is therefore more rapid in areas with warmer winters. In cold environments the density of the snowcover is greater with lower average monthly temperature and higher average monthly wind velocity (Bilello, 1958), but generally does not increase much through winter, apart from a relatively commonly occurring initial period of rapid density increase. In a wind-free situation this could be attributed to difference in crystal size / shape relationships of the falling snow (Mellor, 1964).

Measurements by Kotlyakov (1961; as referred to in Mellor, 1964) show that there is a relationship between wind velocity and the density of newly deposited snow, so that wind velocities of 20 m sec^{-1} may produce initial densities of .4 or more. Drift transport is temperature dependent in the range of 0 to -7^{o} C, decreasing with increasing temperature (Ôura, 1967, p. 1112), and this will affect snow densities if the temperatures reach this range during snowfalls (see Figure 4).

The hardness of the snow surface is related to the size and shape of the surface particles and the amount of cementing between the particles. Crusts may develop on the surface or within the snowpack due to melting. Liquid precipitation and condensation from the atmosphere or from within the snowpack may increase the cementing between surface particles. Sintering (Kuroiwa, 1962; Kuroiwa et al 1967) may be a very important process hardening the surface particularly if the latter consists of wind-blown particles. The logarithm of hardness (kg cm^{-2}) appears to be linearly related to the density of the snow cover (Bilello, 1958) but the relationship is not very close.

The erodibility of snow is closely associated with the hardness of the snow surface and there is a linear relationship between the hardness (kg cm^{-2}) and the wind speed necessary for eroding that snow (Kotlyakov, 1961 as referred to in Mellor, 1964).

2. The surface wind

Airflow of sufficient strength to dislodge snow particles from the surface is turbulent. The velocity, u, of the air at any level, z, above a plane surface may then, according to Prandtl's theory of turbulent flow (Brunt, 1934; von Karman, 1934), be expressed by:

$$u = 5.75\, u_* \log_{10} \frac{z}{k}$$

where the shear velocity $u_* = \dfrac{0.174}{\log_{10} \frac{z}{k}}$

k is the roughness height or the zero intercept of the logarithmic wind profile

and 5.75 is a factor of proportionality relating u_* to the rate of increase in velocity with the log height.

The drag force, τ_o, per unit area parallel to the flow is given by

$$\tau_o = \rho u_*^2$$

where ρ is the density of the air. This force acts parallel to the direction of the flow and is composed of form drag and skin friction. For a homogeneous plain this force is constant for any given wind velocity at a constant height. It is an expression of the transfer of momentum between the airstream and the layer at the ground surface where the velocity is 0 according to extrapolation from the logarithmic wind profile. The air is not necessarily still standing at the level of k. Investigations by Ôura et al (1967) indicated scalar velocities as high as 0.3 m sec^{-1} at 0.5 cm below a snow surface. The height k is thus a characteristic of the roughness of the particular surface and is independent of wind velocity. It may be seen as a direct measurement of the general retarding effect a particular surface has on the airstream, and for a constant velocity at a certain height, it determines the vertical velocity gradient.

The value of k is used for characterizing the roughness when the roughness is uniform. Prandtl found that the value of k is approximately 1/30 of the height of the surface perturbations causing the roughness. The value of k changes when drift transport takes place (Bagnold, 1941) to a value determined by the velocity of the wind and instead of being focused at u = 0, it becomes focused at $u = u_t$ where u_t denotes the threshold wind velocity for drift transport. Thus drift transport increases the energy transfer from the airstream to the ground surface if the velocity at height (z) remains constant, or conversely, if the energy supply is constant, alters the wind profile. The modified wind profile may then be expressed by

$$u' = 5.75\, u'_* \log \frac{z}{k} - u_t \quad \text{(measured at k')}$$

where k' is the threshold intercept of the logarithmic wind profile.

Even a very slight change in roughness height profoundly alters the wind profile, whereas it does not influence the downward transfer of momentum under steady state conditions. However, in rapidly varying roughness conditions, it is likely that it is the wind at some height above the surface which remains unaltered and it is the downward transfer of momentum which changes and fluctuates around a steady average value which is determined by the average roughness conditions over a large area. The transfer of momentum would be proportional to ρu_*^2 near the ground. Since it is this stress which is mainly responsible for the erosion at the surface, changes in surface roughness produce variations in the amount of snow eroded and thus variations in the patterns of snow accumulation.

The Reynolds number relates inertial forces to viscous forces.

It is usually written: $R = \frac{\rho u l}{\mu} = \frac{ul}{v} = \frac{\text{Inertia force per unit area}}{\text{Viscous force per unit area}}$

where μ is the coefficient of dynamic viscosity

v is the kinematic viscosity (μ / ρ)

and l is a characteristic length.

The flow around geometrically similar obstacles behaves similarly if R remains constant even if l or u is changed. The characteristic length may for example be the width of a tube or the diameter of a particle. At a value of R = 1 (inertial force = viscous force) the flow changes from laminar (for R less than 1) to turbulent (for R greater than 1). From being laminar and almost symmetrical around obstacles at a very low Reynolds number, it becomes very irregular with large variations in the instantaneous velocity, with boundary layer separations occurring behind and at times in front of obstacles at higher Reynolds number. The Reynolds number is a dimensionless quantity describing the state of the flow and gives information about the stress relationships within the flow.

3. Transport of snow by wind

Bagnold (1941) distinguishes between three modes of transport of sand grains.

1. Surface creep when particles slide or roll along the surface.
2. Saltation, when particles move along the surface in a series of jumps, rebounding from the surface or ejecting other particles. Saltation is also important in maintaining surface creep.

3. Suspension, when the particles are affected by the turbulence of the airstream.

These three modes of transport have been observed also in the snow drifting. The limits between the different mechanisms are not sharp and the proportional contribution of each mode to the total transport depends on wind velocity. Saltation is always predominant, whereas surface creep ceases when the wind reaches a velocity of 10 m sec^{-1} measured at a height of 5 cm above the snow surface (Dyunin, 1967, p. 1069).

The distance travelled by a particle in one path is related to the height to which the particle rises into the airstream and the wind velocities at the different levels through which the particle passes. Thus in the case of surface creep the distance is in the order of a few centimetres, saltation gives a path length in the order of a few centimetres to a few metres. For particles travelling in suspension the limiting distance is theoretically set only by the distance a particle can travel before it evaporates. In practice, however, the average distance travelled is in the order of a few metres to a few hundred metres. The lower the Reynolds number of the particle, the more likely is it to be affected by the turbulence of the air and deviated from its original path. There is therefore a decrease in particle size with height (Budd, 1966).

The drift density n_z, is defined as the concentration of snow water equivalent (g m^{-3}) at the level z, (Budd et al, 1966). The drift flux F_z is then defined by $F_z = n_z u_z$. The total drift transport Q is obtained by integration of the drift flux over the height range where it occurs. The total drift content N of a vertical column is obtained similarly.

The theory of steady state turbulent drift transport (Shiotani and Arai, 1953; Loewe, 1956; Mellor and Radok, 1960; Dingle and Radok, 1961; Budd, 1966; Budd et al, 1966) treats the vertical transport of particles in analogy with momentum flux. This momentum flux would have the form of a shearing stress

$$\tau = K \, \partial u_z / \partial z$$

where K is the eddy viscosity

$$K = 0.4 \, (\tau / \rho)^{\frac{1}{2}} \, z$$

where the constant 0.4 is von Karman's constant.

The assumption is that the turbulence transporting the momentum also creates the steady upward transport of snow required to balance the steady, gravity dependent fallout of drift snow with the fall velocity w. The balance relation of steady state snow drift is

$$K \, \partial n_z / \partial z = -w n_z$$

For snow of constant fall velocity integration over a range z_1 to z_2 gives the ratio of drift densities at the two levels as

$$n_{z_1} / n_{z_2} = (z_1/z_2)^{-w/0.4u_*}$$

A full account of this theory is given by Budd (1966) and Budd et al, (1966). The theory was modified by Budd (ibid) to account for systematic variations in particle fall velocity with height and conforms well to measurements obtained in the field (Budd et al, 1966).

The total snow transport in relation to wind velocity at some level above the surface has been found to be some power function of the wind velocity. Komarov (1954) quotes a formula by Mel'ink derived from the assumption that Q is proportional to the force of the wind, the formula being

$$Q = Cu^3$$

where Q is the total snow transport in a layer 2 m high in g cm^{-1} min^{-1}

u is the wind velocity measured at 11 m

C is a coefficient equal to 0.0129

another expression given in the same paper is

$$Q = 0.0065\ u^{3.5} - 0.4$$

with u measured at the height of 1 m.

Budd et al (1966) found the following relationship for the drift transport in the lowest 300 m of the atmosphere

$$\log Q = 1.1812 + 0.0887\ u$$

where u was measured at 10 m.

Bagnold (1941) concludes that in the case of sand in saltation the flow Q varies as the cube of $u_*^!$ and also varies as the cube of the excess of wind velocity over and above the constant threshold velocity at which the sand begins to move. For sand in suspension Q varies as u_*^3.

It is apparent that the conditions under which the different relationships were derived may have influenced the results but there is a good general agreement between the different results. It seems reasonable to accept an approximate relation of $Q = c\ u^3$.

The bulk of the transport takes place in the lowest layers of the airstream. Komarov (1954) gives a value of 89.2 per cent of the total transport taking place in the layer 0 - 10 cm above the surface. The total was measured for a layer 2 m high. Ôura (1967) suggests that 90 per cent of the total transport takes place in the lowest 20 cm of the atmosphere.

The low level at which most of the transport takes place implies that the average distance travelled by a snow particle in each movement is short. Therefore if there is a sudden decrease in wind velocity the deposition of snow from the airstream will take place within a short distance.

4. Accumulation of snow

a) Previous studies

The accumulation of wind blown snow in terrain has been described by Shumskii (1964, p. 238):

> "As a rule, during a snow storm, the layers of air near the ground receive a boundless number of snow particles from above or from the snow cover and transport as many of them as can be kept in a suspended state at a given wind velocity and particle size and shape. When the rate of motion increases, more particles are transported, and when it decreases, some snowflakes settle to the surface. Consequently, the acceleration of the wind and not its absolute velocity is the decisive factor in the distribution of snow which is falling from the atmosphere or being transported as blowing snow over the elements of the relief. On a level area, where the average wind velocity is constant with respect to the neighboring areas, the snow removed by the wind is compensated for by snow brought in by the wind. Snow is redeposited, but the average depth remains constant due to drifting. However, the snow cover is blown away from the convex elements of the relief, where the lines of air flow converge and the wind velocity increases, while the snow accumulates and is deposited on the concave elements, where the lines of air flow diverge. Thus, in the case of dissected relief, not only are there regional differences in the depth of the snow cover as a function of the amount of precipitation, but there are very great local differences in the thickness of the entire snow cover and of individual layers deposited by winds of different directions."

The net accumulation of drifting snow was calculated by Komarov (1954) using the relationship:

$$A = Q_1 - Q_2 = C(u_1^3 - u_2^3)$$

This showed good agreement with observational data. A different approach was used by Radok (1968) using the theory of turbulent drift transport but the method did not give a very accurate estimate of the accumulation patterns. In neither of the two approaches were changes of wind patterns due to snow accumulation accounted for. The aerodynamic change of an obstacle due to snow accumulation cannot be ignored since it determines the way in which the snow will continue to accumulate around the obstacle. The results obtained by Komarov (ibid) were from a study of effects of multiple row snow fences, and it is probable, although not stated in the paper, that the study considered the initial stages of accumulation around recently erected fences. Karrer (1921) investigated the accumulation of snow and found that the snow cover may be regarded as a deformable body in a moving fluid and thus accumulates so as to minimize the resistance between airflow and the ground surface.

From Mel'niks formula it may be seen that a change in wind velocity with a factor of two would produce a change in the rate of transport with a factor of eight. The magnitude of the net erosion or deposition would thus be a function of both the initial velocity and the acceleration or deceleration of the airstream. Where accelerations occur there is erosion. Where decelerations occur there is deposition. Erosion will lower the surface and therefore reduce the tendency to accelerate the airstream; whereas deposition will act in the opposite fashion. Therefore as long as the snow surface is homogeneous and easily eroded, the snow would tend to accumulate in a fashion so that accelerations and decelerations of the airstream are avoided. However, the snow supply is limited and this imposes restrictions on the validity of this theory. The supply is limited where the rate of erosion is sufficiently great to erode the newly deposited snow so that old hardened snow, or the ground surface, becomes exposed. In such locations relatively little erosion can take place regardless of acceleration of the airstream. Another factor limiting the snow supply is areas where the surface stress is insufficient for erosion of snow. Downwind of such areas the balance between erosion and deposition weighs towards erosion for some distance before the "limit length of snow storm" has been reached. This expression, proposed by Dyunin (1967), describes the phenomenon that it takes some time before an equilibrium between erosion and deposition is reached when the airstream enters a snow field, even if there is no increase in wind velocity. Bagnold (1941) ascribes this effect to the longer time required for an energy equilibrium to be established when a balance of particle movements and impacts is involved. In the following sections the term "snow fetch" will be used to denote the distance over which erosion has been possible.

b) Accumulation of snow at a snow fence

The literature on accumulation at snow fences is plentiful and a large number of descriptions of drift formations around fences of different constructions are available. The literature on snow fences has been reviewed by Mellor (1965). The sequence of drift buildup around fences of similar construction does not vary much between studies undertaken in widely different areas. Ultimately the drift buildup reaches a stage when the fence is saturated with snow. The ability of the fence to collect more snow is greatly reduced after this stage has been reached.

Figure 5 shows the isotachs of airflow around a fence of 80 to 85 per cent density (after Nägeli, 1953) and the sequential accumulation of snow at a snow fence of 53 per cent density (after Komarov, 1954). The two fences have different aerodynamic characteristics since the densities are different. However, the flow patterns are believed to be relatively similar since the wind velocity in the case of the snow accumulation was probably higher than the wind velocity for the airflow study. This would tend to reduce the difference in Reynolds number.

The sequence of accumulation of snow at the fence may be described as follows:

Stage 1.

Snow drifting along the ground encounters a zone of decreasing wind velocity in front of the fence. Here erosion of snow decreases but deposition of snow already in the air continues for a short distance giving a greater increase in depth in front of and closely behind the fence. Beyond a point approximately 2 h behind the fence, the wind speed near the surface begins to increase. Erosion should therefore occur. This effect is not apparent at this stage, probably because the rough microstructure of the ground across the whole site retains a shallow layer of snow.

Stage 2.

The accumulation near the fence continues to be rapid during the period before the second measurement. A markedly greater part of

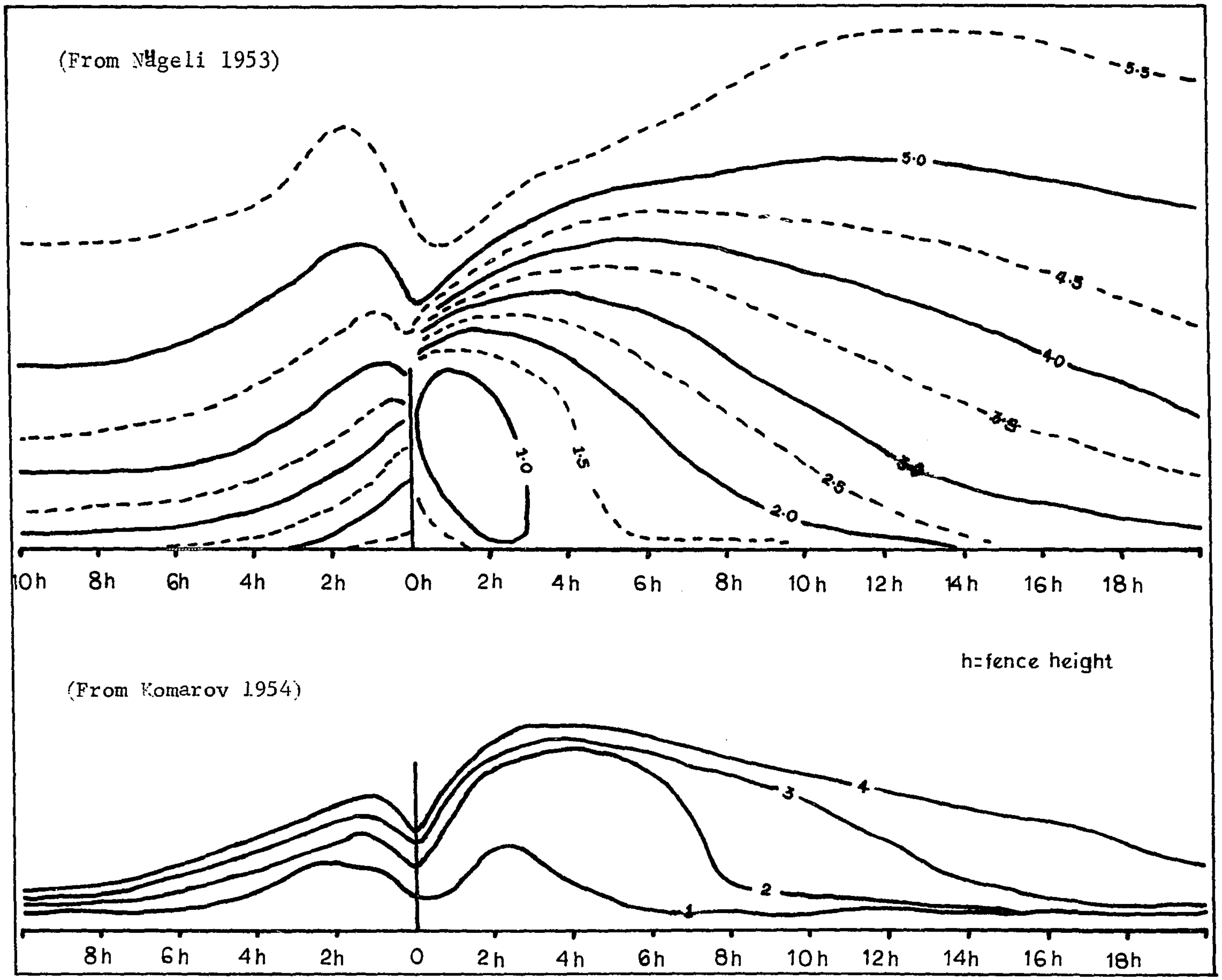


Figure 5 Wind patterns and snow accumulation at a fence

the accumulation takes place behind the fence. The lesser accumulation in front of the fence shows that the decrease in rate of erosion is less pronounced than before. The maximum decrease in erosion now occurs close behind the fence. A greater amount of erosion is now apparent in the zone of increase in wind velocity behind the zone of maximum accumulation. This is shown by the snow surface remaining at a constant level which is apparently controlled by minor surface roughness.

Stage 3.

During the period prior to the third measurement the zone of balance between erosion and deposition extends to a distance of about 4 h behind the fence. The bulk of the accumulation occurred over a short distance beyond this zone. Further downwind, beyond a distance of 20 h from the fence the snow surface remains at a constant level.

Stage 4.

This is the stage when the snow fence approaches saturation. At stage 3 the zone of balance between erosion and deposition extended to about 10 h behind the fence. The accumulation during the period before measurement 4 took place mainly beyond this distance and the accumulation was spread over a longer distance than previously. The zone of no increase in snow depth has been eliminated, indicating that the snow eroded is replaced by deposition of snow from upwind. The outline of the saturated drift is very nearly parallel to the isotach of 4.5 m sec^{-1}. This would indicate that at the stage of saturation the accelerations and decelerations at the snow surface have been largely eliminated and that therefore the balance between erosion and deposition is no longer altered by the presence of the fence.

The sequence shows the following:

1. The bulk of the accumulation takes place within a short distance after a decrease in surface stress.

2. The zone of maximum accumulation progressively moves from windward of the fence to leeward. A "smooth area" develops and maximum accumulation takes place at the windward edge of this area. In this smooth area the balance between deposition and erosion of snow remains largely unaltered from the balance upwind of the fence (left margin of the figure).

3. Erosion prevails downwind of the zone of rapid accumulation.

4. There is a close resemblance between the isotachs and the shape of the drift. However, in view of the differing experimental conditions of the wind and snow observations this may not be regarded as a definite proof that the snow surface aligns itself parallel to the isotachs.

c) The significance of snow depth

Similar shapes of drifts have been obtained with similar snow fences in studies in widely different areas (Finney, 1937; Nøkkentved, 1940; Hallberg, 1943; Shiotani and Arai, 1950; Komarov, 1954; Price, 1960). This shows that the shape of the drift formation is controlled by the roughness created by the fence, whereas water equivalents and densities of the snow may vary. It therefore appears that snow depth rather than water equivalent would be the measureable quality of the snow cover which is most closely associated with wind factors causing its spatial variations. Because it is the association between terrain roughness and snow accumulation that is being investigated, the present study concentrates on the variations in snow depth.

d) Snow accumulation and terrain roughness

In principle snow accumulation in terrain is analagous to the accumulation at a snow fence. Variations in wind stress at the snow surface cause differential snow transport and therefore an uneven accumulation of wind transported snow. The distribution of snowfall is relatively even (Dyunin, 1959).

Four main roughness factors may be regarded as important in the area of the present study.

1. Topographic roughness
2. Trees
3. Brush vegetation
4. Microroughness

The influence of microroughness such as grass, lichens, small stones etc. is demonstrated in the snow fence example above. The microroughness tends to retain a shallow overall cover of snow.

The influence of trees on the airflow has been discussed by Reifsnyder (1955). The modification of the airflow is closely related to height and crown density of the forest, and accordingly results in a lowering of the surface wind velocity. The effect of the decrease in wind velocity is apparent in a greatly increased accumulation of snow at forest edges (Hoover, 1962; Miller, 1965; Hoover and Leaf, 1967; Adams and Findlay, 1966), and in isolated stands or shelterbelts of trees (Komarov, 1954, Figure 2). Small openings in the forest tend to accumulate more snow than the surrounding forest (Hoover, 1962; Miller, 1966), whereas for larger cuttings in the forest the average water equivalent is inversely related to the size of the cutting (Kuzmin, 1960). In the latter case it is an effect of increasing wind transport from larger size openings whereas in very small openings the wind stress at the surface is below threshold for erosion. It is possible that the greater accumulation in small openings results from intercepted snow drifting from the canopy into the opening. A uniform forest greatly reduces the variation in snow accumulation in that it effectively hinders drift transport.

The effect of brush vegetation is similar to that of a forest but on a smaller scale. Whereas a forest does not usually get buried by snow accumulation, the roughness created by brush is eliminated after some time by the greater rates of accumulation in brush covered areas.

The influence of topographic roughness may be seen in snow depth maps by Roy and Barnett for the Ferriman area near Schefferville (Barnett, 1963, p. 77-78). These maps show a general trend of deeper accumulations in concave parts of the terrain and shallow snow in convex areas. The results also show marked differences in accumulation with respect to aspect between different winters. Whereas in 1959-60 there was greater accumulation on northeast facing slopes, the greater accumulations in 1961-62 were found on southwest facing slopes.

Results obtained from studies in Antarctica (Black and Budd, 1964) and Greenland (Mock, 1968) show that on ice caps the accumulation of snow can be closely predicted by use of deviations in altitude from a smoothed terrain profile and deviation in slope.

5. The sequence of accumulation of the seasonal snowcover

The accumulation of snow in terrain may be seen as analogous to the accumulation at a snow fence, in that smooth areas gradually develop where the wind-stress variations are greatly reduced. In contrast to the example of the snow fence the snow fetch is extremely variable in terrain. This influences the balance between erosion and deposition so that when the snow fetch is short, the smooth area constantly loses much of the snow deposited by snowfall. This is because little snow is drifted into the smooth areas from surrounding areas of greater roughness. The snow fetch across smooth areas varies with wind direction and also increases through time with the increase in size of the smooth areas. The greatest rates of accumulation are found at the edges of the smooth areas where there is a reduction in surface wind stress. This accumulation rate increases with snow fetch. In analogy with the accumulation behind a fence the edge of the smooth area is relatively sharp during the earlier stages of accumulation. This is because the greatest rate of deposition occurs immediately behind the point where the reduction in wind stress occurs. The smooth areas are initiated in the parts of the terrain where the wind stress at the snow surface is greatest. These zones are vegetation-free areas such as ridge crests. From these zones the smooth areas grow in all directions at rates depending on snow supply and wind. New centers of growth are initiated as the initial surface roughness is buried by accumulation. At some point in time smooth areas start to merge with one another and since this greatly increases the snow fetch the entire area becomes smooth within a relatively short time period.

After this has occurred the accumulation changes. There are no longer any areas where the surface wind shows sharp increases or decreases and as a consequence the further accumulation of snow becomes more even. Due to the general decrease in surface roughness there is an average increase in surface wind velocities through winter and the eroding capacity of the wind therefore increases through winter, resulting in a much greater mass transport. Nowhere is the surface stress insufficient to cause erosion during snow storms and therefore the balance between erosion and deposition becomes much less variable, producing a pattern of accumulation which is more even than previously.

Thus three major periods of accumulation may be expected:

1. An initial accumulation period when the wind stress at the snow surface is generally low and largely insufficient to cause appreciable erosion. Microroughness traps the snow to give a uniform cover over most of the area. Areas with little microroughness (notably the ridge crests) suffer erosion and the deepest accumulations at this stage are adjacent to such areas.

2. A transition period characterised by growing smooth areas and areas that still retain considerable local roughness. Hence there are areas of very shallow snow, cornice drifts adjacent to ridge crests, and relatively uniform accumulation in valleys.

3. A period when the variations in surface stress have been greatly reduced by snow accumulation, and the snow fetch is large. The deepest accumulations would be found in zones where the wind velocities are locally lower. Larger areas of homogeneous character now show a steady increase related to snowfall, since minor irregularities of the airflow have been eliminated.

CHAPTER III

SNOW ACCUMULATION AT TIMMINS 4

1. Snow depth sampling

The snow sampling program was primarily designed to give information about snow depth variations in different locations through winter so that these may be related to terrain roughness factors.

An aligned square grid was employed in the sampling. A regular sampling design is vulnerable to auto-correlative effects caused by periodic variations in the sample, but the effect is in this case probably small. This is assumed firstly because the alignment of the topography is not strictly parallel, transverse or at a 45 degree angle to the alignment of the sampling grid. Secondly, the spacing between successive ridges and valleys is not regular. Thirdly, only in a limited part of the area are the corrugations of the terrain parallel.

The decision to use this type of sampling grid was also influenced by the following reasons:

1. A systematic grid greatly facilitates both the sampling procedure and the computational procedure (Chapter IV).
2. A systematic grid design ensures an even spatial spread of the sample.
3. Experience from previous snow studies in the Schefferville area shows that weather conditions often make it difficult to find the sampling sites unless these are arranged in a systematic fashion. A systematic arrangement also reduces the time taken for sampling.

The sampling area has a general slope towards the northwest, which gives a bias to the sample. This has effect on the general applicability of results obtained at the experimental site.

Sample size was determined from the results of previous snow surveys at Timmins 1 (Thom, 1969). The grid was designed with 21 rows and seven columns, and with 200 ft. (61 m) interspacing between the sampling points, thus giving a sample size of 147 observations. The grid was surveyed by a team from the Iron Ore Company of Canada. Figure 6 shows the grid and the numbering of the sampling points.

Wooden stakes 240 cm long and 5x5 cm thick were painted white to avoid ablation around the stakes in spring and placed out at the surveyed points. They were marked at 5 cm intervals with distinguishing marks every 50 cm, so that snow depths could be read directly from the stakes. To overcome the problem of zeroing the scale with ground level, the gradation was painted after the stakes had been erected. A stencil was produced by drilling 6 mm holes across an L-profile of aluminum for every 5 cm. Larger size holes were used for the 50-cm marks. In the field the stencil was clamped to the stakes and the gradations were painted using aerosol-packed enamel.

Sampling of snow depths was undertaken so as to trace changes after major snow events. Twenty-five sets of depth samples were taken during the snow season. Table 1 shows the dates of sampling, the mean depths and the standard deviations of the samples.

2. Three periods of snow accumulation

Three periods of accumulation were outlined in Chapter II. These periods are caused by the transition from an initial very rough microstructure of the ground, through a period of large variations in surface roughness, to a period when the microroughness

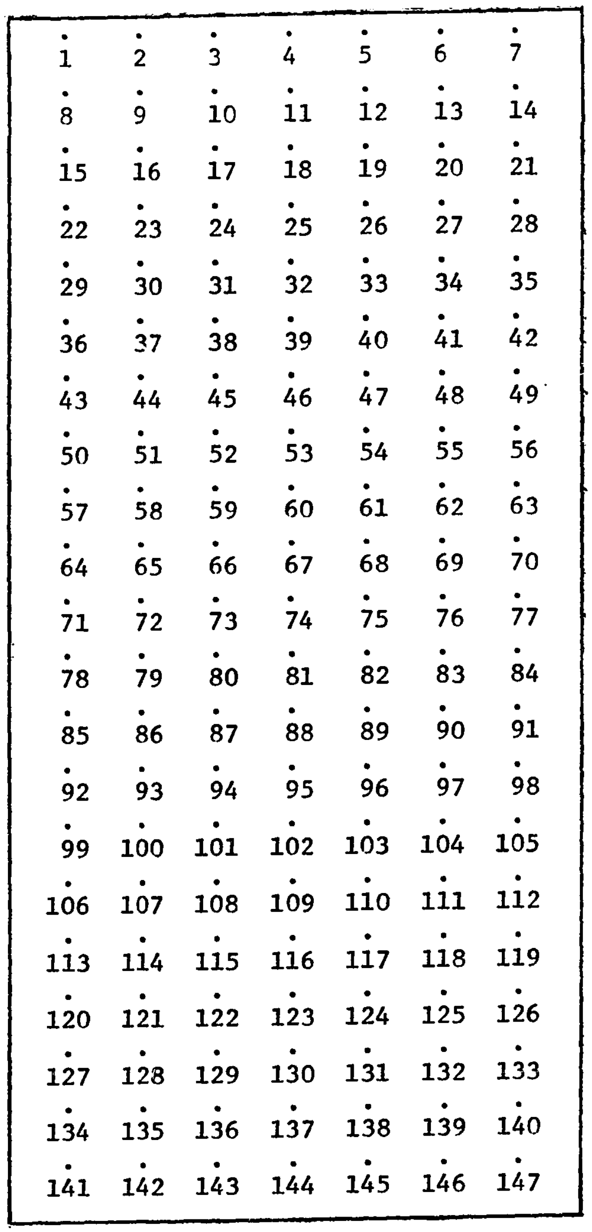


Figure 6 Numbering of sampling points.

TABLE I

DATE OF SAMPLING; AVERAGE DEPTHS AND STANDARD DEVIATIONS

TIMMINS 4, 1968-69

DATE	AVERAGE DEPTH (cm)	STANDARD DEVIATION (non-normalized)
October 27	17.20	9.15
October 31	30.42	13.98
November 12	33.26	12.51
November 24	40.25	20.58
November 28	42.08	21.86
December 4	41.54	21.22
December 14	53.21	35.98
December 21	64.62	47.89
January 17	79.03	56.82
January 27	81.60	57.63
February 9	90.60	64.53
February 16	96.99	69.65
February 25	98.79	70.26
March 13	110.46	74.66
April 21	115.17	74.44
May 8	107.26	75.06
May 14	106.44	73.63
May 21	96.95	72.76
May 28	85.71	70.93
June 4	60.06	65.94
June 11	38.59	58.76
June 18	13.99	38.47
June 25	5.21	21.80
July 2	1.46	8.92
July 9	0.26	2.21

has been eliminated by snow accumulation. The change in surface roughness would influence the variability in depth of the snow accumulated during each of the three periods so that initially the accumulation would be even, due to a generally lower amount of erosion taking place. The second period would be characterised by a spatially very irregular accumulation due to snow drifting from smooth areas and being intercepted at the margins of these areas. In the third period the accumulation would be controlled by larger scale roughness and therefore vary gradually between the extremes.

a) <u>The periods identified</u>

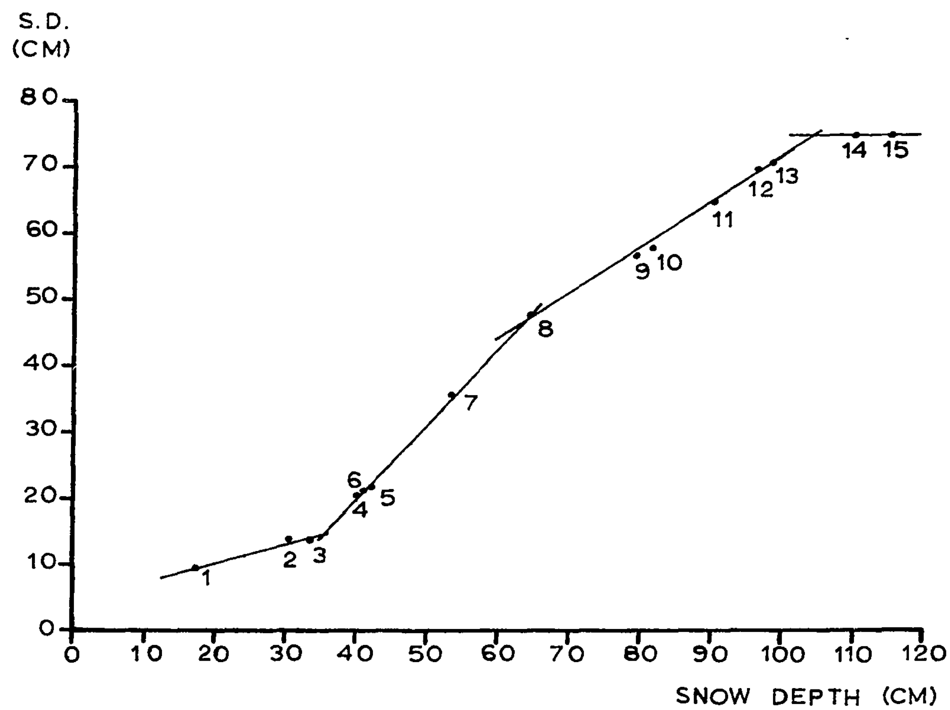
To identify the three periods the standard deviation of snow depth was plotted against the average depth for each survey (Figure 7). The initial period is shown by a slow increase in standard deviation up to November 12. The second period is indicated by a rapid increase in standard deviation, ending on December 21. The third period is shown by a less rapid increase of the standard deviation. A fourth period appears to start in early March when the standard deviation ceases to increase with increasing snow depth. At this point in time the increasing influence of solar radiation and higher daily maximum temperatures induce a general settling of the snowpack due to internal metamorphic processes acting more rapidly at higher temperatures, causing a more rapid depth decrease in zones of deep snow, and thus reducing the variability in depth. Snowfalls at higher temperatures and a more rapid development of crusts reduce the amount of redistribution taking place. The end of the third period and the period of analysis was therefore set at March 13.

FIGURE 7

STANDARD DEVIATION

VS.

AVERAGE DEPTH



b) Depth distributions for the three periods

The average increases in snow depth during the periods indicated by the standard deviations are:

1. October 27 - November 12 33.26 cm
2. November 12 - December 21 31.36 cm
3. December 21 - March 13 45.84 cm

The distributions of accumulation during these periods are shown as histograms in Figure 8.

Period 1.

In the first period the mean falls within the modal group. This group contains 36 per cent of the sample which would indicate that in a large part of the area the redistribution of snow was slight. Only in 3 per cent of the locations was the accumulation deeper than 60 cm. The extreme case was stake 144 which is situated 3 m from a negative step at the southern edge of the upper plain area. Drifting from this plain took place in limited amounts already during the first snowfall due to the only thin cover of brush and caused deep accumulation behind the step in the form of a snow cornice which reached the stake location.

Period 2.

The modal group for the second period is 10 - 20 cm. In all, 63 per cent of the area accumulated less than 20 cm of snow. A second mode, coinciding with the mean (30 - 40 cm) could indicate either that some areas were still protected from wind action or that some areas had developed where erosion and deposition due to drifting

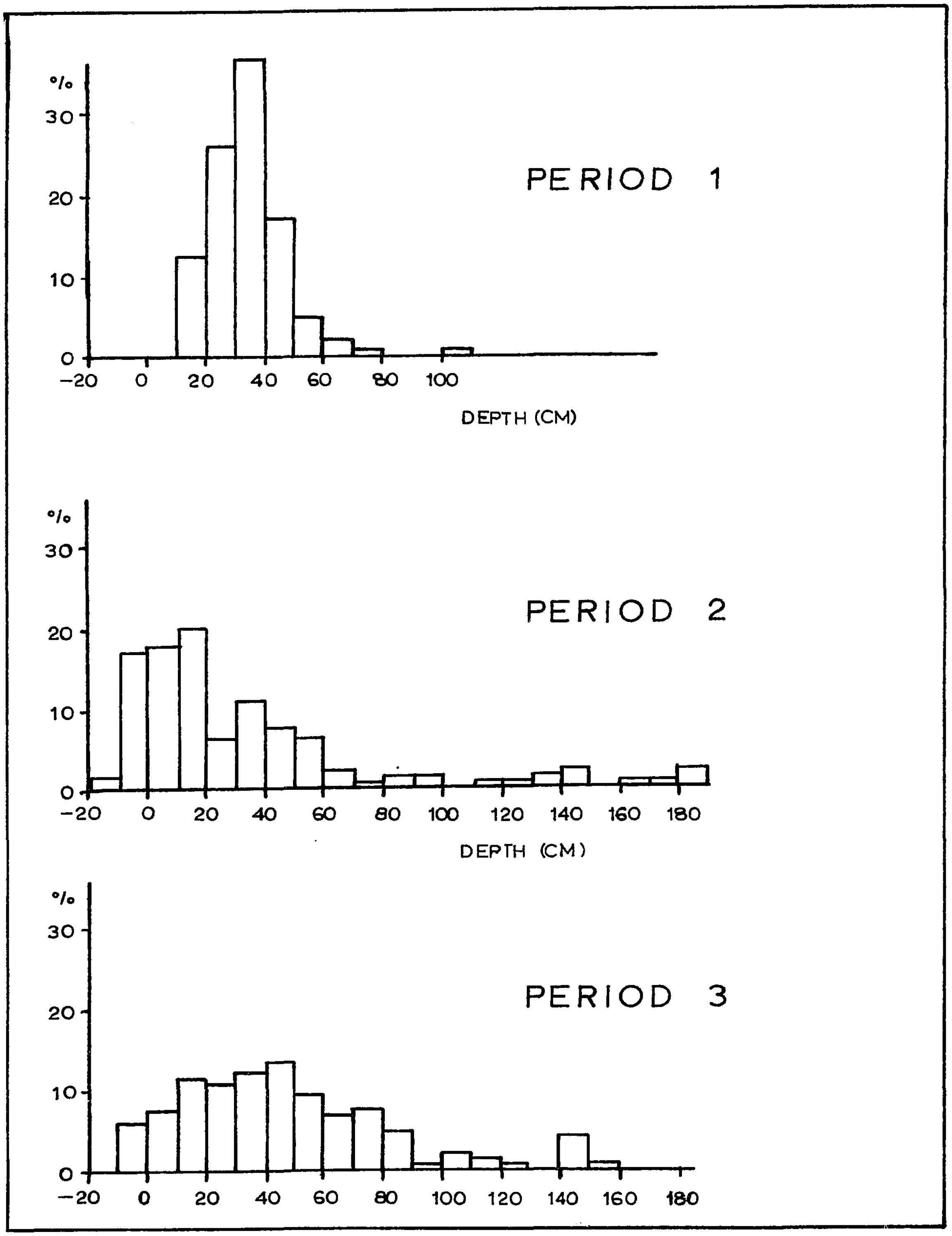


Figure 8 Histograms for depth increase during three periods

were in equilibrium. The strongly skewed distribution shows that erosion prevailed in a large proportion of the area and that very deep accumulations occurred in a limited part of the experimental site. Cornice drifts were commonly observed throughout this period along steeper valley slopes.

Period 3.

In the third period the mean falls within the modal group, showing that in 14 percent of the area there was no net loss of snow due to drifting. Forty-eight per cent of the area has depths less than the modal group, indicating that net erosion tends to be diminished. Thirty-eight per cent of the area shows an increase in depth due to drifting. Despite an average depth increase about fifty per cent greater than during the previous period, the maximum depth increase is less (154 cm). During the previous period five locations experienced deeper accumulations. During period 3 the stakes where accumulations greater than 140 cm occurred (no. 8, 15, 55, 64, 69, 71, 129) were all located at some 40 - 60 m west of prominent ridge crests. Here cornice drifts were observed in the early part of the period which would explain in part the distinctly greater rates of increase in depth. Thus whereas most of the area during the third period was devoid of zones of rapid decrease in wind stress, some such zones still remained at the beginning of the period.

The histograms show that the distribution of the snow accumulated was distinctly different during the three periods. The limits between the periods are not sharp although the standard deviations would suggest this to be the case. It is possible that lighter winds during the first period followed by stronger winds during the second period may have emphasized the differences somewhat but subsequent years' observations show that these distinct periods occur although the spacing in time varies.

c) Spatial variations in accumulation

According to the theory in Chapter II the initiation of smooth areas takes place at ridge crests and in other areas where the microroughness is slight. The smooth areas grow outward through snow accumulation at their edges. When these smooth areas are small, little snow is accumulated within them due to a short snow fetch, which makes erosion prevail over deposition. With increasing snow fetch the balance between erosion and deposition becomes more even and depth increase can occur. Once the smooth areas are interconnected the further accumulation is controlled by more gentle variations in the wind produced by large scale undulations in the considerably smoothed topography.

To investigate the spatial variations in depth increase through winter the maps in Figure 9 were compiled. In order to make the maps more comparable in terms of depth increase the three periods given in the previous section of this chapter were slightly altered to include in the first period the depth increase up to November 24 (40.25 cm). The second period is from November 24 to January 17 (38.78 cm) and the third period from January 17 to March 13 (31.43 cm). The maps are based on linear interpolation between the sampling points which was performed by the computer mapping program in Appendix I.

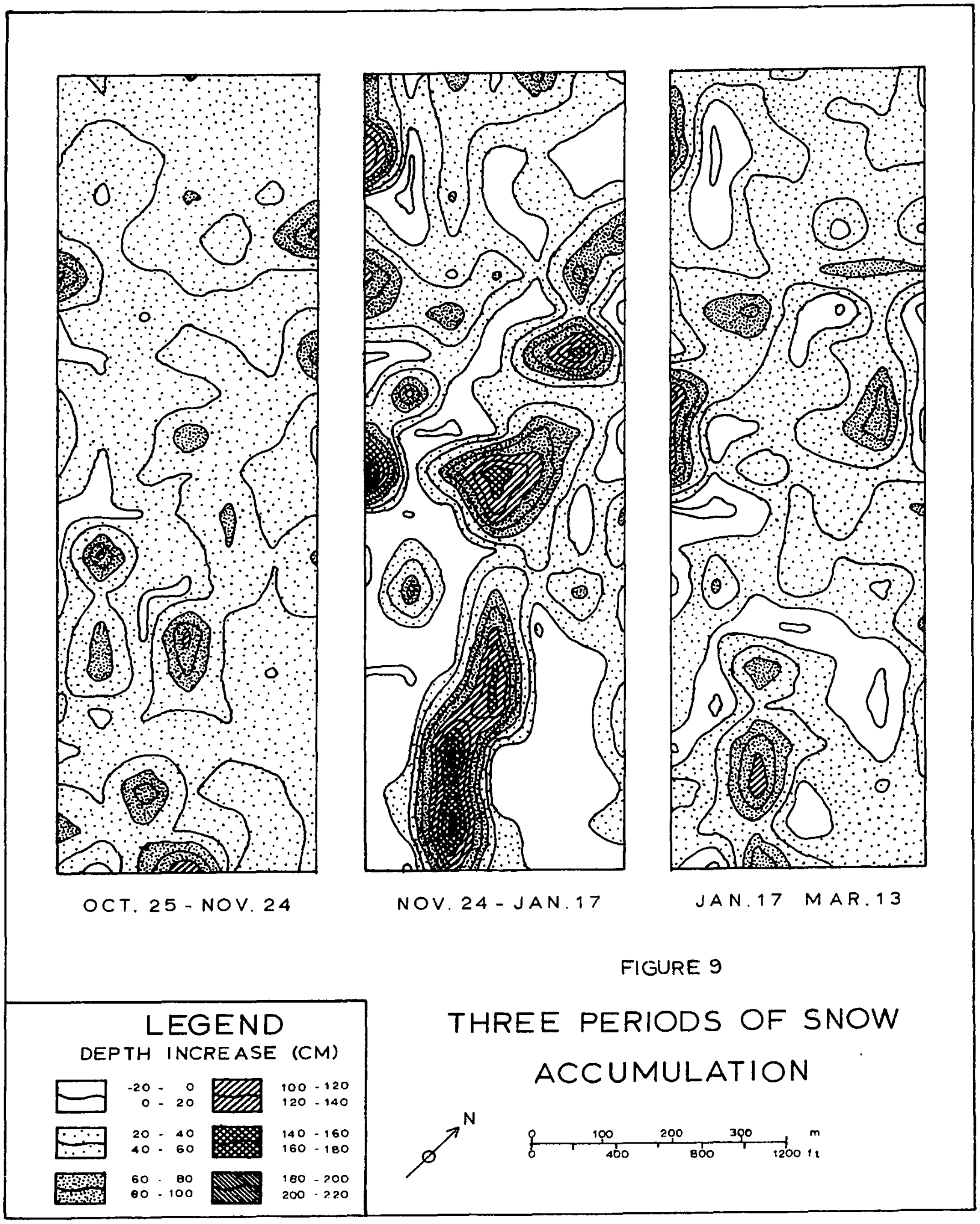


FIGURE 9

THREE PERIODS OF SNOW ACCUMULATION

Map 1. October 27 - November 24

In the first map ridge crest locations exhibit the shallowest snow. A depth below average is shown by the plain area in the south-eastern part of the sampling grid. Deeper accumulations are associated with the edges of this area. An even accumulation, slightly deeper than average, prevails in the deeper brush in the northwestern part of the experimental site. Occasional deep accumulations occur at stake locations near ridge crests.

Map 2. November 24 - January 19

Extensive areas of little or no increase in snow depth are shown by the second map. The plain area did not accumulate much snow during this period, whereas the valley southwest of it has experienced very deep accumulation. Ridge crests and their immediate surroundings show little accumulation whereas at some distance from ridge crests the accumulation is deep. The depth increase in the northwestern part of the area is more uneven after the brush vegetation has been buried and shows a pattern corresponding to topographic undulations.

Map 3. January 17 - March 13

In the third map the depth variation is less pronounced. There is little correspondance between this map and the previous one. In many zones where the accumulation rates were high during the previous period only slight increases occurred during this period. Only in wider valleys and concavities are the accumulations deep. A notable difference from the previous map is that zones on the plain and near ridge crests show continued accumulation. The major valley in the southeastern part can be traced through a greater depth increase but towards the mid-part there is no pattern indicating its existence.

The three maps show that erosion first takes place at ridge crests and areas of shallow vegetation (cf Figure 3). Smooth areas develop as is shown by the second map and the deepest accumulations occur adjacent to these areas. Through winter the zones of small depth increase remain fixed in space, whereas zones of rapid accumulation change location through winter. After the area has become smooth the depth increases most rapidly in larger valleys. The patterns of accumulation become more even, with more gentle increase in accumulation rates from ridge crest to valley bottom.

d) Characteristic trends of depth increase in different locations

As a result of the sequential spatial changes in depth increase each point should, depending on location, experience a characteristic pattern of depth increase through time. The depths of snow through winter for each of the 147 snow stake locations are shown in Figure 10 (foldout map in back cover). Six characteristic types of depth growth may be identified in this map:

Type 1. The depth of snow reaches maximum in the early snowfalls, and thereafter it remains constant or even decreases through winter (e.g. stakes 23, 50, 65).

Type 2. This type is characterised by a rapid initial increase in depth. Thereafter the depth remains constant for longer or shorter periods or increases slowly (e.g. stakes 93, 99, 100).

Type 3. The initial increase is relatively slow. Then follows a period of very rapid increase extended over a period of a few weeks after which the rate of accumulation decreases (e.g. stakes 102, 109, 116).

Type 4. After a short period of initial increase the snow depth remains constant for some time and then increases in an almost exponential fashion (e.g. stakes 57, 62, 64).

Type 5. The depth after the first few snowfalls is constant or slowly increasing and then at some point in time it continues to increase at a greater rate, but distinctly different from Type 4. This new rate of increase is often proportional to snowfall (e.g. stakes 51, 94, 105).

Type 6. The depth increase is approximately proportional to snowfall throughout the accumulation period (e.g. stakes 4, 18, 28).

The average depth increase is shown in Figure 11.

Superimposed on these general trends there are variations due to winds of different directions and temporary drift formations. Stake 56 shows a typical example of when a cornice drift has built out to just beyond the stake location and subsequently is eroded back by a wind of opposing direction. Fluctuations around the general trend are particularly common in the northwestern third of the sampling area. Field observations suggest that the fluctuations are caused by tail drifts behind trees changing location with direction of the latest storm.

For some stakes it is difficult to distinguish the proper group. Particularly group 4 can in some cases occur as part of group 3 where the period of very rapid increase often has the characteristic of group 4. It is also in some cases difficult to draw the limit between depth increase proportional to snowfall and slow increase in snow depth. Table 2 shows which groups the different snow stake locations were assigned to.

Type 1 is characteristic for ridge crests and summits of convex terrain elements. In these locations the wind stress remains locally higher throughout winter which results in a continuing net loss of snow.

Type 2 is found adjacent to ridge crests, in narrow valleys and shallow concavities on otherwise relatively flat areas. The rapid initial increase in depth with the subsequent very slow increase shows that the site is converted from one of initial locally low wind stress to one of locally high wind stress. The low wind stress may be caused by brush vegetation or minor topographic undulations or zones of boundary layer separation near ridge crests. After this minor roughness has been eliminated large scale wind variations control the accumulation. The slow depth increase shows that locations of this type act as source areas of drifting snow. It is these two first types together with types 4 and 5, which constitute the smooth areas in early winter. The stepped increase that may be seen in some cases could indicate the influence of winds of one direction producing accumulations which are not subsequently eroded by winds of other directions. Another explanation could be that temporarily halted snow dunes are hardened sufficiently to withstand erosion.

Type 3 is found near valley bottoms and on slopes at some distance from the crest. The period of very rapid increase is when the edge of the smooth area, often in the form of a cornice, reaches the particular location. The depth continues to increase after the site has been incorporated into the smooth area. This is because the snow

TABLE 2

STAKE LOCATIONS ASSIGNED TO DIFFERENT GROUPS

ACCORDING TO TYPE OF DEPTH INCREASE

Group 1

Stakes No: 16, 23, 30, 47, 49, 50, 65, 70, 72, 79, 92, 101, 103, 111, 114, 118, 120, 142. Total 18 stake locations.

Group 2

Stakes No: 9, 14, 20, 21, 22, 25, 32, 33, 43, 44, 46, 53, 59, 61, 63, 66, 67, 77, 78, 80, 85, 86, 87, 89, 93, 96, 97, 98, 99, 100, 107, 110, 112, 113, 117, 119, 125, 126, 127, 130, 135, 137, 138, 139,140, 141, 144, 145, 146. Total 49 stake locations.

Group 3

Stakes No: 1, 3, 5, 8, 15, 17, 19, 26, 35, 36, 37, 41, 42, 45, 48, 54, 55, 56, 68, 71, 73, 74, 75, 81, 82, 84, 95, 102, 109, 115, 116, 122, 129, 134, 136, 143. Total 36 stake locations.

Group 4

Stakes No: 57, 62, 64, 69, 88, 108. Total 6 stake locations.

Group 5

Stakes No: 2, 11, 27, 38, 51, 52, 60, 83, 90, 94, 105, 106, 121, 123, 124, 128, 132, 147. Total 18 stake locations.

Group 6

Stakes No: 4, 6, 7, 10, 12, 13, 18, 24, 28, 29, 31, 34, 39, 40, 58, 76, 91, 104, 131, 133. Total 20 stake locations fetch is longer in these locations and because of a generally lower wind stress away from the crest.

Type 4 occurs in wider valleys and concavities. The curve indicates that the point goes from an initial stage of low wind stress to a stage of locally higher wind stress and that gradually this changes to a stage when the wind stress is locally lower than for surrounding areas. In this type of location the snow fetch is short because most of the drifting snow is trapped at the edges of surrounding smooth areas. Due to the width of the concavity the point is located in the zone of reattachment after boundary layer separations behind surrounding ridge crests and the flow is therefore accelerated. Snow accumulation gradually eliminates the zones of boundary layer separation and therefore the acceleration becomes less marked and gradually the valley bottom becomes a zone of locally lowest wind velocity. This would explain the almost exponential increase in depth.

Type 5 is not so visibly related to the terrain geometry as the previous groups. The curves show that these sites first lose most of the snowfall but that then there is a change to near balance between the erosion and deposition due to drifting. This suggests that the curves result from an increasing homogeneity with respect to drifting in the area where they are located, so that the accumulation becomes similar to that of an infinite plain.

Type 6 occurs mainly in the lower, northwestern part of the experimental site but is also found in other parts. The sites where it occurs are generally characterised by the absence of major roughness elements in their immediate vicinity. It is a type of depth growth where the balance between drift erosion and deposition remains largely unaltered through the accumulation period. In the lower part of the site there are considerable variations around the trend which are probably a result of nearby trees changing their influence with different wind directions.

The growth curves show that the depth increase through winter is closely associated to the geometry of the terrain. In early winter brush vegetation and roughness features near the point control the depth increase, whereas the later stages of accumulation are influenced by roughness elements at further distances from the point. Increasing homogeneity with respect to drift transport produces further increase in depth in locations where the depth initially remained constant. The curves indicate that changes in the balance between erosion and deposition occur in a large number of cases. Only in a few locations (Type 6) is the depth increase proportional to snowfall throughout winter.

3. Loss of snow from exposed areas

In literature describing snow courses maintained in areas similar to the area of the present study it is frequently reported that large amounts of snow are lost from open areas and that snow depth and snowfall bear little relation to one another. A report from Goose Bay weather station (Pearce and Gold, 1951) states that snow depth reaches maximum in late January and remains fairly constant until the end of April, despite additional snowfall. In the same report, the snow cover in Aklavik, Northwest Territories, is described as reaching maximum depth before December 1st and remaining at that level throughout the rest of the winter. Dry snow, easily disturbed by wind was reported by the observer. Bare patches were observed in open areas throughout winter, whereas in vegetated areas the depth could be six feet (2 m) or more. Observers in Churchill, Manitoba, report very little snow in the open areas, from which deposited snow is soon drifted by the wind into more sheltered, wooded areas. Longley (1960) made observations of snow depth at 21 sites at Resolute, Northwest Territories, twice weekly during the winter of 1957 - 58. On the windswept island on which Resolute is located, many locations were never covered with more than 3 in (7.6 cm) of snow, while in protected locations depths of 20 ft (6.1 m) were common and 100 ft (30 m) ravines filled to the top. The mean depth (ibid. p. 735) shows little or no increase from mid-October despite snowfall. Barnett (1963) made similar observations in the Denault area near Schefferville. This site is an open area surrounded by forest. Barnett reports a period of constant snow depth at eight stakes lasting from the beginning of January to early March despite 86.3 mm water equivalent of snow recorded at the Schefferville weather station. During the week of greatest snowfall for the winter a net decrease of 1 - 2 cm was recorded at each of the eight stakes. These reports all indicate the influence of large scale effects on surface roughness on the average depth of the sample. This is probably to a great extent a result of non-representative sampling. At Timmins 4 similar growth curves would result if a major part of the sample contained sites which experience depth increase according to Groups 1 and 2. The proportional contribution of such sites is 45 per cent of the total sample but it can be seen that if the samples were limited to for example the upper flat area only, the proportion would be greatly increased. This type of growth curve is characteristic when there is a presence of efficient snow collectors nearby. In the report from Resolute such collectors are indicated in the form of ravines and protected locations. In the Denault study a surrounding forest forms an efficient snow sink. Depending on the composition of the sample with respect to locations of different depth growth characteristics, the representativeness of the sample would also vary through winter, particularly in the case of a small sample.

It is therefore of considerable interest to investigate a) the possible difference in accumulation between Timmins 4 and less exposed areas, b) the representativeness of the Timmins 4 snow sample through winter, and c) the amount of snow blown from ridge areas and the distance over which the wind blown snow influences the accumulation in surrounding areas.

a) Snow water equivalents at Timmins 4 and in the Knob Lake drainage basin

To investigate the possible differences in snow water equivalent between Timmins 4 and less exposed areas a water equivalent survey was made at Timmins 4 on March 13. Water equivalents were measured at each of the 147 sampling points using a Mount Rose snow sampler employing standard techniques (Klein et al, 1950). The measuring error of this instrument may be as much as 10 per cent (Beaumont, 1967). The measuring error varies with the type of cutter on the tube and the sampler employed in the survey was equipped with the cutter giving 10 per cent overestimate of the water equivalent.

The Knob Lake drainage basin was selected for comparison.

The survey would then also be of use in a hydrologic study simultaneously conducted in this basin (Penn, 1971). The same equipment was used for the sampling in the basin. Samples were taken at 100 m intervals along traverses transsecting the basin in different directions. In all, 344 samples were obtained. Lake sampling could not be undertaken because of slush present on the ice. In cold temperatures this effectively clogs the sampling tube, making it inoperative for further sampling. Although samples near lake shores are few there may be therefore a slight overestimate by the sample due to drifting from the lakes. The basin survey was undertaken two weeks later than the Timmins 4 survey. Intermediate snowfall as measured by the Nipher gauge was therefore subtracted from the basin figures.

The two surveys yielded very similar water equivalents. For Timmins 4 the average water equivalent was 413 mm and the water equivalent in the basin was 402 mm (corrected for intermediate snowfall). There is no statistically significant difference between the two averages.

This result would indicate that despite Timmins 4 being in a ridge crest location, the amount of snow accumulated is similar to the amount in less exposed areas. However, this presumes that there was no difference in snowfall between the two areas and that each of the two samples are truly representative of their areas.

b) Comparison between measured snowfall and snow depth increase

Underestimates of snowfall by the Nipher gauge

Previous investigators have shown that the Nipher shielded snow gauge at Schefferville tends to underestimate the snowfall. Adams et al (1966, p. 130) give a value of 23 - 28 per cent of true snowfall.

Rogerson (1967, p. 91) gives a value of 30 per cent and suggests that the value may be as much as 35 - 45 per cent. The measured weekly increase in water equivalent of the Schefferville snow course was used as reference. Possible causes of the underestimates have been discussed by Adams et al (ibid. p. 121), who ascribe the difference to ineffecient catch by the gauge due to the influence of wind and possibly losses in water equivalent because snowfalls of less than .01 of an inch are only recorded as 'trace'.

Assuming the averages from the Timmins 4 survey, the basin snow survey and both surveys together to be representative of 'true snowfall' gives values of the undercatch of 42.5, 44.0 and 43.0 per cent respectively. However, a possible overestimate of about 10 per cent by the sampling equipment would reduce the values correspondingly.

Cumulative snowfall and snow depth increase at Timmins 4

If the snow sampling grid at Timmins 4 is representative of snowfall through winter, the average depth increase would closely correspond to the cumulative snowfall measured by the Nipher gauge. This assumes that the catch deficiency of the gauge is approximately constant throughout winter, at least when averaged over a few storms. Changes in density would need to be accounted for. The density would be generally low in early winter, then show a relatively rapid increase during the period of transition from rough to smooth terrain, and continued but slow increase thereafter. This trend results from the snow crystals being broken down into fine fragments during drift transport due to repeated impacts with the snow surface. The fragments pack more closely together than the original crystals and therefore give higher density (Mellor, 1964, p. 49). In early winter the drift transport is limited by microroughness and therefore the density is low. The subsequent increase in density is then a result of either an increasing proportion of finely fragmented snow or pressure compaction of the snow or of metamorphic processes acting within the snowpack. Characteristic trends of density changes through time at Aklavik and Resolute, Northwest Territories are given in a report by Williams (1957), and for different localities in northern North America by Billello (1958, p. 65).

Correcting the Nipher gauge receipts for an undercatch of 42.5 per cent and converting the snowfall water equivalent into depth values using the average density at Timmins 4 on March 13 (.364) yields the estimated depth curve (b) in Figure 11. The density of the snow cover through winter (c) was obtained by assigning the corrected water equivalent measured by the Nipher gauge to the observed depth (a).

The two depth curves show similar trends throughout the accumulation period with one notable exception. The slightly divergent trends suddenly change to convergence between March 13 and the depth sample prior to the density survey. This suggests that the depths recorded by the snow sampler were not the same as those that would have been recorded by the snow stakes. Unfortunately, the snow stakes were not read simultaneously so a direct comparison is not possible. A plausible explanation of the error is that whereas the snow stakes were zeroed at the top of the lichen cover, the depth sampler generally penetrated the lichen mat. This would account for approximately 4 - 6 cm greater depths on the average, which would explain part of the difference. There could also be some difference produced by the

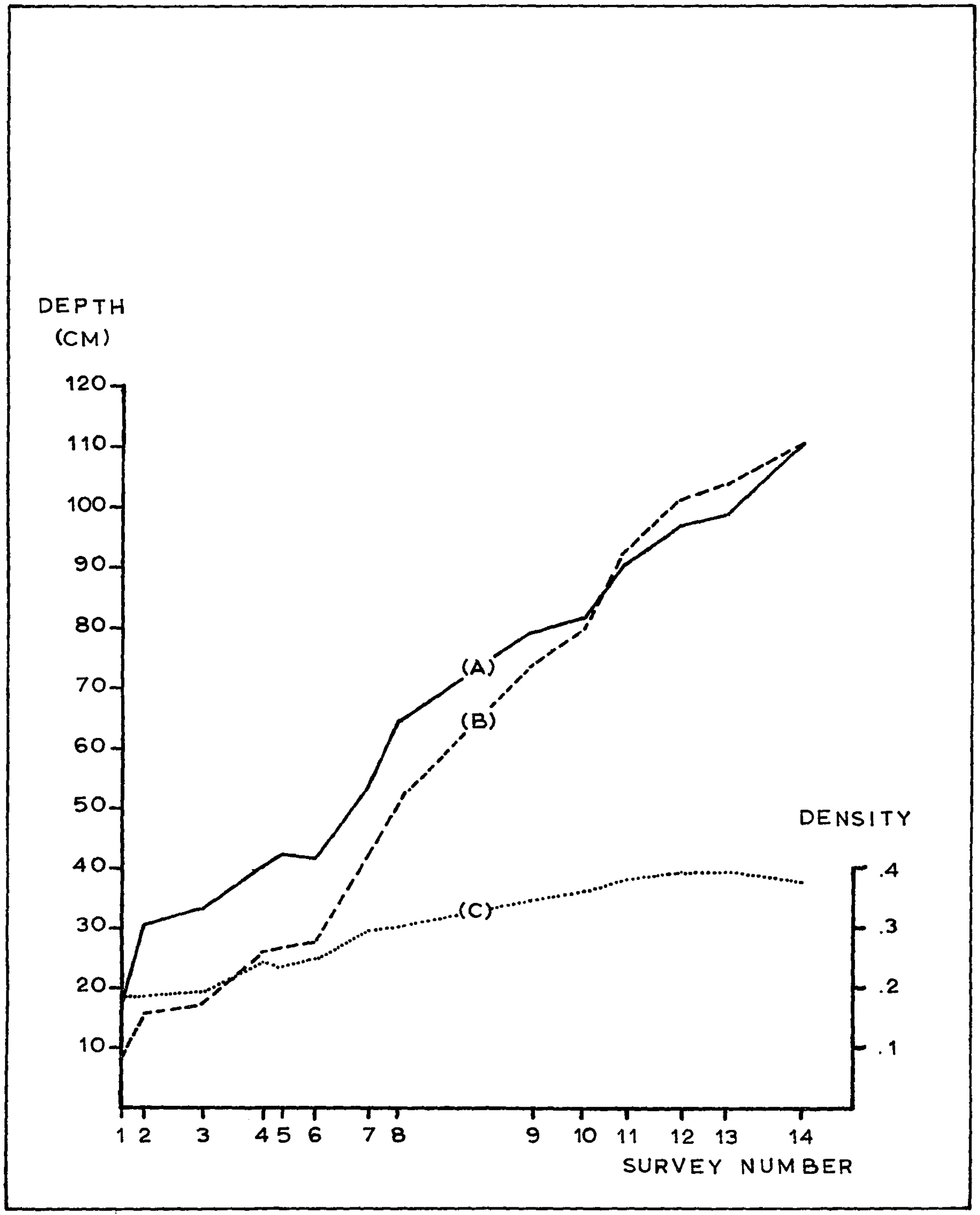


Figure 11 Measured snow depth (A), predicted depth (B) and predicted density (C)

lateral displacement of the sampling points which were moved on the average 50 cm from the stake location.

The density curve shows an increase in density from .188 in early winter to .393 for the survey before the density survey. Since the plugs of lichen were removed before weighing the samples the March 13 density value is affected but not the water equivalent. The densities show an increase which does not indicate any major differences between snowfall receipt and depth increase and are within the expected range. This would suggest that the Timmins 4 snowcourse is representative of snowfall throughout winter. However, there may have been slight gradual changes which are difficult to detect in the absence of intermediate density calibrations through winter.

c) An estimate of snow transport from ridge areas

Although the results so far have indicated that the Timmins 4 site does not have an accumulation significantly different from the accumulation in lower lying, less exposed areas, there is no doubt that appreciable quantities of snow are blown off ridge areas into surrounding lower lying terrain. There is therefore reason to regard the results with caution. To obtain a tentative estimate of the quantity of snow blown off the ridges and to investigate how far from a ridge area the influence of wind blown snow can be noticed, a survey was undertaken, sampling water equivalents 100 m apart along a traverse from south of Irony Mountain (approx. 700 m) to Howells River (approx. 550 m). This traverse runs through a uniformly forested area where the forest is sufficiently dense to prevent erosion within the forest. The forest acts as an efficient snow trap which catches snow blown from the ridge area but does not lose any snow due to drifting. Thus by measuring the water equivalents away from the edge of the ridge area through the forest a decline in water equivalent should be observed and eventually a steady value would be reached at some distance beyond which the influence of blown snow is negligible. The water equivalent in excess of this steady value should then be indicative of the amount of snow blown off the ridge.

The survey was undertaken in late February. At each sampling point three samples were taken 10 m apart. Further away from the edge of the ridge area the number of samples at each point were reduced to two. The survey starts 100 m before the forest edge. The result is shown in Figure 12.

The influence of drifting can be noticed for a distance of approximately 1000 m from the forest edge. The greatest accumulation is found at the forest edge itself. This accumulation is a result of snow particles travelling in the lowest layers of the airstream. Excluding the forest edge accumulation as being a result of mainly local drift transport a tentative estimate of the snow carried in suspension into the forest may be made. The average excess water equivalent is 12.8 per cent. Assuming the samples to be truly representative for each 100 m interval the amount of snow transported into the forest by suspension becomes equivalent to a strip of average snow cover 128 m wide parallel to the forest edge. However, this would account for only the snow drifting with winds entering the forest from the open area. These winds would span a sector from north-northwest to southeast. To account for wind transport in opposing directions the value was doubled, and rounded off to 260 m. If this value is assumed repre-

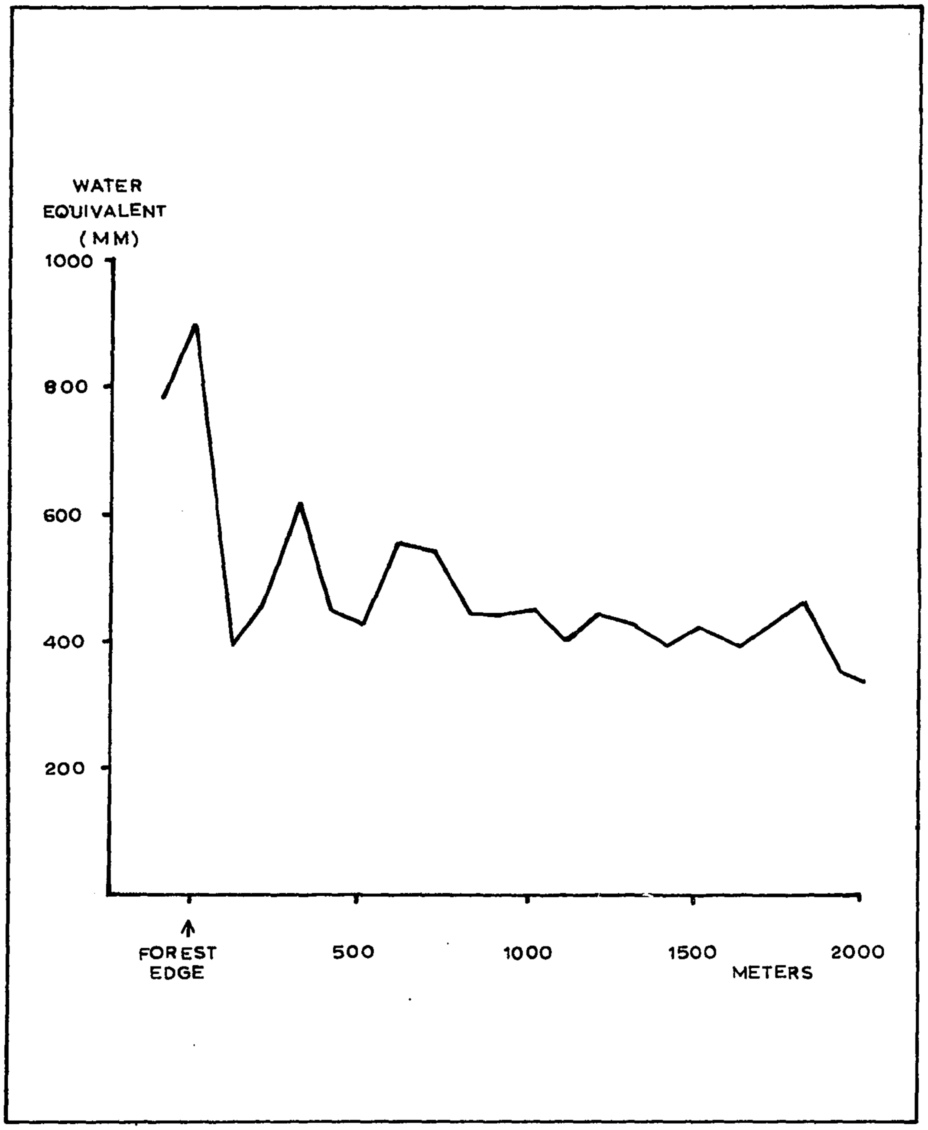


Figure 12 Traverse survey from a ridge area into a forest

sentative of the loss from the general ridge area which is about 6000 m wide where Timmins 4 is located, the average loss would be in the order of slightly less than 5 per cent of snowfall.

The survey shows that the influence of snow blown off ridge areas can be traced for a distance of about 1000 m into dense forest. An estimate of the loss of snow from the general exposed area where Timmins 4 is located suggests that less than 5 per cent of the snowfall is lost due to drifting. It is stressed, however, that these results are very tentative, firstly since the influence of prevailing northwest winds are not properly accounted for. Secondly, the nearby Irony Mountain represents an extremely exposed area at the edge of the general Timmins ridge area and this may have exaggerated the results.

4. Results of density - water equivalent survey at Timmins 4

Water equivalents were measured at Timmins 4 on March 13. The results of the survey are shown in Figure 13. The water equivalent, Y, appears to be closely associated with snow depth, X, and may be approximated by the equation

$$Y = 0.27\ X + 0.0119\ X^2$$

The snow density, Y', being the first derivative of this function is given by

$$Y' = 0.27 + 0.0238\ X$$

There is an apparent greater variability in density for shallower depths. Partly this may be a result of greater influence of error by the sampling equipment when shallow snow is sampled. The hardness of the snow in this range is also more variable with hard crusts within the snowpack which influence the accuracy of the

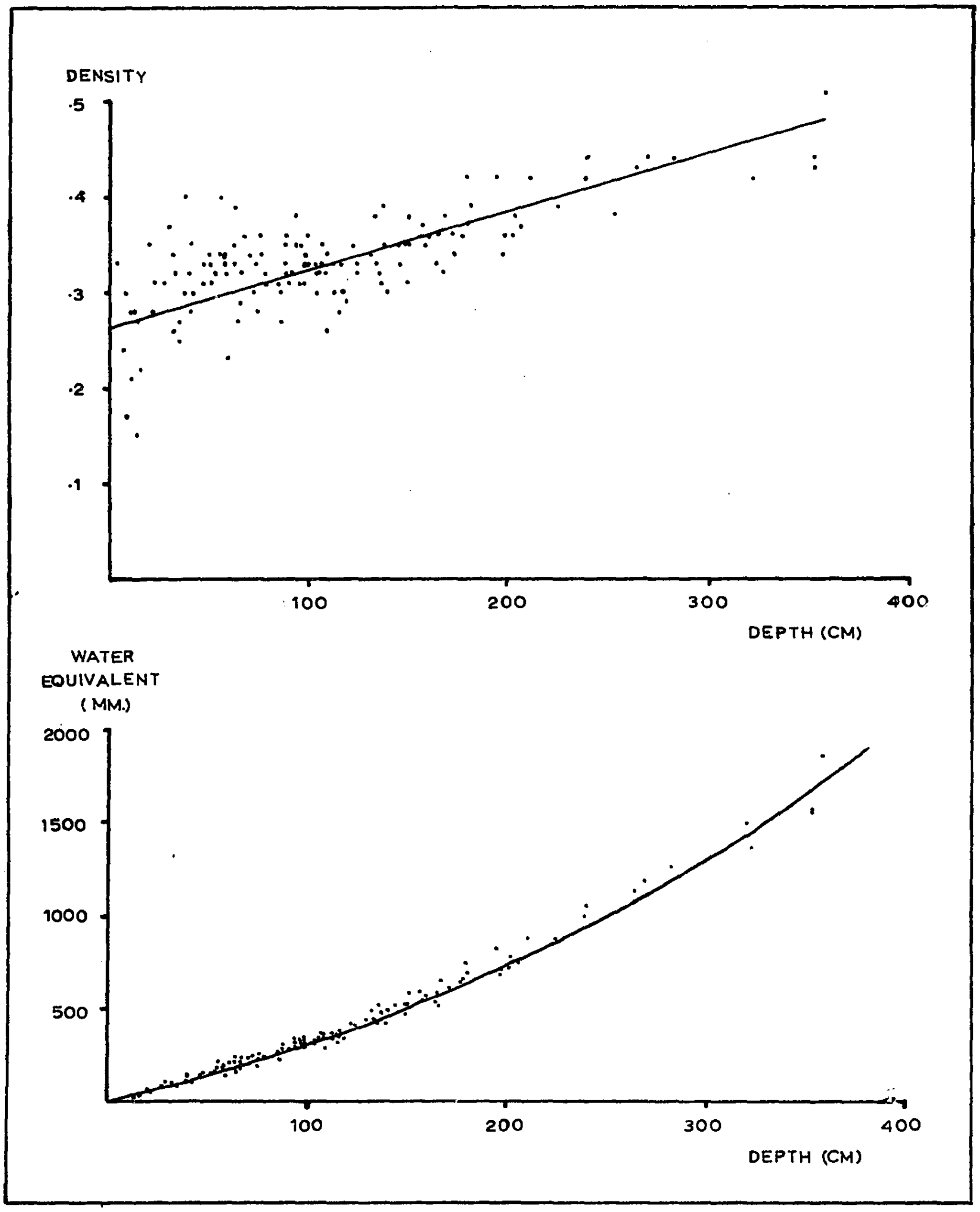


Figure 13 Water equivalent and density versus snow depth on March 13

sampling equipment. There is also some error produced by the penetration of the lichen cover which was discussed in section 3 of this chapter. This would contribute more to the variability in density due to the more irregular occurrence of lichen and shallower depth near ridge crests. Thus a large part of the scatter may be due to measurement error. Therefore, it appears not meaningful to enter a spatial analysis of density variations although probably systematic variations occur which are related to the geographical location.

5. Summary

Analysis of snow depth data from Timmins 4 shows that there are three distinctly different periods of accumulation related to changes in microroughness. This produces a sequence of depth growth which is closely related to firstly the microroughness at any one location and secondly is related to the geometry of surrounding terrain. Thirdly it is a function of the homogeneity of the surrounding terrain with respect to length of snow fetch. It seems that snow accumulation progressively through winter is controlled by terrain roughness of increasing scale.

Water equivalent surveys at Timmins 4 and in the Knob Lake drainage basin do not indicate any significant difference in accumulation between the exposed ridge area and the more sheltered basin. A traverse survey from an exposed ridge area tentatively suggests a loss of only 5 per cent of the snowfall from the general ridge area where Timmins 4 is located. The influence of the snow blown off the ridge area could be traced for a distance of 1000 m. A comparison between the water equivalent surveys at Timmins 4 and in the Knob Lake drainage basin (491 samples) and the snowfall receipt by the Nipher gauge in Schefferville suggest an undercatch of 43 per cent of true snowfall. The error may be somewhat lower due to a possible systematic overestimate of the snow sampler. A comparison between Nipher receipts and snow depth increase at Timmins 4 did not indicate any appreciable changes in the reliability of the snow sampling grid through winter but indicated a discrepancy caused by difference in depth sampling method. Water equivalents and snow depths are closely associated and for the March 13 survey the water equivalents could be closely estimated by a quadratic equation.

CHAPTER IV

A MODEL RELATING SNOW DEPTH TO TERRAIN ROUGHNESS

1. The Model

a) Previous Studies

A major part of the present study was the attempt to develop a method for indirect mapping of the snowcover using stepwise multiple regression for relating snow depth variations to terrain roughness variables. Previous studies have indicated that such an approach is feasible. Black and Budd (1964) found the following expression for the deviation in accumulation, A, from its mean in terms of the deviation in slope, S, and deviation in relative elevation, E:

$$A = -0.85S - 0.21E.$$

where the average profile on which S and E were determined is a 10 mile running mean of elevations.

A similar method was used by Mock (1968) to determine local variations in snow accumulation, the final equation derived was:

$$Y_D = 2.4 - 0.8211X_{ID} - 259.106 \sin a.$$

where Y_D is deviation from predicted accumulation (g/cm^2 year)

X_{ID} is deviation from regional elevation (m)

and a is surface slope in degrees.

Both of these studies were undertaken on ice caps (Wilkes region, Antarctica and Thule Peninsula, Greenland respectively), where the long-term accumulation of snow has caused a considerable smoothing of the original terrain (Swithinbank, 1959). Few attempts have been made to quantitatively relate the variations in accumulation of the seasonal snow cover to terrain parameters. Young (1969) used stepwise multiple regression and a set of variables in analyzing snow accumulation on Axel

Heiberg Island, N.W.T. The variables employed were:

1. Altitude
2. Exposure 5 m
3. Exposure 50 m
4. Exposure 500 m
5. Slope angle
6. Slope azimuth

The measurements representing exposure were obtained using a method by Rodda (1962), and may be defined as the length of an arc of a horizontal circle with its centre at ground level which does not intersect the ground. All of these variables showed a significant correlation with snow depth in one or more of the basins studied although their level of explanation of the variations in snow depth was low. A subsequent study by Young (1970) claims that using 19 independent variables in a multiple regression equation about 90 per cent of the variance in snow depth can be explained. However, this result may be doubted as to its applicability outside the sample, considering the number of sampling points (about 120) and the number of variables employed in the multiple regression equation. This comment is supported by Ezekiel and Fox (1959) in a discussion on the relationship between the reliability of the multiple correlation coefficient, the number of sampling points and the number of variables in the regression equation. The variables used by Young (1970) were obtained by fitting a linear regression plane to altitudes of each snow sampling point and its nearest eight grid intersections on a 100 m square grid. The angle of the maximum slope of this plane was taken as the slope angle of the center point. The slope azimuth was taken as the azimuth of the direction of maximum slope. The amount by which the center point was above or below the regression plane was taken as an index of convexity or concavity. Similar variables were evaluated for different scales.

b) General Model

The following general model was erected for use with stepwise multiple regression:

$$S = f\ (x_1,\ x_2,\ x_3\ \ldots\ x_n)$$

where S is the depth of snow and x_1, x_2, $x_3 \ldots x_n$ denotes terrain roughness variables which by themselves or in any combination contribute significantly to explaining the variation in snow depth. These variables are defined in Section 2 of this chapter, which also describes how the variables were obtained.

Stepwise multiple regression may be used as a tool to select a combination of a few independent variables for prediction of one dependent variable. The selection can be made from a large number of independent variables, which is particularly valuable in problems where there are variables measuring a particular aspect on different scales and the optimal scale is not known.

c) Stepwise multiple regression

The technique of stepwise multiple regression is designed to analyse the relationship between a dependent variable and a series of independent variables. The computational procedure has been described by Efroymsen (1960). Here, only a brief description of the technique will be given.

The technique involves the computation of multiple linear regression equations with additional variables added in a series of separate steps. The criterion of importance is based on the reduction of sums of squares.

The variable added at each step is the one which makes the greatest reduction in the sum of squares of the errors, when superimposed upon the variables already added. Also, it is the variable having the greatest partial correlation with the dependent variable, and which would have the greatest F value.

To limit the number of variables in each equation a significance criterion may be used (e.g. the F value) for deletion of variables already in the equation which have decreased in significance due to the inclusion of additional variables. Similarly a criterion may be used to limit further inclusion of variables when further addition of variables to the equation would no longer increase the accuracy of the prediction.

2. Terrain roughness variables

a) Selection of variables

For the selection of terrain roughness variables an extremely simplified model of airflow in terrain was used. It was simply assumed that the individual velocity variations near the surface are smoothed with height so that at some distance above the surface the effect of individual roughness elements of a particular scale is no longer noticeable. The flow at that level would then be parallel to the general terrain profile and velocity and stress variations near the surface would be represented by variations in the distance between the surface and the undisturbed flow since the velocity gradient would vary in steepness depending on this distance. Divergence between the flow at some height and the flow at the ground would produce decelerations and convergence would produce accelerations. Thus an approximation of the variations in surface wind stress may be obtained. The flow line for the "general flow" at some height may be obtained from a running average of altitudes along a traverse in any chosen direction. Different degrees of generalisation may be obtained by changing the number of altitude points in the running average.

The selection of variables was partly based on previous studies (Black and Budd, 1964; Mock, 1968) whereas the method of obtaining the basic topographic data is similar to that used by Young (1970).

Field observations indicated that in early winter brush vegetation has a major influence on the accumulation patterns. Breakpoints in the terrain appeared to have a significant influence on the depth variations during the second period of accumulation whereas in the third period larger scale topographic roughness features appeared to be most important. Differences in accumulation with respect to aspect are apparent from the snow melt photos (Figure 19 to 23). Different variables were evaluated to account for different stages of winter and some variables were evaluated on different scales to take into account roughness of different magnitudes.

b) Compilation of topographic roughness variables

Detailed topographic maps of scale 1:1200 and with 5-foot contour intervals have been prepared for the mining areas near Schefferville and provide excellent topographic information.

The map of Timmins 4 was produced by Terra Surveys, Ottawa, using photogrammetric techniques. In the following discussions, the measuring units of the map are adhered to, rather than obtaining uneven values by conversion to the metric system. The topographic variables are therefore expressed in feet unless otherwise specified.

By superimposing a grid network with half inch spacing between intersections upon the map of the snowcourse area, sampling points fifty feet apart were obtained. The grid was extended to 400 feet beyond the limits of the sampling grid. Altitudes were obtained by interpolating between contour lines at each of these sample points. Theoretically the error arising from this technique is ±5 feet with respect to the map, but it is estimated that the values are, on the average, correct within ±1 foot. Thus a rectangular matrix of altitude samples, was obtained. The size of the matrix is 41 x 97, with eight sampling points outside the snow sampling grid in all directions, for computational purposes. From this numerical matrix, parallel traverses in eight directions could be analysed and matrices of the different topographic variables generated. Appendix II gives a section of the computer program generating the variables. The different variables are identified by Table 3.

One particular feature of the topographic variables generated from the square data matrix needs special attention. The two different distances between the data points, depending on the direction of the traverses, are important in creating a scale difference between those traverses aligned with the rows and columns, and those at an angle of 45^{o} to the rectangular matrix. Therefore the measurements are not directly comparable, which has to be kept in mind if predictive equations obtained at Timmins 4 are used in other areas. The altitude matrices must be aligned in the same direction as the matrix from which the predictive equation is obtained. This is because the coefficients

TABLE 3

IDENTIFICATION OF VARIABLES

Variable No.	Representative of:		Direction	Fortran Name**
5	Brush depth			
6	Brush density			
7	Tree Density			
8	Altitude			B(KH,8)
9	Deviation from average altitude	3*	N-S	B(KH,9)
10	" " " "	3	NE-SW	B(KH,10)
11	" " " "	3	E-W	B(KH,11)
12	" " " "	3	SE-NW	B(KH,12)
13	" " " "	5	N-S	B(KH,13)
14	" " " "	5	NE-SW	B(KH,14)
15	" " " "	5	E-W	B(KH,15)
16	" " " "	5	SE-NW	B(KH,16)
17	" " " "	7	N-S	B(KH,17)
18	" " " "	7	NE-SW	B(KH,18)
19	" " " "	7	E-W	B(KH,19)
20	" " " "	7	SE-NW	B(KH,20)
21	" " " "	9	N-S	B(KH,21)
22	" " " "	9	NE-SW	B(KH,22)
23	" " " "	9	E-W	B(KH,23)
24	" " " "	9	SE-NW	B(KH,24)
25	" " " "	11	N-S	B(KH,25)
26	" " " "	11	NE-SW	B(KH,26)
27	" " " "	11	E-W	B(KH,27)
28	" " " "	11	SE-NW	B(KH,28)
29	General slope		NE-SW	B(KH,29)
30	" "		N-S	B(KH,30)
31	" "		E-W	B(KH,31)
32	" "		SE-NW	B(KH,32)
33	Deviation from average altitude	15	N-S	B(KH,33)
34	" " " "	15	NE-SW	B(KH,34)
35	" " " "	15	E-W	B(KH,35)
36	" " " "	15	SE-NW	B(KH,36)
37	Deviation from general slope		N	B(KH,37)
38	" " " "		NE	B(KH,38)
39	" " " "		E	B(KH,39)
40	" " " "		SE	B(KH,40)
41	" " " "		S	B(KH,41)
42	" " " "		SW	B(KH,42)
43	" " " "		W	B(KH,43)
44	" " " "		NW	B(KH,44)
45	Delay effect	5	N	B(KH,45)

* Number of altitude points employed

** See Appendix II

continued/......

TABLE 3 (Continued)

IDENTIFICATION OF VARIABLES

Variable No.	Representative of:			Direction	Fortran Name**
46	Delay effect		5	NE	B(KH,46)
47	"	"	5	E	B(KH,47)
48	"	"	5	SE	B(KH,48)
49	"	"	7	N	B(KH,49)
50	"	"	7	NE	B(KH,50)
51	"	"	7	E	B(KH,51)
52	"	"	7	SE	B(KH,52)
53	"	"	7	S	B(KH,53)
54	"	"	7	SW	B(KH,54)
55	"	"	7	W	B(KH,55)
56	"	"	7	NW	B(KH,56)
57	"	"	5	S	B(KH,57)
58	"	"	5	SW	B(KH,58)
59	"	"	5	W	B(KH,59)
60	"	"	5	NW	B(KH,60)

** See Appendix II

obtained are affected by the scale difference.

To avoid zeros and negative values in the variables, a constant was added to all of the independent variables where such values occur.

Altitude

Altitude is intended to represent effects of the general increase in wind velocity with height imposed on the airflow by topography. It was obtained directly from the matrix points coincident with the snow sampling grid.

Deviation from Average Altitude

These variables are intended to represent effects of deviations imposed on the airflow by local topography and were evaluated on six different scales, to account for different stages of winter and for different wind directions. Using the matrix of altitudes the variables were calculated as the difference between the altitude of the selected station and the average of a number of altitudes symmetrically on a line through the station. In general form deviation from average altitude (D) may be written as

$$D_I = A_I - \frac{A_{I-N} + A_{I-(N-1)} + A_{I-(N-2)} \cdots + A_I \cdots + A_{I+(N-2)} + A_{I+(N-1)} + A_{I+N}}{2N+1}$$

where A_I is the altitude of the station for which D_I is calculated and N is the number of altitude points included on each side of the station I along the traverse. Since the sliding average is symmetrical it covers two opposing directions (N-S, NE-SW, etc.) and therefore only four variables of each scale were necessary to cover the eight directions used in the analysis. The six different scales were obtained by using different values of N(1, 2, 3, 4, 5 and 7). The figures given in Table 2 refer to the number of altitude points included in the average.

General Slope

General slope (G) is defined as:

$$G_I = \frac{A_{I-4} + A_{I-3} + \dots + A_I + A_{I+1} + A_{I+2}}{7} - \frac{A_{I-2} + A_{I-1} + A_I + A_{I+1} + \dots + A_{I+4}}{7}$$

Since the variable is symmetrical it was evaluated for four different directions to cover the eight directions of the analysis. It is intended to represent large scale effects of slope on the air flow.

Deviation from General Slope

This variable is intended to represent influence of changes in slope within the nearest 50 or $50\sqrt{2}$ feet windward from the sampling point. Using the variable representing general slope (G) the deviations from this slope (DS) can be written:

$$DS_I = (A_{I-1} - A_I) - G_I$$

Since this variable is not symmetrical it was evaluated for eight different directions. The direction given in Table 2 refers to the direction of the wind.

Delay Effect

These variables are intended to represent systematic differences in accumulation on windward and leeward slopes as a result of winds from one direction prevailing over winds from the opposite direction.

This effect was imitated by a variable (De)

$$De_I = A_I - \frac{A_{I-1} + A_I + A_{I+1} + A_{I+2} + A_{I+3}}{5}$$

and on a larger scale

$$De_I = A_I - \frac{A_{I-2} + A_{I-1} + A_I + A_{I+1} + A_{I+2} + A_{I+3} + A_{I+4}}{7}$$

each for eight different directions. The directions given for these variables in Table 2 refer to the direction of the wind for which the variable is calculated.

c) <u>Compilation of vegetation variables</u>

The aerodynamic significance of the low growing vegetation may be subdivided into two parts. One is the height of the plants, and the second is the density of the plant cover. Both height and density are interrelated in that locations favoring growth of deep brush generally also produce a dense brush cover. It was decided to use two variables, one based on brush depth and measured in the field. The second variable was based on brush density and was obtained by remote sensing.

<u>Brush depth</u>

The depth of brush was measured in the field by using the mean maximum height of individual plants within a circle of 3 m radius centered at the snow stake. The accuracy of this measurement is estimated to be within ± 10 cm. However, in some cases, where there is a systematic decrease in depth of vegetation across the circle the measurement may be somewhat misleading, particularly if the number of plants per unit area also varies.

Brush density

Brush density was found difficult to measure in the field due to large systematic variations in plant size. Since it would be advantageous to find a brush vegetation variable which is easily obtained over large areas it was decided to obtain infrared color aerial photographs of the Timmins 4 site from which different vegetation roughness classes could be delineated.

Kodak Technical Publication M-28 gives a general introduction to infrared color photography. The film used was Kodak Ektachrome Infrared Aero Type 8443, a three layer color film in which two layers operate in the visible range and the third is sensitive to radiation in the near infrared band. All the layers are affected by blue light. A yellow filter (Wratten No. 12) is therefore used on the camera to exclude blue.

In test photographs it was found that it was difficult to obtain a suitable color balance due to an under-representation of the infrared relative to the visible wavelengths. To overcome this problem, two Polaroid filters were used in addition to the ordinary Wratten No. 12. The polarizing effect of these filters decreases sharply in the near infrared range and they may therefore be used at different angles near 90^{o} between the polarizing planes to balance infrared versus visible wavelengths. This method was developed by the author during previous experiments with infrared photography. The results are shown in Figures 14 and 15. Use of the crossed polaroids improved the color separation considerably and made it possible to directly map 5 classes of vegetation.

The high, oblique photographs were rectified using the snow stake grid and ground control photographs in black and white taken at lower altitude. The vegetation map (Figure 3) was produced in this way by hand-plotting the different vegetation classes. The variable representing brush density was obtained from the map by assigning the numerical values of the different vegetation classes to the points of the snow sampling grid falling within each vegetation class. An exception was made for class 5 which was considered to exhibit a roughness comparable to class 2. The accuracy of this variable is therefore not great since it can assume only four different values. Plotting errors may also have some influence on its accuracy.

Trees

The variable representing trees was obtained from counting trees within 200 ft (61 m) squares with the snow stakes as center points. The counting was done from early winter aerial photographs, where trees are easy to distinguish, standing in good contrast to the snow cover. Thus trees are also expressed as a density measurement.

d) Interrelationships between terrain variables

Many of the independent variables are interrelated. There are direct geometric relationships between variables representing

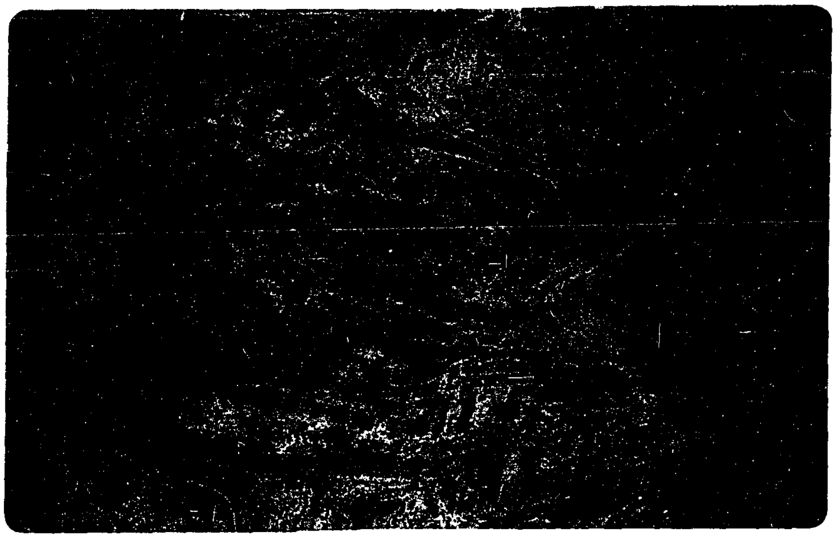


Figure 14. Kodak Ektachrome IR of Timmins 4 with yellow filter only.

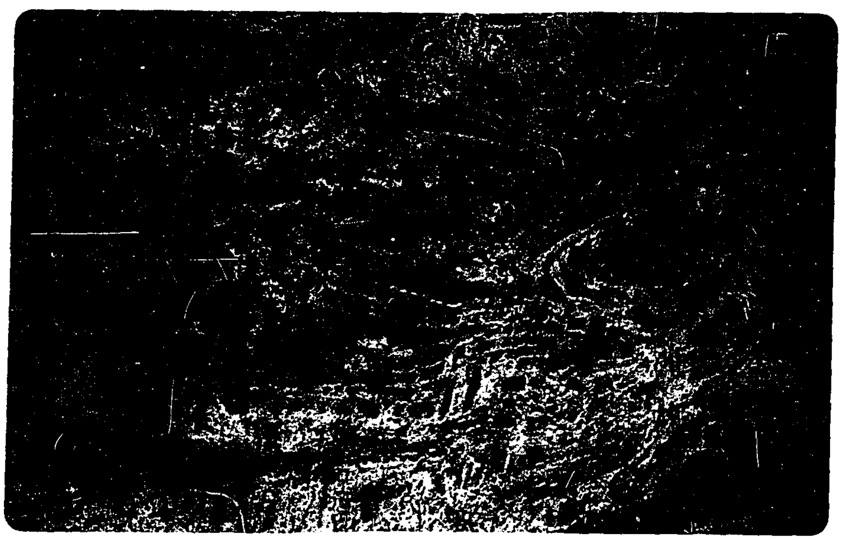


Figure 15. Kodak Ektachrome IR with yellow filter and two polaroid filters at an angle of 85° between the polarizing planes.

different aspects of topographic roughness, and causal relationships between surface geometry and vegetation due to the influence of terrain shape on microclimate, hydrology and soils. This interrelationship between the independent variables may in some cases cause a statistically significant correlation between snow depth and an independent variable without necessarily indicating a causal relationship. Thus the general applicability of the results of the statistical analysis is restricted to areas where similar interrelationships exist between independent variables.

CHAPTER V

TEST OF THE MODEL

1. Methods

Simple correlation was used to investigate how the variables individually related to snow depth through winter. This provides an initial overall view of the potential usefulness of the variables as predictors of snow depth. Rigid conclusions cannot be expected since many of the variables do not comply to the basic requirements for correlation analysis, particularly in that most of the variables are interrelated (Ezekial and Fox, 1959; Ch. 8, Miller and Kahn, 1962).

Five surveys were selected for analysis and for part of the analysis the dependent variable (snow depth) was transformed to normal distributions (Snedecor 1956). The surveys and transformations employed were:

1.	October 27	Cube root
2.	November 10	$\log_{10}$
3.	December 14	Cube root
4.	February 9	Cube root
5.	March 13	Square root

The stepwise multiple regression model was first tested assuming linearity and then tested with the dependent variables transformed for normality. To obtain an approximate estimate of the general applicability of the equation for March 13 a predicted map was produced for the Timmins 4 snowcourse area. This map was compared to a snow depth map for late winter which was produced from sequence aerial photographs during the spring melt period.

2. Simple correlation

a) Expected correlations

The coefficients of correlation between snow depth and the different terrain roughness variables may be expected to vary through winter approximately in accordance with their change in importance to snow accumulation. Thus the correlation with brush vegetation variables would show an early peak and thereafter a decline, since brush vegetation is buried by snow after some time. Deviation from general slope would show a peak correlation during the earlier stages of winter because it is a measurement of small scale roughness. Later in winter it would become overridden by other roughness factors and therefore there would be a decline in the coefficient of correlation. The smallest scales of the variable for deviation from average altitude would show a similar trend, whereas the larger scale versions of this variable would show a continued increase in correlation through winter. General slope would show an increasing correlation through time since systematic transport from windward to leeward slopes would increase with decreasing general roughness. The increased transport would influence the correlations for the delay effects in a similar manner. The increased influence of roughness of larger magnitude towards late winter would produce an increasing negative correlation between snow depth and altitude. Similarly trees would show an increasing positive correlation through time since the roughness created by trees remains largely unaltered by snow accumulation.

b) Observed correlations

The correlation coefficients for the five surveys are shown in Table 4 (non-transformed) and Table 5 (snow depth transformed for normality). The correlation coefficients vary slightly depending on transformation and are generally somewhat higher with the dependent variable transformed. The trends of change in correlation, however, remain similar.

Brush vegetation

The variables representing depth and density of brush do not follow the expected pattern. Both variables start at a low level of correlation and show a rapid increase in correlation up to December 14. After this date there is a continued slight increase in correlation through winter instead of the decrease expected. In the field there is a strong correlation between snow depth and brush in early winter. This relationship is apparently not well represented by either of the two brush variables, as indicated by the low correlation in early winter. The continued increase in correlation through winter indicates the presence of another direct or indirect relationship. The relationship between snow depth and brush vegetation will be further investigated later in the chapter.

Deviation from general slope

The peak correlation is reached for deviations from general slope north and northeast of the sampling points on November 24. The variables for the directions northwest, west and southwest peak on December 14. For the directions south, southeast and east the peak correlations are reached on February 9. The significant correlations are all negative, indicating that if the wind encounters a negative step in the terrain the accumulation behind this step is

TABLE 4

CORRELATION BETWEEN SNOW DEPTH AND TERRAIN

VARIABLES AT FIVE DIFFERENT POINTS IN TIME

(NON-TRANSFORMED)

Variable	October 27	November 24	December 12	February 9	March 13
5	0.210	0.302	0.414	0.453	0.480
6	0.269	0.352	0.466	0.487	0.510
7	-0.005	-0.011	-0.064	-0.026	-0.044
8	-0.015	-0.065	-0.055	-0.140	-0.164
9	-0.245	-0.470	-0.486	-0.506	-0.500
10	-0.256	-0.472	-0.561	-0.545	-0.551
11	-0.221	-0.388	-0.567	-0.585	-0.569
12	-0.042	-0.245	-0.280	-0.272	-0.220
13	-0.285	-0.462	-0.541	-0.586	-0.601
14	-0.289	-0.479	-0.620	-0.642	-0.658
15	-0.266	-0.402	-0.608	-0.678	-0.679
16	-0.089	-0.277	-0.340	-0.321	-0.271
17	-0.263	-0.405	-0.532	-0.598	-0.635
18	-0.317	-0.465	-0.621	-0.679	-0.710
19	-0.283	-0.373	-0.577	-0.682	-0.697
20	-0.131	-0.305	-0.401	-0.393	-0.354
21	-0.256	-0.365	-0.523	-0.598	-0.644
22	-0.314	-0.414	-0.595	-0.681	-0.724
23	-0.278	-0.334	-0.541	-0.665	-0.694
24	-0.166	-0.319	-0.433	-0.447	-0.418
25	-0.252	-0.334	-0.512	-0.591	-0.643
26	-0.303	-0.366	-0.567	-0.672	-0.722
27	-0.251	-0.291	-0.499	-0.632	-0.672
28	-0.185	-0.315	-0.435	-0.465	-0.433
29	-0.069	0.038	0.187	0.298	0.319
30	0.089	0.135	0.256	0.272	0.296
31	-0.129	-0.046	0.063	0.201	0.197
32	-0.149	-0.125	-0.020	0.103	0.079
33	-0.236	-0.293	-0.495	-0.578	-0.625
34	-0.277	-0.299	-0.521	-0.632	-0.685
35	-0.200	-0.237	-0.434	-0.560	-0.605
36	-0.219	-0.306	-0.427	-0.478	-0.460
37	-0.216	-0.475	-0.429	-0.421	-0.446
38	-0.216	-0.438	-0.446	-0.435	-0.447
39	-0.191	-0.341	-0.523	-0.559	-0.545
40	0.056	-0.123	-0.140	-0.159	-0.094
41	-0.088	-0.153	-0.183	-0.210	-0.168
42	-0.115	-0.230	-0.286	-0.242	-0.224
43	-0.110	-0.223	-0.273	-0.239	-0.211
44	-0.090	-0.226	-0.249	-0.185	-0.167
45	-0.265	-0.453	-0.599	-0.662	-0.709

continued/......

TABLE 4 (Continued)

CORRELATION BETWEEN SNOW DEPTH AND TERRAIN

VARIABLES AT FIVE DIFFERENT POINTS IN TIME

(NON-TRANSFORMED)

Variable	October 27	November 24	December 12	February 9	March 13
46	-0.198	-0.422	-0.636	-0.751	-0.788
47	-0.112	-0.277	-0.504	-0.677	-0.679
48	0.050	-0.087	-0.221	-0.308	-0.262
49	-0.283	-0.419	-0.608	-0.676	-0.723
50	-0.242	-0.414	-0.654	-0.780	-0.824
51	-0.153	-0.269	-0.510	-0.690	-0.706
52	-0.006	-0.140	-0.285	-0.372	-0.330
53	-0.179	-0.263	-0.319	-0.369	-0.389
54	-0.289	-0.336	-0.402	-0.404	-0.423
55	-0.293	-0.305	-0.406	-0.415	-0.435
56	-0.213	-0.319	-0.321	-0.239	-0.229
57	-0.137	-0.190	-0.213	-0.241	-0.240
58	-0.263	-0.290	-0.304	-0.267	-0.271
59	-0.268	-0.257	-0.320	-0.284	-0.296
60	-0.205	-0.300	-0.276	-0.176	-0.168

TABLE 5

CORRELATION BETWEEN SNOW DEPTH AND TERRAIN

VARIABLES AT FIVE DIFFERENT POINTS IN TIME

(SNOW DEPTHS TRANSFORMED FOR NORMALITY)

Variable	October 27	November 24	December 12	February 9	March 13
5	0.223	0.310	0.423	0.456	0.491
6	0.211	0.379	0.424	0.514	0.534
7	0.063	0.045	0.021	0.031	0.008
8	-0.103	-0.125	-0.092	-0.203	-0.210
9	-0.304	-0.500	-0.547	-0.570	-0.553
10	-0.340	-0.538	-0.617	-0.597	-0.597
11	-0.304	-0.459	-0.611	-0.617	-0.600
12	-0.087	-0.237	-0.296	-0.304	-0.251
13	-0.357	-0.521	-0.605	-0.645	-0.655
14	-0.376	-0.559	-0.682	-0.697	-0.706
15	-0.354	-0.494	-0.672	-0.718	-0.710
16	-0.145	-0.287	-0.366	-0.365	-0.313
17	-0.333	-0.468	-0.587	-0.644	-0.679
18	-0.407	-0.554	-0.690	-0.731	-0.754
19	-0.375	-0.470	-0.652	-0.725	-0.728
20	-0.193	-0.328	-0.429	-0.439	-0.399
21	-0.324	-0.424	-0.569	-0.632	-0.678
22	-0.405	-0.504	-0.664	-0.726	-0.761
23	-0.371	-0.431	-0.617	-0.707	-0.724
24	-0.230	-0.349	-0.463	-0.492	-0.463
25	-0.313	-0.385	-0.550	-0.615	-0.669
26	-0.392	-0.452	-0.633	-0.707	-0.752
27	-0.338	-0.379	-0.567	-0.667	-0.699
28	-0.251	-0.350	-0.464	-0.506	-0.486
29	-0.127	-0.026	0.093	0.214	0.252
30	0.046	0.131	0.224	0.235	0.262
31	-0.165	-0.115	-0.030	0.126	0.137
32	-0.172	-0.192	-0.098	0.048	0.046
33	-0.285	-0.336	-0.525	-0.585	-0.636
34	-0.353	-0.373	-0.578	-0.652	-0.705
35	-0.275	-0.308	-0.585	-0.582	-0.626
36	-0.287	-0.346	-0.454	-0.516	-0.501
37	-0.245	-0.478	-0.465	-0.467	-0.479
38	-0.251	-0.461	-0.463	-0.474	-0.485
39	-0.221	-0.374	-0.517	-0.562	-0.561
40	0.022	-0.102	-0.129	-0.170	-0.114
41	-0.134	-0.192	-0.233	-0.257	-0.212
42	-0.181	-0.287	-0.339	-0.278	-0.283
43	-0.186	-0.284	-0.332	-0.279	-0.239
44	-0.120	-0.232	-0.282	-0.224	-0.192
45	-0.283	-0.488	-0.620	-0.677	-0.725

continued/.......

TABLE 5 (Continued)

CORRELATION BETWEEN SNOW DEPTH AND TERRAIN

VARIABLES AT FIVE DIFFERENT POINTS IN TIME

(SNOW DEPTHS TRANSFORMED FOR NORMALITY)

Variable	October 27	November 24	December 12	February 9	March 13
46	-0.227	-0.448	-0.625	-0.739	-0.783
47	-0.145	-0.287	-0.480	-0.631	-0.661
48	0.034	-0.045	-0.176	-0.294	-0.264
49	-0.321	-0.470	-0.635	-0.692	-0.740
50	-0.293	-0.459	-0.662	-0.776	-0.823
51	-0.209	-0.307	-0.516	-0.679	-0.695
52	-0.035	-0.113	-0.253	-0.367	-0.340
53	-0.259	-0.314	-0.378	-0.422	-0.440
54	-0.392	-0.440	-0.505	-0.485	-0.489
55	-0.386	-0.420	-0.517	-0.490	-0.493
56	-0.276	-0.384	-0.397	-0.311	-0.286
57	-0.223	-0.243	-0.277	-0.300	-0.293
58	-0.365	-0.390	-0.411	-0.353	-0.340
59	-0.362	-0.374	-0.437	-0.364	-0.355
60	-0.267	-0.368	-0.357	-0.249	-0.222

increased, whereas a positive step decreases accumulation. The decline in correlation indicates that these effects of change in slope near the snow stakes are important at an early stage of winter but that later other aspects of terrain roughness tend to obliterate this effect. The best correlation is shown with changes in slope east of the sampling point from November 24 onwards. This is probably an effect of the greater variation in slope in this direction. It indicates that the effects of winds from the east override the effects of winds from the west, probably because winds with easterly components occur in combination with snowfall, whereas winds from opposing directions generally occur without precipitation. The low correlation for changes in slope south, southwest and west of the sampling points may be explained by overriding effects of the opposing wind directions. The correlations for the northwest and southeast directions are low, although these are the two prevailing wind directions measured in Schefferville. This could be an effect of the lesser roughness in these directions but could also be an indication that the wind directions at Timmins 4 differ considerably from those measured in Schefferville.

Deviation from average altitude

The variables included in this group show a strong general increase in correlation in surveys before February. For all directions there is a tendency for the smaller scale variables to decline in importance towards the later part of the accumulation season whereas the larger scale variables show a continued trend of increase in correlation with increased snow depth. A notable exception is the direction southeast - northwest, where all degrees of the variable show a decline in correlation. Generally the correlations are lower for this direction throughout winter. This case will be further discussed below.

General slope

General slope shows significant correlation with snow depth only for the directions north-south and northeast-southwest and only during the later part of the period investigated. The correlation indicates a greater accumulation on south and southwest facing slopes.

Delay effects

The delay effect variables show a general increase in correlation in successive surveys before February. In February and March the directions northwest, west and southwest show no further increase in correlation. The correlations for the directions north, northeast and east continue to increase through winter. This would indicate that there is a systematic transport of snow from slopes facing these directions to slopes facing opposing directions, for which directions the correlations are lower. The correlations for the northwest and southeast directions are lower than for other directions in February and March.

Altitude

Altitude shows an increasing negative correlation with snow depth through winter. The correlation coefficient, however, does not become significant at the 99 per cent level until late in the period, and then only when the snow depth data is transformed.

Trees

Throughout the period of investigation trees do not show any significant correlation with snow depth. The graph of tree density versus snow depth on March 13 (Figure 16) shows a tendency for less variable snow depth with greater tree density but no direct indication of greater depth due to the greater tree density.

Discussion

The correlation coefficients in most cases follow the expected general patterns through winter. However, there are a few notable exceptions and these will be further investigated in this section.

An increase in correlation of any one variable should result in decrease in the correlation of other variables, unless the variables themselves are interrelated. However, even if the variables are interrelated the one with the closest relationship to snow depth will have the greatest increase in correlation.

In early winter all variables show a low correlation with snow depth. For the topographic variables this was expected and to some extent this is a result of the scale on which these variables are evaluated. Another cause is likely to be the influence of brush vegetation. Neither of the two brush variables, however, gives a high correlation.

The variables for the directions southeast-northwest show a markedly lower correlation with snow depth than equivalent variables representing other directions. This may be expected because of the smaller relief variations in the direction southeast-northwest. These are also the prevailing wind directions. All variables representing

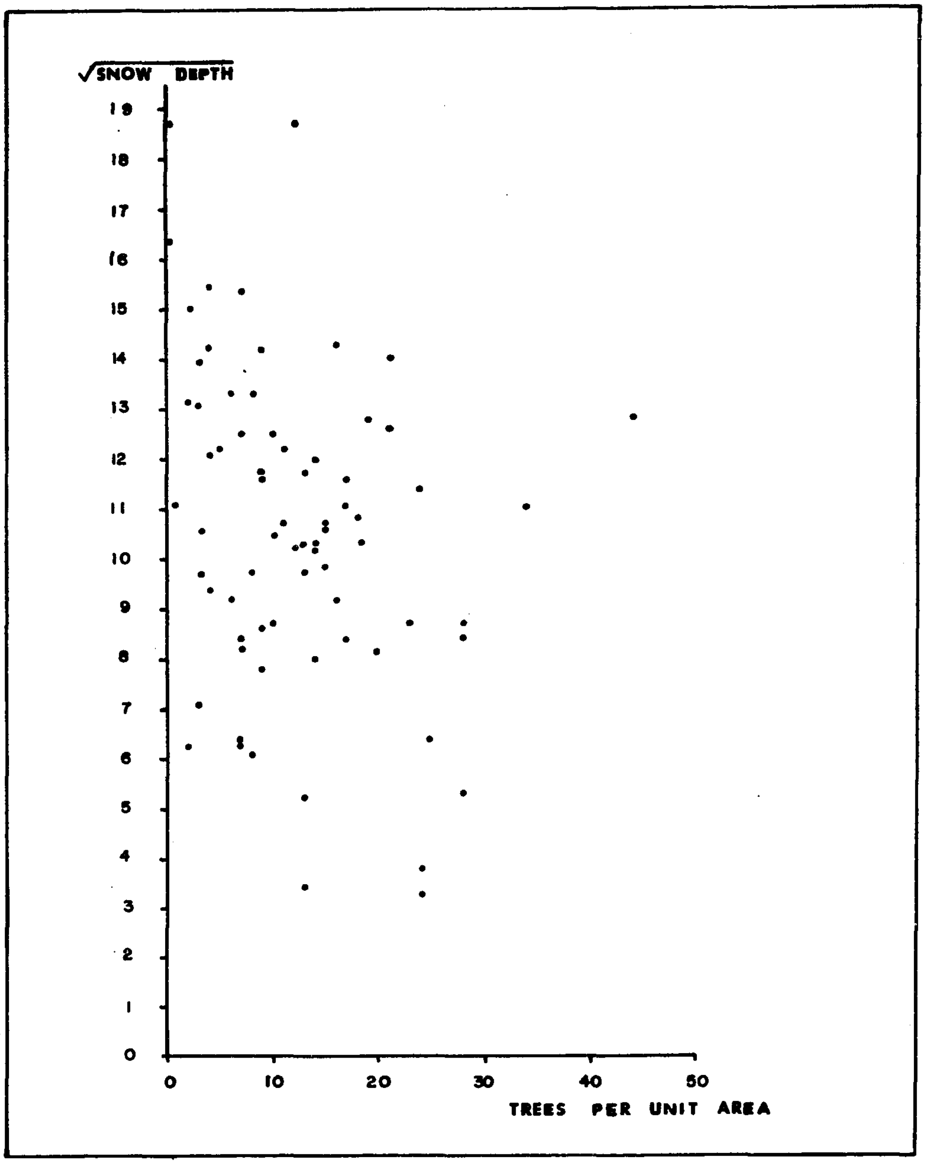


Figure 16 Snow depth versus tree density on March 13

these directions show a constant or decreasing correlation in the late part of the period of investigation. This could indicate that the roughness in these directions is smoothed by snow accumulation early in the winter, and that winds from other directions have greater impact on the patterns of snow accumulation due to the greater variations in surface stress produced by these winds. However, regardless of wind direction the wind stress is higher at the ridge crests than in valley bottoms even if there are no undulations in the direction parallel to the wind. Thus since the greatest relief is east-west the greatest variations in snow depth might be expected east-west. It may be significant that the deviations from average altitude correlate better for the direction perpendicular to the prevailing wind instead of the direction east-west. Similar findings were made in Antarctica by Gow and Rowland (1965) who found a better relationship between topographic undulations and snow accumulations along a traverse perpendicular to the prevailing wind than for a traverse parallel to this wind.

It is not possible to draw any general conclusions about the influence of winds from different directions using the present data, since no measurements of wind directions are available for Timmins 4. Judging from Schefferville wind data winds of directions coincident with snowfall and greater surface roughness have the greatest influence. A hypothesis was put forward by Tout (1964) that the prevailing northwest wind measured in Schefferville would be due to funneling by the northwest-southeast trending ridges. The greater accumulations on southwest and south facing slopes could indicate greater influence of winds from north and northeast than is indicated by the Schefferville measurements. Further research is required with comparable wind data before the relationship between wind directions, snow accumulation and terrain roughness can be further elaborated.

In summary, the correlations are generally as expected from the principles outlined in Chapter II. One major exception is that the correlations with vegetation variables are not as expected. Another exception is that the southeast-northwest variables show lower correlation coefficients, which may be due to the wind characters or it may be due to the nature of the relief.

c) Brush vegetation and snow depth

The brush vegetation variables were expected to show a peak in correlation with snow depth early in the winter. Instead the initial correlation was low and showed an increase through winter. To investigate this case further scatter plots were made of the depth of snow versus the depth of brush (Figure 17). The oblique lines on the figure represent equal depth of brush and snow. The sequence shows the following:

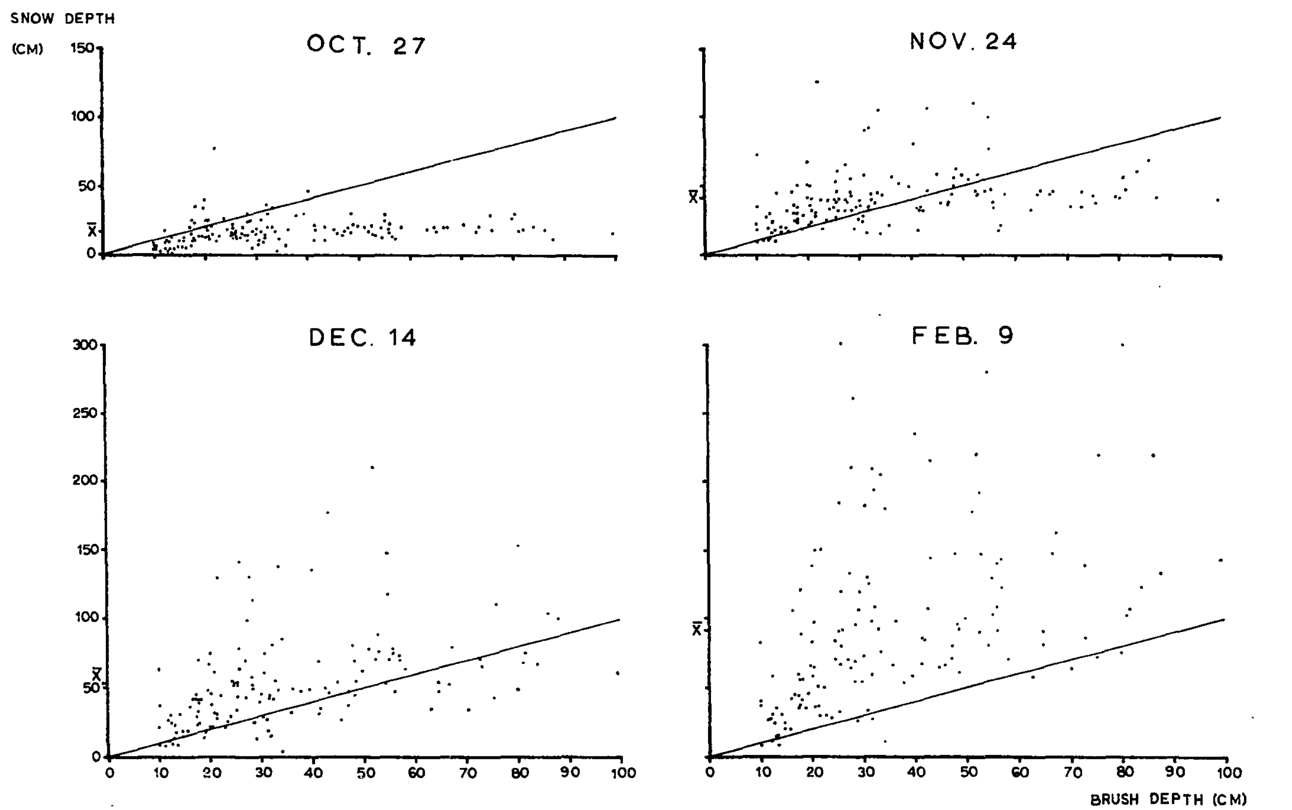
1. Deep brush greatly reduces the redistribution of snow in early winter. This is shown by the limited scatter around the mean depth of snow in early winter (October 27 and November 24) in the deeper range of brush.

2. Wind erosion starts in areas of shallow brush as indicated by the shallower depths and greater scatter in the shallow range of brush (October 27 and November 24).

3. There is a tendency for a linear relationship in the shallow range of brush on October 27 (brush depth less than 30 cm). On November 24 a similar but less strong tendency can be seen in the

FIGURE 17

BRUSH DEPTH AND SNOW DEPTH



deeper range of brush (brush depths greater than 50 cm). This shows that the relationship is linear over only part of the range at any one point in time. The variable is therefore not suitable for the present study.

Two phenomena were observed in the field which indicates that the patterns of snow accumulation are of considerable importance to brush vegetation (primarily Betula glandulosa). Firstly it was observed that near ridge crests the tops of the plants reach a very uniform level. The brush vegetation is very shallow near ridge crests and increases in depth towards valley bottoms where the uniform height of plants is no longer observed. Secondly, after periods of strong winds in conjunction with low temperatures considerable amounts of vegetation debris could be observed on the snow surface. Analysis of this debris confirmed that the bulk of the broken twigs were alive and thus eroded directly from live plants. The diameter of twigs was generally less than 2 mm. It is well known that fresh wood becomes brittle at a temperature of approximately -30°C. The wind force itself is, however, not sufficient to break the twigs but impacts by drifting snow and ice particles may well impose sufficient strain to break off or damage twigs at low temperatures. Since the heavier snow particles travel very close to the snow surface it is when the snow surface is just below the tops of the plants that abrasion is most likely to occur. The snow surface is at a more constant level through the winter near the ridge crests and the patterns of snow accumulation probably vary less from one year to another near the ridge crests. Since abrasion by drifting snow occurs only occasionally, the probability of the snow depth being at a particular level is greater near ridge crests and therefore the tops of the plants are maintained at a relatively constant level in such locations.

3. Multiple regression equations

The equations in Table 6 (linear case)and Table 7 (with snow depths transformed for normality) were obtained using stepwise multiple regression with a criterion of .01 partial explanation for inclusion or deletion of independent variables. The variables are arranged in the equations in descending order of their individual T-values. The T-value is defined as the regression coefficient divided by the standard error of this coefficient.

The improvements given to these equations with vegetation variables included were only slight. In early winter the variable representing brush depth increased the coefficient of multiple correlation with .04 units and in late winter with .02 units. Tree density and brush density did not provide sufficient additional explanation to enter the equation in any one of the five surveys analysed. Therefore the vegetation variables were excluded from the set of independent variables. This allows the equations to be used with input of altitude readings only, which is a considerable advantage for potential applications of the method.

Two cases were investigated, one where linearity was assumed and one where the dependent variables were transformed to normal distributions. A generally better explanation was given by the model with snow depths transformed for normality. However, the difference is not great. A reduction in the standard error in late winter with less than 3 cm resulted from using a square root transformation of snow depth on March 13.

TABLE 6

MULTIPLE REGRESSION EQUATIONS FOR ESTIMATING SNOW DEPTH (LINEARITY ASSUMED)

AT FIVE DIFFERENT TIMES - WINTER 1968 - 1969

Equation Number	Date	R	S.E. (cm)	R^2
1.	27.10.68	0.346	8.82	0.132
$Y_1 = -0.80897\ X_{18} + 0.77616\ X_{48} - 0.49200\ X_{36} + 45.09634$				
2.	24.11.68	0.564	17.234	0.332
$Y_2 = -2.40488\ X_{37} + 1.26863\ X_{30} - 1.13915\ X_{58} - 1.56143\ X_{14} + 236.53474$				
3.	14.12.68	0.703	25.687	0.501
$Y_3 = -5.31008\ X_{50} - 2.85680\ X_{60} - 2.19122\ X_{39} + 577.56812$				
4.	9.02.69	0.804	38.783	0.651
$Y_4 = -9.19558\ X_{50} - 7.50734\ X_{15} + 2.89037\ X_{30} + 798.07690$				
5.	13.03.69	0.845	40.356	0.718
$Y_5 = -10.78660\ X_{50} - 5.75036\ X_{45} - 2.93291\ X_{23} + 1102.03662$				

Note: Index refers to the variable number (Table 3)
For equations of individual variables refer to Appendix II

TABLE 7

MULTIPLE REGRESSION EQUATIONS FOR ESTIMATING SNOW DEPTH (DEPENDENT VARIABLES TRANSFORMED FOR NORMALITY) AT FIVE DIFFERENT TIMES - WINTER 1968-1969

Equation Number	Date	R	S.E. (cm)	R^2 x 100
1.	27.10.68	0.44	8.7	0.20
	$Y_1 = (-0.04993\ X_{18} + 0.04098\ X_{48} - 0.02813\ X_{36} + 4.44781)^3$			
2.	24.11.68	0.63	16.8	0.41
	$\text{Log}\ Y_2 = -0.04028\ X_{14} - 0.01913\ X_{37} - 0.00830\ X_{32} + 0.00718\ X_{30} - 0.00039\ X_8 + 4.83799$			
3.	14.12.68	0.76	23.6	0.58
	$Y_3 = (-0.15389\ X_{18} + 0.05583\ X_{30} - 0.04371\ X_{39} - 0.05722\ X_{44} + 13.76053)^3$			
4.	9.02.69	0.83	35.5	0.69
	$Y_4 = (-0.10487\ X_{50} - 0.12189\ X_{15} - 0.08563\ X_{45} - 0.00211\ X_8 + 21.19156)^3$			
5.	13.03.69	0.86	37.5	0.75
	$Y_5 = (0.59952\ X_{50} - 0.11758\ X_{55} - 0.24733\ X_{45} - 0.00748\ X_8 + 63.01810)^2$			

Note: Index refers to the variable number (Table 3)
For equations of individual variables refer to Appendix II

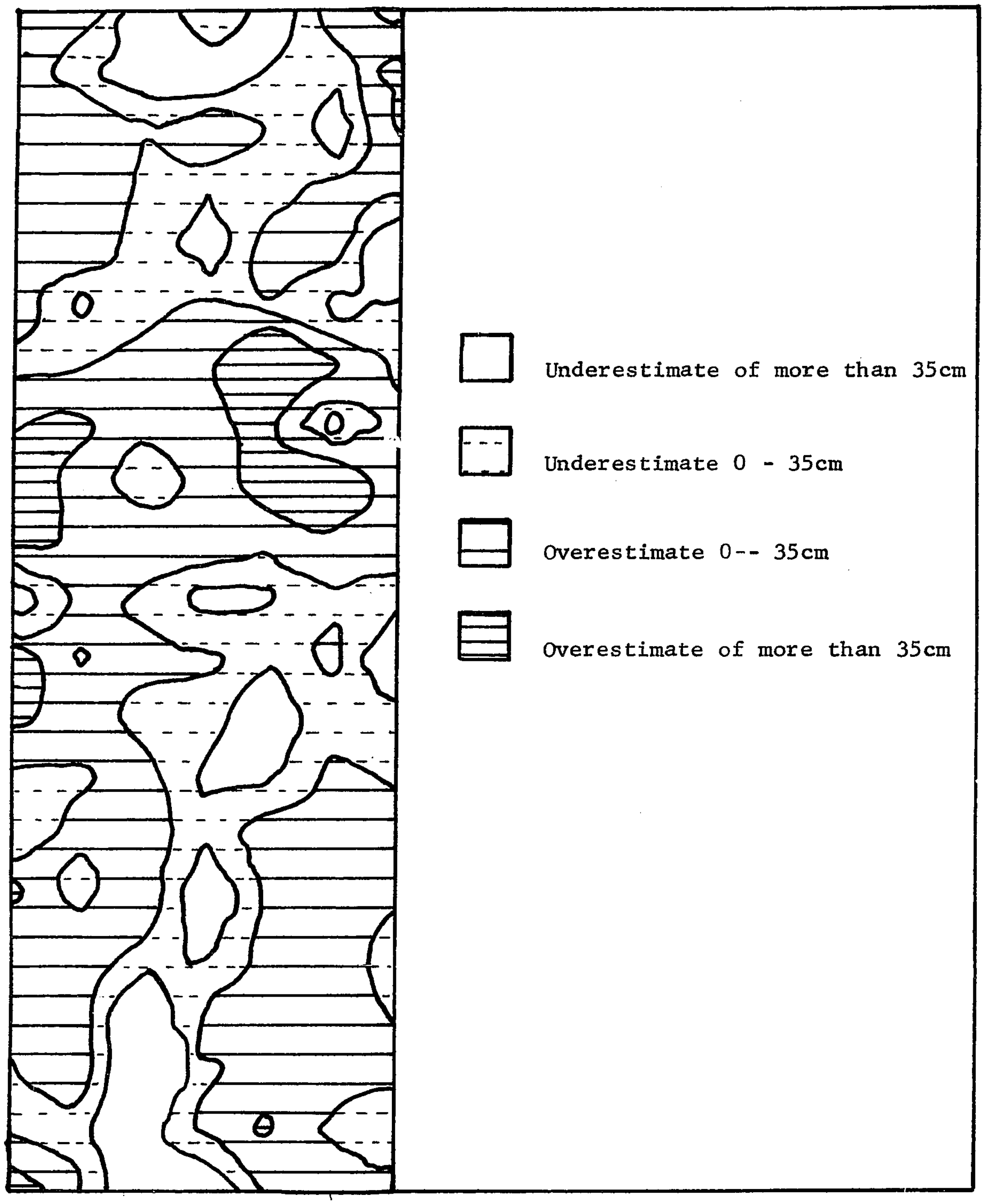


Figure 18 Map of residuals for equation from March 13

(snow depths transformed)

The explanation by the model is low in early winter. This may be expected since brush vegetation is not accounted for. Another factor is the more regular accumulation during the earlier stages of winter. It is possible that an expression of distance from ridge crests could improve the explanation somewhat in early winter but it would be very difficult to account for varying sizes of smooth areas. Part of the cause of the low explanations in early winter could also be the relatively low accuracy and large scale with which the topographic roughness variables were evaluated.

The explanation by the model increases through winter. For March 13 the equation with snow depths transformed gives a coefficient of multiple correlation of .864 which yields an explanation of 75 per cent of the original variation in snow depth. This equation includes three delay effects (7 NE, 7 W and 5 N) and altitude. Mapping the residuals from this equation (Figure 18) shows that the greatest overestimates occur in wide concavities (see Figure 2) where the northwest winds have free entry. The greatest underestimates occur near the bottom of other wide valleys or concavities which are less open to the northwest winds. This would suggest that the effects of northwest winds are not properly accounted for by the present set of variables. Underestimates also occur in the lower part of the experimental site along a zone which could be characterised as the zone where a general increase in tree density occurs. This indicates the need for a variable better representing the effect of trees.

4. Comparison of a map from snow melt photographs with a predicted map

a) Preparation of snow depth map from melt photographs

The depth of the snowcover by the end of the accumulation period was reconstructed for the purpose of testing the results obtained by the step-wise multiple regression technique, using aerial photographs. These were taken over the experimental site at intervals of one to three weeks during the snow melt period. The period between the flights was determined by melt rates and weather conditions.

Snow melt in the Schefferville area takes place at sufficiently uniform rates to provide good estimates of the depth variation of the cover at the beginning of thaw, simply by plotting the outline of remaining snow at the time of the different flights. This uniformity of the melt rates in different parts of the terrain is probably caused by the generally cloudy weather during spring, which reduces the effects of uneven melt due to direct solar radiation. Although additional snowfall was recorded during the melt period, its contribution to variations in melt rates was small, since the snow was wet and therefore relatively evenly distributed. The snow accumulated during the melt period generally melted within a few days after the snowfall, and the flights were timed so as to avoid effects of new snow on the limits of the old snow cover.

The photographs (Figures 19 to 23) were taken using a hand-held Mamiya 6 x 9 cm camera with a 65 mm lens through the side window of a Cessna 172 aircraft. The altitude at which overall views were obtained was approximately 5000 feet above the surface. Pictures for ground control were taken from an altitude of 1000 feet. The photographs from the lower level were used to locate the snow stakes, so that their exact location could be plotted onto the overall view, making it possible to rectify the photographs for errors arising

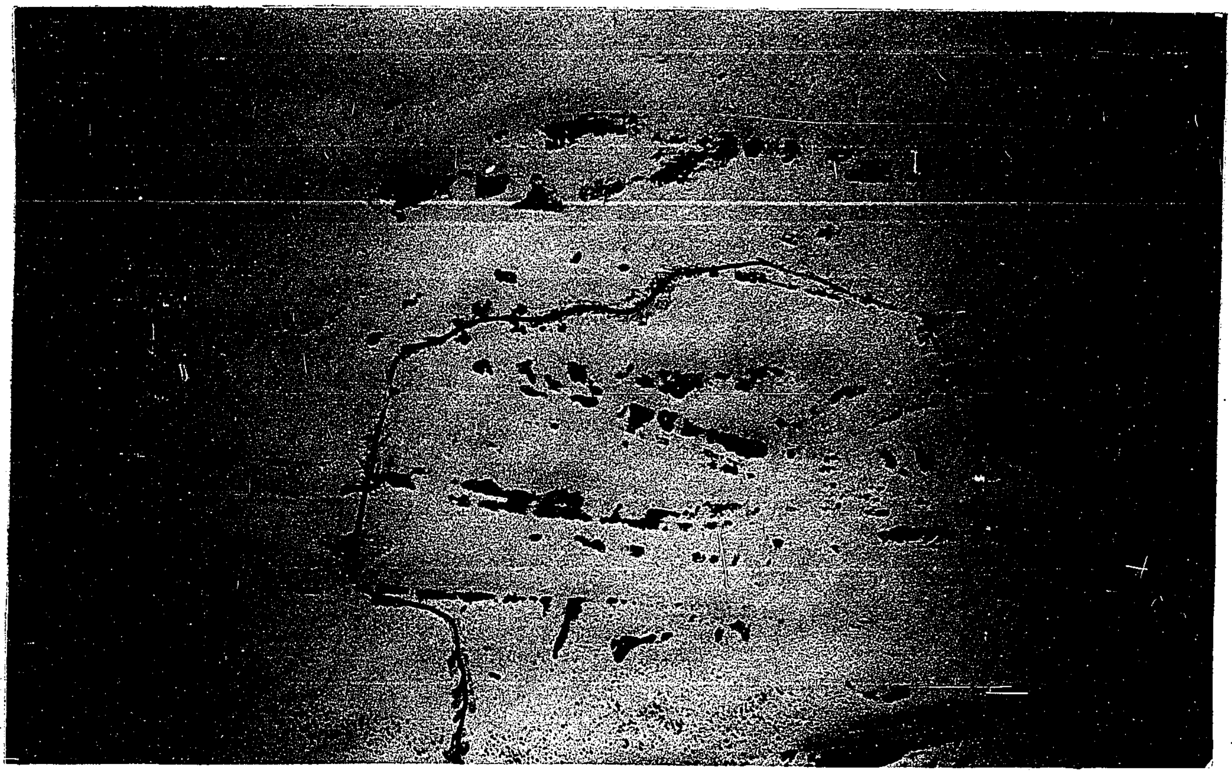


Figure 19. Photo of Timmins 4 on May 14, 1969.

This photo was taken soon after the melt was initiated at the end of April. Twenty centimetres average depth decrease was recorded between the time of maximum cover and May 14. Only ridge crests are exposed, which are visible as dark areas on the photo. Clearance of snow for a road to Howell's River has recently taken place. Deep parts of the snow are indicated by the way the tractor has forced its way through the snowpack, clearing the snow sideways. Analysis of the photograph reveals that 96.7 per cent of the experimental area is snow covered. The snow cover still remains largely unbroken in the north-western third of the experimental site.

Figure 20. Photo of Timmins 4 on June 5, 1969.

Spring melt is well advanced at this stage, three weeks after the second flight. The snow cover has decreased in extent from 96.7 to 65.1 per cent. The snow melt pattern is to some extent affected where the road clearance was undertaken. Melt is now apparent in the north-western third of the area, and the more even accumulation in this part stands in contrast to the very irregular accumulation in the mid-part of the experimental site. There is some indication of runoff, shown as darker areas of water saturated snow and open channels near the middle of the photograph.

Figure 21. Photo of Timmins 4 on June 12, 1969.

About 49.2 per cent of the area is now free from snow. The exposed areas on the ridges have widened, and consumed many of the small patches of thin snow noted in Figure 20. Vegetation begins to appear as black dots in areas of deep snowcover. In the northwestern third of the area, the melt pattern now delineates the convex parts of the terrain. The major valleys and depressions are well depicted through their deep unbroken snowcover. Where slope angles change abruptly the contact between snow and bare ground is sharp and linear. The upper, flat area of Timmins 4 is a good example of the previous, whereas the western edge of the same area shows a sharp snow margin towards the valley.

Figure 22. Photo of Timmins 4 on June 17.

At this stage 28.5 per cent of the survey area remains snow covered. Only in larger depressions and on steep lee slopes is snow still present. Particularly rapid change in surface coverage has taken place in the northwestern third of the area, indicating the more even accumulation of snow. The snow patches in this picture are frequently edged by darker shadings than elsewhere on the bare ground. This darker tone of the ground signifies vegetation different from the type exposed in Figure 20. In the lower left part of the picture, outside the sampling grid, radiation melt around trees is apparent. Within the snow course, where the influence of wind was stronger during the accumulation period, such features are not seen because of the generally deeper snow near the trees due to drift accumulation.

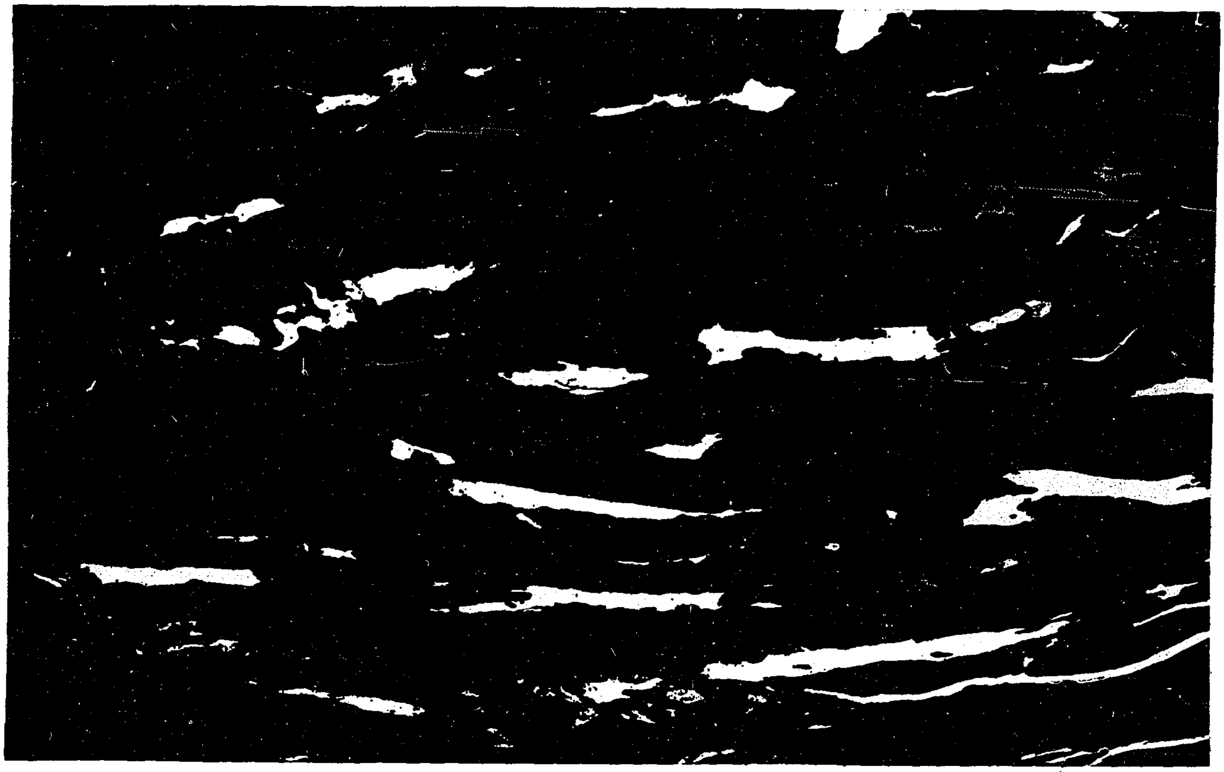


Figure 23. Photo of Timmins 4 on June 26, 1969.

At this stage only 7.5 per cent areal cover of snow remains. Parts of the snow patches are still 150 cm deep. In the field these deep accumulations show an apparent relationship to the local vegetation cover. Most of the snow patches in the picture are located near the base of southwest facing slopes, indicating strong influence of drift transport from east, northeast and north during this particular winter.

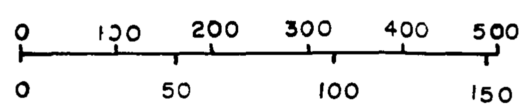
from distortion due to the high angle oblique view. The outline of the snow cover was then transferred by hand to a square grid, to obtain the map shown in Figure 24. The class limits of the map were determined by averaging the depth of snow on March 13 for the stakes laid bare between each flight. The mid-values of each depth increment were used as class limits.

Discussion of the snow depth map

Figure 25 is a plot of depths on March 13 for the stake locations first recorded bare at each flight. Joining the averages of each group of stakes gives a function which represents the cumulative depth decrease through time. The rates of depth decrease are relatively uniform, particularly if it is considered that the actual denudation of stake locations can have taken place at any point in time between flights. The range varies with depth and through time partly as an effect of the time spacing between flights and also partly as an effect of spatial variation in melt rates. Errors arising from the plotting of the map may be in the order of up to 20 m horizontal displacement of the contour lines. These errors are mainly due to difficulty in some cases in locating the snow stakes on the photographs.

The map gives a reasonably good estimate of the patterns of snow accumulation prior to thaw. The error varies through the range of snow depths and is smallest for shallow depths. This makes the method useful for indicating areas of shallow depths of snow, which are important in determining the greatest heat losses during winter. An improvement of the accuracy could be achieved by analyzing the factors influencing the rates of depth decrease of different parts of the area. Relating these factors to terrain variables would supply correction

FIGURE 24 SNOW DEPTH TIMMINS 4 E
FROM MELT PHOTOGRA

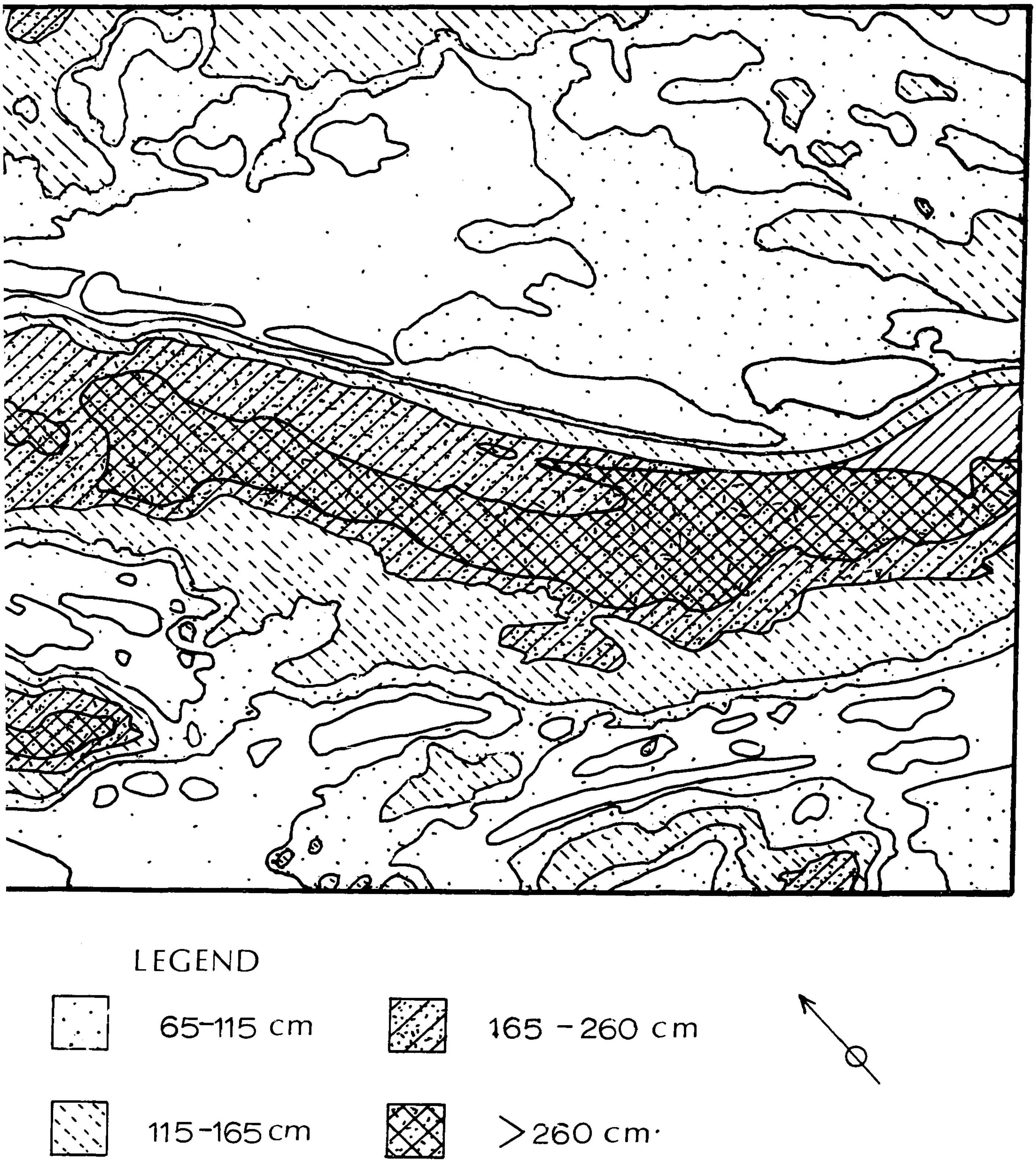


INS 4 ESTIMATED

OTOGRAPHS

LEGE

	0 -35 cm		65-
	35-65 cm		115-


LEGEND
65-115 cm
165 - 260 cm
115-165 cm
>260 cm

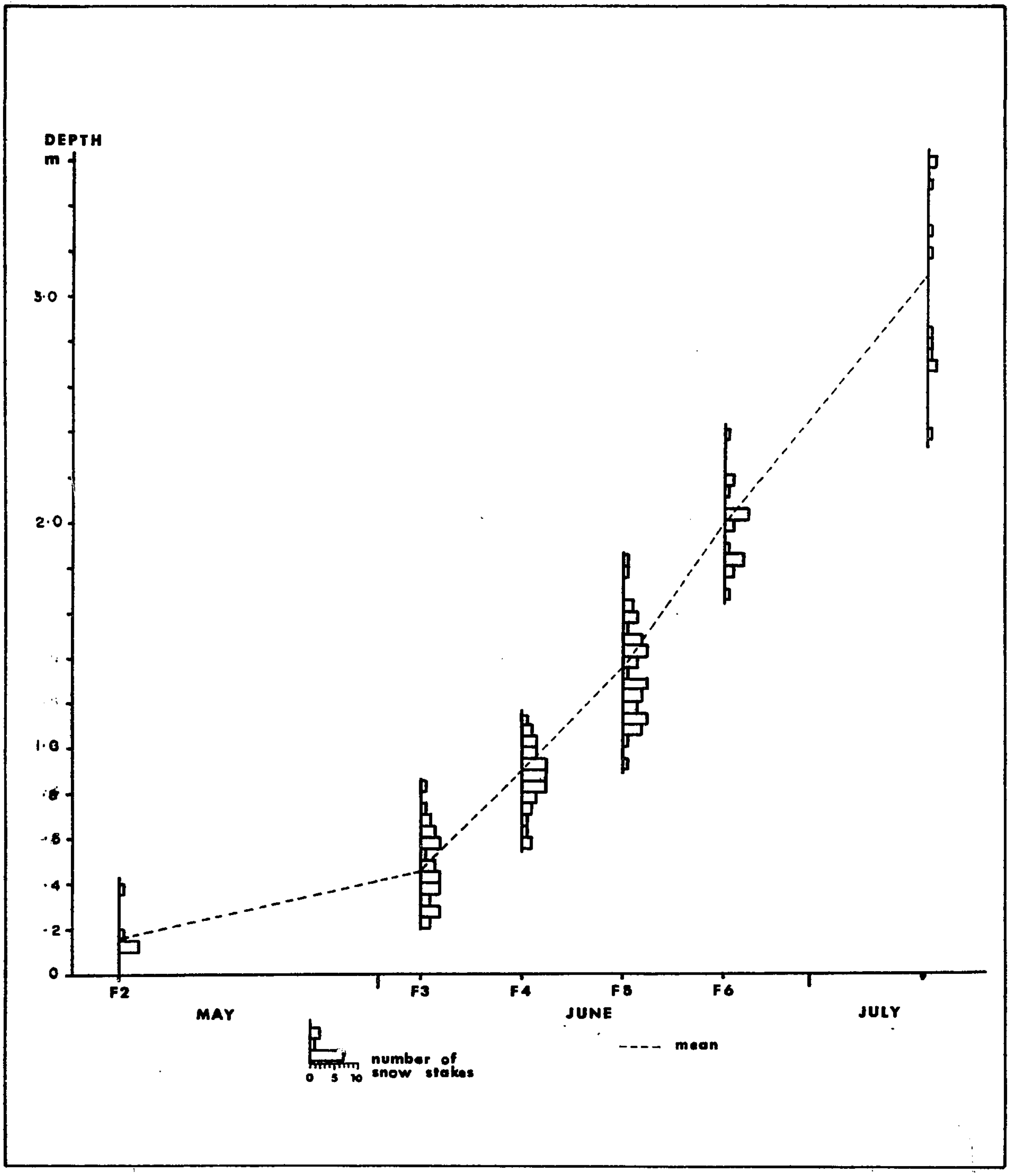


Figure 25 Snow depths on March 13 for stakes melted free between flights

factors for different locations in the terrain, and the map could be correspondingly adjusted to greater accuracy.

b) Preparation of predicted snow depth map for March 13

A predicted map of snow depth on March 13 was produced for the Timmins 4 site using the computer mapping program in Appendix I. A routine which compiles the individual terrain roughness variables for the particular equation and then produces an estimate of snow depth using the equation, was inserted into the mapping program. Snow depths were thus predicted at 50 ft (15 m) intervals on a square grid from an input consisting of a matrix of altitudes (see Chapter IV) and the equation for March 13 (snow depths normalised). In all 2241 snow depth estimates were calculated and the map (Figure 26) was produced by linear interpolation between these estimates. The accuracy of the map with respect to the sample has been indicated in the previous discussion.

c) Comparison between the two maps

The two snow depth maps (Figures 24 and 26) were obtained by different methods and therefore are somewhat different in appearance. The computer map has only four classes whereas the map from snow melt photographs has six. The computer map was produced by interpolations between data points 50 feet (15 m) apart which gives a certain degree of generalisation. This map therefore appears less detailed than the map produced from melt photographs, where the snow cover outline was plotted to as great a detail as possible.

A comparison between the two maps shows a good general correspondence. The deepest parts of the snowcover are similarly

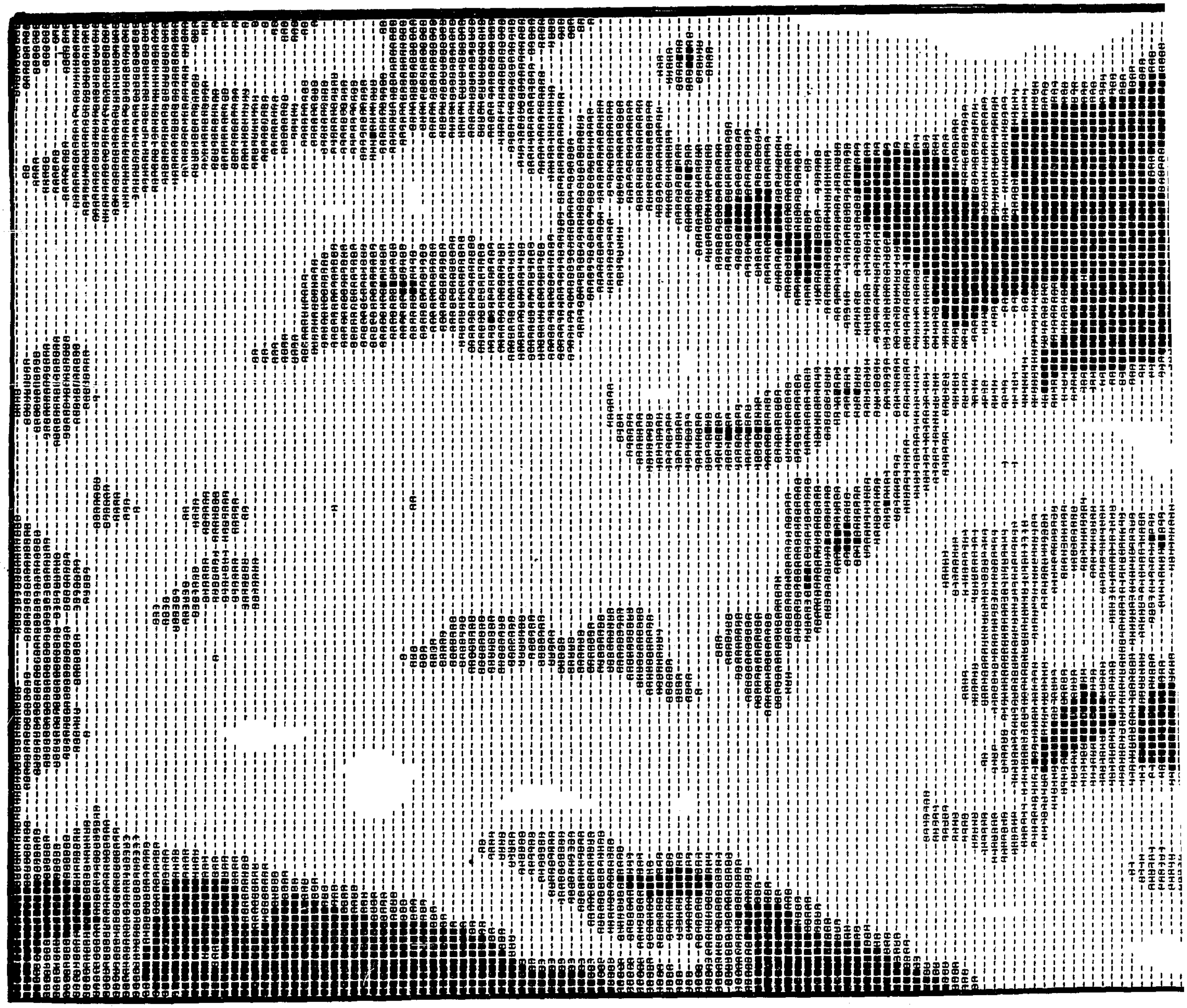
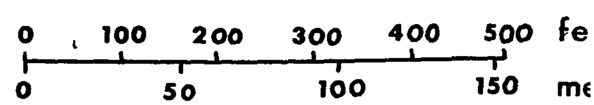
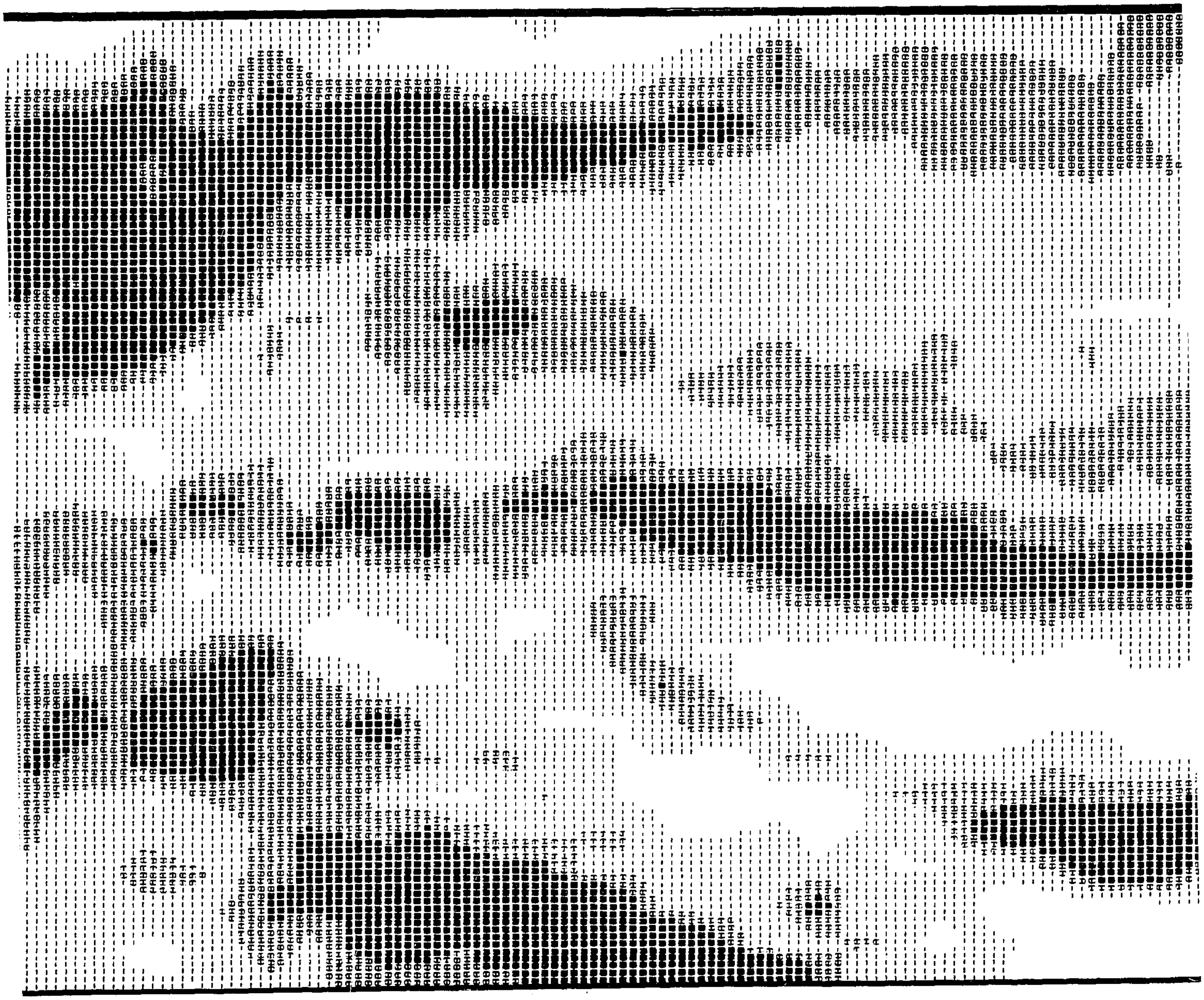


FIGURE 26 ESTIMATED SNOW DE





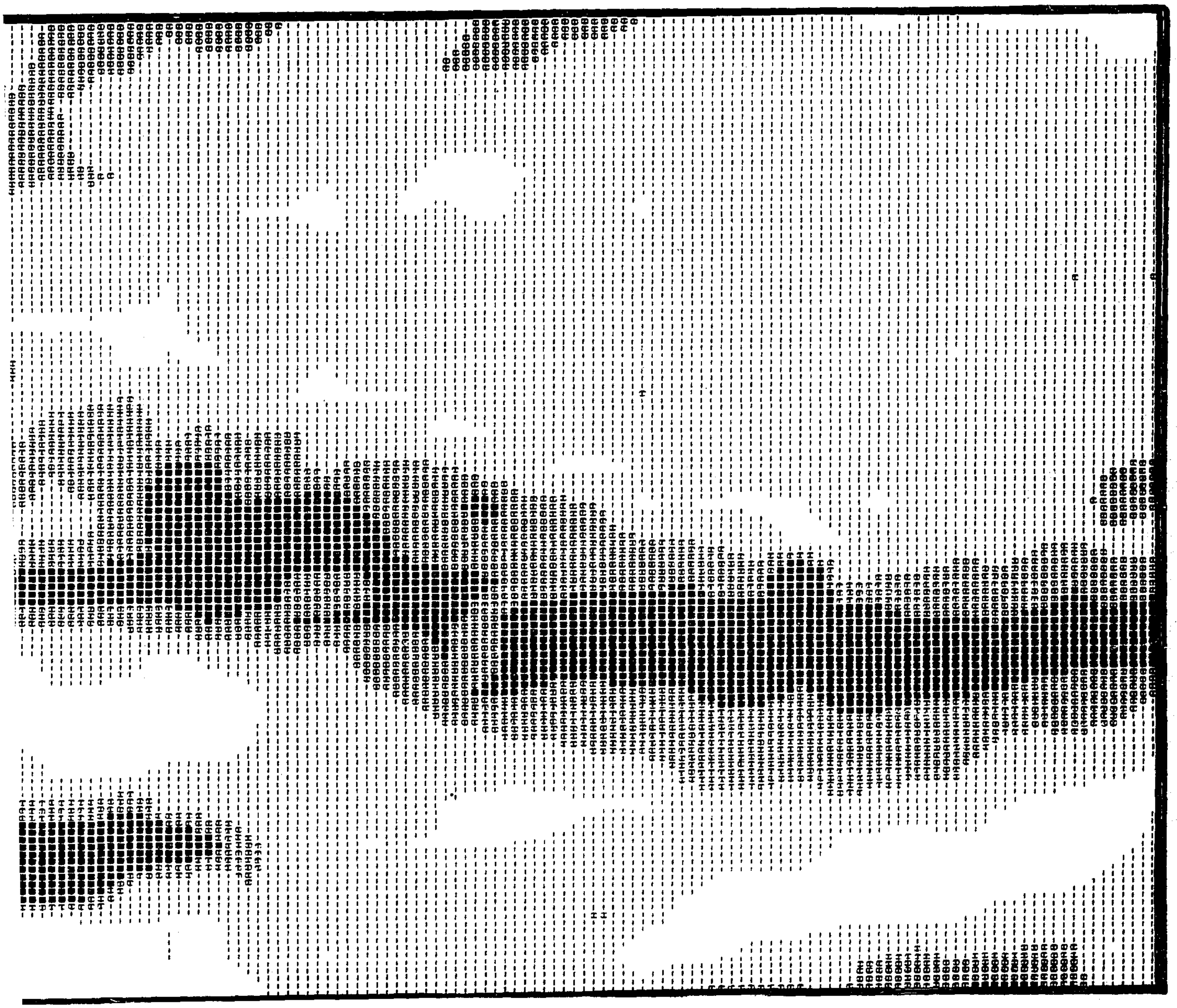
OW DEPTH TIMMINS 4

0-

50

100 500 feet

150 metres



LEGEND.

0–50 cm	100—200 cm
50—100 cm	> 200 cm

shown by both methods and the shallow zones agree well. As noted in the discussion of the residual there is a tendency for underestimate of the actual depth in the major valley in the more elevated part of the experimental site. The computer map gives a narrower zone of deep snow in this valley than the map from melt photographs. In the northwestern part of the experimental site there is a tendency for a slight underestimate by the computer map. The intermediate depths are more difficult to compare due to the differing class limits. The generally good agreement between the two maps suggests that the method of estimating snow depths from terrain roughness yields a good accuracy. To obtain comparable accuracy using actual field measurements it is estimated that more than 500 surveyed depth samples at the critical period would be needed.

A subsequent field application using the method in permafrost research indicates good general validity of the equations obtained at Timmins 4 (ms Nicholson and Granberg 1971).

CHAPTER VI

CONCLUSIONS

Two main results were produced by the present study. Firstly it shows that the accumulation of the seasonal snow cover takes place in a characteristic sequence which is determined by surface roughness. Secondly two methods were developed whereby the late winter snowcover may be mapped with good accuracy.

Three periods of characteristically different patterns of accumulation were identified:

1. An initial period when an overall rough microstructure of the ground surface produces a relatively even accumulation.

2. A transition period when the initially rough ground is gradually replaced by a smooth snow surface. Sharp discontinuities in surface roughness characterise this period. The depth increase is slow in smooth areas but rapid accumulation takes place at their margins.

3. A period when sharp discontinuities in surface roughness have been eliminated. Topographic roughness of a larger scale produces a pattern of accumulation which is more even. The greatest accumulations take place near valley bottoms whereas at ridge crests the depth remains constant.

Six characteristic trends of depth increase were observed, each of which is related to particular terrain locations:

1. Ridge crest locations reach maximum depth during the first snowfall and show little or no increase thereafter.

2. Near ridge crests there is an early period of rapid increase when the location becomes incorporated into the smooth area. Thereafter the depth remains constant or increases only slowly.

3. Further from the ridge crest the initial increase in

depth is relatively slow. A period of very rapid increase in depth occurs as the location becomes incorporated into the smooth area. Thereafter the depth continues to increase but at a reduced rate.

4. In wider valleys and concavities the snow depth may show an initial increase and then remain constant or near constant for some time. Thereafter the depth increases at an increasing rate. These locations exhibit the highest rates of accumulation in late winter.

5. In areas of homogeneous character, where little wind stress variations are produced by larger scale roughness features, the depth of snow begins to increase in some proportion to snowfall once minor roughness has been eliminated.

6. Some areas exhibit a depth increase which is approximately proportional to snowfall throughout winter. This trend is usually associated with relatively flat areas that are homogeneous with respect to vegetation cover and have little microrelief.

A comparison of snow water equivalents at the exposed Timmins 4 site with a survey in the more sheltered Knob Lake drainage basin does not indicate any significant difference in snow accumulation. Comparing the water equivalents obtained by field sampling in March, to snowfall measurements by the Nipher gauge in Schefferville indicates an underestimate by the Nipher gauge of just over 40 per cent. A comparison of snow depth increase at Timmins 4 with water equivalents measured by the Nipher gauge in Schefferville indicates that the snow sampling grid at Timmins 4 was representative for snowfall throughout the period investigated. A traverse survey from the ridge area where Timmins 4 is located into dense uniform forest suggests that the influence of windblown snow may be traced for a distance of about 1 km into the forest. A tentative estimate based on this survey suggests that less than 5 per cent of the total snowfall was lost from the general ridge area.

A stepwise multiple regression model was used in an attempt to quantitatively relate snow depth to terrain roughness. Topographic variables were obtained using altitudes from a map. The topographic variables were evaluated for eight different directions to account for varying influence of winds and surface roughness. Some of the variables were evaluated on different scales to account for different stages of winter. Vegetation variables were obtained by field measurements and remote sensing. Simple correlation was used for an initial scanning of the independent variables and indicated that none of the three vegetation variables gave a close representation of the influence of vegetation on snow accumulation. This conclusion was confirmed by scatter plots of brush depth and tree density. Tests of the stepwise multiple regression model with vegetation variables included, yielded only very slight improvements and the vegetation variables were therefore deleted.

The model shows a low explanation in early winter but the explanation improves through time. With topographic variables only, the model gives a statistical explanation of 75 per cent of the initial variation ($R = .86$) in an equation obtained for mid-March. This equation uses four variables.

A computer mapping program was written which produces a map of estimated snow depths from an input of a matrix of altitudes and the equation for the particular point in time. A map of predicted snow depths was produced for mid-March using this program.

Using sequence aerial photographs of snow melt a different map was produced of the snow depths at Timmins 4 also for mid-March. A comparison shows good agreement between the two maps.

A depth and water equivalent survey on Timmins 4 shows that water equivalents may be closely estimated from snow depth by a second order equation. Thus the depth map may be converted into a map of water equivalents using this equation in the mapping program.

The model shows deficiencies, some of which could be corrected by future work. The low explanation by vegetation variables could possibly be improved. In the case of trees, a variable expressing changes in forest density could possibly be more significant. This variable would need to be evaluated for different directions. The low explanations in early winter is partly due to the discontinuous relation between brush depth and snow depth. A better explanation could be achieved by limiting the range of brush employed in the equation to the range where the relationship exists for each particular point in time. The irregular pattern of the accumulation during the transition period is difficult to predict but a somewhat better result could possibly be achieved using a variable representing distance from ridge crests. Residual mapping indicates the need for a different variable to account for effects of northwest winds. It is apparent that a linear relationship between terrain variables and snow depth cannot be assumed and a further investigation of the functional relationships should be made.

In summary, the study has provided insight into an aspect of snow accumulation which has previously received little attention. A consideration of the sequence in which the snowcover accumulates may aid in explaining a number of phenomena such as patterns of permafrost, semi-permanent snow beds, or variations in runoff from one snowmelt period to another.

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APPENDIX I

A COMPUTER MAPPING PROGRAMME FOR DATA COLLECTED ON A REGULAR GRID

COMPUTER MAPPING PROGRAMME

DESCRIPTION

This computer mapping program is written in Fortran IV. The present version is intended for use on the IBM, System 360/75. The program may, however, be used on other systems with facilities for multiple overprint. The program is useful for mapping any data collected on a square or rectangular grid. Two types of maps may be obtained. The first type is a printout, where values within specified class intervals are represented by a letter symbol. In the second type of map, the mean and the standard deviation of the sample are calculated and the mean, the mean plus one standard deviation, and the mean minus one standard deviation are used as class limits. The output in this case is a greyscale comprising four classes and produced by multiple overprint of different combinations of symbols. If the width of each map does not exceed 65 outprint characters horizontally, both types of maps may be printed side-by-side. If missing data exist within the sample, such points are indicated with a particular symbol on the maps, providing the value of the missing data has been specified on the specification card.

COMPUTATIONAL PROCEDURE

The input data are stored in matrix A, which in the present version, is dimensioned for 21 data points vertically, 7 points horizontally and 13 sets of data to be mapped. The mean and standard deviation of one sample is calculated and this set of data is placed into matrix C, which has the vertical dimension equal to the number of lines comprising the map. The horizontal dimension equals the number of data points horizontally. The data is placed into the matrix, corresponding to their position in the map, and empty spaces between data points are filled with values through linear interpolation vertically. By scanning matrix C and performing equivalent operations horizontally, the operational vector D is generated. Vectors E and F are output vectors, which are filled with symbols corresponding to the values in vector D. These symbols are initially read from the first specification card and stored in storage vector S.

USAGE:

The deck setup is:

1. Program,
2. Two specification cards,
3. Data sets.

The first specification card should be punched as follows:

Col. 1-7 S-should always be -OMW 1 (symbols for map type 2).

8-40S-A B C D E F G H....or any map symbols desired for map type 1.

41-42 N-Number of data points horizontally on map.

43-44 M-Number of data points vertically on map.

45-46 NH-Number of characters per data point horizontally (i.e. D.....D.....D gives 05 in Col. 45-46).

47-48 NV-Number of characters per data point vertically.

49-50 MN-Number of maps with corresponding data sets.

51 MT-Map type, 1 of type 1, 2 of type 2, 3 if side-by-side printout is desired.

52-55 DM-Indicator of missing data (same as on data card).

56-59 CW-Classwidth for map type 1.

60-80 FO-Output for map (always 1H , 131A1).

The second specification card contains two output formats and the input format for data.

Col. 1-12 FOO-Multiple overprint format (1H+, 131A1).

13-32 FOD-Output format for side-by-side outprint (1H+, 61A1, 5X, 61A1).

33-80 FI-Input format for data.

SCALING OF MAPS:

The desired scale of the maps is obtained through changing the values of NV and NH. Ordinary IBM printers have 10 characters per inch horizontally and 6 characters per inch vertically, but other vertical spacings may be obtained if specified on //EXEC. SETUP-Card.

DIMENSIONING:

Only matrix A and C need to be adapted to particular problems. The dimensioning is done as outlined in the description of the computational procedure.

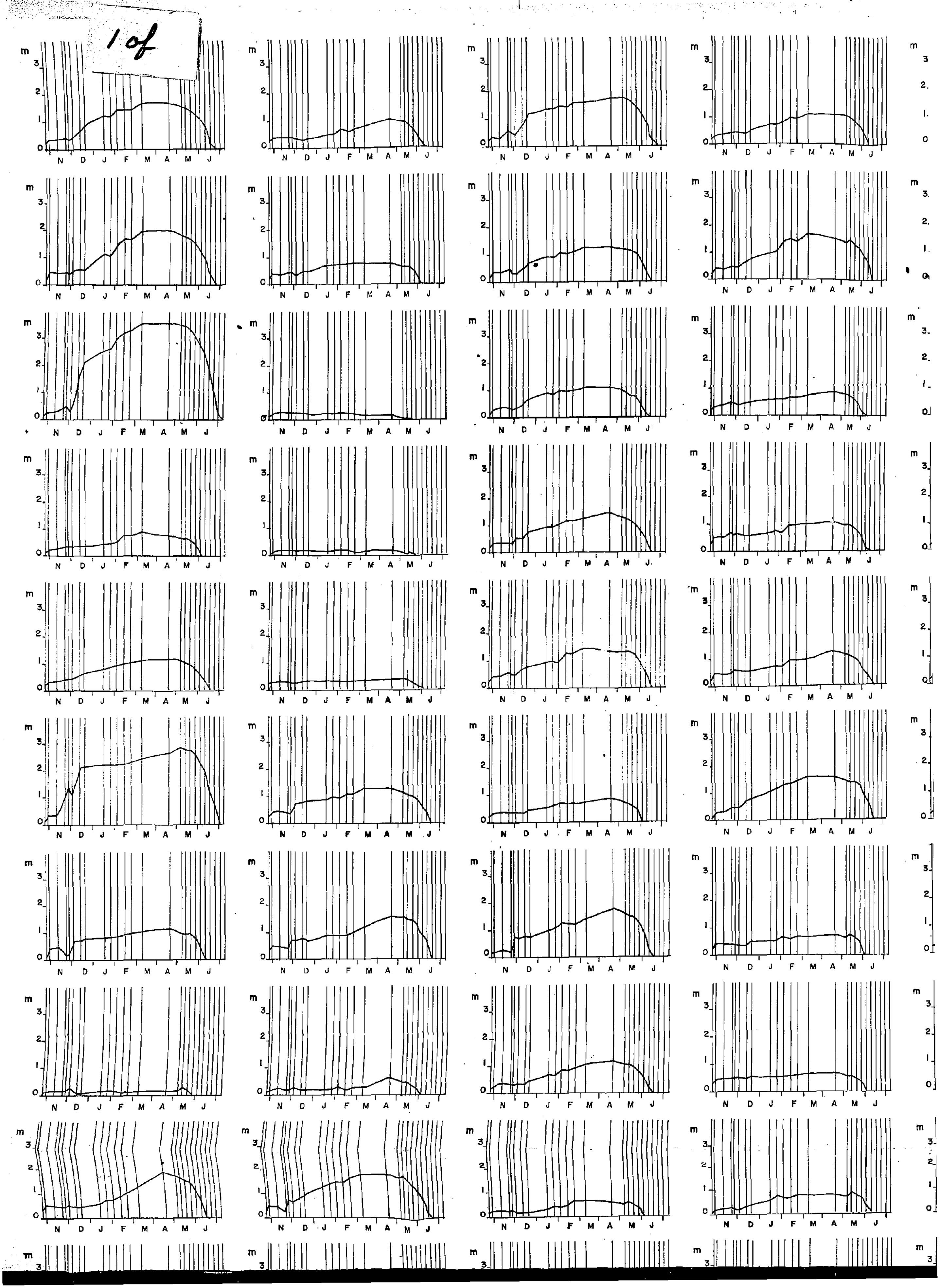
```
      DIMENSION A(21,7,13),C(161,7),D(131),E(131),F(61),S(40),B(4),FO(3)
     1,FOO(3),FI(12),FOD(5)
      READ(5,1)(S(I),I=1,40),N,M,NH,NV,MN,MT,DM,CW,(FO(K),K=1,3),(FOO(KA
     1),KA=1,3),(FOD(I),I=1,5),(FI(I),I=1,12)
    1 FORMAT(40A1,5I2,I1,2F4.1,3A4/3A4,17A4)
      NH1=NH-1
      NH2=2*NH-1
      NV1=NV-1
      NV2=2*NV-1
      ME=NV*M-NV1
      MEV=NV*M-NV2
      NE=NH*N-NH1
      NEV=NH*N-NH2
      READ(5,FI)(((A(I,J,MAP),MAP=1,MN),J=1,N),I=1,M)
      DO 10 MAP=1,MN
      DNO=M*N
      SUMB=0.0
      SUMB2=0.0
      DO 2 I=1,M
      DO 2 J=1,N
      IF(A(I,J,MAP).EQ.DM) DNO=DNO-1.0
      SUMB=SUMB+A(I,J,MAP)
      SUMB2=SUMB2+A(I,J,MAP)**2
    2 C(NV*I-NV1,J)=A(I,J,MAP)
      B(2)=SUM3/DNO
      BSD=SQRT(DNO*SUMB2-SUMB**2)/DNO
      B(1)=B(2)-BSD
      B(3)=B(2)+BSD
      B(4)=B(3)
      WRITE(6,3)MAP,B(2),BSD,B(1),B(2),B(3)
    3 FORMAT(1H1,11HMAP NUMBER ,I2,10H    MEAN = ,F7.2,24H    STANDARD DEV
     1IATION = ,F5.2,/,22H CLASS LIMITS ARE:    <,F8.2,3H  -,F8.2,3H  -,F
     28.2,3H  <,//)
      DO 4 I=1,MEV,NV
      DO 4 J=1,N
      DO 4 N1=1,NV1
      CN=N1
    4 C(I+N1,J)=((FLOAT(NV)-CN)*C(I,J)+CN*C(I+NV,J))/FLOAT(NV)
      DO 10 I=1,ME
      DO 5 J=1,N
    5 D(NH*J-NH1)=C(I,J)
      DO 6 J=1,NEV,NH
      DO 6 M1=1,NH1
      CM=M1
    6 D(J+M1)=((FLOAT(NH)-CM)*D(J)+CM*D(J+NH))/FLOAT(NH)
      IF(MT.EQ.2) GO TO 8
      DO 7  J=1,NE
      JC=((D(J)+1.0)/CW)+17.0
      E(J)=S(5)
    7 E(J)=S(JC)
      IF(MT.EQ.3) GO TO 20
      GO TO 12
   20 DO 21 J=1,NE
      F(J)=S(5)
   21 F(J)=F(J)
    8 JB=1
   18 DO 9 J=1,NE
      E(J)=S(5)
      IF(B(JB).LT.D(J)) E(J)=S(JB)
      IF(D(J).EQ.DM) E(J)=S(6)
    9 CONTINUE
      IF(JB.EQ.1.AND.MT.EQ.3) GO TO 11
      IF(JB.GT.1) GO TO 13
   12 WRITE(6,FO)(E(J),J=1,NE)
      GO TO 40
   11 WRITE(6,FOD)(E(J),J=1,NE),(F(J),J=1,NE)
      GO TO 40
   13 WRITE(6,FOO)(E(J),J=1,NE)
   40 JB=JB+1
      IF(JB.LE.4.AND.MT.GT.1) GO TO 18
   10 CONTINUE
      STOP
      END
```

APPENDIX II

SECTION OF COMPUTER PROGRAM GENERATING TOPOGRAPHIC VARIABLES

```
DO 30 J=4,114
KH=KH+1
B(KH,8)=A(I,J)
B(KH,9)=(A(I,J)-((A(I,J)+A(I+1,J-1)+A(I-1,J+1))/3.0))+50.0
B(KH,10)=(A(I,J)-((A(I,J)+A(I,J-1)+A(I,J+1))/3.0))+50.0
B(KH,11)=(A(I,J)-((A(I,J)+A(I-1,J-1)+A(I+1,J+1))/3.0))+50.0
B(KH,12)=(A(I,J)-((A(I,J)+A(I-1,J)+A(I+1,J)))/3.0)+50.0
B(KH,13)=(A(I,J)-((A(I,J)+A(I+1,J-1)+A(I-1,J+1)+A(I+2,J-2)+A(I-2,J
1+2))/5.0))+50.0
B(KH,14)=(A(I,J)-((A(I,J)+A(I,J-1)+A(I,J+1)+A(I,J-2)+A(I,J+2))/5.0
1))+50.0
B(KH,15)=(A(I,J)-((A(I,J)+A(I-1,J-1)+A(I+1,J+1)+A(I-2,J-2)+A(I+2,J
1+2))/5.0))+50.0
B(KH,16)=(A(I,J)-((A(I,J)+A(I-1,J)+A(I+1,J)+A(I-2,J)+A(I+2,J))/5.0
1))+50.0
B(KH,17)=(A(I,J)-((A(I,J)+A(I+1,J-1)+A(I-1,J+1)+A(I+2,J-2)+A(I-2,J
1+2)+A(I+3,J-3)+A(I-3,J+3))/7.0))+50.0
B(KH,18)=(A(I,J)-((A(I,J)+A(I,J-1)+A(I,J+1)+A(I,J-2)+A(I,J+2)+A(I,
1J-3)+A(I,J+3))/7.0))+50.0
B(KH,19)=(A(I,J)-((A(I,J)+A(I-1,J-1)+A(I+1,J+1)+A(I-2,J-2)+A(I+2,J
1+2)+A(I-3,J-3)+A(I+3,J+3))/7.0))+50.0
B(KH,20)=(A(I,J)-((A(I,J)+A(I-1,J)+A(I+1,J)+A(I-2,J)+A(I+2,J)+A(I-
13,J)+A(I+3,J))/7.0))+50.0
B(KH,21)=(A(I,J)-((A(I,J)+A(I+1,J-1)+A(I-1,J+1)+A(I+2,J-2)+A(I-2,J
1+2)+A(I+3,J-3)+A(I-3,J+3)+A(I+4,J-4)+A(I-4,J+4))/9.0))+50.0
B(KH,22)=(A(I,J)-((A(I,J)+A(I,J-1)+A(I,J+1)+A(I,J-2)+A(I,J+2)+A(I,
1J-3)+A(I,J+3)+A(I,J-4)+A(I,J+4))/9.0))+50.0
B(KH,23)=(A(I,J)-((A(I,J)+A(I-1,J-1)+A(I+1,J+1)+A(I-2,J-2)+A(I+2,J
1+2)+A(I-3,J-3)+A(I+3,J+3)+A(I-4,J-4)+A(I+4,J+4))/9.0))+50.0
B(KH,24)=(A(I,J)-((A(I,J)+A(I-1,J)+A(I+1,J)+A(I-2,J)+A(I+2,J)+A(I-
13,J)+A(I+3,J)+A(I-4,J)+A(I+4,J))/9.0))+50.0
B(KH,25)=(A(I,J)-((A(I,J)+A(I+1,J-1)+A(I-1,J+1)+A(I+2,J-2)+A(I-2,J
1+2)+A(I+3,J-3)+A(I-3,J+3)+A(I+4,J-4)+A(I-4,J+4)+A(I+5,J-5)+A(I-5,J
1+5))/11.0))+50.0
B(KH,26)=((A(I,J)-(A(I,J)+A(I,J-1)+A(I,J+1)+A(I,J-2)+A(I,J+2)+A(I,
1J-3)+A(I,J+3)+A(I,J-4)+A(I,J+4)+A(I,J-5)+A(I,J+5))/11.0))+50.0
B(KH,27)=(A(I,J)-((A(I,J)+A(I-1,J-1)+A(I+1,J+1)+A(I-2,J-2)+A(I+2,J
1+2)+A(I-3,J-3)+A(I+3,J+3)+A(I-4,J-4)+A(I+4,J+4)+A(I-5,J-5)+A(I+5,J
2+5))/11.0))+50.0
B(KH,28)=(A(I,J)-((A(I,J)+A(I-1,J)+A(I+1,J)+A(I-2,J)+A(I+2,J)+A(I-
13,J)+A(I+3,J)+A(I-4,J)+A(I+4,J)+A(I-5,J)+A(I+5,J))/11.0))+50.0
B(KH,29)=(((A(I,J)+A(I,J+1)+A(I,J-1)+A(I,J+2)+A(I,J-2)+A(I,J+3)+A(
1I,J+4))/7.0)-((A(I,J)+A(I,J+1)+A(I,J-1)+A(I,J+2)+A(I,J-2)+A(I,J-3)
2+A(I,J-4))/7.0))+50.0
B(KH,30)=(((A(I,J)+A(I-1,J+1)+A(I+1,J-1)+A(I-2,J+2)+A(I+2,J-2)+A(I
1-3,J+3)+A(I-4,J+4))/7.0)-((A(I,J)+A(I-1,J+1)+A(I+1,J-1)+A(I-2,J+2)
2+A(I+2,J-2)+A(I+3,J-3)+A(I+4,J-4))/7.0))+50.0
B(KH,31)=(((A(I,J)+A(I+1,J+1)+A(I-1,J-1)+A(I+2,J+2)+A(I-2,J-2)+A(I
1+3,J+3)+A(I+4,J+4))/7.0)-((A(I,J)+A(I+1,J+1)+A(I-1,J-1)+A(I+2,J+2)
2+A(I-2,J-2)+A(I-3,J-3)+A(I-4,J-4))/7.0))+50.0
B(KH,32)=(((A(I,J)+A(I+1,J)+A(I-1,J)+A(I+2,J)+A(I-2,J)+A(I+3,J)+A(
1I+4,J))/7.0)-((A(I,J)+A(I+1,J)+A(I-1,J)+A(I+2,J)+A(I-2,J)+A(I-3,J)
2+A(I-4,J))/7.0))+50.0
B(KH,33)=(A(I,J)-((A(I,J)+A(I+1,J-1)+A(I-1,J+1)+A(I+2,J-2)+A(I-2,J
1+2)+A(I+3,J-3)+A(I-3,J+3)+A(I+4,J-4)+A(I-4,J+4)+A(I+5,J-5)+A(I-5,J
1+5)+A(I+6,J-6)+A(I-6,J+6)+A(I+7,J-7)+A(I-7,J+7))/15.0))+50.0
B(KH,34)=(A(I,J)-((A(I,J)+A(I,J-1)+A(I,J+1)+A(I,J-2)+A(I,J+2)+A(I,
1J-3)+A(I,J+3)+A(I,J-4)+A(I,J+4)+A(I,J-5)+A(I,J+5)+A(I,J-6)+A(I,J+6
1)+A(I,J-7)+A(I,J+7))/15.0))+50.0
B(KH,35)=(A(I,J)-((A(I,J)+A(I-1,J-1)+A(I+1,J+1)+A(I-2,J-2)+A(I+2,J
1+2)+A(I-3,J-3)+A(I+3,J+3)+A(I-4,J-4)+A(I+4,J+4)+A(I-5,J-5)+A(I+5,J
2+5)+A(I-6,J-6)+A(I+6,J+6)+A(I-7,J-7)+A(I+7,J+7))/15.0))+50.0
B(KH,36)=(A(I,J)-((A(I,J)+A(I-1,J)+A(I+1,J)+A(I-2,J)+A(I+2,J)+A(I-
13,J)+A(I+3,J)+A(I-4,J)+A(I+4,J)+A(I-5,J)+A(I+5,J)+A(I-6,J)+A(I+6,J
2)+A(I-7,J)+A(I+7,J))/15.0))+50.0
B(KH,37)=((A(I,J)-A(I-1,J+1))-(((A(I,J)+A(I-1,J+1)+A(I+1,J-1)+A(I-
12,J+2)+A(I+2,J-2)+A(I-3,J+3)+A(I+3,J-3))/7.0)-((A(I,J)+A(I-1,J+1)+
2A(I+1,J-1)+A(I-2,J+2)+A(I+2,J-2)+A(I-3,J+3)+A(I-4,J+4))/7.0)))+50.
30
B(KH,38)=((A(I,J)-A(I,J+1))-(((A(I,J)+A(I,J-1)+A(I,J+1)+A(I,J-2)+A
1(I,J+2)+A(I,J-3)+A(I,J+3))/7.0)-((A(I,J)+A(I,J+1)+A(I,J-1)+A(I,J-2
2)+A(I,J+2)+A(I,J+3)+A(I,J+4))/7.0)))+50.0
B(KH,39)=((A(I,J)-A(I+1,J+1))-(((A(I,J)+A(I+1,J+1)+A(I-1,J-1)+A(I+
12,J+2)+A(I-2,J-2)+A(I+3,J+3)+A(I-3,J-3))/7.0)-((A(I,J)+A(I+1,J+1)+
2A(I-1,J-1)+A(I+2,J+2)+A(I-2,J-2)+A(I+3,J+3)+A(I+4,J+4))/7.0)))+50.
30
B(KH,40)=((A(I,J)-A(I+1,J))-(((A(I,J)+A(I+1,J)+A(I-1,J)+A(I+2,J)+A
1(I-2,J)+A(I+3,J)+A(I-3,J))/7.0)-((A(I,J)+A(I+1,J)+A(I-1,J)+A(I+2,J
2)+A(I-2,J)+A(I+3,J)+A(I+4,J))/7.0)))+50.0
B(KH,41)=((A(I,J)-A(I+1,J-1))-(((A(I,J)+A(I-1,J+1)+A(I+1,J-1)+A(I-
12,J+2)+A(I+2,J-2)+A(I-3,J+3)+A(I+3,J-3))/7.0)-((A(I,J)+A(I-1,J+1)+
2A(I+1,J-1)+A(I-2,J+2)+A(I+2,J-2)+A(I+3,J-3)+A(I+4,J-4))/7.0)))+50.
30
```

```
      B(KH,42)=((A(I,J)-A(I,J-1))-(((A(I,J)+A(I,J-1)+A(I,J+1)+A(I,J-2)+A
     1(I,J+2)+A(I,J-3)+A(I,J+3))/7.0)-((A(I,J)+A(I,J+1)+A(I,J-1)+A(I,J-2
     2)+A(I,J+2)+A(I,J-3)+A(I,J-4))/7.0)))+50.0
      B(KH,43)=((A(I,J)-A(I-1,J-1))-(((A(I,J)+A(I+1,J+1)+A(I-1,J-1)+A(I+
     12,J+2)+A(I-2,J-2)+A(I+3,J+3)+A(I-3,J-3))/7.0)-((A(I,J)+A(I+1,J+1)+
     2A(I-1,J-1)+A(I+2,J+2)+A(I-2,J-2)+A(I-3,J-3)+A(I-4,J-4))/7.0)))+50.
     30
      B(KH,44)=((A(I,J)-A(I-1,J))-(((A(I,J)+A(I+1,J)+A(I-1,J)+A(I+2,J)+A
     1(I-2,J)+A(I+3,J)+A(I-3,J))/7.0)-((A(I,J)+A(I+1,J)+A(I-1,J)+A(I+2,J
     2)+A(I-2,J)+A(I-3,J)+A(I-4,J))/7.0)))+50.0
      B(KH,45)=(A(I,J)-((A(I,J)+A(I-1,J+1)+A(I+1,J-1)+A(I-2,J+2)+A(I-3,J
     1+3))/5.0))+50.0
      B(KH,46)=(A(I,J)-((A(I,J)+A(I,J-1)+A(I,J+1)+A(I,J+2)+A(I,J+3))/5.0
     1))+50.0
      B(KH,47)=(A(I,J)-((A(I,J)+A(I-1,J-1)+A(I+1,J+1)+A(I+2,J+2)+A(I+3,J
     1+3))/5.0))+50.0
      B(KH,48)=(A(I,J)-((A(I,J)+A(I-1,J)+A(I+1,J)+A(I+2,J)+A(I+3,J))/5.0
     1))+50.0
      B(KH,49)=(A(I,J)-((A(I,J)+A(I-1,J+1)+A(I+1,J-1)+A(I-2,J+2)+A(I+2,J
     1-2)+A(I-3,J+3)+A(I-4,J+4))/7.0))+50.0
      B(KH,50)=(A(I,J)-((A(I,J)+A(I,J-1)+A(I,J+1)+A(I,J-2)+A(I,J+2)+A(I,
     1J+3)+A(I,J+4))/7.0))+50.0
      B(KH,51)=(A(I,J)-((A(I,J)+A(I-1,J-1)+A(I+1,J+1)+A(I-2,J-2)+A(I+2,J
     1+2)+A(I+3,J+3)+A(I+4,J+4))/7.0))+50.0
      B(KH,52)=(A(I,J)-((A(I,J)+A(I-1,J)+A(I+1,J)+A(I-2,J)+A(I+2,J)+A(I+
     13,J)+A(I+4,J))/7.0))+50.0
      B(KH,53)=(A(I,J)-((A(I,J)+A(I-1,J+1)+A(I+1,J-1)+A(I-2,J+2)+A(I+2,J
     1-2)+A(I+3,J-3)+A(I+4,J-4))/7.0))+50.0
      B(KH,54)=(A(I,J)-((A(I,J)+A(I,J-1)+A(I,J+1)+A(I,J-2)+A(I,J+2)+A(I,
     1J-3)+A(I,J-4))/7.0))+50.0
      B(KH,55)=(A(I,J)-((A(I,J)+A(I-1,J-1)+A(I+1,J+1)+A(I-2,J-2)+A(I+2,J
     1+2)+A(I-3,J-3)+A(I-4,J-4))/7.0))+50.0
      B(KH,56)=(A(I,J)-((A(I,J)+A(I-1,J)+A(I+1,J)+A(I-2,J)+A(I+2,J)+A(I-
     13,J)+A(I-4,J))/7.0))+50.0
      B(KH,57)=(A(I,J)-((A(I,J)+A(I-1,J+1)+A(I+1,J-1)+A(I+2,J-2)+A(I+3,J
     1-3))/5.0))+50.0
      B(KH,58)=(A(I,J)-((A(I,J)+A(I,J-1)+A(I,J+1)+A(I,J-2)+A(I,J-3))/5.0
     1))+50.0
      B(KH,59)=(A(I,J)-((A(I,J)+A(I-1,J-1)+A(I+1,J+1)+A(I-2,J-2)+A(I-3,J
     1-3))/5.0))+50.0
      B(KH,60)=(A(I,J)-((A(I,J)+A(I-1,J)+A(I+1,J)+A(I-2,J)+A(I-3,J))/5.0
     1))+50.0
  300 CONTINUE
      WRITE(6,33)(HEAD(JJ),JJ=1,10)
      DO 203 I=1,M
      DO 203 J=1,N
  203 A(I,J)=B(I,J)
      LA=0
      CALL RPRINT(A,N,M,LA)
      XM=M
      DO 234 J=1,N
      DO 204 JN=1,10
  204 AA(JN)=0.0
      JN=0
      ISWT=0
      ISI=0
      ISKIP=0
  205 XXX=0.0
      DO 206 I=1,M
  206 XXX=XXX+A(I,J)
      DO 207 K=1,15
  207 Q(J,K)=0.0
      Q(J,1)=XXX/XM
      DO 208 I=1,M
      Q(J,2)=Q(J,2)+((A(I,J)-Q(J,1))**2)
      Q(J,3)=Q(J,3)+((A(I,J)-Q(J,1))**3)
  208 Q(J,4)=Q(J,4)+((A(I,J)-Q(J,1))**4)
      Q(J,5)=Q(J,2)/(XM-1.0)
      Q(J,6)=(XM*Q(J,3))/((XM-1.0)*(XM-2.0))
  209 Q(J,7)=(XM*(((XM+1.0)*Q(J,4))-((3.0*(XM-1.0))*((Q(J,2)**2)/XM))))/
     1((XM-1.0)*(XM-2.0)*(XM-3.0))
      Q(J,8)=(Q(J,6))/(Q(J,5)*SQRT(Q(J,5)))
      Q(J,9)=Q(J,7)/(Q(J,5)**2)
      Q(J,10)=((6.0*XM)*(XM-1.0))/((XM-2.0)*(XM+1.0)*(XM+3.0))
      Q(J,11)=((24.0*XM)*((XM-1.0)**2))/((XM-3.0)*(XM-2.0)*(XM+3.0)*(XM+
     15.0))
      Q(J,12)=SQRT(Q(J,10))
      Q(J,13)=SQRT(Q(J,11))
      Q(J,14)=Q(J,8)/Q(J,12)
```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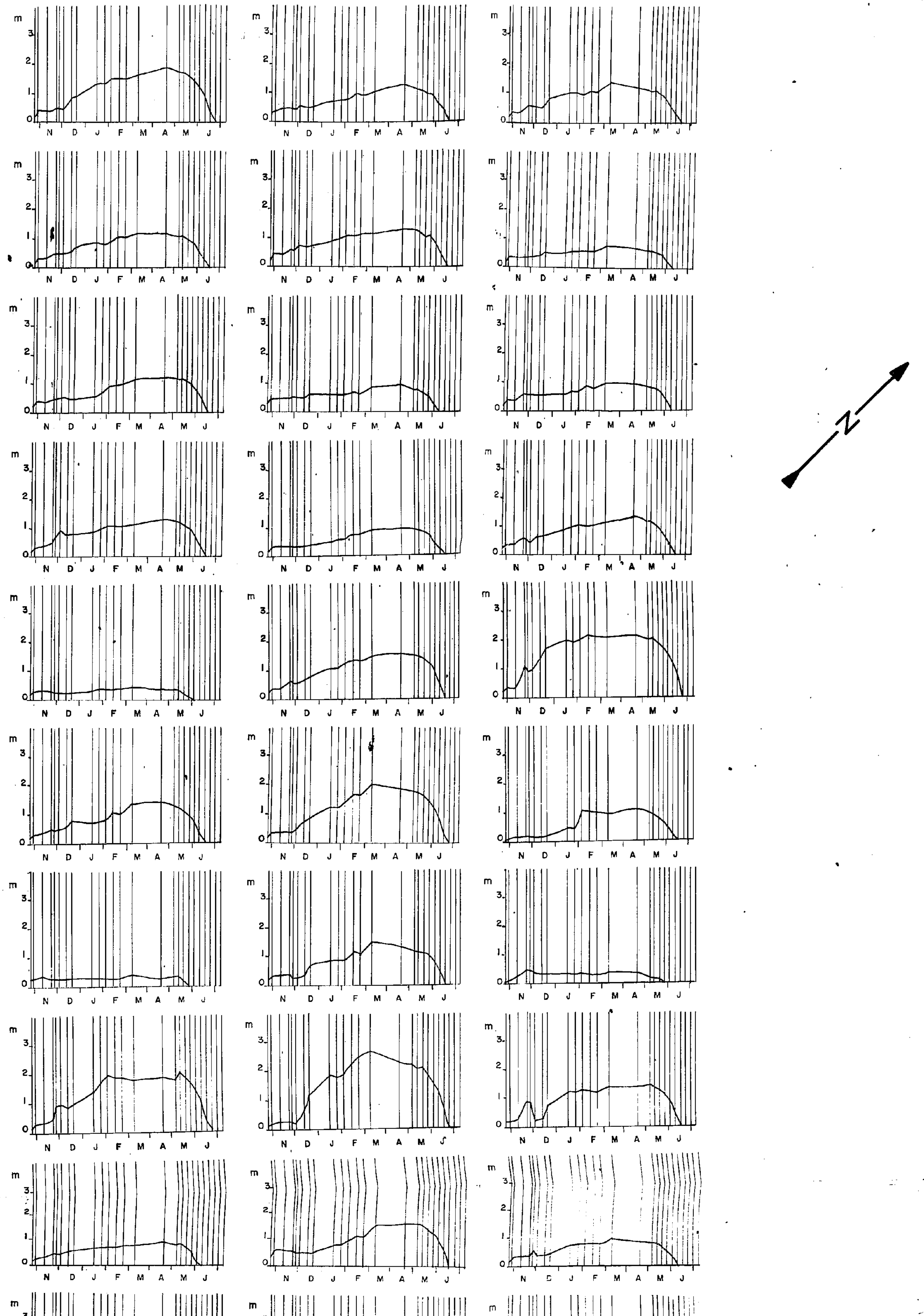
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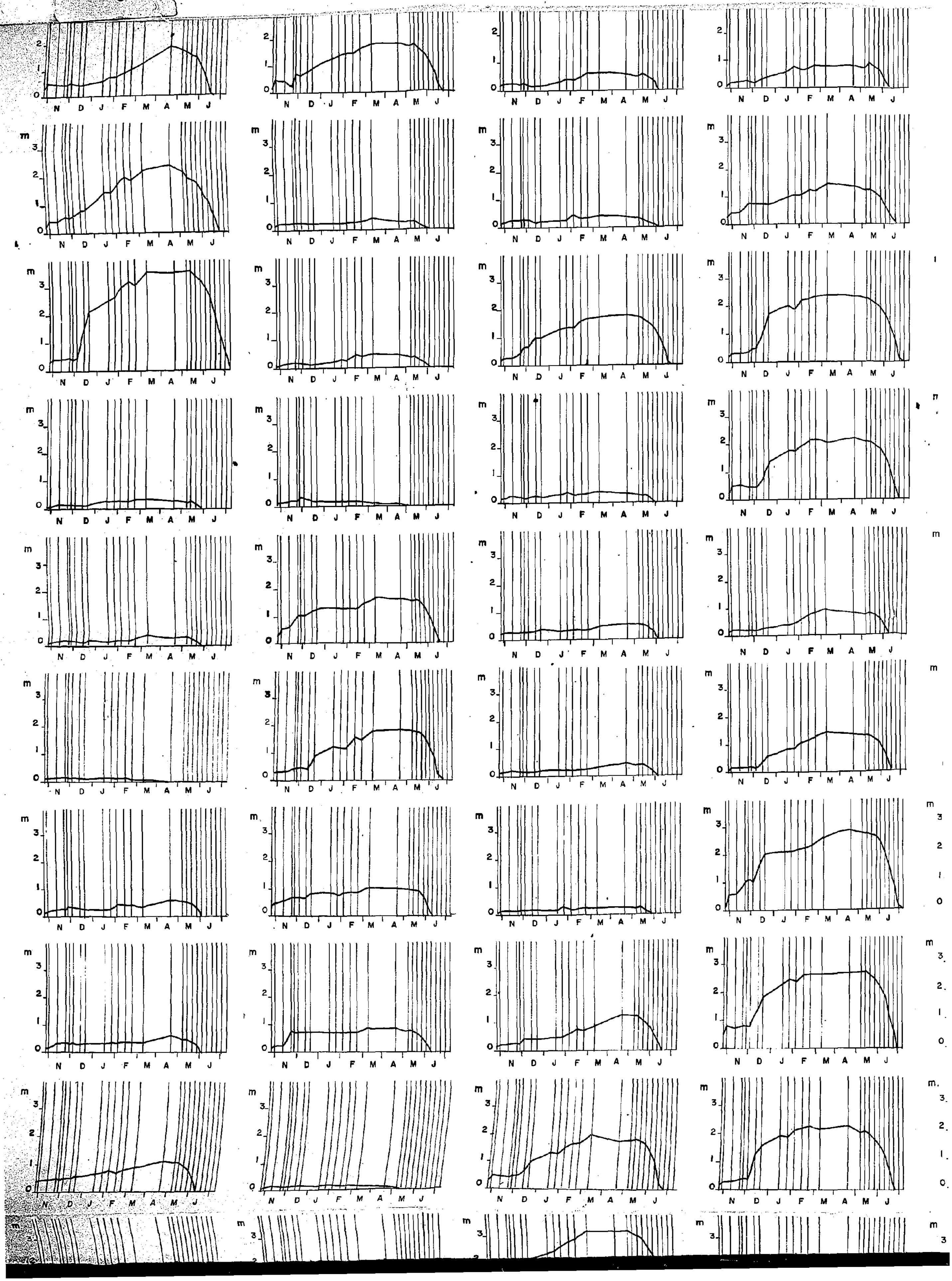

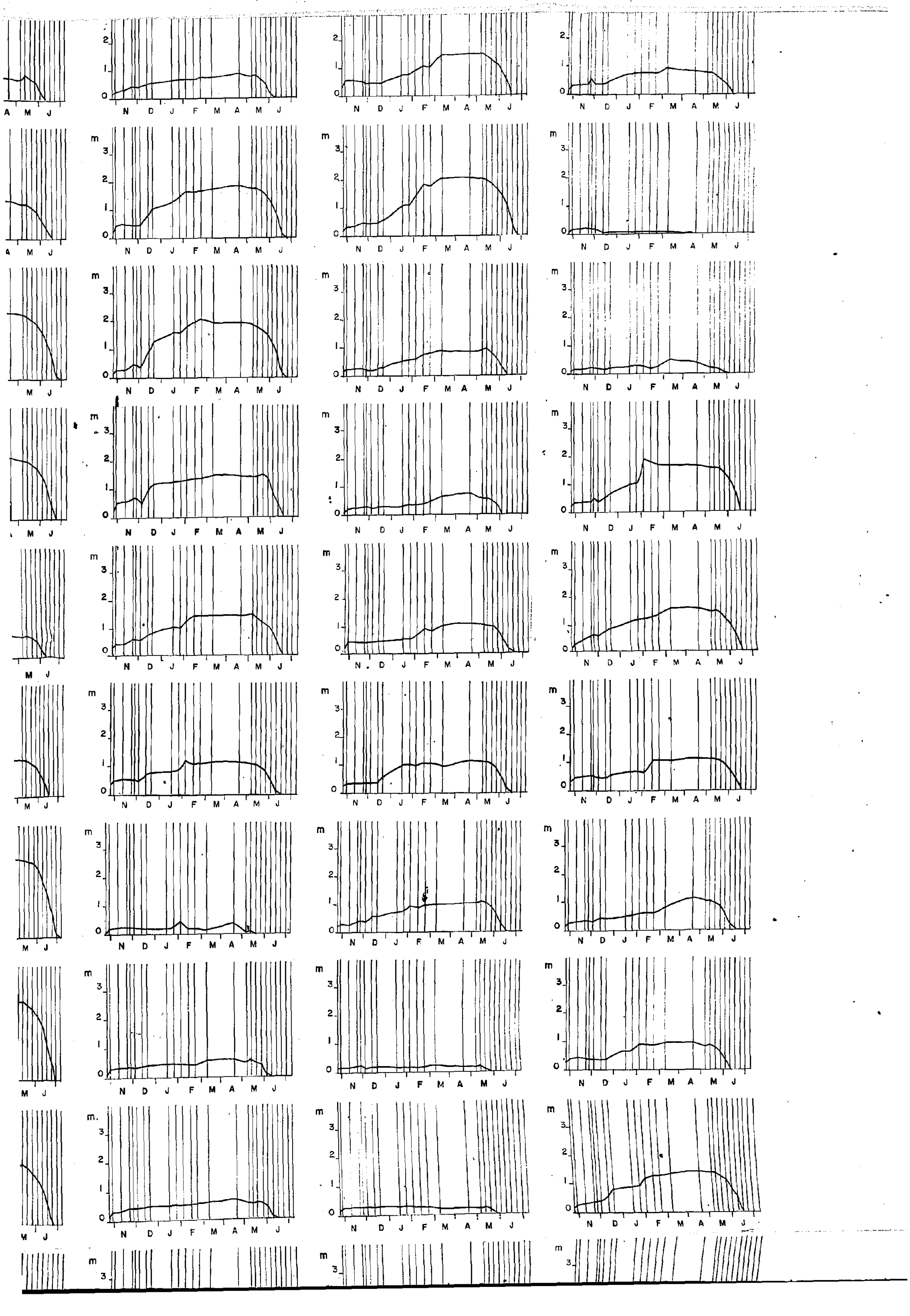

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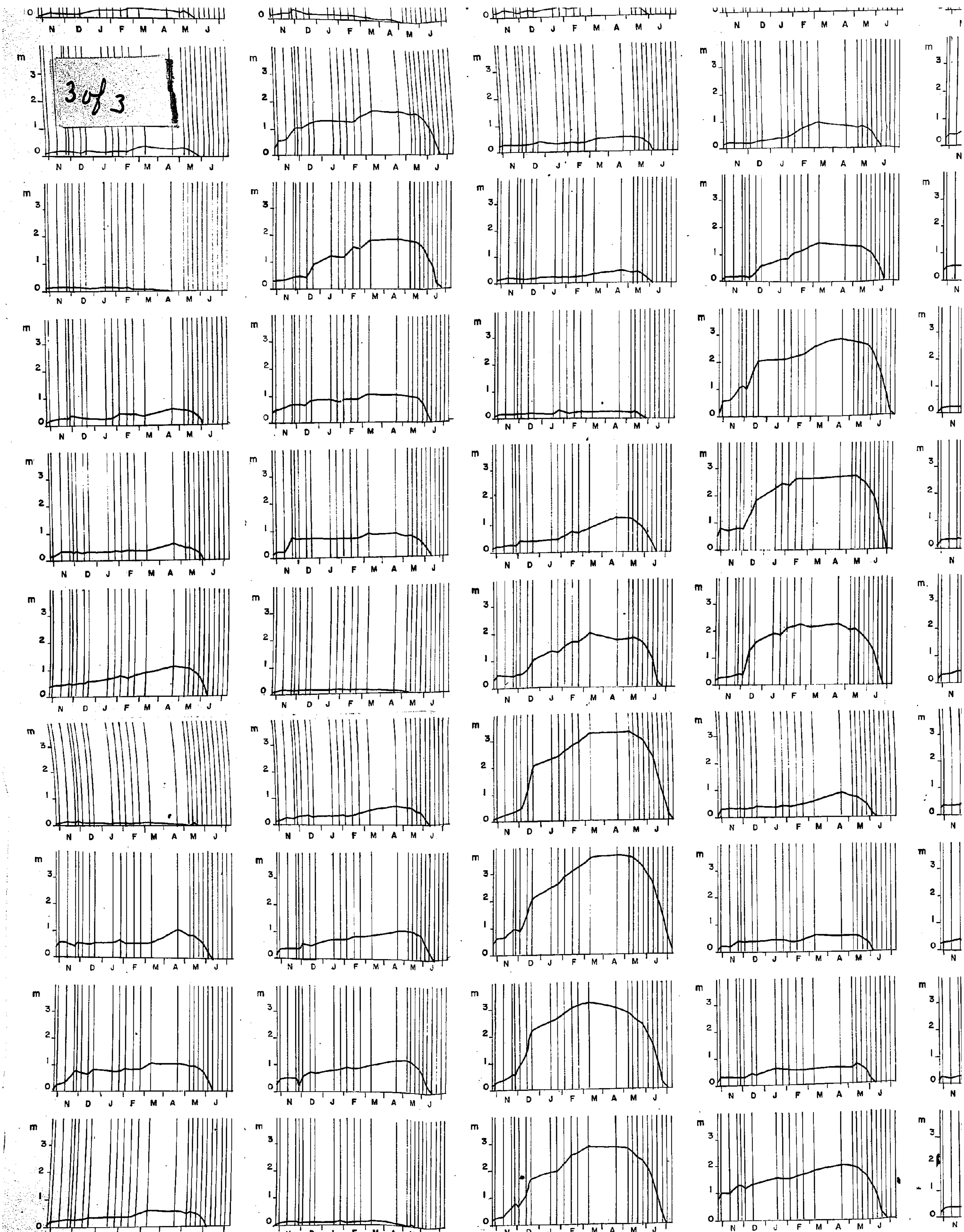
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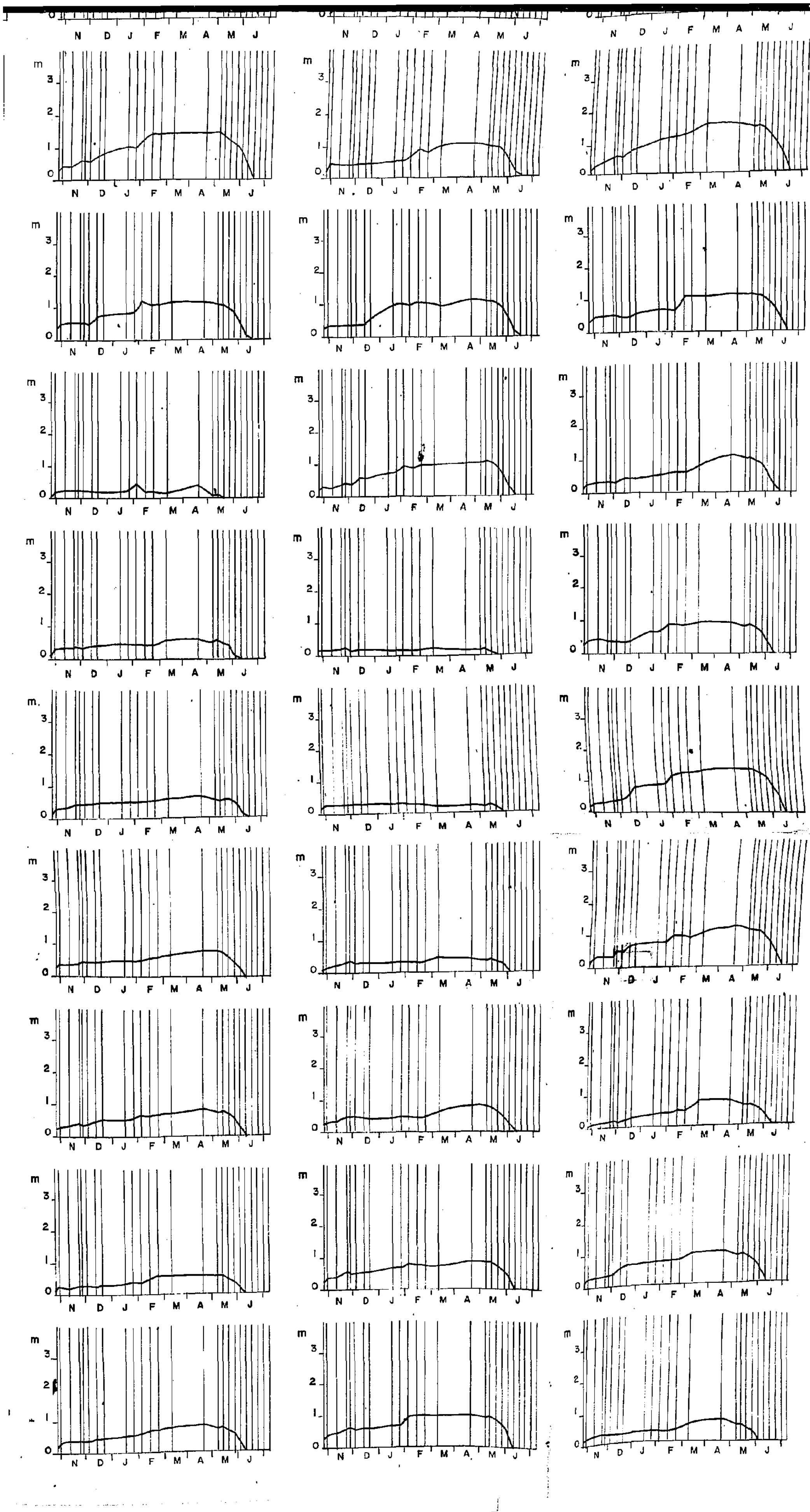

Figure 10. Depth increase at 147 snow stakes.