Simulation of cirque glacier distribution

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Abstract

SIMULATION OF CIRQUE GLACIER DISTRIBUTION

Jasbir S. Gill

This study is concerned with an examination of the manner in which local topographical and climatological factors interact to control glacier balance. The approach used is one of developing a deterministic computer simulation model which, with given inputs of local topography and monthly climate, locates sites of positive mass balance and cirque glacier generation in mountainous ranges of temperate areas.

Topographical parameters (altitude, gradient, aspect, profile and plan convexity) are calculated, by means of a terrain-analysis program, at each grid intersection of an altitude matrix. The interaction of local topography with regional climate defines the variation of local climatic variables (air temperature, snowfall, wind speed and direction).

Separate modelling of the accumulation and ablation processes leads to the simulation of glacier mass balance at all points of the altitude matrix. Emphasis is placed on modelling, as far as possible, the actual processes involved in local glacier development. Visual displays of the simulated results, by means of line-printer maps, can be specified at any time during the model development.

Model validation is attempted by comparing the results of mass balance simulation with observations of snowpatch and local glacier distribution in northern Iceland. The success of this modelling approach reveals the critical role of the topographic parameters in controlling spatial variations of local glacier balance.

Computer simulation experiments enable the effects of changing climatic inputs on glacier balance to be examined. They reveal the glaciological sensitivity of northern Iceland to small changes in climate; incipient cirque glaciers being generated with a 2°C. drop in air temperature. The location of these glacier generation sites in concavities adjacent to plateau remnants, permitting snow drift accumulation, is shown to be significant. Such experiments are useful for the verification of palaeoclimatic interpretations regarding local glacier distributions in the past and for the prediction of future glacierisation patterns with postulated climatic trends.

SIMULATION OF CIRQUE GLACIER DISTRIBUTION

by

Jasbir S. Gill

A thesis submitted in partial fulfilment of
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to Thana and my mother

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Eggert Ólafsson (Icelandic naturalist 1726-1768)

on the formation of new glaciers

- being one of the earliest accounts of a climatological origin for a cirque glacier.

> "When we passed here [Mófell on the northern side of Skardsheidi] on August 6th [1752] , we noticed high up on the mountain a fairly large patch, which looked like glacier ice. The owner of the nearest farm, Mófellsstadir, in reply to our question whether the ice on Mófell did not melt in summer, not only answered no, but added that when he was a boy he never saw any ice there at all; but that when he after many years absence returned some years ago he had noticed a beginning of snow accumulation, and that gradually less of it melted away in the summer. The place is facing N.W., and the ice already shows cracks of a green colour due to the refraction usual on thick glaciers. This indicates that the ice may increase and new glaciers form, even on moderately high mountains in this neighbourhood, provided the periodical cold winds persist year after year".

- translated and quoted by

Thorarinsson (1960), from Vice-Lavmand Eggert Ólafssens og Landphysici Bjarne Povelsens Reise igiennem Island. Soröe 1772. (Travels in Iceland 1752-1757 by Eggert Ólafsson and Bjarni Pálsson).

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No material contained in this thesis has previously been submitted for a degree in this or any other university.

CHAPTER I

INTRODUCTION

1.1 The Glaciological Problem

Glaciology can be defined as the scientific study of the origin, distribution and physics of snow and ice (Marcus, 1964). Within this broad field the relationship between glaciers and climate has occupied the attention of research workers for over two centuries[1].

The importance of climatic control on glaciers is well established (e.g. Ahlmann, 1948a; Hoinkes, 1964; Meier, 1965). Sharp (1960, p.20), for example, notes that "glaciers are utterly dependent upon elements of the climatic environment for birth and for sustaining life". A recognition of this dependence has led to a considerable amount of literature being devoted to the development of relations between glacier fluctuations and variations in both present and past climates (reviewed in chap. II).

The great number of studies concerning glacier-climatic relationships may give the impression that these are quite well understood. However, the complexity and a lack of detailed understanding of the relationship between glacier behaviour and climatic events has been pointed out by several workers in this field (e.g. Meier, 1965; Paterson, 1969; Andrews et al., 1970). For example, the response of the glacier terminus to climatic changes is complicated by the

[1]The first statement of a correlation between glaciers and climatic variations being made by J. Walcher in 1773, according to Hoinkes (1964, p.391). For an earlier account of a climatological origin for a cirque glacier by an Icelandic naturalist, see the front of this thesis (page v).

dynamics of ice flow and assumptions of synchronisity between glacier responses and climatic events are clearly invalid (Meier, 1965; Løken, 1972; Kuhn, 1978).

In practice, the response of glaciers to changes in climate is controlled by numerous inter-related glacial and meteorological factors, such as the dimensions of the glacier, thermal and other physical characteristics of the ice, and the amplitude and period of the climatic fluctuations. These factors lead to individual glaciers responding with different time lags to the same climatic change. Differences in the individual response-characteristics of glaciers may result in opposite tendencies of movement, with some retreating and others advancing, being observed in adjoining glaciers (see examples provided by Paterson, 1969, p.233 and Kuhn, 1978, p.444).

The complexity of glacier-climate relationships is further revealed by analysing the factors responsible for local glacier distribution (discussed in chap. III). With small glaciers, for example, the local topography (often neglected in glacier-climate studies), orientation and exposure, relation to prevailing winds, and local climatic conditions all result in local differences in the process of accumulation and ablation of snow and ice, giving rise to an "endless variety in the conditions of development of glaciation" (Shumskiy, 1950, p.17).

With the realisation that the relation of glacier snout fluctuations to local or regional climate cannot reveal much about the intricate glacier-climate relationships, attention has been focused on the mass balance. The latter, being a "fundamental hydrometeorological variable in the glacier theory" (Lliboutry, 1971, p.88), should provide a much better basis for the evaluation of glacier-climate relationships. Furthermore, the mass balance forms a critical link between glaciers and their interaction with the environment.

This interaction can be visualised as a chain of distinct processes as shown in Fig. 1.1. These generalised series of steps serve to identify three major problem areas in the relation of glaciers and the environment.

The net mass balance forms a vital link between (a) the 'glaciometeorological' problem, concerned with the way in which factors of climate and topography control the mass and energy exchange processes at the glacier's surface and, through these, produce temporal and spatial variations in the mass balance; and (b) the 'glaciodynamical' problem, concerned with the manner in which the glacier adjusts to changes in its mass balance, for example by causing changes in the thickness, rate of flow and the dimensions of the glacier (Meier and Tangborn, 1965; Paterson, 1969; Lliboutry, 1971). A third area of research, concerning the relation between the glacier's dynamical response and its effects on the underlying land surface, through the processes of glacier erosion and deposition can be distinguished as the 'glaciogeomorphological' problem.

All the three problem areas, defined above, include consideration of some aspect of the glacier-climate relationship. Within the glacio-geomorphological problem, though glacial geomorphologists have for long used reconstructions of past glacier distributions to yield information about palaeoclimates (sec. 2.3.5), current interest is focused on understanding the development of landscapes in terms of the glacial processes involved, and in relating the latter to relevant environmental factors, particularly climate (Sugden, 1977a). Although some theoretical advances have been made regarding our understanding of glacial processes (e.g. Boulton, 1974; 1975), we are still a long way from determining landform - process interactions in much detail (International Glaciological Society, 1979; 1981).

Our understanding of the glaciodynamical problem, however, is relatively well advanced, with numerous theoretical contributions (sec. 2.3.3). While these advances enable us to solve equations for the dynamical response of glaciers to changes in mass balance, the question of the detailed effect of climatic changes on mass balance variations, the basic glaciometeorological problem, remains to be fully understood.

Although numerous determinations of glacier mass balances and their relation to local and regional climate have been made (see sec. 2.3.3), the details of glacier-climate relations have not been revealed since any given mass balance can be produced by an infinite combination of accumulation and ablation values. For a clearer understanding, the processes involved in these mass balance components need to be analysed separately and related to climatic variables. Furthermore, an understanding of the topographical factors controlling these mass balance processes and, in turn, local glacier distribution remains a fundamental task for glacial geomorphologists in their attempts at studying both past and present conditions of glacierisation.

This study examines the manner in which local topographic and climatic factors interact to control local glacier balance. Prior to defining the aims and scope of this work, consider some of the main difficulties that have limited the advancement of our understanding of glacier-climate relationships.

1.2 Difficulties Hindering Glacial Research

The variable nature of the different processes affecting the glacier balance and the existence of complex interrelations between them have impeded the understanding of glaciometeorological relations. For example, many of the processes involved in the accumulation of snow, such as snow drifting and avalanching, are quantitatively only poorly understood. Uncertainty also exists about a number of glaciological processes, such as ice flow and the effect of the glacier, itself, on altering the climatic elements in its immediate surroundings.

Many of these problems in understanding the interaction of glaciers and the environment are caused simply by a lack of adequate data. Direct observations of the glacier/rock-bed interface are relatively rare since they involve long term and expensive tunnelling or drilling operations. Most of the direct knowledge of glaciological processes is, therefore, based upon a limited number of sporadic and random empirical observations, making it difficult to guide and test theoretical considerations.

Surface observations of mass and energy balances are usually only carried out for short periods (during the ablation season) at a limited number of points on a few selected glaciers. The sparse and unrepresentative network of weather stations in glaciated mountains areas makes the climatic data unsuitable for detailed glacier-climate analysis. These data restrictions make the wider application of any conclusions highly questionable.

A related problem to the inadequacy of data is the difficulty of defining the appropriate scale at which to study the glacier-climate interactions. In particular the lack of meso-scale glacier-climate studies, i.e. between the micro-scale (heat budget) and the macro-scale (glacier variations related to long-term regional climatic averages) studies, has restricted an understanding of local glacier balance controls.

1.3 Aim, Approach and Scope of Present Study

Given the need to understand the glaciometeorological problem (sec. 1.1), and bearing in mind the difficulties facing the researcher in this field (sec. 1.2), this study is aimed at the prediction of local glacier distribution from considerations of topographical and climatological factors. This involves an identification of the various factors affecting glacier balance and the formulation of quantitative relationships between them, leading to a synthesis of topographic - climatic - glacial relations.

The approach adopted is one of developing a deterministic simulation model which, with given inputs of local topography and climate, will locate areas of positive mass balance and local glacier generation. In order to make the model generally applicable to temperate mountainous regions, an attempt is made, as far as possible, to model the actual physical processes controlling the spatial and temporal variations in mass balance.

For a clearer understanding of the controls on the mass balance, values of accumulation and ablation are obtained separately and the net mass balance calculated as the difference between them. The use of climatic indices to estimate accumulation, ablation and mass balance, without a proper consideration of the processes governing these measures, is not satisfactory since such indices can only be applicable to localised areas and they do not reveal the underlying causes for temporal and spatial variations in the mass balance.

Where lack of quantitative information makes the physical modelling approach difficult, as in modelling the variation of climatic elements, wind drifting and avalanching, it is necessary to resort to rather indirect and statistical considerations. Given the difficulties of understanding glacier systems (sec. 1.2) we cannot, at present, hope to physically model the whole process of glacier development; however, we can usefully make an attempt at the numerical modelling of the glacier balance, even if we are forced to use relatively crude simplifications (Glazyrin, 1975).

An important feature of the approach used in this study to simulate local glacier development is the emphasis on measures of local topography, e.g. altitude, slope, aspect, profile and plan convexity, based on the hypothesis that variations in local topography essentially determine the local climate and consequently the glacier balance. Thus, if the controlling links between topography, local climate and glacier balance can be defined then a model of local glacier distribution can be arrived at.

It is hoped that such a modelling approach will lead to a better understanding of the processes controlling glacier balnace. By enabling the effects of climatic and topographic variations on glacier balance to be analysed the model will enable the prediction of glacier balance, from easily available input parameters. Furthermore, the use of computer simulation provides a means of experimenting with different climatic inputs and observing the relative importance and particular effects on glacier balance, of changes in individual climatic variables.

1.4 Limitations and Simplifications

This study is restricted to a consideration of local mountain glaciers, especially cirque glaciers. A number of other associated small glaciers, differing to varying extent in form and setting from the typical cirque glacier, have been distinguished; e.g. niche, crater, hanging, slope, wall, ice-apron, and glacieret (for a description of these refer to the report on Perennial Ice and Snow Masses by UNESCO/IASH, 1970a). In this study the term 'cirque glacier' is used in its broadest sense, including all variety of small mountain glaciers and perennial ice masses.

The choice of studying local cirque glaciers was determined by the fact that, being indicators of marginal conditions of glacierisation, cirque glaciers are most constrained by conditions of local topography and climate. Furthermore, having much shorter response times compared with other larger ice-masses (Table 1.1), they respond sensitively to relatively minor changes in climate. Thus, any climatic warming will first affect cirque glaciers. Conversely, cirques and similar topographic hollows, being favoured sites for the accumulation of snow and ice, will be the first to be occupied by glaciers in the event of a climatic deterioration. It should be noted that the concern of this study is with the factors leading to cirque occupation by ice, rather than with the development of cirques, themselves.

This study is further limited to a consideration of 'temperate' glaciers, as opposed to those of the 'polar' and 'sub-polar' types (after Ahlmann, 1935). A temperate glacier is defined as one that is at the pressure melting point throughout except for a thin surface layer subject to seasonal cooling. Limitation of the study to glaciers of this type is due primarily to the fact that the bulk of current knowledge of cirque glacier processes is drawn from the observations of temperate glaciers.

The assumption of the whole glacier being at pressure melting point has been questioned (e.g. Paterson, 1972; Harrison, 1972), since in detail a glacier usually consists of zones having differing temperature and hydrological characteristics (Müller, 1962). However, the concept of temperate glaciers is useful in allowing a number of simplifying assumptions to be made about the behaviour of ice and its energy exchange with the atmosphere: this is discussed further in chap. VIII.

Some further general simplifications made in this study concern the effect of glaciers in altering the local climate and topography. Given the small size of the glaciers being considered and the relatively short temporal periods over which the glciers are being modelled, we can probably neglect the influence of glaciers on the climate and of changes in topography induced by glacier occupation. A model of the complete glacier system (Fig. 1.1) would, of course, have to take these feedback effects into account. However, in this study, the prime aim is to analyse the spatial variations in the mass balance of snow and ice, and to locate sites for the generation of local glaciers.

Since small glacierised basins will normally be situated above the treeline, we can also disregard the effect of vegetation or forest cover in affecting snow and ice distribution. Other more detailed simplifications and assumptions will be considered in the sections appropriate to the modelling of the glacier balance processes.

1.5 Layout of Succeeding Chapters

Although previous work on the modelling of the local glacier balance is relatively limited, a review of the literature reveals numerous ways in which the relationships between glaciers and their environment have been studied. A great variety of studies have been concerned with examining some aspect of the glacier-climate relationship or with analysing the factors controlling the distribution of snow and ice. These studies provide useful qualitative and quantitative information for the modelling of local glacier distribution and are reviewed in chapter II.

The factors relevant for the determination of cirque glacier balance are identified in chapter III. This is followed by an outline of the procedure for modelling cirque glacier balance (chap. IV). A description of the area in which the model was applied and developed (in northern Iceland) is also included in chapter IV.

Chapters V and VI, following the outline of the simulation model, are concerned with a discussion of the manner in which the input variables of topography (chap. V) and climate (chap. VI) were obtained. Chapter V includes a discussion on the use of a 'terrain-analaysis program' to derive local topographic measures from an altitude matrix. The relationships between these measures of local topography and the regional climate to derive the variation of local climate are discussed in chapter VI.

Modelling of the various processes of the glacier mass balance is considered in chapter VII (snow accumulation and densification) and chapter VIII (ablation). In each of these chapters the importance of the particular aspect being discussed to the glacier balance is shown, followed by an explanation of the modelling procedure; illustrated with examples from an area in N.W. Iceland.

Chapter IX is concerned with the application of the mass balance model to simulate snow cover and local glacier distribution. Model validity is tested by comparing the simulated mass balance results with observations of snow cover and local glacier distribution in two sample Icelandic study areas. Simulation experiments, with varying climatic inputs, are also carried out to develop an understanding of glacier-climate relationships. Finally, chapter X provides the conclusion and suggestions for further study.

TABLE 1.1

Topographic Constraints and Response Time Scales For Different Glacier Types

(modified after Sugden and John, 1976)

Increasing topographic constraint

Glacier type	Approximate Response time scales for climatic change (years)
Ice sheets	100,000 - 10,000
Ice caps	
Ice fields	10,000 - 1,000
Outlet glaciers (draining plateau ice caps)	
Valley glaciers (originating in cirques)	1,000 - 100
Cirque and slope glaciers	100 - 10
Perennial snowfields	10 - 1

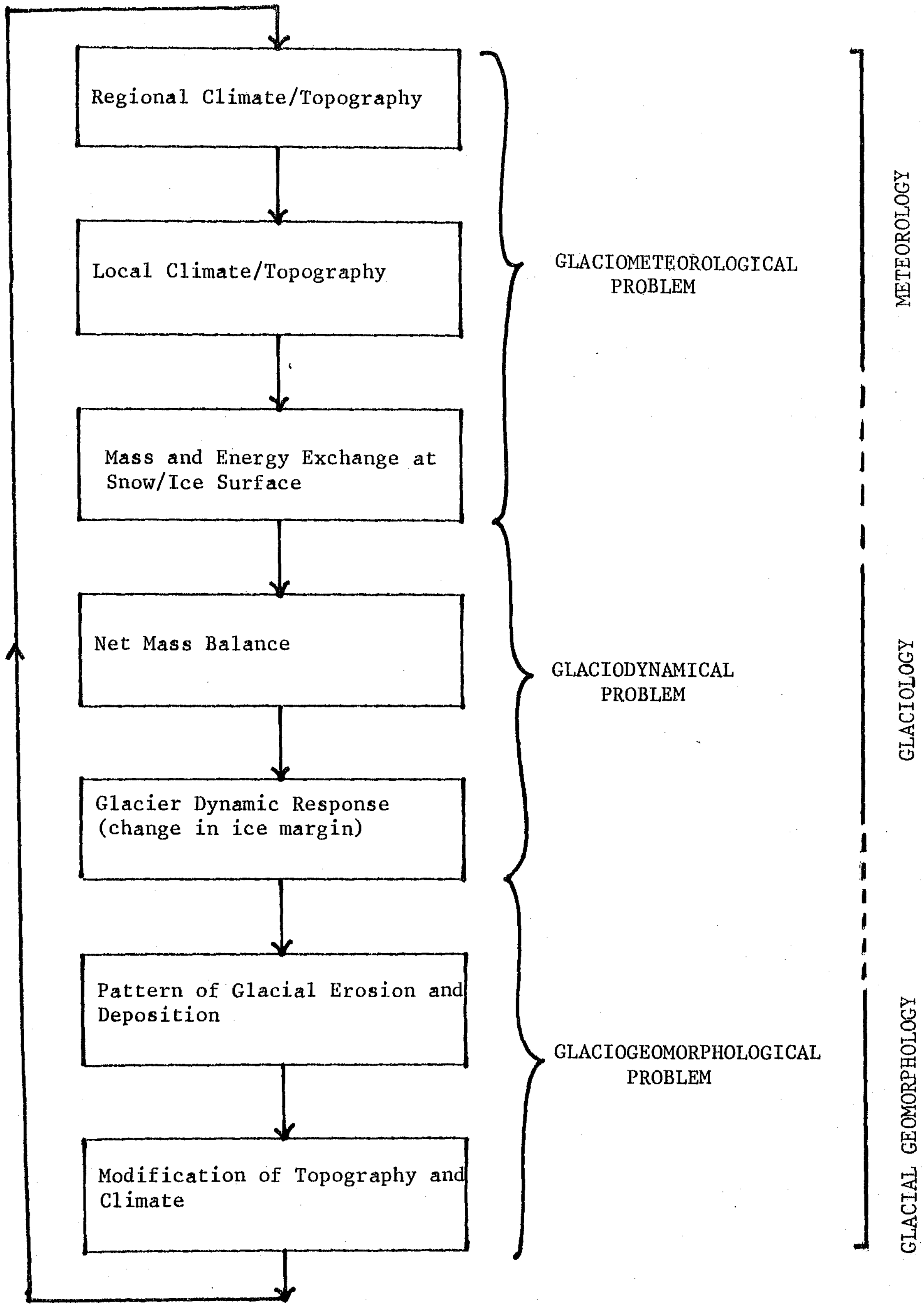

FIGURE 1.1

Major Steps Involved in a Complete Study of the Interactions Between Glaciers and Their Environment

CHAPTER II

REVIEW OF STUDIES RELATING TO GLACIER-CLIMATE RELATIONSHIPS

An awareness of the importance of glaciers in both academic and applied fields has led to a considerable amount of literature being devoted to the development of relationships between glaciers and climatic or other controlling variables (e.g. review of glacier meteorology studies by Hoinkes, 1964). A familiarity with the many ways in which these relationships have been studied and the conclusions arrived at, is important from the point of yielding information regarding the relative significance of different factors in controlling glacier balance. This information is a necessary prerequisite for the development of a glacier balance simulation model, since a review of such studies, not only allows us to assess the state of our understanding of glacier-climate relations, but further, provides guidelines for the incorporation of relevant variables and relations in the modelling of local glacier distribution.

The reasons for glacier studies achieving relevance and importance are considered first. This is followed by a brief outline of the development of glacier-climate studies, providing a historical perspective to contemporary glacier studies.

2.1 Importance of Glacier Studies

Glacier ice at present covers about 10% of the land surface. Though the bulk of this is concentrated in the two polar ice-sheets smaller ice masses, in the form of cirque and valley glaciers, are an integral part of many glaciated alpine landscapes. Furthermore, during the maximum Quaternary extent, glaciers and ice-sheets covered over 30% of the earth's land-area.

Given the great extent of ice distribution, both at present and in the relatively recent geological past, glaciers assume considerable importance in moulding a substantial portion of the earth's relief. They have also been responsible indirectly for changes in local and regional climates, eustatic variations of sea-level and isostatic depression and recovery in those areas affected by the accumulation and removal of ice. Thus, a study of the present-day characteristics of glaciers, together with their interrelationships with factors of topography and climate, is essential for an understanding of the geomorphology and climatology over a large portion of the earth's surface.

Further impetus for glaciological research has come from the effects of glacier fluctuations on human activities. We have only to read early accounts for the destruction of farmlands and settlements by glacier advances in Iceland and the Alps to appreciate the close interrelation between glaciers and human occupation (e.g. Thorarinsson, 1956). In view of the considerable debate over the impending new Ice Age, such historical accounts assume contemporary importance, promoting attempts to determine the magnitude of climatic change necessary for the initiation of glacierisation (see sec. 2.3.5).

A fact of great economic importance is that about 75% of the world's fresh water is stored as glacier ice. With the ever increasing demands for fresh water supplies, glaciers are of special interest hydrologically, especially in areas such as the American West, Central Asian ranges, Caucasus Mountains, the Alps and in Norway, where communities rely largely on the snowmelt from small temperate mountain glaciers and snowpatches for their water supply.

In these areas snow and ice hydrology has achieved great significance, since accurate estimates of the volume of water released by snow and ice-covered areas are needed for such purposes as water supply and flood forecasting, hydro-electricity, design of hydrologic and hydraulic structures, irrigation, recreation and domestic supplies. With the need to conserve and regulate water supply from snow and ice reserves, various watershed management schemes have been developed (see sec. 2.3.6).

The need for intensive studies of the processes controlling the accumulation and melting of snow and ice for the proper planning and management of these resources has led to considerable advances in our understanding about the nature of temperate snowpacks and small glaciers. With the recent proposals of towing icebergs from polar areas to more temperate latitudes (e.g. Husseiny, 1978; International Glaciological Society, 1980b), the use of glacier-ice as a fresh water resource has assumed an even greater importance.

It will be seen that a proper analysis of any of the above applications would involve the combination of a number of disciplines, e.g. meteorology, geology, hydrology, physics and mathematics. This has led to the emergence of modern glaciology as an interdisciplinary science, incorporating the efforts and interests of many specialists from other disciplines. Within this broad framework the question of 'glacier-climate relationships' has been of central concern to most glacier research workers.

2.2 Historical Perspective

Following the first descriptions of glaciers in the middle of the 16th Century, it was not until the 18th Century that glaciers attracted scientific attention in the Alps, Iceland and Scandinavia. In the latter part of the 19th Century systematic measurements of glacier variations were initiated through the setting up of the Commission International des Glaciers in 1894.

The C.I.G. resolved to "encourage and to collect observations on glaciers all over the world, with the special object in view of discovering a relation between the variations of glaciers and of meteorological phenomena" (quoted by Baird, 1958 p.253). Numerous measurements of glacier variations and other glacier characteristics were conducted under the auspices of the C.I.G. and subsequent international bodies[1].

The detailed study of glacier-climate relationships, however, stems from the classic work of Hans W. Ahlmann and his co-workers in the middle 1930's. Their work summarised in two reports, (Ahlmann, 1948a; 1953), was to lay the foundations of modern glacier research. Beginning in 1919 Ahlmann studied in detail some glaciers around the North Atlantic coasts in Scandinavia, Iceland, Spitsbergen and N.E. Greenland. He formulated various methods for studying the glacier regime which were later employed by many other glacier researchers, such as Wallén, Hoinkes, Orvig and Hubley. With the realisation of a need for an interdisciplinary approach together with international co-operation for worthwhile advances in glacier-climate understanding, the work of individual researchers has been superseded by international research programmes.

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[1]The C.I.G. was dissolved in 1927 and its work is at present conducted by the Commission on Snow and Ice of the International Association of Scientific Hydrology set up in 1948.

Many advances in glacier studies were made during the International Geophysical Year (I.G.Y., 1957-1958) and more recently the International Hydrological Decade (I.H.D., 1965-1974). The latter programme was launched by U.N.E.S.C.O. to "promote international co-operation in research and studies and the training of specialists and techniques in scientific hydrology" with the purpose of enabling "all countries to make a fuller assessment of their water resources and a more rational use of them as man's demands for water constantly increase in face of developments in population, industry and agriculture" (UNESCO/IASH, 1970a p.5).

Due to the obvious hydrologic importance of snow and ice (sec. 2.1), a major concern of the I.H.D. programme has been with studies concerning glaciers and other major snow and ice bodies. In particular, research has been aimed at three major areas - (a) world inventories of perennial and annual ice and snow masses (sec. 2.3.1); (b) measurements of glacier variations on a world-wide basis (sec. 2.3.2); and (c) combined water,ice and heat-balance measurements at selected representative glacier basins (sec. 2.3.3).

In addition to the advances in glacier-climate relationships made by the various I.H.D. programmes, a number of other long-term projects have yielded important information; e.g. the multi-disciplinary and interdisciplinary studies of alpine environments conducted in the Icefield Ranges of the St. Elias Mountains since 1967, (e.g. Wood, 1963; Bushnell and Ragle, eds., 1969-1972; Bushnell and Marcus, eds., 1974); and the allied High Mountain Environment Project (1966-1970) concerned, in part, with climatological and glaciological research at alpine sites in the Wrangell Mountains in Alaska and on Mt. Logan in Yukon (Marcus, 1974a). The latter project was supported by the U.S. Army Research Office and the University of Michigan.

Baffin Island is another area which has been the subject of concentrated study by J.T. Andrews and his colleagues from the Institute of Arctic and Alpine Research (I.N.S.T.A.A.R.) and the University of Colorado. These workers have focussed their main attention on the past and present glaciological responses to climate (e.g. Andrews et al. 1970; 1972).

The nature of glacier-climate studies carried out during the above mentioned projects and other subsequent research programmes is now reviewed.

2.3 Studies Relating to Glacier-Climate Relationships

(refer to Table 2.1)

2.3.1 Distribution of Snow and Ice Masses

One of the first important considerations in a study of the factors controlling glacier development is a proper understanding of the distribution of present-day perennial snow and ice masses. This is being achieved in a number of ways:

(i) Snow and Ice Surveys. Accurate estimates of the mass of snow cover in mountainous regions, through the conduction of snow surveys, are important for providing seasonal streamflow forecasts. In the western United States, for example, where a major portion of the stream water is derived from mountain snowpacks, snow surveys have been performed on a systematic basis since the early 1900's. The use and value of snow surveys and early developments in snow surveying have been discussed by Church (1933), one of the pioneers of snow surveying. Snow surveys represent the first systematic snow accumulation studies and provide useful evaluation of such factors as wind drifting, topography, forest cover, temperature variations, and evaporation on affecting the distribution of snow cover.

Following the early work of Church, snow surveys have become a regular activity in many of the snow-covered regions of the world, providing valuable information for streamflow forecasting. Conventional field and aerial surveys are being replaced by mapping through satellite photography (Barnes and Bowley, 1968) which overcomes problems of accessibility (found in field surveys) and poor flying weather (restricting aerial surveys).

(ii) Glacier inventories and mapping. An important requirement of the I.H.D. programme has been the recording of all permanent ice and snow masses according to the recommendations provided in the U.N.E.S.C.O. guide, "Perennial Ice and Snow Masses", (UNESCO/IASH, 1970a).
As part of the I.H.D. programme and subsequent work numerous inventories of glaciers have been published for regions such as the Axel Heiberg Island (Ommanney, 1969); Southern Norway (Østrem and Zeigler, 1969); North Cascades (Post et al., 1971); Northern Scandinavia (Østrem et al., 1973); Western Alps (Vivian, 1975); Switzerland (Müller et al., 1976); and various parts of the Soviet Union (Inventory of glaciers of the U.S.S.R.), the work being carried out by the Gidrometsluzhby SSSR (the Hydrometeorological Service of the U.S.S.R.) and other research bodies (Grosval'd and Kotlyakov, 1969).

A review of the I.H.D. inventory programmes in various countries is provided by Ommanney (1974). These inventories include information referring to the location, aspect, size, variations, level of equilibrium line and other morphometric details concerning the ice and snow masses. This information is very useful for the analysis of glacier distribution with the various controlling variables of regional climate and topography.

A closely related and necessary development to the one of compiling glacier inventories has been that of glacier mapping. Accurate glacier maps are an essential prerequiste for most glaciological work. For example, they are necessary for determining the effect of local variations in topography and climate on the distribution of snow and ice, for revealing variations in glacier extent, and for obtaining quantitative data on thickness, volume, mass balance, movement and surface features.

(iii) Glacier distribution studies - relation to controlling factors of climate and topography. In addition to the compilation of glacier inventories, regional descriptions of mountain glaciers, with an analysis of the controlling topographic and climatic factors on their distribution, are widespread. For example, cirque glaciers are described from the mountains of Soviet Union - esp. Urals, West Caucasus, Pamir, Tien Shan, Altai and East Siberia (Dolgushin, 1961a, 1961b; Grosval'd and Kotlyakov, 1969; Vinogradov and Konovalova, 1973); United States - Rocky Mountains (Meier, 1961a; Outcalt and MacPhail, 1965; Graf, 1977); Canada - Baffin Island (Andrews et al., 1970); Sweden - Kebnekajse Massif (Schytt, 1959); Norway - Jotunheimen (Lewis, ed., 1960); and a comprehensive account of mountain glaciers in the Northern Hemisphere (Field, ed., 1975). These descriptions show that cirque glaciers are a common feature of all glacierized mountain regions of the world.

The above-mentioned glacier studies reveal that, on a regional basis, glaciers can be shown to be situated according to the factors of latitude, continentality and altitude (e.g. studies by Leighly, 1949; Chorlton and Lister, 1971). Further climatic indices can be used to distinguish different types of 'glacier landscapes', e.g. oceanic, maritime, temperate, continental, etc. (Golubev and Kotlyakov, 1978). On a local basis, however, the distribution of glaciers in any particular area is a complex function of climatological and morphological factors (chap. III). Apart from local glaciers, snowpatches, being particularly sensitive indicators of climatic change, have also attracted attention in helping to reveal subtle controls of climate and topography on glacier distribution (e.g. Manley, 1971; Tsuchiya, 1974).

2.3.2 Glacier Front Fluctuations - Relation to Climatic Variations

Measurements of glacier-front fluctuations and their correlation with corresponding climatic fluctuations constitute one of the most common ways in which the glacier-climatic relationship has been studied. As noted in sec. 2.2, frontal measurements of glaciers have a long history. Beginning with the first systematic observations in the Alps, the work of glacier measurement has been extended to most glacierised areas of the world. With the adoption of standard terminology and recording methods, as advocated by the I.H.D. programme, comprehensive reports of the variations in the positions of glacier fluctuations in various countries have been produced by the Permanent Service on the Fluctuations of Glaciers (e.g. Kasser, 1967, 1973; Müller, 1977).

One of the main aims of these systematic measurements of glaciers is to improve our understanding of the relationship between climatic trends and glacier fluctuations. For this purpose numerous researchers have attempted to correlate historical variations of glacier-fronts with corresponding climatic variations, e.g. for the Karsa Glacier, Sweden (Wallèn, 1949); glaciers in the Washington Cascades and Olympic Mountains (Hubley, 1956); Lemon Creek Glacier, Alaska (Heusser and Marcus, 1964); glaciers in the Austrian Alps (Posamentier, 1977); and those in the Bernese Oberland, Switzerland (Messerli et al., 1978). General reviews of the relations between glacier and climate fluctuations have been carried out by such workers as Manley (1950), Ahlmann (1953), and Hoinkes (1968).

Table 2.2, compiled from the results of a number of such studies, illustrates the type of climatic variables found to be important in accounting for glacier variations. It is seen that changes in the temperature and precipitation regimes, particularly during the ablation season, are thought to be responsible for most glacier variations. While some controversy has always existed as to the relative importance of temperature and precipitation variables (e.g. illustrated by the exchange of letters between Seligman, 1944 and Callendar, 1944), it is more generally true to contend that a combination of temperature and precipitation data is needed to improve our understanding of the climatic preconditions for observed glacier variations (Messerli et al., 1978). Furthermore, given the importance in glacier balance of the many other climatological factors, such as the wind speed, wind direction and cloudiness, together with the effects of the local topographical setting, it is both difficult and unrealistic to develop simple correlations between glacier variation and temperature or precipitation records.

Apart from the identification of individual climatic variables, attempts to study broad glacier-climate relationships have included the analysis of glacier variations in relation to the frequencies of groups of weather types ('Grosswetterlagen'). Further, to avoid problems of determining the relative effect of particular climatic elements, research workers have turned to the development of relations between glacier behaviour and large-scale atmospheric circulation patterns (e.g. Hoinkes, 1968).

As noted in sec. 1.1, uncertainty about the nature of the glacier response to climatic change, complexity of glacier dynamics, and the involved nature of the factors affecting glacier balance reduces the value of the studies linking glacier snout variations with climatic fluctuations. Furthermore, such studies, while throwing some light on the importance of the various climatic variables, do not contribute to a detailed understanding of the processes controlling the glacier-climate relationship. With the realisation that, for a more complete understanding, changes in the whole glacier and not simply the snout need to be considered, attention has come to be focussed on mass balance studies which provide more direct relationships between 'glacier health' and climate.

2.3.3 Mass Balance Studies - Relation to Climate

As noted in sec. 1.1, glacier mass balance, representing the net gain or loss of snow and ice in terms of water equivalent calculated over a balance year, provides a critical link between glacier variations and climatic changes (Fig. 1.1). The realisation of this fact has led to increasing attention being devoted by research workers to the measurement and estimation of glacier mass balances.

Following the early work of Ahlmann and his collaborators on the study of glaciers around the North Atlantic coasts, (summarised in Ahlmann, 1948a), which provided a system of studying glacier regimes and related climatological variables, mass balance measurements have been conducted on glaciers in numerous areas. Many of these mass balance studies were initiated in response to the International Geophysical Year and the International Hydrological Decade programmes, e.g. the I.H.D. Combined Heat, Ice and Water Balance project, whose specific objective is "to obtain sufficient information to define and understand heat, ice and water balances and how they change with time at a number of glacier basins situated in widely differing environments in many parts of the world"[1] (UNESCO/IASH, 1970b p.9).

Most of the glacier mass balance studies have been carried out for short and differing periods of time; with the exception of glaciers such as Storbreen, Norway, where mass balance details have been known since 1948 (Liestøl,1967) and Störglaciären, Sweden, where measurements exist from 1945 (Schytt, 1962). In an effort to overcome this problem the I.H.D. committee for the Combined Balances project recommends that mass balance measurements should be undertaken for "comparable periods during a number of years at selected, representative glacier basins distributed through many different climatic regimes" (UNESCO/IASH, 1970b p.10).

Apart from mass balance measurements various related characteristics have been defined to describe the glacier's 'state of health', e.g. the elevation of the equilibrium line (where the mass

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[1]A list of mass balance stations known to be operating is given in Appendix 1 of the U.N.E.S.C.O. guide for mass balance measurements (UNESCO/IASH, 1970b).

balance is zero); accumulation and ablation gradients (Haefeli, 1962; Schytt, 1967); the vertical net balance gradient; glacier ratio - being the ratio of the accumulation area to the ablation area; 'accumulation area ratio' - being the ratio of the accumulation area to the whole glacier surface (Meier, 1962); the 'glacierisation energy' (Shumskiy, 1950); and the 'activity index' (Meier, 1961b), the latter two being expressions of the vertical mass balance gradient at the equilibrium line.

A number of studies have employed these measures to facilitate the comparison of glacier regimes, especially in areas where direct mass balance measurement are difficult (e.g. Meier and Post, 1962; Grosval'd and Kotlyakov, 1969). These generalised measures, though useful in providing an indirect estimation of the mass balance, do not have much use as predictors of glacier response to climatic fluctuations (Dugdale, 1972). Furthermore, due to the numerous micro-environmental effects of local topography on the mass balance of small local glaciers, these general characteristics do not serve as reliable indicators of the mass balance of such glaciers (Outcalt, 1965).

As with glacier-front fluctuations (sec. 2.3.2),glacier-climate studies have further been concerned with attempts at analyzing the relationships between climatic change and mass balance variations. Mass balance parameters, not being influenced by the dynamic response of the glacier, should provide better bases for the evaluation of glacier-climate relationships.

Following the early work of Ahlmann and his collaborators, (Ahlmann, 1948a), in which they showed the importance of ablation processes and particularly temperature as the dominant factor influencing the regime of glaciers in the North Atlantic region, various workers have related net mass balance to climatic variables; e.g. Outcalt and MacPhail (1965); Higuchi (1975); Bradley and England (1978b); and Tangborn (1980). These studies reveal that climatic variables similar to those associated with glacier fluctuations (Table 2.2) correlate highly with mass balance changes, e.g. summer temperature, length of ablation season, degree months above freezing (sum of monthly means above $0^{o}C$), winter precipitation and amount of summer snowfall.

Apart from the correlation of mass balance with single climatic elements, some research workers have extended their attention to a study of mass balance variations with combinations of climatic elements - 'Witterungscharakter', (e.g. Hoinkes and Rudolph, 1962). Furthermore, in certain cases, non-meteorological causes like avalanches (Lossev, 1967) and volcanic dust (Bradley and England, 1978a), together with the effect of local topographic variations (Young, 1974a) may have important controls on mass balance variations.

The mass balance of a glacier represents the balance between accumulation and ablation for a given budget year and the same mass balance could be produced by an infinite number of combinations of accumulation and ablation values (as noted in sec. 1.1). It will be realised, then, that although the development of correlations between net mass balance and climatic elements contributes to our general understanding of glacier-climate relations, for a clearer understanding of the mass balance processes it is necessary to analyse the accumulation and ablation processes individually.

Only by their separate measurement and relation to appropriate climatic and topographic controls can the relative importance of the different mass balance processes be evaluated. The realization of this fact has led to the incorporation, in many mass balance studies, of separate accumulation and ablation measurements in their study of mass balance (e.g. Sagar, 1964; Hoinkes and Rudolph, 1962; Østrem, 1964a;

LaChapelle, 1965; Anonymous, 1967; Wendler et al., 1972; and Thompson and Kells, 1974).

The various processes involved in snow and ice accumulation and their relation to environmental controls are discussed in chapter VII. The role of ablation processes in the glacier mass balance has been considered to be especially critical (e.g. Ahlmann, 1948a) and an important consequence of this has been the considerable attention of research workers being devoted to determining the relationship between ablation and various meteorological factors. This aspect of glacier study, known as 'glacial-meteorology', has involved the determination of energy balance components on numerous temperate glaciers and snow surfaces. These provide an insight into the energy transfer processes between the atmosphere and the glacier, and serve as an important means of evaluating the melt and ablation rates.

A further application of energy balance modelling and glacier balance studies has been in the prediction of snowmelt from glacier-covered drainage basins (see sec. 2.3.6). The energy balance at the glacier's surface together with other aspects of ablation are discussed in chapter VIII.

The results of detailed mass balance studies reveal considerable local variations in snow cover (e.g. on maps of mass balance variations on Peyto Glacier, Alberta by Østrem, 1966a and Young, 1971). These local variations in mass balance are caused by the interaction of numerous environmental factors controlling the accumulation and ablation processes;

attempts have been made to identify the importance of these factors through the numerical modelling of the glacier mass balance (e.g. Williams, 1974; Allison and Krauss, 1977).

Along with the development of mass balance - climate relations important attempts have been made at understanding the dynamical response of glaciers to climatic changes. Inadequacies of available field measurement, (sec. 1.2), have meant that progress in this field has largely been through the development of theoretical ideas. The main contributions have been by J.F. Nye, who in a series of papers developed his theory of glacier response to changes in mass balance (e.g. Nye, 1960 and subsequent papers). An outline of his theory can be found in Paterson (1969).

Nye's kinematic wave theory enables one to calculate, in principle, the response of any glacier to changes in mass balance. This theory has been confirmed with limited field observations relating to variations in glacier length and mass budget (e.g. with data provided for South Cascade Glacier, Washington by Meier and Tangborn, 1965). Although this points to considerable progress in this aspect of glaciology, alternative theories of glacier dynamics (e.g. Weertman, 1964; Lliboutry, 1968) reveal some disagreement among those working on this problem and so the manner in which glaciers adjust to changes in their mass balance remains a critical problem in glaciology.

With the recent improvements in field measurement techniques, such as the use of radio echo-soundingto determine thicknesses and velocities of temperate glaciers (e.g. Robin, 1975; Watts and England, 1976) it is hoped that verification and improvement of the various theories on the dynamical response of glaciers will be possible in the near future.

A final aspect of climate-mass balance studies is the possible effect of the glacier on altering the climate in its immediate surroundings. It is recognised that after a glacier is established its evolution may be determined by the glacier characteristics themselves (e.g. size, and nature of glacier surface - roughness, albedo). For example, the presence of snow and ice cover will lead to the lowering of air temperature since the heat lost by reflection of solar radiation, together with the heat used for snowmelt, is not available for transfer to the atmosphere, resulting in local cooling[1]. Various research workers have attempted to describe and quantify this effect of glaciers on climate (e.g. Tronov, 1961; Hess, 1973; Rannie, 1977). However, as noted in sec. 1.4, given the small size of the glaciers being simulated in this study, the effect of glaciers on climate can be ignored as a first approximation.

2.3.4 Cirque Distribution and Morphometry - Relation to Climate

A parallel development to that of studies concerning present-day glacier characteristics and their relation to climatological elements has been the rise of studies relating to the distribution and morphometry of cirques. Bearing in mind certain provisos, discussed by Derbyshire and Evans (1976), such as the much longer period of cirque development compared with periods of local glacier occupance, the fact that individual cirques may be formed by different glacier types, and the local importance of lithological and structural controls on cirque formation, the distribution of glacial cirques may be assumed to relate closely to that of former local glaciers.

[1] Such a cooling on a large scale has been suggested as a possible reason for the initiation of ice-sheets (Williams, 1978).

The close relation between cirques and glaciers has promoted the belief that a study of cirque form should enable inferences to be made regarding the factors affecting local glaciers. This has led to numerous studies of cirque distribution and morphometry being carried out; (e.g. Seddon, 1957; Andrews, 1965; Temple, 1965; Peterson, 1968; Sugden, 1969; Goldthwait, 1970; Andrews and Dugdale, 1971; Unwin, 1973; Williams, 1975; Graf, 1976; Trenhaile, 1976; Evans, 1977; Gordon, 1977; and Vilborg, 1977). These studies, though recognising the importance of structure, relief and duration of glaciation on cirque location, have devoted much of their attention to determining the controls of climate on cirque form, particularly as revealed by a study of cirque altitudes and orientations.

Ever since Flint's suggestion (1957) that the cirque snowline lies at or slightly above the cirque floor, one of the main aims of cirque studies has been in the estimation of palaeosnowlines from the floors of abandoned cirques. These estimates have, in many cases, been used to infer palaeoclimates by calculating the difference in temperature, at an assumed lapse rate, between the elevation of the present snowline and that estimated for the palaeosnowline. This question is discussed further in sec. 2.3.5.

Regarding cirque aspects, the identification of controls on cirque orientation by such factors as prevailing wind direction and the effects of insolation has led to much attention being devoted to the estimation of glacial climates through the study of cirque aspects. Evans (1977) has compiled available information on world-wide variations in the direction and concentration of cirque aspects and has attempted to interpret these in relation to the relevant climatic effects.

Morphometric parameters, such as the shape, size and curvature of the cirque, have also been shown to have an important control on cirque glacierisation. For a discussion on the various morphometric descriptors of cirques, refer to Andrews and Dugdale (1971), Evans (1974) or Derbyshire and Evans (1976). The results of cirque morphometry studies, revealing the importance of topographic factors in controlling the nature of glaciological processes and consequently local glacier distribution, are discussed in chapter III.

The interpretation and comparison of results from cirque studies have been hindered by the rather subjective nature of cirque definition, giving rise to inconsistencies in the delimitation of cirques and their morphometric characteristics by different workers in the same area. The geomorphometry group of the British Geomorphological Research Group has suggested a standardisation of the definition of cirques; they propose that a cirque should be defined as a "hollow, open downstream but bounded upstream by the crest of a steep slope ('headwall') which is arcuate in plan around a more gently-sloping floor. It is glacial if the floor has been affected by glacial erosion while part of the headwall has developed subaerially, and a drainage divide was located sufficiently close to the top of the headwall for little or none of the ice that fashioned the cirque to have flowed in from outside" (Evans and Cox, 1974 p.151).

The general adoption of such a broad definition of cirques will enhance the value of cirque morphometry studies and further advance their role as indicators of the spatial variations in the past and present levels of local glacierisation. It needs to be remembered, however, that given our present lack of detailed understanding of the quantitative controls of environmental factors on glacierisation, the use of cirque distribution and morphometry to infer palaeoclimates needs to be viewed with caution (Unwin, 1973).

2.3.5 Palaeoclimatic Estimates Based on Former Glacier Distributions

A major component of glacier-climate studies has been concerned with the determination of palaeoclimates through studying the evidence pertaining to former glacier distributions. This has involved the determination of palaeosnowlines and the reconstruction of former glacier distributions through glacial geomorphological evidence. These studies of palaeoclimates also serve to provide clues as to the magnitude of climatic change necessary for future glacierisation.

(i) Palaeosnowlines. A long-established manner in which glacier-climate relations have been examined has been through the determination of both modern and palaeo snowlines. The former help in identifying particular levels of glacierisation while the latter, when compared with modern snowlines, serve to indicate the changes in such factors as temperature and precipitation.

Given the considerable variation in the distribution of snow-cover due to local effects of snowdrifting and avalanching, a major difficulty is in the determination of a suitable regional measure of the snowline. The use of such general and variable concepts as the 'climatic' and 'mean' snowline has been shown to be invalid for any climatic comparison to be made (e.g. Hoshiai and Kobayashi, 1957). Many other indirect methods such as the use of independent cirque floors, the equilibrium line altitude on temperate glaciers, the relation of the accumulation to the ablation area, and the determination of a glaciation limit have been proposed to estimate the level of present and palaeo regional snowlines.

The problems in selecting a suitable expression of the regional snowline have been exaggerated by the confusion in the terminology and definitions used to describe these methods (Andrews and Miller, 1972). The common methods of estimating snowlines have been reviewed by Østrem (1966b) and Osmaston (1975) so only brief consideration is given here to three popular methods.

As noted in sec. 2.3.4,the altitudes of cirque floors have been widely used for the estimation of the heights of palaeosnowlines (e.g. Porter, 1964; Andrews, 1965; Peterson and Robinson, 1969; Andrews et al., 1970; Robinson et al., 1971; Unwin, 1973; Trenhaile, 1975; Zwick, 1980). A number of problems are, however, inherent in this method of snowline estimation.

The variation of snowline with aspect means that studies need to be based, either on cirques having similar aspect (e.g. Porter, 1964) or, preferably, the location of snowlines needs to be examined through variations in both the altitude and the aspect of cirques (e.g. Williams, 1975). If one is interested in obtaining the regional snowline during glacial maximum, then only the lowest (probably north-facing in the Northern Hemisphere) cirques should be selected. Furthermore, as has been emphasised by Flint (1971) Evans (1974), and Osmaston (1975), since the assumption of the snowline being near the cirque floor is only true for small independent cirque glaciers, careful selection of isolated cirques, not formed at the heads of valley glaciers, is necessary for valid snowline reconstruction.

Finally cirque altitudinal levels may be locally controlled by pre-glacial topography and other structural considerations rather than being simply related to palaeosnowlines (Trenhaile, 1977). Thus considerable breadth of coverage is needed in such studies. All these above considerations cast considerable doubt on the widespread use of cirque-floor levels to indicate the regional trend of former snowlines. Attention has instead been directed at two other indirect methods of snowline estimation, namely, the equilibrium line altitude and the glaciation limit (level).

The equilibrium line altitude (ELA) is the height at which the accumulation and ablation of snow and ice are balanced, i.e. mass balance is zero. Thus, variation in the ELA reflect changes in the mass balance of a glacier. Given a stable climate over a period of years, the ELA may approach a theoretical steady-state value.

While the contemporary ELA can be obtained from mass balance studies, the palaeo ELA has to be estimated from some proportional method utilising the accumulation and ablation areas of former glaciers, as derived from glacial geomorphological data. For example, (as mentioned in sec. 2.3.3), a commonly used index - the 'accumulation area ratio' (AAR) is calculated as the ratio of the accumulation area to the whole glacier surface (Meier, 1962). For steady state glaciers the AAR usually attains a value of between 0.6 - 0.7 (Meier and Post, 1962; Grosval'd and Kotlyakov, 1969).

Using such assumed AAR's, maps of ELA's have been plotted for both present (e.g. Andrews and Miller, 1972) and past (e.g. Porter, 1975a) distributions of glaciers. The various proportional methods used for the estimation of the ELA, together with their inherent problems and errors, are discussed by Osmaston (1975).

The glaciation limit or level (GL) has become a popular means to express the critical elevation that divides the glacierised from nonglacierised summits (e.g. mapping of GL by Ahlmann, 1948a; Østrem, 1964a, 1964b, 1966b, 1972; Andrews and Miller, 1972; Miller et al., 1975; Porter, 1975b).

The GL is usually calculated by the summit method which involves taking the arithmetic mean of the highest ice-free summit and the lowest glacierised summit (Østrem, 1966b). It would appear from this definition that the GL is essentially the reflection of the balance between winter accumulation and summer ablation of snow and ice. However, since some high summits may be too steep to support glaciers, the likelihood of snow and ice accumulating on a mountain is also controlled by considerations of summit morphology (Manley, 1955; Andrews et al., 1970).

An additional complication in the use of the GL concept is found in those areas where the glaciation is so intensive that few ice-free summits can be located. However, in areas of 'alpine topography' and marginal glacierisation, the relative ease and objective nature of the GL concept, as compared with other indirect methods of snowline estimation, has led to its wide application (refer to the above-given references).

Whichever of these methods for estimating the 'regional snowline' are employed, their value lie in the conclusions that can be made regarding the occurrence of present or former distribution of glaciers and their climatological conditions. The GL and the ELA, in particular, provide integrated, regional climatic indices which form an important measure of the state of glacierisation in an area and are also useful for indicating the sensitivity of an area to changes in climate (Miller et al., 1975).

Many studies have been undertaken to evaluate the climatic factors controlling these 'snowline estimates'. They have been found to correlate most closely with temperature, precipitation and other climatic factors closely related to these elements. For example, Weidick (1968) found that the GL in west Greenalnd is closely related to the summer temperature (r = 0·76). Furthermore, Andrews and Miller (1972) found the July $0^{o}C$ isotherm over Baffin Island to correlate with the height of the GL (r = 0·64 - 0·84) while over the Canadian High Arctic, Koerner (1970) and Bradley (1975) found high correlations (r = 0·82 to 0·96) between the ELA and the height of the mean July freezing level.

In contrast, the altitude of the GL in the Cascade Range of Washington is found to correlate most strongly (r = 0·93) with the winter precipitation (Porter, 1977). Further, in examining the variation of the GL in British Columbia, Alberta and southeastern Alaska, Østrem (1966b, 1972) found that its height was also closely related to the precipitation distribution. This close correlation with precipitation points towards the importance of continentality in determining glacier distribution.

The distribution of the ELA and GL is related to the glacier mass balance which is a function of both accumulation and ablation. These measures should thus reflect the combined effects of precipitation and temperature variables. In order to examine the nature of this combined effect, Porter (1977) ran a stepwise correlation program with the GL, in the Cascade Range of Washington, given as a function of five controlling climatic parameters (namely - annual, ablation-season and July temperature; accumulation-season and annual precipitation). He found that 90.4% of the variance was accounted by the accumulation-season precipitation and the estimated mean annual temperature at the GL. The inclusion of the latter variable is interesting as it contrasts with the importance attributed to the mean summer temperature by many researchers.

Differences of opinion regarding the importance of particular climatic elements in controlling the distribution of the GL and the ELA are a reflection of a number of problems inherent in such studies. Firstly, with small mountainous glaciers being particularly sensitive to minor changes in their local topographical and climatological environments, it becomes difficult to make valid general statements regarding their regional distribution (chap. III). Secondly, a proper evaluation of the climatic controls on the GL and the ELA is made difficult due to a lack of climatic stations at high altitudes.

Thirdly, it is by no means certain that the distribution of the present-day 'snowlines' are in equilibrium with the contemporary climatic conditions, due to the lag response of glaciers to climatic changes (see Table 1.1). It is, then, probable that 'snowlines' based on small cirque glaciers reflect the climate of the past several decades while larger ice blocks may still be responding to the climate of several centuries ago (Miller et al., 1975).

Finally, the use of different, not directly comparable, methods of snowline estimation further reduces the value of snowline - climate relationships. In the light of these problems, attempts to determine the climatic controls of the GL and the ELA can, at best, only reveal the influence of broad scale effects of altitude, latitude and continentality on glacier distribution.

Apart from the relation of modern snowlines to the variation of regional climatic factors, snowline-climate relations have also been studied through the determination of palaeosnowlines and their comparison with the positions of modern snowlines. The estimates of palaeosnowline lowering, thus gained, help to yield the magnitude of climatic change, since the snowline lowering is usually converted to a temperature change by applying an environmental lapse rate.

In this manner numerous estimates have been made of the palaeotemperatures during past glacial phases (such as the Little Ice Age - discussed in the succeeding section). For example, the 200 m. lowering of the palaeosnowline, determined from the distribution of cirques in Okoa Bay, northern Cumberland Peninsula, Baffin Island is equivalent to a temperature lowering of about 1.2 - 1.4°C. (Andrews and Dugdale, 1971). The extreme glaciological sensitivity of Baffin Island is further illustrated by the formation of permanent snowbanks and incipient glaciers in the northern Cumberland Peninsula, following the climatic fluctuations since the 1940's with increased winter precipitation and lower summer temperatures. (Andrews et al. 1972; Bradley and Miller, 1972).

In another region, sensitive to small changes of climate, Sugden (1977c) postulates a drop of 1.6 - 2.0°C in annual temperature for cirque glacierisation in the Cairngorms during the Little Ice Age. Some estimates of the annual temperature lowering, (average of 4.7°C), from various regions around the world based on changes in the regional snowline during the Quaternary are given by Andrews (1975, p.5). This method of palaeotemperature estimation is, thus, useful for indicating the sensitivity of particular areas to climatic change and for revealing the magnitude of climatic change necessary for glacierisation to occur.

In spite of its widespread use, the method of climatic interpretation from snowline lowering is not without its problems. Apart from the problems associated with the use of indirect methods for determining palaeosnowlines and the difficulties encountered in developing relations between snowline patterns and the climatic situation, discussed above, Williams (1975) has outlined theoretically- and empirically-tested reasons that the change in snowline altitude and change in mean temperature are not simply related. Furthermore, lapse rates (used to convert the snowline lowering to a temperature change) differ from place to place and vary over different altitudinal ranges. Climatic interpretation is further made difficult since an infinite number of possible combinations of temperature and precipitation could be responsible for the observed snowline differences.

The correct way to examine the temperature estimates would be to use an energy balance approach which includes the consideration of both accumulation and ablation factors (Williams, 1974, 1975). In the light of all these problems, specific causes for snowline variations cannot be isolated and comparison of estimated palaeosnowlines with present-day ones, at best, only provides a measure of the gross difference in climate between the present and earlier periods of glacier advance.

(ii) Reconstruction of former glaciers ('Little Ice Age')

Along with the mapping of palaeosnowlines, attempts at the detailed local reconstruction of former glaciers, from geomorphological and other evidence, have yielded further information on glacier-climate relations. Sugden (1977c), for example hypothesised about the former distribution of cirque glaciers in the Cairngorms using evidence regarding the positions and nature of the moraines at the cirque lips. Detailed reconstruction of local glaciers using geomorphological evidence has also been attempted for the south-east Grampian Highlands of Scotland by Sissons and Sutherland (1976). They analysed the distribution of the reconstructed glaciers by reference to such factors as direct radiation, effect of shading by higher ground, avalanching and snow drifting. They infer average July and January sea-level temperatures of 6°C and -8°C respectively,with precipitation similar in total to present-day but rather differently distributed. Regional palaeoclimatic inferences from former glaciers in Scotland and the Lake District are further discussed by Sissons (1979, 1980).

Many of the attempts at local reconstruction and palaeoclimatic studies are concerned with the conditions during the 'Little Ice Age'[1]. Interest in the events of this period is high due to the fact that this is the last time that many northern latitudes achieved a condition close to intensive glacierisation. In fact many of the present-day local glaciers originated and achieved their maximum extent during this period. Thus, in order to ascertain the orders of magnitude of climatic change necessary for marginal glaciation, numerous attempts have been made to

[1]Little Ice Age - a period of climatic cooling generally dated from 1500 - 1920 A.D. (Sugden and John, 1976, p.114).

reconstruct the glaciers and climate during this period.

This has involved the use of evidence gathered from a wide variety of sources, e.g. the use of long instrumental climatic records (especially for the latter part of the 'Little Ice Age'). Estimates of the Central England temperatures from 1659 have been made by Manley (1974) while temperature records from Holland date from the beginning of the 18th century (Ahlmann, 1948b), from Stockholm since 1756 (Liljequist, 1949), and from Stykkisholmur, Iceland since 1845 (Eythórsson, 1949).

The extent of sea ice off Iceland has been correlated with temperature changes (e.g. Schell, 1961; Bergthorsson, 1969). Indications of palaeoclimatic variations have also been provided by changes in the isotopic composition of oxygen in the Greenland ice-cores (Dansgaard et al., 1975), mapping of lichen-covered areas to indicate the extent of 'Little Ice Age' perennial snow and ice cover (e.g. in the Baffin Island - Ives, 1962; Locke and Locke III, 1977), tree-ring data related to climatic and glacier fluctuations (e.g. Matthews, 1977), mapping of terminal moraines and other geomorphological evidence for the reconstruction of glacier fluctuations (e.g. reviews by Denton and Porter, 1970; Denton and Karlén, 1973), and by correlation of glacier fluctuations with seasonal temperature trends (e.g. Manley, 1950).

The evidence from these and other sources for the climatic conditions during the 'Little Ice Age' has been reviewed by Ladurie (1972) and Lamb (1977). In general, it appears that temperatures during the 'Little Ice Age' were 1-3^{o}C lower than present values.

2.3.6 Snow and Ice Management for Water Resources

The understanding of glacier-climate relations has been further advanced by attempts at snow and ice management for water resources. The importance of glacier ice as a source of fresh water has already been mentioned in sec. 2.1. Given this importance many glacier studies, discussed above, have taken on an applied aspect with the aim of estimating, predicting and controlling the amounts of water stored and released by snow and ice masses. For example, snow surveys (sec. 2.3.1) are valuable for forecasting snow cover runoff (e.g. Tsuchiya, 1974) while detailed mass balance studies (sec. 2.3.3) provide similar information on river runoff from glaciers (e.g. Østrem and Pytte, 1968 on the work of the Norwegian Water Resources and Electricity Board).

Furthermore in areas such as the temperate alpine mountain zones, where snow and ice constitute an important component of the water balance, snow management techniques involving the construction of snow fences, intentional avalanching, terrain and albedo modification have gained wide application for controlling the distribution, size and rate of melting of snowpacks and glaciers (Martinelli, 1966, 1975).

One of the main aims of snow management is to increase the summer streamflow originating from snow and ice masses. Given a considerable amount (approximately 60% according to Meiman and Grant, 1974) of seasonal precipitation in alpine areas may be lost through evapo-sublimation by wind transport of snow to lower elevations, attention of snow management schemes has been focused on the more effective trapping and storage of snow.

While, in the high elevation forested watersheds of the subalpine zones, the effect of forest management practices, such as timber harvesting, on the accumulation and melt of snow is dominant (e.g. reviewed by Leaf, 1975 for the Rocky Mountain Subalpine Zone), in true alpine areas wind transport of snow from exposed areas and its deposition in the lee of terrain breaks and other obstacles is important.

It has been found that the size and shape of snowfields varies with the size, shape and orientation of the barriers behind which they form. Thus an obvious technique for improving water yields is the construction of snow fences upwind of natural snowfields to increase the amount of snow held in these areas until late summer. Experiments with snow fences of different materials, sizes and shapes in alpine areas have shown that snow depths can be substantially increased by this technique (Martinelli, 1965, 1973; Tabler, 1971; Slaughter et al., 1975). Of related interest is the designing of suitable barriers for the protection of transportation facilities from the effects of snow drifting and avalanching (e.g. Price, 1961; Tabler, 1980).

Further trapping and storage of snow can be achieved by terrain modification at a local scale, through shaping snow accumuation areas into more efficient configurations for snow deposition, e.g. by the deepening of natural depressions. Intentional avalanching of snow deposits forms another potentially useful snow management practice. This involves the periodical removal of snow, from cornices and other potential avalanche deposits, located in areas where it is liable to considerable evapo-sublimation loss, into sheltered confined localities, e.g. cirques, where it can be stored and released during the summer. A study by Santeford (1972) has shown the economic viability of this snow management technique.

Since snowpack and glacier melt is extremely sensitive to changes in the reflectivity of their surfaces, albedo modification through artificial means provides a simple technique for controlling snowmelt. Experimental studies have shown that melting rates can be substantially increased by spreading a thin layer of dark powder, such as coal dust, soot or ash on the snow surface (e.g. Arnold, 1961). Conversely thick layers of insulating materials such as sawdust, soil or sheets of polystyrene foam (Higuchi, 1973) can be used to reduce the melting rates. The relative ease and efficient manner of this snow-management technique have led to its widespread application (e.g. Slaughter, 1969; Kotlyakov and Dolgushin, 1973). A comprehensive, but dated, literature review on albedo modification has been compiled by Slaughter (1969).

Snowmelt runoff forms the bulk of stream discharge in many of the alpine and subalpine zones (Leaf, 1975; Martinelli, 1975). Thus, along with the development of the various management techniques to control snow accumulation and melt, considerable attention has been paid to attempts at forecasting the amounts of meltwater from snowpacks and glaciers. The importance of these forecasts for various purposes, (mentioned in sec. 2.1), has led to the emergence of 'snowmelt hydrology' as a separate specialised study area within the disciplines of glaciology and hydrology.

The development of snow-hydrology studies has considerably aided our understanding of the processes involved in the melting of snow and ice (sec. 8.1). This increase in our knowledge of snowmelt processes, together with the need to improve runoff forecasts, has led to the development of computer simulation routines of the snowmelt process and their incorporation into runoff models such as the Stanford Watershed Model (Crawford and Linsley, 1966) and the National Weather Service River Forecasting System (Anderson, 1973b). A further trend associated with the increased use of snow management techniques has been the growth of models specifically designed to simulate watershed management practices and their resultant effects on hydrologic system behaviour (e.g. Leaf and Brink, 1973; 1975). These various aspects related to the modelling of snowmelt are discussed more fully in chapter VIII.

TABLE 2.1

Studies Relating to Glacier-Climate Relationships

Studies relating to glacier-climate relationships

(1) Distribution of Snow and Ice Masses - surveys; inventories; relation to controlling factors

(2) Glacier Front Fluctuations - relation to climatic variations

(3) Mass Balance Studies - Relation to Climate - mass balance measurements; accumulation and ablation studies; correlation with climatic variables; effect of glaciers on climate.

(4) Cirque Distribution and Morphometry - relation to climate.

(5) Palaeoclimatic Estimates Based on Former Glacier Distribution - palaeosnowlines; magnitude of climatic change necessary for ~~glacierisation~~; ~~reconstruction~~ of former glaciers.

(6) Snow and Ice Management for Water Resources - snowmelt hydrology; snowfences; avalanche control; albedo modification.

TABLE 2.2

Relationships Between Glacier and Climate Fluctuations

Study Area	Climate Variables Responsible for Glacier Fluctuation	Reference
Austrian Alps (c.1898-1969)	Regression equations: (a) involving several temperature and precipitation variables for the ablation months; accounting for 67% of the variation in Austrian glaciers. (b) mean temperature for June-August for the concurrent plus preceding 7 years; accounting for 71% of the variation.	Posamentier (1977)
Karsa Glacier, Sweden (1942-1948)	Recession of glacier due to: increased summer temperature, increased summer humidity, prolonged ablation season (possibly increased wind velocity).	Wallén (1949)
Pasterze Glacier Glockuer Alps (c.1796-1950)	Average summer temperature; average autumn, winter and spring precipitation. Radiation. Number of summer snowfalls (affecting albedo). Local/regional winds.	Schwartz (1974)
Jan Mayen glaciers (c.1950)	Advance of glaciers due to: annual precipitation increase; winter precipitation increase, slight drop in summer temperature.	Lamb et al.(1962) Sheard (1965)
Washington Cascade and Olympic Mountains of Washington State (c.1920-1930).	Decreasing temperature; increasing precipitation	Hubley (1956)
Bernese Oberland Switzerland (1600-1975)	Advance of glaciers associated with: a succession of short term (5-7 years) periods of cold and wet summers.	Messerli et al. (1978)

CHAPTER III

FACTORS AFFECTING LOCAL GLACIER BALANCE

The variety and scope of glacier-climate studies have been examined in the previous chapter. It is seen that many of these studies (e.g. snow surveys, glacier inventories, glacier fluctuations and palaeosnowlines) are concerned with the examination of glacier-climate relations at a large scale. While such studies are useful in permitting broad generalisations to be made regarding the factors affecting glacier distribution, many of the relationships developed will break down at the local scale of cirque glacier distribution. This is due to the important influence of local topography on glacier balance.

In this chapter the main environmental factors affecting the local glacier balance are identified. This will help to determine the structure of the mass balance simulation model, outlined in chapter IV.

3.1 Local Glacier Studies and the Role of Topography

Glacier studies in areas of marginal glaciation, where the effect of local variations in topography and climate is most apparent, best reveal the complexity and diversity of the factors affecting glacier distribution. For example, research workers from the Institute of Arctic and Alpine Research have been concerned with the examination of factors affecting cirque glacierisation in eastern Baffin Island (e.g. Andrews et al., 1970, 1972; Andrews and Dugdale, 1971; Williams, 1972, 1975). They stress the importance of various 'indirect' measures of the cirque environment, such as cirque altitude, size and aspect in controlling the glacier mass balance.

In particular, differences between glacierised and non-glacierised cirques could be explained by differences in cirque aspect rather than the usual 'snowline elevation interpretation' (sec. 2.3.5). As noted by Derbyshire and Evans (1976, p.453), "...asymmetry is stronger in lightly glaciated mountains, i.e. where the snowline is only just low enough to support glaciers. As glaciation becomes less marginal, glaciers from on all aspects..."

In Baffin Island the importance of aspect for glacier growth is attributed to insolational advantages of the north-facing cirques. Thus, Williams (1972) found the separation of glacierised and non-glacierised cirques on the basis of elevation and solar radiation to be particularly effective. Apart from solar radiation,the importance of aspect at the local scale is manifested by its control on such processes as snow drifting and avalanching.

Among studies emphasising the importance of 'directional control' on glacier balance, with small changes in cirque aspect being responsible for substantial changes in mass balance, is that of Alford (1973) who examined the mesoscale aspects of cirque glaciers in the Colorado Front Range. He postulates that variations in the winter balance are largely the result of variations in the angular relationship between cirque orientation and the mean winter storm path trajectory. Variations in the summer balance can be ascribed to orientation controls on the amounts of shortwave solar radiation receipts. Alford expresses these controls by developing equations linking the orientation of glaciers with their expected mean specific mass balance.

Likewise, with respect to the glaciers of the Central Asian ranges, Dolgushin (1961a, b) notes that the degree of glacierisation of the mountain slopes is directly dependent on their aspect. Favourable aspects lead to additional accumulation through windblown and avalanched snow, lowering the snowline by many hundreds of metres below the theoretical 'climatic snowline'. Such contrasts in the relative positions of both present and palaeo snowlines (sec. 2.3.5) on slopes of differing aspects illustrate the influence of topo-climatic effects on local glacier balance.

The influence of the various topoclimatic effects on mass balance, and in turn glacier aspect, has been examined in detail by Evans (1974, 1977). He discusses the tendency for the production of shade, lee- and east-ward facing glaciers due to the effects of radiation, wind, and morning: afternoon contrasts respectively.

Apart from aspect, other topographic measures - such as slope gradient and surface concavity, also reveal a close correlation with the distribution of snow and ice cover. Young (1973, 1976), for example, has developed a method, employing a grid-square technique by which maps of snow accumulation and ablation can be produced from a few sampling points through utilising relationships between terrain characteristics and snow depth. He applied the method to Peyto Glacier, Alberta and presented maps of topographic parameters used together with predicted snow accumulation and ablation for the entire glacier area (Young, 1971). The results show the importance and feasibility of surface terrain parameters for predicting glacier balance.

Further evidence of the important role of topography in controlling local glacierisation comes from work on cirque morphometry (sec. 2.3.4). Analysis of seventeen variables describing the shape, location and geometry of cirques located in Okoa Bay, northern Cumberland Peninsula, Baffin Island by Andrews and Dugdale (1971) reveal the importance of cirque altitude and aspect in determining the state of cirque glacierisation.

Graf (1976), by comparing the sites of cirques containing glaciers with those that are empty in the Rocky Mountains, has further shown the importance of cirque morphometry and cirque environment in influencing glacier location. For glacier occupance, the effect of shading from insolation offered by the northerly aspects and high walls of cirques, snow accumulation through avalanching from the surrounding steep slopes and snow drift from plateau surfaces upwind are found to be important factors. Based on these effects Graf considers an optimum glacier location for the Rocky Mountains to be "a large cirque facing northeast, with a planimetric shape of width greater than length, high steep walls, a pass located to the windward, and a peak to the southwest".

3.2 Field Observations of Local Glaciers in Iceland and Italy

The close control of snow and ice distribution by local topography was also evident in the field observations made by the author during visits to glacierised regions in northern Iceland and northern Italy. As a member of the Durham University Vestfirðir Project (described by John and Alexander, 1975) observations of snowpatches and cirque glaciers were made in north-western Iceland (Vestfirðir region) during the summers of 1976 and 1978. Glaciers in northern Italy

(Ortles-Cevedale range) were observed in the summer of 1976 and in north-central Iceland (Tröllaskagi region) during the summer of 1978.

3.2.1 Vestfirðir

Vestfirðir is a typical Tertiary basalt district consisting for the main part, of a gently undulating plateau with an average altitude of 500-800 m. above sea level. This plateau is intersected by a number of steep-sided fjords and fjord systems, e.g. Isafjardardjup and Arnarfjordur. The western part of Vestfirðir is distinguished by an area of 'alpine topography', with considerable development of cirques and trough-ends occupying the valley sides (Figs. 3.1, 3.2, 3.3). Between the cirques and troughs dissecting this region, numerous plateau remnants still remain (Fig. 3.4). John (1975) has recognised eight major glacier erosional landscape types in Vestfirðir (Fig. 3.5).

The glaciation level varies in Vestfirðir from about 600 m. in the extreme northern part of the region to over 1000 m. above sea level in the south. At present the glacierisation in this region consists of the Drangajökull ice cap in the north-east and a number of small cirque glaciers in isolated northern localities (see Fig. 3.5). Vestfirðir is a glaciologically marginal area which has remained as a relatively isolated peninsula, being removed from the direct influence of the Pleistocene Icelandic ice-sheets.

The extreme sensitivity of Vestfirðir to climatic changes is evident from a study of the numerous climatologically induced variations of the Drangajökull outlet glaciers since 1700 (Eythórsson, 1935). Furthermore, according to historical records, during the middle of the 18th century the Glama plateau in the central part of the region was covered by a relatively large ice cap (Glamajökull). However according to Thorarinsson (1943, p.16) the Glama plateau was free of ice in 1893 and, on the whole, has remained ice free up to the present times. Thus Glamajökull seems to appear and disappear in response to medium scale climatic changes (John, 1976).

This glaciological sensitivity is further illustrated by observations of snow and ice distribution in the western 'alpine' area. In the absence of glacier ice in the majority of the cirques, due to the glaciation level being slightly higher than the cirque altitudes, attention was first directed at the distribution of snow-patches in early August. Numerous local snowpatches are found to occupy sheltered localities provided by small concave hollows, gullies or the stepped topography formed by different basalt flows.

For local glacier generation, however, it is the rather larger (c. 100 m. length) snowpatches that are important. These are invariably located in the cirque floors at the foot of steep avalanching slopes. This can be seen in Fig. 3.6 which is a field sketch of snowpatch distribution in August, 1978 in an 8 km^2 area near Nupur. Sheltered from wind deflation and the direct rays of insolation, the snow in these cirques can survive well into the ablation season (Fig. 3.7). In fact some of these snowpatches appear to be perennial. Their location in cirques adjacent to plateau remnants may be an important controlling factor in their distribution.

The plateau remnants act as source areas for snowdrift which is swept into the leeward cirques and other suitable concavities. This importance of snowdrift accumulation for the survival of major snowpatches is shown by the fact that in Fig. 3.6, the cirque in the extreme north-eastern part of the area mapped is devoid of any snow, reflecting the lack of a substantial adjacent plateau remnant.

The importance of cirque location leeward of plateau areas is further substantiated upon consideration of some small cirque glaciers that occur at the head of Seljalandsdalur and Hattardalur valleys on the northern edge of Lambadelsfjall (Fig. 3.5). These glaciers are located in north and north-east facing concave hollows backing into the plateau edge, between 600-700 m. above sea level (Fig. 3.8). Similar glaciers situated at lower altitudes (c. 400 m) are also found north of Isafjardardjup (Fig. 3.5). They probably owe their existence to the avalanching of snow from cornices at the plateau edge, fed in turn by snowdrift from the plateau. Distinct neoglacial moraines can be observed in front of these glaciers indicating their past activity and larger areal coverage than at present.

3.2.2. Tröllaskagi

Tröllaskagi, located between the Skaga and Eyja fjords, in northern Iceland, is another marginally glacierised area characterised by the development of numerous discrete local glaciers. The region, extending for about 55 km. from east to west and more than 75 km. north to south, represents a deeply-incised mountain range. The highly rugged and irregular terrain dissects a 1000 - 1500 m. high plateau. In contrast to the Vestfirðir area, plateau remnants are few and small, the divides being formed by sharp ridges and high peaks. The greatest altitudes (> 1300 m.) occur in the central region along the main water-shed, between Hörgardalur and Skidadalur (Fig. 3.9). The numerous cirques and trough-ends in this region are often separated by steeply-towering ridges many hundreds of metres high.

According to data gathered from aerial photographs (taken by the U.S. Air Force in 1960), 1:50,000 photogrammetric maps (prepared by the U.S. Army Map Service), and field observations of the Iceland Unit (Young Explorer's Trust) involved in developing a North Iceland Glacier Inventory (Escritt, 1974, 1976), there are approximately 110 perennial snow and ice masses in this region (Fig. 3.9). Most of these can be distinguished as small localised glaciers or glacierets of the cirque and hanging variety.

The majority of the cirque glaciers have a northerly aspect. The glaciation limit lies at an altitude of about 1000 m. in the northern parts of the mountain range and ascends to over 1300 m. in the central part. Given the markedly irregular topography, avalanching of snow off the steep enclosing slopes and its accumulation in the cirque floors appears to be the main factor responsible for the survival of these local glaciers.

3.2.3 Ortles-Cevedale Group

The third area of mountain glacierisation in which observations of glaciers were carried out is the Ortles-Cevedale range in northern Italy. The Ortles-Cevedale mountains form part of the Rhaetian Alps and are much more intensively glacierised than the northern Icelandic regions. The group consists of four major ranges radiating out from the central node of Mt. Cevedale (3778 m.) as shown in Fig. 3.10. Mt. Ortles (3899 m.), on the western range forms the highest point of the group and many other peaks exceed 3000 m.

The distribution and characteristics of the Ortles-Cevedale glaciers have been studied in detail by Desio and his collaborators (Desio et al., 1973). They describe the occurrence of 132 glaciers in the region covering an area of approximately 96 km^2. The glaciers are located in three hydrographical basins - the Adda basin with 48 glaciers, the Oglio basin with 2 glaciers and the Adige basin with 82 glaciers. As a result of the varied landscape forms a great variety of Alpine glacier forms can be distinguished in this region. The majority (87%), however, can be classified as local 'slope glaciers', which suggests the control of topography on glacier location.

With regard to glacier distribution, the climatic snowline in the Ortles-Cevedale region is calculated at around 3000-3100 m. However, the control of slope aspect is particularly striking for glacier location, with northerly aspects being most dominant. From a survey of glaciers carried out in 1961, Desio et al. (Table 6, p.39) determined that 58% of the glaciers or 61·2% of the glacierised area have northerly components in their aspects.

The northerly aspects become progressively more dominant at altitudes increasingly below the climatic snowline (i.e. in conditions of more marginal glaciation). Thus the largest glacierised area for the northern orientation is situated at 2900 m. above sea level while for the southern orientation it is at 3070 m. Below (to 2210 m, the absolute minimum altitude for glaciation) only N, NNW and NNE aspects occur.

Desio et al. (p.46 ff) considered the further influence of local topography in glacier location by calculating a rough 'orographical coefficient', taken to be the altitude (in metres) by which the average altitude of a glacier falls above or below the snowline because of the influence of the orographical factors, in particular the shape of the hollow containing the glacier and its orientation. The orographical coefficient is made up of two elements, namely the orographic coefficient and the morphological coefficient.

The orographic coefficient was calculated by Desio et al. as the average difference (positive or negative) between the average altitude of the area of the glacier, (corresponding to the altitude of the orographical snowline), and the altitude of the climatic snowline (calculated as 3050 m.); the data being related to the orientation of the glacier. The results reveal a clear influence of aspect, with northerly component having negative values and southerly components positive values.

The morphological coefficient depends solely on the morphology of the glacier and its determination was based on an examination of the type of alimentation received by each glacier. The average value of the average altitudes of the glaciers, for each of the different types of alimentation, relative to a height of 3050 m. and to the heights obtained using the orientation coefficient were determined by Desio et al. (Table 15, p.51) for the 1961 glacier survey as follows: for glaciers with direct alimentation (+30 m.); direct and from avalanches (-10 m.); from avalanches and reconstructed (-138 m.). Based on these considerations an average morphological coefficient was determined for different glacier types; e.g. valley glaciers (+22 m.), slope glaciers (+100 m.), cirque glaciers (-2 m.), ridge glaciers (+41 m.), and summit plateau glaciers (+391 m.).

The above discussion serves to illustrate that spatial variations in the distribution of snow and ice at the local scale are closely related to the variability of the topographic factors, especially local altitude, aspect, and surface concavity. Thus an accurate portrayal of the topography is essential for the successful modelling of local glacier distribution (the nature of topographic inputs is discussed in chap. V). A second important control on glacier mass balance is provided by the factors of local climate.

3.3 The Role of Climate

The crucial role of climate for glacier distribution has been stressed in the various studies concerned with developing broad glacier-climate relationships, reviewed in chap. II. The use of regional macroclimatic indices to define areas of local glacier ice growth is, however, unjustified since local topography may allow the growth of ice, sheltered from the summer insolation and fed by snow drift, in areas where the regional macroclimate data does not offer any clues for the presence of glacier ice.

Studies of local glacier environment (e.g. Alford, 1973) further reveal that mass balances may vary greatly within the same regional climate. Clearly, then, for local glacier distribution, consideration of local variability in climatic elements is necessary. This may be obtained by determining the manner and extent of modification by terrain of regional climatic values (chap. VI).

Fundamental climatic elements such as precipitation, air temperature, cloud cover, wind speed and direction will, all, obviously, have an important control on glacier mass balance (sec. 2.3.3). For example, consideration of air temperature is necessary for such purposes as determining the proportion of precipitation that will fall as snow at any altitude, whether the snow is 'dry' enough for substantial drifting to take place, and for determining the effectiveness of the energy exchange processes for ablation to take place. The importance of summer snowfalls for glacier balance, through the reduction of the ablation season and increased reflection of radiation by the snow surface, has been shown by Tronov (1962) in his study of the glaciers in the Central Altai region.

The characteristics of the wind near the earth's surface are of major importance in controlling the amount of snow movement and so determining the scour and depositional patterns through the effects of snow drifting and avalanching. In addition, transport of snow by wind will result in changes of snow density and other physical properties. Furthermore, wind speeds will influence the amount of turbulent heat exchange taking place between the glacier surface and the atmosphere. The radiation component of the energy exchange will be affected by the degree of cloudiness.

These various climatic elements experience a great deal of spatial variability in mountainous areas and are closely linked to the topographic factors mentioned earlier (sec. 3.2). Most of them show clear variation with altitude in the first instance and are further modified by the influence of slope aspect, gradient and surface concavity. For example precipitation amounts generally increase with altitude and also vary with broad scale aspect effects, windward slopes experiencing higher precipitation than leeward ones (though at the local scale this effect is reversed by the drifting of snow to leeward locations). The variability found in the climatic elements, their influence on the mass balance and the manner in which they are related to the local topography are all discussed further in chapter VI.

3.4 The Importance of Processes in Mass Balance

While studies linking variations in glacier mass balance to particular topographic and climatic factors, mentioned above, are useful in indicating their important role in controlling the distribution of snow and ice cover, the variability of the mass balance processes leads to variation between topographic and climatic indices applicable in different areas of glacierisation. Thus, the coefficients in Young's (1971) equations linking snowdepth and terrain parameters (sec. 3.1) will vary from year to year and on different glaciers since the method is not process-based but relies on field sampling of the density and depth of snow.

Such procedures provide little understanding about the nature of the interactions between the different processes affecting glacier mass balance. For example, while summer temperature is regarded as a key factor in determining ablation amounts because of its close correlation with melting, in actual fact there is no simple relationship between air temperature and glacier mass balance. This is because summer melting is not determined by air temperature but by the amount of energy absorbed at the glacier surface. This, in turn, is determined by the amount of incident radiation, the albedo of the glacier surface material and the turbulent structure of the lowest part of the atmosphere. All of these factors experience great spatial and temporal variability.

Thus if the amount of accumulation or ablation is simply related to topographic or climatic indices without regard to the processes involved, the model will be applicable only in the area where the indices were formulated. If the intensity or type of climatic variable or the topographic environment changes, there will be corresponding changes in their relationships with mass balance. Clearly, to develop a mass balance model of general applicability, providing information on the dynamic effects of the processes responsible for inducing changes in the mass balance, consideration of the actual processes of both snow accumulation and ablation is necessary (sec. 1.3).

Although some attempts have been made to model the processes affecting glacier ablation, through consideration of the energy exchange at the glacier surface (e.g. Williams, 1974, 1975), little attention has been directed to such processes as snow drifting and avalanching. These latter processes may be crucial for the growth and survival of small cirque glaciers, e.g. in the Colorado Front Range (Outcalt and MacPhail, 1965), Urals (Dolgushin, 1961a), and northern Iceland (sec. 3.2). Difficulties in modelling these processes have been mentioned in sec. 1.2; however, their close control by factors of local topography and climate points to the feasibility of modelling local glacier balance through establishing quantitative links between mass balance processes and relevant topographic and climatic variables.

Snow drifting, for example, is affected by such climatic controls as changes in winter wind velocity and direction together with the temperature and mobility of the freshly fallen snow. The crucial levels of wind velocity, thickness of initial snow cover and its temperature for the commencement of drift can be determined. Local topographic control is also evident in the necessity for exposed plateau areas to permit snowdrift, and concavities or gentle leeward slopes for its accumulation. The build-up of wind drifted snow in certain localities (such as slopes with steep gradients or convex profiles) will result in its avalanching. In the process of snow transportation some of it may be lost through sublimation, depending on temperature, wind speed, humidity and insolation conditions. The nature of snow accumulation processes and the manner in which they can be related to topographic and climatic variables is discussed in chapter VII.

For ablation also, spatial and temporal variations in the absolute and relative magnitude of the energy exchange processes are controlled by the relevant climatic and physiographic factors. For example, solar radiation receipt (often the most important component in energy exchange) varies with altitude, slope gradient, aspect and the duration of radiation; the latter being controlled by the time of sunrise and sunset expected at the slope of the surface, and modified by the effect of surrounding terrain blocking the sun's rays. Cloudiness further modifies the amount of direct solar radiation while its effectiveness for melting will be controlled by air temperature (sec. 8.2). The turbulent transfer of heat between the snow and the atmosphere will in turn depend on the temperature, wind and humidity conditions above the glacier surface. The role of glacier ablation processes together with the modelling of the energy balance components is considered in chapter VIII.

Apart from consideration of topography and climate in controlling mass balance processes, 'glaciological factors' - concerned with the physical properties of the glacier surface, also play an important role in mass balance determinations. For example, the albedo of the glacier surface is a critical factor in controlling the amount of solar radiation absorbed by the surface. The albedo varies with the angle of incidence of solar radiation. It is also dependent on the density and age of the snow or ice. Thus fresh summer snowfalls can have a decisive effect on glacier regime by reducing ablation through increasing albedo (Tronov, 1961). Further, such variables as the surface temperature, roughness and emissivity of snow and ice must be considered for energy balance calculations; while density and water equivalent are important indices depicting the physical state of the material forming the glacier.

3.5 Identification of Relevant Mass Balance Factors and their Interactions

From the brief discussion above of the variables important in the determination of glacier mass balance, it is clear that three sets of controlling variables need consideration, namely topographical, climatological and glaciological factors (Table 3.1).

The relevance and importance of these different factors for glacier mass balance determination will depend on the particular scale of study employed. Glacier distribution can be visualised at a macroscale, emphasising regional variability; mesoscale, emphasising local or within-region variability; or at a microscale variability level. From the various inventories of local glacier distribution in alpine areas (sec. 2.3.1), it can be seen that the sizes of most cirque glaciers vary from less than 0·5 sq. km. in area to about 2-3 sq. km.

Determination of the scale at which most glaciological and climatological processes act is more difficult. Some glacier characteristics such as albedo and density, together with such climatological effects, as wind speed and direction, may change over short distances; while other controlling variables, such as the distribution of precipitation, may reveal the operation of larger scale effects as is evident from differences in precipitation amounts between windward and leeward sides of mountain ranges. However, studies of snowcover accumulation and ablation patterns reveal that the majority of the processes affecting local glacier mass balance operate at the mesoscale, i.e. with characteristic linear distance variability of 10^2 to 10^3 m.

It is precisely at this scale of variability that spatial variations in the mass balance processes are closely determined by variations in local topography. Thus, factors important for the regional variability of glaciers, such as latitude and continentality, and those depicting microscale effects concerning differences within distances of 10 to 10^2 m. are not relevant at this scale.

It is interesting to note that the majority of glacier studies have been concerned with either mass balance measurements over small areas of individual glaciers or with large scale studies of the regional distribution of glaciers. Relatively little work has been done at the mesoscale level of variability.

Having identified the factors important for mass balance determination, the next step is to formulate the interrelationships between them to facilitate the assessment of their individual influences on glacier mass balance. This has been attempted in a general way in

Fig. 3.11. It is clear that, even in general outline, the scheme presented reflects the complex pattern of the interaction of climate and topography with local glacier balance.

Some of the controlling variables, e.g. solar radiation, air temperature and snowfall are not mutually independent. Others (e.g. albedo) are affected by feedback mechanisms operating between them and the mass balance processes. The influence on climatological variables of local topography is also portrayed.

The close control of topography over local climate, together with the accumulation and ablation processes, suggests that a spatial model of local glacier distribution can be developed by relating mass balance factors to topographic variables. Since these topographic variables can, in turn, be derived from altitude values (sec. 5.3.1), glacier mass balance variation can be estimated from the input of local altitude and regional climate variables alone, provided the interactions between the various mass balance components in Fig. 3.11 can be quantified. A model which attempts to simulate these interactions is presented in the following chapter.

TABLE 3.1

Some Critical Factors Affecting Local Glacier Balance

Topographical	Climatological	Glaciological
Altitude	Solar radiation	Surface temperature
Gradient	Air temperature	Albedo
Aspect	Precipitation	Density
Surface concavity	Wind speed	Water equivalent
Relief	Wind direction	Surface roughness
Area of neighbouring steep slopes	Cloudiness	Emissivity
	Humidity	
Shading by surrounding topography		

FIGURE 3.1
Looking N.N.E. into Nupsdalur. Note the cirques (Hrútaskál and Geldingadalur) cutting into the plateau margins.

FIGURE 3.2
Looking S.E. from Head of Nupsdalur across Mouths of the Miðhvilft and Selijahvilft Cirques (right), and across Nupsdalur to Klukkulandschvilft.

FIGURE 3.3
Looking West into Geldingadalur Cirque About 9 km. South of Thingeyri.

FIGURE 3.4
Undulating Plateau with Blockfields in the Nupur Area.

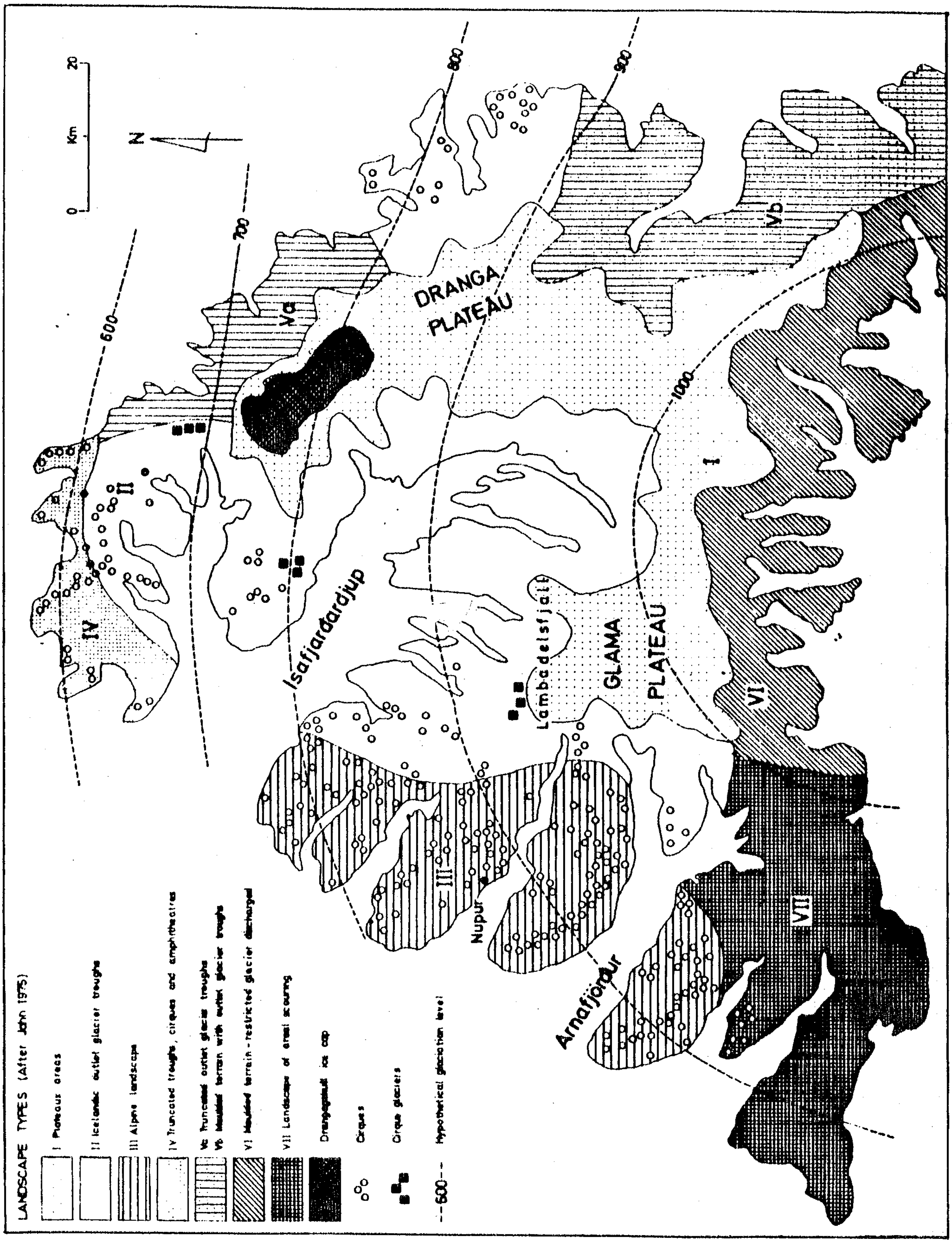

FIGURE 3.5
Sketch-Map of the Glacierisation and Landscape Types Found in Vestfirdir.

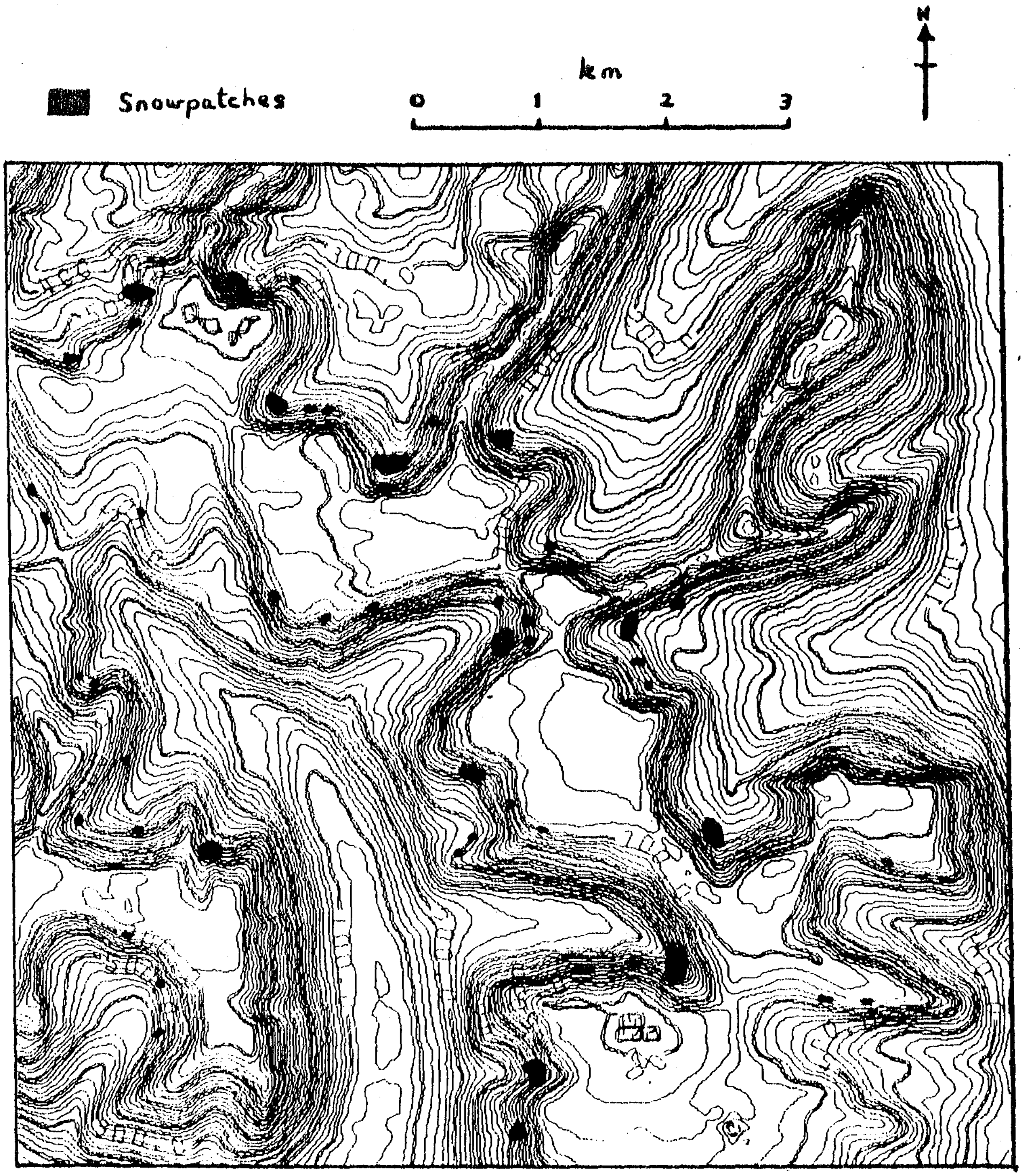

FIGURE 3.6
Snowpatch Distribution in the Nupur Area (August 1978).

FIGURE 3.7
Snowpatch at the Head of Geldingadalur Cirque in the Nupur Area, 12th August 1978.

FIGURE 3.8
View of Cirque Glaciers at the Head of Hattardalur on the Northern Edge of Lambadelsfjall (Vestfirdir).

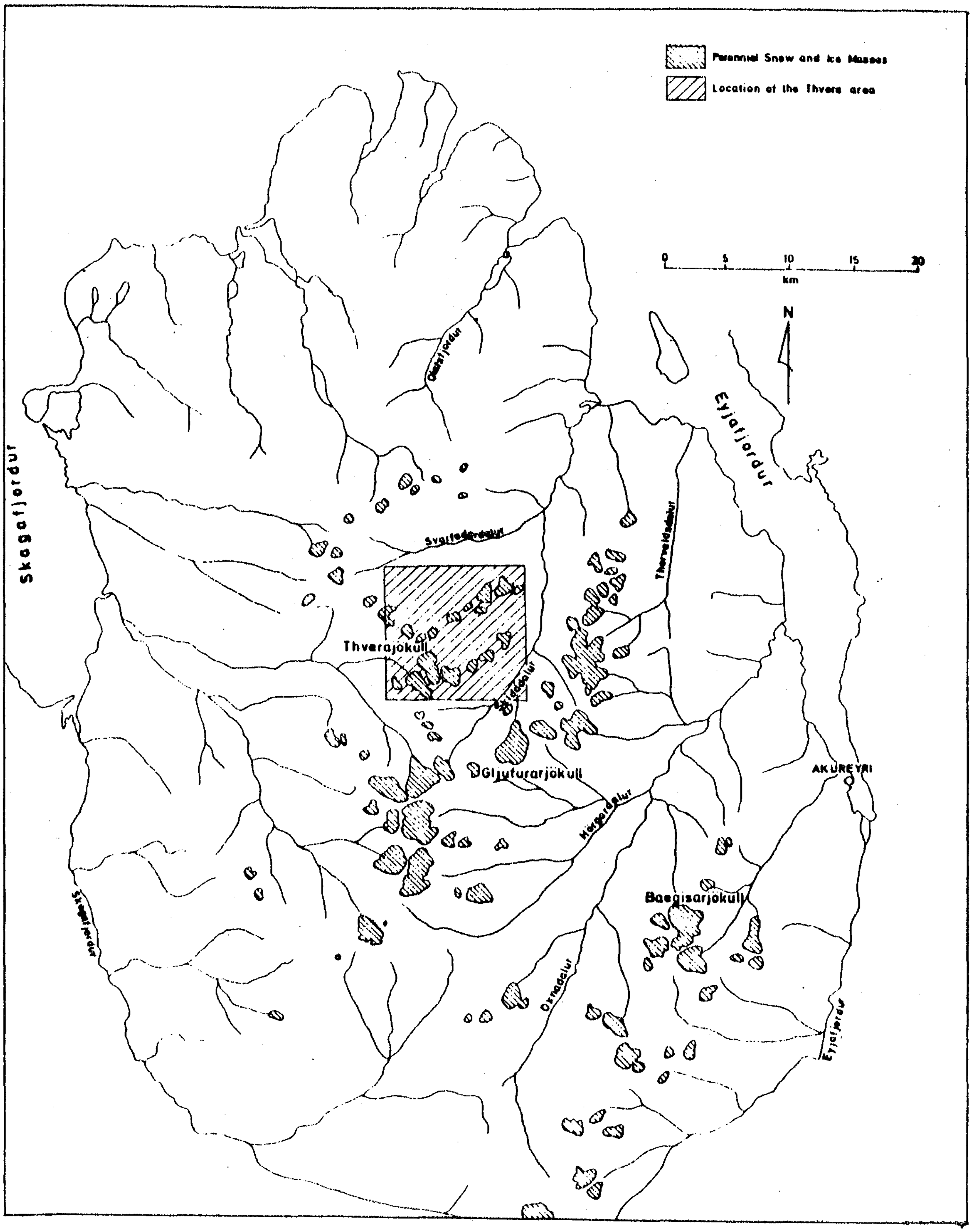

FIGURE 3.9
Local Glaciers of Tröllaskagi.

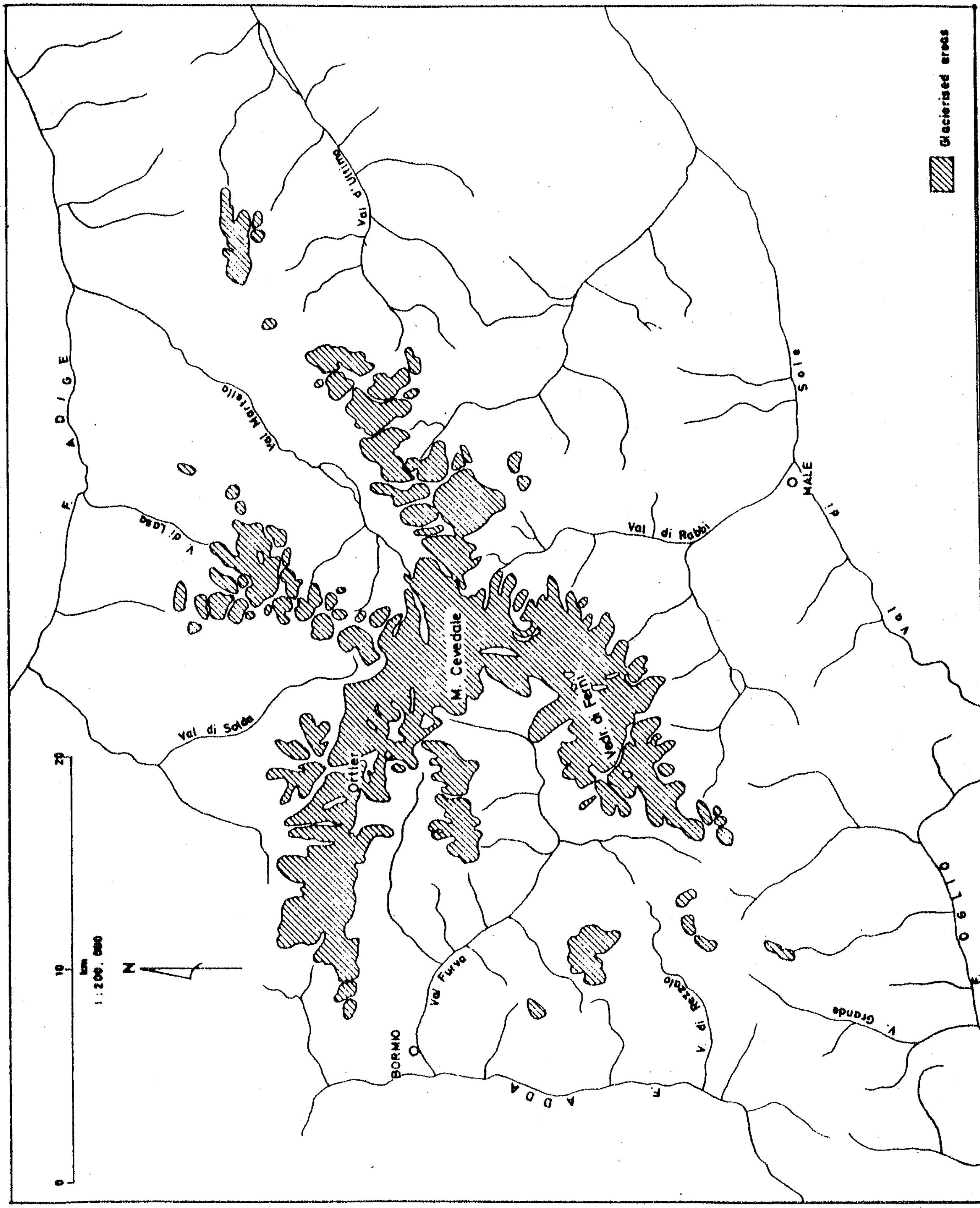

FIGURE 3.10
Glacierisation of the Ortles-Cevedale Group

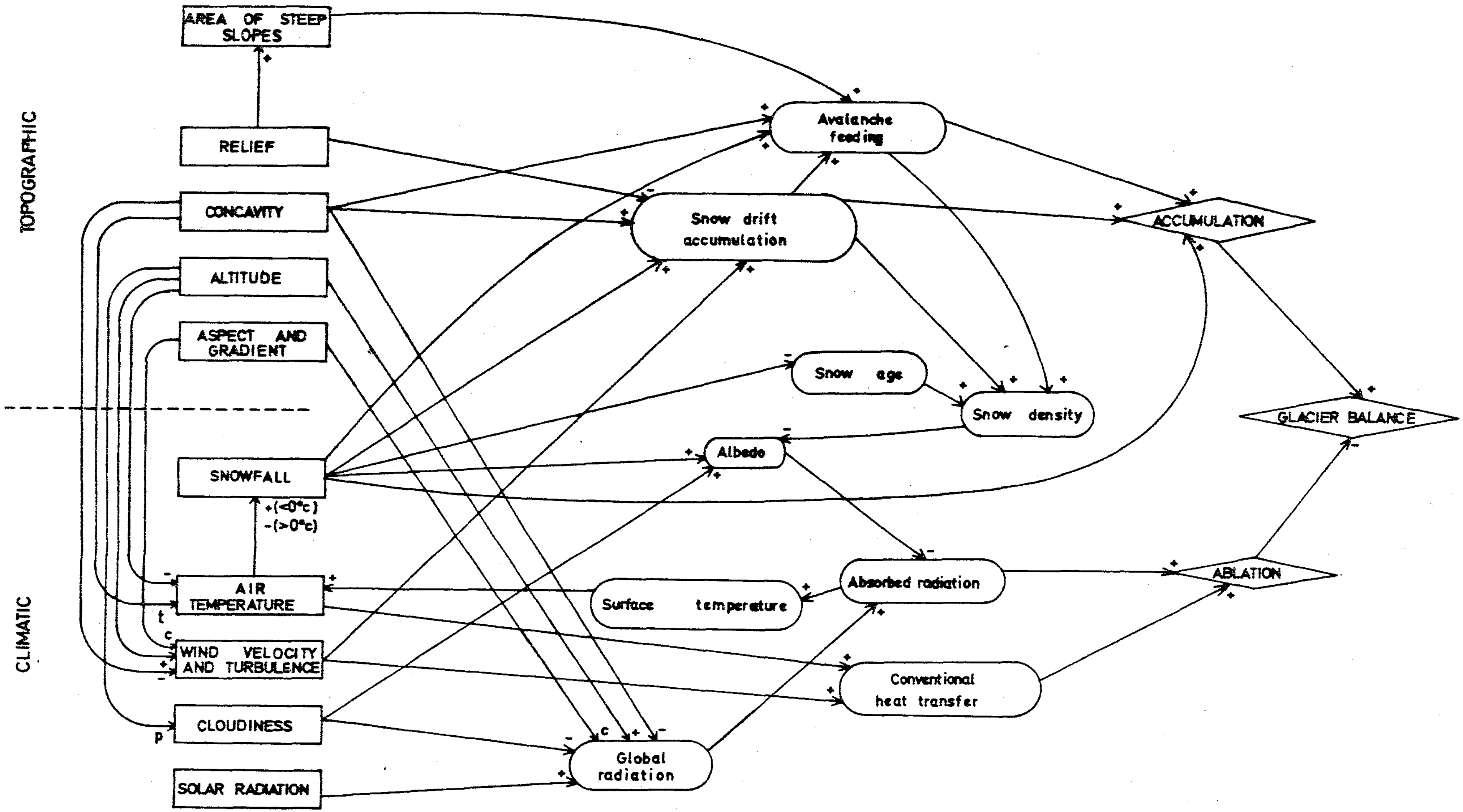

FIGURE 3.11
Interaction of Factors Affecting Glacier Mass Balance (modified after Evans, 1974 p.67)

Notes. c-circular relation. p-parabolic relation. t-thermal inversions.

CHAPTER IV

SIMULATION MODEL OUTLINE

As seen in chapter III, spatial variations in glacier mass balance are a complex function of the distribution and amount of snow accumulation, the nature and amount of incoming energy, and the utilisation of this energy at the glacier surface during the summer ablation period. Interactions between the relevant climatological and topographical factors, including variations in the mass balance processes, would appear to be best studied by means of computer simulation. A model of glacier mass balance simulation will be outlined in this chapter, after initial consideration of the nature and advantages of computer simulation.

4.1 Nature of Computer Simulation

Simulation techniques are being increasingly applied to a great variety of problems in many diverse disciplines. This has resulted in many different definitions of 'simulation' depending on the aims and nature of the study being undertaken. A simple and broad definition would be "Simulation is essentially a technique that involves setting up a model of a real situation and then performing experiments on the model" (Naylor et al., 1966 p.2). In this sense simulation is basically a working analogy and, as such, it includes all sorts of simulation exercises, like military war games, hardware scale models of beaches or river basins, wind tunnel experiments, as well as mathematical models.

The majority of simulation studies, however, use digital computers to study mathematical models which usually represent some active system involving the dynamic interaction of variables and parameters. In view of this, the above definition of simulation can be narrowed to the following:

> "Computer simulation is a numerical technique for conducting experiments on a digital computer; this technique involves certain types of mathematical and logical models that describe the behaviour of dynamic systems (or some component thereof) over periods of time". (adapted from Naylor et al., 1966 and Maisel and Gnugnoli, 1972).

Having defined the simulation technique, in the sense that it is used in this study, we should consider the type of problem for which it is most appropriate, with particular attention to the study of glacier balance. Computer simulation is used in cases where:

(i) it is either difficult or costly to observe certain processes in the real world. For example, in this study of glacier balance determination, the many problems involved with direct observation of glaciers coupled with the slow rates of glacier processes, (mentioned in sec. 1.2), make field study and physical experimentation on glaciers difficult.

(ii) the observed system may be so complex that both mathematical analysis and experimentation with prototypes may not be possible. Following on from the discussion in sec. 3.5, it is clear that the many factors involved in the determination of glacier balance and the poor expression in quantitative terms of their interrelations make analytical solutions impossible while, the construction of hardware models results in problems of scaling and in difficulties of controlling relevant variables. Hence in this case only numerical solution is possible.

Computer simulation permits the study of the complex interactions of a natural system in a systematic and controlled manner. It is, thus, ideally suited to a situation where an outcome depends on many variables and parameters, and involves numerous computation steps. This is the case with the simulation of glacier balance. It should be noted that simulation is normally used as a last resort after all other methods of solution have been shown to be inappropriate.

The use of computer simulation has several other advantages. It enables the study of a system in compressed time; thus slow-acting effects such as the glacier response to climatic change can be readily analysed. Furthermore, it enables the effects of certain modifications in the system model on system behaviour to be studied. Thus, in the present study, the effect of changing certain climatic parameters on glacier balance can be studied through computer simulation experiments. This ability to predict events under specified conditions of the system forms one of the most important advantages of computer simulation.

The development of a simulation model involves a synthesis of the knowledge about the real system and, as such, the "ability to accurately predict behaviour is a severe test of the adequacy of knowledge in any one subject" (Crawford and Linsley, 1966 p.2). There is always the possibility that the simulation methodology of imitating a system through a numerical abstraction may lead to a better understanding of the system, with the highlighting of those areas where data or theory are lacking, and the provision of ideas for further improving the simulation model.

The above advantages of using simulation are partly offset by the fact that a long time is usually involved in the development of complex simulation models and the process may turn out to be rather costly. A potential limitation is imposed by the great number of simplifying assumptions that are usually incorporated in the computer program and these could get hidden in the simulation process. As noted by Frenkiel and Goodall(eds.1978,p.12),"the power of simulation models lies in the complexity of the system they can examine; their weakness lies in the simplistic assumptions that may be used in some or all of the component parts".

Furthermore, being an experimental technique, simulation gives output only for the particular values of the independent variables and the system parameters used in that experiments. The functional relationships between the output and the independent variables and parameters, thus, have to be realised by numerous 'simulation runs' with one or more parameters set at new values.

The widespread application of simulation to a great variety of systems and the differing nature of simulation studies have resulted in many variations in the way a simulation study proceeds. A number of common procedural steps can, however, be identified. These are discussed in most standard references on computer simulation (e.g. Naylor et al., 1966). The basic steps are shown in the form of a flow chart in Fig. 4.1. It can be seen that computer simulation is concerned especially with the analysis of a physical system which is expressed as a collection of mathematical terms and parameters. This mathematical model is verified by simulating system behaviour with known input and output. If the simulated output does not agree with the observed system response changes are made in the mathematical representations.

This process is continued until the simulation model is judged to be an adequate representation of the physical system. The various steps involved in the simulation process are discussed at greater length with reference to the development of the glacier mass balance simulation model (secs. 4.3 - 4.6).

4.2 Previous Applications of Simulation in Glacier Studies

With the study of most natural systems being handicapped by problems of direct observation and experimentation, the complexity of controlling factors, and the slowness of process rates, computer simulation is often the only feasible technique available. The widespread use of computer simulation in earth - sciences as a whole, (e.g. Harbaugh and Bonham - Carter, 1970; Rayner, 1974; Fleming, 1975), is reflected by the increasing application of the technique to glacier-related problems. For example, considerable attention has been focused on the dynamics of glaciers and ice-sheets (e.g. reviews by Budd and Radok, 1971; Budd and Jenssen, 1975). This knowledge of ice dynamics has been used to simulate the flow of valley glaciers (e.g. Allison and Kruss, 1977) and, at a larger scale, the growth of ice sheets (Mahaffy, 1974). The latter model has been used to simulate the growth of the Laurentide ice sheet (Andrews and Mahaffy, 1976).

Further attempts at reconstructing the characteristics of former large scale ice masses include the models developed by Mahaffy (1976) for the Barnes ice cap in Baffin Island; Sugden (1977b) for the Laurentide ice sheet at its maximum; and Gordon (1979) for northern Scotland.

It can be seen that many of the numerical simulation models developed for glacier study involve large temporal and spatial dimensions. Relatively little attention has been directed at the simulation of local glaciers. Exceptions include the reconstruction of former glacier-climate conditions in the eastern Grampian Mountains by Sissons and Sutherland (1976); the mass balance simulations during an ablation season by Williams (1974, 1975); and the qualitative comments on the interaction between topography, climate and mountain glaciers by Glazyrin (1975). Williams' (1974) model, using an energy balance approach, is the most detailed attempt to simulate local glacier balance. However, lack of any consideration of snow drifting and avalanching, important processes in mountain glacier growth, means that the model is of limited application in determining the distribution of local glaciers.

A field of related interest to glacier balance determination, in which simulation models have been extensively used, is that of 'snowmelt hydrology' (sec. 2.3.6). Here computer simulation, with its ability to predict and experiment with different input conditions, has been found to be ideally suited, both as a runoff forecasting tool and as a means of improving the understanding of snowmelt processes. This has led to the development of numerous snowmelt routines which form important components of runoff models (sec. 8.1).

4.3 Characteristics of the Glacier Mass Balance Simulation Model

The nature of the glaciometeorological problem has been discussed in sec. 1.1, and from sec. 4.1 it is seen that such a problem is appropriately analysed by computer simulation techniques. Thus, it was decided that the question of local glacier distribution and its relation to climate and topography is best approached through the development of a deterministic simulation model which, with given inputs of local topography and climate, will locate areas of positive mass balance and local glacier generation. Such a model will also make it possible to identify and study the effects of the many different combinations of climatic parameters on glacier mass balance. The aims, approach and the scope of this study have been outlined in sec. 1.3. The succeeding sections in this chapter are concerned with the development of the mass balance simulation model in accordance with the procedural steps shown in Fig. 4.1.

4.3.1 System Description and Model Structure

Having defined the study problem, the next step is to collect and process real world data in order to develop a model of the system being studied. This involves a description of the system in terms of the relevant components, their relationships to one another, and the establishment of system boundaries.

A workable structure of the glacier mass balance system can be established from the analysis of published reports (chap. II), and through the identification of the relevant factors affecting local glacier balance (chap. III). A simplified glacier mass balance system, with the identification of its major components and relationships, visualised as a result of the interaction between exogenous topographic and climatic factors was presented in sec. 3.5.

Following the definition of the system, the next step is to model it. A model can be simply thought of as an 'abstraction of a real system that can be used for the purpose of studying the system'. Naylor et al., (1966 p.10) point out that in order to be useful a model must necessarily try to be both realistic and simple at the same time. They note that, on the one hand, it should serve as a reasonably close approximation to the real system and on the other hand, it must not be so complex that it is impossible to understand and manipulate.

A simplified structure of the glacier mass balance model is presented in the form of a flow chart in Fig. 4.2. The manner in which the various aspects of the mass balance have been modelled is discussed in the succeeding chapters; however, without delving into a detailed discussion of systems and modelling theory, a number of general considerations pertinent to the development of the mass balance model are discussed in the following sections.

4.3.2 Choice of Variables

The choice of which and how many variables to include in the model is based on the identification of those parts of the system judged to be important. This choice is further governed by such considerations as the particular spatial and temporal scales employed in the study, computer memory capacity, and the computer time available to complete a model 'run'. In the model structure a distinction is made between exogenous and endogenous variables. In particular, the exogenous variables representing the local topographical and climatological input need to be specified prior to mass balance simulation. The topographical variables relevant for glacier mass balance determination are defined in chap.V. Local climate is obtained from the interaction of local topography with regional climate (chap. VI). Endogenous variables, being the dependent or output variables of the system, consider the effects of the various accumulation and ablation processes (chaps. VII and VIII respectively) resulting in the net mass balance at the end of the simulation time period.

4.3.3 Nature of Relationships

In developing the simulation model it is useful to distinguish between the various types of relationships that are used to express the interactions between the model components. These range from deterministic to stochastic on the one hand and from empirical to theoretical on the other.

In the present study it is attempted to establish deterministic relations, as far as possible, in order to reduce model complexity. In trying to model the actual processes of mass balance, however, problems occur in those parts of the system where, due to a lack of quantitative information, it is difficult or impossible to quantify the variables and relationships affecting system behaviour. As noted in sec. 1.3, this is especially the case with analysing the areal variation of local climatic factors and in calculating the amounts of snow being drifted and avalanched. Furthermore, it is clear from the review of glacier studies in chap. II that much of the work on glacier climatology is empirical in nature, being based on mass and energy balance studies at selected studies on individual glaciers over limited durations of time. The consequence of this is that, although the direction of the mass balance relations, as portrayed in Fig. 3.11 may be known, the understanding of their quantitative effect is still vague.

This problem of developing a model in spite of gaps in the knowledge of the system is common to most studies concerned with the simulation of natural systems. In order to develop the simulation model, permitting progress in the understanding of the system as a whole, it is necessary to use 'empiricisms' for those parts which are relatively little understood. By 'empiricisms' I mean the practice of fitting a curve to observed data as opposed to rational application of dynamic principles to predict a pattern of behaviour.

Such empirical relations can be viewed statistically. There is the problem, of course, that the use of empirical relations unsupported by dimensional analysis may lead to risky extrapolation. However, they can be "used to bridge gaps in knowledge, pending confirmation or replacement, to avoid stalling progress in the development of the model". (Holtan, 1970 p.14). In this manner, by highlighting those aspects of the system that warrant special attention and from which parameters should be measured, even a first relatively crude simulation should help in the design of experiments necessary for better understanding of the system under study.

In the mass balance model, then, many of the relationships between model variables are based on rather vague and ill-defined empirical and semi-empirical relations. Inevitably some parts of the simulation model (e.g. energy balance calculations) will be more detailed than others (e.g. snow drift calculations) which are not adequately described mathematically. As knowledge about the system increases it will be possible to reduce this empiricism and express important processes by more fundamental physical relations.

A further simplifying characteristic of the mass balance model is the utilisation, in a deterministic fashion, of a number of important thresholds in the glacier balance system. For example, a temperature threshold is used to distinguish between the amounts of liquid and solid precipitation, a combination of topographical and climatological thresholds are used to distinguish those areas which will experience snow drifting and avalanching, a threshold of snow and ice thickness is used to determine when flow will occur on a given gradient, etc.

Consideration is also given to such feedback relations as the effect of changing albedo and surface roughness conditions (varying with changes in the state of glacial material from snow → firn → ice) in controlling the amount of energy available for the melting of glacial material. These threshold and feedback relations form an important aspect of the simulation study and are discussed in the appropriate sections of the succeeding chapters dealing with the modelling of the mass balance components.

4.4 Temporal and Spatial Scales of Study

The particular temporal and spatial scales adopted should form important considerations in the development of any simulation model. In attempting to model the system characteristics as closely as possible, the ultimate in modelling would be the utilisation of continuous time and infinitesimal space increments. This is, however, prevented by many practical limitations and, in any case, would make the model far too complex to be useful.

In general, model complexity is dependent on the magnitude of the temporal and spatial increments utilised, since with the application of large increments, effects of phenomena which change over relatively small increments of space and time can be neglected. Reducing temporal and spatial increments would require more detailed definition of the small-scale effects, thus leading to a further increase in the requirements of computer capacity and capability.

4.4.1 Representation of Time

An important consideration in simulation is the representation of time,since all processes are time-dependent. Variation of time may be treated in a continuous manner or, as is more usual, in a series of discrete steps. In reality most operations are continuous but it is necessary to break them into a sequence of steps to simulate them on a digital computer. In digital simulation, the passage of time (referred to as 'clock time') could be represented by using a critical-event approach (the system being viewed as proceeding from one event to another until a prescribed sequence of events is completed) or by an interval-oriented approach (the system being viewed as changing in all its aspects over time; its state being updated, usually in fixed time intervals until a prescribed amount of time has elapsed).

The passage of time in the mass balance model is represented by using the interval-oriented approach (Fig.4.3); the clock being advanced by one-month intervals. Consideration of the size of time-interval is important since this will affect the model output. An important constraint on the size of time-interval is imposed by the total computer time necessary to execute one 'run' (i.e. one movement of the system from the beginning to the end of the simulated time period).

In the case of glacier balance simulation, it may take 3-5 years for temperate ice formation and about 30-40 years for glacier flow to begin. Now with a time-interval of one month the mass balance model needs approximately 102 C.P.U. seconds (for a 4x8 km^2 area) and approximately 287 C.P.U. seconds (for a 10x10 km^2 area) to simulate the mass balance for one year.[1] Given these relatively large amounts of computer time needed to run the simulation model, reducing the time-interval further would prevent the use of the computer in terms of the cost and availability of computer time.

The use of monthly time-intervals for model input and mass balance computations also enables the seasonal variation of climatic elements to be taken into account (as monthly means of the basic climatic elements are usually readily available). Furthermore, by using monthly time-intervals the mass balance model keeps in line with the way in which field measurements of glacier mass balance are usually made. Thus the simulation can be started at the conventional break in summer/winter balance (i.e. around September/October in the northern hemisphere). The results of mass balance simulation can be obtained at monthly intervals or after any given number of months.

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[1]The computations being carried out on the N.U.M.A.C. (Northumbrian Universities Multiple Access Computers) linked system consisting of IBM 370/168 and 370/67; and, with minor modifications, on the IBM 370/148 system at Universiti Sains Malaysia.

4.4.2 Representation of Space

The analysis of spatial relationships forms an important aspect of the present study. A square-grid technique was used to divide the study area into a large number of grid squares. The study area can, thus, be considered as consisting of a matrix of squares which provide a convenient system for storing, processing and retrieving mass balance information. The various model inputs of climate and topography need to be defined for each grid-intersection. All model computations are also carried out with reference to the grid intersections, thus providing an objective basis for the examination of spatial variations in glacier mass balance.

The grid square technique has been widely used for the storing and analysis of physiographic, climatic and hydrologic data, (e.g. Solomon et al., 1968; Pentland and Cuthbert, 1971; Obedkoff, 1976), while Young (1976) has shown its suitability for glacier mass balance studies. The technique has been popularised for use in the spatial extrapolation of hydrometeorological data because of its adaptability for computer operations.[1] It further facilitates the tabulation of results and makes mapping procedures fast and simple.

A grid interval of 100m is used in the mass balance model. It is important to discuss the choice of this interval rather than any other. Ideally it is desirable, given the great variability of snow

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[1] Computer languages such as FORTRAN readily provide for the representation of grid locations in the form of two-dimensional arrays.

and ice cover in mountainous regions, to have as fine a grid spacing as possible. However, a number of factors restrict the choice of grid interval.

Firstly, the quality of the topographic maps determines the scale at which altitudes (which form the basic model input variable) can be read with confidence. Secondly, there is the question of computer memory capacity and time allocation. For a study area of, say 10 km^2, a grid interval of 100 m will involve mass balance computations at 10,000 grid intersections. With the use of many matrices necessary for defining the topographic and climatic inputs, the action of various accumulation and ablation processes, and to keep account of net mass balance changes at each grid intersection; the storage space needed approaches the maximum limit that can be conveniently sought from most computer systems. Reducing the size of grid interval, leading to an increase in the number of grid intersections and program running times, would make mass balance computations unmanageable given the limitations of computer time and storage capacity.

Thus the choice of a 100 m grid interval is based more on convenience for computation rather than any other criterion. However, bearing in mind the size of glaciers being considered and the meso-scale characteristics of the local glacier balance processes (sec. 3.5), it is believed that a grid interval of 100 m is suitable for the present study. Furthermore, Young (1974b, 1976) in relating mass balance measurements to terrain characteristics notes that many characteristics of the glacier surface are adequately described by using a grid spacing of 100 m. It would also appear that the use of a 100 m grid interval is in keeping with field measurement of mass balance, since Østrem and Stanley (1969, p.20) recommend that a density of 100 points per km^2 is desirable to measure snow depth variability.

Based on the above considerations regarding the nature of model variables, relationships and scales, it is seen that simulation models can be of several different types (Fig. 4.4). The particular type of simulation model developed will depend on the goal and purpose of the simulation study. The classificatory characteristics of the mass balance simulation model developed in the present study are defined by a broad line in Fig. 4.4.

4.5 Formulation of the Computer Program

Having developed the model of the real system, the next step is to construct a flow diagram showing the structure of the model interrelations. A simplified flow chart of the mass balance model, representing the various components and processes to be simulated, is presented in Fig. 4.2. This flow chart, outlining the logical sequence of the major events in the simulation study, facilitates the rather time-consuming task of writing the computer program.

Computer programming involves a number of considerations, the foremost of which is the choice of computer language. There are a wide variety of computer languages to choose from. These are conventionally divided into general purpose languages such as FORTRAN, ALGOL or COBOL; and special purpose simulation languages such as SPSS, SIMSCRIPT or GASP. The latter have been developed to simplify and facilitate the task of writing simulation programs for certain types of systems. These simulation languages differ considerably in the extent to which they can be applied to particular types of systems and the extent to which they ease programming difficulties. For a detailed comparison of various languages see, for example, Tocher (1965) or Teichroew and Lubin (1966).

The decision as to which programming language to use will depend on several factors of which knowledge and availability are the most critical. Most simulation programs are still written in FORTRAN, being the best known and readily available general-purpose language. Given this wide availability, flexibility to deal with simulation modelling, and the author's familiarity with the language, it was decided to adopt FORTRAN IV for programming the mass balance model.

4.5.1 Glacier Mass Balance Program (GSP1) Description

Program GSP1 comprises a main routine and 18 subroutines. Table 4.1 contains the names and functions of the subroutines.

The program listing is contained in Appendix A.

Main routine. The main routine (Fig. 4.5) provides overall control of the computational procedures simulating mass balance. It is used to:

- initialise the system.
- define and provide storage space for the various arrays containing the mass balance variables.
- read input topography and mass balance cover data.
- call up appropriate subroutines modelling the processes of snow and ice accumulation, ablation and flow.

- compute necessary densification and net mass balance changes for snow, firn and ice.
- provide output data, revealing mass balance distribution over the altitude matrix, in a suitable form for input to line-printer mapping programs.

Computing notes: (refer to program listing in Appendix A).

(i) Size of matrix

GSP1 has been written to compute the mass balance variations over a 40x80 rectangular matrix (with 100 m grid spacing). A study area of any other size can be modelled by changing the storage space allocated for the various arrays (as defined in the COMMON and DIMENSION statements).

(ii) List of arrays

The various arrays used in the main routine are defined in Table 4.2.

(iii) Period of simulation

The length of time simulated in any one computer run depends on the stages at which mass balance output is required and the amount of computer time available for running a program at any one time. As noted in sec. 4.4.1, GSP1 requires about 102 C.P.U. seconds to simulate mass balance for 1 year (for a 4x8 km^2 area). It was found convenient to simulate in 5-year periods, (using approximately 510 C.P.U. seconds per run), to trace the development of mass balance simulation.

(iv) Input/Output characteristics (Fig. 4.6)

The various logical input/output units utilised in GSP1 are listed below. The output units (3, 6 and 7), denoted as optional, are particularly useful in the development stages of the program.

1 - Topographical input data. This is obtained by use of the terrain analysis program - OMY8. This program and the nature of the topographic input are discussed in chap.V. To save on computer memory space, the input is read in an unformatted form by use of the program TPGINP1 (listed in Appendix B1).

The unformatted version constituted a saving of 67-71% on disk storage compared with the formatted version.

5 - Period of simulation. This input unit is used to read in the month in which the simulation is to be started (A1) and finished (A2) in a particular computer run.

2 - Initial mass balance cover. This is used to specify the state of the accumulation cover at the start of the simulation. The water equivalent and depths of snow, firn and ice and the total water equivalent and depth are read in for each grid intersection, together with the calendar month (X) to which these values refer. The data are unformatted by use of program SIMINP1 (listed in Appendix B2).

3 - Accumulation state map (optional). A simple line-printer map can be drawn at the end of each month's simulation to show the points at which there is snow, firn, ice or no accumulation.

6 - Simulation progress checks (optional). A check on the progress of the simulation program can be kept by printing out messages after the completion of each major task in the program.

7 - Output of mass balance values at specified intervals (optional). This unit is used to print out values of mass balance variables (above a critical thickness of water equivalent) at specified intervals (in any combination of months). Such output is in a form suitable to draw line-printer maps showing the variation of mass balance cover at regular intervals (e.g. every five years).

4 - Mass balance cover at the end of the simulation period. The unformatted values of the water equivalent and depths of snow, firn and ice, total water equivalent and depth at each grid intersection are printed to specify the state of the accumulation cover at the end of the simulation run. If the simulation is to be continued further these values would form the initial mass balance cover, being the input through the I/O unit numbered 2.

(v) Display of simulation results

Simulation studies usually produce a great volume of data to show the variation of the modelled components in time and space. The display of this information, thus, becomes an important consideration. The results of GSP1 are displayed by using a series of variable density line-printer maps (discussed in chap. V). Simulation output data to print these maps can be obtained from GSP1 (using I/O unit 7) at the end of any month's simulation, thus permitting the analysis of mass balance results before proceeding with the simulation.

(vi) Action of program subroutines

The nature of the various mass balance subroutines (listed in Table 4.1), is discussed in the relevant sections dealing with the modelling of glacier mass balance components in the succeeding chapters.

Having written the computer program the relevant input data, regarding the initial state of the system, are provided and the first simulation run can be made (chap. IX). If the program runs successfully the output can be analysed and one of the most important and difficult stages of the simulation process is reached, that of model validation.

4.6 Model Validation and Simulation Experiments

Validation is a measure of the extent to which the model satisfies its design objectives. It usually involves comparing simulated model output with corresponding known and reliable field data about the real system. Such calibration and testing of the model with a comprehensive range of environmental data is necessary for its use in prediction. If the simulated output does not compare very closely with the real world data, then, those parts of the model which may be causing the discrepancies are modified and the program re-run. This process of trial-and-error is continued until the simulation model is judged to be an adequate representation of the real system. Though this validation procedure might appear to be straight-forward enough, for most physical systems it can be a very difficult one, involving practical, theoretical, statistical and philosophical complexities (Naylor, et al. 1966).

A major problem regarding the study of natural systems is that suitable data, against which the simulated results can be compared, may simply not be available. Given the problems associated with the choice, observation and measurement of relevant variables describing the behaviour of natural systems, the data gathered about such systems tends to be sparse, unreliable and unrepresentative of the system variability. There is the additional problem of equifinality, in that any given form in the real world may result from a range of possible causes.

Furthermore, the lack of quantitative understanding of natural systems leads to the use of inevitable simplifications, through the inclusion of experimental, empirical and ad hoc assumptions, to assist in defining the necessary system relationships (sec. 4.3.3) making objective comparison of model output with real-world data even more difficult. The resulting interpretation problems of simulation models are discussed by Howard (1971) with reference to the modelling of geologic and geomorphic systems. He points to the need to exercise extreme caution in the interpretation of such simulation models since, in the case of modelling most natural systems, the validation procedure has to rely on a rather qualitative comparison of simulated and real-world data, to see if the results "look right" (Harbaugh and Bonham - Carter, 1970 p.34).

In view of the above comments the validation procedures for the mass balance model are rather limited. Lack of quantitative data about certain aspects of the system, such as the variability of climatic elements, drifting, avalanching, densification rates, and parts of the energy balance computations, means that the model must be calibrated with whatever observed data available for net mass balance variations. Model validation, has thus been restricted to the following:

(i) Individual calibration of those model routines for which some form of quantitative data are available (e.g. topographic parameters, radiation balance).

(ii) Comparison of the snow cover distribution for an area in north-west Iceland, as mapped from aerial photographs (July 1959), with that simulated by the mass balance model using real climatic data for that particular year.

A similar comparison of the major snowpatches found in the area at the end of the ablation season, as determined by field mapping (August 1978) with the simulated snow cover for the same period, using the appropriate climatic input for that year.

(iii) Simulation of local glacier generation sites for a marginally glacierised region, in north-central Iceland, using suitable climatic input values thought to be representative of the climate prevailing during the last period of glacier development. The simulated sites are compared with the actual glacier distribution in the region as determined from aerial photographs and field observations.

The above validation procedures, though far from satisfactory, do provide some general indication of the validity and usefulness of the mass balance model. The areas in which the simulation model is tested are defined in sec. 4.7, while the mass balance simulations for the validation analysis are described in chap. IX.

The final stage in the simulation study, after the computer model has been shown to be valid, is to design and run simulation experiments to observe the behaviour of the system under different inputs or changed parameters. With regard to the mass balance - climate relations, obviously an infinite number of combinations with various climatic parameters could be tried to see their effect on mass balance. Since one of the main concerns of glacier-climate studies (chap. II) has been to determine the effect of changes in temperature and precipitation on glacier balance, a number of simulation runs were made with the mass balance model using different combinations of these climatic parameters - e.g. changes in annual temperature, 'summer' temperature, 'winter' temperature and precipitation. The results of these simulation experiments are discussed in chap. IX.

4.7 Sample Areas for Model Application

The mass balance model was developed and tested with reference to two sample areas in northern Iceland, viz. Nupur and Thvera, located in the Vestfirðir and Tröllaskagi districts respectively (described in secs. 3.2.1 and 3.2.2).

The Núpur area lies in the western part of Vestfirðir on the northern side of Dýrafjörður (Fig. 4.7). A plot of 8 km^2 was selected to examine the mass balance variations (Fig. 4.8). This area forms part of the 'alpine topography' found in western Vestfirðir (Fig. 3.5) and is characterised by the development of several steep-sided valleys (e.g. Nupsdalur, Tungudalur, Hgarðardalur and Galtardalur) and numerous well defined cirques dissecting a gently undulating plateaux (Figs. 3.1, 3.2 and 3.3). The cirque floors lie at an average of about 360 metres above sea level, while the plateau remnants compose , for the most part, extensive blockfields (Fig. 3.4) at altitudes of 600 - 800 metres.

Such a deeply dissected topography would be expected to exert a strong influence on mass balance variations. At present the snow and ice cover in this region is characterised by the development of local snowpatches (Figs. 3.6 and 3.7); however, the effect of any climatic deterioration resulting in local glacier generation can be examined through simulation experiments (chap. IX).

In order to reduce the number of computations necessary in the model and permit a quick portrayal of the action of various mass balance subroutines, the effects of the latter are examined with reference to the central half of the Nupur area (Fig. 4.8). This 8x4 km^2 area is referred to as the 'test matrix'.

The Thvera area lies in the central part of Tröllaskagi about 40 km north-west of Akureyri (Fig. 3.9). The mass balance model is applied to a plot, 10 km^2 in area, characterized by highly rugged and irregular terrain (Fig. 4.9). In contrast to the Nupur area, altitudes are higher (the mountain peaks being over 1200 m) and the plateau remnants are very restricted. Furthermore the region is marginally glacierised with the development of numerous local glaciers.

In addition to the suitability of their topography for developing the mass balance model, the choice of these two sample areas was influenced by three other considerations:

(i) the availability of good photogrammetric maps at a scale of 1:50,000 with a contour interval of 20 m prepared by the American Army Service in 1949 (Figs. 4.8 and 4.9).

(ii) aerial photographs at a scale of about 1:36,000 taken by the U.S.A.F. during 1957 - 1960 (Figs. 4.10 and 4.11).

(iii) field experience gained as a result of visits to the Nupur area in 1976 and 1978 (with members of the Durham University Vestfirđir Project) and to Tröllaskagi in 1978 (with members of the Science Institute at the University of Iceland).

TABLE 4.1

GSP1 : List of Subroutines and Functions

Subroutine		Function
		For each grid intersection of the study area matrix the various subroutines determine:
MTEMP	:	monthly mean temperature
MSNOW	:	monthly mean snow cover
MWIND	:	monthly mean wind speed
WEXP TPSW	:	'exposure' as due to surrounding topography (for use in wind speed calculations).
MDAVAL PLAV	:	areas (and amounts) from which snow avalanching on to potential drifting situations will take place.
MDRIFT DEP DRWIND	:	monthly redistribution of snow cover due to snow drifting.
MAVAL AVDIR	:	monthly redistribution of snow cover due to snow avalanching.
MELTPR	:	potential 'melt' periods.
SWRAD GLSR SHADE (optional)	:	monthly short-wave radiation balance (effect of shading due to surrounding topography can also be taken into account).
LWRD	:	monthly long-wave radiation balance.
TBHT	:	monthly heat exchange due to turbulent heat transfer (both sensible and latent heat).
FLOW (optional)	:	whether a critical thickness of ice has accumulated for flow to take place.

TABLE 4.2

GSP1: List of Arrays Used in the Main Routine

Topography :	Altitude (ALT) Aspect (ASP) Slope (SLOP) Profile convexity (PROFC) Plan convexity (PLANC)
Climate :	Initial snow cover (SNOW) Initial snow cover after drifting (SNOW1) Initial snow cover after avalanching (SNOW2) Temperature (TEMP) Wind (WIND)(WRED)
Melt Periods :	Number of hours per day (DMHRS) Mean Temperature (TEMPML) Number of days (DAYSML)
Energy Balance :	Global radiation (GRBL) Longwave radiation (XLWR) Total radiation balance (TRBL) Sensible heat (SBHT) Latent heat (TLHT) Net energy balance (ENGBL)
State of Accumulation Cover :	Density of snow (DSNOW) Density of firn (DFIRN) Density of ice (DICE) Water equivalent of snow (WSNOW) Water equivalent of firn (WFIRN) Water equivalent of ice (WICE) Total water equivalent (TWEQ) Total depth (TDEPTH)
Flow :	Thickness of ice (THIC) Critical thickness of flow (CRIT)

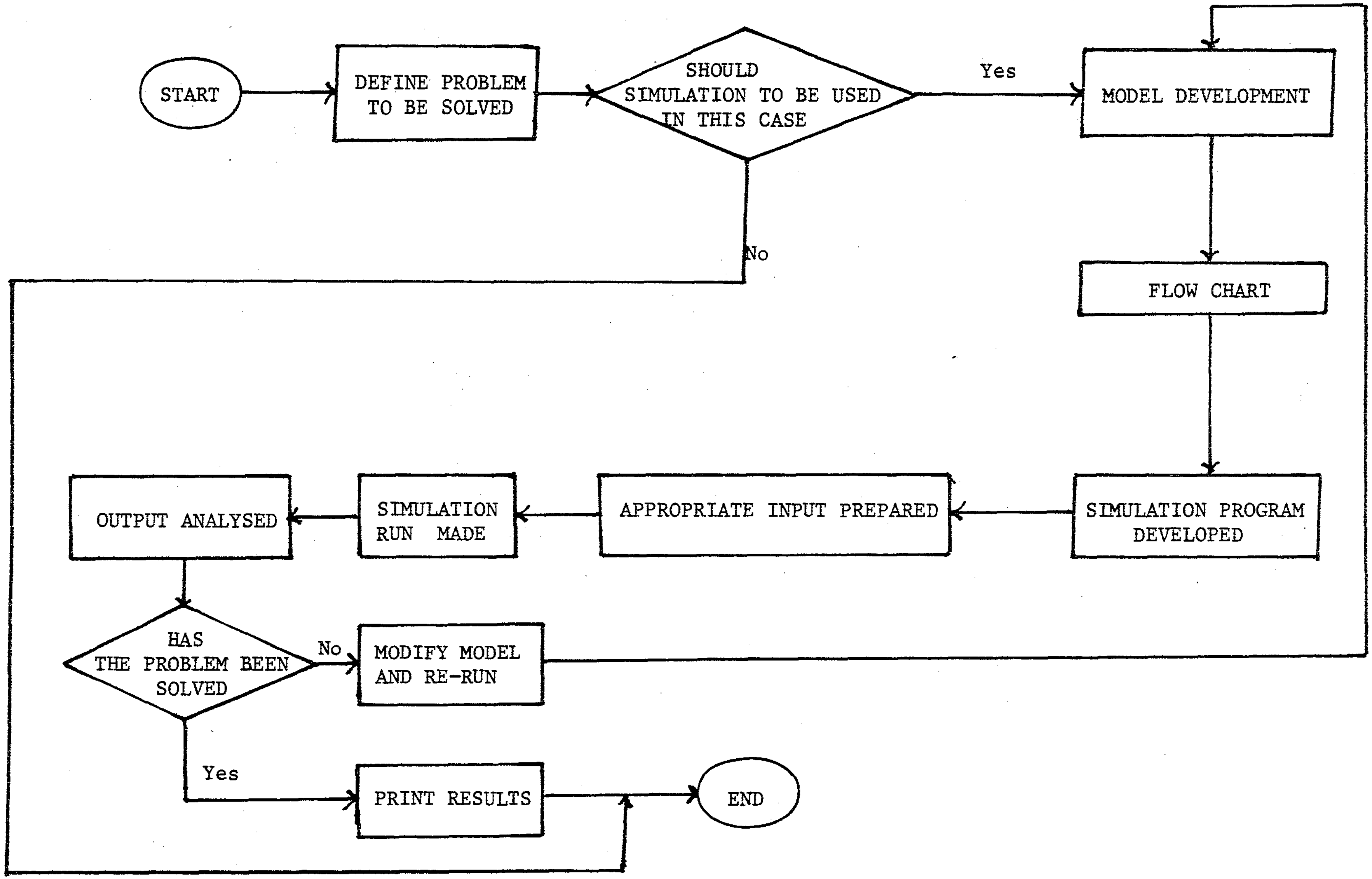


FIGURE 4.1

Flow Chart Describing the Use of Simulation

Exogenous Input Variables

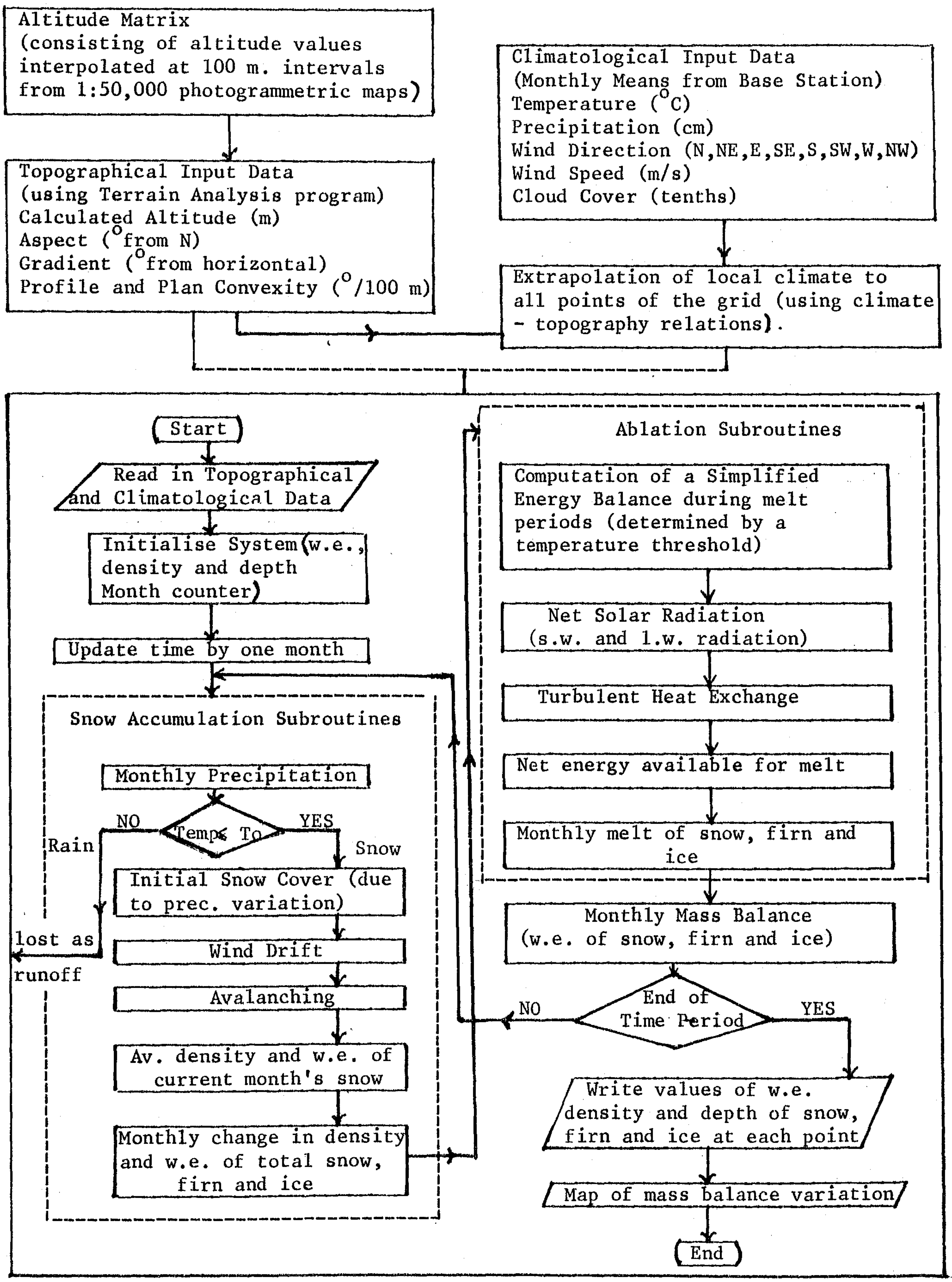


FIGURE 4.2

Simplified Flow Chart of Glacier Mass Balance

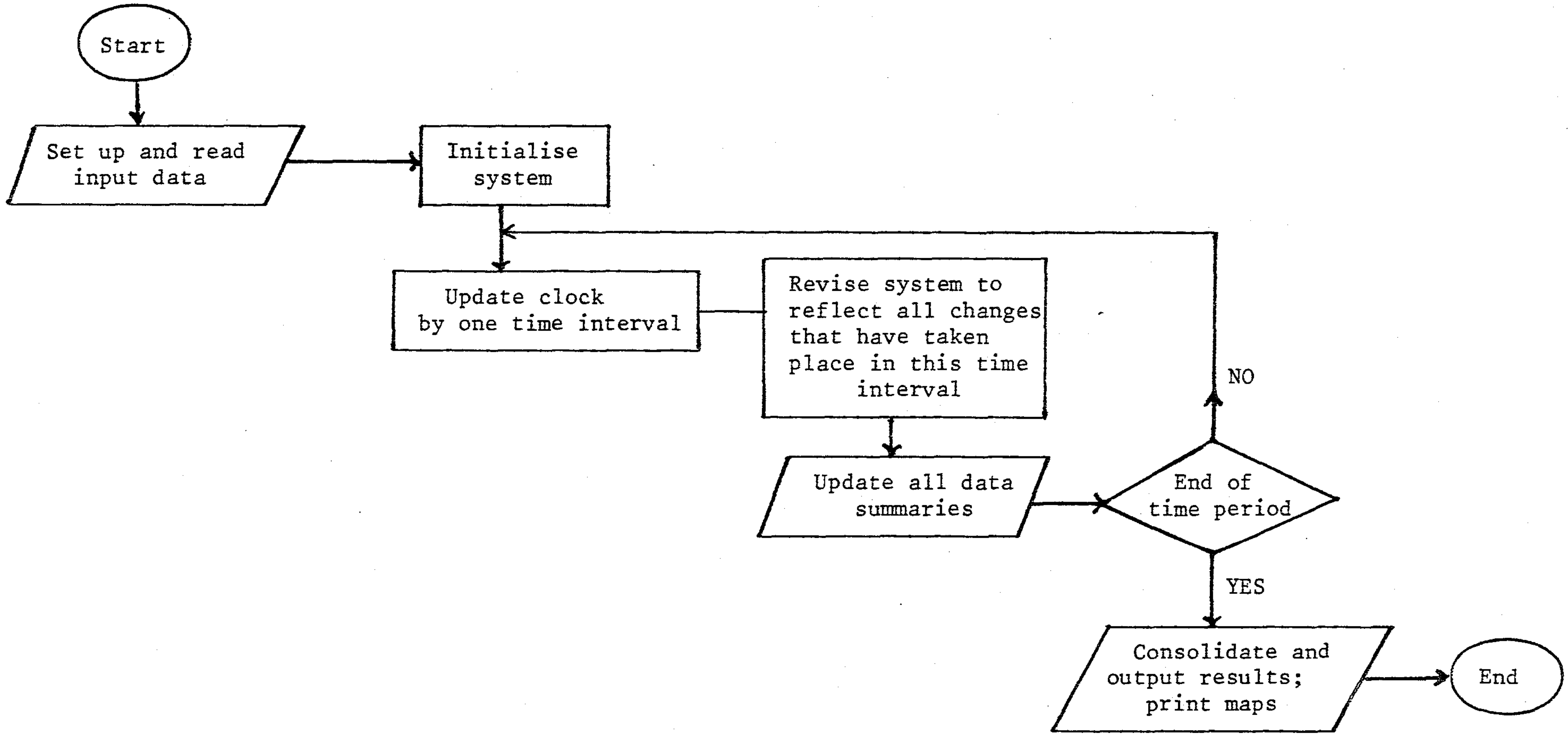


FIGURE 4.3

Simulation Using the Interval-Oriented Approach

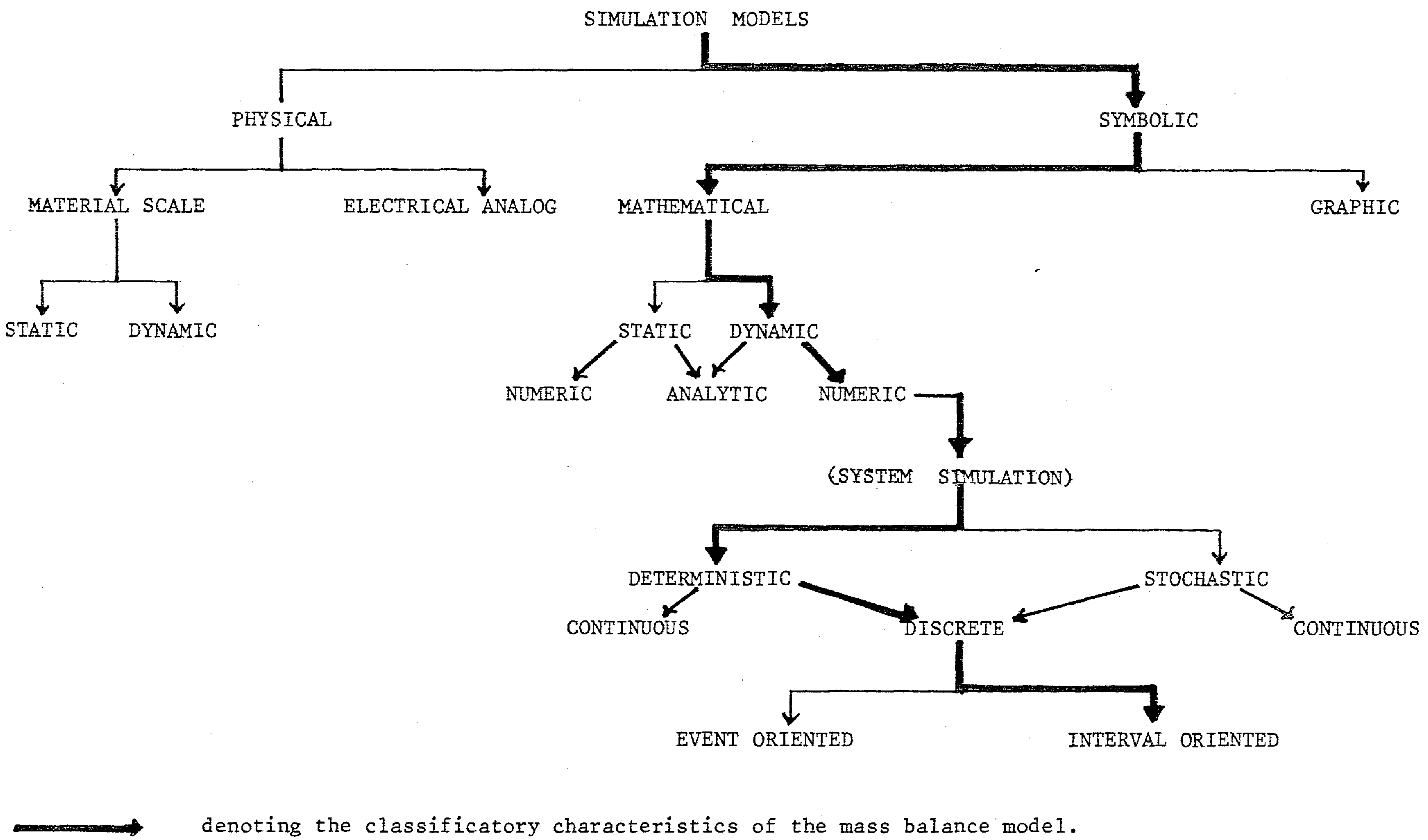


FIGURE 4.4

Classification of Simulation Models

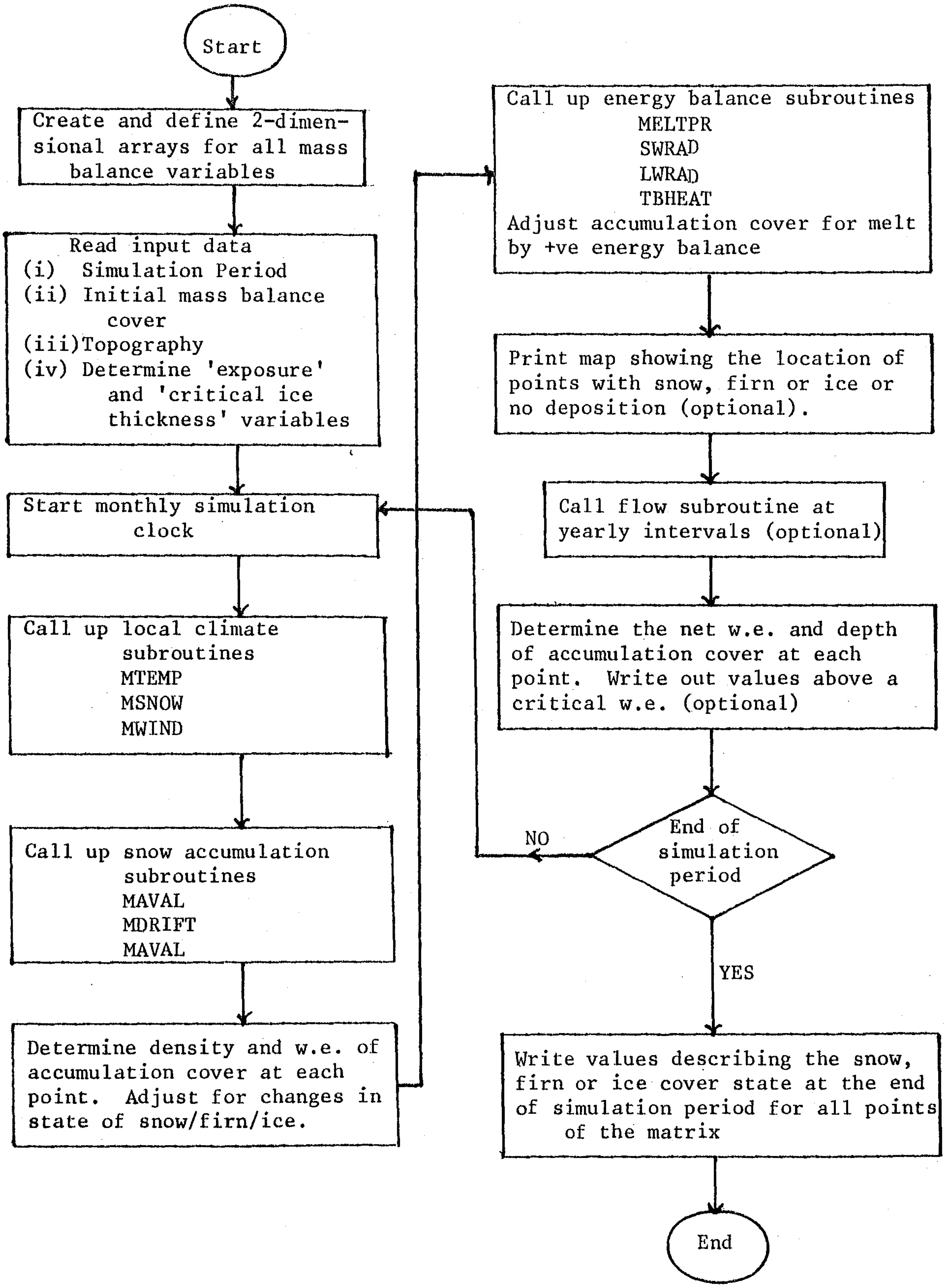


FIGURE 4.5

Generalised Flow Chart of the Main Routine in Program GSP1

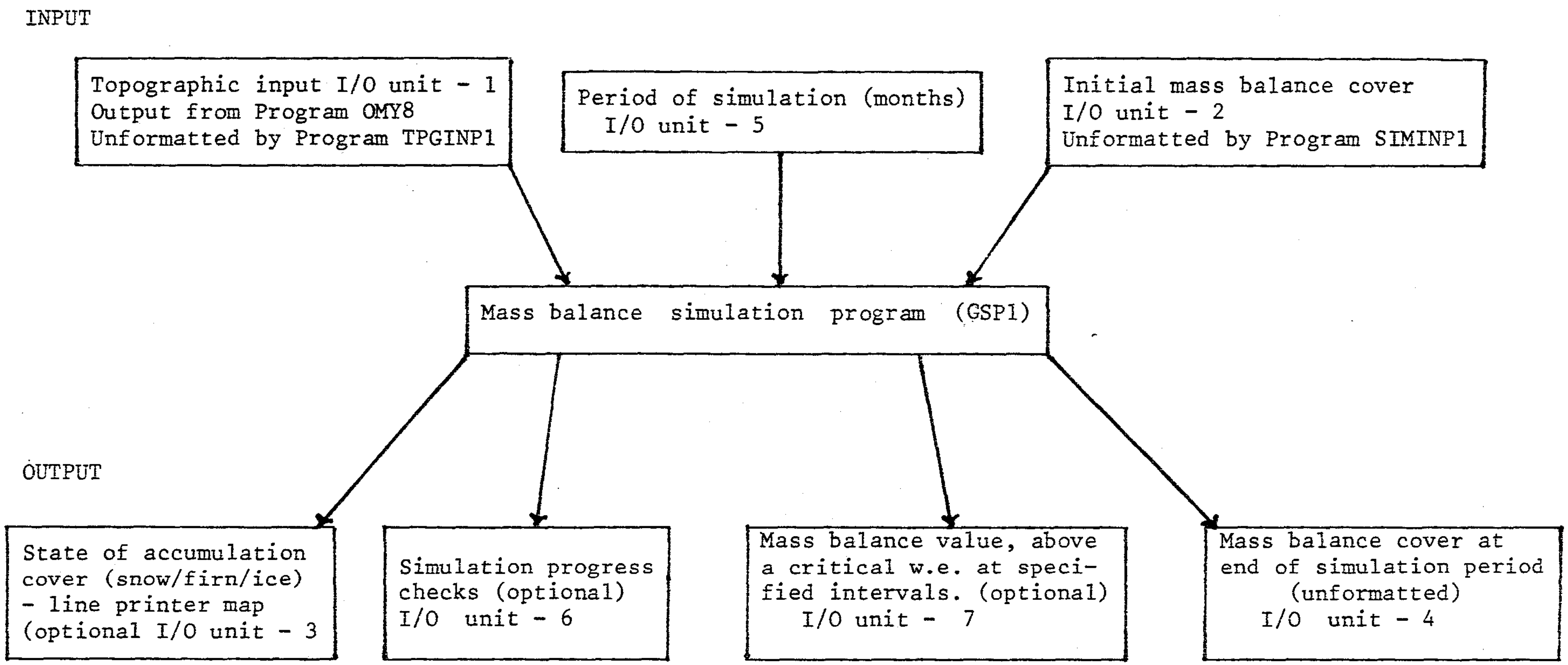


FIGURE 4.6

Input/Output Units Used in GSP1

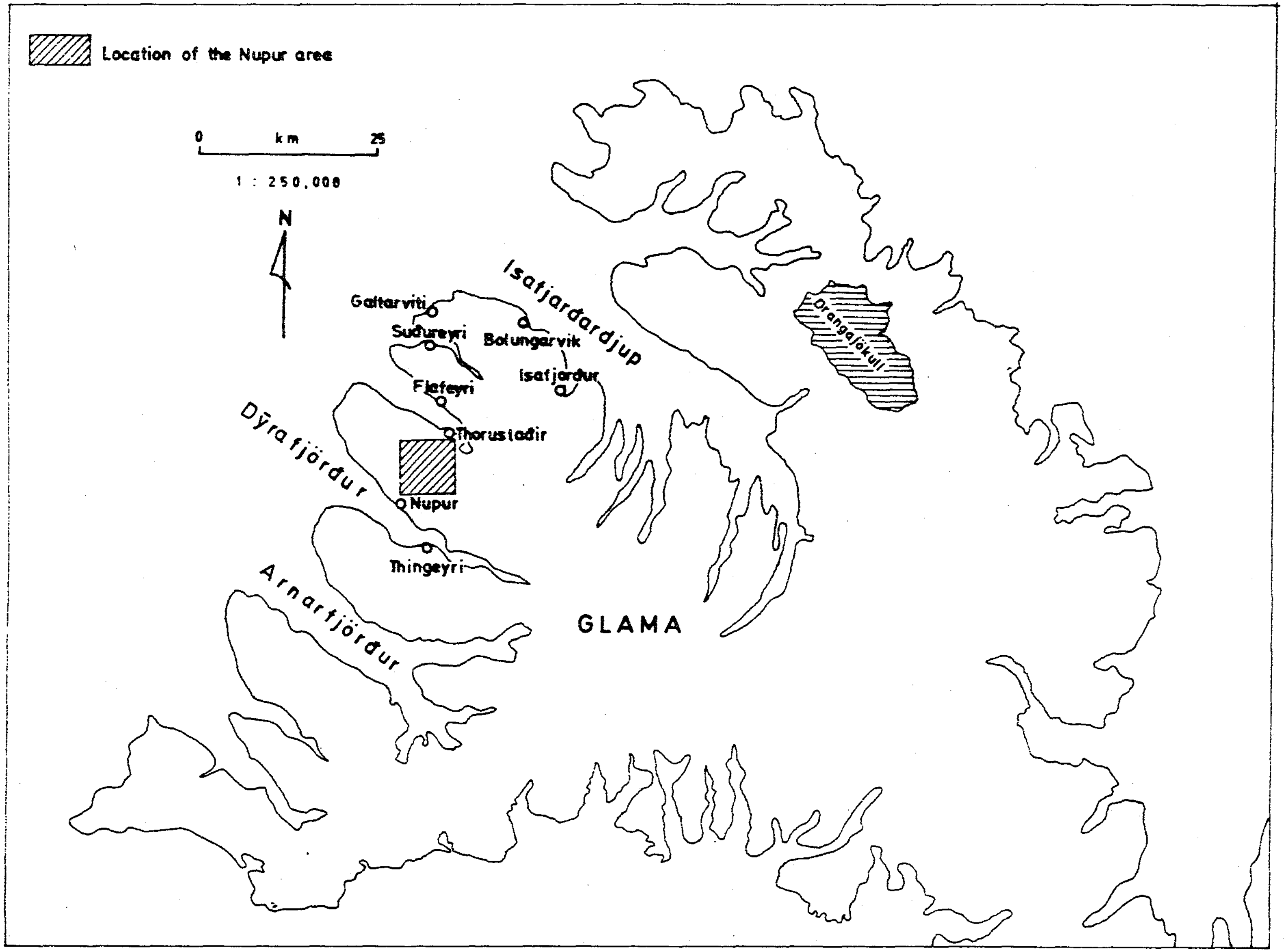
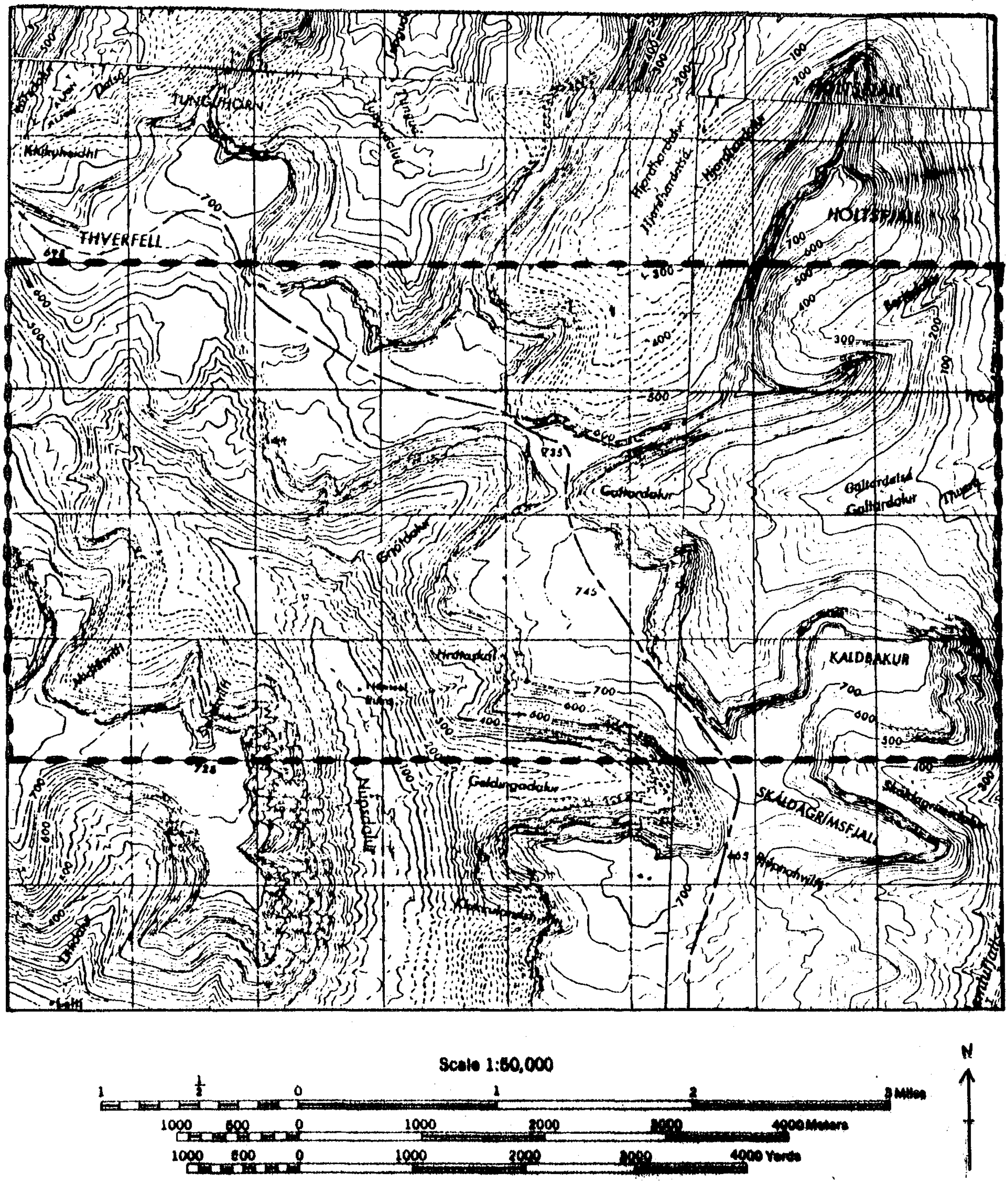


FIGURE 4.7
Location of the Nupur Area in Vestfirđir

NÚPUR



CONTOUR INTERVAL 20 METERS

FIGURE 4.8
The Nupur Area (Central half of the area is defined by the dashed line).

THVERÁ

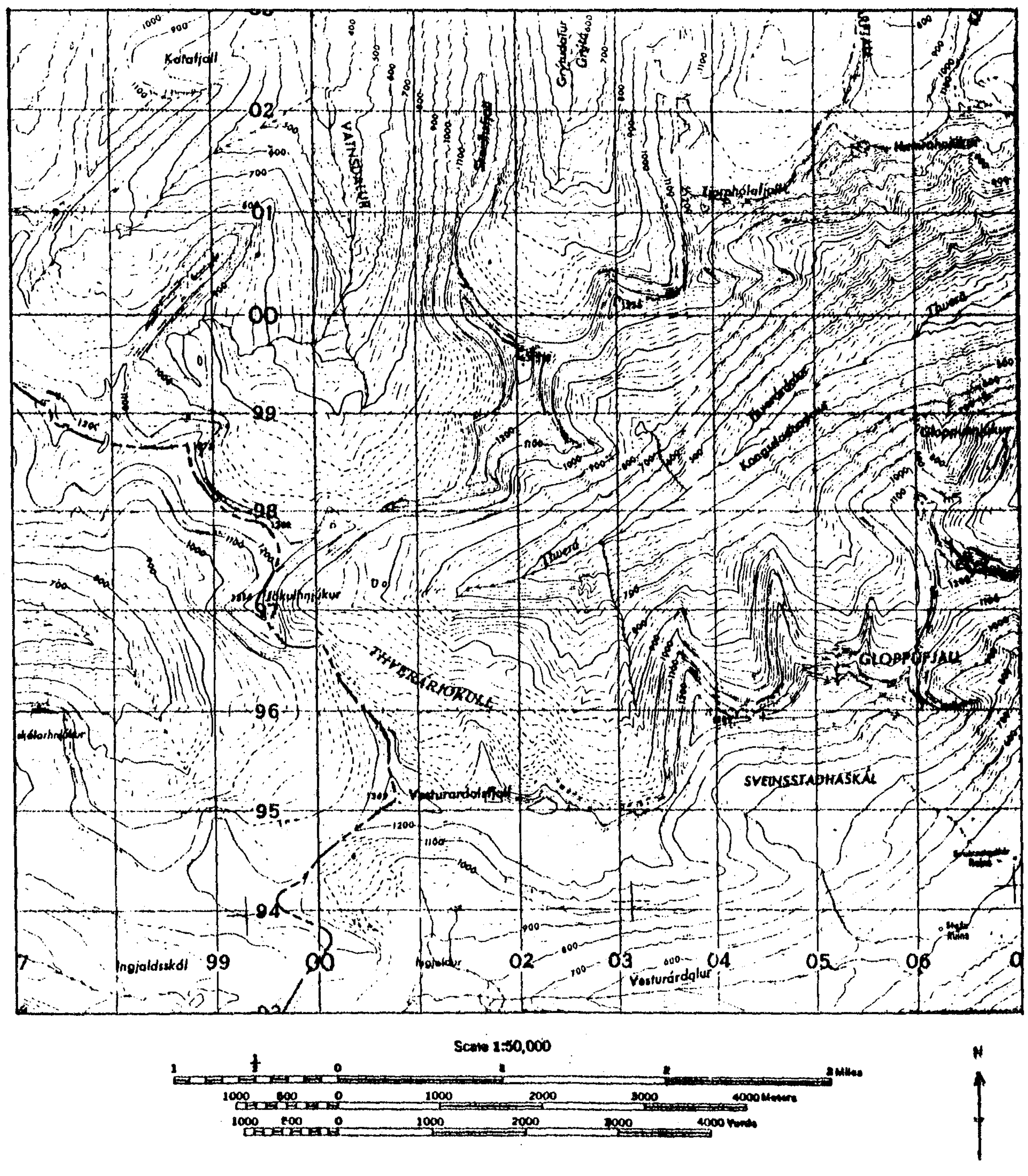


FIGURE 4.9
The Thvera Area

FIGURE 4.10
Aerial Coverage of the Nupur Area

FIGURE 4.11
Aerial Coverage for the Eastern Part of the Thvera Matrix (Note position of Gljúfurárjökull).

CHAPTER V

TOPOGRAPHIC INPUT

The preoccupation of glaciologists and glacial geomorphologists with the question of the link between glaciers and climate (chap. II) has rather overshadowed the critical role of topography for local glacier generation. The importance of the latter is evident from studies undertaken to determine the factors affecting local glacier distribution (sec. 3.1). Furthermore the modification of local climate by topography (chap. VI) and the close control of snow accumulation and ablation processes (chaps. VII and VIII) by topographic parameters illustrates the importance of both direct and indirect effects of local topography on glacier balance. An accurate assessment of local topography is, thus, essential to the success of the modelling approach adopted in this study. This chapter is concerned with a discussion of the manner in which the various topographic parameters have been derived.

5.1 Choice of Topographic Parameters

Discussion relating to the nature of the topographic influence on glacier balance and local climate (chaps. III, VI, VII and VIII) reveals that surface altitude, gradient, slope aspect, and a measure of local convexity are the significant topographic parameters that warrant consideration in a model of glacier mass balance. In a wider context, such topographic measures are commonly used in general geomorphometry[1] as a basis for the quantitative comparison of landscapes.

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[1]being the "measurement and analysis of those characteristics of landforms which are applicable to any continuous rough surface" (Evans, 1972 p.18).

For example, Evans (1972), following a review of previous approaches to general geomorphometry, suggests that the basic parameters of general geomorphometry can be defined in terms of altitude and its derivatives. These are slope gradient and aspect (being the first vertical and horizontal derivatives of the altitude surface), profile (downslope) and plan (cross-slope) convexity (the second vertical and horizontal derivatives).

Gradient is commonly defined as the angle (varying between 0 and 90 degrees) between the horizontal and a plane tangent to the surface at any given point. Combined with aspect, expressed as degrees (0-360) measured clockwise from north, it defines the slope at a point. Surface curvature, denoted by the term 'convexity' (concavity being treated as a negative value of convexity), is divided into profile and plan convexity. Profile convexity refers to the rate of change of gradient while plan convexity is the rate of change of aspect (i.e. the degree of contour curvature). Both components of convexity are defined in degrees per 100 m.

The usefulness of these variables in defining particular elements of the landscape of interest to glacier mass balance computation is illustrated in sec. 5.6, while, their general importance in geomorphology and related fields has been discussed by Evans (1979). Given their widespread application and relevance in controlling glacier mass balance processes, these five variables have been chosen to provide the topographic input for the mass balance model.

Following the selection of the relevant topographic parameters, the next step is to devise a suitable terrain analysis program to generate values for these parameters, at a suitable scale, for each sample point in the study area. Various terrain analysis programs have been developed which define suitable topographic parameters from an input of a regularly spaced array of altitudes i.e. an 'altitude matrix' (e.g. Young, 1973; Grender 1976). The system used in this study was developed at Durham University by Evans and his collaborators. It is the outcome of a research project designed to "implement and evaluate techniques for the statistical characterization of altitude matrices by computer" and is described in a report entitled "An Integrated System of Terrain Analysis and Slope Mapping" (Evans, 1979).

5.2 Construction of Altitude Matrix

As noted in sec. 4.4.2 the most suitable method of representing space for computer manipulation is the square grid technique. The terrain analysis program, thus, has been designed to accept a regular array of altitudes as input data. Such altitude matrices can be directly generated from aerial photos in stereoplotters or indirectly from digitised contours (using an interpolation program). In the absence of such facilities, they can be constructed manually by visual interpolation from accurate contour maps.

The altitude matrices for the two Icelandic sample areas of interest in glacier mass balance simulation, viz. Nupur (Fig. 4.8) and Thvera (Fig. 4.9), were generated manually by laying a square grid with an interval of 100 m over 1:50,000 photogrammetric contour maps. Altitudes were then interpolated at all grid intersections. For the Nupur area, the matrix consists of 82 rows and 82 columns (resulting in topographic parameters being generated over 80x80 points since the peripheral rows and columns are not considered). The slightly larger Thvera area has a matrix size of 102x102 points.

The size of grid mesh used forms an important consideration for the calculation of topographic parameters (sec. 5.5). While the terrain analysis program can accept altitude matrices of any grid mesh, a grid interval of 100 m was used for the sample matrices for reasons of computer manipulation and map accuracy (sec. 4.4.2). The method of manual interpolation is subject to errors of interpolation by the operator and those associated with the construction of the original map. In the case of the sample matrices, given the close parallel pattern of contours, it is thought that interpolation errors are within $\pm$ 5 m (for unglacierised regions).

5.2.1 Ice Thickness Approximations in Glacierised Areas

In the Thvera area the interpolation of altitudes over the glacierised areas (Fig. 4.11) presented a problem since the topography beneath the ice cover had to be estimated. In the absence of any ice thickness data for glaciers in this region, a rough approximation of glacier thickness was obtained from relationships developed between glacier area and glacier thickness in areas of mountain glacierisation. For example, Ommanney (1969, p.46), in compiling the inventory of glaciers in Axel Heiberg Island presented the results of area and mean glacier depth for a number of different glacier types. The expected mean depths for small mountain glaciers are 15 m (for a glacier 0-1 km^2 in area); 20 m (1-2 km^2); 35 m (2-5 km^2) and 50 m (5-20 km^2).

Similar relationships have also been developed from the Swiss Alps and recorded in the inventory by Müller et al. (1976, p.11). They indicate that for glaciers between 0·5 and 23·0 km^2 in area (A) their mean depth (H in metres) can be estimated from the equation: $H = 5{\cdot}2 + 15{\cdot}4\sqrt{A}$. Since the glaciers in the Thvera area vary from about 0·2 - 2·8 km^2 in area, these relationships would indicate their mean depths to be between 11 - 31 m. Based on these approximations, together with a study of the topography in the unglacierised cirques, the contours over the glacierised areas were smoothly displaced by between 0·5 to 2 contour spacings to account for the ice thickness.

It was intended to carry out some spot determinations of ice thickness on these glaciers to assess the accuracy of these estimated values. For this purpose cooperation was sought from the Icelandic glaciologist, Helgi Björnsson (from the Science Institute, University of Iceland), who wished to test the recently designed radio echo-sounding equipment for temperate glaciers. The Thvera area was visited in August 1978 but, due to problems of accessibility and transportation of the equipment over the rough terrain, it was not possible to carry out thickness determinations on any of the glaciers in the Thvera area. The nearest readily accessible glacier on which radio echo-sounding could be carried out was Gljufurárjökull, about 2 km from the south-eastern corner of the Thvera matrix (Figs. 3.9, 4.11 and 5.1).

Gljufurárjökull, being more extensive with a well defined snout projecting into the valley(Gljufurá) below the glacier, is not directly comparable to the more confined glaciers in the Thvera matrix. However thickness determinations on Glujufurárjökull would give some idea as to the upper limit of ice thickness in this region. The radio echo equipment used operates at 2 to 10 MHz and has a range of 30 to 400 m with a range resolution of 8 m. The details of the equipment are provided by Sverrisson et al. (1980). On Gljufurárjökull spot thickness determinations were possible on the relatively uncrevassed lower portion of the glacier (Figs 5.2 and 5.3). The equipment functioned well with the easy identification of pulses reflected from the bedrock. The ice depths were calculated to vary from about 30 m near the glacier margins to a maximum of 100 to 115 m in the central part; the mean depth being rather greater than that assumed for the glaciers in the Thvera matrix. Given the more confined nature of the latter, the use of the above ad hoc corrections for mean glacier thickness can be regarded as being approximately correct.

5.3 Description and Application of the Terrain Analysis Program

In this section the terrain analysis system of programs (Young, 1978; Evans, 1979) are described. These programs, in addition to calculating values of the selected topographic parameters at the grid intersections of the altitude matrix, determine summary statistical characteristics for these parameters by using SPSS routines. The results of these programs are displayed in a variety of ways, viz. histograms, scatter diagrams, line printer density maps and graph - plotter maps. The programs are written in FORTRAN and have been run under the MTS system on the NUMAC IBM 370/168 computer. As an example of their use, the programs are applied to the Nupur area (described in sec. 4.7) to derive the topographic input parameters for mass balance computations.

5.3.1 Calculation of Topographic Parameters

The main terrain analysis program (OMY8) utilises altitude matrix data to derive descriptive characteristics for slope gradient and aspect, profile and plan convexity, in addition to calculated altitude, for the central point of each 3x3 submatrix consisting of adjacent grid points (Fig. 5.4). This is done by fitting a quadratic surface using the least squares method and solving for its first and second horizontal and vertical derivatives at the central point. The exact mathematical procedures used for the calculation of these derivatives are contained in the report by Young (1978).

The altitude matrix is arranged so that the altitudes are equally spaced in the N-S and E-W directions, the first row being the northern most and is read from west to east. Other input specifications for the terrain analysis program are the distance in metres between grid points, the number of columns in the data matrix and the scaling factor to convert the altitudes to metres. An example of the input for program OMY8 is given in Appendix C1.

In order to interpret the results of the terrain analysis program it is necessary to construct suitable maps depicting the spatial variation of the topographical parameters. For this purpose variable density line printer maps were found to be the most convenient. In these, each data point in the matrix is represented by one printer position and its value is indicated by the density of printing (variable density classes being formed by overprinting characters). The mapping programs, OYSM and OYSMA, are described by Young (1978). OYSM produces a 6-density map while OYSMA, used to map aspect data, produces an 8-density map.

The input data values for the mapping programs need to be in the form: row number, column number and value. The output of topographic parameters from the terrain analysis program is in this form, enabling its direct input into the mapping programs. Other input requirements for the mapping programs are listed with reference to the OYSM program in Appendix C2. In addition to the line printer programs a slope plotting program (OMYL) is available for producing a map showing the direction and magnitude of the gradient at each grid point.

5.3.2 Application in the Nupur Area

The variation of altitude, gradient, aspect, profile and plan convexity in the Nupur area is depicted by Figs. 5.5 - 5.9 respectively (constructed by using the line printer programs, OYSM and OYSMA, with input of topographic values generated by the terrain analysis program). It should be noted that due to a difference in the size of the line printer characters horizontally (giving 10 columns per inch) and vertically (8 rows per inch), the square Nupur matrix has been altered to a rectangular one in these line printer maps. This problem of vertical elongation can be resolved by using a square line printer (not available at Durham). Alternatively, the SYMAP mapping package can be used to produce a square map by interpolating extra columns resulting in a much larger map. The highly photo-reduced maps (originals are 4.7 x these)of the topographic characteristics in the Nupur area, produced by SYMAP are presented in Figs. 5.10 - 5.14 with the key to the symbols used in Fig. 5.15.

While the SYMAP maps have the advantage of reproducing a square matrix, the interpolation of extra characters around each grid point leads to undesirable effects. Firstly, for a large-sized matrix, several separate computer pages of output are required in producing one map. In the case of the Nupur area, the matrix had to be divided into 9 square plots (or 3 vertical strips) and separate maps produced for each plot before being combined and photo-reduced. This rather cumbersome process defeats the object of producing a rapid output of maps in a convenient form for analysing the results from the terrain analysis program. Furthermore, interpretation of the topographic parameter value at a particular grid point is made difficult by having several interpolated points for each grid intersection location. For these reasons the single-sheet maps produced by the OYSM and OYSMA programs, in which each grid point is represented by a single printed character, are preferred both for their speed of production and convenience of interpretation. Their close similarity with the SYMAP maps further reveals that the portrayal of the topographic characteristics is only slightly affected by their departure from a square matrix.

To permit comparison of the line printer maps with the original contour map of the Nupur area, the latter has been suitably redrawn (using the GPCP package program[1]) with appropriate vertical and horizontal scales to match the line printer maps (Fig. 5.16). By superimposing this contour map (with locations of grid intersections marked)on to the line printer maps, the value of any topographic characteristic at a particular grid point (or an impression of its variability over the whole area) can be easily obtained. (Note: these comparisons are facilitated by use of the transparent overlay for the Nupur study area, enclosed in the pocket bound in the back of this thesis).

It can be seen that the calculated altitude map (Fig. 5.5) bears a close resemblance to the contour map of the Nupur matrix (Fig. 5.16). The gradient map (Fig. 5.6) highlights the steep slopes forming the cirque and valley walls. It also clearly identifies the

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[1]Contour maps drawn by Calcomp's GPCP (General Purpose Contouring Program) as implemented on NUMAC IBM 370/168 comuter running under MTS.

flat summits of the plateau remnants. The aspect map (Fig. 5.7) indicates northerly aspects by the darkest shading and southerly aspects with the lightest, while easterly and westerly directions have printer characters different in form but similar in density. The contrasting aspects of slopes in the cirques are well portrayed in this map.

In order to depict the variation of both slope gradient and aspect in one map, the slope plotting program (OMYL) is used to produce a map showing the direction and magnitude of gradient at each grid point (Fig. 5.17). This is done by using the graph plotter to produce 'arrow plots'. The length of the arrow depends on the magnitude of the gradient (in 5 classes), while its direction represents the slope aspect. In Fig. 5.17 the sites of the cirques and plateau summits are clearly depicted.

The profile convexity map (Fig. 5.8) is useful in indicating the main convex ridge-lines present in the area. Interpretation of the plan convexity map (Fig. 5.9) is rather difficult, however, consideration of the first class (< - 40.0), indicated by blank areas, reveals the pattern of the major concavities formed by valley and cirque floors.

5.3.3 Summary Statistics

In addition to the individual values of the topographic characteristics at each grid intersection, the terrain analysis program provides numerical summary statistics for the whole altitude matrix. Those for the Nupur matrix are presented in Table 5.1. Moment statistics, viz. mean, standard deviation, skewness and kurtosis (the latter modified so that the Gaussian expectation is zero), are calculated for altitude, gradient, profile and plan convexity. Since moment statistics are not appropriate for aspect, the central tendency for the latter is measured by the direction of the resultant vector and the strength of this tendency is given by the vector strength. It should also be noted that zero gradient points are given special consideration since the values of the characteristics for these points are incomplete (aspect being indeterminate). Two separate summary tables are thus constructed, the first excluding and the second including the zero points.

From Table 5.1, it can be seen that the glaciated Nupur area is characterised by a high variability of altitudes (with a small negative skewness reflecting the extensive areas of flat summit plateaus and high cirque floors); high mean and variability of gradient, positive skewness of profile convexity and an excess of strong positive plan convexities. For a more detailed consideration of these summary statistics the reader is referred to the discussion in chapter III of the report describing the terrain analysis system (Evans, 1979).

5.4 Test of Accuracy

In view of the importance attached to the role of topography in glacier mass balance determinations, it is desirable to have a check on the accuracy of the topographic parameters calculated by the terrain analysis program. A limited test for calculated altitude, maximum gradient and slope aspect was carried out by comparing the calculated topographic values with those determined manually from the contour map. 75 points were randomly selected from the Thvera matrix for this purpose.

Figs. 5.18 - 5.20 illustrate the comparison between the calculated and measured values. It can be seen that in all three cases there is close correspondence between the two sets of topographic values. This is further confirmed by calculating least squares regressions between calculated (C) and measured (M) values (Table 5.2). In each case, the regression coefficient is within 1.5 standard errors of its expected value of 1.0, and the standard error of the estimate is low in relation to the range of values encountered.

5.5 Effect of Changing Grid Mesh

While, for reasons outlined in sec. 4.4.2, the topographic parameters have been calculated using a grid mesh of 100 m, it should be emphasised that certain climatological and glaciological processes may operate at other spatial scales. In view of this, it is interesting to reflect on the effect of a change in grid resolution on the calculation of the topographic parameters.

One way in which this can be considered is to obtain results from the terrain analysis program at increasingly coarser grid intervals. This has been done for the central Nupur 'test' matrix (Fig. 5.21) at grid intervals of 100, 200 and 300 m. The number of points for which topographic parameters can be determined is reduced as the grid mesh is enlarged. Thus, for the 42x82 'test matrix', topographic parameters are yielded for 3200 points with a 100 m grid interval; 2964 points with 200 m and 2700 points with 300 m. The resulting effects of grid mesh on the topographic characteristics are presented in Figs. 5.22 - 5.27 and Table 5.3.

Altitude statistics, as seen from Table 5.3 and Fig. 5.22, are the least affected by grid mesh since they are not based on derivatives. Maximum gradient shows a progressive decrease in mean gradient and standard deviation with increasing grid mesh. Fig. 5.23 reveals the increasing generalisation of gradient with the use of 200 and 300 m grid mesh. The narrow belts of steep gradient (>36^{o}), formed by the slopes encompassing the cirques, at the 100 m scale are replaced by broader belts of gentler gradients (24-30^{o}) at the 300 m scale. Apart from this the basic spatial pattern is retained at all three scales.

The main effect of increasing grid mesh for the aspect of maximum slope is the evening out of local variations, resulting in a more simplified portrayal of the aspect pattern (Fig. 5.24). This effect is seen more clearly in Fig. 5.25 in which all points with gradients less than 20 degrees have been excluded (importance of aspect being greater with steep slopes). At the 300 m scale larger blocks of points with the same aspect are found than at the 100 m scale.

The convexity measures are most sensitive to changes in grid mesh, as can be seen from the sharp reduction in the extreme values of profile convexity, resulting in the ridge pattern being less clearly defined at the 300 m scale compared with the 100 m scale (Fig. 5.26). The moment statistics (Table 5.3) illustrate the instability of the plan convexity measures,however,the maps (Fig. 5.27) reveal a more consistent impression. The latter are of particular interest from the point of identifying concavities formed by valley and cirque floors. It is seen that at the 100 m scale extreme values of plan convexity (< - 40.0 degrees per 100 m) shown as blank areas reveal the sites of all local concavities but at the 300 m scale, only the major valleys and cirque floors are picked out.

The effect of increasing grid mesh, then, results in the increasing generalisation of these topographic parameters. It would appear that, from the point of studying local variations in glacier balance, it is the 100 m grid (or finer) which is the most relevant. The various effects of grid resolution and replicability of topographic parameters have been further considered by Evans (1979).

5.6 Utilisation of Topographic Parameters

The topographic parameters determined from the terrain analysis program are used at all stages of glacier mass balance determination; e.g. in the estimation of local climate by developing relationships between local topography and regional climate, in assessing the effects of snow accumulation processes such as drifting and avalanching, and in the determination of the spatial variation of energy balance components. The five basic topographic parameters are particularly useful in helping to identify certain topographic situations relevant for the computations of mass balance. As an example of these, the following elements of the landscape can be delimited in the Nupur area by selecting suitable limiting values of individual or combined topographic parameters.

(i) Plateau areas

The plateau summits may act as temporary storage areas from which fallen snow can be drifted into the nearby concavities. Trials with different combinations of altitude and gradient values determined that the limits used in Fig. 5.28 best identify the location of these plateaus summit areas and major ridges (compare with the contour map of the Nupur area in Fig. 5.16).

(ii) Concavities

The identification of suitable concavities in which snow and ice can accumulate is a particularly relevant exercise for glacier mass balance determination. A combination of negative profile and plan convexity values would appear to be most suitable for this purpose. Experiments with different limits of these measures resulted in Fig. 5.29 which attempts to delimit concavities formed by cirques, trough-heads and other landscape features. A choice of different limits for the convexity measures picks out the lineations formed by the main valley and cirque floors found in the area (Fig. 5.30).

It should be noted that the size of the concavity identified will depend on the size of grid interval used. As noted in sec. 5.5, the concavities located with a grid mesh of 100 m would seem to be the most appropriate for studying local glacier generation. However further experimentation, especially with grid intervals of less than 100 m, is necessary for the determination of the scale most suitable for the identification of concavities enabling sufficient snow and ice accumulation to generate local glaciers.

(iii) Ridges

The identification of the main ridges present in the area may be useful because of their effects on certain climatological elements (e.g. modification of wind speeds). A suitable limiting value of profile convexity is sufficient for portraying the main convex ridge-lines, formed by cirque and trough development, in the Nupur area (Fig. 5.31).

(iv) Areas of Steep Slopes

Given the importance of snow avalanching for mountain glacier development, it is important to distinguish areas with steep slopes. Maps showing slopes above certain threshold gradients can be constructed for this purpose (Fig. 5.32). Along with gradient, consideration of the aspect of such steep slopes is important for calculating the radiation balance and in the estimation of basic climatological elements. Given the importance of aspect with increasing gradient, these two measures can be combined to highlight aspects above a particular gradient threshold (Fig. 5.33).

It should be noted that in attempting to delimit landscape elements of glaciologic interest in any given region, the choice of the particular threshold limits adopted needs individual consideration, since the scale of these landscape features may vary from one region to another.

TABLE 5.1
Summary Statistics for the Nupur Matrix

NUPUR DATA
NO. OF ROWS = 80
STATISTICS FOR 6013 POINTS WITH NON ZERO GRADIENT

	EST. ALT.	GRADIENT	PROFC	PLANC
MEAN	440.313	21.875	0.391	-0.673
SDEV	191.938	12.426	11.888	52.723
SKEW	-0.246	0.187	1.313	2.271
KURT	-0.961	-0.969	9.033	60.669
MAX	776.111	59.138	112.178	998.855
MIN	10.000	0.477	-116.556	-687.550

VECTOR MEAN ASPECT ANGLE 94.616
VECTOR STRENGTH (PROPORTION) 0.113
GRADIENT WEIGHTED VECTOR MEAN ASPECT ANGLE 79.042
GRADIENT WEIGHTED VECTOR STRENGTH (PROPORTION) 0.044

CORRELATION COEFFS

	EST. ALT	GRADIENT	PROFC	PLANC
EST. ALT.	1.000	0.025	0.467	0.179
GRADIENT	0.025	1.000	-0.054	0.034
PROFC	0.467	-0.054	1.000	0.133
PLANC	0.179	0.034	0.133	1.000

STATISTICS INCLUDING ZERO GRADIENT POINTS
EST ALT AND GRADIENT FOR ALL 6084 POINTS
PROFC AND PLANC FOR 6083 NON ZERO AND PLAIN POINTS
WHERE PLANC IS TAKEN AS 0.0 FOR PLAIN POINTS

	EST. ALT	GRADIENT	PROFC	PLANC
MEAN	451.627	21.620	0.386	-0.666
SDEV	193.028	12.574	11.819	52.419
SKEW	-0.252	0.178	1.322	2.284
KURT	-0.967	-0.967	9.175	61.409

TABLE 5.2

Relationship between Manually Measured (M) and Calculated (C) Values of Altitude, Gradient and Aspect at 75 Points in the Thvera Matrix

	Regression	r^2	S.E.E.	S.E.B.	S.D.
Altitude	C = 1.003M - 2.373	0.9997	3.59m	0.002	191m
Gradient	C = 0.984M + 1.416	0.9840	1.27^{o}	0.015	10.1^{o}
Aspect	C = 1.009M - 1.630	0.9962	5.74^{o}	0.007	91^{o}

r^2 = coefficient of determination

S.E.E. = standard error of the estimate of C.

S.E.B. = standard error of the regression coefficient

S.D. = standard deviation of M.

TABLE 5.3

The Effect of Grid Mesh on Statistics for the Nupur Test Matrix

Altitude

MESH	MEAN	SD	SKEW	KURT	MAX	MIN
100m	471	196.2	-0.367	-0.88	776	29
200m	475	192.6	-0.369	-0.88	786	34
300m	482	185.4	-0.365	-0.89	806	56

Gradient

100m	21.13	12.13	0.178	0.97	55.32	0.00
200m	19.36	9.56	0.102	0.81	44.05	0.00
300m	17.23	7.71	0.114	0.58	38.82	0.45

Profile Convexity

100m	0.635	11.21	1.556	5.90	100.6	-42.2
200m	0.619	6.84	0.836	0.98	38.8	-21.2
300m	0.634	4.78	0.503	0.04	20.1	-10.4

Plan Convexity

100m	-0.023	52.7	3.70	80	999	-688
200m	1.937	98.4	45.46	2322	5042	-359
300m	0.551	37.6	-3.48	127	418	-823

FIGURE 5.1
Gljúfurárjökull at the Head of Gljúfurá
(View to the South)

FIGURE 5.2
Setting up the Radio Echo-Sounding Equipment on Gljúfurárjökull with Members of the Science Institute (University of Iceland).

FIGURE 5.3
Ice Thickness Determination on the Central Part of Gljúfurárjökull by Radio Echo-Sounding (View to the North with Gljúfurá in the background).

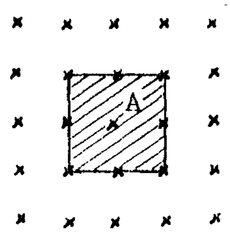
Interpolation of altitudes from photogrammetric contour maps

INPUT

Construction of altitude matrix

Terrain Analysis Program

- fits quadratic surface in each 3x3 submatrix (e.g. for central point A)



Surface solved for its first and second horizontal and vertical derivatives at the central point

OUTPUT

For each grid intersection:

(1) Calculated Altitude (m)

(2) Maximum Gradient (in degrees from horizontal, 0-90°)

(3) Aspect of slope (in degrees clockwise from North, 0-360°)

(4) Profile Convexity (degrees/100 m)

(5) Plan Convexity (degrees/100 m)

FIGURE 5.4

Use of Terrain Analysis Program

VALUES BETWEEN 100.000 220.000 340.000 460.000 580.000 700.000

NUPUR N.W.ICELAND
CALCULATED ALTITUDE (METRES)

\- BLOCK 1

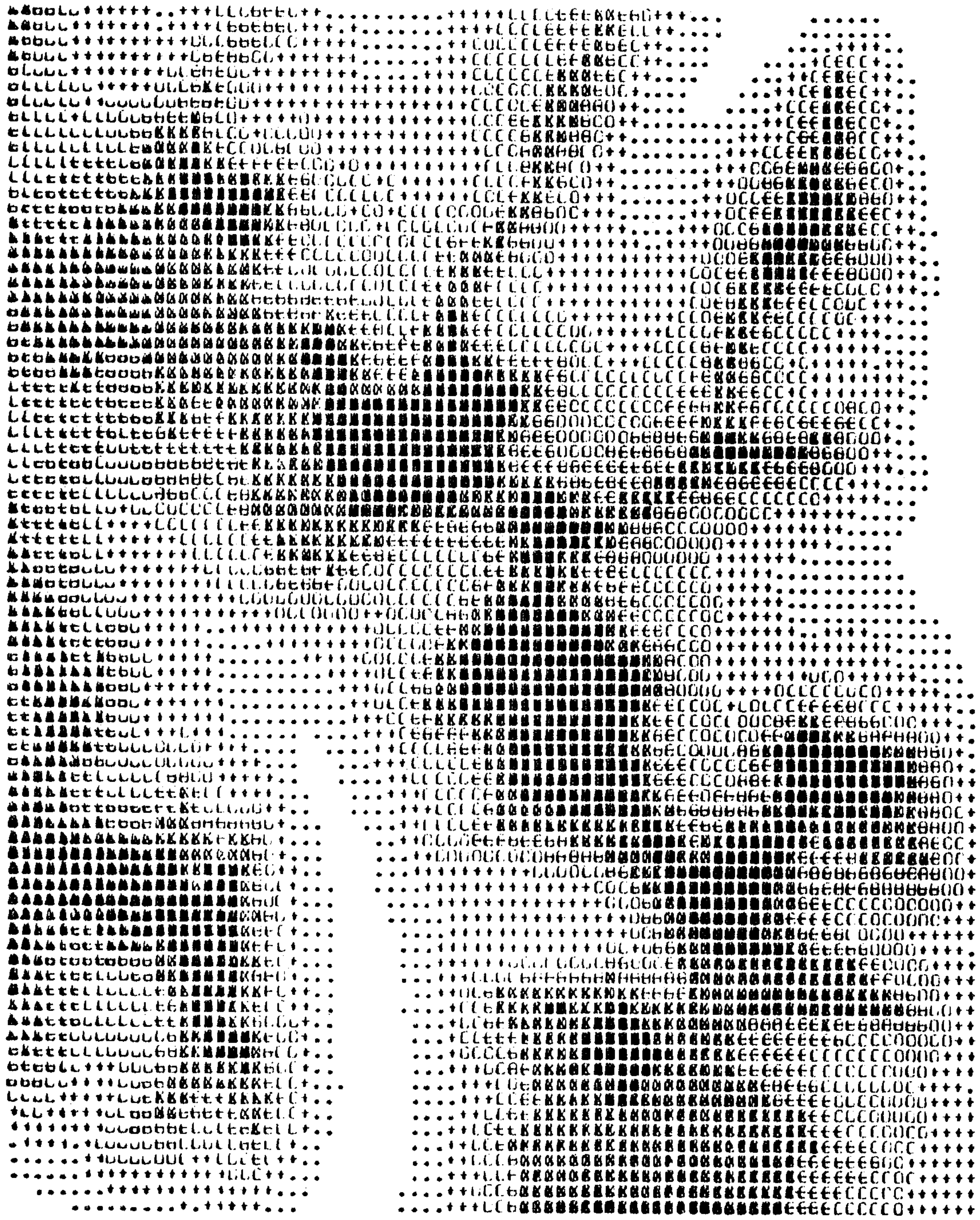


FIGURE 5.5
Nupur: Map of Altitude (using OYSM)

FIGURE 5.6
Nupur: Map of Gradient (using OYSM)

FIGURE 5.7
Nupur: Map of Aspect (using OYSMA)

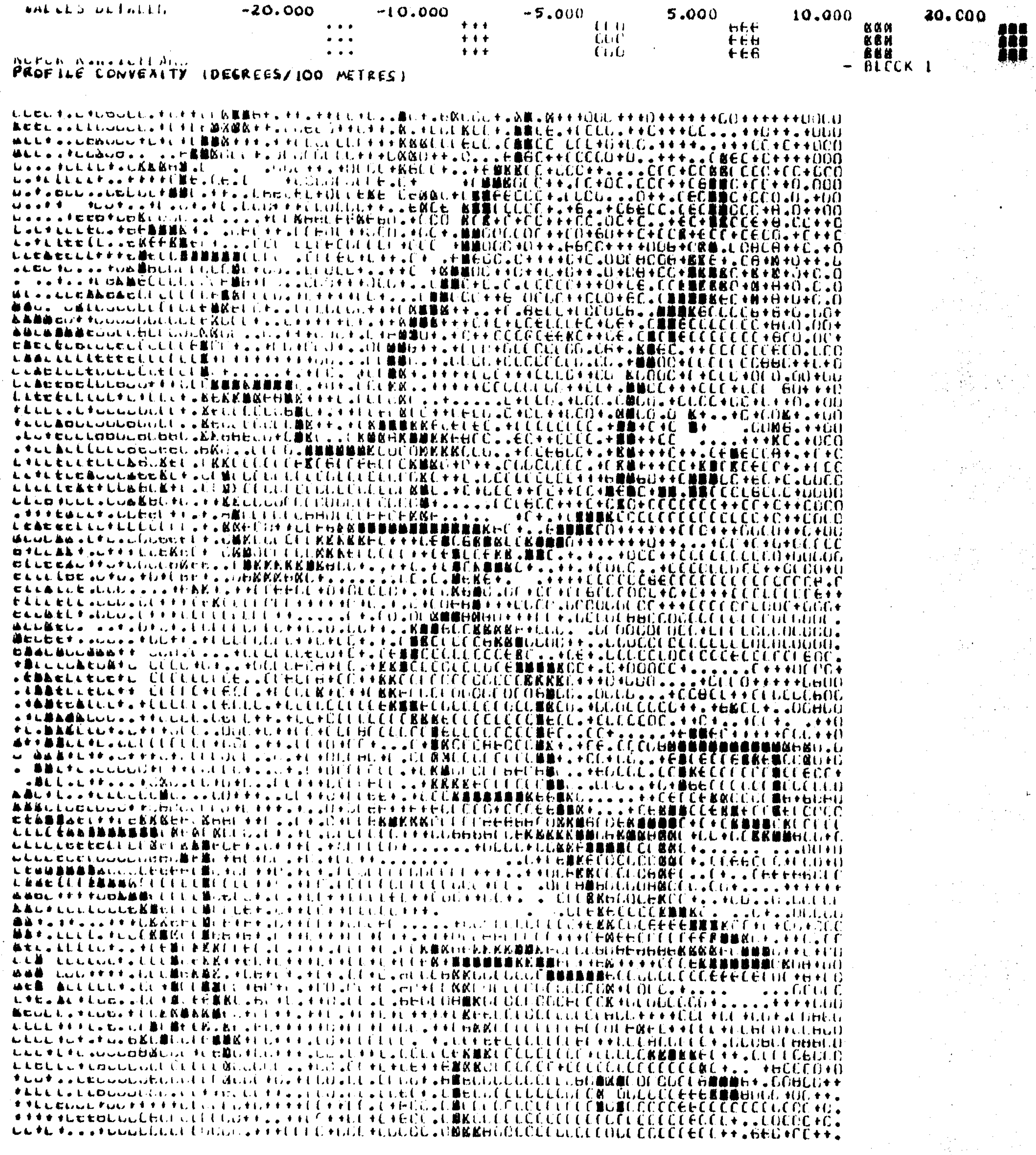


FIGURE 5.8
Nupur: Map of Profile Convexity (using OYSM)

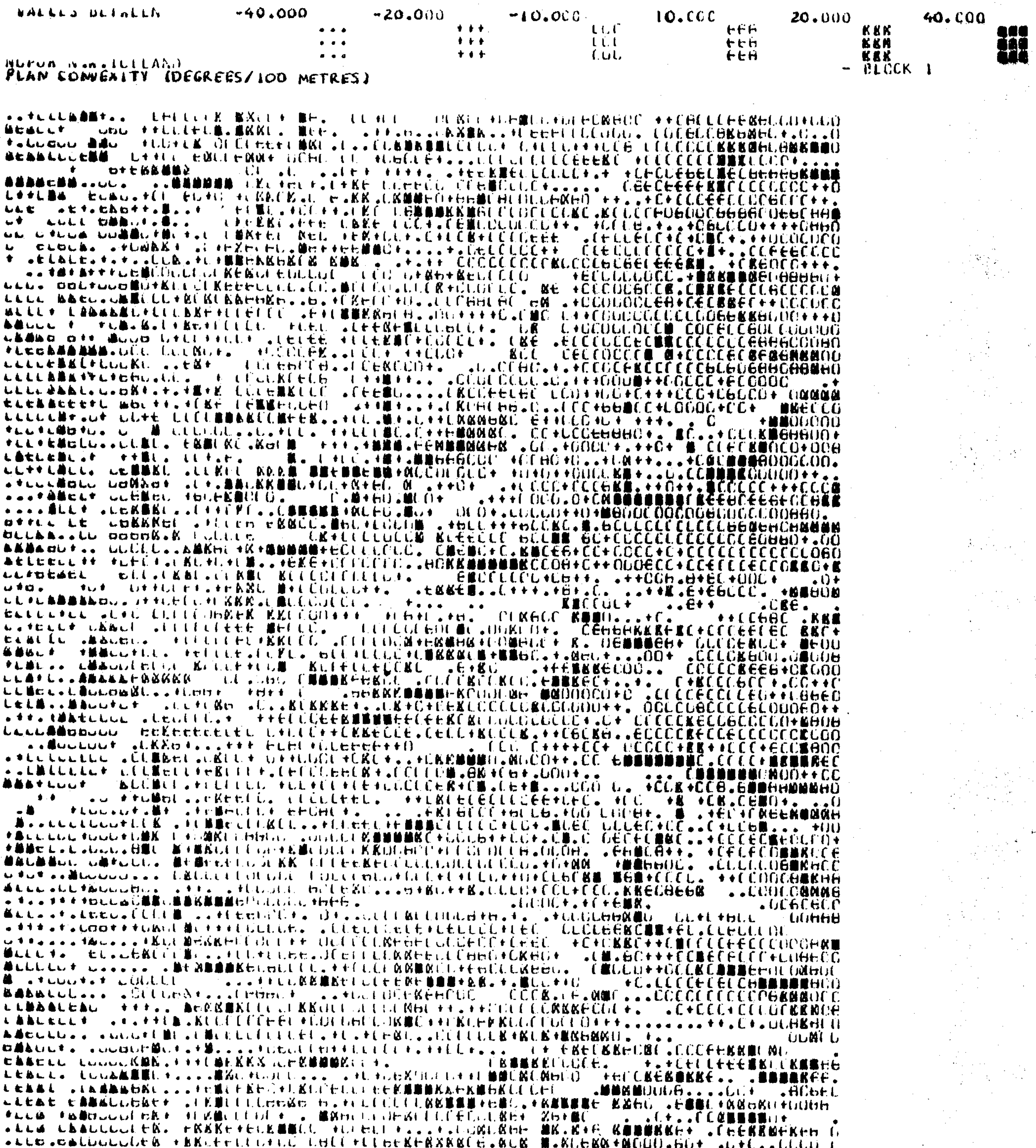


FIGURE 5.9
Nupur: Map of Plan Convexity (using OYSM)

FIGURE 5.10
Nupur: Map of Altitude (using SYMAP)

FIGURE 5.11
Nupur: Map of Gradient (using SYMAP)

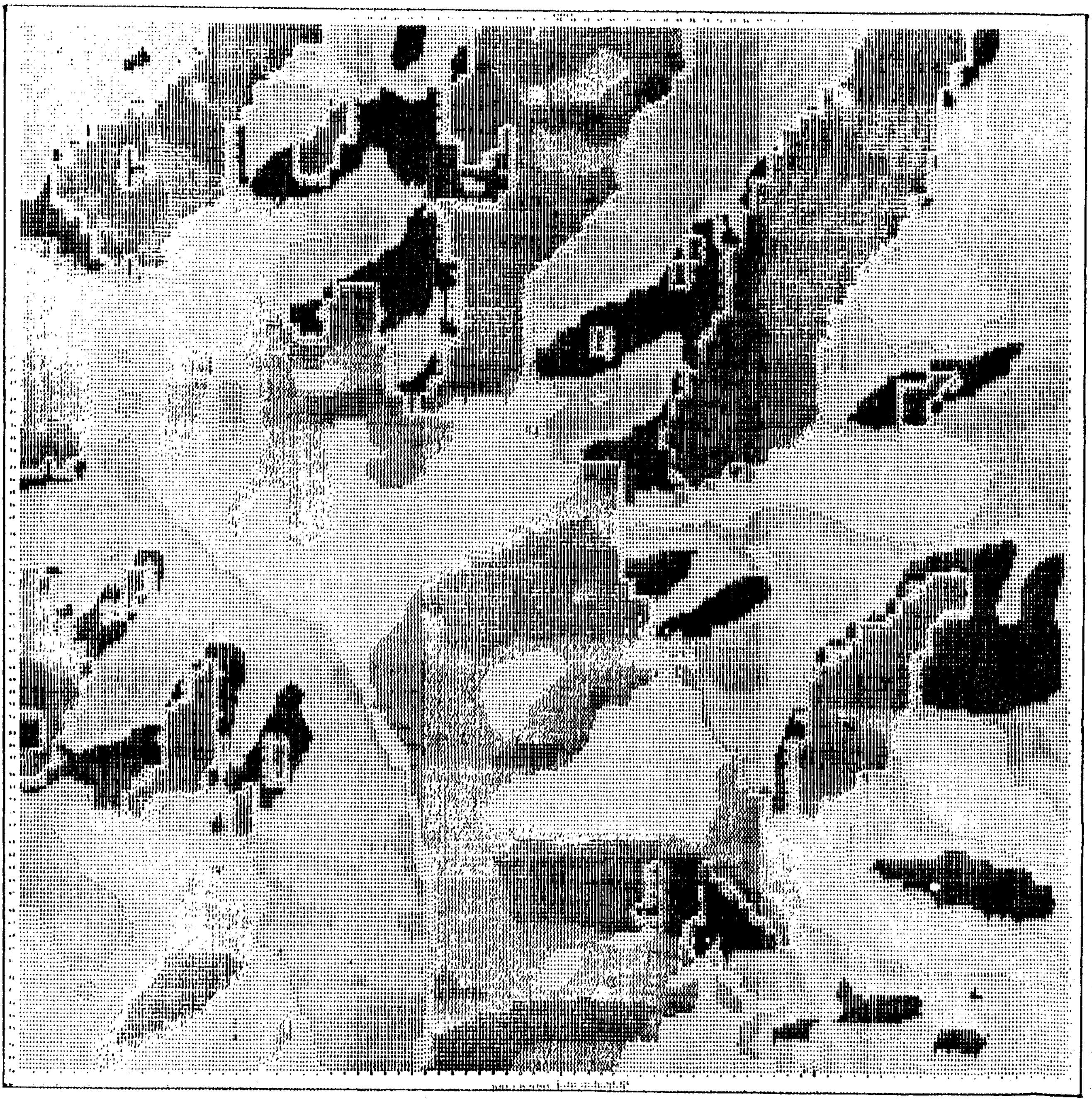


FIGURE 5.12
Nupur: Map of Aspect (using SYMAP)

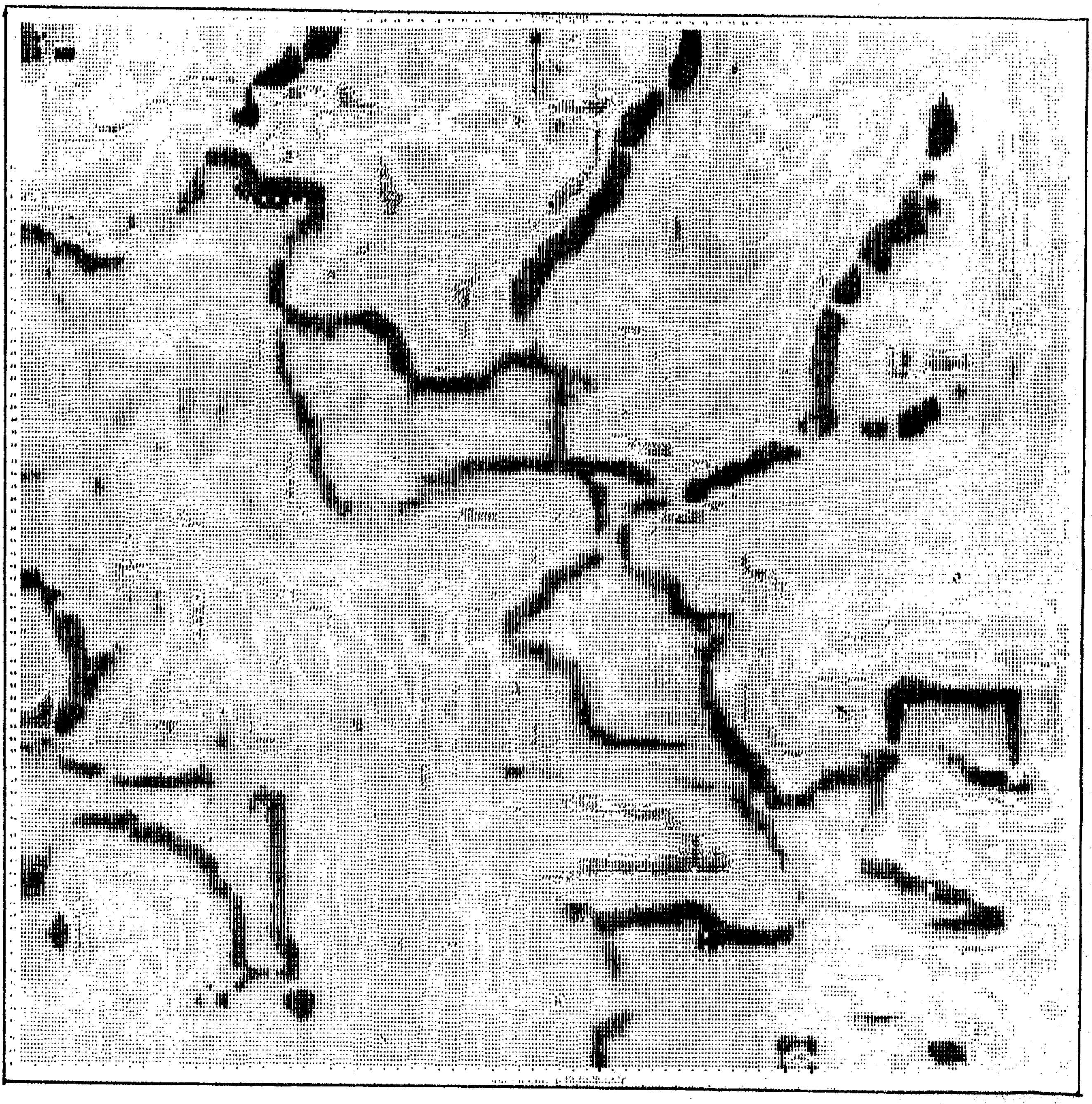


FIGURE 5.13
Nupur: Map of Profile Convexity (using SYMAP)

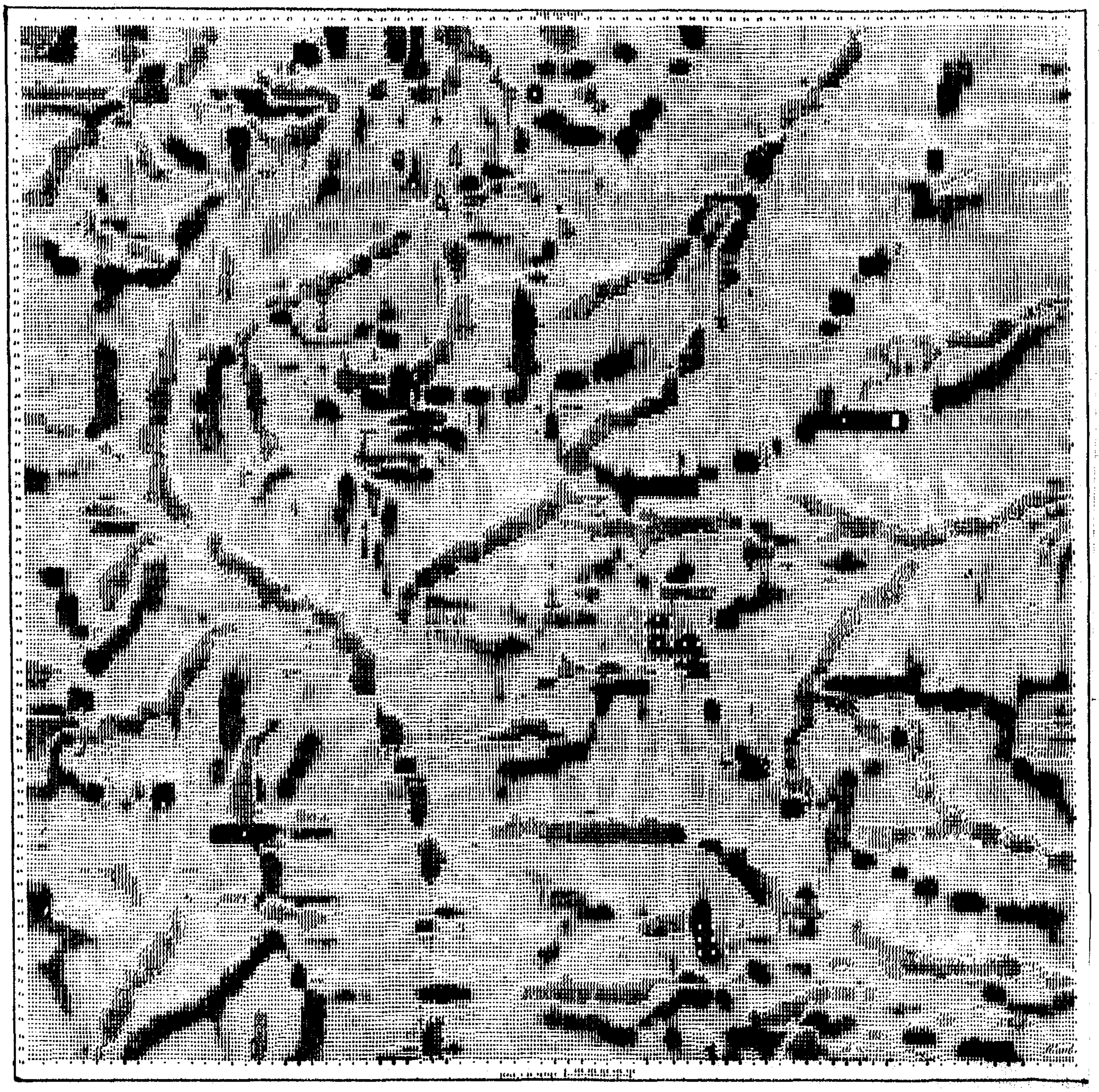


FIGURE 5.14
Nupur: Map of Plan Convexity (using SYMAP)

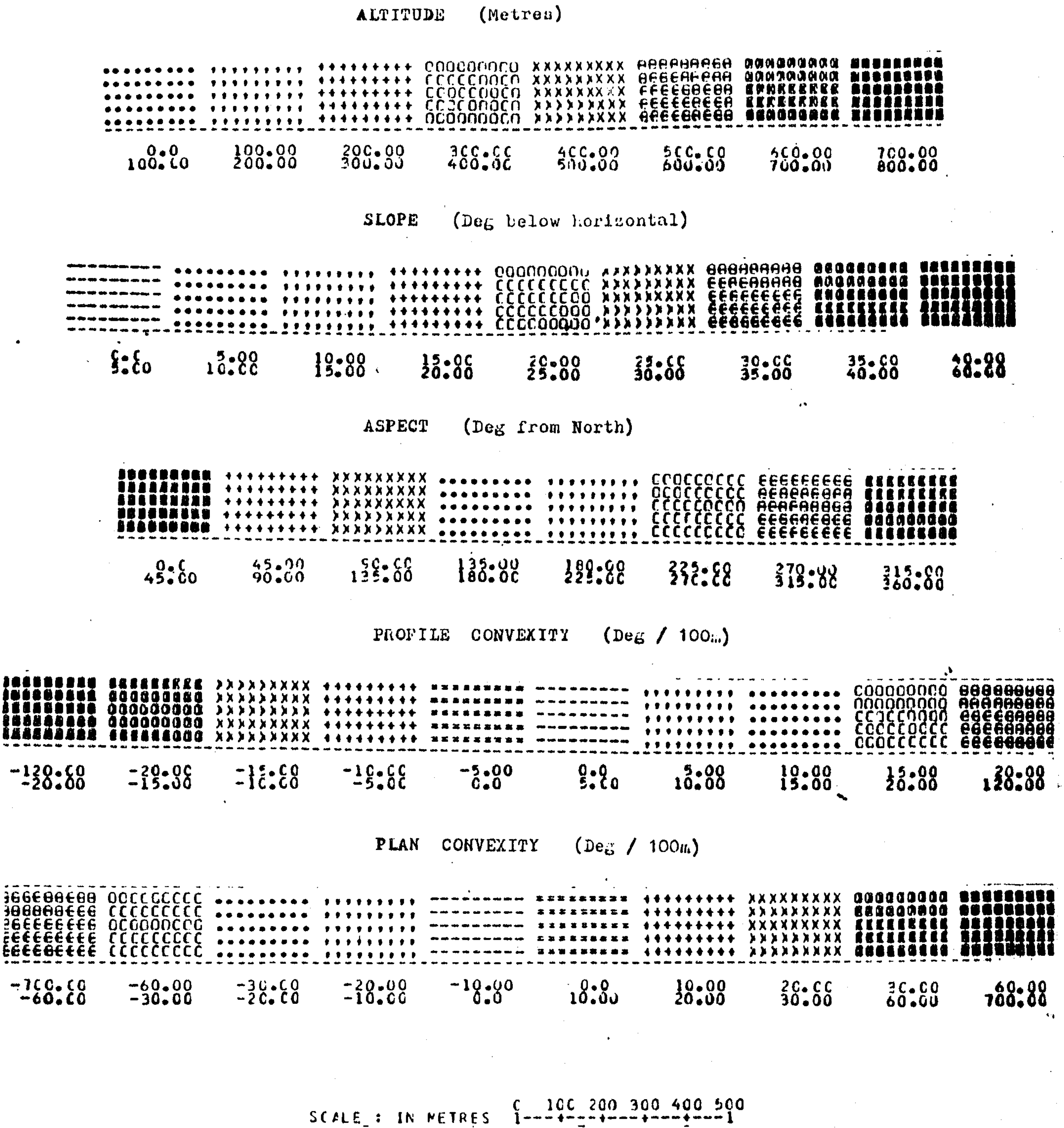


FIGURE 5.15
Key to SYMAP Maps (Figs. 5.10-5.14)

FIGURE 5.16
Nupur Matrix Modified to Match the Line Printer Maps.
(contoured by GPCP with a grid interval of 20 m.).

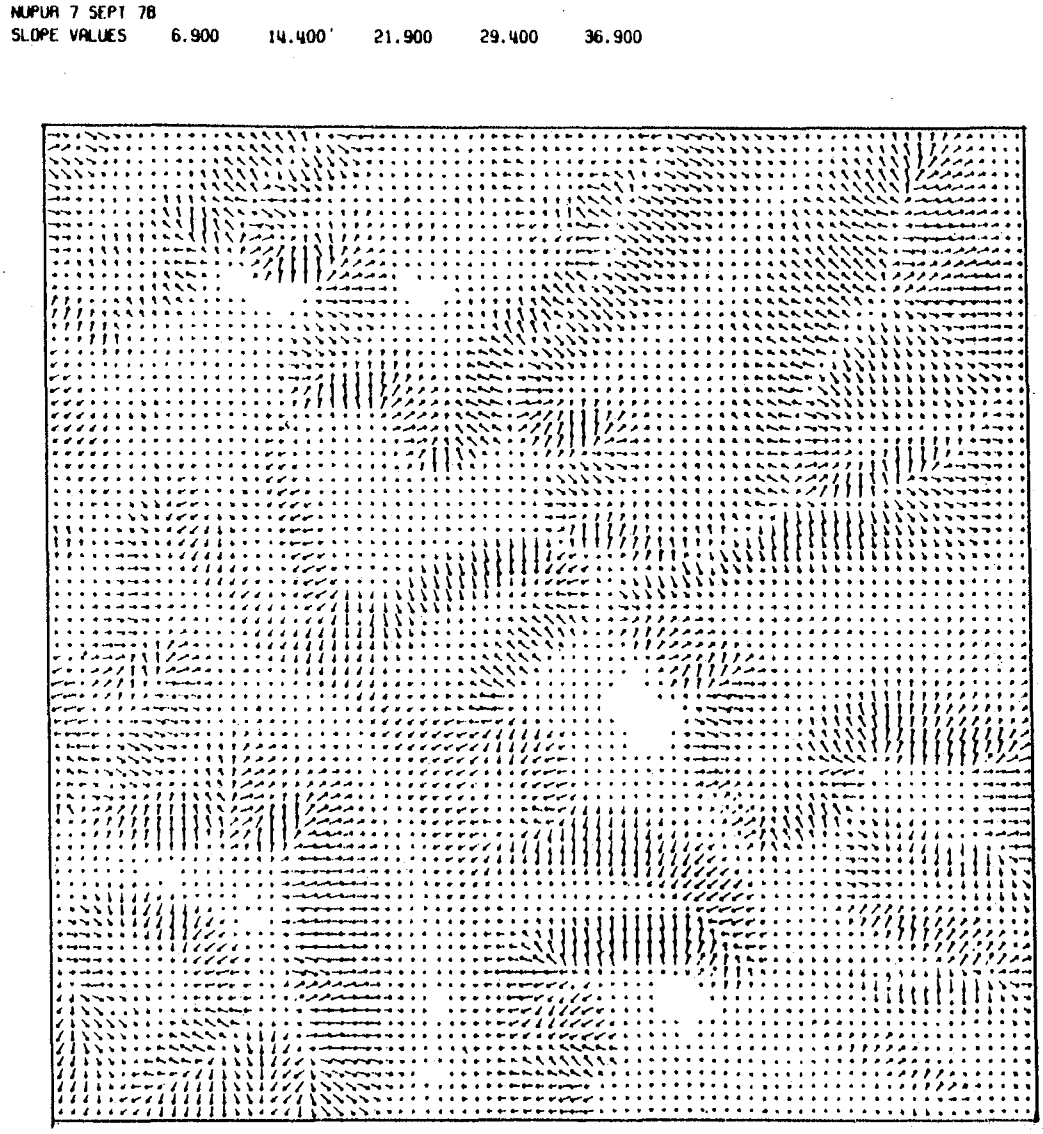


FIGURE 5.17
Nupur: Slope Map (using OMYL)

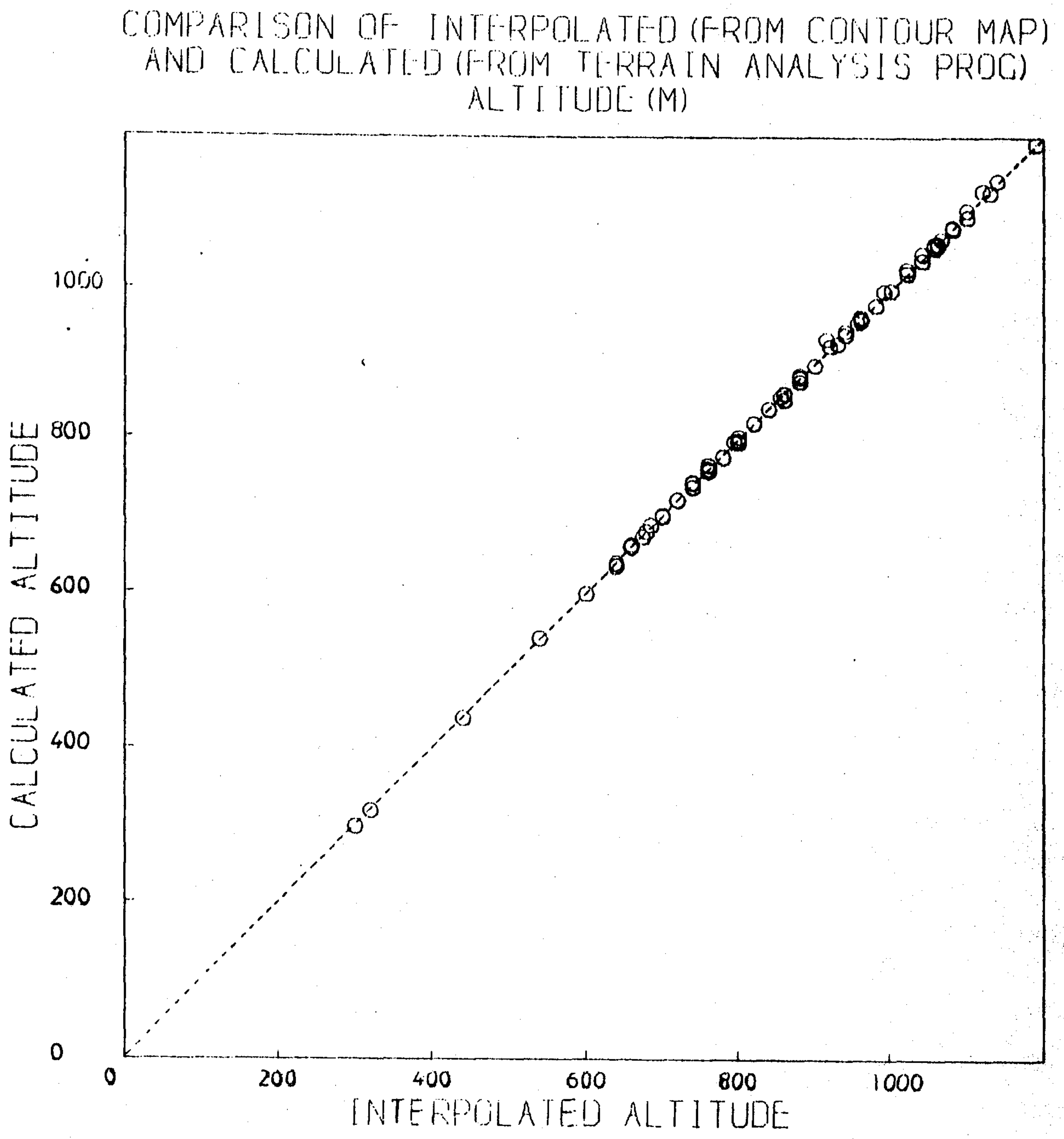


FIGURE 5.18
Test of Accuracy for Altitude

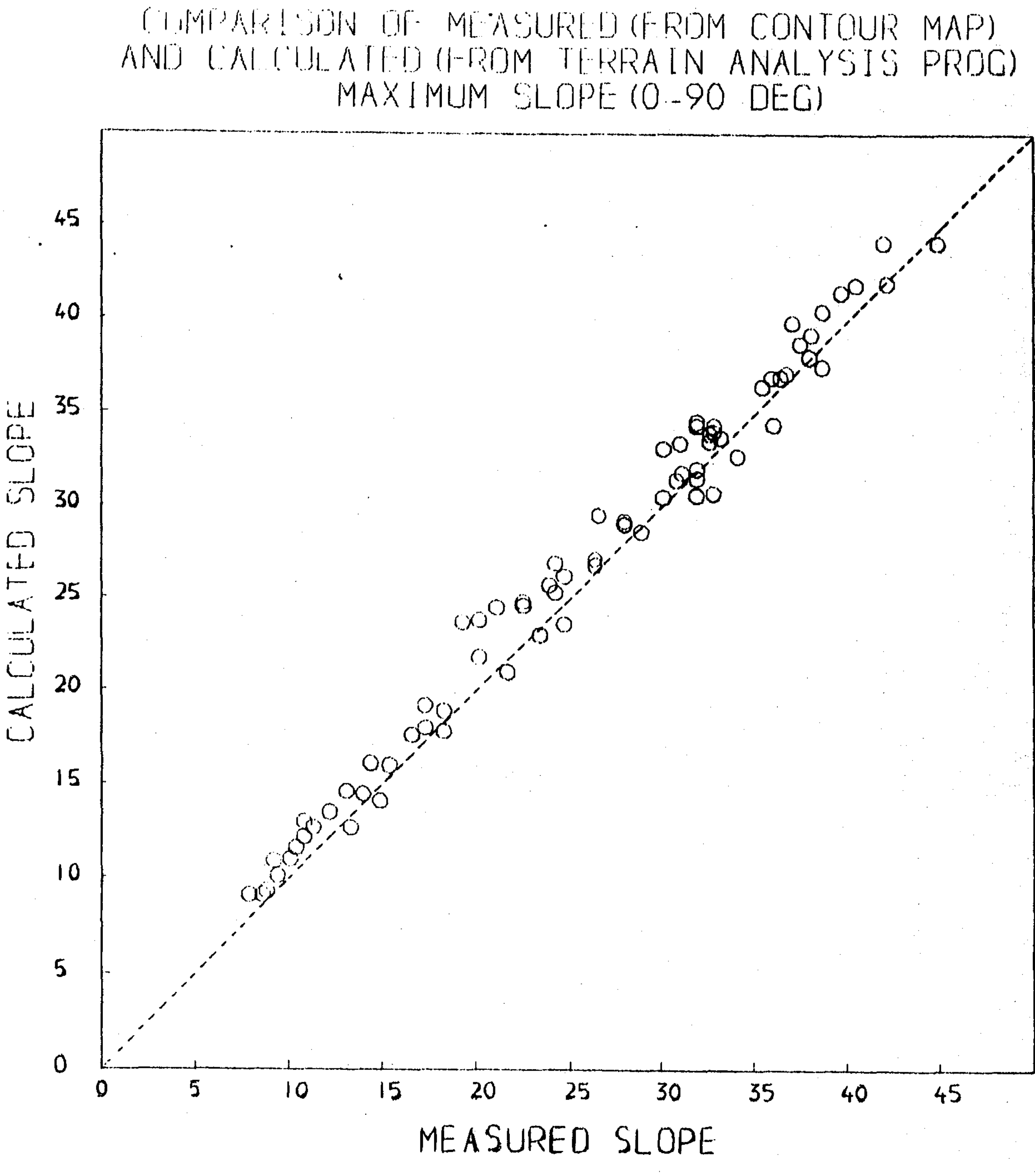


FIGURE 5.19
Test of Accuracy for Gradient

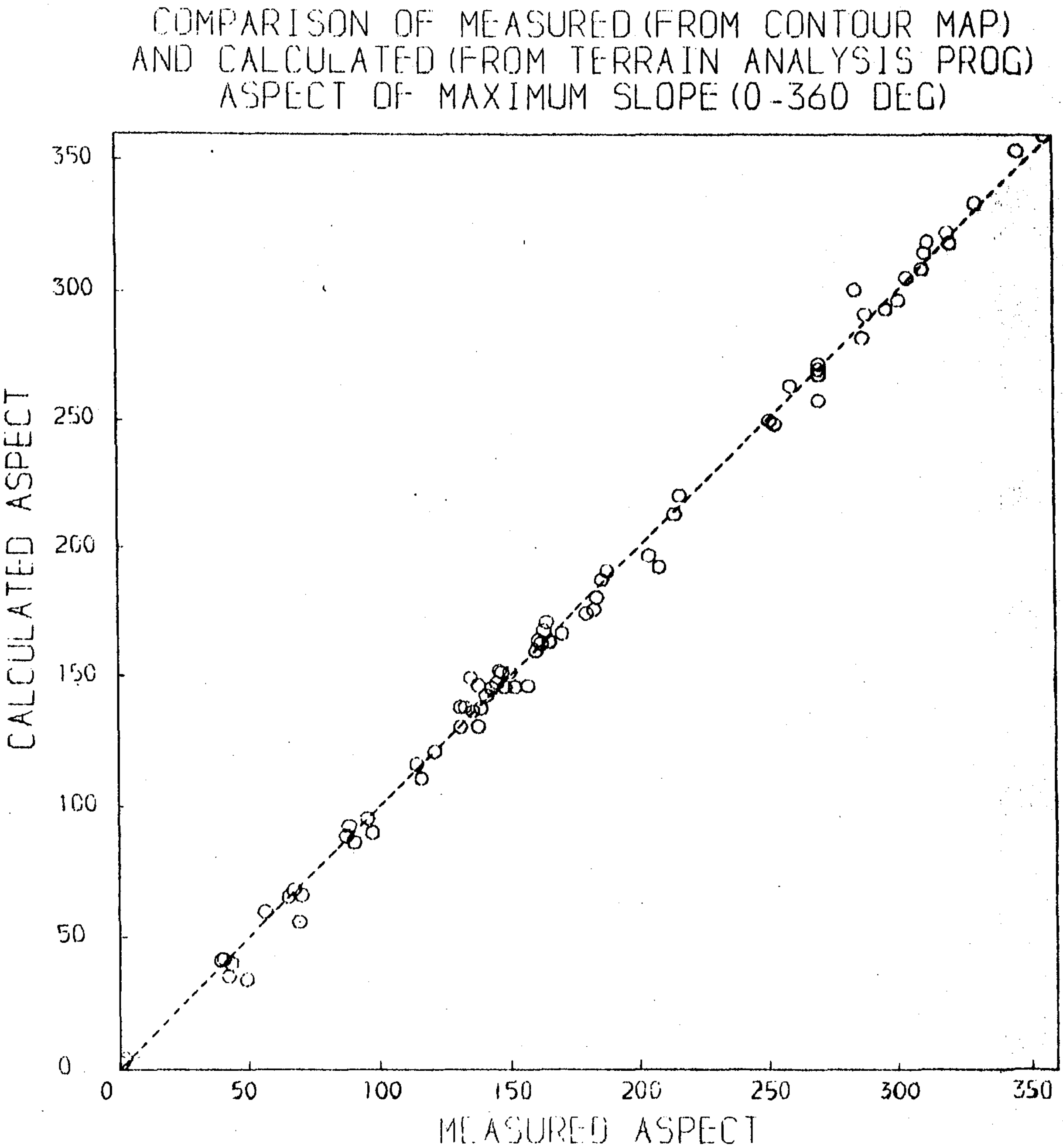


FIGURE 5.20
Test of Accuracy for Aspect

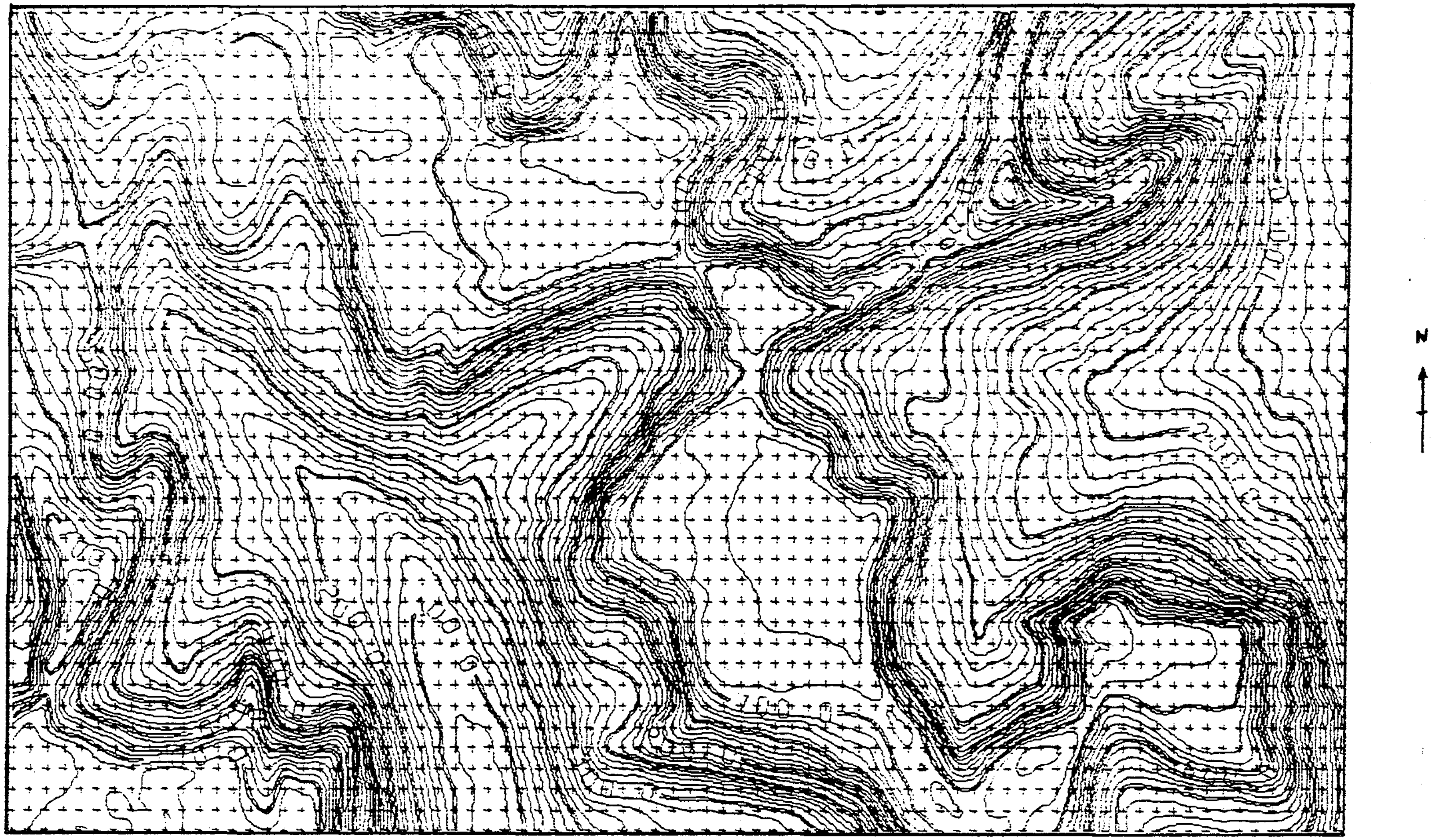


FIGURE 5.21
Central Nupur 'Test Matrix'

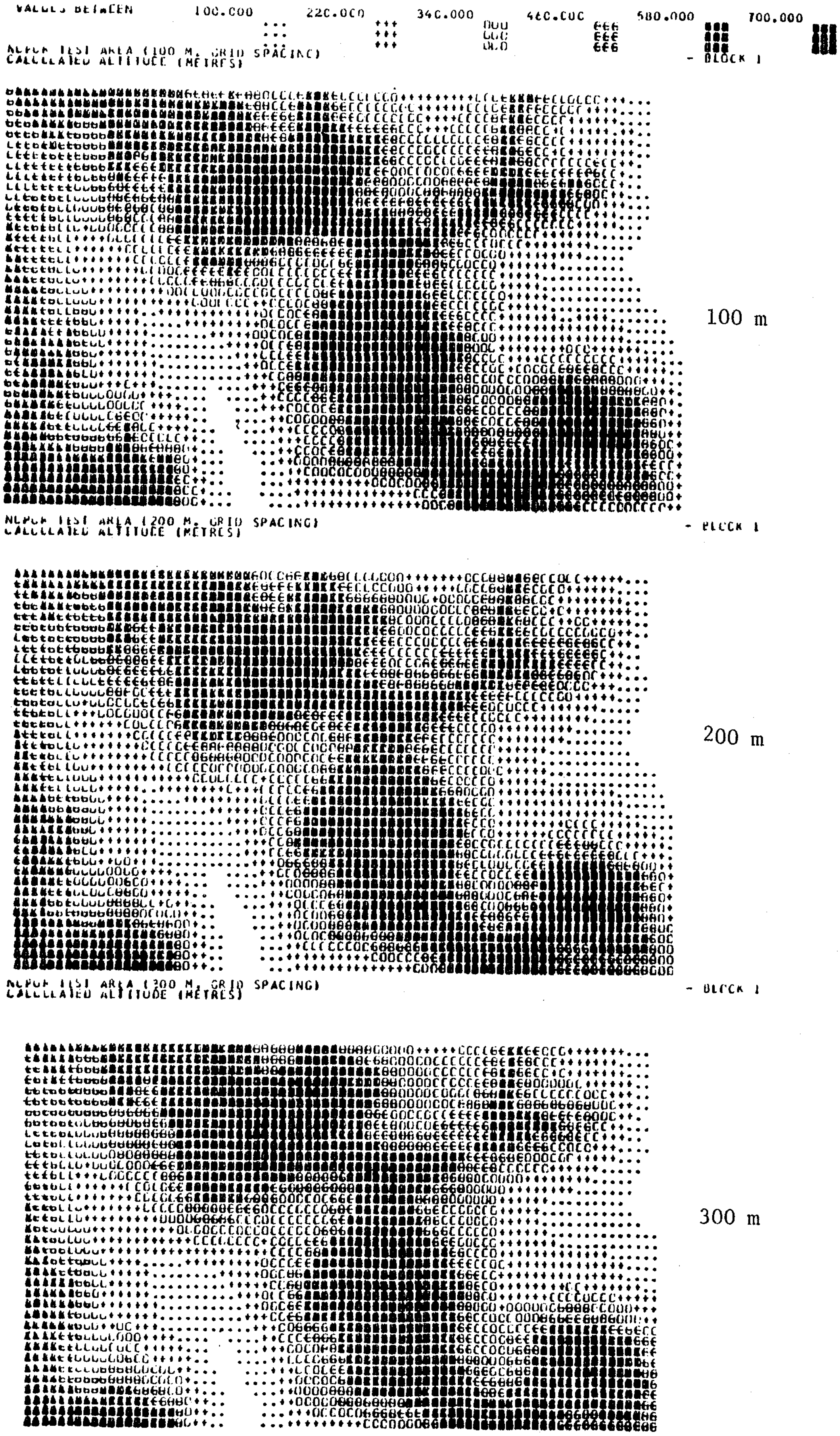


FIGURE 5.22
Effect of Grid Mesh on Altitude

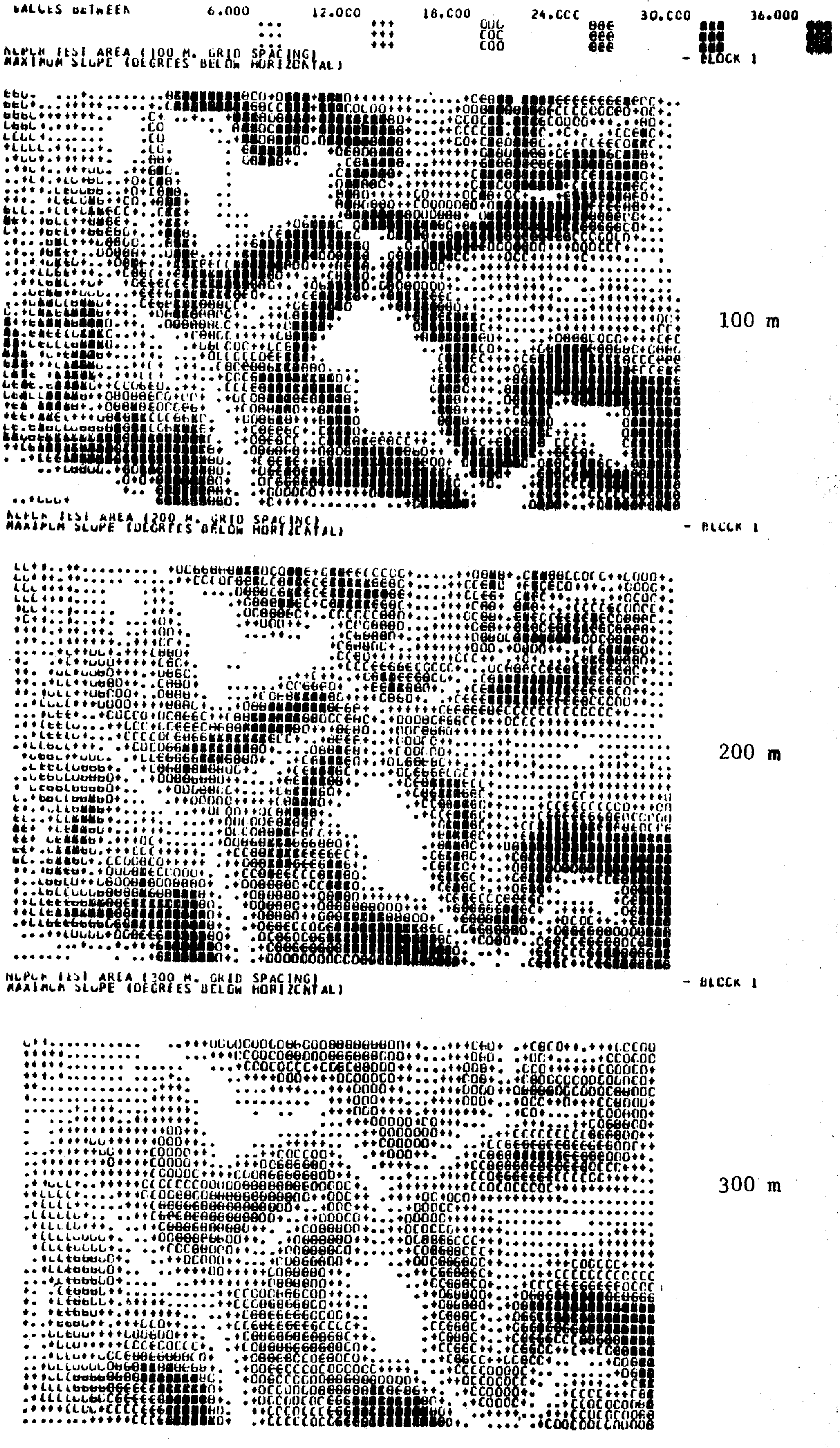


FIGURE 5.23
Effect of Grid Mesh on Gradient

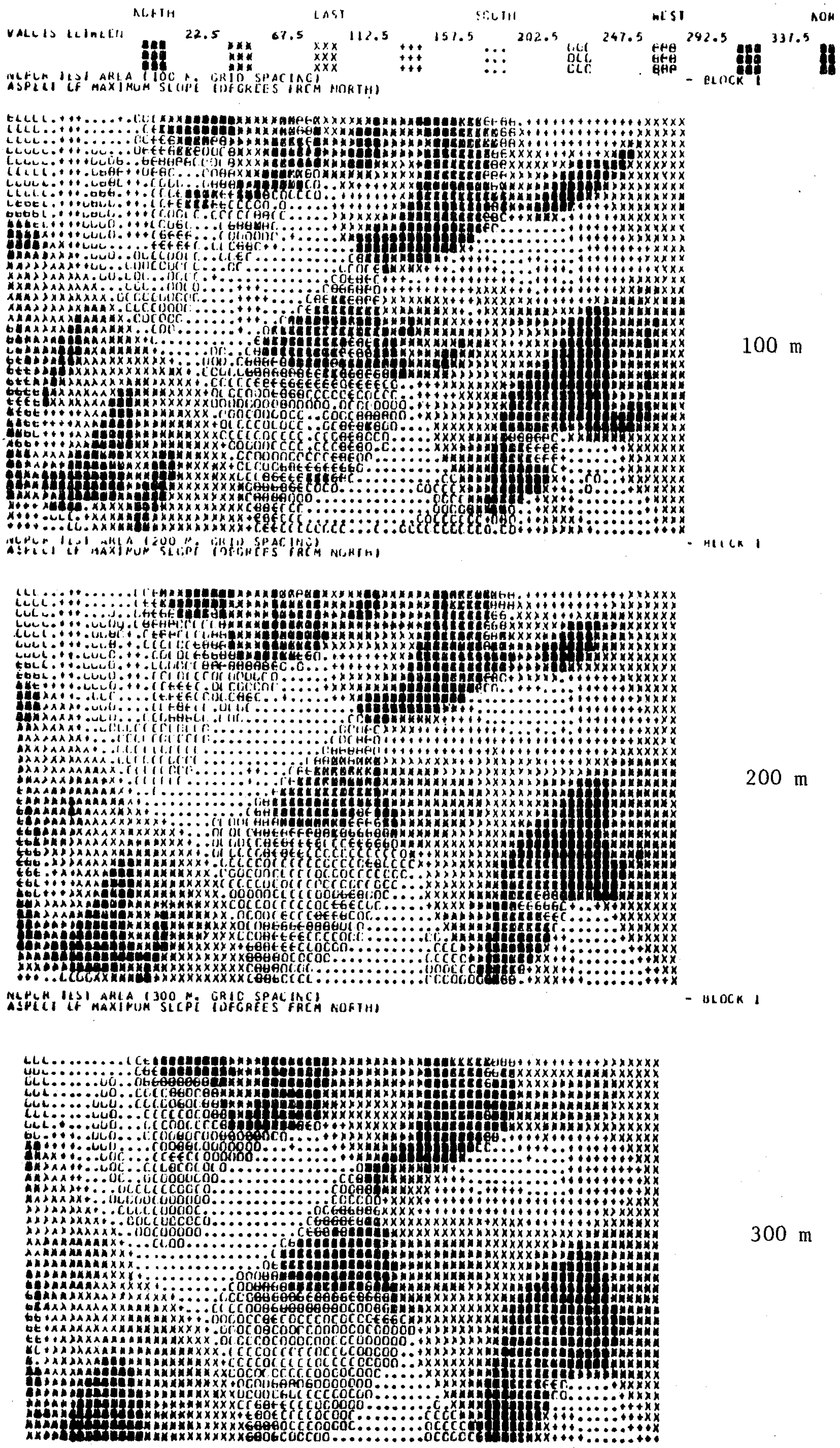


FIGURE 5.24
Effect of Grid Mesh on Aspect

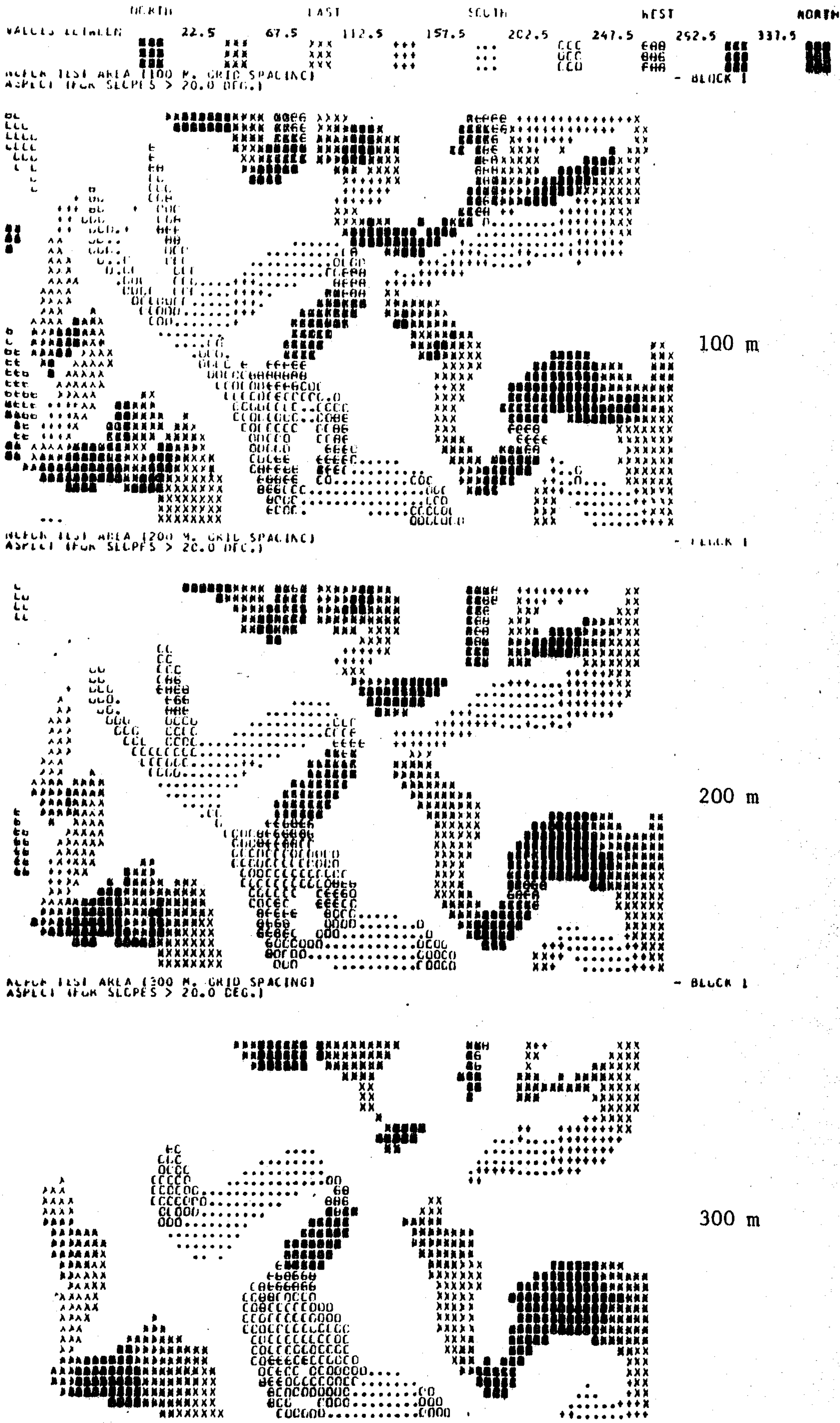


FIGURE 5.25
Effect of Grid Mesh on Aspects of Steep Slopes

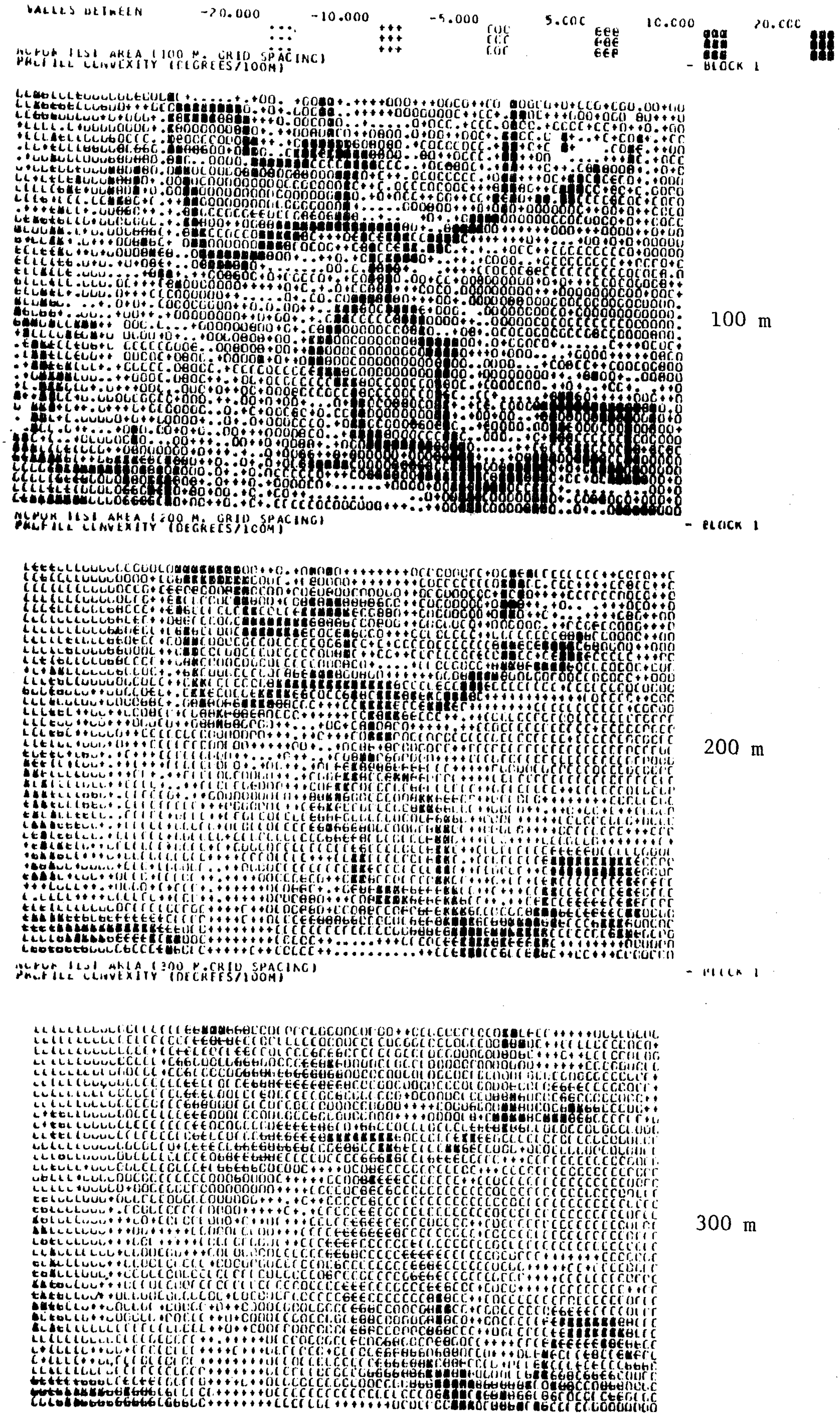
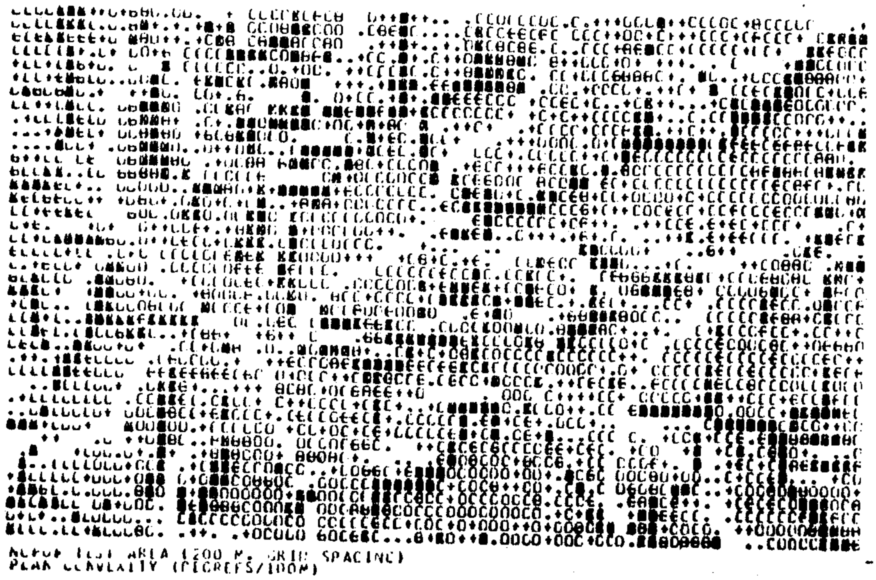


FIGURE 5.26
Effect of Grid Mesh on Profile Convexity

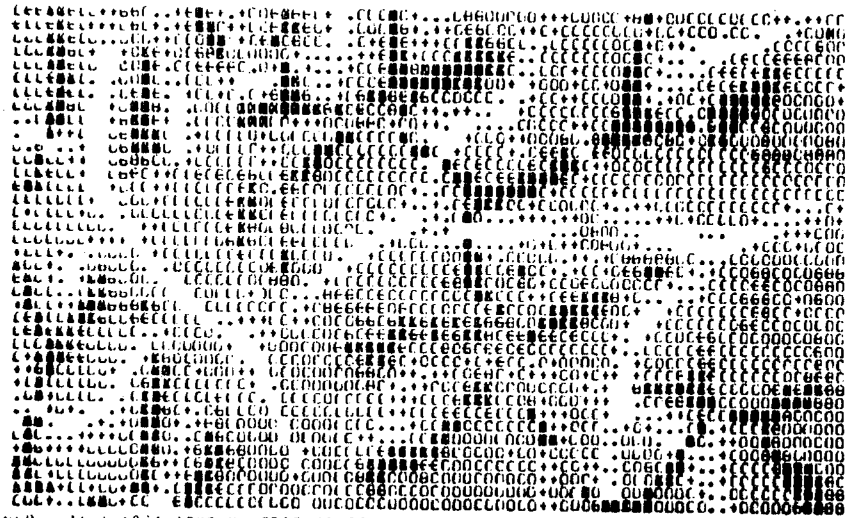
VALUES BETWEEN -40.000 -20.000 -10.000 10.000 20.000 40.000

NCFOR TEST AREA (100 M. GRID SPACING)
PLAN CONVEXITY (DEGREES/100M) - BLOCK 1



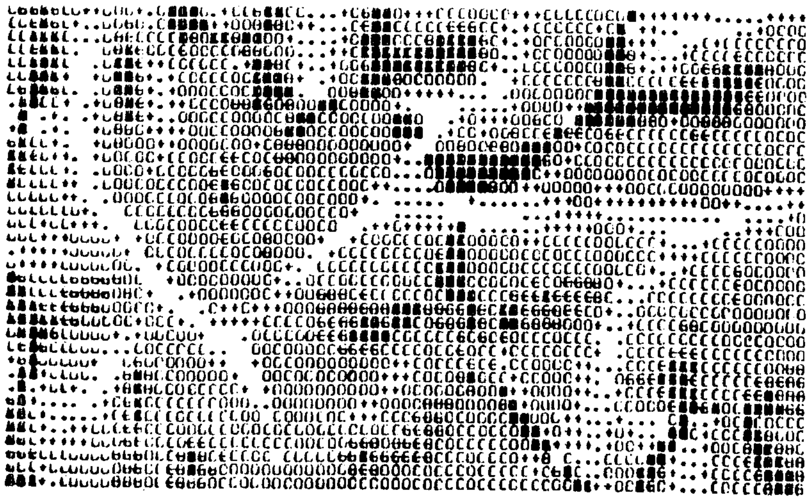
100 m

NCFOR TEST AREA (200 M. GRID SPACING)
PLAN CONVEXITY (DEGREES/100M) - BLOCK 1



200 m

NCFOR TEST AREA (300 M. GRID SPACING)
PLAN CONVEXITY (DEGREES/100M) - BLOCK 1



300 m

FIGURE 5.27
Effect of Grid Mesh on Plan Convexity

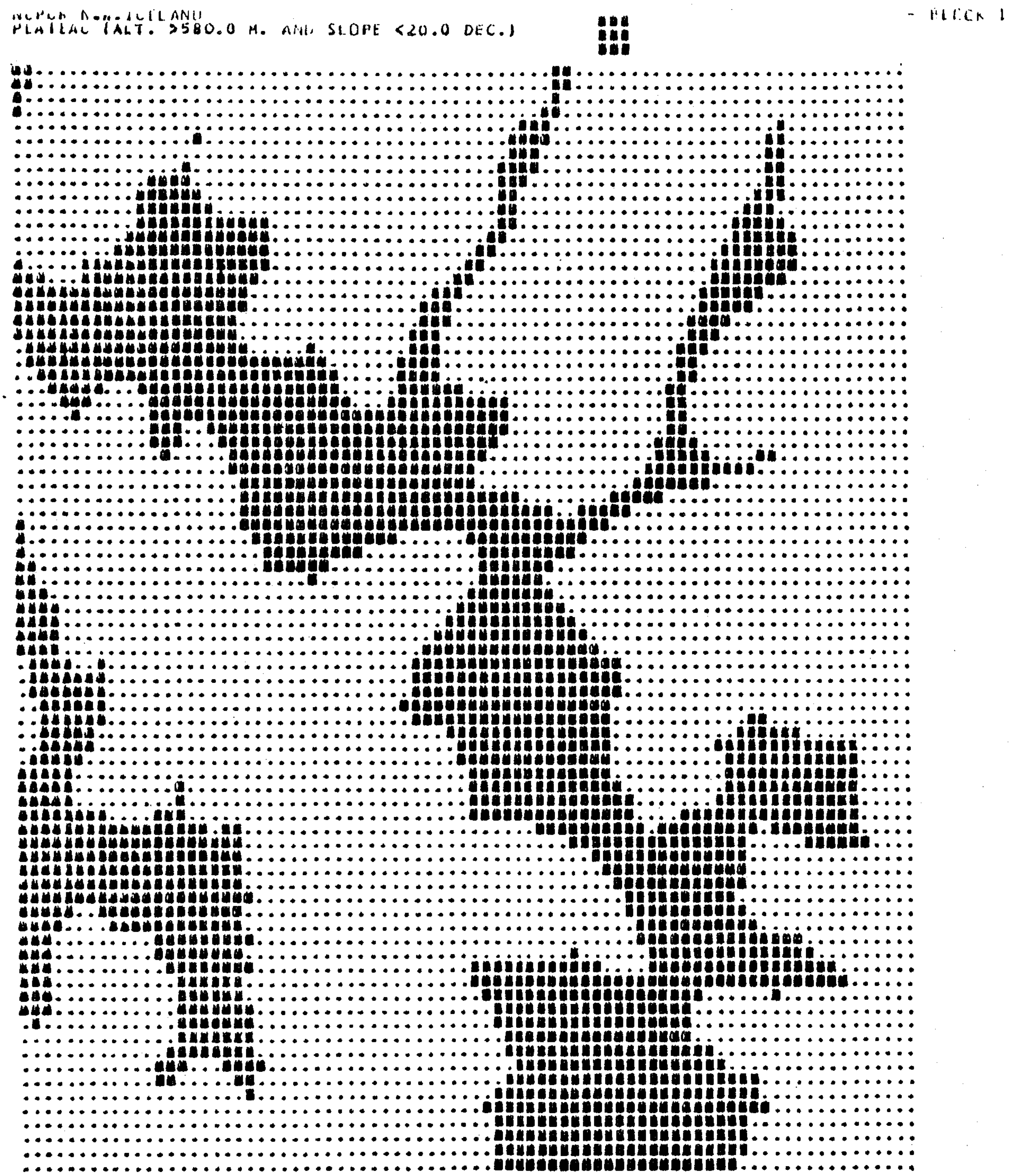


FIGURE 5.28
Nupur: Plateau Areas

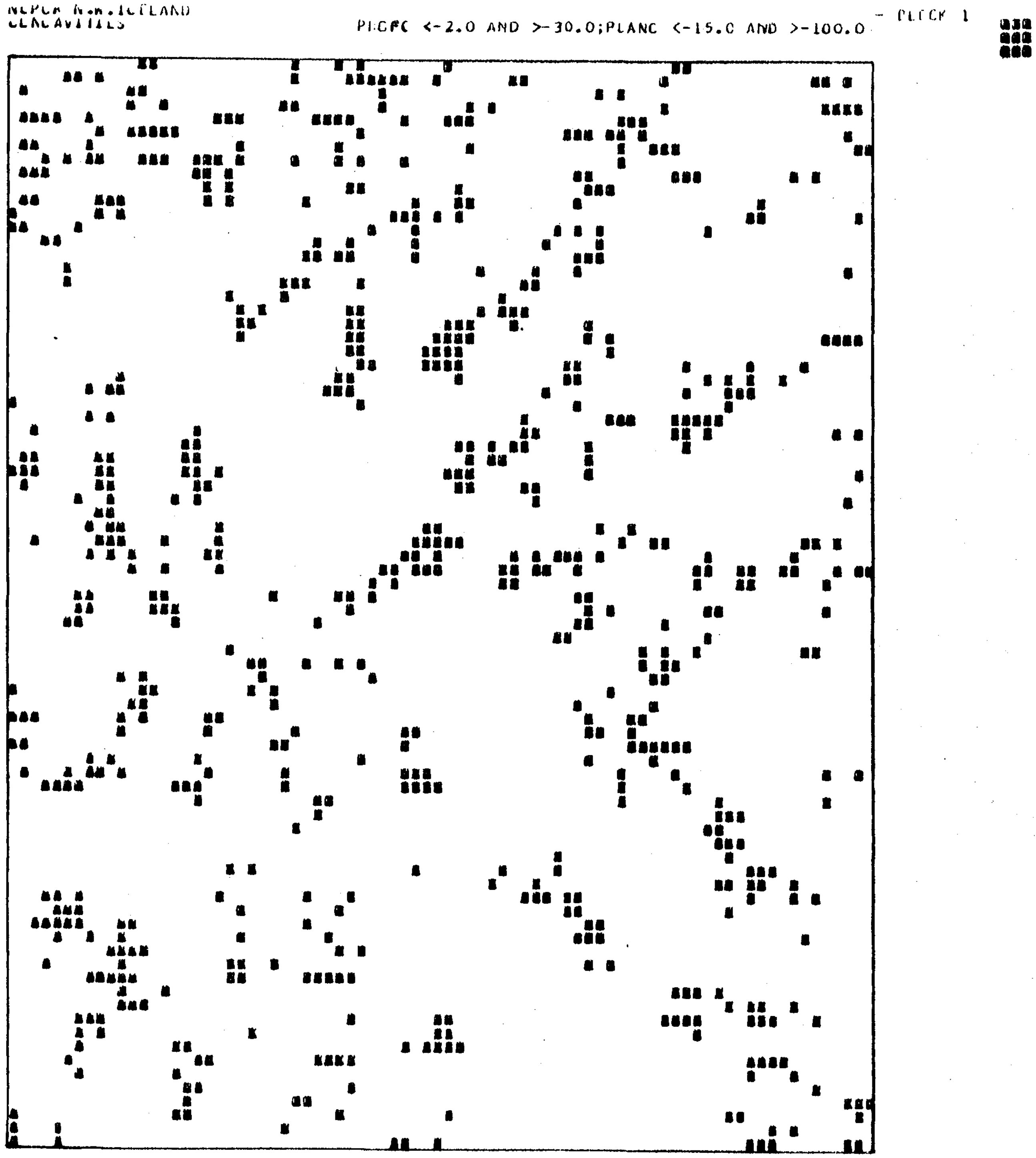


FIGURE 5.29
Nupur: Concavities

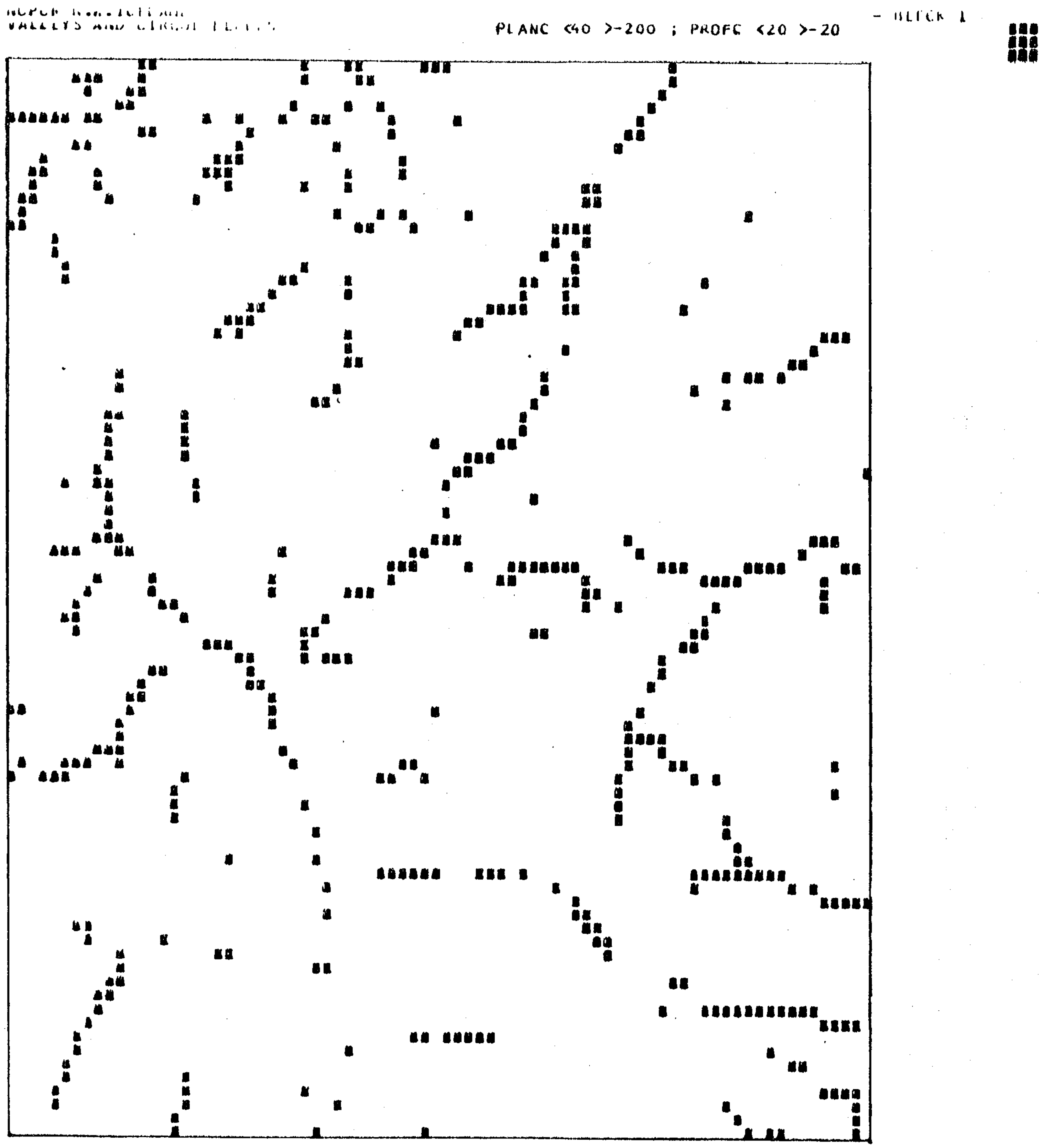


FIGURE 5.30
Nupur: Cirque and Valley Floors

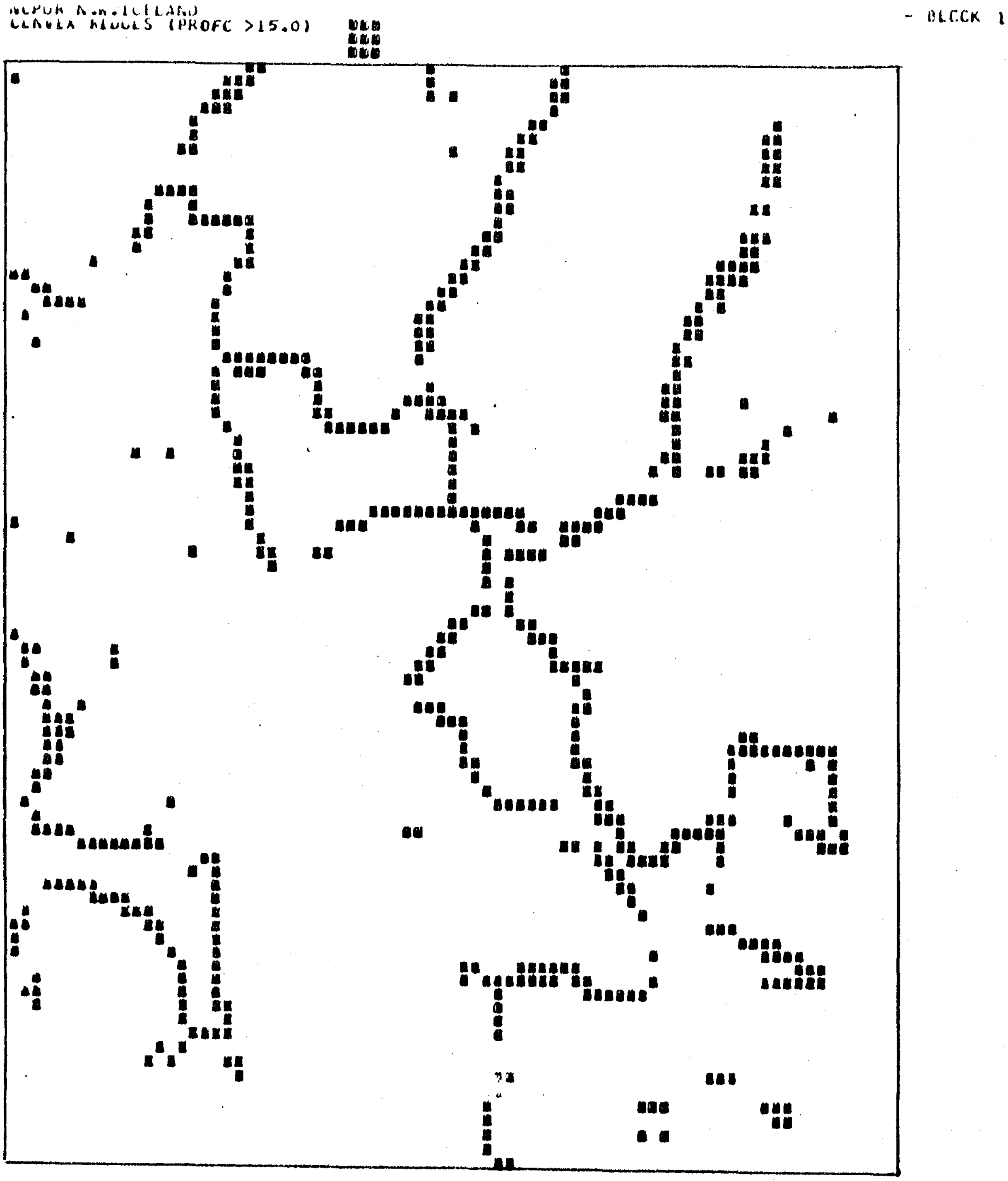


FIGURE 5.31
Nupur: Ridges and Plateau Edges

VALUES BETWEEN 25.000 35.000 45.000

NUPUR N.W.ICELAND
AVALANCHING SLOPES (25-35,35-45,>45 DEG.) - BLOCK 1

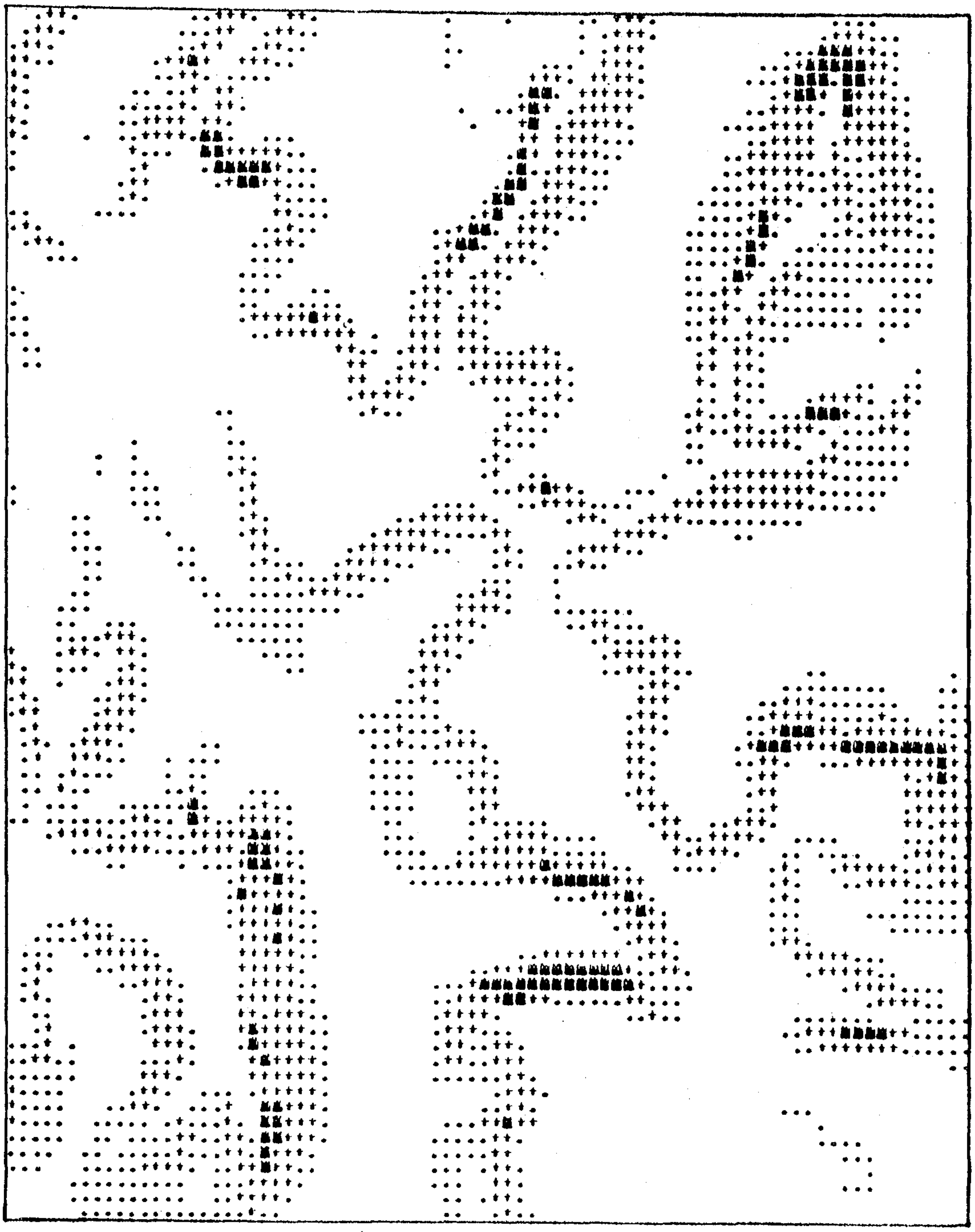


FIGURE 5.32
Nupur: Areas of Steep Slopes

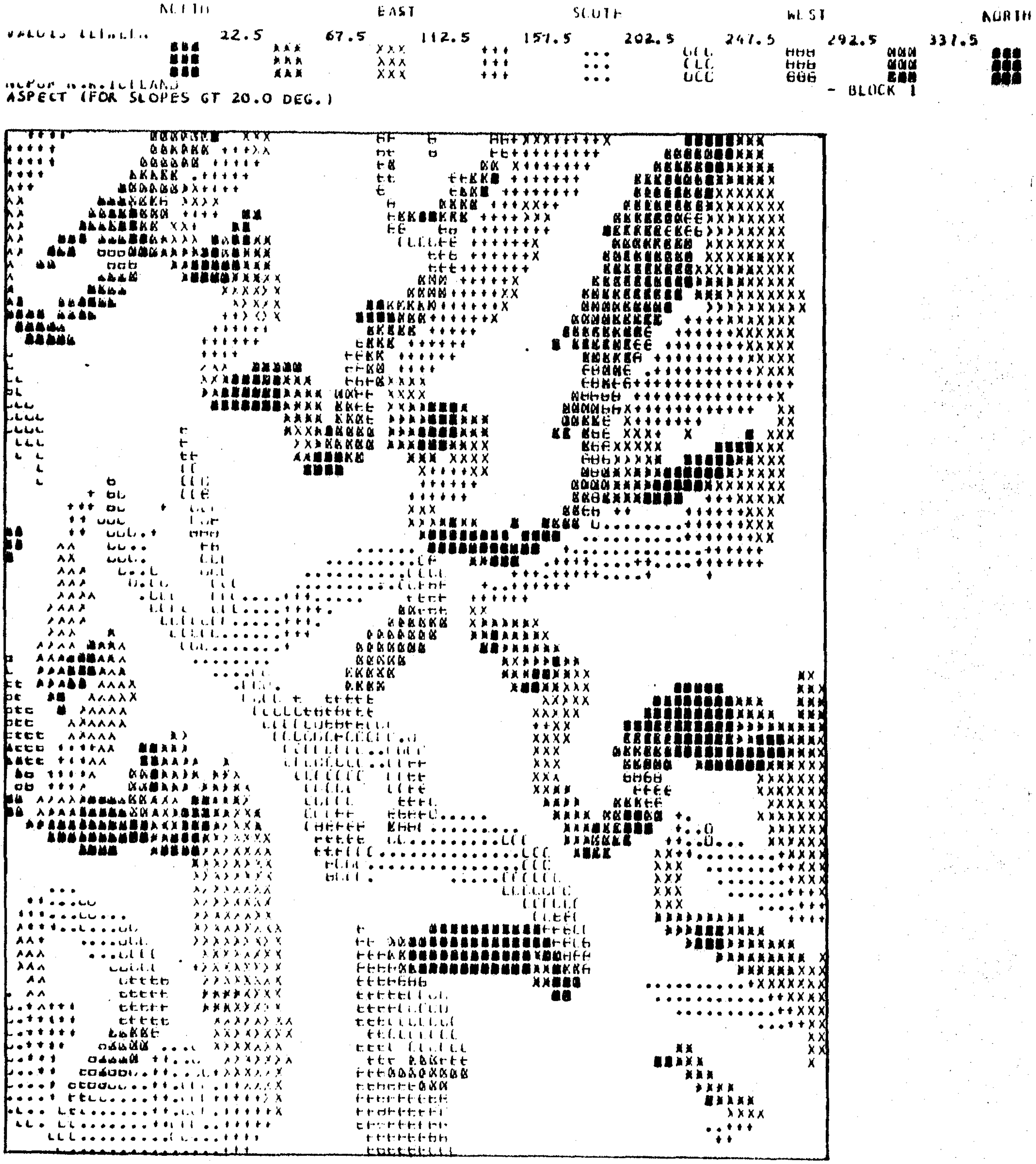


FIGURE 5.33
Nupur: Aspects of Steep Slopes

CHAPTER VI

CLIMATIC INPUT

The importance of climatic control in the determination of glacier mass balance is evident from the review of glacier-climate studies in chap. II and the discussion of factors affecting local glacier balance in chap. III. While the great volume of literature devoted to glacier-climate studies has enabled general cause and effect patterns to be developed, the nature of local spatial variations in the snow and ice cover remain little understood. This is largely a reflection of the paucity of relevant climatological data at the local-and meso-scales. The manner in which the local climatic input for the mass balance model can be estimated is described in this chapter, with reference to the Nupur area.

6.1 Choice of Basic Climatological Elements

While the availability of climatic records imposes an important constraint on the climatic input, studies of glacier mass balance (sec. 2.3.3)indicate that data on air temperature, precipitation, cloud cover, wind speed and direction are necessary for a meaningful study of climate - mass balance relations. The importance of these climatic elements in controlling glacier mass balance has been outlined in sec. 3.3.

The climatic input for the mass balance model is in the form of monthly means (averaged over a number of years) of the basic climatic elements. While data over shorter time intervals (e.g. days) are necessary for analysing the effect of 'weather' inputs on local mass

The seasonal variability in the climatic elements can be defined by fitting curves, through Fourier series, as has been done in Fig. 6.1[1]. The advantage of using such generalised curves is that the effect of changing climate on glacier balance can be considered by simply altering the amplitude or base level (a_1) of the curves.

Having obtained the regional climate, the next step is to estimate the climate data at all points of the altitude matrix. Since the base station is invariably located in a low-lying region, while the area of interest for mass balance study is characterised by mountainous terrain, such estimation presents serious problems. First it is necessary to examine the nature of climatological conditions prevailing in mountainous regions.

6.2 Local Climatic Conditions in Mountainous Regions

A consideration of the local climatic conditions in mountainous regions is fundamental for the determination of glacier mass balance (sec. 3.3). However, the study of mountain climates is beset with

---

[1]The curves fitted to the climatic data were defined according to a Fourier series given by the following relation:

$$y_x = \tfrac{1}{2}a_1 + \sum_{k=2}^{n} \{a_k \cos(\pi(k-1)(x-1)/n) + b_k \text{ Sin } (\pi(k-1)(x-1)/n)\}$$

where y is the particular variable being considered,
x is the month (1-12),
a_k and b_k are the Fourier Cosine and Sine coefficients.

numerous problems due to the immense variability of climatic elements and the availability of only limited and unrepresentative data.

The varying combinations of altitude, slope gradient and aspect result in the production of a wide range of local climates and microclimates in mountainous regions. Furthermore unfavourable weather conditions, sparse population, transportation problems and the physical isolation of such regions pose numerous difficulties in the selection, equipment and maintenance of observation stations. As a result there is a lack of sufficient meteorological stations at high altitudes. For example, only 3.9% of all meteorologicl stations in Iceland and only 4.1% in Britain are situated above 300 m. The few stations that do exist at high altitudes only observe a few of the climatic variables.

The adverse environmental conditions also lead to many measurement problems in the recording of individual climatic elements (e.g. discussed by Rowe, 1975). The analysis of mountain climates is further constrained by 'problems of scale', due to the "rapid changes in primary and secondary climatic controls and processes which occur over short distances" (Marcus, 1974b p.17). Lack of quantitative work on the scales at which these climatic controls act, makes the choice of any particular scale for the study of local climate variability rather arbitrary and difficult (sec. 5.5).

In view of these observational problems, our knowledge of mountain climates is based on the collection of rather fragmented and empirical evidence. Comprehensive compilations of studies concerned with the nature of local climates have been made, for example, by Geiger (1965, 1969) and Yoshino (1975). Many of the early studies cited by these authors were prompted by the close relationship between plant growth and local climate on mountain slopes (e.g. reviews by Cantlon, 1953; Ayyad and Dix, 1964).

Studies concerned with the temporal and spatial variation of the atmospheric environment in the surface layer, within a limited area are commonly presented under the term microclimate, local climate, or mesoclimate. However, since the topography of the land is one of the main factors that are more directly responsible for local differences in climate, the term 'topoclimate' (proposed by Thornthwaite in 1954) is preferable for studies of local climates in mountainous regions. Furthermore, while opinions differ, such studies generally involve a region of about 10 m. to an order of 10 km (Yoshino, 1975).

Apart from studies of local climate, further insight into the nature of highland climates is gained from the results of high mountain observations, for example in the St. Elias Mountains in Yukon and Alaska (Marcus, 1974 b), European Alps (Karrasch, 1973), and the New Zealand Alps (Coulter, 1967). In this respect, records from mountain stations which operated during the latter half of the 19th century continue to form an important primary source for the analysis of mountain climates, e.g. Ben Nevis in Scotland (Paton, 1954), Mt. Blanc and Sonnblick in the European Alps (Talman, 1934) and Mt. Washington and Pike's Peak in North America (Stone, 1934).

Information from high mountain observations can be complemented by that obtained from free-air soundings. The latter provide a useful indication of mountain winds and temperature conditions, especially for the climate of mountains for which no surface data are available (Coulter, 1967; Barry and Van Wie, 1974). However corrections should be made for any systematic differences between surface and free-air conditions.

All studies of mountain climates emphasise the topographic effect at the local scale in controlling the variability of climatic elements. For example, much of the contrast in air and ground temperatures with varying slope gradients and aspects can be related, in the first instance, to the different angle of incidence of solar radiation on sloping ground. Further modifications of temperature conditions can be due to the 'shadow' effect as a result of the local topography cutting off the direct solar radiation (sec. 8.3.1). The direction and speed of the wind is also affected by the local topography while the establishment of local wind circulations, as a result of topoclimatic contrasts, modify both temperature and precipitation régimes.

In spite of the numerous studies concerned with an examination of the climatic modifications due to topographic factors, our understanding of such effects is rather fragmentary. Most of the research is in the form of qualitative case studies of short duration, dealing with the climate of rather isolated and limited regions. Furthermore, little is known about the quantitative effect of such topographic features as cirques, which exert an important control on local climate and, in turn, glacier mass balance (Derbyshire and Evans, 1976).

To observe adequately the highly varied nature of the local climate in mountainous regions, would require the setting up of a station network of sufficient density, to allow the continuous areal recording of climatic elements on slopes in all directions and at different altitudes. Such systematic studies of mountain climates are lacking, largely because the considerable finance required to set up such dense networks of observing stations has not yet been provided.

The problem of defining the climatic input for a region of mountainous terrain can be overcome in part by the use of expected relationships between the climatic elements and basic terrain parameters, such as altitude, slope gradient and aspect. As a first approximation, much of the spatial variability in mountain climates can be related to the distribution of altitudes. For example, an increase of solar radiation, wind speed, precipitation and percentage of snow in total precipitation with altitude; and a decrease of air temperature with altitude are commonly observed.

The strong dependence of climatic elements upon altitude has, in fact, led to the identification and extension of climatic data for representative 'elevation bands'. However, the square-grid technique provides a better basis for the storing and correlation of climatologic and physiographic characteristics of an area (sec. 4.4.2).

Following the effect of altitude, further spatial modifications in local climate due to the influence of such factors as slope gradient, aspect, and exposure can be considered. In the following sections,then, an attempt is made to estimate point values of air temperature, snowfall and wind conditions over the Nupur matrix, using postulated relations between local topography and climate.

It should be noted that, in view of our present understanding of topoclimates, such estimates involve the use of rather crude generalisations. Furthermore, since our knowledge of local climates may vary greatly from one area to another, individual consideration of each study area is necessary for the development of topoclimatic relations. However, in the event of future advances in our understanding of the relationship between topography and local climate, the appropriate subroutine can be easily modified to accommodate such advances.

6.3 Estimation of Air Temperature

An estimation of monthly air temperature values at each grid intersection of the altitude matrix is needed in the mass balance model for determining: (i) proportion of precipitation falling in the form of snow (sec. 6.4); (ii) effectiveness of snow drifting, 'dry' snow being more important for drifting (sec. 7.2); (iii) suitable densification rates (sec. 7.4.2); (iv) potential periods during which snow and ice melt can take place (sec. 8.2); and (v) turbulent heat exchange (sec. 8.4).

The decrease of air temperature with altitude is well documented and it is generally recognised that it tends to follow a specific lapse rate (usually between 0.6°C/100 m. and 1.0°C/100 m.). However several investigators have shown that such lapse rates are subject to great variability depending on local pressure and circulatory conditions (e.g. Yoshino, 1966; Steinhauser, 1967).

Lapse rates usually vary throughout the year with different seasons. While the evidence for these seasonal fluctuations is varied and conflicting, generally lower mean monthly lapse rates are observed during the winter months as opposed to the summer months. Furthermore, the rates are found to differ with altitude, within the same mountain range (Steinhauser, 1967).

In the absence of high altitude stations, necessary to determine lapse rates for mountainous regions, it may be possible to use aerological observations from the free atmosphere (Coulter, 1967). It should be noted, however, that mountain temperatures are on average about 1-2^{o}C cooler than corresponding free air measurements in winter, and 1-2^{o}C above the temperature in the free air during summer (Samson, 1965; Ingham, 1967). These differences generally disappear at altitudes of about 1-1.5 km. (Coulter, 1967, Steinhauser, 1967).

Regarding the temperature input for the Nupur matrix, the monthly variation in the lapse rates was obtained from the curve in Fig. 6.2. This curve is drawn from data on upper air observations of air temperature at different levels up to 2000 m. for a period of 15 years, from 1954-1968 (Eythörsson and Sigtryggsson, 1971 p.16).

From Fig. 6.2 it can be seen that the lapse rate varies from a maximum of 0.7^{o}C/100 m. in May to a minimum of 0.55^{o}C/100 m. in January and February. The data also indicated some variation in lapse rate with altitude, the monthly lapse rates being slightly higher in the 50-1000 m. range than for 1000-2000 m. The curve in Fig. 6.2 is for the lower altitudinal range, since this corresponds with the altitudes found in the Nupur area.

The appropriate monthly lapse rate can be used to extrapolate the monthly mean air temperatures at the base station (Fig. 6.1a) to all points of the altitude matrix by use of the following relation:

$$t(x, y) = -L. \ a(x, y) + C$$

where

$t(x, y)$ is the mean air temperature at a point (x, y) in ^{o}C.

$a(x, y)$ is the altitude at the point (x, y) in m.

L is the mean monthly lapse rate ($^{o}C/m.$)

C is the base constant determined by:

$C = T + (L . A)$; T and A being the base station temperature and altitude respectively.

The altitudinal variation of air temperature needs to be modified by consideration of slope aspect.

Near surface air temperature variations according to slope aspects are essentially brought about by differences in ground temperatures which, in turn, reflect differences in the varying amounts of radiation received on sloping terrain. Microclimatological observations of air and ground temperatures with varying slope aspects have been made by a number of researchers (e.g. Aikman, 1941; Cantlon, 1953; Fuh, 1962; Steinhauser, 1967; Wymore, 1974).

Most of these studies emphasise the contrasts in air temperatures between north and south slopes. Fuh (1962), for example, in a study concerned with the effect of slope orientation on microclimate in the Nanking Mountains (China), found that the average air temperatures at 1·5 m. on the southern slopes to be 0·6°C greater than those on the eastern and western slopes, and about 1·0°C greater than those on the northern slopes. However data on slope gradients at the measurement sites were not given. More extreme contrasts were observed at lower levels (2·3°C at 5 cm above the surface), and with maximum air temperatures, (1–2°C at the 1·5 m. level and 5–8°C at the 5 cm. level), since the latter are more closely related to the radiation balance.

Studies of temperature contrasts with different slope aspects also reveal that the temperature maximum follows the sun from S.E. via S. to S.W. (in the northern hemisphere). Maximum temperatures are, thus, observed on the S.W. facing slopes, since the greater part of the solar radiation received during the morning is spent on the desiccation of soil; and it is only after the evaporation of the soil moisture (when the radiation is directly over the S.W. slopes) that the solar energy can be absorbed to raise the ground temperatures.

Given the greater importance of radiation in surface heating at higher altitudes, it can be expected that temperature differences between slopes also increase with altitude. However the few studies of slope differences in soil temperatures at varying altitudes (e.g. as reported by Barry and Van Wie, 1974) present rather conflicting evidence.

A seasonal variation in the temperature contrasts between slopes of differing aspects is also indicated by some studies. Steinhauser (1967), for example, in his topoclimatic survey of the Austrian Alps notes that north-facing slopes may be 0·8°C colder during the winter and $0{\cdot}3^{\circ}C$ during the summer than south-facing slopes on clear days. According to Steinhauser, this difference could be due to the fact that with greater radiation totals during the summer months, even the northern slopes are more heated than in winter relatively to the southern slopes.

Measurements by Cantlon (1953) on the Cushetunk Mountain (New Jersey) indicate a more marked seasonal variation between north- and south-facing 20 degrees slopes. Fig. 6.3 is drawn from the data provided by Cantlon (Table 2, p.248) on the monthly averages of mean temperatures, at various levels above the surface, sampled on the north and south slope stations. It depicts the magnitude of temperature differences between these slopes at the 5 cm. and 2 m. levels. It can be seen that the temperature differences are greatest near the surface. Furthermore, the differences are greater during autumn (about $4{\cdot}5^{\circ}C$ at 5 cm. and $1{\cdot}2^{\circ}C$ at 2 m.) and least during the summer. To some extent these differences can be explained by changing vegetational patterns (leaf fall and expansion periods).

The majority of studies concerned with examining the variation of air temperature with slope aspect are of short durations and have been conducted in rather scattered localities, making any generalisation difficult. However it is desirable to modify the altitude-adjusted air temperature values for differences due to slope aspect.

In the absence of definite air temperature - slope aspect relationships, it is necessary to postulate the probable relations.

For example Wymore (1974) provides aspect temperature correction factors (Fig. 6.4) for use in the Piceance Creek Basin, Colorado. These were developed by the use of minimum tempeature differentials that would provide a logical distribution of evapotranspiration rates in combination with variations in solar radiation rates using the modified Jensen-Haise method (described in Wymore, 1974).

Based on the above discussion and other specific aspect adjustments suggested by research on soil and air temperatures (e.g. Geiger, 1965; Yoshino, 1975) postulated seasonal air temperature corrections for slope aspect, at the 2 m. level, are applied to the altitude-adjusted temperatures at all grid intersections of the altitude matrix (Fig. 6.5). Though data are very limited, some researchers (e.g. Fuh, 1962) have pointed out that these postulated contrasts in air temperature with varying slope aspects are only evident on steep slopes. Thus the aspect-corrections are applied to well-defined slope aspects (above a threshold gradient of 20 degrees) which highlight major topographic elements, such as the cirque and valley walls in the central Nupur area. (Fig. 5.25). A lack of any detailed information on air temperature contrasts on slopes with varying aspects and gradients prevented a more comprehensive treatment of local air temperature adjustments.

These considerations of air temperature estimation are programmed in the subroutine MTEMP. Results from MTEMP depicting the spatial variation in the May and July air temperatures in the central Nupur area can be seen in Fig. 6.6.

6.4 Estimation of Snowfall

An important input requirement for the simulation of glacier mass balance is the initial snow cover distribution due to snowfall. The incidence of snow in highland areas is, however, particularly sensitive to the interplay of air circulation, temperature and precipitation patterns as modified by topographical effects. The resulting complexity makes the specification of snowfall distribution very difficult. Furthermore, precipitation (especially snowfall) measurements used to represent conditions at a point may be subject to considerable errors (e.g. Peck, 1972; Rechard et al., 1974; Larson and Peck, 1974).

The approach used in this section is to estimate monthly point values of snowfall, for all grid intersections, using expected relationships between regional precipitation, local air temperatures and local topographic parameters.

While precipitation distribution, especially over short time periods, is more directly related to meteorological characteristics (e.g. atmospheric stability, amount of precipitable moisture, cloud cover, wind flow, etc.), a lack of observational data on these aspects has led to an interest in estimating precipitation amounts from topographic considerations. A number of researchers have shown that it is feasible to describe the distribution of long-term (i.e. monthly and annual) spatial variations in precipitation by developing statistical relations with various topographic parameters (e.g. using regression analysis or graphical correlation techniques).

Spreen (1947), for example, conducted a notable study relating altitude, rise, exposure, orientation and position to the distribution of monthly and annual precipitation in western Colorado. Similar studies have been carried out by Burns (1953) in California, Schermerhorn (1967) in western Washington and Oregon, and Hutchinson (1968) in Otago, New Zealand. In general these studies indicate that a consideration of altitude together with aspect and exposure to rain-bearing winds is necessary to adequately specify the precipitation distribution in mountainous areas.

The important influence of altitude on precipitation distribution is particularly well documented. In addition to the above-mentioned studies, numerous workers have considered the correlation of precipitation with, both spot altitudes (e.g. Donley and Mitchell, 1939; Lull and Ellison, 1950; Smallshaw, 1953; Dickson, 1959; Rodda, 1962; Peck and Brown, 1962; Curry and Mann, 1965; Unwin, 1969; Bleasdale and Chan, 1972) and mean altitudes (e.g. Schermerhorn, 1967; Gow and Lockwood,1974). These studies indicate that precipitation - altitude relationships can be regarded as linear for the most part.

However, with the quantity of moisture in the atmosphere decreasing with increasing altitude, a zone of maximum precipitation may be expected. The evidence from temperate areas is not conclusive and the precipitation is usually found to increase progressively up to the mountain peaks. In the humid tropics, however, there appears to be more evidence for an upper maximum limit below the highest peaks (de Boer, 1950; Weischet, 1969).

Some studies further indicate that the precipitation - altitude gradients may vary seasonally due to differences in precipitation type (e.g. Lull and Ellison, 1950; Rodda, 1962; Curry and Mann, 1965; Hutchinson, 1968). In particular, the relationship may become weaker or break down in areas where the rainfall type is convectional for the main part. However, the latter usually forms such a minor component of mountainous precipitation regimes that mean gradients with altitude are scarcely affected at all. For example, Ahlmann (1927) found that 93% of even the summer precipitation in the Horung Massif, Jotunheim was due to cyclones with a strong orographic influence; while the importance of cyclones for winter precipitation has been commonly observed (e.g. Burns, 1953; Hutchinson, 1968). The insignificance of convention as an agent for vertical air movements during the winter months means that it will not have an important influence for snowfall estimation.

Regarding the precipitation variation in the Nupur area, use was made of the relationship between annual precipitation and altitude shown in Fig. 6.7a. Due to a lack of high altitude precipitation stations in Vestfirðir, this relationship was developed from an area of similar annual precipitation regime and altitudinal range in south-west Iceland, viz. the area around Lake Hvalvatn (N.E. of Reykjavik). This area has the most intensive network of precipitation stations in Iceland and records from 14 stations, covering the period 1950-1971, were available from the Icelandic Meteorological Office (Fig. 6.7 b).

The correlation coefficient for the precipitation-altitude relation was found to be 0·72; thus altitude accounts for 52% of the variation in annual precipitation. Furthermore, this relationship indicates an annual increase of about 132 mm. for every 100 m. rise in altitude. Taking the long-term mean annual precipitation at the base station in the Nupur area (994 mm at Thorustaðir, altitude 20 m.), the annual precipitation at all grid intersections in the Nupur matrix can be estimated from the following relationship:

$$p(x,y) = 1{\cdot}32\ a(x,y) + 967{\cdot}6$$

where

p(x,y) is the mean annual precipitation at a point (x,y) in mm.

and a(x,y) is the altitude at the point (x,y) in m.

The altitudinal effects on the precipitation distribution are modified by the secondary influence of slope aspect. The importance of slope aspect and exposure, in relation to the prevailing wind direction, has been stressed by most of the studies concerned with an examination of the topographic influence on precipitation distribution (e.g. Spreen, 1947; Burns, 1953; Hutchinson, 1968).

Slope aspect effects on precipitation distribution operate at both broad and local scales. At the broad scale, differences can be expected to arise btween windward and leeward slopes, with the former receiving higher precipitation due to the effects of orographic lifting. It should be noted,however, that on steep mountain ridges the 'spillover' effect can result in greater precipitation on leeward slopes (Hovind 1965). Furthermore, at the local scale, these differences in precipitation distribution may be modified by the effects of wind drifting, especially in the case of snow (sec. 7.2).

In the absence of any long-term observations concerning the effect of slope aspect on precipitation distribution, the nature of this influence is postulated from common sense, though rather simplified and arbitrary, considerations. For example, an indication of the expected windward-leeward precipitation differential may be provided by a consideration of the frequency of winds from different directions (e.g. Reid, 1973). While slopes facing the prevailing winds can be expected to receive higher amounts of precipitation than those which are leeward, the problem lies in the fact that annual precipitation data may include precipitation occurring with a number of wind directions.

This is especially the case in Iceland, for example, where the moisture content of winds from all directions is high due to the strong maritime influence. In view of this, it is assumed that the amount of precipitation received with winds from a particular direction is proportional to the frequency with which the winds blow from that direction. Furthermore, to account for the variation of precipitation with slope aspect, it is assumed that twice as much precipitation falls on the windward slopes compared with those leeward to the prevailing wind direction.

Given the regional distribution of the prevailing winds in the Nupur area (sec. 6.5), the postulated differences in the percentage of precipitation received by slopes of different aspects can be determined (Fig. 6.8). For example, if north-westerly winds blow for 25% of the time, then it is postulated that 16.7% of the precipitation will fall on slopes facing north-west and 8.3% on the south-east facing slopes. Similarly, if the winds from south-east blow for 14% of the time, then slopes facing south-east are expected to receive 9·3% of the precipitation and those which are facing north-west will receive 4·7%. The appropriate windward and leeward components of the precipitation from different wind directions are then summed to give the net precipitation expected with any slope aspect (e.g. 21·4% for slopes with N.E. aspects and 17·6 for those facing S.W. in the above example).

The point precipitation values calculated by use of the precipitation altitude relation are considered to be applicable for the theoretical case in which the frequency of winds from different directions is the same (i.e. 12·5% from each of the eight main directions considered). Any departure from this theoretical value and the postulated percentage of precipitation according to the actual frequency of wind direction and slope aspect considerations, is then taken as the correction for the altitude - determined precipitation value (Fig. 6.8).

Since these aspect -corrections are more relevant for the major steep slope units, they are only applied to slopes with gradients of over 20^{o}, thus excluding minor variations in slope aspect (as in the case of air temperature - estimation). Furthermore, in order to exclude points which, due to the effects of local topography, may not be exposed to winds from particular directions, the aspect correction is only applied to those points which do not have any land higher, than the point in question, for a distance of 1 km. in the particular direction of the wind being considered. This indication of exposure in different directions is determined by the subroutine TPGW.

Following the estimation of the total annual precipitation in the above manner, the monthly precipitation at each grid intersection can be obtained by dividing the annual precipitation according to the expected percentage contribution of each month. Fig. 6.9 depicts the curve for the mean contribution of each month to the annual precipitation, as determined from long-term precipitation records at Thorustaðir.

Finally, the amount of snowfall in the monthly precipitation totals needs to be determined.

Daily maximum and minimum air temperatures, in relation to a base temperature, have been commonly used to determine the form of precipitation (e.g. Willen, et al., 1971; Leaf and Brink, 1973; Auer, 1974). Estimation of the precipitation form for monthly totals is more difficult.

Bogdanova (1976) presented a method for calculating solid precipitation amounts from monthly precipitation totals. She noted from numerous empirical measurements, that the relationship between monthly solid precipitation (P_s) and mean monthly temperature (T) can be approximated by a linear equation of the form:

$$P_s = a - b\,T.$$

This relationship varies regionally according to the following parameter:

$$\delta = \left(\frac{t_{max} - t_{min}}{t_{ann} + 50}\right) t_{ann}$$

where t_{max} is the mean temperature of the warmest month.

t_{min} is the mean temperature of the coldest month.

and t_{ann} is the long-term average of the annual temperature.

The coefficients, a and b are related to δ by empirically determined curvilinear relationships for the range of values encountered by Bogdanova (Fig. 6.10).

Now for Thorustaðir, with the mean temperature of warmest month, July, being 10·5°C and that of the coldest month, February, being -1·6°C, while the long-term (1931-60) mean annual temperature is 3·6°C, the value for $\delta = 0{\cdot}813$; $a = 33{\cdot}246$; $b = 5{\cdot}565$. Thus, the percentage of solid precipitation (0-100%) for any month, at any grid intersection in the Nupur matrix, can be determined from the following relation:

$$P_s = 33{\cdot}246 - 5{\cdot}565\ T.$$

(valid for T values in the range -12 to 5·9°C).

A limited check on the computation from monthly total precipitation totals was possible using snow depth measurements for 1976 at Akureyri and Thorustaðir, provided by the Icelandic Meteorological Office. The monthly estimations of snow according to this method are compared with the actual measurements of snow depths in Fig. 6.11. It can be seen that, apart from the apparent overestimation of the snow depths for February and November at Thorustaðir, the comparison is quite good. The difference between the computed and measured snow for these two months appears to be due to an unusually large proportion (75-85%) of measured precipitation for these months being classified as 'sleet'.

The above considerations for the estimation of monthly snowfall are programmed in the subroutine MSNOW. Results of MSNOW for January and October snowfall over the central Nupur matrix are shown in Fig. 6.12.

6.5 Estimation of Wind Conditions

Of all the climatological elements, wind is probably subject to the greatest variations in mountainous regions. A variety of local wind systems may be generated by orographic, frictional, gravitational, and thermal effects (Geiger, 1965; Flohn, 1969; Yoshino, 1975). The great variability in wind conditions is a consequence of both the passive effects of the landscape, resulting in variations of wind speed, direction, and turbulence; and active effects due to local contrasts in the energy exchange at the surface which establish thermally driven local circulations.

In view of this complexity, the modelling of point wind conditions in an area is a major and problematic undertaking which is outside the scope of this study. Some indication of wind conditions is, however, necessary for the estimation of glacier mass balance. In particular, the regional distribution of wind directions is required for determining variations in the precipitation distribution (sec. 6.4) and for the direction of snow drifting from the summit plateau areas (sec 7.2.2). Also an indication of the variation in mean wind velocities during the summer months is needed for the calculation of turbulent heat exchange (sec. 8.4).

The regional pattern of wind directions in the Nupur area was obtained from the long-term (1931-60) records at the base station (Thorustaðir). Both annual (for precipitation estimation) and winter (for snow drift estimation) mean wind directions were determined (Figs. 6.8 and 6.13 respectively). In the absence of meteorological stations on the plateau summits, some confidence in the use of the records at Thorustaðir to indicate regional wind directions is provided by the general consistency of wind directions in N.W. Vestfirðir. Long-term records from Suðureyri (Fig. 6.13) and short-term records from other climatological stations in the region indicate that winds from the sector between N.E. - S.E. are particularly frequent.

With regard to the estimation of mean wind speeds, the scarcity of data on wind speed variations in mountainous regions necessitates the use of rather general relationships between wind speed and topography. In the first instance, it is commonly observed that wind speeds increase with altitude (e.g. Manley, 1952, p.188; Yoshino, 1975, p.233).

In the absence of field measurements, the variation of wind speed with altitude can be obtained from aerological data (Coulter, 1967; Marcus and LaBelle, 1970). However, corrections according to the effects of topography need to be applied to the free-air estimates of wind speeds. In general, summit windspeeds may be quite closely correlated with those in the nearby free atmosphere; though in certain instances, with light winds, the wind speed on mountain peaks may be less than that observed in the free air (Marcus and LaBelle, 1970; Barry and Van Wie, 1974). With strong winds, however, the orographic barriers tend to converge the flow of air, causing higher wind speeds to be observed on exposed mountain summits and ridges.

The effect of sheltered localities on wind speed reduction is more consistent. For example, Derbyshire and Blackmore (1974) report that wind velocities in cirques can be reduced by as much as 50-70% compared with free-air measurements.

In general, wind speeds are found to be greater over exposed convex areas than in sheltered concave areas. It appears, then, that an approximate correction based on consideration of the degree of exposure at any grid point could be applied to the wind speeds determined in the free-air.

The manner in which the variations in summer wind speeds, (for use in turbulent heat exchange calculations), over the altitude matrix were estimated is discussed with reference to the Nupur area. First an approximate relation of summer wind speed with altitude was developed using aerological data. In Iceland a number of measurements of wind speed variation in the free air have been made at Adalvik, Akureyri, Reykjavik and Keflavik. Although these measurements are for differing and rather short durations, a general trend of increasing wind speeds with altitude can be discerned (Fig. 6.14). The mean gradient of wind speed increase with altitude is taken as 0·142 ms^{-1} per 100 m. Using this gradient and the appropriate long-term, monthly wind speed value at the Nupur base station, Thorustaðir (Fig. 6.1d), the mean monthly wind speed at all points of the matrix can be determined.

The wind speed values determined from free-air observations are then modified for the effect of local topography by consideration of the degree of exposure present at all grid intersections of the altitude matrix. The exposure at any grid point is determined by examining each successive point (within a radius of 1 km) for all directions of wind flow, in turn, to see if there is any land higher than the point in question. The monthly wind speed is then reduced according to the degree of exposure in relation to the frequency of the winds from different directions. Thus, if the point lies on the summit plateau and is totally exposed to all directions of wind flow, then there is no reduction in wind speed from the value determined by free-air considerations; if, for example, the point is blocked by higher land in the north (summer wind frequency of 6%) and north-east (summer wind frequency of 23%) the mean velocity is reduced by 29%. In this manner, it is found that major concavities such as cirques, which may be sheltered from winds in three cardinal directions, will experience wind speed reductions of 60-70%.

The computations for wind speed reduction according to exposure and regional wind direction need only be done once for a particular area. In the program GSP1, they are, thus, carried out by the subroutines WEXP and TPGW at the start of the program, prior to the calculations for monthly climate and mass balance. The results from these subroutines are stored in the matrix WRED (I, J). The variation of mean wind speeds for January and July over the central Nupur matrix are shown in Fig. 6.15.

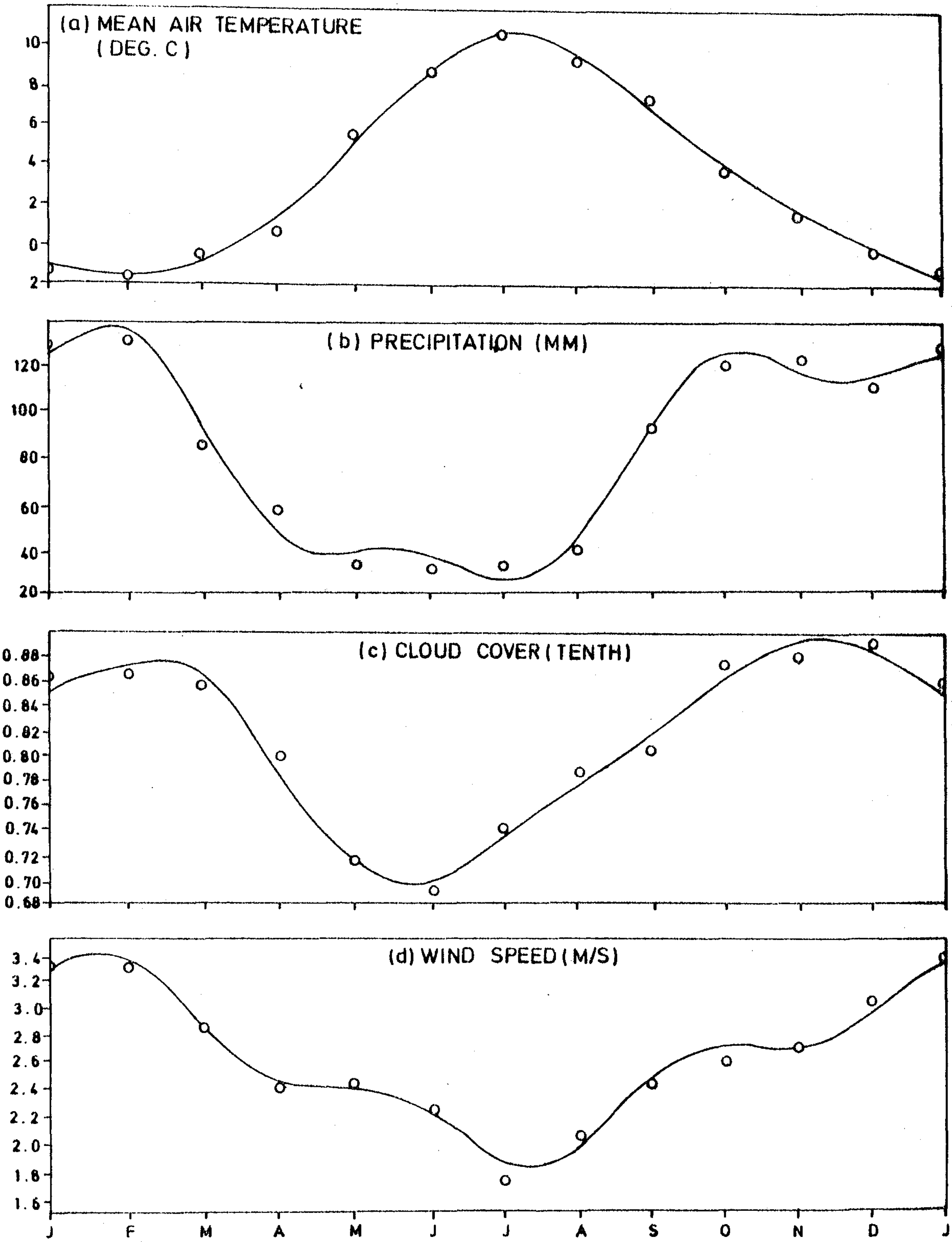


FIGURE 6.1
Variability of Basic Climatological Elements at Thorustaðir (1931-60). Source: Veðrattan, 1931-60. Monthly and yearly summaries from the Icelandic Meteorological Office.

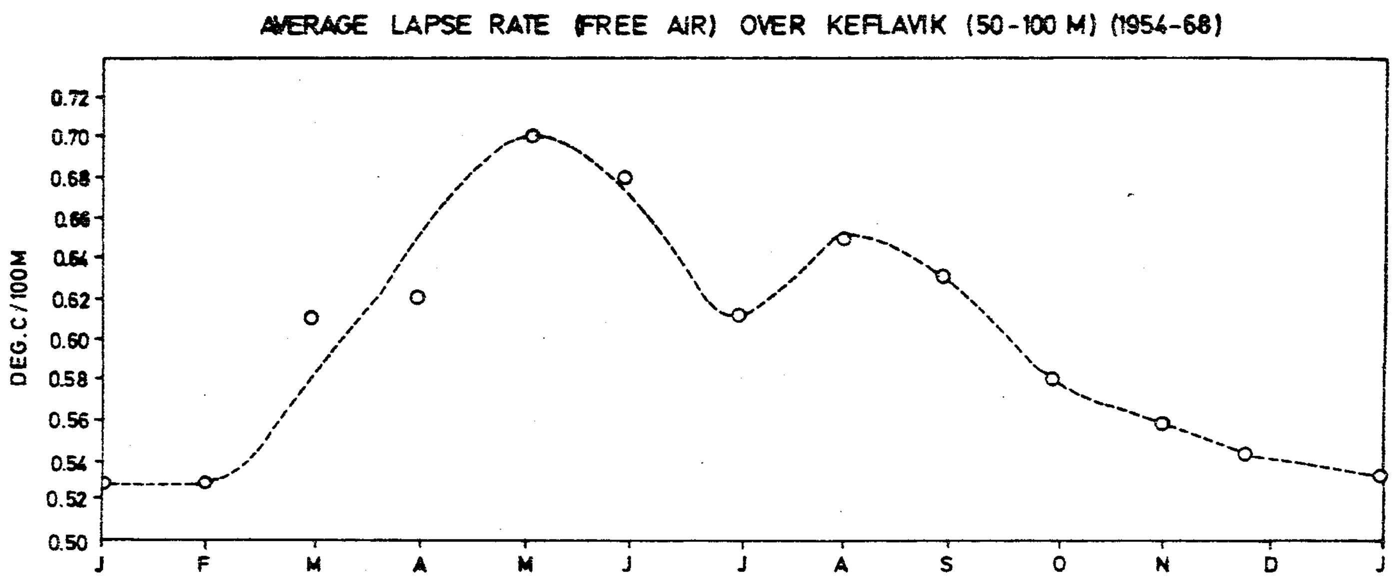


FIGURE 6.2
Variation of Monthly Lapse Rate (Free Air) Over Keflavik, Iceland.
Source: Eythórsson and Sigtryggsson (1971 p.16).

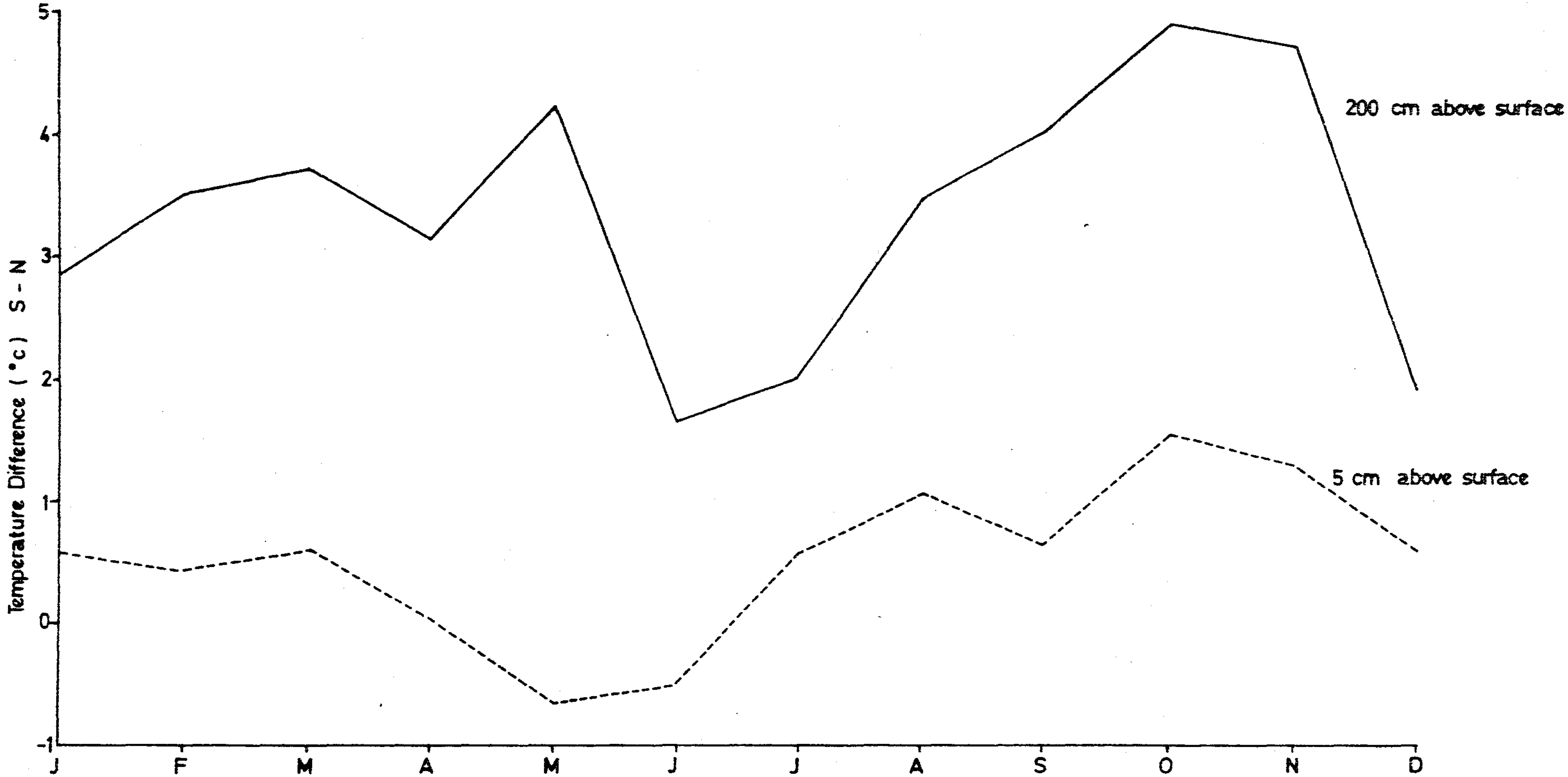


FIGURE 6.3
Seasonal Variation of the Temperature Difference(°C) between South and North Slopes on the Cushetunk Mountain, New Jersey. Data provided by Cantlon (1953 p.248).

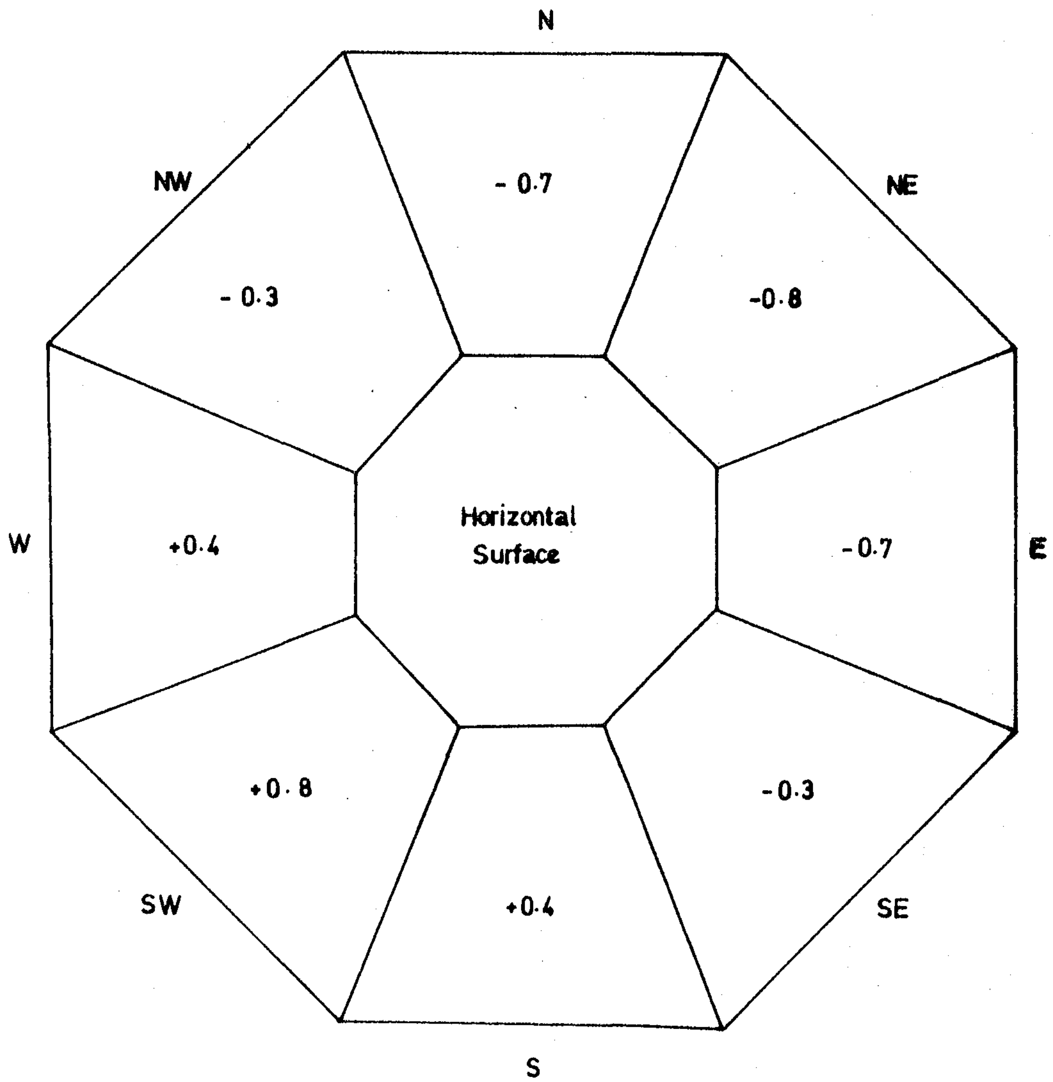


FIGURE 6.4
Modification of Mean Monthly Air Temperature (^{o}C) Due to Slope Aspect as Suggested by Wymore (1974) for Use in the Piceance Watershed, Colorado.

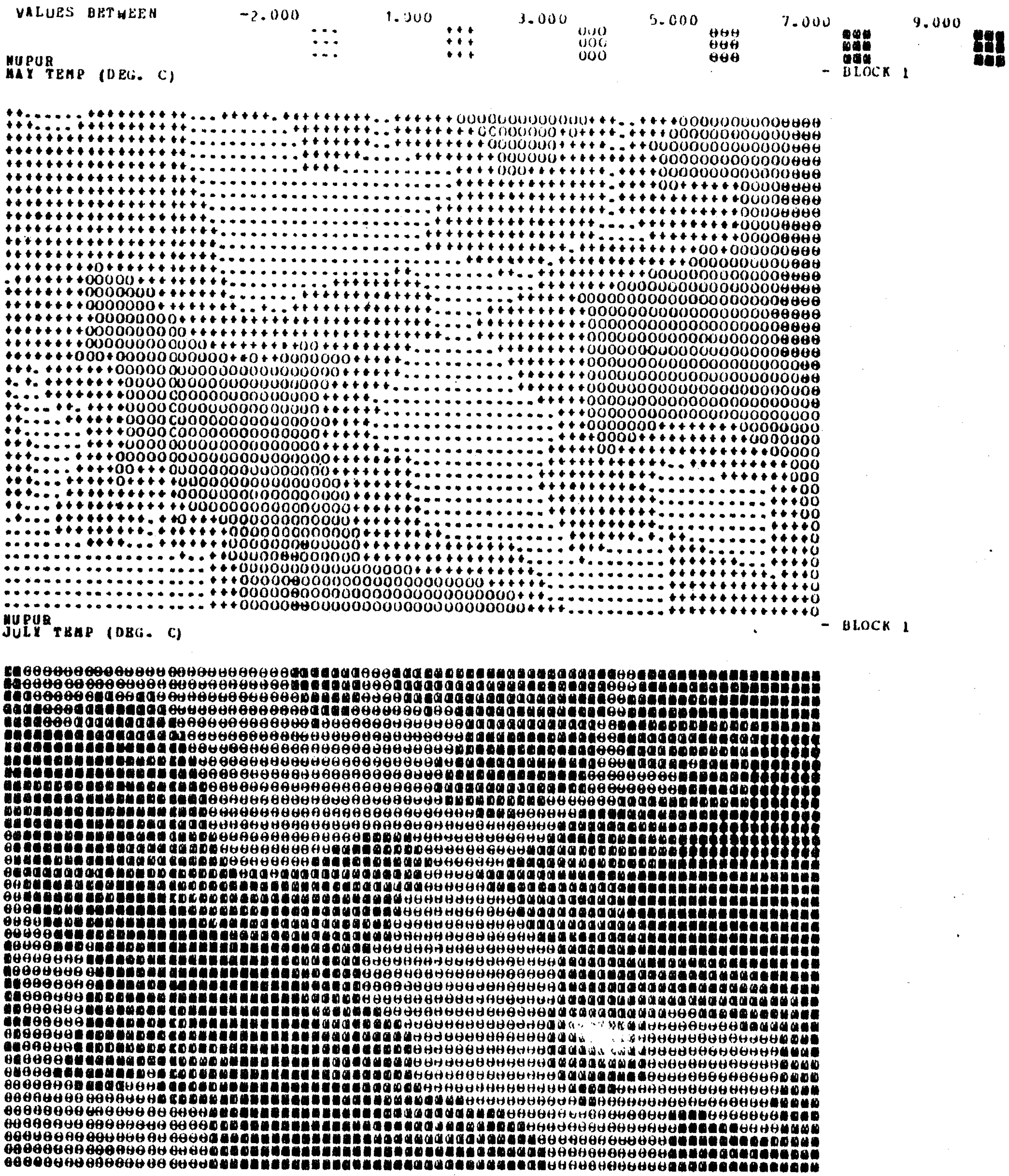
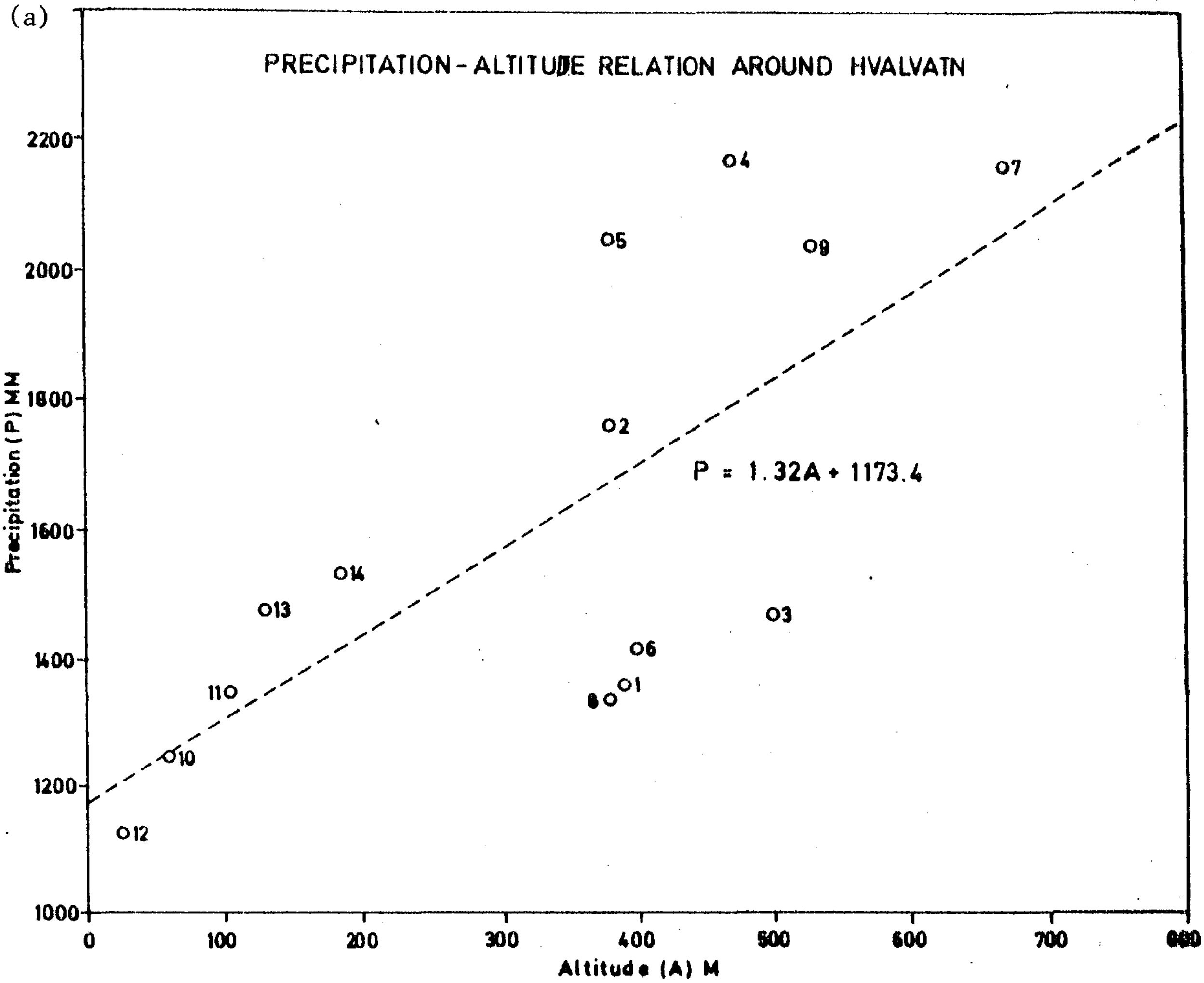


FIGURE 6.6
May and July Air Temperatures Over the Central Nupur Matrix as Determined by the Subroutine MTEMP.

(a)



(b)

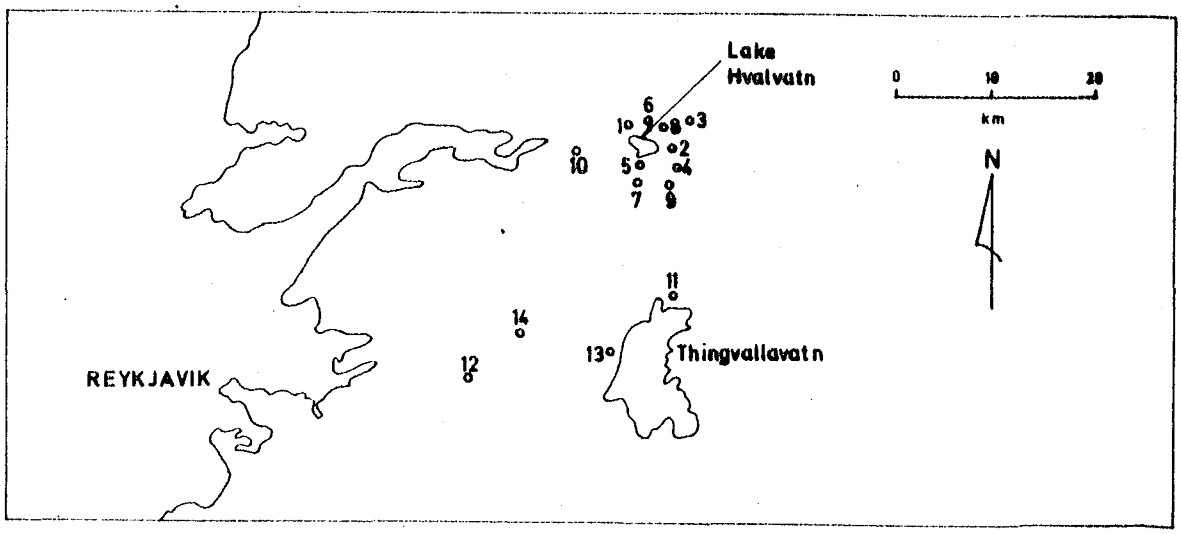


FIGURE 6.7

(a) Precipitation - Altitude Relation in the Area Around Lake Hvalvatn (S.W. Iceland).

(b) Location of Precipitation Stations.

Source: Vedrattan (1950-1971). Icelandic Meteorological Office.

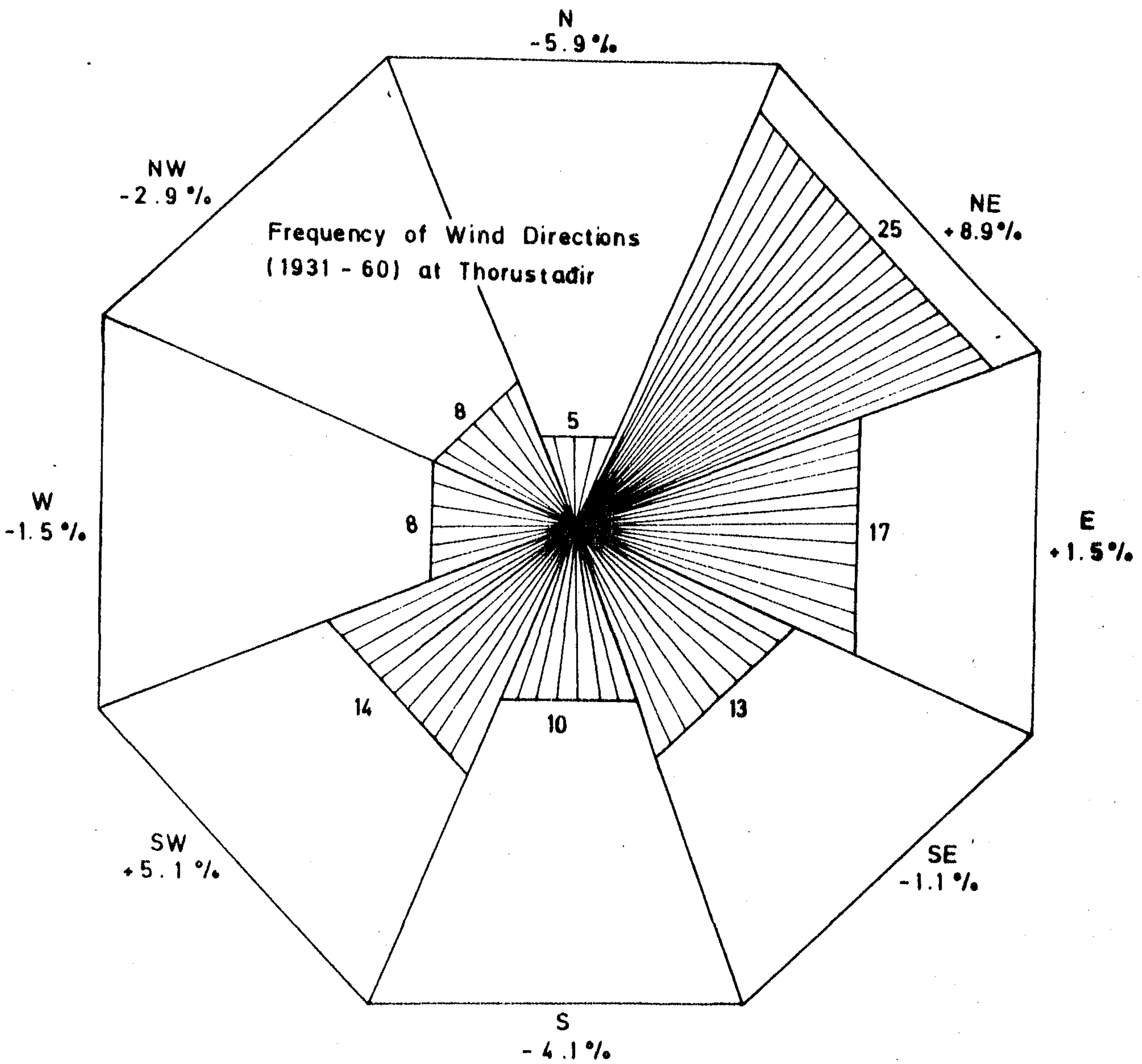


FIGURE 6.8

Modification of Annual Precipitation Amounts According to Slope Aspect in Relation to Prevailing Wind Direction for the Nupur Area. (see text for explanation).

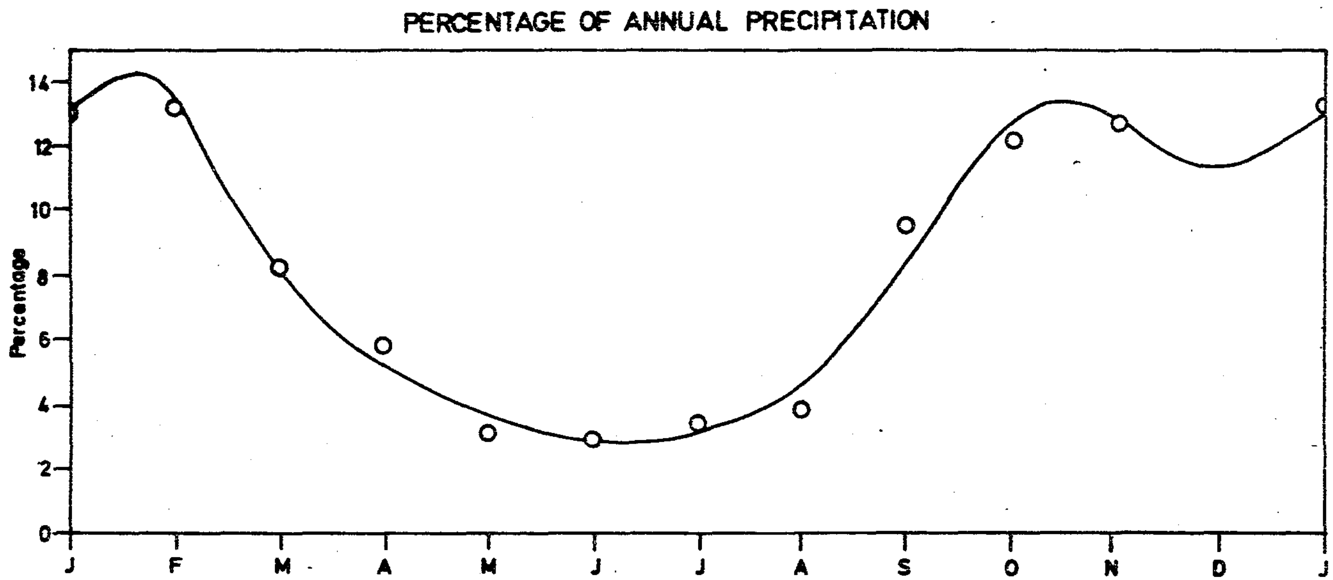


FIGURE 6.9
Mean Monthly Contribution to the Annual Precipitation at Thorustaðir.
Source: Icelandic Meteorological Office.

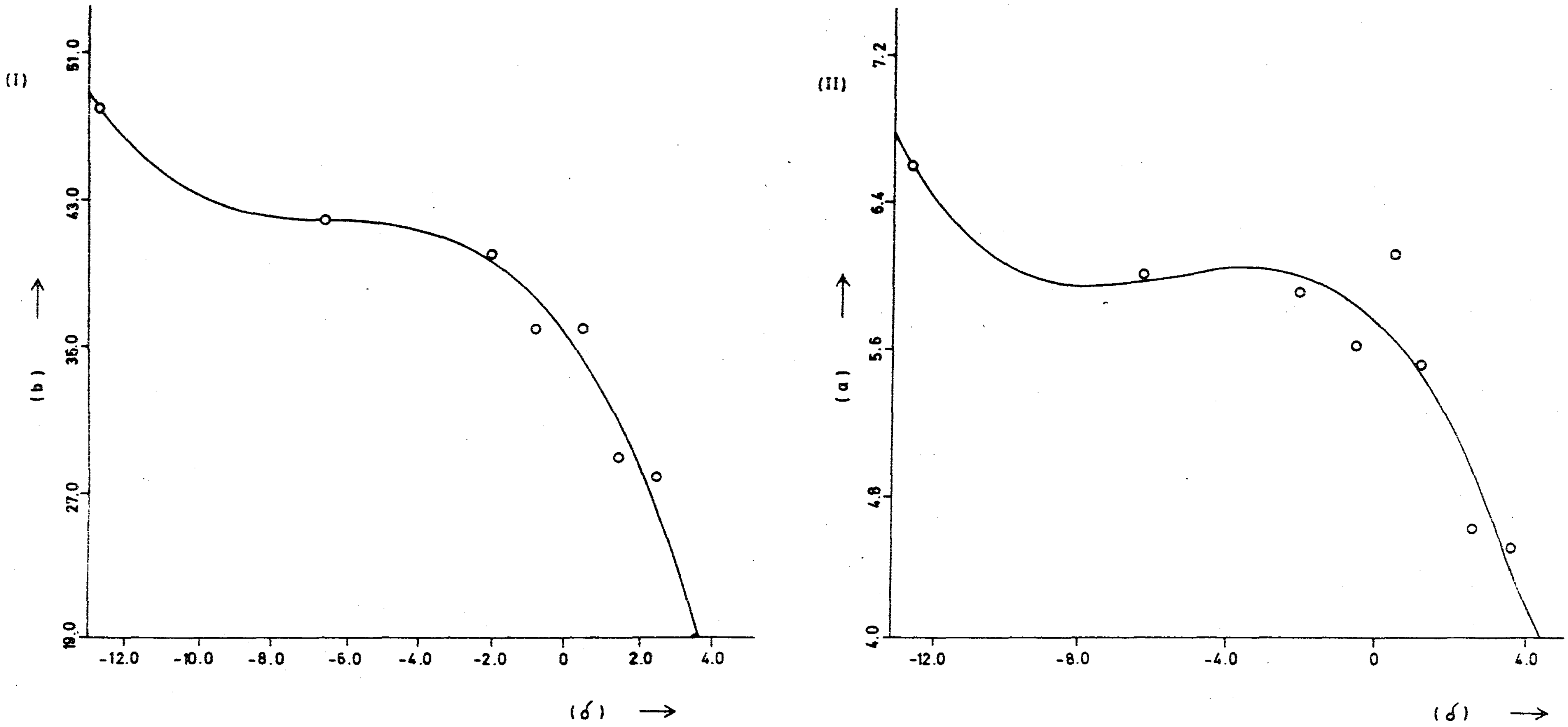


FIGURE 6.10
Relationship Between Coefficients a and b to δ . Data from Bogdanova (1976). See text for explanation.

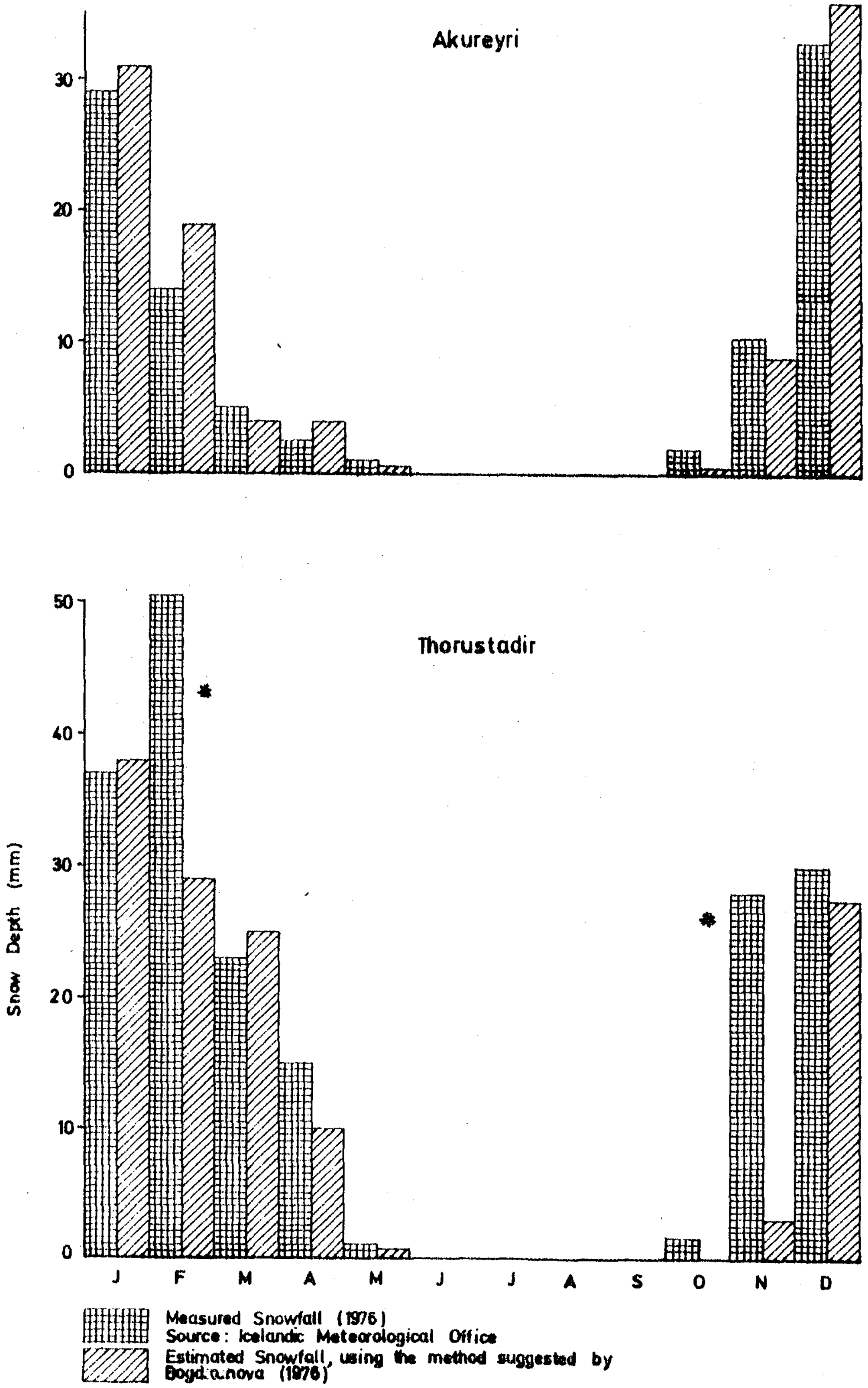
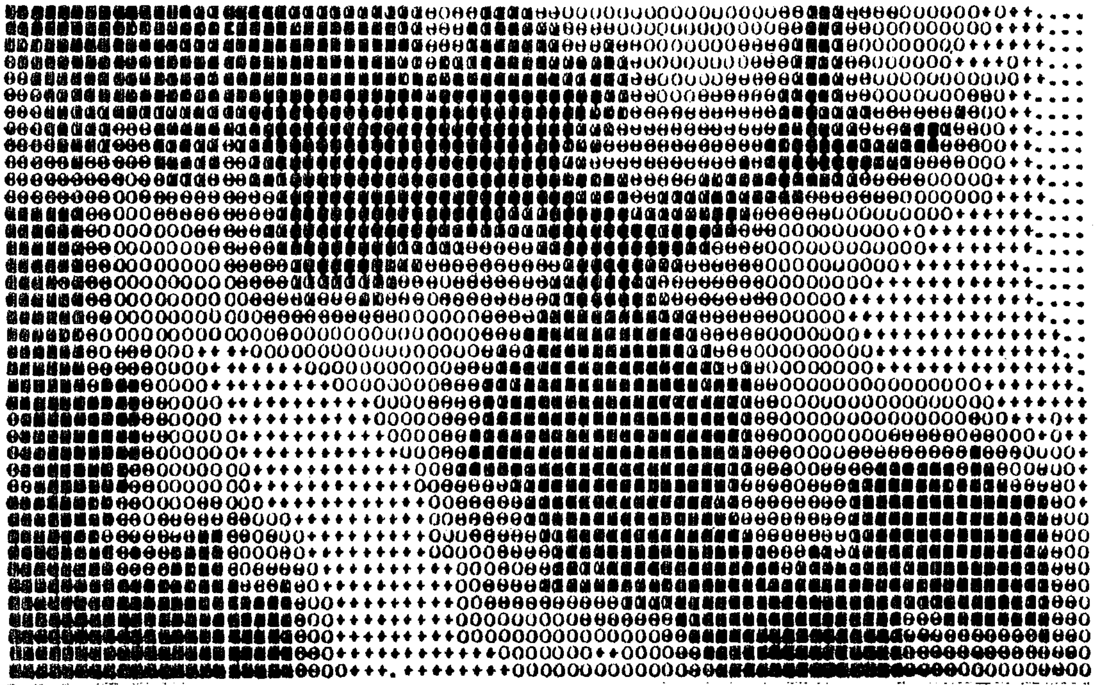


FIGURE 6.11
Comparison of Measured and Estimated (from total precipitation) Snow Depths at Akureyri and Thorustadir (1976).
* Note: the difference appears to be due to an unusually large proportion (75%-85%) of measured precipitation being classified as 'sleet'.

VALUES BETWEEN 4.000 6.000 8.000 10.000 12.000 14.000

NUPUR
JAN SNOW (CM) - BLOCK 1



NUPUR
OCT SNOW (CM) - BLOCK 1

FIGURE 6.12
January and October Snowfall Over the Central Nupur Matrix as Determined by the Subroutine MSNOW.

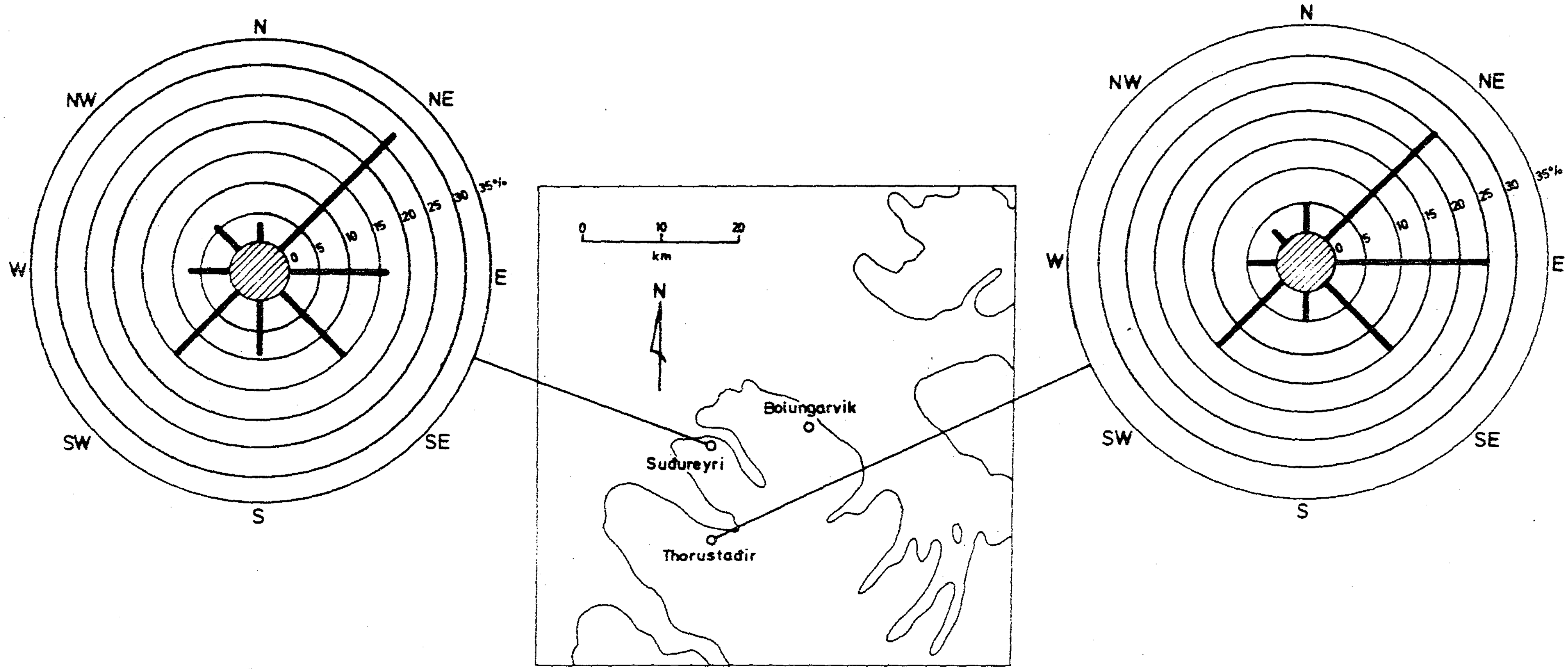
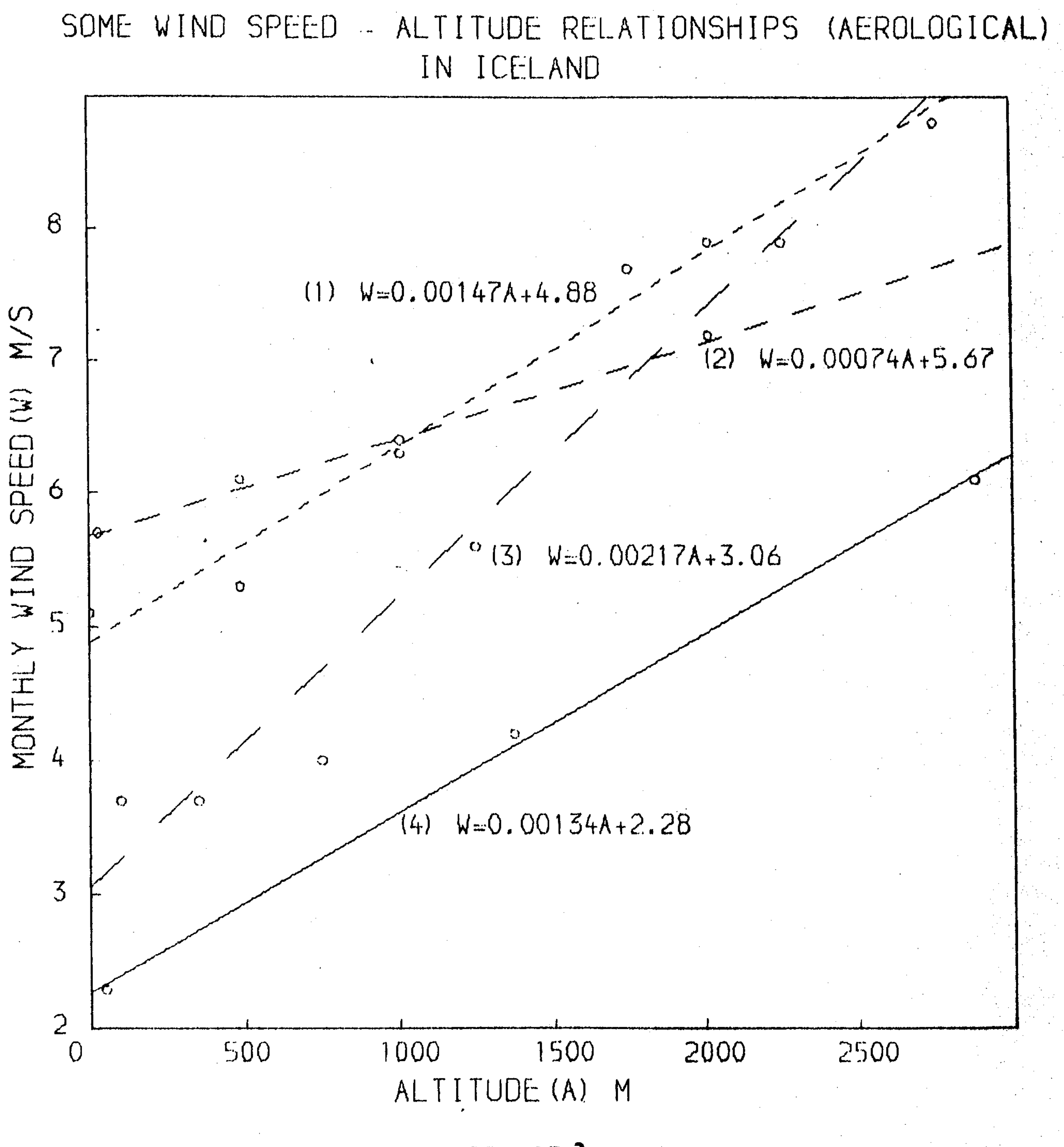


FIGURE 6.13
Frequency of Winter (Nov-Apr) Wind Direction (1931-60) in N.W. Vestfirðir.
Source: Veðrattan: Icelandic Meteorological Office.

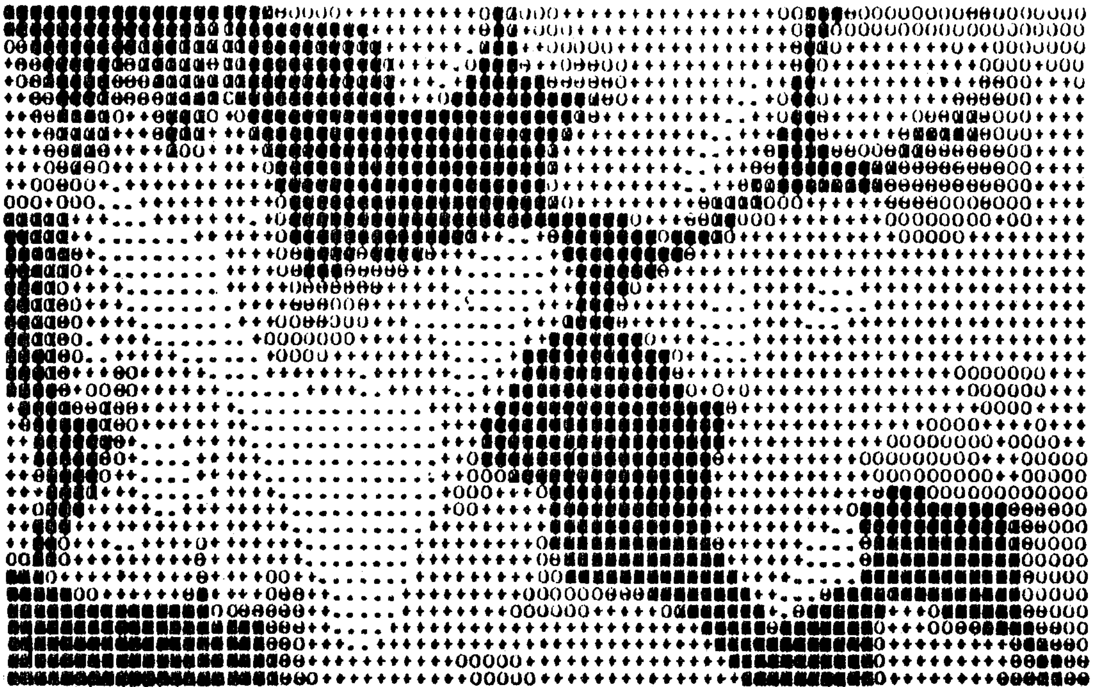


(1) ADALVIK (JUNE-JULY, 1926-27) } U.S. Hydrographic Office (1943)
(2) REYKJAVIK (MAR-AUG, 1921-33) }
(3) AKUREYRI (MAR-AUG, 1909-27) = Georgi and Kuhlbrodt - Raehder (1932)
(4) KEFLAVIK (MAR-AUG, 1965-74) - World Meteorological Organisation and USNOAA (1965-1974)

FIGURE 6.14
Aerological Wind Speed - Altitude Relationships in Iceland

VALUES BETWEEN 0.500 1.500 2.500 3.000 3.500 4.000

NUPUR
JAN WIND (M/S) - BLOCK 1



NUPUR
JULY WIND (M/S) - BLOCK 1

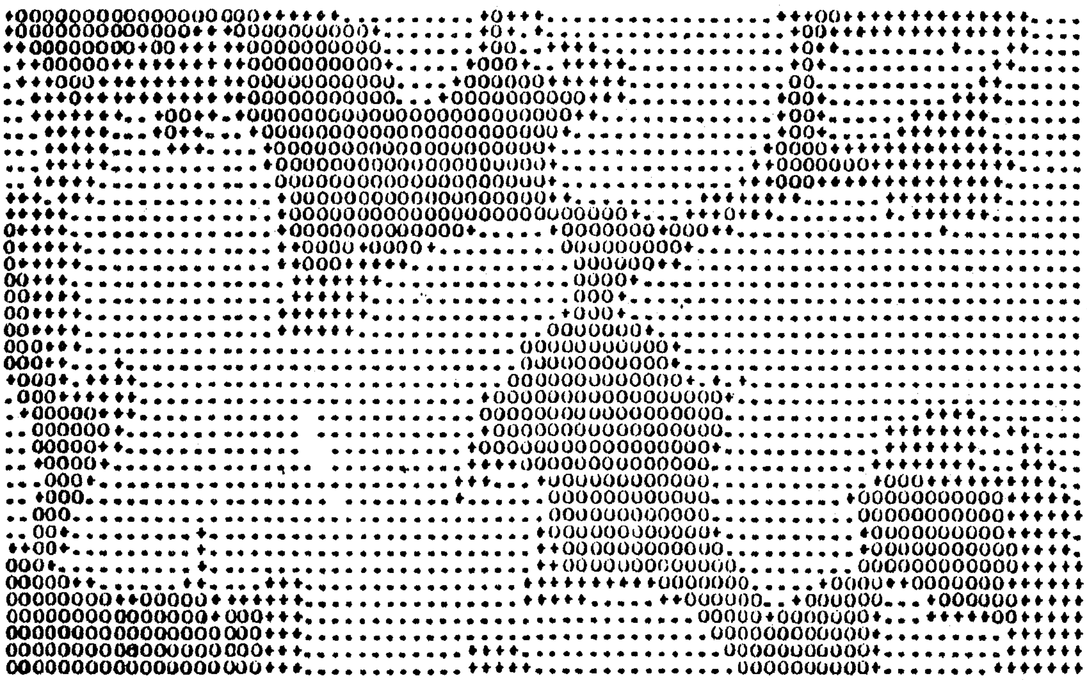


FIGURE 6.15
January and July Wind Speeds Over the Central Nupur Matrix as Determined by the Subroutine MWIND.

CHAPTER VII

SNOW ACCUMULATION AND DENSIFICATION

Following the specification of topographic and climatic inputs, the mass balance processes need to be modelled. As noted in sec. 2.3.3, in order to develop a model of general applicability and to obtain a clearer understanding of the factors controlling glacier mass balance, individual consideration of the accumulation and ablation processes is necessary.

The spatial variation in snow and ice accumulation due to such processes as snow drifting, avalanching and changes in densification of the snow material are dealt with in this chapter. Those processes resulting in the ablation of snow and ice are considered in the succeeding chapter.

7.1 Characteristics of Snow Accumulation

7.1.1 Previous Studies

A characteristic of snow accumulation in mountainous areas is its extremely local distribution. Substantial differences in snow-cover over short distances arise as a result of complex interactions between local winds and topography resulting in the redistribution of snowfall. A recognition of the highly variable nature of snow accumulation has led Meiman (1968, p.35) to comment that snow accumulation "is probably the weakest link in our knowledge of the snow hydrologic cycle".

The complexity of snow transport and deposition processes prevents a complete physical understanding of snow cover variations. Out of necessity, therefore, researchers have resorted to the use of empirical relations (established through standard regression techniques) to define the spatial distribution of snow accumulation. These relationships make use of the association between snowcover and various physiographic and meteorological factors, including altitude, gradient, aspect, vegetative cover, snowfall amount, air temperature, wind speed and direction.

The influence of these factors on snow accumulation variability at the meso- or local-scale has been examined by a number of researchers. Meiman (1968) provides a comprehensive review of the North American literature concerned with the role of altitude, slope aspect, and forest vegetation on snow accumulation. Among the studies emphasising the influence of terrain characteristics, he quotes the work of Mixsell et al. (1951); Grant and Schleusener (1961); Leaf (1962); Court (1963); Anderson and West (1965); and Gary and Coltharp (1967). In this respect the work of Salamin (1961), Anderson H.W. (1968), Golding (1970), and Super (1971) is also relevant.

The important effects of altitude and aspect for snow accumulation have also been examined by the snow investigations from the U.S. Army Corps of Engineers (1956) while Kuzmin (1963) has emphasized the interactions of terrain and wind in establishing snow accumulation patterns. A recent review of the factors affecting snow accumulation and distribution has been undertaken by Gray et al. (1979).

Many of these studies have observed linear associations between snow accumulation and altitude. However, large variations are encountered in the rates of snow accumulation increase, due to variations between major physiographic areas and also spatial and temporal variations within a given area. Some researchers have attempted to improve the correlations by including other terrain features. Young (1974 b, 1976), for example, has suggested the use of a regression equation based upon altitude, gradient, and local relief to estimate snow accumulation at each point of a grid covering a given area, with sampling points chosen to cover the ranges of these variables as well as possible (sec. 3.1).

These studies of snow accumulation related to physiographic and other environmental factors provide little in the way of physical understanding of the spatial variations in snow accumulation. For the development of a model of general validity, therefore, it is important to consider the main processes responsible for snow cover variations (sec. 3.4).

7.1.2 Accumulation Processes

Glacier accumulation in mountainous areas can be regarded as the result of the following processes:

Accumulation = snowfall ± wind-drifted snow
± snow avalanching from adjacent areas
∓ condensation of atmospheric moisture onto the glacier surface (rime; hoar-frost)
+ formation of superimposed ice (freezing of rain and meltwater).

With the limitation of this study to a consideration of 'temperate' glaciers at pressure melting point throughout (sec. 1.4), the role of rime and superimposed ice is not considered to be important for glacier mass balance. This simplification is justified to some extent by field studies of temperate glaciers which reveal the contribution from these two processes to be negligible (e.g. Meier and Tangborn, 1965).

Glacier accumulation on local glaciers in temperate mountainous regions is, thus, considered to be a function of the variations in solid precipitation, snow drifting and avalanching. It was noted earlier (sec. 3.1) that the redistribution of fallen snow by wind drifting exerts an important control on the asymmetry of local glacier distribution. Numerous studies of mountain glaciers, with reference to their local climatic and geomorphic environments, further reveal the great extent and importance of wind-blown and avalanche-transported snow for glacier-feeding.

Dolgushin (1961a), for example, noted that many cirque glaciers in the Urals owe their existence completely to wind drifted and avalanced snow. As a result these 'Ural-type' glaciers are situated well below the theoretical snowline. Such glaciers have also been identified in other parts of the U.S.S.R., viz. Altai, Eastern Siberia and Caucasus. On the glaciers in Zailiysky Alatau it is estimated that 15 to 20% of the accumulation is brought in from the surrounding slopes (according to Soviet researchers quoted by Hoinkes, 1964).

Kotlyakov (1973) reported the results of snow accumulation measurements on mountain glaciers in the Polar Urals, Caucasus and Pamir-Alai. He noted that snow drifting increases accumulation by about 15% on valley glaciers and by more than 50% on cirque and slope glaciers. While it is difficult to distinguish between the relative contributions of snow drifting and avalanching, (especially since snow drifting is a most important factor for avalanche formation), Kotlyakov suggested that the share of glacier feeding by avalanching is about one-third of that by snow drifts. The important role of avalanching for mountain glacier nourishment in the Soviet Union has also been discussed by Iveronova (1966), Lossev (1967), Tushinsky (1975) and Zalikhanov (1975).

Studies of local glaciers in the U.S.A. also emphasise the importance of snow drifting and avalanching for glacier mass balance. For example, mass balance studies on glaciers in the North Cascade Range (Washington), revealed that between 15-35% of the total mass influx is due to snow drifting and avalanching (Tangborn et al., 1977; Dethier and Frederick, 1981). For glaciers in the Colorado Front Range, Outcalt and MacPhail (1965) estimated that the contribution of drifted snow to be between four and eight times the regional snow precipitation.

Clearly, then, a consideration of snow drifting and avalanching is important for the modelling of glacier mass balance in mountainous areas. The complex nature of snow transport and deposition processes and the large number of factors affecting these processes, however, defy the development of a generalised, physical-based mathematical model of snow cover accumulation. In view of this, the modelling of the snow accumulation processes in this study is based on rather generalised and deterministic relations (sec. 4.3.3). In particular, the close control of the accumulation processes by the input topographic and climatic variables is emphasised in the modelling approach (sec. 3.4).

The manner in which the initial snowfall distribution was determined has been discussed in sec. 6.4. The modelling of snow drifting and avalanching, with respect to the Nupur area, is discussed in the two succeeding sections.

7.2 Snow Drifting

Snow drifting takes place in its most complicated form in mountainous areas, where wind speed and direction, local topography, and the properties of snow cover are most variable. The mechanics of snow drifting have been reviewed by Mellor (1965), Ōura et al. (1967), and Radok (1977); while Kuzmin (1963) has examined the interactions between local topography and wind speed in controlling the transport and redistribution of snow.

7.2.1 Identification of Snow Drift - Erosion and Deposition Areas

Snow drifting may result in a decrease or increase of snow cover depending on the local wind speed which, in turn, is determined by local obstructions to air flow (sec. 6.5). In general, wherever surface winds increase, as on the crest of a ridge or plateau, erosion occurs. Conversely, wherever winds decrease, as on the lee side of ridges, along plateau margins, or in suitable depressions, snow is deposited by the air stream.

A first consideration in the redistribution of snow by wind drifting is the identification of potential snow-drift source areas. An examination of the Nupur area (Fig. 4.8) reveals the presence of extensive plateau remnants which can be expected to act as snow-drift catchment areas. The occurrence of high wind speeds on these exposed summit areas would lead to the drifting of snow in the direction of the prevailing winds and its deposition on the sheltered leeward slopes along the plateau margins. The latter would, thus, become the main areas for the collection of snow-drift and the formation of avalanches (sec. 7.3).

The distribution of the potential drift-source areas, in the form of summit plateau remnants, in the Nupur area can be determined by considering the degree of exposure present at every grid point (e.g. by using the subroutine TPGW) or, more simply, by using a combination of altitude and slope gradient threshold values, as has been done in Fig. 5.28 (sec. 5.6).

Apart from favourable topographic location, a number of other conditions need to be fulfilled before snow drift can occur, as has been pointed out by Pugh and Price (1954). Firstly a sufficient thickness of snow must be present to even out the surface irregularities on the plateau. Only after the plateau surface becomes aerodynamically smooth, can it be expected to act as a supply area for drifting snow. From field observations of the plateau summit in the Nupur area (Fig. 3.4) it was estimated that an average snow thickness of about 0.5 m. is necessary for the minor roughness to be filled in. However, following validation procedures (chap. IX), this critical snow thickness value was set at 0.3 m. Thus, in the mass balance model, the snow cover due to previous accumulation (given by matrix TDEPTH(I, J)) and the current month's snowfall (in matrix SNOW (I, J)) must exceed 30.0cm. before any snow drift can occur.

A further condition for appreciable drifting is that the snow must be dry. Thus, snow drifting is only considered to be important at grid points with mean monthly temperatures of less than or equal to $0^{o}C$. This is because at below freezing temperatures, the inter-particle cohesive forces in newly fallen snow can be considered negligible, resulting in snow transport with low wind speeds. Related to this factor is the condition that before any movement of snow can occur, the wind must attain a suitable threshold shear velocity to overcome the snow particle weight and the inter-particle cohesive forces (Schmidt, 1980). This critical velocity is highly variable depending on the size, shape and weight of the snow crystals and the cohesive forces, the latter being dependent on the wetness of the snow. This aspect of the critical wind velocity needed to initiate snow transport is discussed further in sec. 7.2.2.

Any grid point which satisfies the above condition in the altitude matrix can be identified as a drift-erosion point. The next step is to determine the location of the grid points receiving snow-drift. In the first instance the snow from the drift-erosion points is considered to be transported to the slopes at the plateau margins, leeward to the actual distribution of wind directions. In theory, then, (given the variable direction of winds at the plateau) all the points lying adjacent to the plateau edge can be considered as potential drift-deposition points. These points are identified by means of the subroutine DEP which considers each point in the altitude matrix (with the exception of those forming the plateau remnants) to determine if any of its eight neighboring points lie on the plateau.

Following the determination of the drift-erosion and potential drift-deposition points, a simple line-printer map (optional) can be constructed to depict the drift-situation at each point of the altitude matrix (Fig. 7.1).

7.2.2 Determination of the Amount and Direction of Snow Drift

The next stage in the determination of the snow drift pattern is concerned with estimating the amount of snow drift received by the potential drift-deposition points in any given month. In view of the lack of data on local wind conditions (sec. 6.5), a number of simplifying assumptions are necessary for computing the amounts of drifted snow.

(i) It is assumed that the winds on the plateau, during the winter months, blow long enough and with sufficient intensity to remove all the fallen snow above a thickness of 0.3 m. This assumption is justified for the Nupur area in the following manner:

Snow drift measurements reveal that an initial wind velocity of around 3 to 5 $m.s^{-1}$ (at the 1 m. level) is necessary to initiate snow transport (Kuzmin, 1963; Mellor, 1965; McKay, 1968). Furthermore, a number of empirical formulae relating snow drift transport to wind speeds have been published. For example Kuzmin (1963 p.30) lists the relationships derived by Mel'nik, Dyunin and Komarov. The results obtained from these relationships compare closely with total drift transport calculated from theoretical considerations (e.g. using the theoretical model of Dyunin, quoted in Kuzmin, 1963 p.33) and with other empirical measurements (e.g. Imai, 1969). However differences in snow drift transport estimated from published empirical formulae can arise because of the different snow conditions and stages of development at the various localities that snow drift measurements were made (Takeuchi, 1980).

In the following discussion reference is made to the widely used relationship developed by Komarov, which expresses the total drift intensity of a snow wind current, in an infinite vertical layer, Q in g. $cm.^{-1}$ $min.^{-1}$ as a function of the wind velocity, at the one metre level, U in m. s^{-1}

$$Q = 0{\cdot}0065\ U^{3{\cdot}5} - 0{\cdot}4$$

Using this relationship to express the total drift transport intensity of blowing snow from the plateau areas, the approximate time for which winds of a certain force need to blow to remove all the monthly snowfall accumulated in any given month, can be estimated.

An examination of the drift source plateau areas in Nupur (Fig. 4.8) reveals that the maximum distance the snow has to be drifted in any given direction to reach the plateau margins is about 2·0 km. From sec. 6.4, it can be estimated that the maximum monthly snowfall in any winter at the level of the plateau is about 22·0 cm. Furthermore, although measurements of wind speeds at the plateau are lacking, records of wind speeds at the base meteorological station indicate that extremely high wind speeds are common in N.W. Iceland. For example, the average number of days (mean for 1931-1960), that wind speeds over or equal to 9 Beaufort (i.e. 22·6 m. s^{-1}) were recorded at Suđureyri (Fig. 6.13) is 15·6 per year, 12·1 of which occurred during the winter months (Eythōrsson and Sigtryggsson, 1971 p.44). The frequency for days with high wind speeds at the level of the plateau summits can clearly be expected to be much greater than that recorded at this low-lying base station.

Now the rate at which the snow surface (of density ρ g.cm^{-3}) over a plateau of length (L cm) is lowered by snow drifting with an intensity of Q g. cm.$^{-1}$ min^{-1} can be expressed as:

$$\frac{dh}{dt} = \frac{Q}{L.\rho} \text{ cm.min}^{-1}.$$

The time (t) taken to remove a snow cover thickness of h cm is given by:

$$t = \frac{L.\rho.h.}{Q} \text{ min.}$$

Taking the density of settled snow as 0·25 g.cm^{-3}, the maximum plateau length as 2·0 km, and the snow depth as 22·0 cm, the approximate time taken to remove all the accumulated snow on the plateau remnants with wind speeds of 22·6 m s^{-1} can be estimated as:

$$t = \frac{2{\cdot}0.10^{5}.0{\cdot}25.22{\cdot}0}{(0{\cdot}0065.22{\cdot}6^{3\cdot5}) - 0{\cdot}4}$$

$\simeq$ 51 hours (i.e. for approximately 7·1% of the month)

In view of the above comments, the plateau summits in Nupur are likely to experience high wind speeds for periods of longer duration than necessary for the complete removal of the monthly snowfall. Furthermore, wind speeds of lower than 22·6 m. s^{-1} but of sufficient intensity to enable snow drifting (i.e. > 3 m. s^{-1}) can be expected to occur for longer durations.

For example, free air observations during the winter months (1921–1933) at a level of about 1000 m. over Reykjavik reveal that wind speeds of around 4·7 ms^{-1} were observed for 45·8% of the time, 11·1 ms^{-1} for 40·0%, and 18·0 ms^{-1} for 11·3% (U.S. Hydrographic Office, 1943) There may, however, be systematic differences between ground conditions and free air observations. If these values are used to calculate the potential snow drift from the Nupur plateau summits, it is estimated from the above relationships that about 26·2 cm. of snowfall can be removed each month. This is greater than the monthly snow accumulation of 22·0 cm expected on the plateau areas of Nupur. It can, thus, be safely assumed, from the above considerations, that the monthly snowfall from the plateau remnants is all removed by snow drifting.

(ii) The second assumption refers to the amount of snow sublimated during snow drifting. Some studies on snow transport have indicated that, under certain conditions, a large proportion of the snow may sublimate when it is transported by wind (e.g. Santeford, 1972; Tabler and Schmidt, 1973; Tabler, 1973; Meiman and Grant, 1974).

The average distance a snow particle travels before it completely sublimates is referred to as the "transport distance", Rm (Tabler, 1973, p.75). Schmidt (1972) has developed a sublimation model of wind - transported snow which examines all the factors contributing to the sublimation of a single blowing particle. Based on this model Tabler and Schmidt (1973) presented a method of calculating the transport distance (Rm in metres) using the following relation:

$$R_m = \frac{0{\cdot}715\ (\bar{u}_2 - 5{\cdot}87)\ \bar{x}^{-3}}{(d_m/d_t)_{\bar{x}}}$$

where $(d_m/d_t)_{\bar{x}}$ is the average sublimation rate of a snow particle of average diameter, $\bar{x}$ (in cm).

$\bar{u}_2$ is the average wind speed at the 2 m level (in m. s^{-1}).

$(d_m/d_t)_{\bar{x}}$ is calculated from the following equation:

$$(d_m/d_t)_{\bar{x}} = \frac{8{\cdot}95.10^{-4}\ \bar{x}(1+37\bar{x})(1-RH/100) + 5{\cdot}22.10^{-7}(85{\cdot}6-T)\ \bar{x}^{-2}.S}{T^2 - 8{\cdot}2T + 254}$$

where RH is the relative humidity (%)

T is the air temperature ($^{\circ}$C)

S is the solar radiation (cal. cm.$^{-2}$hr.$^{-1}$)

$\bar{x}$ is the average sublimation particle diameter (cm).

The above relations were used to estimate R_m for drifting snow, on the plateau remnants of Nupur, during the winter months(November to April). The mean monthly winter conditions,necessary for calculating R_m, were specified by data for the following: RH (provided by the Iceland Meteorological Office), T (estimated from the method described in sec. 6.3) and S (from Einarsson, 1969). $\bar{x}$ was taken as 0·05 cm (being the average size of snow particle drifted according to Kuzmin, 1963; Mellor, 1965). Using the appropriate values for each winter month the average sublimation rate for drifting snow in the Nupur area was determined to vary from $9{\cdot}736.10^{-8}$ (for November with a mean temperature of $-2{\cdot}1^{\circ}$C) to $8{\cdot}723.10^{-8}$ (for February with a mean temperature of $-5{\cdot}2^{\circ}$C).

By substituting these sublimation rates into the equation for the determination of Rm, the variation of the average transport distance of drifting snow particles (R_m in metres) in the Nupur area can be expressed as a function of the mean wind speed ($\bar{u}_2$ in m.s.$^{-1}$). This relation is depicted for each winter month in Fig. 7.2. It can be seen from this figure that, with wind speeds of 8·0 m.s.$^{-1}$ and above, the sublimation of snow particles being drifted to the plateau margins (max. distance of transport being about 2·0 km) is likely to be low.

The consideration of sublimation of snow particles during drift-transport is, thus, left out of the mass balance model. However, sublimation losses could still occur, during periods of high intensity winds, by some snow being blown beyond the areas immediately leeward of the plateau margins into lower levels with higher temperatures. Studies of wind blown snow in Central Colorado (Santeford, 1972; Meiman and Grant, 1974) indicate that considerable amounts of snow may be lost by being transported over the ridgeline. The proportion of drifted snow sublimated in this manner is, however, difficult to estimate without continuous measurements of wind speeds and snow drift intensity at the plateau edge. Following model validation procedures (chap. IX) such losses of snow are set at about 20% of the total snow drifted from the plateau drift-source areas.

(iii) The third assumption is concerned with calculating the amount of drift received at the potential drift-deposition points. Now the direction and amount of snow drifted from each of the drift-erosion points will be controlled by the local wind speed and direction during the period of snow drifting. Lack of any short-term data on wind conditions and snow transport, however, makes it impossible to develop any specific relations between snow drift intensity and wind speed and direction.

For the computation of monthly totals of snow drift, it is assumed that the amount of snow drifted in any particular direction is directly proportional to the frequency with which winds are estimated to blow in that direction during the winter months (Fig. 6.13). It is conceivable, however, that the snow drift pattern in a particular month could be determined by the occurrence of high-intensity winds from one dominant direction leading to the drifting of all accumulated snowfall. Consequently for the remainder of the month, little or no snow may be available for drifting with winds from other directions. Clearly, this problem cannot be resolved without information on the relative timing of snowfall and the occurrence of high intensity winds within a given month. It seems reasonable to assume, however, that the snow redistribution pattern at the end of the accumulation season should bear a close resemblance to the relative frequency of winds from different directions; if not in a single season, at least as a long term average.

Having considered the basic assumptions in the calculation of snow drift, the amount of snow received at each drift-deposition point is determined by the subroutines MDRIFT and DRWIND in the following manner. Taking each drift-deposition point in turn, the subroutine MDRIFT considers snow drifting from grid points on the plateau remnants in all possible directions (maximum of eight for a drift-deposition point completely surrounded by drift-erosion points). For each drift-direction the subroutine DRWIND calculates the amount of snow drift received from each successive drift-erosion point on the plateau, in accordance with the frequency of winds from that direction, until there are no further grid points contributing snow drift or the limit of the study area is reached (Fig. 7.3). The total snow-drift received at each drift-deposition point, from all possible directions, is shown by the matrix DRIFT (I, J).

A check on the calculation of monthly snow drift is provided in the subroutine MDRIFT, by computing the total amount of snow being drifted off the plateau remnants (TPL), excluding the portion lost through sublimation,the total amount of snow received at the plateau margins by the drift-deposition points (TDR), and the amount of snow drifted outside the study area limits (OUTDR).

It was attempted, as far as possible, during the construction of the altitude matrix to reduce boundary effects by considering whole mountain ranges. In the Nupur area, however, the extension of the plateau remnants beyond the southern and western margins (Fig. 5.28) would lead to some inaccuracy in the calculation of snow drifting. An estimation of the boundary effects on snow-drift computations was made for the 'test' matrix by calculating the difference between the total snow drifted from the drift-erosion points (using a value of 30.0 cm for TDEPTH) and that received by the drift-deposition points within the study area (Table 7.1). It can be seen that, for each of the drifting months, the portion of snow drifted to points outside the study area is less than 10%. The average is 9.0, and thus the effect of the study area boundary on the determination of glacier mass balance is considered small.

The monthly snow cover at all grid interactions of the altitude matrix, following redistribution of initial snowfall by snow-drifting, is recorded in the matrix SNOW1 (I,J). Fig. 7.4b illustrates the effect of this redistribution in the 'test' matrix for the drifting of initial snowfall during February (Fig. 7.4a) to the leeward slopes at the plateau margins. The computational procedures for the determination of the monthly show drift pattern are summarised in Fig. 7.5.

7.3 Snow Avalanching

7.3.1 Characteristics of Snow Avalanching

Following the redistribution of initial snowfall by wind drifting, consideration is given to the effects of snow avalanching.

Snow avalanches have been studied from many aspects and basic information on their characteristics, classification, mechanics, control, forecasting and protection can be found in the monographs by Mellor, 1968; Castelberg et al. (Eds.), 1975; and Perla and Martinelli (1976). These aspects have also been discussed at a number of international symposiums held at Davos, Switzerland (I.A.S.H., 1966); Cambridge, England (International Glaciological Society, 1977); Banff, Alberta (Perla, ed. 1978); and Fort Collins, Colorado (International Glaciological Society, 1980a). A useful bibliography of recent non-Russian avalanche literature (1950-1977) has been compiled by the World Data Center A for Glaciology (Shartran (Ed.), 1977).

In order to permit some degree of avalanche prediction and control, a number of investigators have attempted to analyse the quantitative relationships existing between terrain, climate, snow cover properties, and avalanche formation (e.g. Judson and Erickson, 1973; Armstrong and Ives, 1976). However, such attempts are complicated by the great number of factors controlling avalanching occurrence.

The magnitude, frequency and timing of avalanches, at a given site, is largely a function of the meteorological variables (e.g. snowfall rate and amount, wind speed and direction, air temperature, insolation etc.) which interact with the local topography to determine the character, thickness and state (e.g. density, free-water content, porosity and friction) of the snowpack and therefore its strength and stability. In spite of the great deal of attention devoted to avalanche studies, the variable and complicated nature of the relationships existing between these controlling factors has meant that "even the most fundamental measures of a snow avalanche remain uncertain" (Perla, 1980 p.457).

The determination of the location and timing of a particular avalanche event would necessitate the excavation of snow pits to examine the stability conditions of the snow-pack under specified topographic and meteorological conditions. However, such an analysis is not possible in the present study due to the generalised nature of the climatic input variables in the mass balance model. In any event, the main concern in this aspect of the model is to (i) identify the potential avalanching sites together with the areas in which the avalanched snow could be re-located, and (ii) estimate the amount of snow being avalanched from, or received by, any grid point in the altitude matrix within a particular month.

The controls of snow avalanching vary according to the type of avalanche being considered. While a variety of avalanche types have been distinguished according to several classification schemes (de Quervain et al., 1973), observations of snow avalanches suggest that direct-action avalanches (in the form of dry loose-snow and soft slab avalanches) are predominant in mountainous areas (Judson, 1967; Barbat, 1973; Abbi and Pareek, 1974).

An analysis of snow avalanches in Iceland by Jónsson and Rist (1971), Rist (1975), and Björnsson (1980) reveals that Vestfirðir and Tröllaskagi (the two regions in which the study areas of Nupur and Thvera are situated) are the main areas of avalanching in Iceland. The majority of avalanches in these regions consist of dry snow and occur during the winter months (Nov. - April). They are associated with the occurrence of high winds and moderate snowfall (10-25 $mm.d^{-1}$).

Direct-action avalanches usually occur during snowfall or within the following few days. Thus, with the determination of snow redistribution by avalanching over monthly periods in the mass balance model, the actual timing of avalanche events does not form an important consideration and the avalanche events in any particular month can be assumed to be related to the snowfall occurring within that month.

A further characteristic of direct-action avalanches is their low degree of internal cohesion and, as a first approximation, failure can be assumed to occur for many areas when the angle of repose is exceeded (Perla, 1980). This points to a close correlation between the occurrence of direct-action avalanches and local terrain features, especially the 'critical slope gradient for avalanching'.

Local terrain factors exert an important control in delimiting the spatial distribution of snow avalanches (Schaerer, 1972; Akifeva and Kravtsova, 1973; Martinelli, 1974; Luckman, 1977). Many researchers have examined the relation between slope gradient and avalanche formation. It has been established that the optimal gradient for the development of slab avalanches lies between 25 and 50°, particularly between 30 and 45° (Fraser, 1966; Judson, 1967; Mellor, 1968, Maksimov, 1969; Schaerer, 1972; Martinelli, 1974; de Quervain, 1975; Mears, 1976; Perla and Martinelli, 1976; Björnsson, 1980). On slopes steeper than about 45°, snow tends to sluff off soon after falling and the development of loose-snow avalanches is common. Thus we are not concerned with an upper gradient limit and in general avalanches initiated on starting zones steeper than about 30° will tend to run through over track gradients of about 15-25° and the snow will be deposited in runout zones with gradients of less than 15°.

Apart from suitable slope gradient, another important consideration for avalanche formation is a sufficiently thick snow cover. For a given site, the depth of snow accumulation has to cover the surface obstacles on the slope before avalanching can occur. On non-vegetated smooth slopes direct action avalanching is usually associated with a snow cover thickness of about 30 cm (Akkouratov, 1966). Observations by de Quervain (1975) suggest that a thickness of new snow of about 10-30 cm is sufficient for frequent loose-snow avalanching on slopes greater than 45°, 30-50 cm of snow for the development of local slab avalanches on slopes of greater than 35° and a snow thickness of greater than 50 cm for widespread slab avalanching on more gentle slopes of between 25-30°.

In general, then, a combination of steep slopes and adequate snowfall will result in avalanching. One topographic situation in which these conditions are commonly met is on leeward slopes adjacent to ridge or plateau margins, where large amounts of snow may be concentrated by wind drifting (as in the Nupur area). In such situations snow cover distribution by wind becomes the leading meteorological factor for the formation of avalanches (Akkouratov, 1966; Kotlyakov and Plam, 1966). In addition to snow drift accumulation, the lee slopes are often overhung with cornices which fall and provide effective triggers for avalanching on the lower slopes.

7.3.2 Determination of Snow Cover Distribution by Avalanching

From the above discussion (sec. 7.3.1) it can be seen that, as a first approximation, the redistribution of the snow cover by avalanching can be determined through the use of suitable threshold terrain and snow cover conditions.

Following a similar approach to that adopted in the calculation of snow drifting (sec. 7.2), the grid points satisfying the conditions for snow avalanching and snow deposition are identified in the matrix AVAL (I, J). In any particular month, the snow is considered to avalanche from those grid points which fulfill any of the following conditions: (i) gradient greater than 45° and snow depth greater than 10 cm; (ii) gradient between 35-45° and snow depth greater than 30 cm; (iii) gradient between 25-35° and snow depth greater than 50 cm. These threshold values can, to a certain extent, be verified by use of equations relating to the stability of snow deposits on slopes (e.g. Kuroda 1967). However the great variability found in the shear strength of snow under different conditions, together with the effects of snow metamorphism in changing the mechanical strength of the snow deposit, prevent a more general treatment of factors affecting snow stability and the occurrence of avalanching. As a first approximation, then, the above empirically-based threshold values of slope gradient and snow thickness are used in the mass balance model to locate the sites of snow avalanching. Avalanched snow is considered to be deposited at those grid points which have gradients of less than 15^{o}, or which form major concavities (sec. 5.6ii).

A simple line-printer map (optional) depicting the avalanche situation, in terms of the points from which the snow is being avalanched, potential sites for the deposition of avalanche snow, and points not relevant for the computation of avalanching, can be produced for the particular month under consideration (Fig. 7.6).

The redistribution of the snow cover by avalanching is then determined by the subroutines MAVAL and AVDIR in the following manner. Taking each avalanching point, in turn, the subroutine AVDIR is used to determine the amount and direction of avalanched snow. From the point of avalanching, the snow (that part above 10 cm depth; the remainder is left to cover minor irregularities of slope) is considered to be transported to an adjacent grid point in a direction given by the aspect of the maximum gradient at the avalanching point. If the point receiving the avalanched snow also satisfies the avalanching conditions, then the snow is further transported to next grid point in line with the aspect of maximum gradient at the second avalanching point. The snow will continue to be avalanched in this manner until (a) a grid point satisfying the conditions for snow deposition is reached or (b) the limit of the study area is reached. Some of the snow may be lost by sublimation during the transportation process, by being blown to lower levels. This loss was set at about 20% following validation procedures (chap. IX).

A check on the calculation of monthly snow avalanching is provided in the subroutine MAVAL by computing the total snow being avalanched (AMAV), the total amount of snow being received by the deposition points (AMDP), and the portion transported outside the study area (OUTSN).

As in the case of snow drifting, the boundary effects on mass balance computation are considered slight in Nupur since, for all avalanching months, the quantity of snow transported outside the study area was less than 10% of the toal snow avalanched (Table 7.2).

The snow cover at each grid point following redistribution by avalanching is recorded by the matrix SNOW2 (I, J). The sequence of snowfall→ drifting→avalanching is illustrated in Fig.7.4 for the 'test' matrix.

A particular avalanching situation deserves special consideration. The avalanching routine, described above, is concerned with the determination of snow cover distribution by avalanching following the consideration of snow drifting effects. In the Nupur area this does not create any problems since all potential avalanching points lie below the summit plateau remnants important for snow drifting. However, in other alpine areas, potential avalanching sites may also exist above the level of snow drift areas (e.g. in the case of mountain peaks). Snow avalanched from these sites may be initially deposited in areas acting as sources for snow drift and subsequently be drifted to the leeward slopes (when conditions suitable for snow drifting occur), where it may be involved in further avalanching. Snow avalanching from such grid points, lying above potential snow-drift areas, needs to be considered prior to the computation of snow drifting.

Separate avalanche routines (MDAVAL and PLAV) were thus developed to compute the amounts of avalanched snow being received by potential drifting points. These routines are placed before the drifting routine (MDRIFT) and the main avalanching routine (MAVAL) in the mass balance program.

7.4 Densification of Glacial Material

Following the determination of snow cover distribution by the snow accumulation processes, an important consideration in the mass balance model is the determination of the density and water equivalent of the glacial material at each grid-intersection of the altitude matrix.

Density is one of the key indicators of the physical state of the glacial material as it is related to grain size, thermal history and mechanical properties. It also serves as a useful indicator of the gradual development of glacial ice from newly fallen snow. The changes in the state of the glacial material (snow→firn→ice) deserve consideration in the mass balance model since they affect the computation of the energy balance (e.g. through different albedo values for snow and ice) and consequently the amount of glacier melting (chap. VIII).

In the mass balance program (Appendix A) the densification procedure is incorporated in the main routine following the determination of monthly accumulation at each grid point. It is considered in two parts: (a) estimation of the density and water equivalent of the 'current' month's snow cover (i.e. the particular month being simulated at any given time); and (b) monthly densification and changes in state of the existing glacial material at each grid intersection of the altitude matrix.

7.4.1 Density and Water Equivalent Variations in the Current Month's Snow Cover

The initial snow density of freshly fallen snow depends on the particular combination of air temperature, crystal form, wind speed and the rate of snow deposition. Several investigators have attempted to link variations in the density of new snow to meteorological (primarily temperature) and topography (primarily altitude) controls (e.g. Bossolasco, 1954; Wilson, 1955; LaChapelle, 1962 ; Martinec, 1966; Grant and Rhea, 1974). However comprehensive definition of density controls has yet to be clearly established.

The average density of newly fallen snow is usually taken as 0.1 $g.cm^{-3}$. Within a few weeks of deposition settled snow achieves an average density of about 0.25 $g.cm^{-3}$ (Seligman, 1936). This is confirmed by density - age relationships derived from long-term field data (Keeler, 1967; Sadvakasov, 1970). Subsequent modification depends on site, shelter and local weather conditions.

Alford and Keeler (1968) note that temperature increases tend to increase the instability of the initial crystal forms, resulting in the formation of a grain with few crystallographic faces and greatly reduced surface area per unit mass. The effect of wind is to increase densification by mechanically breaking up the original snow crystals into forms which are more susceptible to close packing.

The importance of wind packing has been stressed by Seligman (1936) who distinguishes between the formation of wind crusts (snow deposit packed by wind) and wind slabs (formed by the drifting and deposition of snow). The latter are usually involved in avalanching on leeward slopes and consist of compact drift with a density of around $0{\cdot}3\ g.cm^{-3}$.

Avalanched snow will generally have increased density, compared with that subject to wind drifting, depending on the type of avalanche. Observations by Schaerer (1967) of avalanched snow at selected sites in Glacier National Park in British Columbia, indicate that the density of snow subjected to dry-snow avalanching in winter varies from about 0·35 - 0·45.

In the mass balance model the mean density of snow accumulated in any particular month is considered to be a function of the extent to which it has been subject to the processes of wind drifting and avalanching. At any particular grid point the amount of snow left undisturbed is considered to have a density of $0{\cdot}25\ g.cm^{-3}$, that which has been brought in by wind drifting has a density of $0{\cdot}30\ g.cm^{-3}$ and any snow received through avalanching is given a density of $0{\cdot}40\ g.cm^{-3}$.

The amounts of snow (thickness in cm) received by initial snowfall (P), drifting (P1) and avalanching (P2) are determined by reference to the three matrices containing the values of snow cover distribution after initial snowfall - SNOW (I, J); after drifting - SNOW1(I,J); and after avalanching-SNOW2(I,J), as shown in Table 7.3.

The snow cover relations at each grid intersection of the altitude matrix must meet one of the eight conditions listed in Table 7.3. As an example of this procedure, the number of points representing each density category, during the course of mass balance simulation in the Central Nupur Matrix, for the month of January are listed in Table 7.3. A visual check can also be made by producing line-printer maps depicting which of these conditions are being fulfilled at any particular grid intersection (Fig.7.7).

Following the determination of the proportions of snow received by snowfall, drifting and avalanching, the mean water equivalent at each grid intersection is computed in the following manner:

$$\text{W.E. (I,J) in cm} = (0{\cdot}4.P_2 + 0{\cdot}3.P_1 + 0{\cdot}25.P)$$

The mean density is given by:

$$\text{DENSITY (I,J) in g.cm}^{-3} = \frac{\text{W.E.(I,J)}}{(P_2 + P_1 + P)}$$

The matrices SNOW1 (I,J) and SNOW (I,J) are converted to record the values for the water equivalent and density respectively. The variation of water equivalent over the altitude matrix in any given month can be depicted by means of line-printer maps (e.g. Figs. 9.1 to 9.13).

7.4.2 Monthly Densification and Changes in the State of Existing Glacial Material

Following the determination of the current month's density and water equivalent, consideration is given to the densification of the pre-existing glacial material at each grid point. The water equivalent of the current month's snow needs to be incorporated into any existing snow, while the monthly increase in densification and any changes in the state of the glacial material need to be determined.

Over time, deposited snow undergoes metamorphism which transforms the original new snow into granular firn (density range taken as 0·58–0·85) and eventually glacier ice (density 0·85–0·91). This transformation results from the interaction of several processes which include settling and compaction, sintering, sublimation, recrystallisation, movement along the internal glide planes of crystals, and refreezing of percolating meltwater. These processes have been classified under several types of metamorphic changes affecting the densification of snow and ice (e.g. destructive, constructive, and melt metamorphism) and have been discussed in detail by several researchers (e.g. de Quervain, 1963; Anderson and Benson, 1963; Shumskiy, 1964).

The net result of these metamorphic changes is a progressive increase in the density of the glacial material with time. The actual densification rate depends on the climate and type of metamorphism involved. However, several studies of density - age relations for snow and ice in temperate areas reveal some similarity in monthly densification rates.

Bilello (1967) in an examination of snow cover densification at various localities in North America found a mean monthly density increase of 0·01 $g.cm^{-3}$ during the winter months. Much higher densification rates are reported during the summer months due to fast settling rates associated with higher air temperatures and increased refreezing of meltwater percolating down from the snow surface. Schytt (1973), for example, noted a rapid density increase at the start of the ablation season on Storglaciärren (Kebnekaise, Sweden). His observations reveal a monthly density increase of 0·046 $g.cm^{-3}$ $month^{-1}$ during the summer months.

Measurements of density changes in the firn of the Great Aletsch Glacier (Switzerland) by Hughes and Seligman (1939) suggest a summer monthly densification rate of 0·04 $g.cm^{-3}$. Similar densification rates during the summer months are reported by Sharp (1951) for firn on the Upper Seward Glacier (St. Elias Mountains, Canada).

These measurements of densification rates roughly correspond with observations of the time period taken for the transformation of snow to ice in temperate glaciers. For example, observations on ice formation on the Central Tuyuksu Glacier (Zailiysky Alatau) by Makarewich and Tokmagambetov (1961) and on small glacierets in the Japanese mountains (Tsuchiya, 1974) reveal that the transformation from snow to glacial ice takes about 2-3 years. On Seward glacier, the process may take 3-5 years and is accomplished within a depth of about 13 m. (Sharp 1951).

Based on this literature survey, monthly densification rates of 0·01 $g.cm^{-3}$ (firn), 0·02 $g.cm^{-3}$ (snow) during the winter months (i.e. air temperature $\leq 0^{o}C$) and 0·04 $g.cm^{-3}$ (firn), 0·05 $g.cm^{-3}$ (snow) for the summer months (air temperature $> 0^{o}C$) are used in the mass balance model. The density of snow and firn at grid points where accumulation of glacial material is taking place is adjusted at monthly intervals using the appropriate densification rates. Changes in the state of glacial material (snow→firn→ice) are made when it obtains a critical density of 0·58 (for firn) and 0·85 (for ice).

The density and water equivalent of any snow, firn and ice present at each grid intersection are recorded in the matrices-DSNOW (I,J), WSNOW (I,J), DFIRN (I,J), WFIRN (I,J), DICE (I,J) and WICE (I,J). These aspects of monthly densification and changes in the state of glacial material are carried out in the main routine of the mass balance program. (Appendix A).

TABLE 7.1

Effect of Study Area Boundary on Snow Drift Computation (Central Nupur Matrix)

Month	TPL(cm)	TDR(cm)	OUTDR(cm)	% OUTDR
October	1576.7	1458.4	118.3	7.5
November	10101.8	9172.6	929.2	9.2
December	10657.6	9674.7	982.9	9.2
January	12136.9	11015.5	1111.4	9.2
February	14616.8	13265.6	1351.2	9.2
March	9415.5	8546.8	868.7	9.2
April	5049.1	4584.9	464.2	9.2
May	191.8	174.5	17.3	9.0

TPL = total snow drifted off the plateau areas.

TDR - total snow received by the drift-deposition points.

OUTDR - total snow transported out of the study area.

% OUTDR - OUTDR expressed as a percentage of TPL.

TABLE 7.2

Effect of Study Area Boundary on Snow Avalanche Computation (Nupur Matrix)

Month	AMAV (cm)	AMDP (cm)	OUTSN (cm)	% OUTSN
October	9590.0	8880.9	709.1	7.4
November	19590.7	18123.9	1466.8	7.5
December	21546.1	19894.2	1651.9	7.7
January	25031.2	23087.0	1944.2	7.8
February	30324.3	27960.5	2363.8	7.8
March	19415.1	17915.9	1499.2	7.7
April	10057.3	9299.8	757.5	7.5
May	2899.6	2684.9	214.7	7.4

AMAV = total snow avalanched

AMDP = total avalanched snow deposited within the study area

OUTSN = avalanched snow transported out of the study area

% OUTSN = OUTSN expressed as a percentage of AMAV

TABLE 7.3

Procedure for the Determination of Snow Amounts Received by Snowfall (P), Drifting (P1), and Avalanching (P2) at Each Grid Intersection

TDEPTH (I,J)

(snow cover due to previous accumulation)

SNOW (I,J)
(current month's snowfall)

SNOW1 (I,J)
(snow cover after drifting)

(SNOW1 = SNOW) (not affected)	(SNOW1 < SNOW) (SNOW1 = 30.0 - TDEPTH) (drift erosion)	(SNOW1 > SNOW) (drift deposition)

SNOW2 (I,J)
(snow cover after avalanching)

(SNOW2 = SNOW1) (not affected)	(SNOW2 < SNOW1) (SNOW2 = 10.0 - TDEPTH) (avalanche erosion)	(SNOW2 > SNOW1) (avalanche deposition)

Each grid point must meet one of the following conditions:

1. No snowfall and point not involved in drifting or avalanching.
   i.e. SNOW2 = SNOW1 = SNOW = 0.0
   therefore P = P1 = P2 = 0.0

2. Snowfall but point not involved in drifting or avalanching.
   i.e. SNOW2 = SNOW1 = SNOW > 0.0
   therefore P = SNOW2.

3. Snow drifted from point but not involved in avalanching.
   i.e. SNOW2 = SNOW1 = 30.0 - TDEPTH < SNOW
   therefore P = SNOW2.

4. Snow drift received but point not involved in avalanching.
   i.e. SNOW2 = SNOW 1 > SNOW
   therefore P = SNOW
   P1 = SNOW1 - SNOW

TABLE 7.3(contd.)

5. Point received both snow drift and avalanched snow.
   i.e. SNOW2 > SNOW1 > SNOW
   therefore P = SNOW
   P1 = SNOW1 - SNOW
   P2 = SNOW2 - SNOW1

6. Point received avalanched snow only.
   i.e. SNOW2 > SNOW1 = SNOW
   therefore P = SNOW
   P2 = SNOW2 - SNOW1

7. Point involved in avalanche erosion
   i.e. SNOW2 = 10.0 - TDEPTH < SNOW1
   therefore P = SNOW 2

8. Drift source point received avalanched snow which is subsequently drifted.
   i.e. SNOW2 > SNOW1 = 30.0 - TDEPTH < SNOW
   therefore P = SNOW 2

Number of points representing each category during simulation of glacier mass balance in the Central Nupur Matrix (40x80) for the month of January.

Category	1	2	3	4	5	6	7	8
Number of points	-	928	654	120	2	338	1158	-

DRIFT SITUATION MAP
MONTH=JANUARY

DRIFT EROSION POINT (*)
POTENTIAL DRIFT DEPOSITION POINT (O)
NOT INVOLVED IN DRIFT ROUTINE (.)

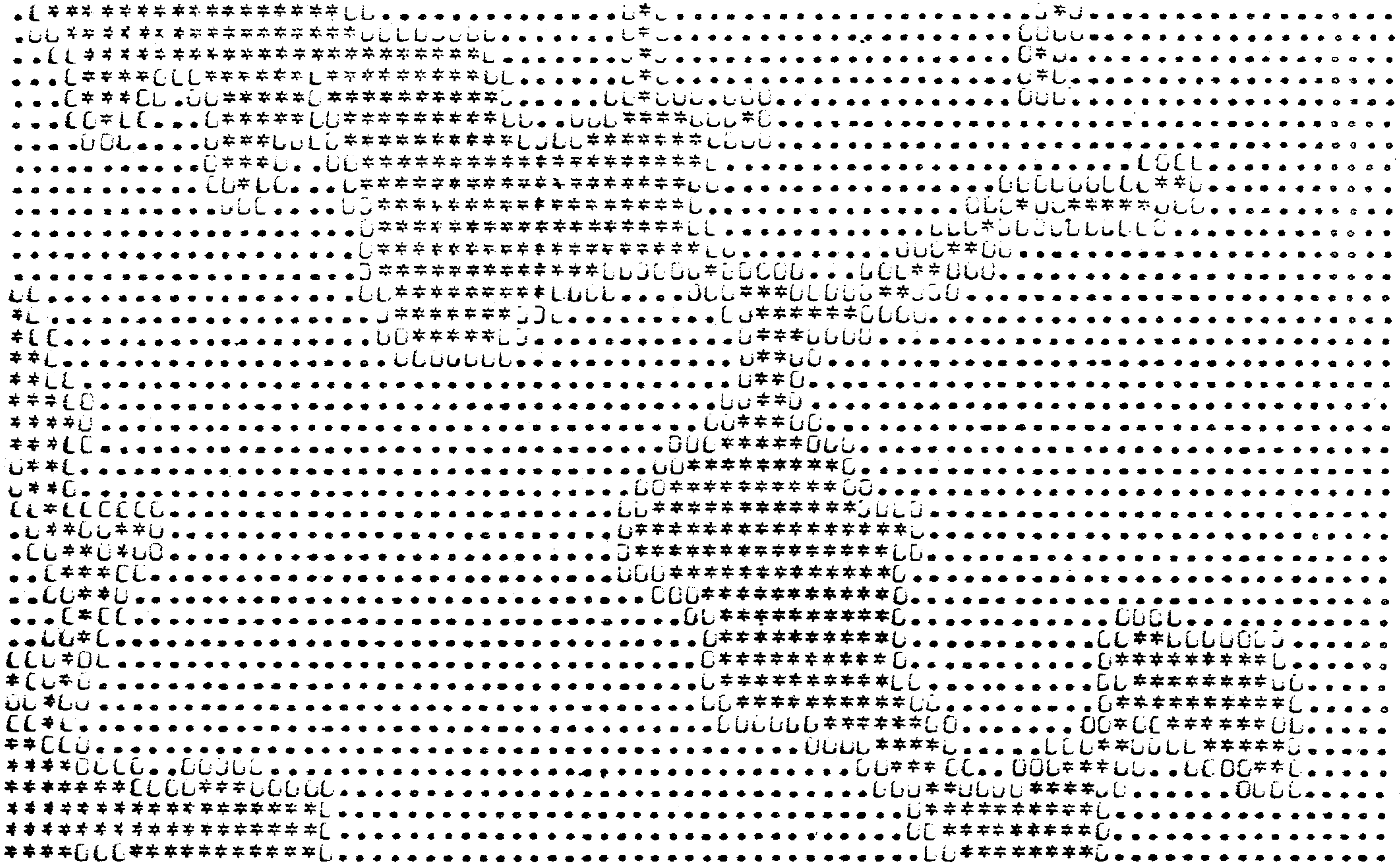


FIGURE 7.1
Snow Drift Situation Map Over the Central Nupur Matrix (for January)

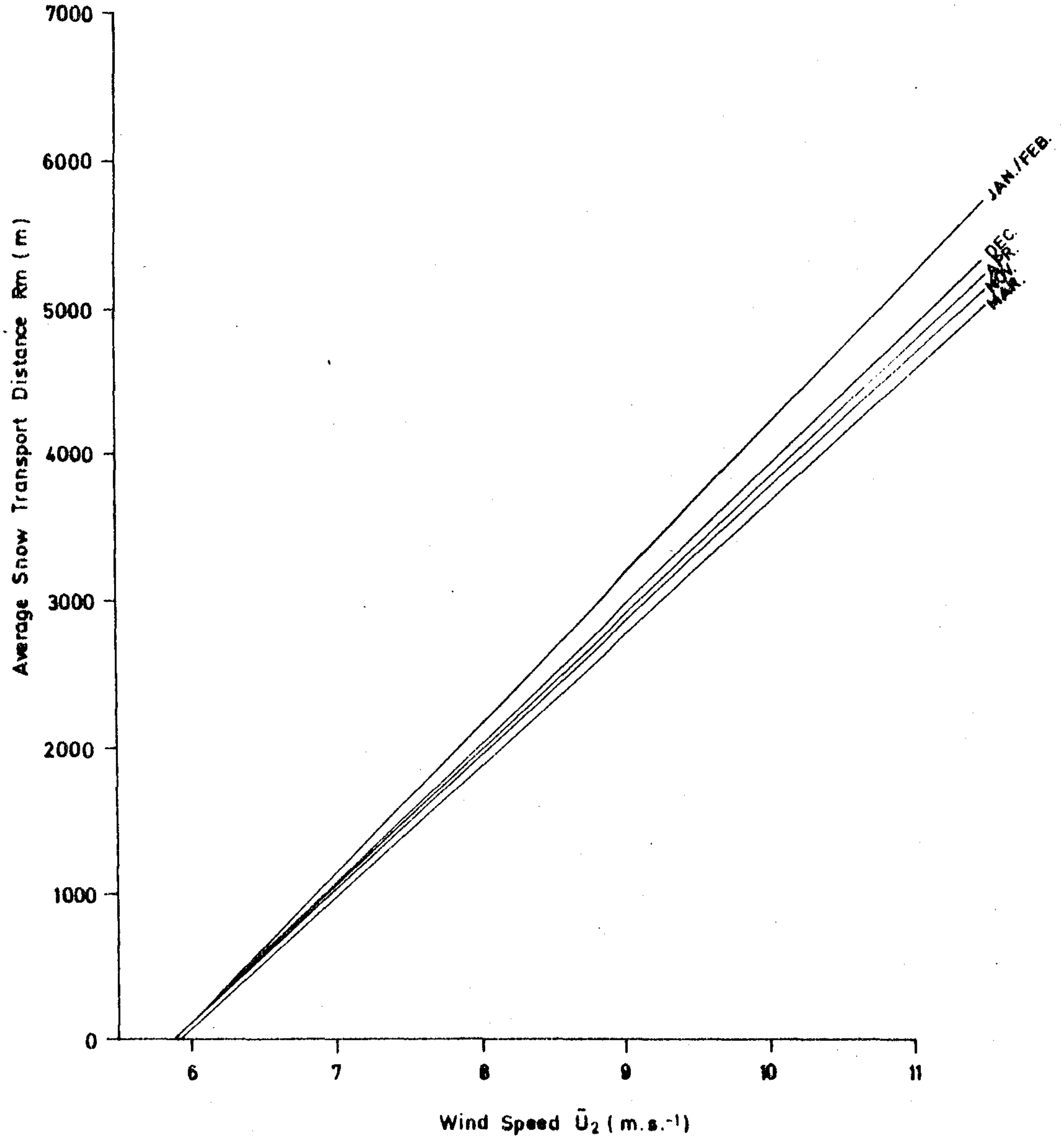


FIGURE 7.2

Snow Transport Distance as a Function of Wind Speed. Using monthly mean values for winter conditions during snow drifting from the Nupur plateau areas. (see text for explanation of computational procedure).

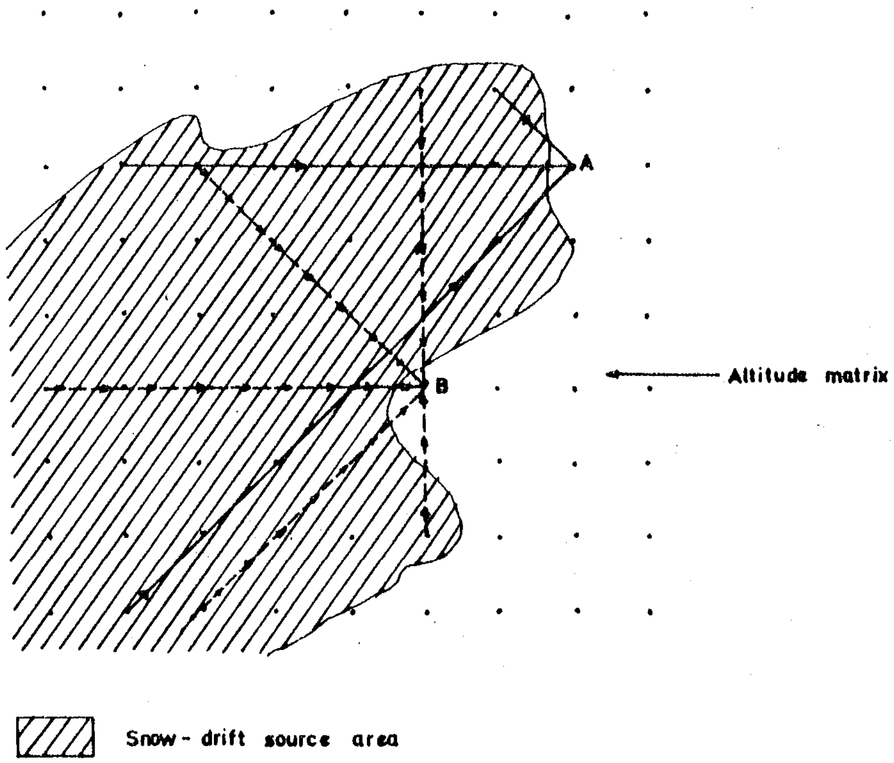


FIGURE 7.3

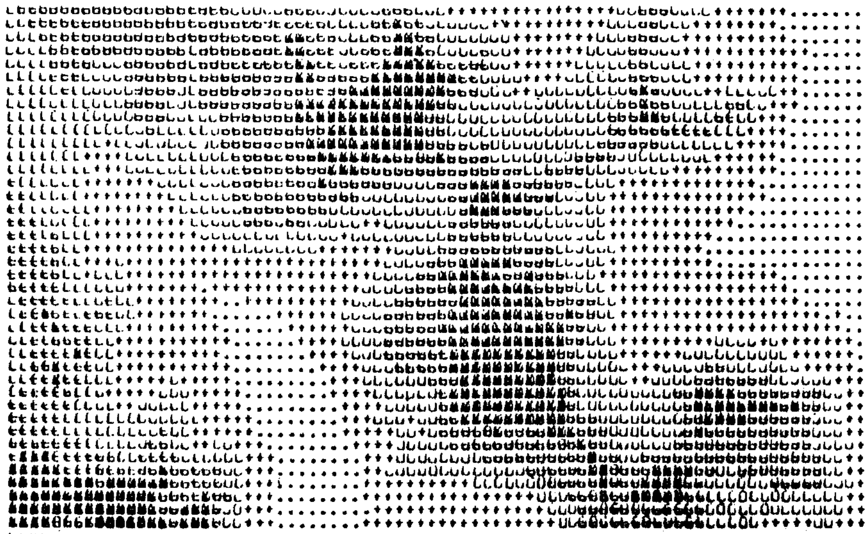
Manner of Snow Drift Computation

Point A on the plateau margin receives snow drift from three possible wind directions (SW, W, NW) while point B receives snow drift from five possible directions (S, SW, W, NW, N).

VALUES BETWEEN 5.000 8.000 11.000 14.000 17.000 20.000

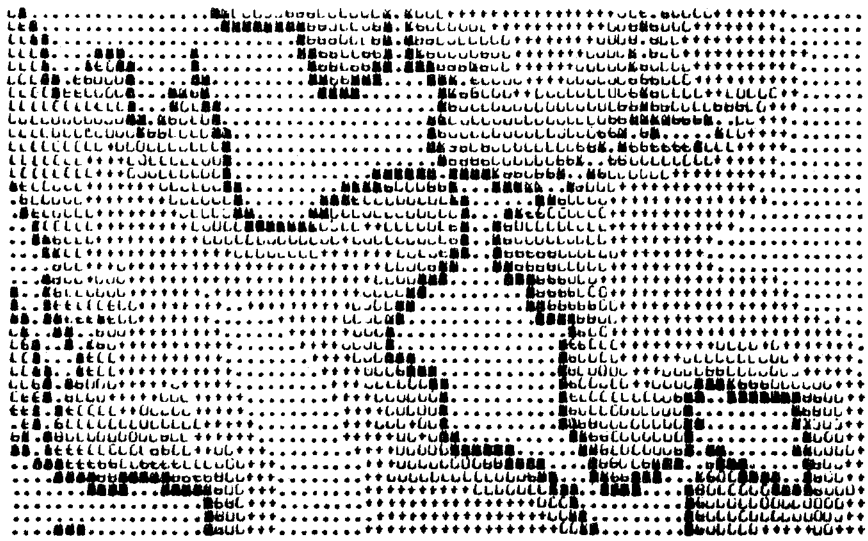
AEPER
FEB. SNOW COVER AFTER SNOWFALL (CM) - BLOCK 1

(a)



AEPER
FEB. SNOW COVER AFTER DRIFTING (CM) - BLOCK 1

(b)



AEPER
FEB. SNOW COVER AFTER AVALANCHING (CM) BLOCK 1

(c)

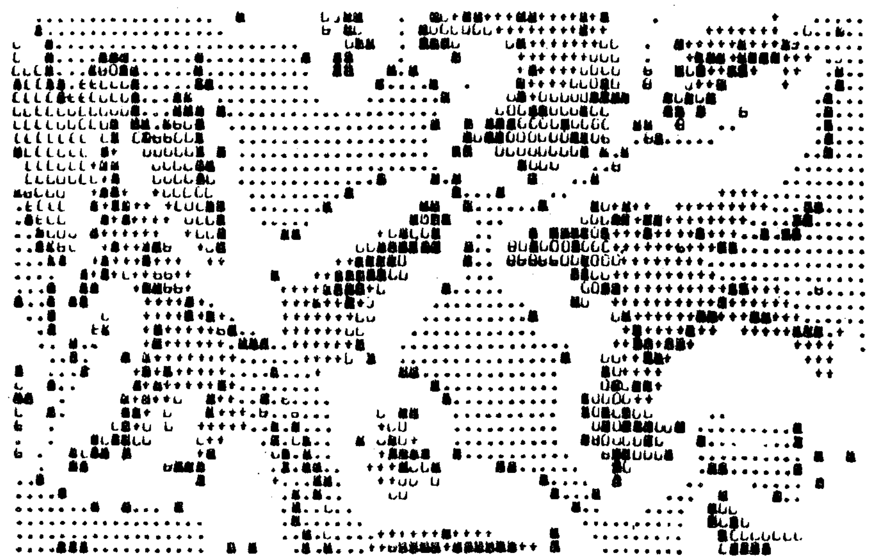


TABLE 7.4
February Snow Cover Distribution Over the Central Nupur Matrix Following (a) Initial Snowfall, (b) Snow Drifting, and (c) Snow Avalanching.

Determination of drift-erosion points (from topographic considerations)

↓

Pre-conditions for snow drifting to occur (sufficient snow thickness; dry snow)

↓

Determination of potential drift-deposition points using subroutine DEP (from topographic considerations)

↓

Drift-situation map (optional)

↓

Calculation of snow drift amounts received at each potential drift-deposition point according to the frequency of winds from different directions (using subroutine DRWIND).

↓

Check of drift computations and consideration of boundary effects (optional)

↓

Adjustments of snow cover after redistribution of initial snowfall by drifting. Results contained in matrix SNOW1 (I,J).

FIGURE 7.5

Computational Procedure for the Determination of Monthly Snow Drift Using Subroutine MDRIFT

AVALANCHE SITUATION MAP
MONTH=JANUARY

AVALANCHING POINT (*)
POTENTIAL AVALANCHE DEPOSITION POINT (O)
NOT INVOLVED IN AVALANCHE ROUTINE (.)

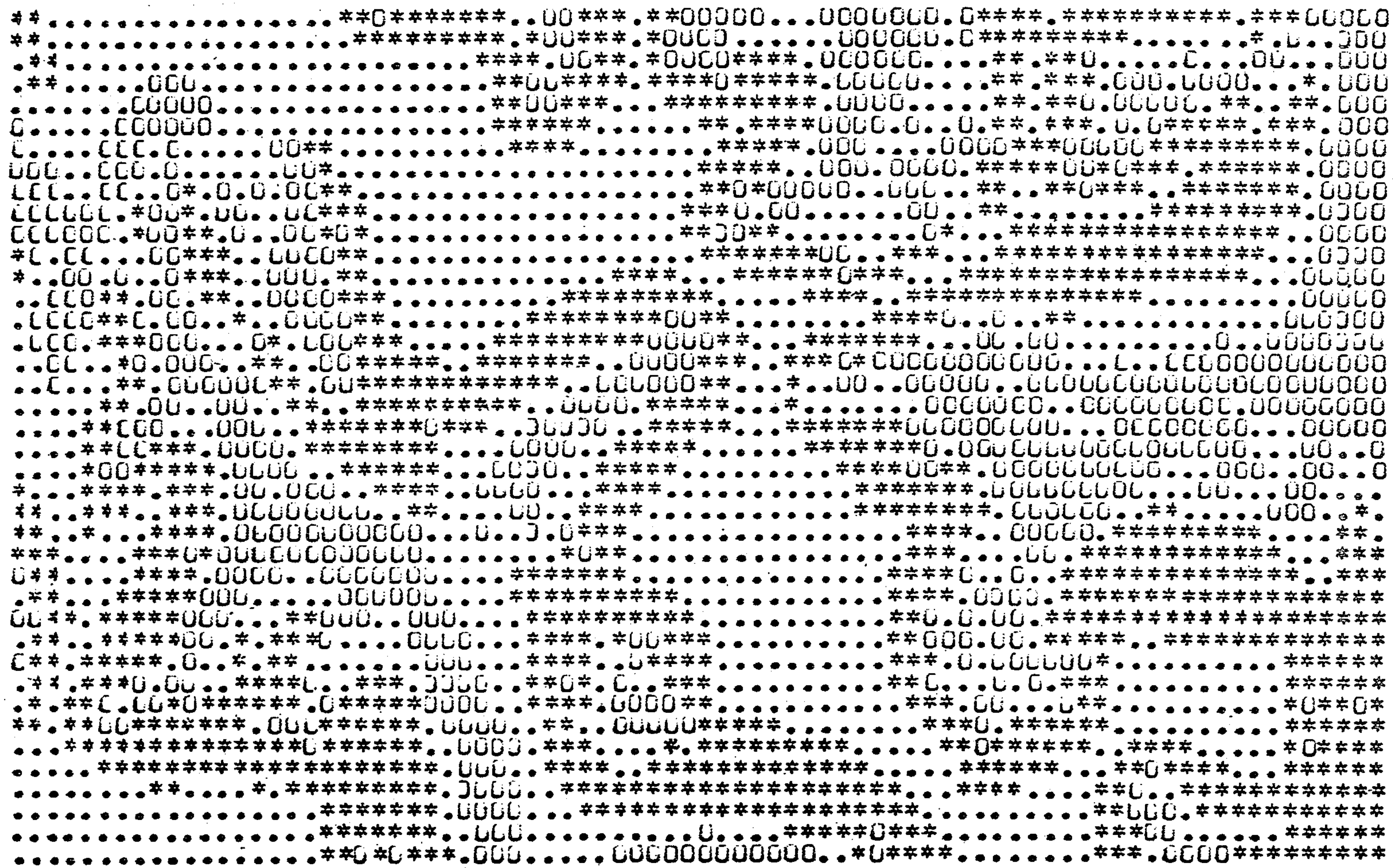


FIGURE 7.6
Snow Avalanche Situation Map Over the Central Nupur Matrix (for January).

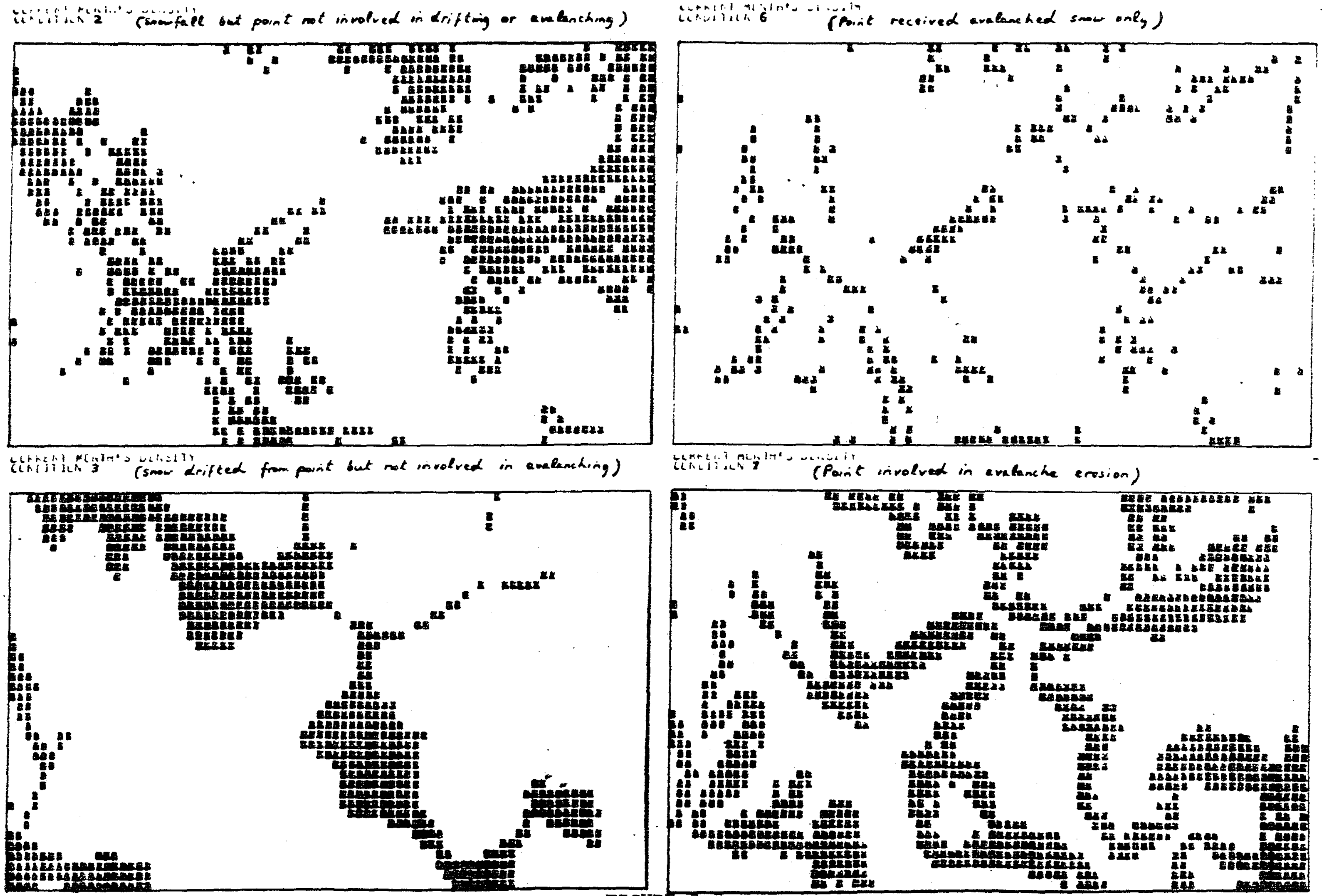


FIGURE 7.7
Distribution of the Grid-Points Over the Central Nupur Matrix which Satisfy Conditions 2, 3, 6 and 7 for the Determination of Snow Density (see Table 7.3).

CHAPTER VIII

ABLATION OF GLACIAL MATERIAL

Following the determination of the state and water equivalent of the glacial material at each grid intersection of the altitude matrix, consideration is given to the amount of material ablated during the particular month being simulated.

8.1 Characteristics of Snow and Ice Ablation

Glacial ablation encompasses all material which is removed from a glacier by melting, evaporation, sublimation, calving, wind action and avalanching; the latter two processes have been considered already, in the redistribution of glacial material by the accumulation processes. In areas where calving is not relevant, melting is the main process by which 'temperate' snow and ice is ablated.

The amount of material lost by evaporation is commonly only a fraction of that removed by melting (less than 5% of the total ablation according to Paterson, 1969) and on some glaciers these evaporation losses are offset by condensation. In the mass balance model, therefore, the determination of glacier ablation is essentially considered to be a question of calculating the amounts of snow and ice being melted.

As in the case of accumulation (sec. 7.1.1), the ablation of glacial material involves numerous causal factors resulting from the complex interactions between terrain and meteorological conditions.

For example, Hendrick et al. (1971), have related spatial and temporal variations in snowmelt to various aspects of the watershed environment including altitude, slope-aspect and forest cover. Whiteley et al. (1973) emphasised the effects of various hydrometeorological variables, e.g. net radiation, air temperature, vapour pressure of air and wind speed in determining snowmelt patterns. A recent review of the factors affecting snowmelt in mountainous environments has been undertaken by Obled and Harder (1979).

Since the melting of solid H_2O requires a high calorific input from radiation, sensible and latent heat exchange processes, the interaction of all the variables affecting snow and ice melt can be envisaged through an examination of the net energy exchange taking place between the glacial material and its environment. Following the fundamental work of Sverdrup (1935, 1936), Ahlmann (summarised in 1948a), and Wallén (1948), numerous studies have been undertaken to analyse the heat balance of glaciers in middle and high latitudes As noted in sec. 2.3.3, these studies of 'glacial meteorology' provide important information for assessing the variation in ablation rates.

Of particular interest are attempts at determining the relative importance of the various energy balance components at the glacier surface. Paterson (1969, pp.58-61) has collated thirty two such measurements, and results from some more recent studies are reported in Table 8.1. Generalisations from these results are made difficult by the variable and short duration of the studies, differences in definition of the energy sources, differences in the state of the glacier surface and by problems of calibration and measurement errors (Müller and Keeler, 1969). In general terms, however, these results confirm the predominance of solar radiation over convection and condensation as the main sources of heat for ablation.

The relative importance of net radiation appears to increase with increasing latitude, continentality, and altitude. The effect of the latter two variables can be demonstrated by reference to the energy balance measurements from eight Norwegian glaciers as reported by Haakensen and Wold (Eds.) 1981, p.45. A plot of altitude versus the percentage of heat supplied by net radiation for glacier ablation is shown in Fig. 8.1. Given the narrow latitudinal range of these glaciers, altitude (with the secondary influence of continentality) appears to be the main factor responsible for the variation in net radiation amounts. This is also evident from the results tabulated by Paterson (Evans, 1974 Fig.3.5).

Heat exchange by convection and condensation decreases with increasing altitude due to the decreasing temperature and moisture content of the air. However, in areas experiencing a strong maritime influence, with high air temperature, wind speed and humidity conditions, the proportion of heat provided by convection and condensation for glacier ablation can be as much as that contributed by net radiation (e.g. Streten and Wendler, 1968; Björnsson, 1972; Poggi, 1977; Anderton and Chinn, 1978). It should be emphasised that the relative importance of the various heat transfer processes can vary considerably during the course of the ablation season depending on local weather conditions (Wallén, 1948).

A development related to the measurement of energy balance components on glacier surfaces has been the emergence of studies concerned with 'snowmelt hydrology' (sec. 2.3.6). These studies have resulted from the need for an accurate understanding of snow and ice melt processes in order to forecast the runoff from glacierised and snow-covered regions. Such runoff forecasts are of great importance for water management and the design of water control structures. Studies of snowmelt are characterised by a number of approaches, viz. (i) degree-day method; (ii) multiple linear regression method; (iii) formulation of generalised empirical snowmelt equations involving a consideration of the energy exchange processes.

(i) A well established, though oversimplified, approach is to regress daily melt against the number of degree-days. The resulting regression equation can be used to predict future melt from an estimate of the degree-day totals (Collins, 1934; Linsley, 1943). Many simple models of meltwater prediction are based on this method (e.g. Zimmerman, 1972; Martinec, 1975). However, this is a rather unsatisfactory method since the melt coefficients will vary with local conditions of weather, snow state, and terrain. These problems are enhanced by differences in the calculation of mean daily temperatures (Arnold and MacKay, 1964).

(ii) Some researchers have attempted to improve snowmelt prediction by developing heat indices which go beyond the use of air temperature alone; or through more sophisticated statistical analyses involving the consideration of such meteorological variables as vapour pressure, net radiation and wind conditions (e.g. Psyklywec, 1968; Zuzel and Cox, 1975). In an effort to understand the factors affecting glacier ablation, glacial hydrologists have also commonly used multiple regression and other statistical methods to relate glacier runoff with meteorological parameters (e.g. Gudmundsson and Sigbjarnarson, 1972; Goodison, 1972; Jensen and Lang, 1973; Østrem, 1973).

(iii) The above statistical approaches to predict melt and runoff from snow and ice suffer from the problem that the results from a given region are usually not transferable to other areas. Furthermore, these approaches reveal little about the actual process of ablation and provide no idea about the relative influence of the different energy components involved in the melting process. Clearly, the only strictly correct way of computing the amount of melt and, at the same time, gaining a physical understanding of the ablation process is to use an energy balance approach.

Early attempts at a physical reasoning of the energy balance processes affecting snow melting (e.g. Wilson, 1941; Light, 1943) were followed by comprehensive snowmelt studies conducted by the U.S. Corps of Engineers and the U.S. Weather Bureau (U.S.Army Corps of Engineers 1956). These studies have contributed significantly to a more fundamental understanding of snow hydrology by deriving generalised snowmelt equations based on the theory of heat transfer to a melting snowpack. It should be noted,however, that the physical theory employed in these studies involves many approximations due to the unavailability of suitable data for a complete physical analysis of the energy balance.

Following the findings of the U.S. Corps of Engineers and especially with the advent of digital computer techniques which are particularly suited to the problems of snow hydrology (sec. 4.2), a number of conceptual simulation models of varying sophistication and detail have been developed to predict snowmelt runoff (see review by Anderson, 1973a). These models are based on the physical process of energy exchange between the snowpack and its environment. However, they differ according to the mathematical relationships used to model each component of the energy balance. Some of these models use an index (usually air temperature) to estimate the snow cover energy exchange (e.g. Rockwood, 1964; Anderson and Crawford, 1964; Riley et al. 1969; Willen et al. 1971; Eggleston et al., 1971; Anderson, 1973b; Bengtsson 1976; Quick and Pipes, 1976). Others incorporate a more complete theoretical treatment of the energy balance (e.g. Anderson E.A. 1968; 1976; Price and Dunne, 1976; Obled and Rosse, 1977).

While some of these models lack sufficient testing against experimental data, they provide a much better insight (than do the purely statistical approaches) into the processes of energy exchange and the interaction of the various factors controlling snowmelt. The energy balance approach has also been utilised in attempts at the modelling of the ablation processes on glaciers (Williams, 1974, 1975).

8.2 Simplified Energy Balance for 'Melt-Periods'

It has been established from the above discussion that glacier ablation is best estimated through a consideration of the various processes of energy transfer. In the present study, given the need to estimate melting of glacial material from rather generalised input values of climate and topography, the approach adopted is to determine the total energy available for melting (in monthly periods) after solving a simplified energy balance at each grid intersection where glacial material is present.

The restriction of the model to a study of 'temperate' glacier formation enables certain simplifying assumptions to be made regarding the energy exchange processes. It is assumed that the glacial material is isothermal at $0^{o}C$ and that its storage potential (i.e. free water capacity) has been satisfied. Thus any gain of heat will result in the melting of the glacial material.

Now the energy balance for an isothermal snow or ice body at melting point can be written as:

$$Q_M = Q_R + Q_H + Q_E + Q_P + Q_C$$

where Q_M = the heat available for melting

Q_R = the net radiative heat flux

Q_H = the sensible heat flux

Q_E = the latent heat flux (caused by condensation, evaporation or sublimation)

Q_P = the heat gained from precipitation

and Q_C = the conduction of heat from underlying ground

All the heat fluxes are expressed in cal.cm^{-2} hr^{-1}. The sources of energy received by the glacial material are conventionally defined as positive and the sinks are regarded as negative.

This equation has been widely applied in studies of the energy balance over melting snow and ice (sec. 8.1) and has been found to be generally satisfactory for determining melt at a point. The ground heat flux and the precipitation heat flux are usually negligible in the consideration of isothermal glacial material at $0^{\circ}C$ and are omitted in the present study. The heat available for melting (Q_M) is, thus, largely a function of the net radiative flux and the turbulent heat exchange (sensible and latent) at the snow or ice surface; i.e. $Q_M = Q_R + (Q_H + Q_E)$.

The computation of the energy balance is restricted to the 'melt-periods' when air temperature is greater than $0^{\circ}C$. Numerous studies of glacier ablation (sec. 8.1) show that surface melting does not occur until the temperature of the free air rises above the freezing point. Although the receipt of solar radiation is independent of air temperature, it has been pointed out (e.g. Callendar, 1952) that any surplus of solar radiation available for melting is highly dependent on air temperature. At below freezing temperatures, convective cooling of the glacial material prevents the radiation from being effective for melting.

The mass balance model, thus, utilises air temperature as an index for the determination of potential melt-periods. These are defined for each grid point by the subroutine MELTPR through the following considerations:

(i) an estimation of the number of days in the month that above freezing temperatures can be expected - DAYSML (I,J).

(ii) An estimation of the number of hours per day that temperatures greater than $0^{\circ}C$ occur in different months - DMHRS (I,J).

(iii) Determination of the mean temperature during that part of the day when temperatures are greater than $0^{o}C$ - TEMPML(I,J).

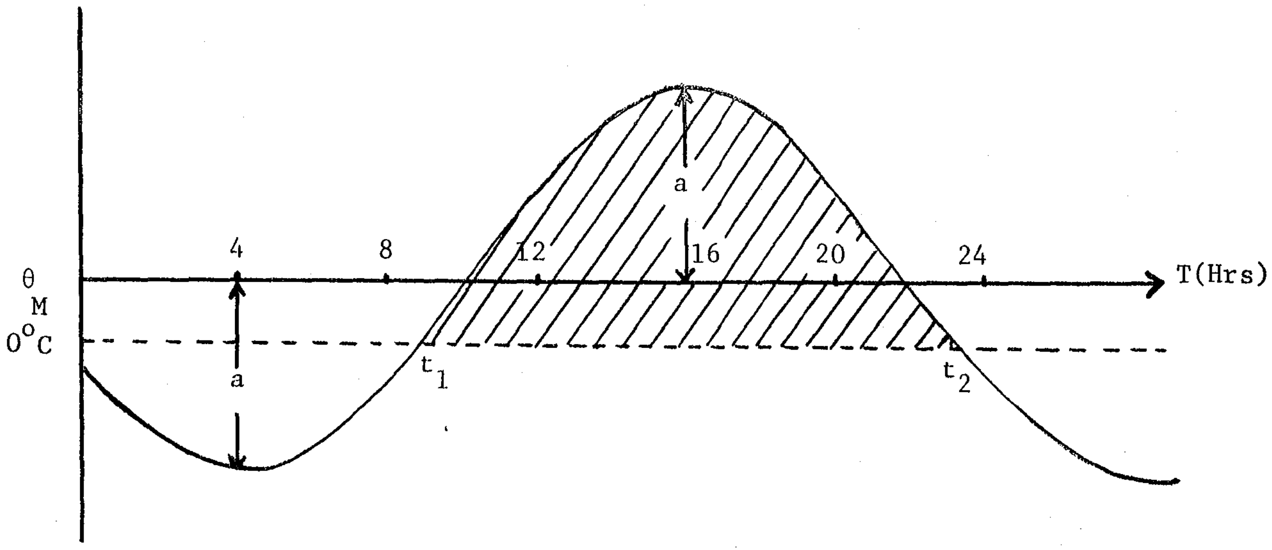
In the event of difficulty in obtaining information on these aspects of air temperature variations directly from climatological records, use was made of other indirect sources. These are discussed with reference to the Nupur matrix.

An indication of the diurnal variation of air temperatures in Vestfirðir is provided by the base meteorological station at Bolungarvik (location shown in Fig. 4.7). The diurnal air temperature curves for different months depicted in Fig. 8.2 are drawn using data on bi-hourly averages of air temperature for the period 1946-1953 provided by the Icelandic Meteorological Office. It can be seen (Fig. 8.3a) that the diurnal amplitude of air temperature is very small during mid-winter (being less than $1{\cdot}0^{o}C$ for the period November-February) and is greatest in mid-summer (greater than $2{\cdot}7^{o}C$ for the period May-August).

In the present study it was assumed that the pattern of the diurnal variations in air temperatures shown in Fig. 8.2 applies to all grid intersections of the altitude matrix since no information on the variations in the diurnal amplitude of air temperatures with altitude was available.

In Fig. 8.2 the mean monthly air temperature corresponding to each month's curve of the diurnal variation of air temperature is given in brackets. As an approximation, the relationship between the mean monthly temperature (θ_M) and the diurnal variation of air temperature (θ)

in any month can be represented by the following curve:



The equation of this curve can be expressed as :

$$\theta = \theta_M - a \cos t$$

where $t = \frac{\pi}{12}(T-4.0)$

Now it is clear from this curve that (i) if the mean monthly temperature (θ_M) is equal to or less than $-a^oC$, then it can be assumed that the air temperature (θ) throughout the day will be below 0^oC; (ii) conversely, when θ_M is equal to or greater than a^oC it can be assumed that the diurnal temperature is always greater than 0^oC; and (iii) if θ_M is between $\pm a^oC$ then the diurnal temperature curve will cut across the 0^oC axis (as shown in the above diagram).

The first case, with $\theta_M \leqslant -a^oC$, is not of interest for the computation of the energy balance since both DMHRS (I,J) and DAYSML (I,J) can be expected to equal zero while TEMPML (I,J) will be 0^oC or below freezing. For the other cases with $\theta_M \geqslant -a^oC$, an estimation needs to be made, in the first instance, for the mean number of days in the month on which air temperatures of greater than 0^oC can be expected. In view of the difficulty of obtaining this information from summary climatological tables, use was made of data concerning the number of days with frost at the base climatological station (Fig. 8.3b). The number of days with frost in different months can be correlated with the corresponding mean monthly air temperatures to provide an indication of the expected number of days with below freezing temperatures in any month (Fig. 8.4). Using the relationship developed in Fig. 8.4, the approximate number of days with air temperatures above 0^oC (DAYSML) can be obtained by subtracting the number of frost days from the total number of days in the month.

Furthermore, in the case with $\theta_M \geqslant a^oC$, DMMRS (I,J) can be taken as being equal to 24 hours and, thus, TEMPML (I,J) will be equal to the mean air temperature, TEMP (I,J) or θ_M. The determination of DMHRS and TEMPML is more complicated in the case with θ_M being between $\pm a^oC$. With reference to the curve representing the diurnal variation of air temperature (in the above diagram), it can be seen that the times (t_1, t_2) between which the air temperature (θ) is above 0^oC can be obtained as follows:

$$\theta_M - a \cos t = 0$$

$$\therefore \quad t = \arccos (\theta_M/a)$$

$$\text{i.e.} \quad t_1 = \arccos (\theta_M/a)$$

$$\text{and} \quad t_2 = (2\pi - t_1)$$

Now the average temperature (TEMPML), during the period of positive air temperatures represented by these times, is given by integrating the equation of the curve with limits of t_1 and t_2. Thus,

$$\text{TEMPML} = \frac{1}{t_2 - t_1}\int_{t_1}^{t_2} (\theta_M - a \cos t)dt$$

$$= \frac{1}{t_2 - t_1} \Big[\theta_M t - a \sin t\Big]_{t_1}^{t_2}$$

$$= \frac{1}{t_2 - t_1} \{ \theta_M(t_2 - t_1) - a(\sin t_2 - \sin t_1)\}$$

$$= \theta_M - \frac{a}{t_2 - t_1} (\sin t_2 - \sin t_1)$$

By substituting $t_2 = 2\pi - t_1$,

$$\text{TEMPML} = \theta_M + \left(\frac{a \sin t_1}{\pi - t_1}\right)$$

The duration (in hours) during which the temperature is above 0°C (DMHRS) is given by:

$$\text{DMHRS} = \frac{12}{\pi} (t_2 - t_1)$$

Following the determination of these 'melt-period' characteristics for each grid intersection, consideration is given to the solution of the simplified energy balance. In any month the energy balance is only determined at those points where TEMPML (I,J) is greater than 0°C and where glacial material in the form of snow, firn or ice is present. The heat available for melting is summed for that part of the day when temperatures are above 0°C (DMHRS) and for the number of days that the mean temperature is expected to be above $0^{\circ}C$ (DAYSML). If needed, a simple line printer map depicting those points with TEMPML $> 0^{\circ}C$ can be constructed to reveal the distribution of potential areas experiencing melting conditions in any month (e.g. Fig. 8.5 for April in the 'test' matrix).

The manner in which the net radiation and turbulent heat exchange was computed is discussed in the following sections. The radiation balance, consisting of shortwave and longwave radiation components, is determined by the subroutines SWRAD and LWRAD respectively. The turbulent energy exchange, consisting of the transfer of sensible and latent heat fluxes, is determined by the subroutine TBHEAT. The net energy exchange by these processes is summed at each grid intersection to determine the monthly totals of energy (Q_M) available for melting the glacial material. The variation of Q_M for July over the Nupur 'test' matrix is shown in Fig. 8.6d.

The amount of monthly melt (M) taking place, at each grid point with a positive value of Q_M, is then determined from the following relation:

$$M(cm) = Q_M/L$$

where L = the latent heat of fusion (80 cal.gm^{-3}).

The melting of glacial material is considered in the sequence (1) snow (2) firn (3) ice. The amount of material melted in any of these states will depend on the water equivalent of material present in each of these states and the net energy available for melting. First the available energy is used to melt any snow present at the grid point in question; if any surplus energy is left after all the snow has been melted, consideration is given to the melting of any firn and finally ice that may be present, taking into account the different albedo of the newly revealed material. Following the melt computations, appropriate adjustments of the monthly mass balance are made to determine the amounts of snow, firn and ice (if any) remaining at each grid intersection. These aspects of the mass balance are programmed in the main routine of GSP1 (Appendix A).

8.3 Computation of the Net Radiation Balance

The net radiative flux, Q_R, in the simplified energy balance equation (sec. 8.2) can be expressed as the balance of shortwave and longwave radiation components:

$$Q_R = R_G(1-\alpha) + R_L$$
$$= (D + d)(1-\alpha) + L\downarrow - L\uparrow$$

where R_G = the incoming 'global' shortwave radiation consisting of both direct sunshine (D) and diffuse radiation (d) that has been scattered and reflected by the atmosphere and clouds.

α = the albedo of the surface material for shortwave radiation. The product $(D+d)(1-\alpha)$ represents the net global radiation absorbed by the glacial material.

R_L = the net longwave radiation consisting of incoming longwave radiation from the atmosphere ($L\downarrow$) and the upward flux of longwave radiation emitted by the surface of the glacial material ($L\uparrow$).

In the mass balance model the two major components of the net radiation balance, R_G and R_L are computed by separate subroutines. The shortwave radiation balance is determined by the subroutines SWRAD, GLSR and

SHADE while the longwave radiation balance is determined by the subroutine LWRAD.

8.3.1 Shortwave Radiation Balance

The amount of global radiation received at any point on the earth's surface varies with the latitude, date (declination of the sun), time of day (hour angle), cloudiness, transparency of the atmosphere, altitude, slope gradient and aspect, and the effect of shading by surrounding topography. Various empirical relations have been developed between solar radiation receipt and such meteorological factors as air temperature, amount of cloudiness, cloud type, and duration of sunshine (e.g. Vowinckel and Orvig, 1962; Chang and Root, 1975), while tables providing information on potential radiation amounts for specified times, slopes and latitudes have been compiled (e.g. Frank and Lee, 1966). However, a general solution to the problem of solar radiation receipt is best obtained through the computation of shortwave radiation amounts from fundamental principles based on a combination of astronomical, meteorological and topographical information.

Garnier and Ohmura (1968, 1970) presented a useful method for the calculation of solar radiation, allowing for the declination of the sun, time, latitude of the site, and the gradient and aspect of the surface slope at each point. This method has been widely applied for the areal estimation of solar radiation amounts (e.g. Goodison, 1971; Greenland and Clothier, 1975). Williams et al. (1972) have developed the method further to analyse the influence of global radiation on mountain glacierisation.

In the present study the method of Garnier and Ohmura, together with its refinement by Williams et al., has been adopted to estimate the direct and diffuse components of solar radiation at each grid intersection of the altitude matrix during the course of the 'melt-periods'.

The computational formulae used to calculate clear-sky direct and diffuse radiation, together with the procedure for determining the effect of topographic shading on radiation receipt are described first. This is followed by a discussion on the selection of various parameters necessary for the estimation of monthly global radiation amounts, e.g. sun's declination, duration of daylight and atmospheric transmissivity. Finally the manner in which the absorbed global radiation at each grid point was determined, after allowing for the effects of altitude, cloudiness and surface albedo, is outlined.

<u>Direct solar radiation</u>. According to Garnier and Ohmura (1970) daily totals of direct shortwave radiation (D) falling on a slope can be calculated by integrating the following equation over small increments of time (H):

$$D = I_o/r^2 \int_{H_1}^{H_2} p^m \cos(\chi \Lambda S)\, dH$$

where I_o = the solar constant

r = the radius vector of the earth's orbit

p = the mean-zenith-path transmissivity of the atmosphere

m = the optical air mass

H = the hour angle measured from solar noon (+ve after noon), the integral being taken over the duration of sunlight on the slope

$\chi \Lambda S$ = the angle of incidence of the solar beam on the slope

$$\cos(\chi\Lambda\ S) = \Big[(\sin\phi\cos H)(-\cos A\ \sin\theta) - \sin H\ (\sin A\ \sin\theta\ ) + (\cos\phi\cos H\ \cos\theta)\Big]\cos\delta + \Big[\cos\phi\ (\cos A\ \sin\theta) + \sin\phi\ \cos\theta\Big]\sin\delta$$

where ϕ = the latitude

θ = the slope gradient

A = the slope aspect

δ = the declination of the sun

A computer program for the computation of direct radiation receipt by using the above expression has been provided by Fuggle (1970). This program was modified to calculate global radiation amounts, in increments of sixty minutes, after including expressions for the computation of the diffuse radiation component and for the effect of shading by surrounding topography.

Diffuse solar radiation. The accurate estimation of the diffuse radiation flux is complicated (for a detailed discussion refer to Obled and Harder, 1979). As a simplification, it is usually assumed to be isotropic (valid for slope gradients up to 30^o according to Kondratyev and Manolova, 1960 and hence for most glacier surfaces) and is thus regarded as being independent of slope aspect. Its variation on differing slopes can be determined from the following equation (Kondratyev, 1969):

$$d_s = d_o \cos^2(\theta/2).$$

where d_s = the diffuse radiation received on a slope of gradient θ.

and d_o = the diffuse radiation received on a horizontal surface.

Measurements of diffuse radiation are, however, rarely available and it is necessary to obtain d_o from theoretical considerations. A commonly used expression for calculating this flux under clear sky radiation is given by List (1966, p.420 attributed to Fritz):

$$d_o = 0{\cdot}5\ (1 - a_w - a_o)\ I_t - I_d$$

where a_w = the proportion of radiation absorbed by water vapour (assumed to be 7%)

a_o = the proportion absorbed by ozone (assumed to be 2%)

I_t = the extraterrestrial radiation

$$= \frac{I_o}{r^2} \int_{H_1}^{H_2} \cos z_s \, dH$$

I_d = the direct radiation on a horizontal plane

$$= \frac{I_o}{r^2} \int_{H_1}^{H_2} p^m \cos z_s \, dH$$

z_s = the zenith angle of the sun

and 0·5 expresses the assumption that half of the radiation is scattered forward and half backward. Thus, the combined expression for the estimation of the diffuse radiation flux on a slope (d_s) becomes

$$d_s = 0{\cdot}5 \cos^2(\theta/2) \frac{I_o}{r^2} \int_{H_1}^{H_2} (0{\cdot}91 - p^m) \cos z \, dH$$

Although the validity of this expression is limited by the assumption of isotropic conditions, since the intensity of clear sky diffuse radiation is greatest in the vicinity of the sun and near the horizon (Kondratyev, 1969), the comparison of calculated results,obtained by using this expression, with measured values has been shown to be reasonable by Williams et al. (1972).

The effect of shading by topography. A further complicating factor in the determination of global radiation in mountainous terrain is the possible effect of shading by topographical features around the grid point being considered. This effect can achieve a marked significance during early morning and evening when the sun is less than 30^{o} above the horizon, during low sun periods in mid-latitudinal winters, and in high latitudes (over 60^{o}) at all seasons. For example, Wendler and Ishikawa (1974) report that screening by mountains around the McCall Glacier in Alaska (latitude 69^{o} 18'N.) reduces direct solar radiation amounts by about 13% during the ablation season (with a solar declination of 20^{o}), by 68% at the equinox and by more than 90% at a solar declination of 10^{o}.

Computational procedures for the estimation of the shading effect have been developed by a number of researchers (e.g. Ohmura, 1970; Ferguson et al., 1971; Goodison, 1971; and Williams et al. 1972). In the present study a computer subroutine (SHADE) based on the method described by Williams et al. (1972 p.528) has been written to consider the effect of shading by surrounding topography. Basically it considers the altitude at each grid intersection and searches along the line of the sun's azimuth (for a distance of 1 km. along the ground) to determine the presence of any point high enough to block the sun. If the grid point under consideration is found to be in shadow, the computation of the direct solar radiation component is omitted for the appropriate increment of the hour angle being considered. Further details of this method can be obtained by referring to the discussion in Williams et al.(1972) or by examining the listing of the SHADE subroutine in Appendix A.

In the Nupur area, the effect of topographic shading on solar radiation receipt was found to be insignificant at the height of the ablation season (with solar declination greater than 15^{o}). For example, the reduction in solar radiation received at all grid points in the 'test' matrix in July as a result of topographic shading was found to be less than 14% of the radiation received without considering the shadow effect (Fig. 8.7). This is similar to the findings of Wendler and Ishikawa (see above). The greatest reduction (greater than 8%) was found on the steep north-facing cirque slopes. Over the greater part of the matrix, however, the percentage reduction of solar radiation is low, perhaps due to the limited relief amplitude. As a consequence, differences in monthly solar radiation receipt (for the height of the ablation season) due to the shading effect are hardly discernable (Fig. 8.8).

In the mass balance model, the energy balance computations are only of interest for the determination of glacial material melting during the ablation season, i.e. when the effect of shading on solar radiation is least significant. The slight additional accuracy in solar radiation computations as a result of including the shadow routine was not thought to be justified by the increase in computer time and thus the shading effect was not considered in the Nupur mass balance simulations.

The above procedures for the determination of the direct solar radiation, the diffuse radiation, and the effect of shading by topography (optional) constitute the major components of the subroutine GLSR, concerned with the determination of daily global radiation totals. In addition to these procedures, a number of other expressions relating to the computation of the monthly net shortwave radiation absorbed, by glacial material at any grid point, need to be specified. Considerations relating to the selection of the declination angles of the sun, duration of daylight, transmissivity of the atmosphere, and the albedo of the glacial material are defined in the main 'driving' subroutine for short-wave radiation, viz. SWRAD.

Selection of sun's declination, duration of daylight, transmissivity and albedo. In order to reduce computer time, the computations for solar radiation were performed for the 8th and 24th of each month for the duration of the 'melt-period' at each grid intersection of the altitude matrix. (With the use of large matrices - e.g. 100 x 100, the increased computer time needed for mass balance computations, makes it impractical to compute solar radiation on more than one representative day in the middle of each month). Values for the sun's declination and duration of daylight on these days, for the appropriate latitude of the study area (65^{o}N in the case of the Nupur matrix), can be obtained from the Smithsonian Meteorological Tables (List, 1966 Tables 169 and 171) and their variation throughout the year is depicted in Fig. 8.9. The values for the 8th of the month are taken to correspond with the first half of the 'melt-days' in the month (i.e. DAYSML (I,J)/2), while the declination and duration of daylight for the 24th of each month is considered to be representative for the remainder of the days in the month on which melting of glacial material is expected.

Values for atmospheric transmissivity are not readily available from tables or theoretical considerations since they can vary considerably from place to place, being a function of the local weather and air mass characteristics. Measurements of atmospheric transmissivity reveal that it usually varies seasonally, being highest in winter and lowest in summer, because the atmosphere contains more water vapour and dust during the summer months. For the Icelandic study areas, an indication of the monthly variation in atmospheric transmissivity was obtained from a comparison of calculated values (using the subroutine SWRAD) and measured values (from the Icelandic Meteorological Office) of global radiation at Reykjavik (Fig. 8.10).

The albedo of the surface glacial material is a critical factor in the determination of absorbed solar radiation amounts and estimation of its value at each grid point of the matrix is essential for the computation of the shortwave radiation available for melting. Several investigations have shown that the albedo of snow and ice surfaces is influenced by the surface characteristics of the glacial material (e.g. density, water-content, size, shape and structural arrangement of the snow or ice crystals), time since deposition, cloudiness, time of day (solar altitude), season, latitude and altitude, as well as the aspect and gradient of the surface slope (e.g. Hubley, 1955; U.S. Army Corps of Engineers, 1956; Slaughter, 1969; Korff and Vonder Haar, 1972; O'Neill and Gray, 1973; Dirmhirn and Eaton, 1975; Petzold, 1977; and Bolsenga, 1977).

The largest variations in albedo values are usually attributed to changes in the state of the glacial material (snow → firn → ice), cloudiness, and solar altitude. In the mass balance model the initial value of the albedo of glacial material present at any grid point is selected by considering the state of the surface material. Based on a review of the albedo values for different surfaces reported in the literature, the surface material is given a mean albedo value of 0·3 for ice, 0·4 for firn and 0·5 for old snow. These values refer to cloud-free conditions and for the times when the sun's rays are perpendicular to the surface.

To account for the increase of albedo with cloudiness, this mean value is increased by 0·01 for a one-tenth increase in cloud cover (as indicated by several of the above-mentioned studies). The monthly mean cloud cover is obtained from the base meteorological station (Fig. 6.1c). Lack of data on the variation of cloud conditions diurnally or with altitude prevented a more detailed consideration of the effect of cloudiness on albedo.

The variation of albedo with the angle of incidence of sun's rays is estimated (in subroutine GLSR) by constructing approximate curves relating the albedo of snow, firn, and ice to changes in solar altitude, in a similar manner to those presented by Geiger (1965, p.17) for sand and water surfaces. In Fig. 8.11, the albedo values for snow, firn and ice (α) at any solar altitude (θ) can be estimated from the following relation:

$$\alpha = (1 - \alpha_o)\, e^{-k\theta} + \alpha_o$$

where α_o = the mean albedo for the surface material at a solar altitude of 90^o.

and k = a constant which equals 0·0768 (for snow); 0·0796 (for firn) and 0·0819 (for ice).

These curves indicate that the albedo values decrease with increasing solar altitude to about 40^{o} (Fig. 8.11). The estimated albedo values for particular solar altitudes are generally in agreement with measured albedo values reported in the literature for glacial material at corresponding solar altitudes.

<u>Computational procedure for the determination of absorbed global radiation.</u> Having discussed the major components of the short-wave radiation routines, the computational procedure for the determination of net global radiation absorbed by glacial material at each grid point is described with reference to the Nupur area.

In any given month, the potential melting period (DAYSML), for the grid points satisfying the conditions for melting to occur (sec. 8.2), is divided into two equal parts and the shortwave radiation balance is determined for one representative day in each part using the declination, duration of daylight and transmissivity values for the 8th and 24th day of the month. For each of these representative days the subroutine GLSR determines the mean daily shortwave radiation received at each grid point by summing sixty minute increments of radiation receipt by both the direct and diffuse components. The total time period over which the radiation balance is calculated is controlled by the duration of daylight or the period for which air temperatures are expected to be greater than $0^{o}C$ (i.e. DMHRS).

The procedures used for computing, the clear-sky global radiation have been discussed above. It should be noted that the selection of the solar constant (varying from 1.94 to 2.07 cal. cm^{-2} min^{-1}) is dependent on the variation of the sun's declination. Furthermore, the determination of the optical air mass (m) using the secant approximation ($m = \sec z_s$) has been shown to be in error with large zenith angles (Garnier and Ohmura, 1968, 1970). Thus, a separate determination of the optical air mass for zenith angles greater than 70^o is made using the data provided in the Smithsonian Meteorological Tables (List, 1966 Table 137).

The clear-sky global radiation received for each sixty minute increment is corrected for the effect of altitude and cloudiness. An increase in the potential solar radiation with altitude is expected under cloudless sky conditions due to a decrease in optical air mass, the shorter path traversed by the solar beam, and changes in atmospheric transmissivity brought about by decreased quantities of moisture content and dust pollution at high altitudes. While the rates of increase will vary from place to place, a mean increase in incident solar radiation of about 1% per 100 m increase in altitude is usually quoted (e.g. Kuzmin, 1972 p.123; Barry and Van Wie, 1974 p.73; Barry and Chorley, 1976 p.57).

This increase in radiation receipt with altitude will be offset by any increase in cloudiness with altitude. However, a lack of information concerning the variation of cloud cover with altitude prevents this aspect from being incorporated in the mass balance model. Instead, the global radiation is adjusted for mean cloud cover by use of the following relation developed by Ångstrom-Savinov (Kondratyev, 1969):

$$R_C = R_G\ (1-(1-k)n)$$

where R_C = the cloudy-sky global radiation

R_G = the clear-sky global radiation

k = an empirical coefficient characterising shortwave radiation transmission by clouds

n = the cloud cover (in tenths)

Kondratyev (1969, p.468) provides a table for obtaining the empirical coefficient, k, value for different latitudes. For the Nupur area (65^{o}N) a value of k = 0.45 was selected.

The net incoming global radiation is further adjusted for the effect of the surface albedo to determine the amount of solar radiation absorbed by the glacial material during each time increment. The values from all the time increments are summed to yield the daily total of absorbed global radiation for the two representative days in each month. The mean daily radiation values for each half of the monthly melt-period are multiplied by the number of 'melt-days' in each part of the month to determine the total monthly global radiation absorbed at each relevant grid intersection (given by the matrix GRBL (I,J)).

An indication of the accuracy of the global radiation calculations was provided by comparing the results obtained from the subroutine SWRAD with measurements of global radiation conducted on Baegisárjökull in northern Iceland (Fig. 3.9) by Helgi Björnsson during the summer of 1968 (Björnsson, 1972 and personal communication). Daily calculated (using actual daily cloudiness) and measured values of global radiation for the month of July are compared in Fig. 8.12. The fit is thought to be reasonable with the total monthly global radiation calculated to be 13,550 cal.cm^{-2}.month^{-1} compared with the measured value of 14,210 cal.cm^{-2}.$month^{-1}$ for July. In the mass balance global radiation is only estimated for the 8th and the 24th of each month; when the calculated global radiation values corresponding to these days (in the above example, 478 and 380 cal.$cm^{-2}day^{-1}$ respectively) are multiplied by half the number of days in the month and summed, the total monthly global radiation receipt for July is estimated to be 13,250 cal.cm^{-2} which again corresponds fairly closely to the measured value of 14,210 cal.cm^{-2}.

An illustration of the computational method for the estimation of absorbed global radiation is provided in Table 8.2 by listing the diurnal variations (in 60 minute time intervals) of the various SWRAD subroutine components. These estimates of solar radiation are for the 1st of July in the above example.

The variation of the shortwave radiation receipt over the Nupur 'test' matrix is illustrated by means of line-printer computer maps in Fig. 8.6a. The combined effect of slope gradient and aspect, in particular, is clearly portrayed.

8.3.2 Longwave Radiation Balance

The shortwave radiation absorbed by the earth's atmosphere is re-radiated in all directions as atmospheric radiation. Snow and ice surfaces can be regarded as black bodies with respect to longwave radiation, absorbing all such incident radiation and emitting the maximum possible radiation corresponding to their temperature.

The net longwave radiation at the surface of the glacial material is, thus, the difference between the downward atmospheric radiative flux or 'counterradiation' (L↓) and the upward flux of 'terrestrial' radiation (L↑). Over melting snow and ice L↑ is usually greater than L↓ and the net longwave radiation acts as a heat sink. However, during overcast conditions (with a cloud cover of greater than about eight tenths) the contribution of longwave radiation for melting may become important (e.g. Hubley, 1957; Ambach, 1974).

The magnitude of both L↓ and L↑ is usually determined by reference to the Stefan-Boltzmann's law which states that a radiation flux, F, is proportional to the temperature (T) of the radiating body:

$$F = \sigma \varepsilon T^4$$

where σ = Stefan-Boltzmann constant ($4{\cdot}92 \times 10^{-9}$ cal.cm^{-2}hr^{-1} ^{o}K^{-4})

ε = infrared emissivity (usually taken as unity for snow and ice)

Since the glacial material is assumed to be at melting point (T = 273^{o}K), its radiative flux (L↑) is easily determined from the above relation.

The incoming longwave radiation flux (L↓), emitted from carbon dioxide and water vapour in the atmosphere, is a function of numerous factors including air temperature, vapour pressure, cloud cover, exposure to the sky hemisphere, and the dust content of air. Due to difficulties in measuring some of these factors, a number of empirical equations have been proposed to estimate clear-sky atmospheric radiation from surface air temperature and vapour pressure alone, e.g. by Ångstrom, Brunt and Swinbank (see Sellers, 1965 pp.53-54).

Following Williams (1974) the equation proposed by Swinbank (1963) was used to estimate the net longwave radiation under clear skies (R_o) from surface temperature alone:

$$R_o = \sigma \varepsilon T^4 (9 \cdot 35.10^{-6} T^2 - 1)$$

where T = the screen height air temperature (usually at 2 m)

Sellers (1965 p.61) notes that in the event of a large difference between the air temperature (T) at screen height and the surface temperature of melting snow and ice (T_s), it is necessary to apply a correction factor to the above equation:

$$R_L = R_o + 4\varepsilon T^3 (T - T_s).$$

Furthermore, the atmospheric longwave radiation has been shown by Kondratyev (1969, p.681) to vary with slope angle (θ) in the following manner:

$$L_s\downarrow = L_h\downarrow \cos^2(\theta/2)$$

where $L_s\downarrow$ = the longwave radiation received by a slope surface

$L_h\downarrow$ = the longwave radiation received by a horizontal surface.

Finally the longwave radiation has to be adjusted for cloudy conditions since the presence of cloud will tend to increase the amount of counterradiation. Various types of relationships have been suggested to estimate longwave radiation recorded under cloudy conditions. These usually take the form:

$$R_c = R_o(1-kn^2)$$

where R_c = the net longwave radiation under cloudy conditions

k = an empirical constant

n = the cloud cover (in tenths)

The value of k varies according to cloud type and height. However, satisfactory results of the longwave radiation balance have been found to occur in the Alps with a value of 1·4 for k (e.g. Hoinkes and Untersteiner, 1952). This value has also been adopted for estimating longwave radiation exchange on glaciers in Iceland and Norway (e.g. Liestøl, 1967; Messel, 1971; Björnsson, 1972).

The net longwave radiation balance (R_L) at the surface of the melting glacial material can thus be estimated from the combination of the above equations:

$$\begin{aligned} R_L &= R_o(1-kn^2) + 4\sigma\varepsilon\, T^3(T-T_s) \\ &= \sigma\varepsilon\, T^4((9.35.10^{-6}T^2)(\cos^2\theta/2) - 1)(1-kn^2) + 4\sigma\varepsilon T^3(T-T_s) \end{aligned}$$

Williams (1974 Fig.1 p.27) shows that satisfactory results were obtained by using this expression when compared with daily measurements of longwave radiation on 'Boas' glacier in Baffin Island. This expression also yielded similar values of daily and monthly longwave radiation totals to those computed by Björnsson (1972 and personal communication) for the energy exchange on Baegisárjökull in northern Iceland.

In the mass balance model, the above expression is used to estimate the net longwave radiation balance at each grid point where glacial material is present. TEMPML is used to indicate the mean air temperature for the period (DMHRS) that diurnal air temperature is expected to be above $0^{o}C$. The diurnal net longwave radiation balance for this period is multiplied by the number of melt-days (given by DAYSML) to determine the monthly longwave radiation balance at each relevant grid point.

The variation of net radiation (balance of shortwave and longwave components) for July in the 'test' matrix is depicted in Fig.8.6b. The longwave radiation balance was found to be negative (with a mean cloud cover of 0·74) at all points of the matrix.

8.4 Estimation of the Turbulent Energy Exchange

The turbulent transfer of energy taking place at the glacial material surface is a function of the sensible (H) and latent (E) heat fluxes. These processes can contribute significantly to the melting of glacial material since during 'melt-periods', with air temperatures above $0^{o}C$, the sensible heat flux will always be directed towards the snow and ice surface (assumed to be at $0^{o}C$). Similarly the latent heat flux also acts as a heat source for the melting of glacial material since positive air temperatures coupled with high relative humidity values (as in Iceland) result in the vapour pressure increasing with height leading to the predominance of convection over evaporation.

These turbulent energy exchange processes are, however, most difficult to determine in a mountainous environment due to the uncertaihity in specifying the spatial variation of point wind speed and air temperature values (chap. VI). In studies of the energy exchange over melting snow and ice (see references in sec. 8.1) it is common practice to estimate the turbulent heat exchange by using conventional aerodynamic equations based on the analysis of temperature, vapour pressure and wind velocity gradients in the lower atmosphere. The development of these aerodynamic equations from the theory of transfer coefficients within the surface boundary layer is discussed elsewhere (e.g. Sellers, 1965 chap.10; Anderson, 1976 p.8 ff; Price, 1977 p.17 ff).

In the mass balance model rough estimates of sensible (H) and latent (E) heat transfer are obtained by using the following simplified approximations of the turbulent heat transfer equations:

$$H = \rho.c_p.K_h\ (T_a - T_s)\ \text{cal cm}^{-2}\ \text{hr}^{-1}$$

$$E = \rho.L.K_w(0.622/p)(e_a - e_s)\ \text{cal cm}^{-2}\ \text{hr}^{-1}$$

where K_h = the transfer coefficient for heat (cm/hr)

K_w = the transfer coefficient for water vapour (cm/hr)

ρ = the air density (gm.cm^{-3})

c_p = the specific heat of air at constant pressure (taken as $0{\cdot}24\ \text{cal.gm}^{-1}.^{\circ}\text{C}^{-1}$).

L = the latent heat of vaporisation of water (taken as 597·3 cal.gm^{-1}).

p = the atmospheric pressure (mb)

T_a = the air temperature ($^{\circ}$C) at a height z (taken as 2m) above the surface

T_s = the surface temperature ($^{\circ}$C) of the melting glacial material (taken as 0°C)

e_a = the vapour pressure of the air (mb)

e_s = the vapour pressure of the melting glacial material (taken as 6.11 mb)

These equations are based on the assumption that the shear stress and vertical fluxes of heat and water vapour are constant with height within the surface boundary layer. Furthermore they assume that the coefficients for heat (K_h) and the coefficient for moisture (K_w) equal the coefficient for the vertical transfer of momentum (K_m) such that:

$$K_h = K_w = K_m = \frac{k^2 u_z}{[\log_e (z/z_o)]^2} \text{ cm.hr}^{-1}$$

where u_z = the wind velocity at height z (cm.hr^{-1})

k = von Karman's constant (taken as 0·4)

z = the height above the surface (taken as 200 cm)

z_o = the roughness parameter (cm)

The assumption regarding the equality of the transfer coefficients is generally valid under neutral and near-neutral stability conditions (i.e. when the air temperature gradient is equal to or nearly equal to the dry adiabatic lapse rate). In such conditions the wind velocity can be assumed to vary with the logarithm of the height.

The above method of calculating the turbulent heat flux is thus, strictly only applicable in an atmosphere which is neutral. While some researchers (e.g. Grainger and Lister, 1966) believe that the logarithmic law may be valid for both neutral and extremely stable atmospheres, other studies (e.g. de La Casinière, 1974) reveal that these assumptions are not valid under the generally stable conditions predominating over melting snow and ice surfaces.

Under stable conditions the air near the surface is cooled and assumes a higher density. This air tends to maintain its position when it is disturbed by turbulence, resulting in decreasing rates of turbulent exchange. A stability correction, thus, needs to be applied for the calculation of the transfer coefficients. Following Price and Dunne (1976), the following simple correction for stable conditions given by Monteith (1957) is used in the mass balance model:

$$K_s = K_h/(1 + \sigma R_i)$$

where K_s = the transfer coefficient under stable conditions

K_h = the transfer coefficient under neutral conditions

σ = a constant with a value of 10·0 (according to Webb, 1970)

R_i = the Richardson number

The Richardson number (positive for stable, zero for neutral, and negative for unstable conditions) is given by:

$$R_i = \frac{g.\ z.\ \Delta T}{[T_{abs}(\Delta u)^2]}$$

where g = the acceleration due to gravity (cm hr^{-2})

ΔT = the temperature difference between the surface and the height z (^oK)

T_{abs} = the temperature of the air layer (^oK)

Δu = the difference in wind speed between the surface and height z.

The solution of these equations is further simplified by known boundary conditions at the melting snow and ice surface. The temperature (T_o) at the surface of the glacial material can be regarded as 0^oC while the vapour pressure (e_o) can be assumed to equal the saturation vapour pressure of water (6·11 mb). The wind speed (u) near the surface is assumed to be equal to zero, such that $u = 0$ at $z = z_o$.

The roughness length (z_o) can be extremely variable, differing with the changing snow and ice surface conditions during the course of the ablation season. For a given surface, however, z_o can be expected to remain constant especially with wind speeds below those leading to snow drifting. It is usually estimated from detailed measurements of wind profiles over the surface in question. Since this was not possible in the present study, mean values conventionally used for the different states of the glacial material were obtained from the literature as z_o = 0·4 cm. (for snow), 0·25 (for firn) and 0·1 (for ice). In view of uncertainty in the accuracy of these values it is, perhaps, fortunate to note that even a difference of about 0·5 cm in the value of z_o is not very critical for the turbulent heat calculations (Price, 1977 p.30).

Using these boundary conditions, the turbulent heat exchange can be calculated from values of wind speed (U_a), temperature (T_a) and vapour pressure (e_a) at only one height above the glacial material surface. The estimated input values of air temperature and wind speed at each grid point of the altitude matrix (secs. 6.3 and 6.5) are considered to be representative for the 2m level. The mean vapour pressure at each grid point can be determined from the mean relative humidity and saturation vapour pressure at the monthly mean air temperature by using the following expressions:

$$e_a = r.e_s/100$$

where e_a = the vapour pressure (mb)

r = the relative humidity (%)

e_s = the saturation vapour pressure (mb)

The monthly variation of the mean monthly relative humidity at the base meteorological station is shown in Fig. 8.13a.

The relation between the saturation vapour pressure (e_s) and mean air temperature (T in ^{o}C) can be expressed as:

$$e_s = 6{\cdot}11.10^{7{\cdot}5T/(T+237{\cdot}3)}$$

(Haurwitz, 1941 p.9 after Tetens).

The vapour pressure (e_a) can be adjusted for altitude by using the following empirical formula developed in mountainous areas (Kuzmin, 1972 p.122):

$$e_a = e_o.10^{-ma}$$

where e_a = the vapour pressure at altitude a (mb)

e_o = the vapour pressure at sea level (mb)

a = the altitude (km)

m = an empirical constant determined to be 1/6·3 for mountain regions

The density of air will also vary with altitude according to changing pressure conditions. An approximate relation for the variation of air density (ρ) with atmospheric pressure (P) is given by List (1966 p.290):

$$\rho = 3{\cdot}4838.10^{-4}.P/T^*$$

where ρ = the air density (g.cm^{-3})

P = the atmospheric pressure (mb)

T^* = the virtual temperature of air ($^{\circ}$K)

$$T^* = T/(1-0{\cdot}379(e_a/P))$$ (Haurwitz, 1941 p.8).

where T = the air temperature ($^{\circ}$C)

The variation of atmospheric pressure (P) with altitude can be obtained from the relation provided by List (1966, p.268).

$$P_a = P_o\left[1-(0{\cdot}0065a/288)^{5{\cdot}256}\right]$$

where P_a = the mean atmospheric pressure at altitude a (mb)

P_o = the mean atmospheric pressure at sea level (mb)

a = the altitude (m)

The monthly variation of the atmospheric pressure at the base meteorological station is shown in Fig. 8.13b.

Using the above relations, the net hourly turbulent energy exchange due to the sensible and latent heat fluxes can be calculated for each relevant grid intersection of the altitude matrix. The monthly estimates of turbulent heat transfer are made by multiplying the mean hourly values by the duration of the monthly 'melt-period' as indicated by DMHRS and DAYSML. The computational method is programmed in the subroutine TBHEAT and Fig. 8.6c depicts the spatial variation of the turbulent energy balance for July over the Nupur 'test' matrix.

It should be emphasised that the above method of estimating turbulent energy transfer is rather approximate due to uncertainties in the accuracy of the spatial variations of temperature and wind speed estimates (chap. VI), the degree of validity of the various assumptions for the turbulent transfer process, and the problem of heat brought in by horizontal advection.

A potentially serious error is introduced in the estimation of the turbulent energy exchange by neglecting the heat flux due to local and large scale advection processes. The possible importance of heat brought in from areas adjacent to the glaciers has been noted by Ohmura (1971) while, in areas of patchy snowcover, Gray and O'Neill (1974) found that significant amounts of heat may be advected from snow-free areas and used to melt the snow in adjacent snow-covered areas. Under such continuously changing boundary layer characteristics, the above aerodynamic formulae need to be used with caution.

A correct theoretical approach to this problem would be to develop a two- or three-dimensional method of calculating the energy exchange process (e.g. Weisman, 1977). However, with the limited information available, such an approach was not feasible in the present study and the computation of the turbulent transfer of heat has been simplified by assuming a vertical one-dimensional heat flux.

The averaging of the turbulent energy balance over long periods of time is also questionable, in view of the fact that stability conditions may change over short time intervals. However, for the duration of the period for snow and ice melting, it is thought likely that similar conditions prevail for much longer periods.

The application of the above method for the calculation of sensible and latent heat fluxes on Baegisárjökull, using the July 1968 meteorological values, yielded comparable daily values of the turbulent energy exchange to those estimated by Bjornsson (1972 and personal communication).

TABLE 8.1

Relative Importance of Heat Sources for Glacier Ablation

(from some recently reported studies) cf. Paterson (1969, Table 4.2)

Glacier	Latitude	Altitude (m)	Surface	Dates	% contribution of			Reference
					Rad.	Conv.	Cond.	
McCall Glacier, Alaska	69°18'N	1730	snow/ice	13 June-28 Aug 1970	74	26	-	Wendler and Weller (1974)
Boas Glacier, Baffin Island	67°35'N	1140	snow	9 June-7 Aug 1970	60	18	8[1]	Jacobs et al. (1973)
Baegisárjökull, North Iceland	65°36'N	1100	snow	July, Aug 1967	51	33	16	Björnsson (1972)
Storbreen, Norway	61°35'N	1600	snow/ice	6 July-8 Sept 1955	54	32	14	Liestøl (1967)
Worthington Glacier, South Eastern Alaska	61°10'N	850	ice	16 July-1 Aug 1967	51	29	20	Streten and Wendler (1968)
Capps Glacier, Alaska	61°N	1743	ice	30 June-4 Aug 1979	81	12	7	Aufdemberge (1974)
Omnsbreen, Norway	60°39'N	1540	snow/ice	3 June-8 Sept 1968	50	34	16	Messel (1971)
				3 June-8 Sept 1969	55	31	14	
Lemon Creek Glacier, Alaska	58°23'N	1200	firn	5 Aug-20 Aug 1968	49	43	8	Wendler and Streten (1969)
Peyto Glacier, Canada	51°39'N	2510	snow/ice	1 July-15 July 1970	44	48	8	Föhn (1973)
Ampere Glacier, Kerguelen Islands	49°S	200	ice	8 Jan-2 Mar 1976	58	26	16	Poggi (1977)
Ivory Glacier, New Zealand	43°08'S	1457	ice	5 Jan-14 Feb 1972	53	33	11[2]	Anderton and
				28 Jan-14 Feb 1973	51	25	22[3]	Chinn (1978)

1 : Refreezing of meltwater supplied 14% of heat

2 : Refreezing of rain supplied 3% of heat

3 : Refreezing of rain supplied 2% of heat

TABLE 8.2

Illustration of the Diurnal Variation in the Individual Components of the SWRAD Subroutine

Time Increment (60 minutes)	H	θ	α	D	L	R_C	$R_C(1-\alpha)$	TR_C	$TR_C(1-\alpha)$
1	-10.278	2.235	0.918	0.002	2.065	0.930	0.076	0.930	0.076
2	- 9.230	5.918	0.809	0.383	5.270	2.544	0.485	3.474	0.561
3	- 8.183	10.638	0.706	2.961	8.297	5.066	1.491	8.540	2.052
4	- 7.136	16.138	0.622	8.060	10.691	8.438	3.188	16.977	5.240
5	- 6.089	22.133	0.562	15.695	12.106	12.511	5.482	29.488	10.722
6	- 5.042	28.315	0.522	24.409	12.916	16.796	8.034	46.284	18.757
7	- 3.994	34.330	0.496	33.043	13.347	20.876	10.511	67.160	29.268
8	- 2.947	39.762	0.481	40.643	13.553	24.388	12.647	91.548	41.915
9	- 1.900	44.112	0.473	46.458	13.636	27.042	14.252	118.590	56.166
10	- 0.853	46.845	0.469	49.948	13.663	28.625	15.203	147.214	71.369
11	0.194	47.533	0.468	50.803	13.667	29.011	15.435	176.226	86.804
12	1.242	46.051	0.470	48.948	13.657	28.172	14.931	204.398	101.736
13	2.289	42.658	0.476	44.547	13.614	28.173	13.727	230.570	115.463
14	3.336	37.844	0.486	37.995	13.495	23.171	11.908	253.741	127.370
15	4.383	32.142	0.504	29.912	13.220	19.410	9.619	273.150	136.990
16	5.430	26.020	0.535	21.127	12.670	15.208	7.079	288.359	144.069
17	6.477	19.869	0.582	12.873	11.563	10.996	4.601	299.355	148.670
18	7.525	14.023	0.650	5.835	9.916	7.088	2.478	306.443	151.147
19	8.572	8.781	0.742	1.624	7.273	4.004	1.032	310.447	151.179
20	9.619	4.418	0.850	0.073	4.043	1.852	0.277	312.299	152.456
21	10.666	1.178	0.955	0.000	1.089	0.490	0.022	312.789	152.478

H = hour angle, θ = solar altitude, α = albedo, D = direct solar radiation, d = diffuse solar radiation, R_C = Global radiation corrected for cloud cover, $R_C(1-\alpha)$ = absorbed global radiation, TR_C = cumulative total of R_C, $TR_C(1-\alpha)$ = cumulative total of $R_C(1-\alpha)$.

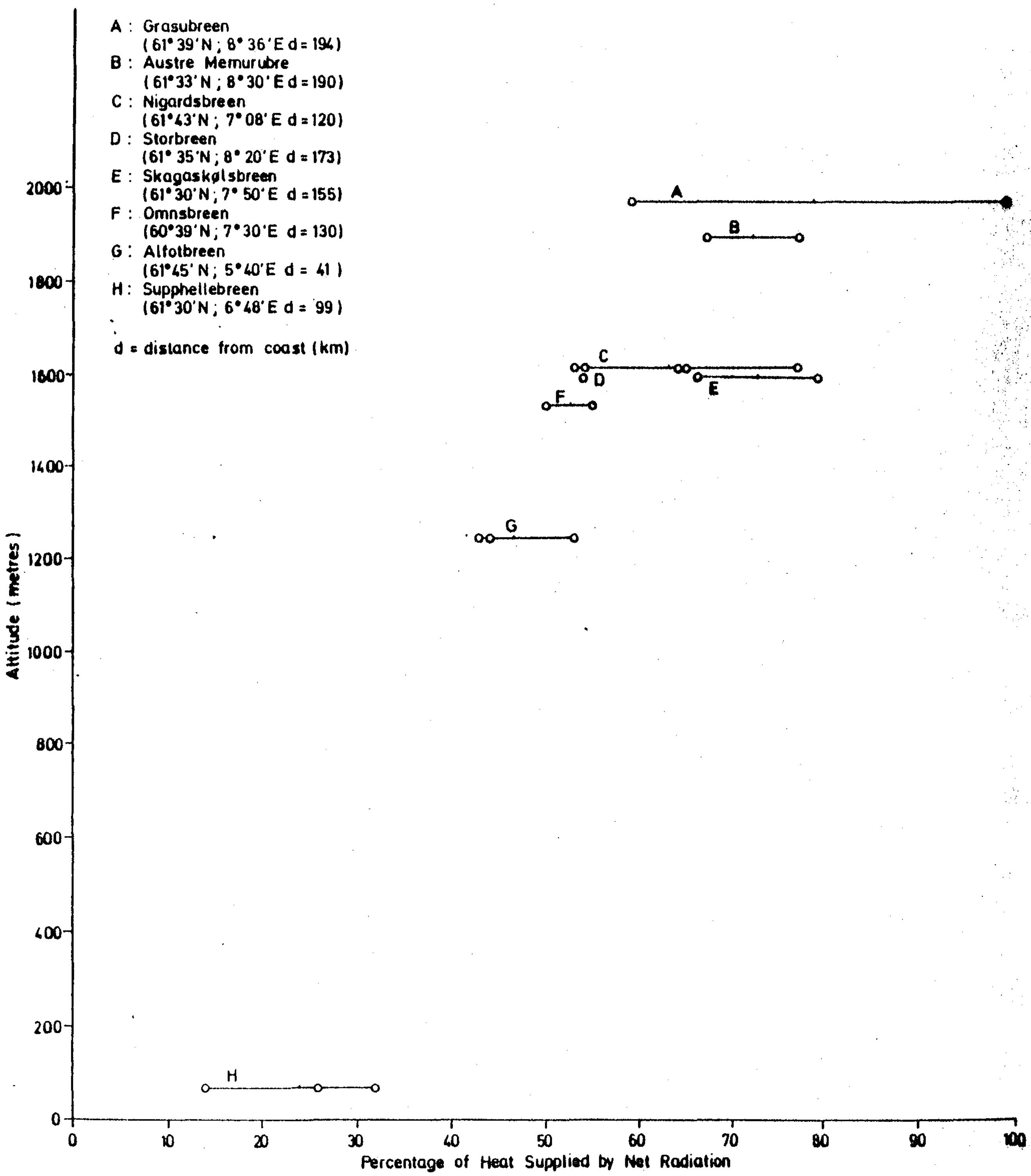


FIGURE 8.1
Effect of Altitude on the Contribution of Net Radiation to the Glacier Heat Balance. (data from Haakensen and Wold (Eds.), 1981 p.45).

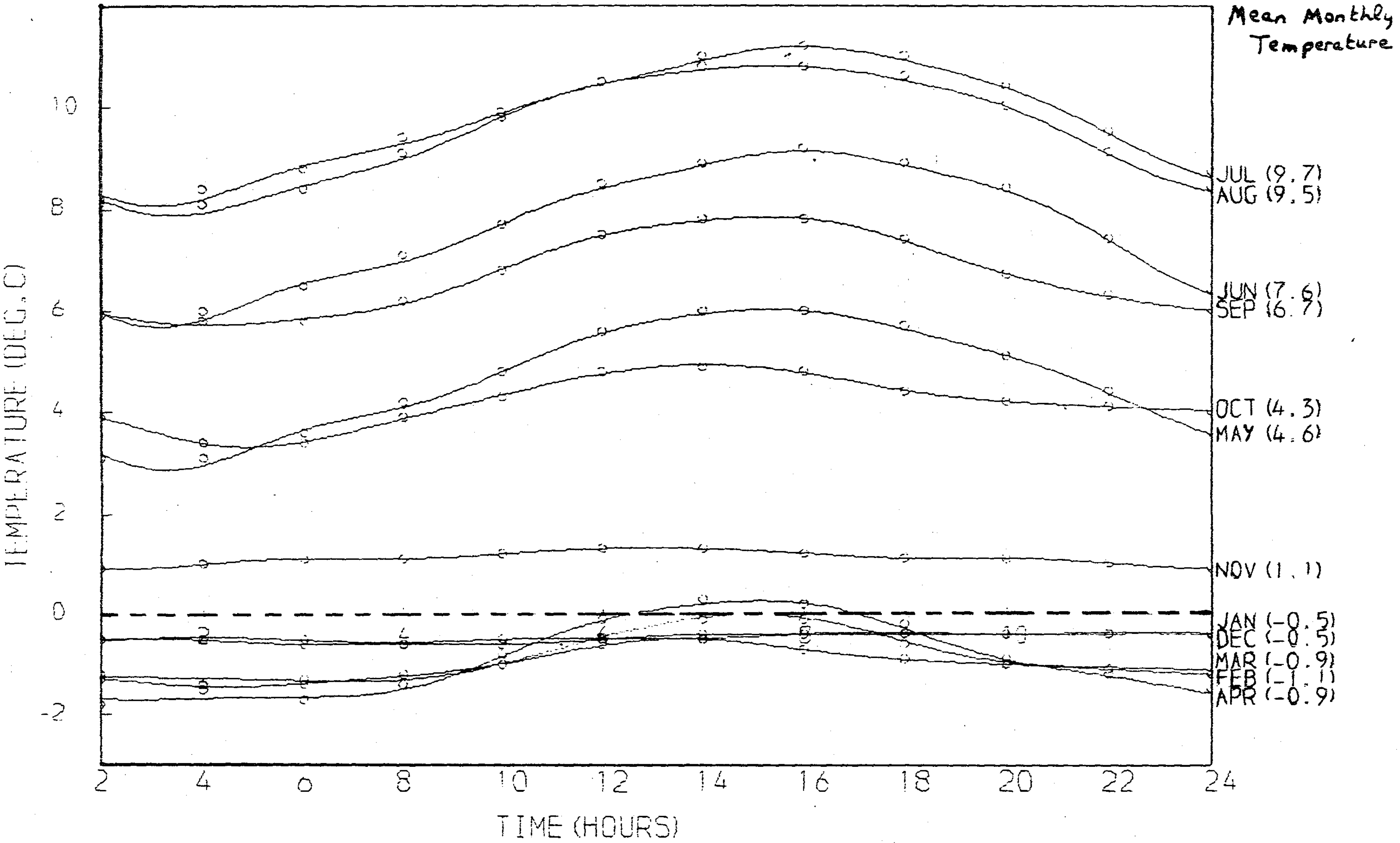


FIGURE 8.2
Diurnal Variation of Temperature at Bolungarvik (1946-1953)
Source: Veðrattan. Icelandic Meteorological Office.

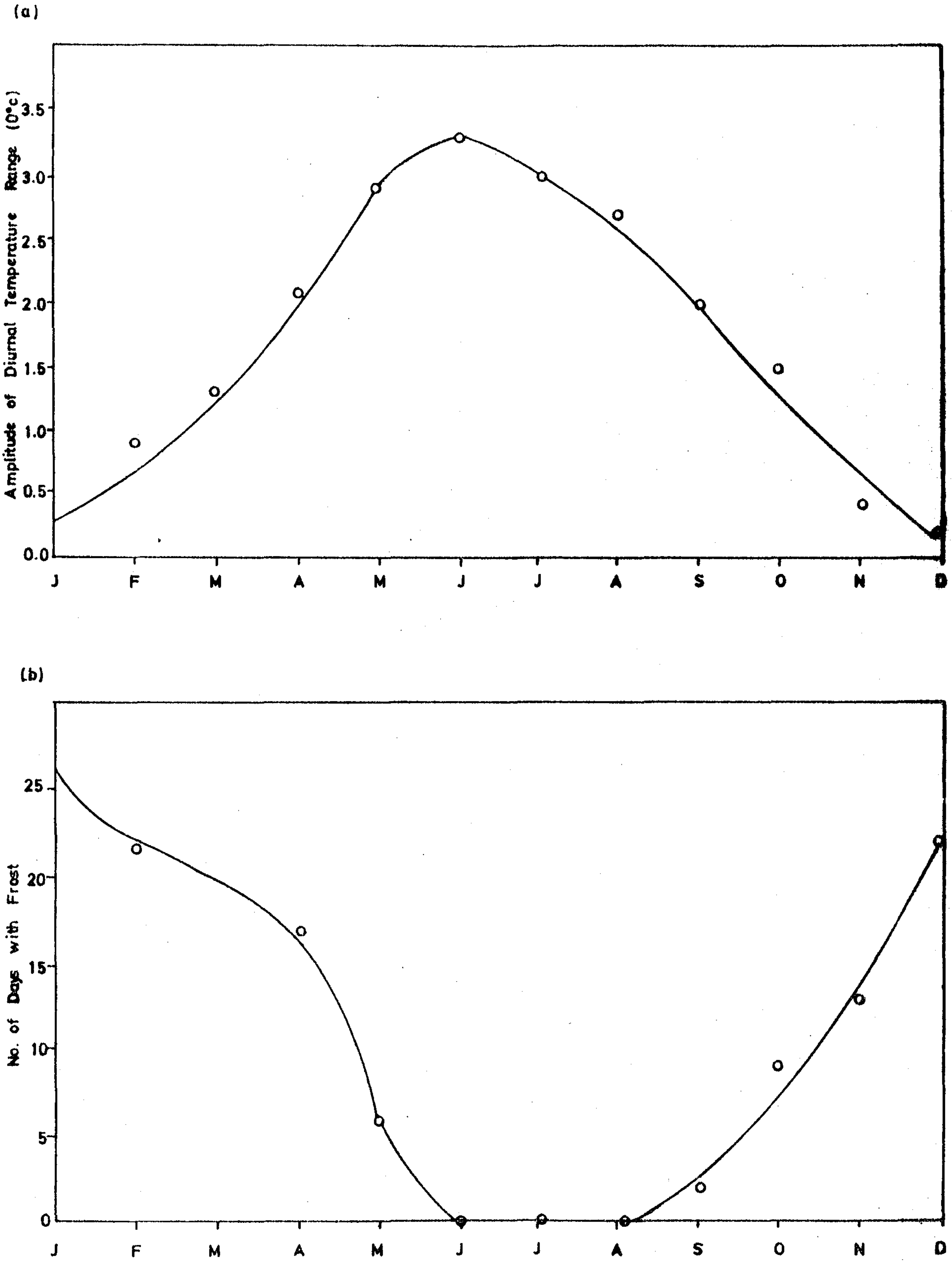


FIGURE 8.3

(a) Monthly Variation in the Amplitude of Diurnal Variation Range (°C) at Bolungarvik (1946-1953).

(b) Number of Days with Frost at Sudureyri (1951-1960)

Source: Icelandic Meteorological Office.

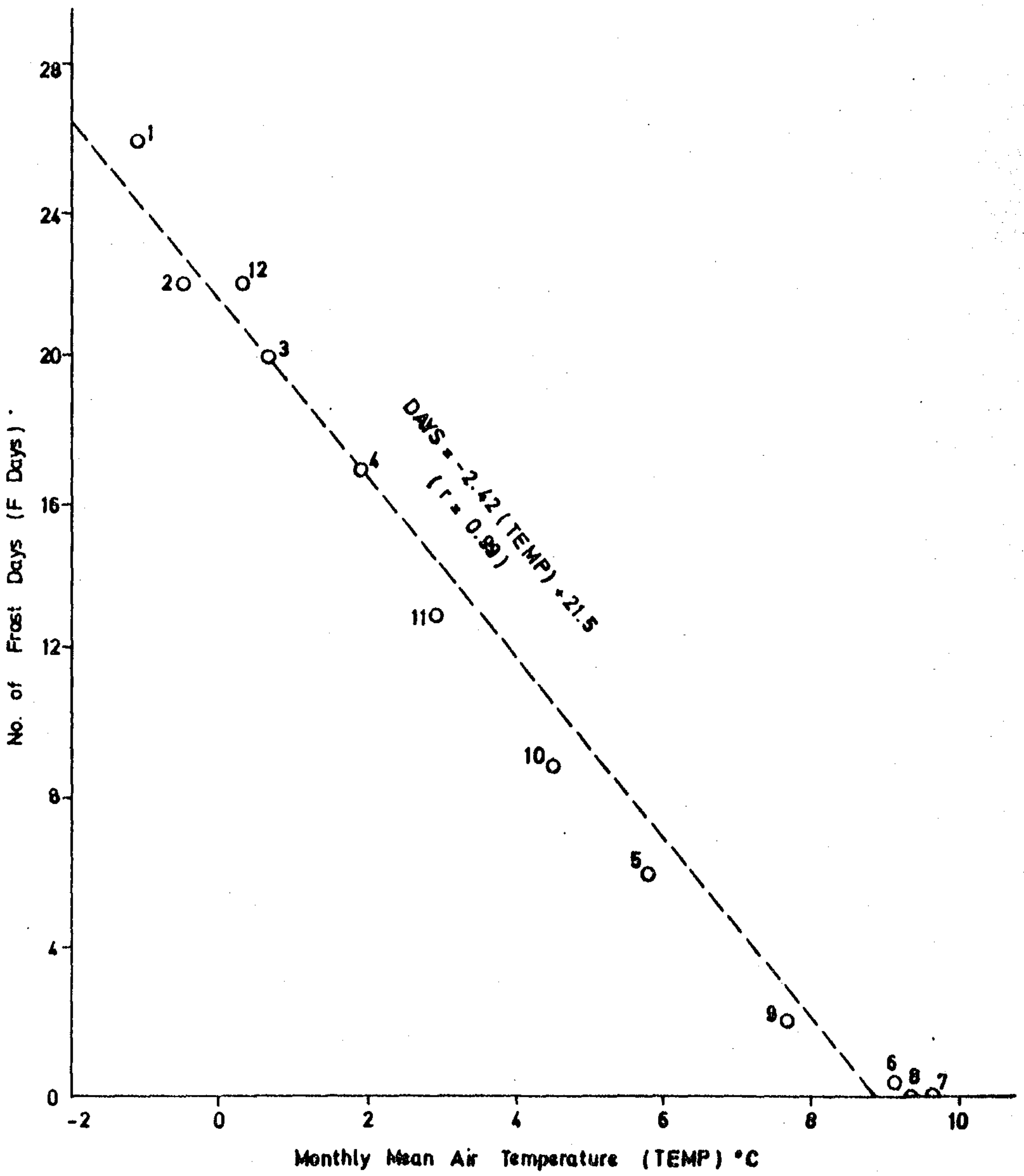
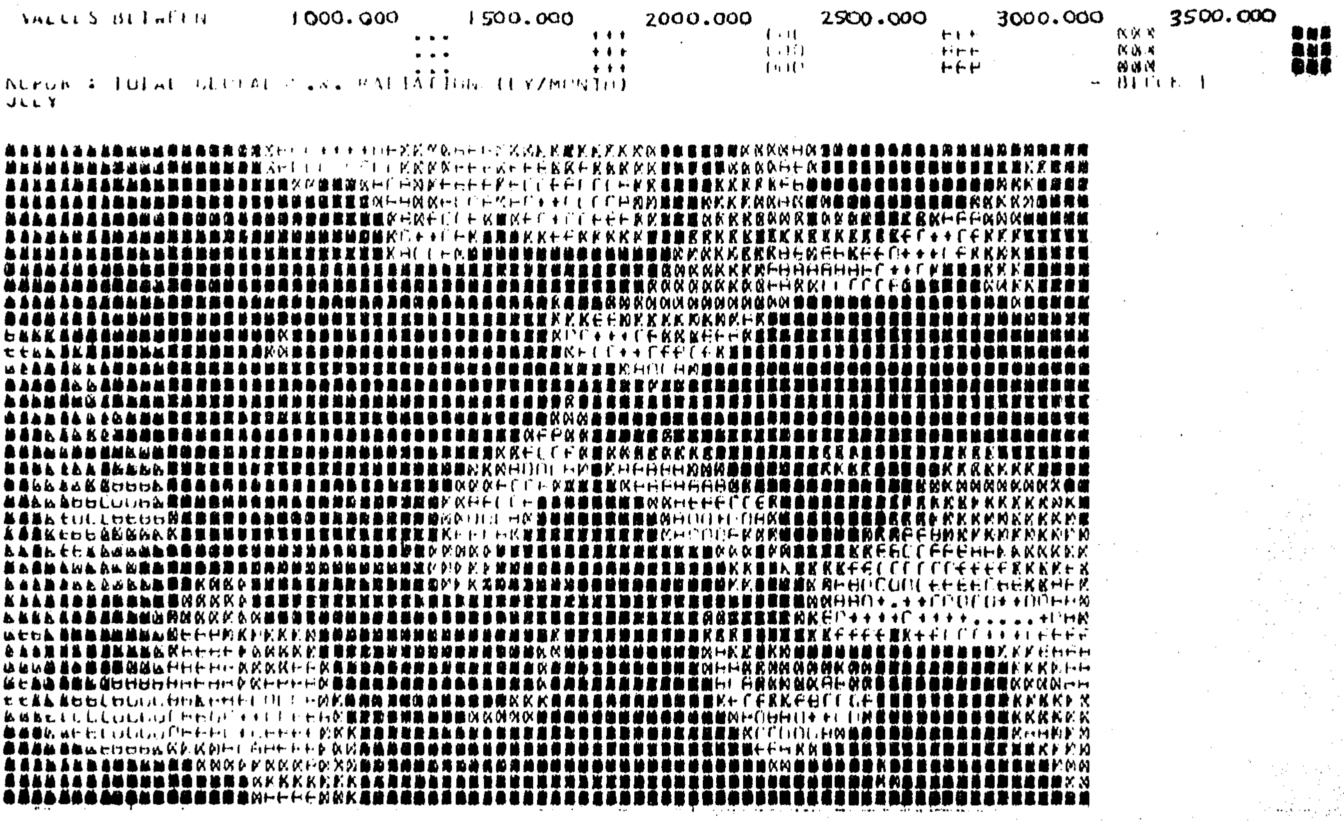


FIGURE 8.4
Number of Frost Days as a Function of the Monthly Mean Air Temperature at Sudureyri (mean of 1951-1960).
Source: Icelandic Meteorological Office.

APRIL
POTENTIAL MELT AREAS (TEMPAL > 0.0)

FIGURE 8.5
Distribution of Potential Melt Areas Over the Central Nupur Matrix (for April)

(a)



(b)

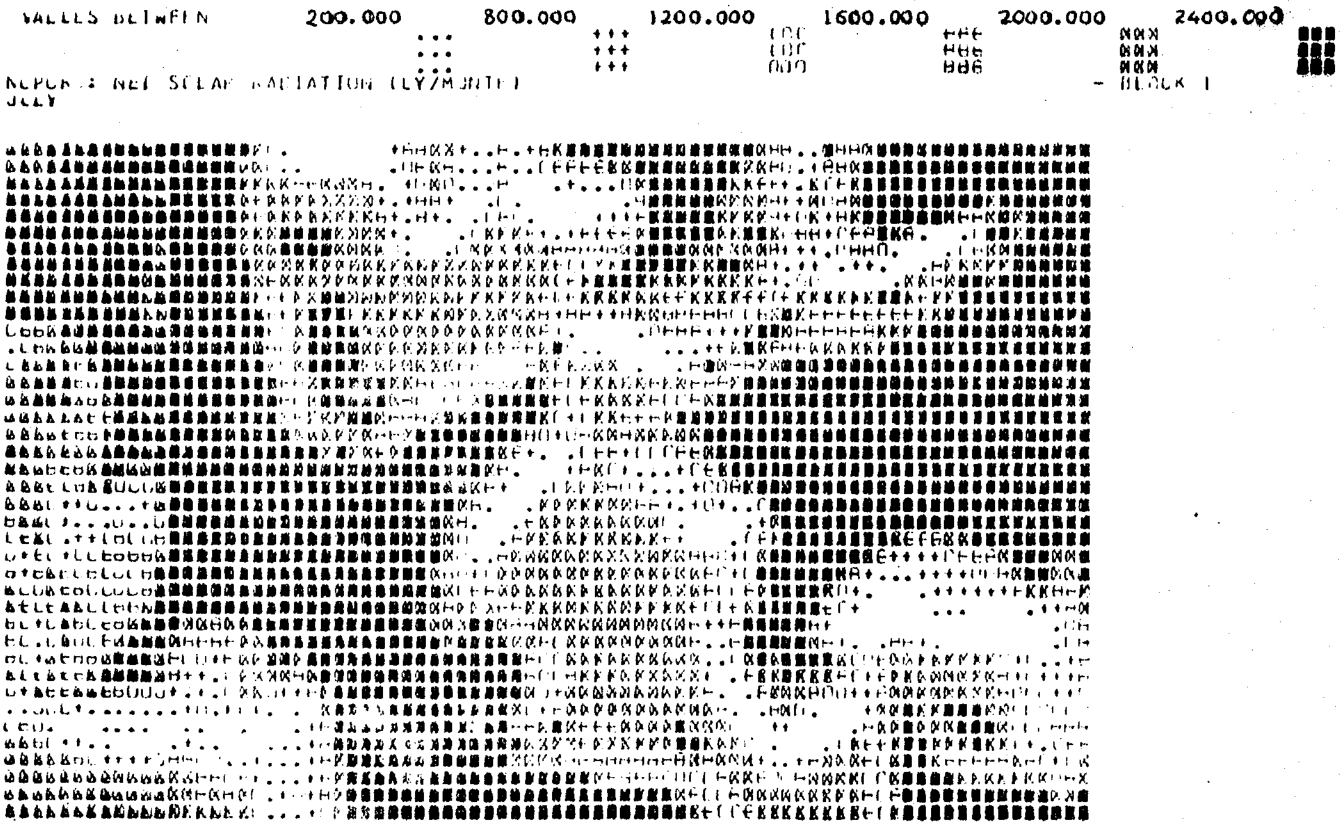


FIGURE 8.6
Variation of Energy Balance Components Over the Central Nupur
Matrix (for July)

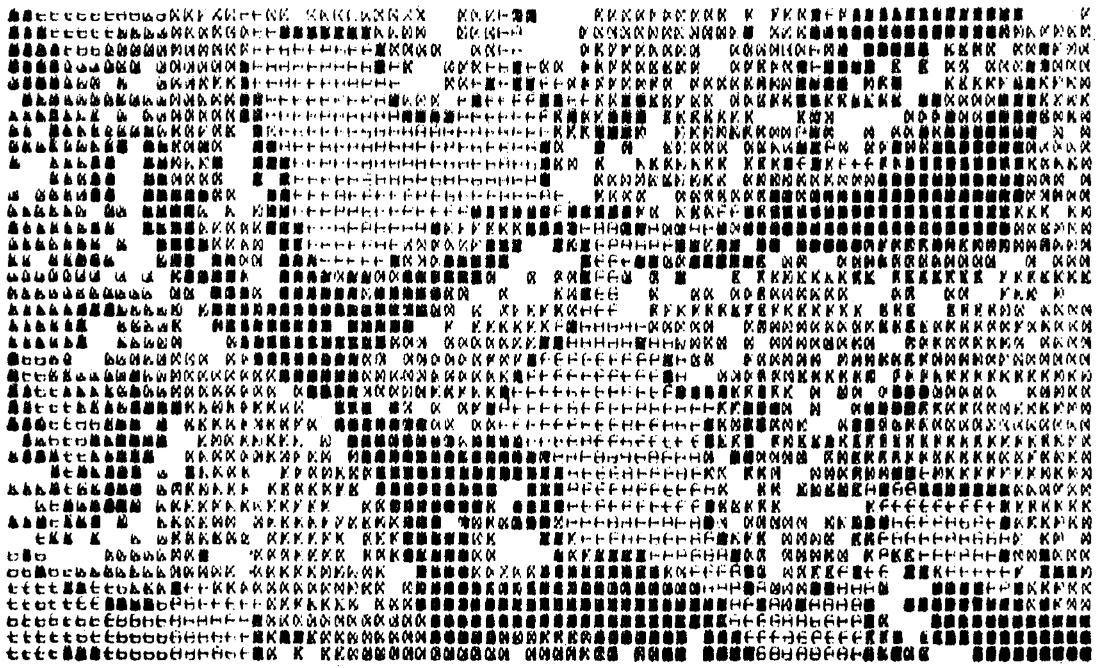
(a) Global Shortwave Radiation

(b) Shortwave + Longwave Radiation

(continued on the next page)

VALUES BETWEEN 3200.000 3600.000 4000.000 4400.000 4800.000 5000.000

(c) NEPER : TOTAL SENSIBLE HEAT (LY/MONTH) - BLOCK 1
JULY



VALUES BETWEEN 4500.000 5250.000 6000.000 6750.000 7500.000 8250.000

(d) NEPER : NET ENERGY BALANCE (LY/MONTH) - BLOCK 1
JULY

FIGURE 8.6 (cont.)
(c) Turbulent Heat
(d) Net Energy Balance

VALUES BETWEEN 0.000 4.000 6.000 8.000 10.000

PERCENTAGE REDUCTION OF S.W. RADIATION
(DUE TO SHADING BY SURROUNDING TOPOGRAPHY)

\- BLOCK 1

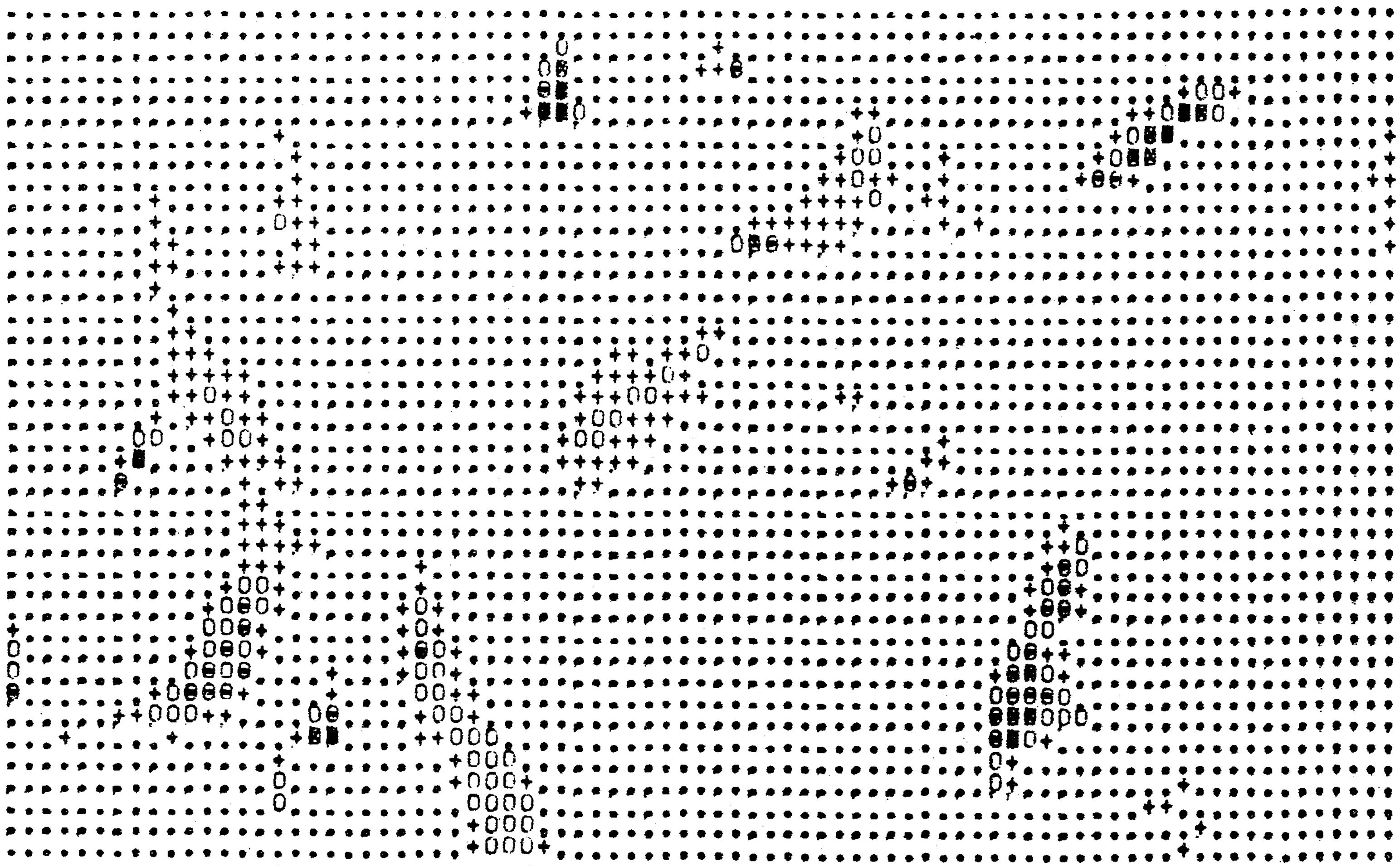


FIGURE 8.7
Percentage Reduction of Shortwave Radiation Receipt (for July) as a Result of Topographic Shading Over the Central Nupur Matrix.

VALUES BETWEEN 2500.000 3000.000 3500.000 4000.000 4500.000 5000.000

(a) S.W. RADIATION (USING SWRAD) - BLOCK 1
(WITHOUT SHADOW ROUTINE)

(b) S.W. RADIATION (USING SWRAD) - BLOCK 1
(WITH SHADOW ROUTINE)

FIGURE 8.8
Shortwave Radiation Receipt (July) Over the Central Nupur Matrix.
(a) Without Shadow Routine (b) With Shadow Routine.

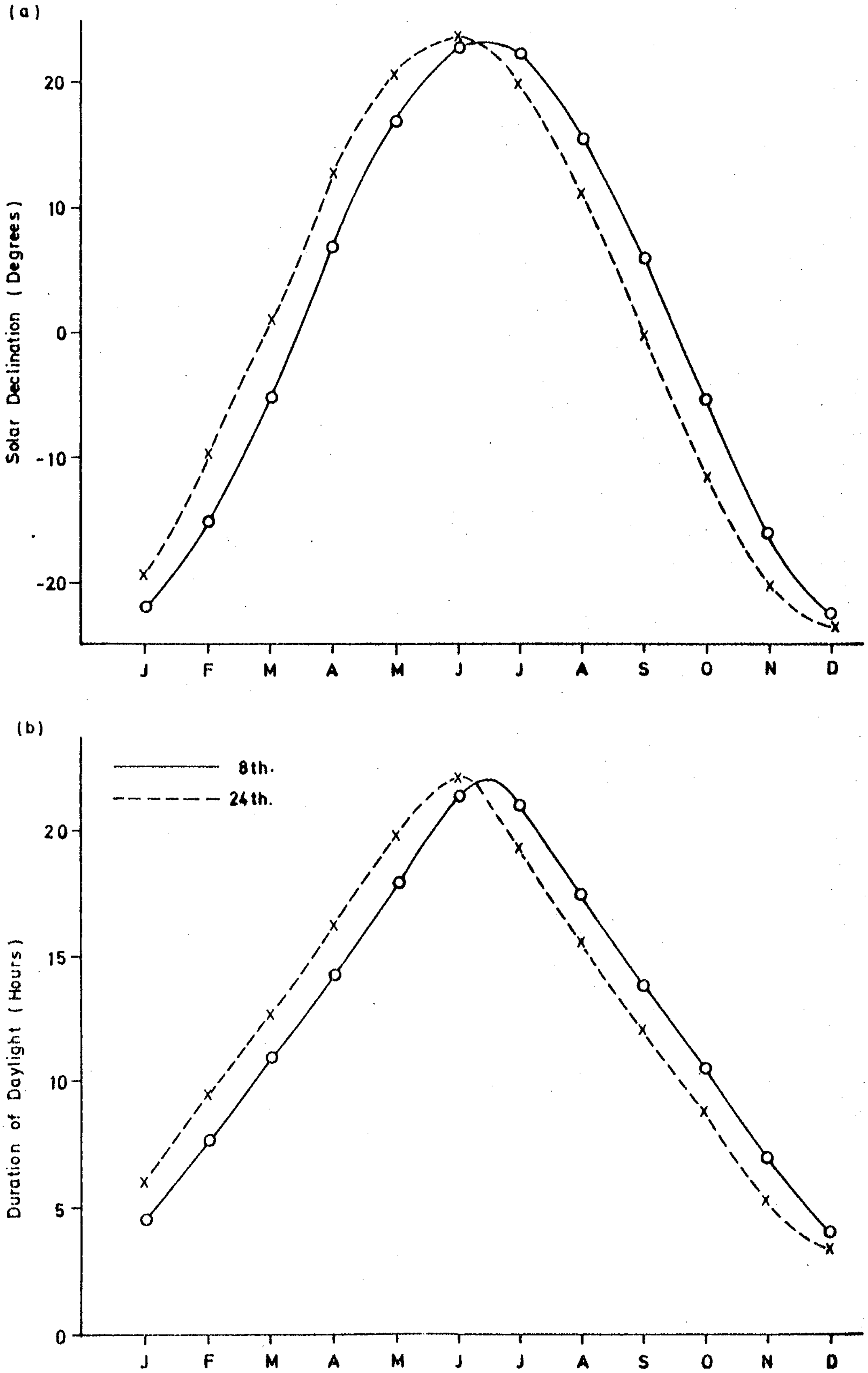


FIGURE 8.9
(a) Annual Variation of Solar Declination (Degrees)
(b) Duration of Daylight at 65^{o}N (Hours)

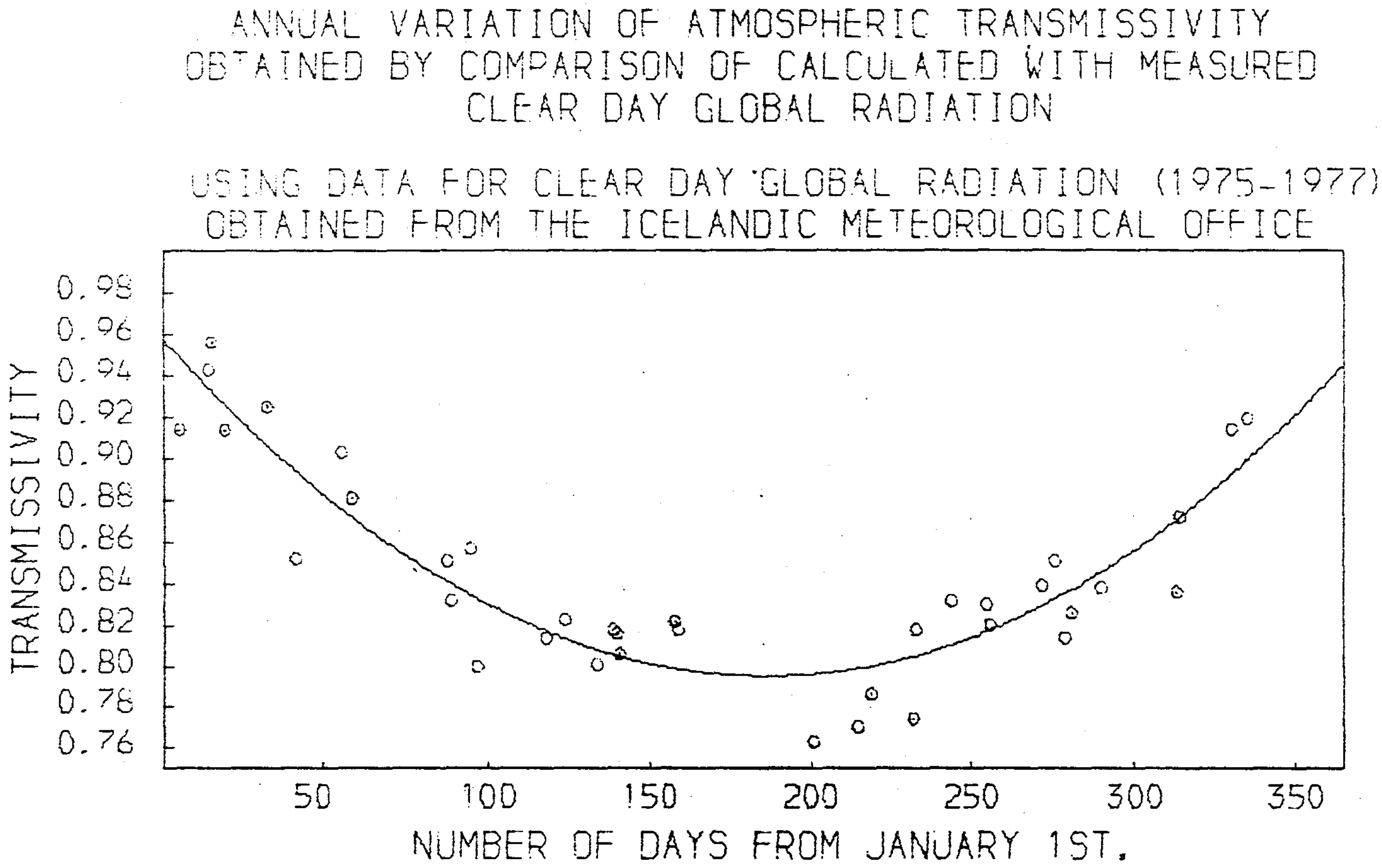


FIGURE 8.10

Variation of Atmospheric Transmissivity (Iceland)

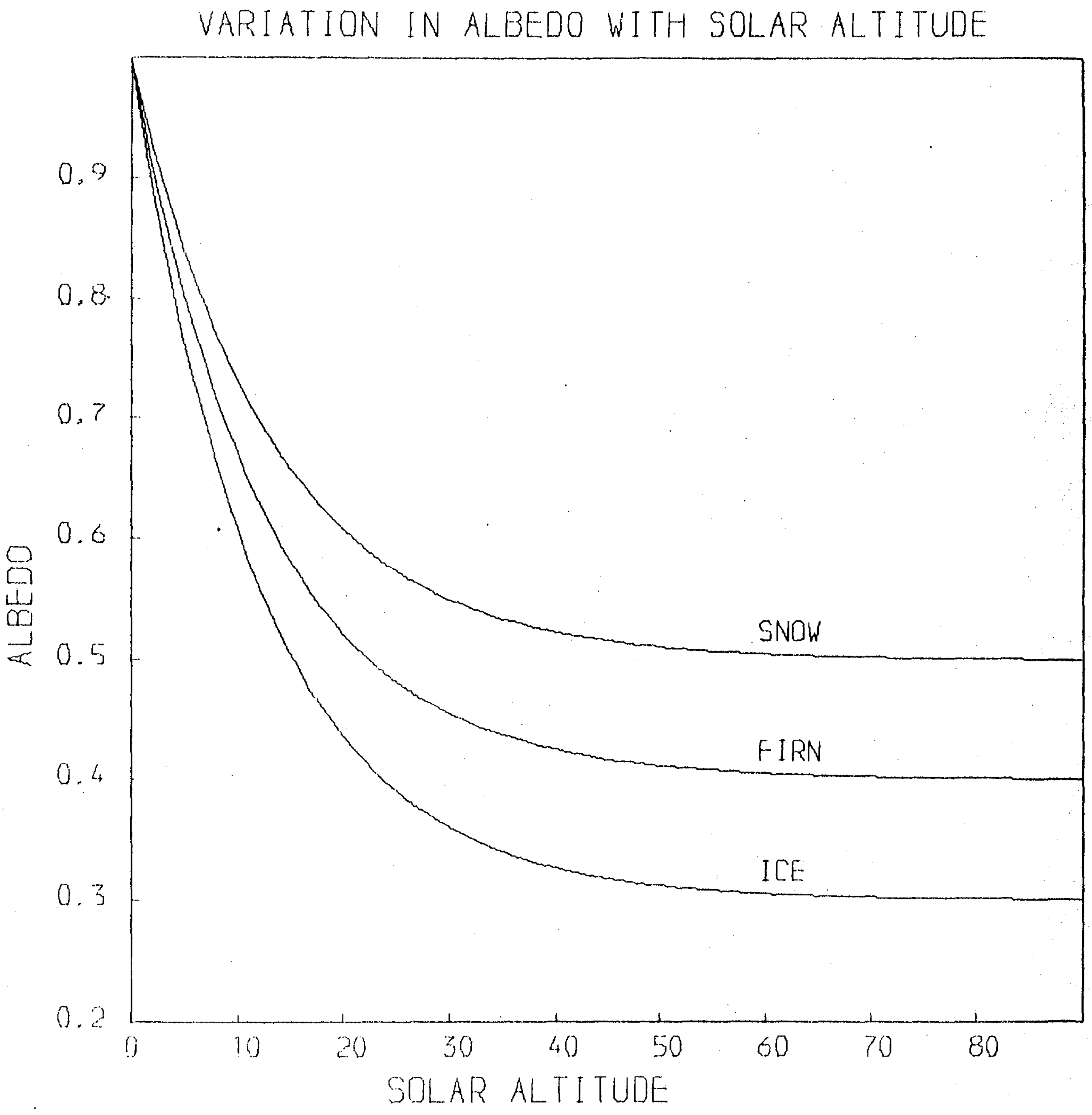


FIGURE 8.11
Variation of Glacial Material Albedo With Solar Altitude

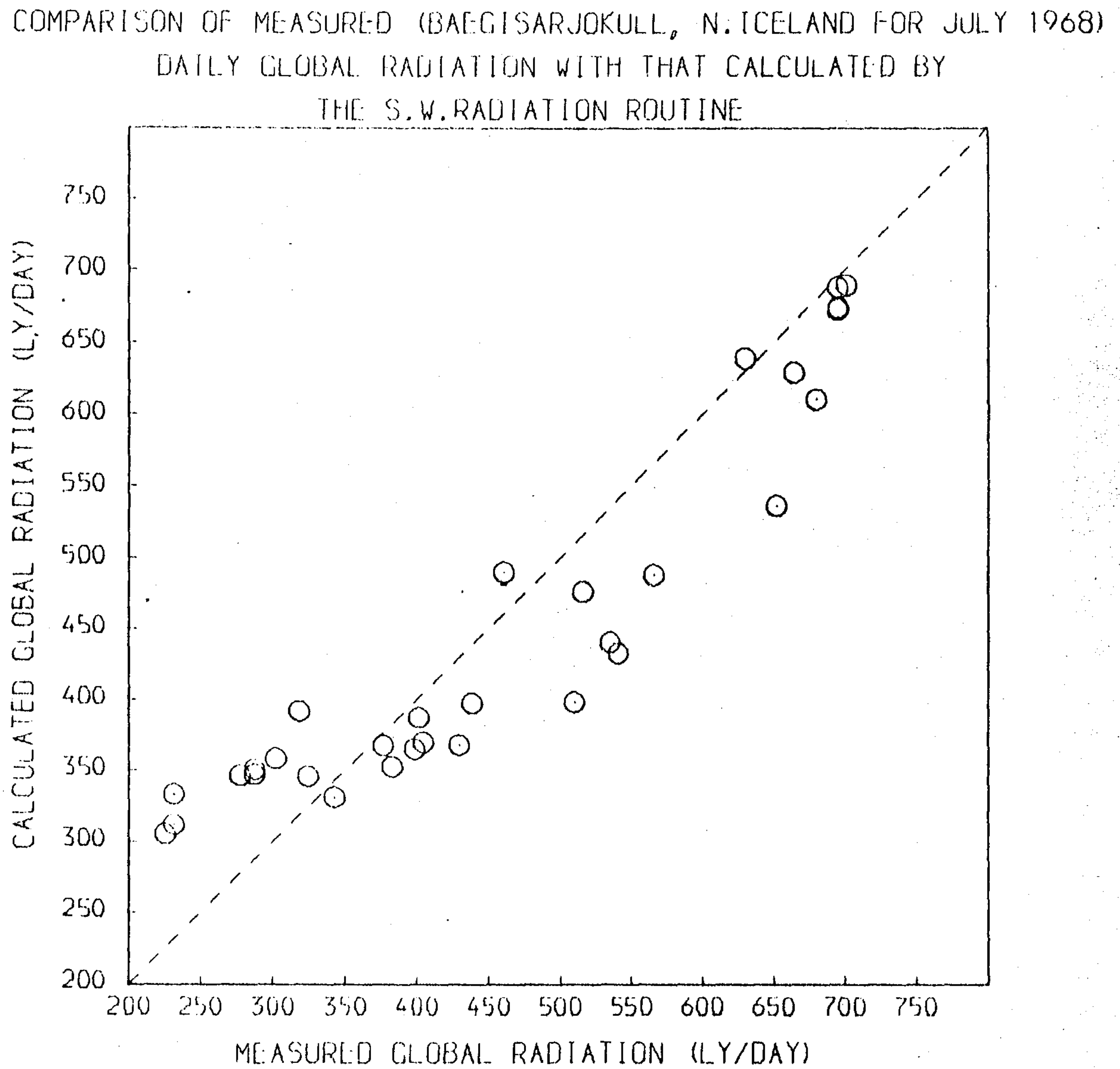


FIGURE 8.12
Test of Accuracy for the Calculation of Global Solar Radiation

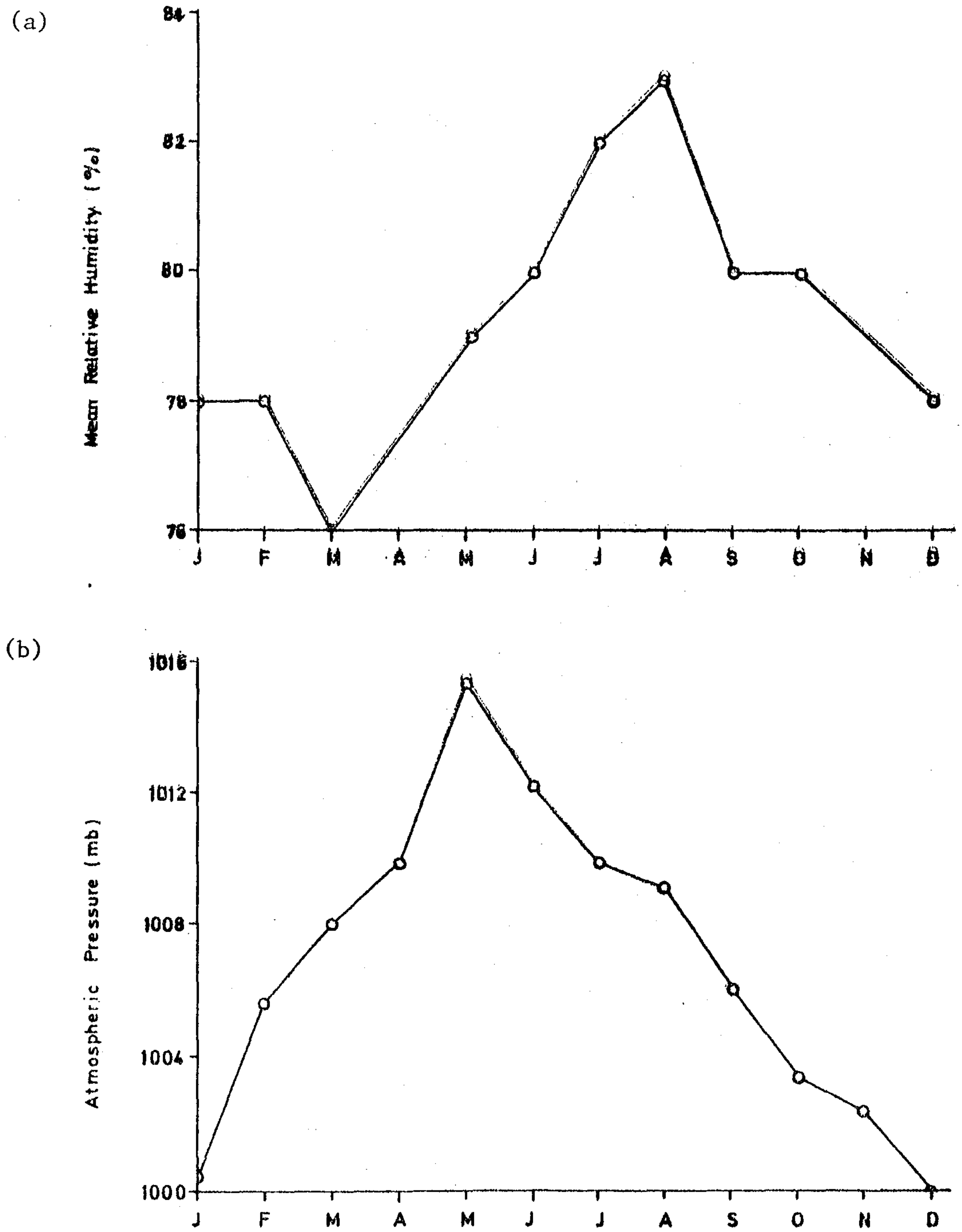


FIGURE 8.13

(a) Variation of Mean Monthly Relative Humidity at Galtarviti (1958-1967).

(b) Variation of Monthly Atmospheric Pressure at Galtarviti (1931-1960).

Source: Veðrattan: Icelandic Meteorological Office.

CHAPTER IX

MASS BALANCE SIMULATION

Following the modelling of accumulation and ablation processes, (described in the preceding chapters), the net mass balance of glacial material present at each grid point of the altitude matrix, at the end of any month's simulation, can be determined. This is specified in the mass balance model by recording both the total depth and water equivalent of glacial material, (given by the matrices TDEPTH and TWEQ respectively), and that part present in the form of snow, firn and ice.

The application and validity of the mass balance model is tested in this chapter by simulating mass balance variations in the two Icelandic sample areas (sec. 4.7).

9.1 Nupur Mass Balance (1931-60)

The mass balance model was first applied to the 80x80 Nupur matrix (Fig.5.16). Using the 1931-60 mean climatic input for the Nupur area (Fig. 6.1), the monthly simulation of mass balance is depicted by means of line printer maps in Figs. 9.1 to 9.13. The spatial variations in snow cover can be related to the topographic variations in the Nupur area by comparing the mass balance results with Fig 5.16 or by utilizing the transparent overlay for the Nupur matrix (provided in the pocket bound in the back of this thesis).

The simulation was started in October, at the beginning of the accumulation season. The snow cover is seen to extend from the highland areas in October (Fig. 9.1) to the whole matrix by December (Fig. 9.3).

Snow drifting from the plateau areas (once minor roughness has been covered), and avalanching off the steep slopes becomes evident in January (Fig. 9.4). The transported snow is deposited in suitable concavities at the foot of avalanching slopes. This trend of increasing snow cover in favoured localities continues through February (Fig. 9.5), and March (Fig. 9.6), sees the maximum snow accumulation in the Nupur area.

The effect of the ablation routines becomes apparent in April (Fig. 9.7) with the melting of snow in the low-lying valley areas, which experience positive energy balance conditions. Melting of snow extends to higher altitudes as the ablation season advances (Figs. 9.8 to 9.10). Towards the end of the ablation season (Figs. 9.11 to 9.12) pockets of snow remain only in localised concavities, sheltered from the effects of solar radiation and wind erosion. This annual cycle of mass balance variations recommences at the start of another accumulation season in October (Fig. 9.13).

While the above pattern of mass balance changes in the Nupur area corresponds in general fashion with observations of snow cover distribution (during the summer months at least), some form of test of model validity is desirable.

9.2 Model Validation

The problems of obtaining suitable field data for testing the validity of the mass balance model were discussed in sec. 4.6. It was pointed out that only limited testing of model validity was possible by comparing the simulated mass balance results with observations of snow cover and local glacier distribution in the sample Icelandic study areas.

9.2.1 Nupur Snow Cover Distribution

(i) Comparison of the Nupur snow cover distribution as mapped from aerial photographs (July 1959) with that simulated by the mass balance model using the climatic input for the balance year preceding July 1959.

An attempt was made to map the major (>100 m.) snowpatches in the Nupur area from aerial photographs taken at a scale of about 1:36,000 (Fig. 4.10). Using the appropriate climatic input for the mass balance year preceding July 1959 (provided by the Icelandic Meteorological Office) the simulated snow cover at the end of July is shown in Fig. 9.14. The distribution of the observed snowpatches is also shown in the same figure. An approximate validation test of the mass balance model is provided by comparing the simulation results with the distribution of snowpatches mapped from the aerial photographs.

(ii) Comparison of the Nupur snow cover distribution determined by field mapping (August 1978) with that simulated by the mass balance model using the climatic input for the balance year preceding August 1978.

Due to difficulties in the accurate determination of the position and size of snowpatches from aerial photographs, a better indication of the Nupur snow cover towards the end of the ablation season (August 1978) was obtained by mapping the distribution of snowpatches in the field itself (Fig. 3.6). As in the above case, sec. 9.2.1(i), the distribution of the major snowpatches can be compared with the simulated mass balance at the end of August (Fig. 9.15), determined by using the appropriate climatic input for the mass balance year preceding August 1978 (provided by the Icelandic Meteorological Office).

These tests of model validity reveal a fair degree of agreement between the simulated mass balance results and the observed snowpatch distribution. In particular, the August 1978 pattern of major snowpatches, determined by field mapping, matches closely with the simulated distribution of grid squares with positive mass balance (Fig. 9.15). Only two out of the thirteen observed snowpatches are not represented by sites of simulated positive mass balance in the corresponding or adjacent grid squares. In both cases the snowpatches are found on rather steep slopes ($36{\cdot}8^{\circ}$ and $44{\cdot}3^{\circ}$) without any marked concavity being present; in fact both sites have slightly positive values ($1{\cdot}1$ and $2{\cdot}9^{\circ}/100$ m.) for profile convexity. Furthermore only three out of the fourteen predicted sites of positive mass balance are not represented by field observations. In all three cases, however, the predicted accumulation is relatively minor (less than 50·0 cm. water equivalent).

The observed and predicted snowpatch locations in the Nupur area are typically found to lie at altitudes of between 460 - 530 m., on slope gradients of between 20 and 38 degrees, with slope aspects of NE-N-NW. They are situated in concavities with the values of profile convexity being between -2 and -15 degrees per 100 m., while those of plan convexity being between -20 and -80 degrees per 100 m.

The correspondence between simulated and observed snow cover is less satisfactory in the case of the June 1959 results (Fig. 9.14). This is due, at least partly, to the problems of estimating the importance and position of snowpatches from aerial photographs. In the majority (75%) of the cases, however, the observed snowpatches have been predicted by positive mass balance in the corresponding or adjacent grid square position; while in 74% of the cases, simulated sites of positive mass balance correspond to the observed distribution of snowpatches.

The above tests of model validity also provided some indication of model sensitivity by observing the effect of changes in model components on the simulated mass balance results. For example, the inclusion of the shadowing subroutine (SHADE) for direct shortwave radiation computations, was found to have little effect on the predicted snow cover distribution (sec. 8.3.1). A more important consideration was found to be the selection of the snow thickness threshold, to cover minor roughness on the plateau areas, before substantial snow drifting could take place. The original estimate of 50.0 cm was found to be too high. The closest comparison of the simulated snow cover distribution, at the end of the ablation season, with the observed pattern of snowpatches was obtained by using a snow drift threshold thickness of 30.0 cm. Furthermore, setting both the percentage of snow lost by being blown off the plateau edge during the drifting process (sec. 7.2.2) and that assumed to sublimate during the avalanching process (sec. 7.3.2) to 20% of the total snow being transported in each case, was found to give satisfactory results in the above simulations.

9.2.2 Thvera Local Glacier Distribution

Further validation of the mass balance model was attempted by simulating glacier mass balance on the Thvera matrix (Fig. 4.9).

The present-day distribution of local glaciers in the Thvera area (shown in Fig. 9.16) was determined from aerial photographs and field observations (Nigel Griffey, personal communication). These glaciers are believed to have originated or reached their maximum extent during the 'Little Ice Age'.

Initially the mass balance distribution in the Thvera area was simulated by using the mean climatic input for the period 1931-60, (obtained for the base climatological stations. Skriðuland and Hólar i Hjaltadal), and appropriate combinations of topographic parameters (derived from the Thvera altitude matrix) to delimit significant topographic features, such as plateau areas and concavities (sec. 5.6). Following this, attempts were made to simulate the generation sites of the local glaciers in the Thvera area by using mean monthly temperatures reduced from the 1931-60 averages in an attempt to represent temperature conditions during the Little Ice Age (sec. 2.3.5).

It was found that incipient cirque glaciers simulated with a 2.0^{o}C drop in mean air temperature (Fig. 9.17) best represented the expected generation-sites of the Thvera local glaciers. The glacier simulations were carried out in 3-year periods and the mass balance results presented in Fig. 9.17 portray the grid squares with greater than 10 m. of ice accumulation after 36 years of simulation. A transparent overlay for the Thvera area is provided (in the pocket bound in the back of this thesis) to relate these glacier mass balance variations to the local topography in the Thvera area.

The greatest concentration of simulated ice is found in the numerous cirques enclosed by steep avalanching slopes. To a lesser extent some accumulation of glacial material is evident in the flat-lying valley floors and at the foot of avalanching slopes.

A comparison of the simulated results (Fig. 9.17) with the observed local glacier distribution in Thvera (using the transparent overlay for Thvera) reveals that, with the exception of the south-western part of the Thvera matrix, the predicted sites of glacier ice development generally correspond with the expected generation-sites of the observed glaciers. The north-facing cirque in the extreme south-western corner of the matrix, adjoining a plateau remnant, which was predicted to contain glacier ice, however, remains rather surprisingly unglacierised. In contrast, the expected generation site of the adjacent (eastwards) glacierised cirque is not evident in the simulation results. In the latter case, the discrepancy appears to result from the lack of a well-defined source area (in terms of avalanching slopes for example) for substantial snow accumulation to be simulated. This could be a consequence of difficulty in determining the topography beneath the ice cover (sec. 5.2.1), due in this case, to the rather complicated pattern of ice flow over the ridge to the south-east.

In contrast to the Nupur mass balance simulations (secs. 9.1 and 9.2.1), the greater relief amplitude present in the Thvera matrix (1105 m. compared with 766 m. in the Nupur matrix), suggested a need to include the shading effect in modifying solar radiation. This is borne out by Fig. 9.18 which depicts the reduction (%) in direct solar radiation received in July at each grid point of the Thvera matrix. It was found that topographic shading resulted in a reduction of about 30% on some sheltered north-facing cirque slopes compared with less than 14% in the Central Nupur matrix (Fig. 8.7). In view of this, the SHADE subroutine was incorporated in the Thvera glacier balance simulations. To offset the increase in computer time, the radiation computations were conducted for one representative day in each month (using the sun's declination, duration of daylight and transmissivity values for the 15th of each month).

The primary aim of the mass balance model is to determine the sites of local glacier generation. While the latter are indicated by areas of positive net mass balance, their determination can be further enhanced by identifying those grid squares with sufficient ice accumulation for glacier flow to occur. The detailed modelling of glacier flow is beyond the scope of this study; however, an attempt was made to incorporate a simple routine into the mass balance model for the determination of ice flow, based on its variation with ice thickness and slope gradient.

Assuming perfect plasticity, the critical thickness at which the ice at any grid square will begin to flow can be estimated from the well known equation:

$$\tau_o = \rho \; g \; h \; \sin\alpha$$

where τ_o = the basal shear stress (taken as 1 bar)

ρ = the density of ice

g = the gravitational acceleration

h = the ice thickness

α = the slope of the upper ice surface

In the present study the mass balance cover over the altitude matrix is defined in terms of 100 m. grid squares. If it is assumed that the glacial material is evenly distributed within each grid square and that the ice thickness does not differ greatly in adjacent squares, then the ice surface slope can be regarded as being the same as that calculated for the ground surface. Furthermore the direction of ice movement can be taken to correspond with the aspect of the maximum gradient for any particular grid square.

The critical thickness (CRIT(I,J)) for ice flow at each grid square can be calculated from the above equation (given by $h = \tau_o/\rho g \sin\alpha$). These critical ice thickness determinations need only be carried out once (for each study area to which the model is applied) at the start of the simulation program, prior to the mass balance calculations.

Having determined the critical thickness for ice flow at each grid square, the question of the rate at which glacial material is transferred from grid squares where the ice has achieved its critical flow thickness to other adjacent grid squares, needs to be resolved. For this a consideration of ice flow velocities in cirque glaciers is necessary.

Now basal sliding has been shown to be the predominant type of ice movement in temperate cirque glaciers (e.g. McCall, 1952 p.62). From published measurements of sliding velocity, Paterson (1970) obtained the following relationship between glacier thickness (h in m.) and the annual mean velocity of basal sliding (U_b in ma^{-1}):

$$U_b = 0{\cdot}11\ h^{0{\cdot}95}$$

This empirically-derived relationship was used to provide an indication of ice flow rates expected with varying ice thicknesses. Given a constant rate of annual net mass balance increase, the initial ice mass in any grid square which has achieved the critical flow thickness, can be considered to be transferred to an adjacent grid square (in the direction given by the aspect of maximum gradient) after it has moved a distance of 100-141 m.

Now the critical flow thickness at the predicted sites of glacier generation in the Thvera matrix was generally found to vary between 35 and 55 m. Using the above relation between U_b and h, these critical thickness values correspond to sliding ice velocities of 3·2 and 5·0 m a^{-1}

respectively. Thus it would take between 20 and 44 years, following the establishment of critical flow conditions at any grid square, for the ice mass to be transferred to an adjacent grid square. An examination of the sites of local glacier generation in the Thvera matrix further revealed that, within this period, an ice thickness equivalent to about two times the critical flow thickness would have been achieved.

Thus, in order to simplify the flow computations in the mass balance model, it is considered that once a grid square attains an ice thickness corresponding to twice its critical flow thickness, then half of the total water equivalent (composed of ice, firn, and snow) present is transferred to an adjacent grid square in line with the expected direction of flow. The water equivalent, density and thickness of the glacial material at the latter grid square are then recalculated, taking into account the amount of glacial material received and that already present from previous accumulation. If the ice thickness at the grid square receiving this additional glacial material is, in turn, found to exceed by more than two times its corresponding critical flow thickness, then half of its total water equivalent is transferred to next grid square in line with the 'flow' direction. This process is continued until the ice thickness at the grid square receiving glacial material does not exceed the 2 x critical thickness threshold value, or the limits of the study area are reached.

The above procedure for the determination of 'glacier flow' is programmed in the subroutine FLOW which is called by the main program at yearly intervals of mass balance simulation. This routine provides a rough approximation of the manner in which the ice is initially expected to 'flow' from areas of positive mass balance. The initial direction of flow, from grid squares which have achieved critical flow conditions, is indicated for the simulated Thvera glacier in Fig. 9.17. In order to model the development of these glaciers up to their present limits, a more sophisticated treatment of glacier flow, incorporating the theoretical ideas of such workers as Nye, Weertman, Lliboutry, Budd and Kamp is needed. Work towards this end, in a related study to the present one, is in progress (Richard Hindmarsh, personal communication).

9.3 Simulation Experiments

Following the above attempts at testing model validity, the application of the mass balance model was illustrated further by conducting a number of simulation runs using varying climatic inputs. While there can be no unique solution to the problem of defining the climatic conditions necessary to produce a given pattern of glacier mass balance distribution, changes in glacierisation are usually expressed in terms of changing temperature and precipitation conditions (sec. 2.3.5). Thus the effect of climatic changes on glacier balance in the Nupur 'test' matrix (Fig. 5.21) was simulated by the following changes (from the 1931-60 averages) in the air temperature and precipitation conditions at the base station:

(i) 2 degree overall temperature drop.

(ii) 2 degree 'winter temperature' drop (representing the drop in temperature during the accumulation months, Nov. - April).

(iii) 2 degree 'summer temperature' drop (representing the drop in temperature during the ablation months, May - Oct.).

(iv) 50% increase in precipitation.

(v) 100% increase in precipitation.

(vi) 4 degree overall temperature drop.

(vii) 2 degree temperature drop + 50% increase in precipitation.

The distribution of glacier mass balance (ice thickness above 10 m) after 30 years of simulation for each of the above cases is illustrated in Figs. 9.19 and 9.25. The variation of simulated mass balance over the 'test' matrix can be analysed by use of the transparent overlay for the Central Nupur matrix (provided in the pocket bound in the back of this thesis).

An analysis of the simulated results reveals that with a 2 degree drop in air temperature, glacier ice is predicted to attain sufficient thickness for incipient glaciers to form in a number of the concavities found in the Central Nupur matrix (Fig. 9.19). The location of these concavities adjacent to plateau remnants where they can trap drifting snow appears to be a significant factor in the formation of these local glaciers (as noted in sec. 3.2.1 with reference to snowpatch distribution). For example, the absence of glacierisation in the cirque located in the north-eastern corner of the matrix, viz. Berjadalur, (refer to Fig. 4.8), can be explained by its location beneath a sharp ridge with limited area compared with the more extensive stretches of upland plateau adjacent to the other glacierised concavities.

The importance of such summit plateau areas for snow-drift nourishment and the survival of cirque glaciers has also been emphasised by researchers in the Colorado Front Range (Outcalt and MacPhail, 1965), Rocky Mountains (Graf, 1976), Urals (Troitskiy et al., 1966), and Cairngorms (Sugden, 1977c).

The relative effects of a 2 degree 'winter' cooling compared with a 2 degree 'summer' cooling are shown in Figs. 9.20 and 9.21 respectively. It can be seen that a greater number of grid squares with ice thickness of 10 m or more are predicted with a 2 degree temperature drop during the summer months compared with a temperature drop of the same magnitude during the winter months. This appears to reflect the fact that a reduction in the ablation of glacial material (as a result of summer cooling) is more critical for glacier balance in the Central Nupur matrix than is increased accumulation (as a result of winter cooling).

It is interesting to note that a 50% increase in annual precipitation at the base station results in a pattern of glacierisation (Fig. 9.22) similar to that achieved with a 2 degree drop in overall air temperature (Fig. 9.19). Simulating glacier balance with more extreme climatic deteriorations, e.g. 100% precipitation increase (Fig. 9.23) and a 4 degree temperature drop (Fig. 9.24) results in further sites becoming available for glacier generation in accordance with the "law of decreasing glacial asymmetry with increasing glacial cover" (Evans, 1977 p.169). In the above simulation experiments the most widespread distribution of glacier ice was produced by the combined effect of a 2 degree drop in air temperature and a 50% increase in annual precipitation (Fig. 9.25).

It should be noted that the presence of any substantial glacier ice in Berjadalur is only effected with rather extreme climatic changes (e.g. 100% precipitation increase or a 4 degree drop in air temperature) to offset the lack of snow-drift accumulation. These simulations also reveal that in Grjótdalur, it is only the southern part of the cirque (i.e. the north-east facing slopes) that is predicted to contain glacier ice. This appears to reflect the differential income of solar radiation on slopes of differing aspects.

These simulation experiments serve to illustrate how the effect of a given climatic change on glacier mass balance can be analysed.

VALUES BETWEEN 0.100 5.000 10.000 20.000 35.000 55.000

NUPUR MASS BALANCE SIMULATION (1931-60 CLIMATE) - BLOCK 1
WATER EQUIVALENT (CM) OCTOBER (A=1)

FIGURE 9.1
Nupur Mass Balance Simulation (1931-60): October (A=1)

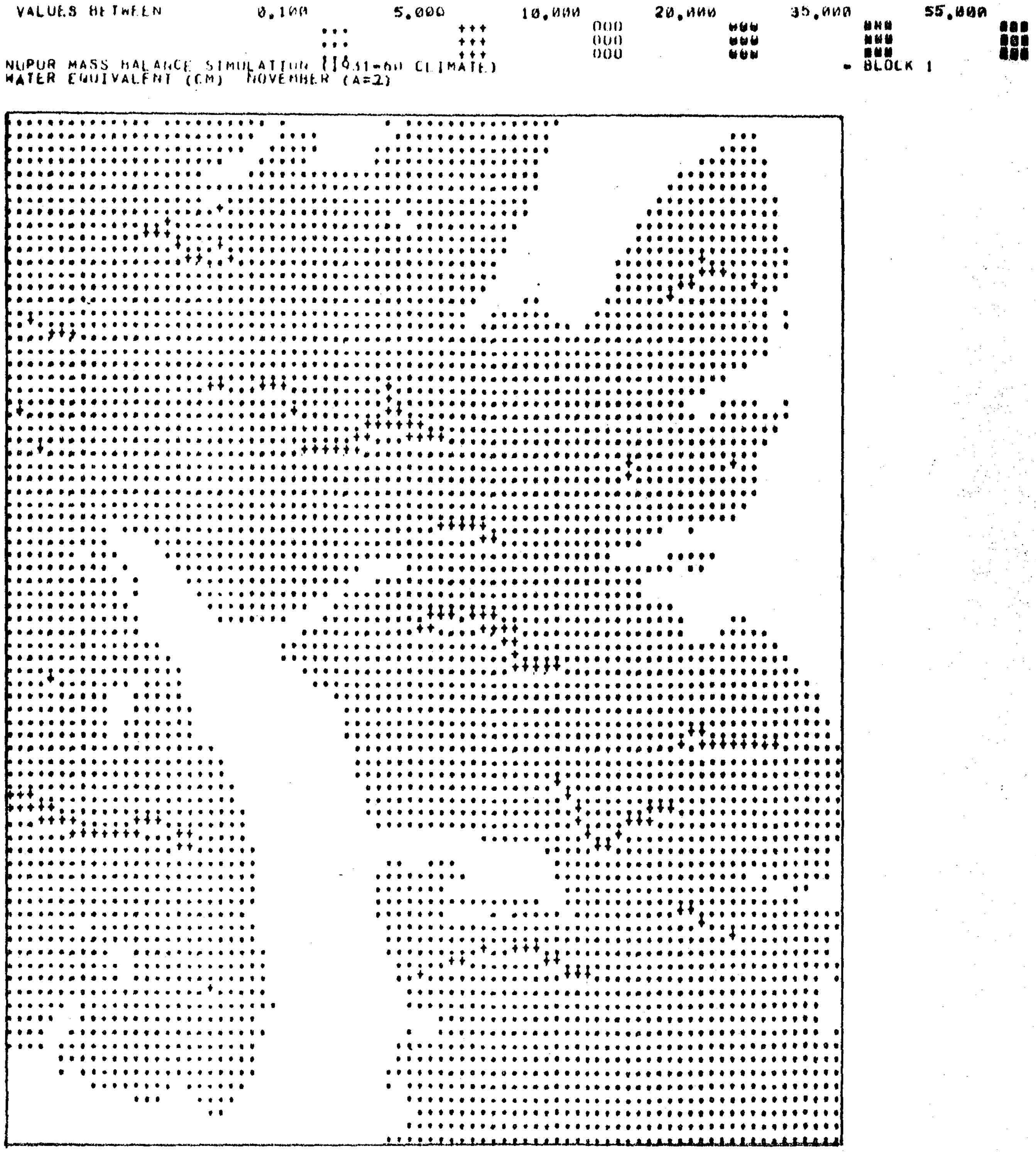


FIGURE 9.2
Nupur Mass Balance Simulation (1931-60): November (A=2)

VALUES BETWEEN 0.100 5.000 10.000 20.000 35.000 55.000

NUPUR MASS BALANCE SIMULATION (1931-60 CLIMATE) - BLOCK 1
WATER EQUIVALENT (CM) DECEMBER (A=3)

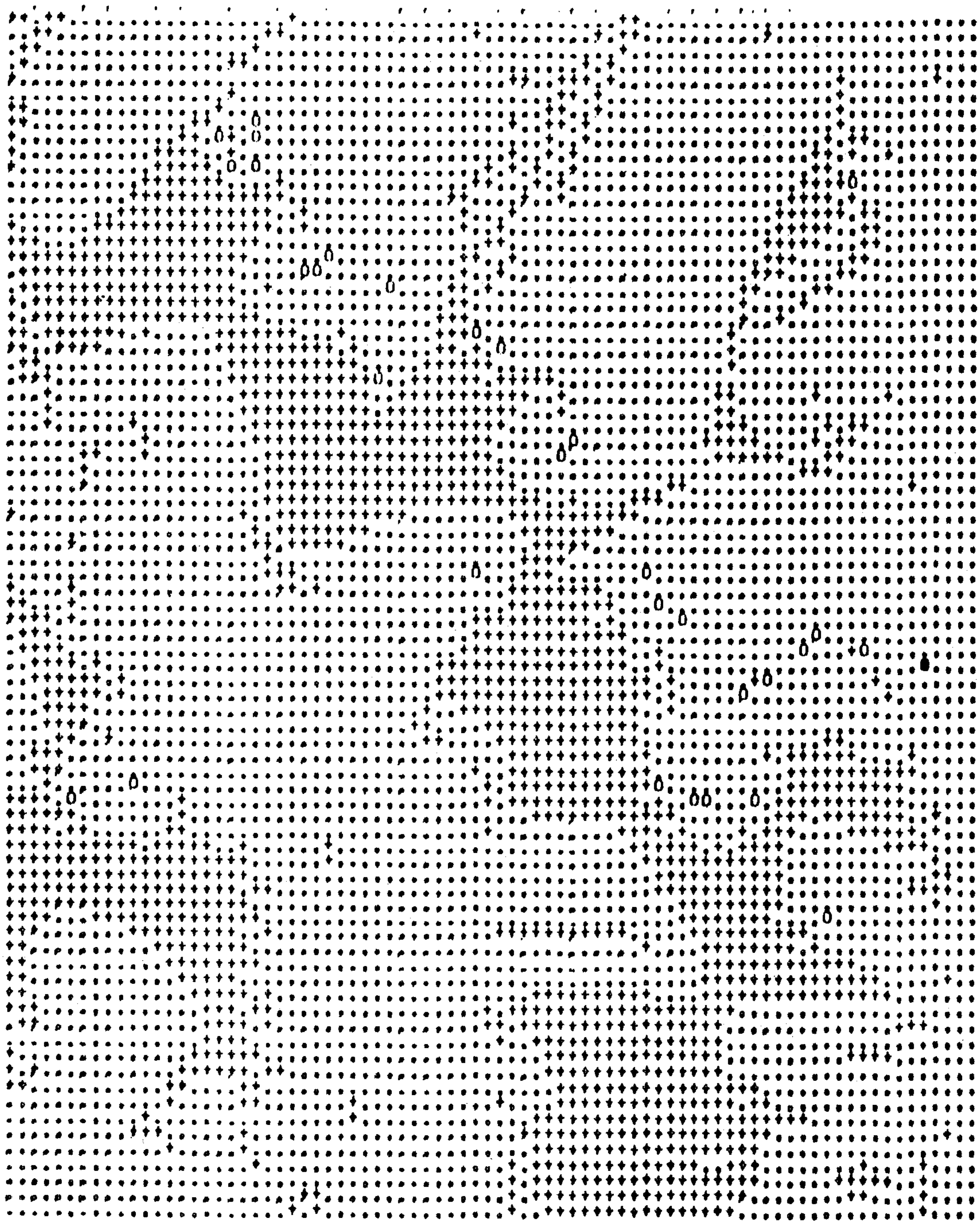


FIGURE 9.3
Nupur Mass Balance Simulation (1931-60): December (A=3)

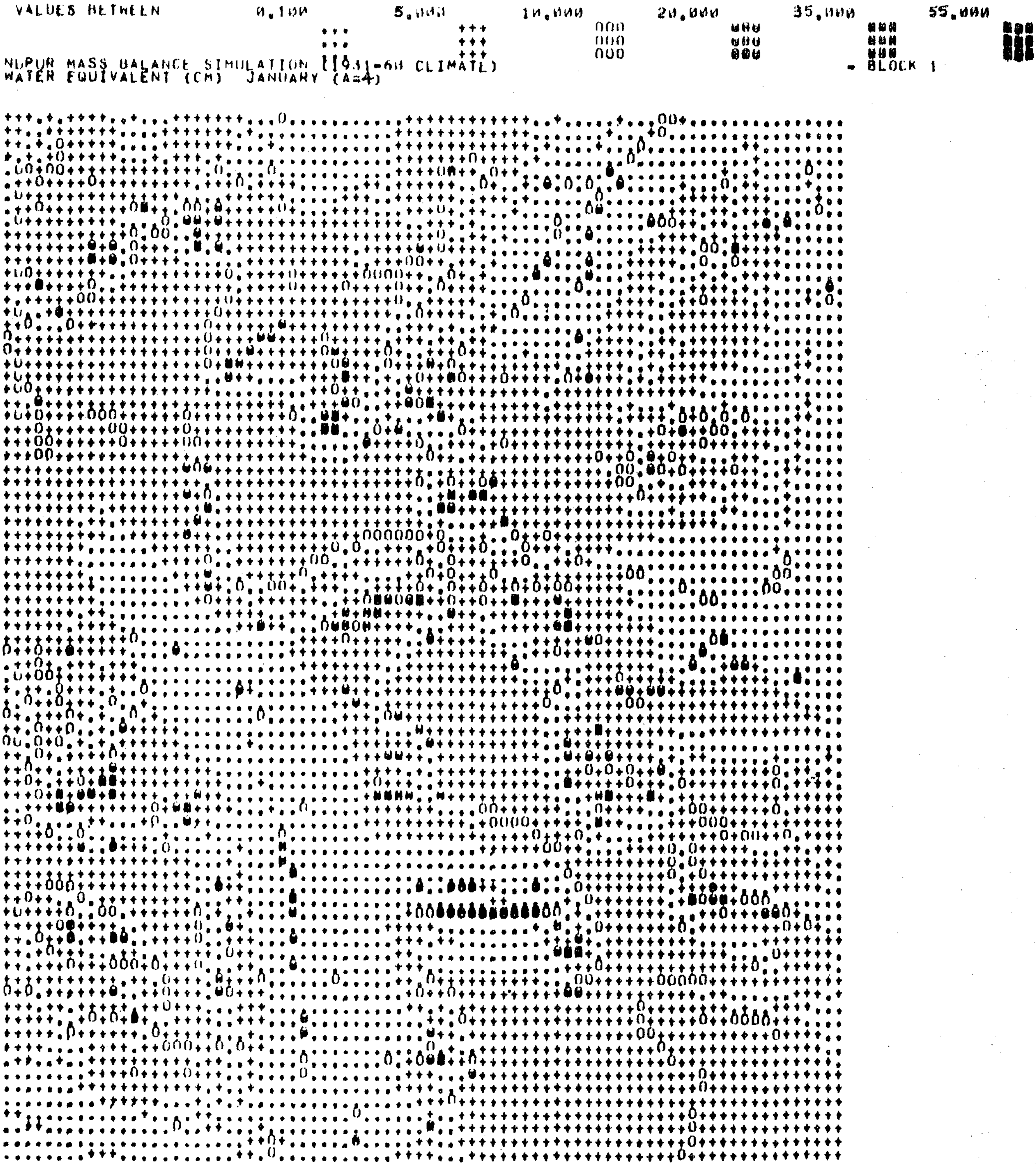


FIGURE 9.4
Nupur Mass Balance Simulation (1931-60): January (A=4)

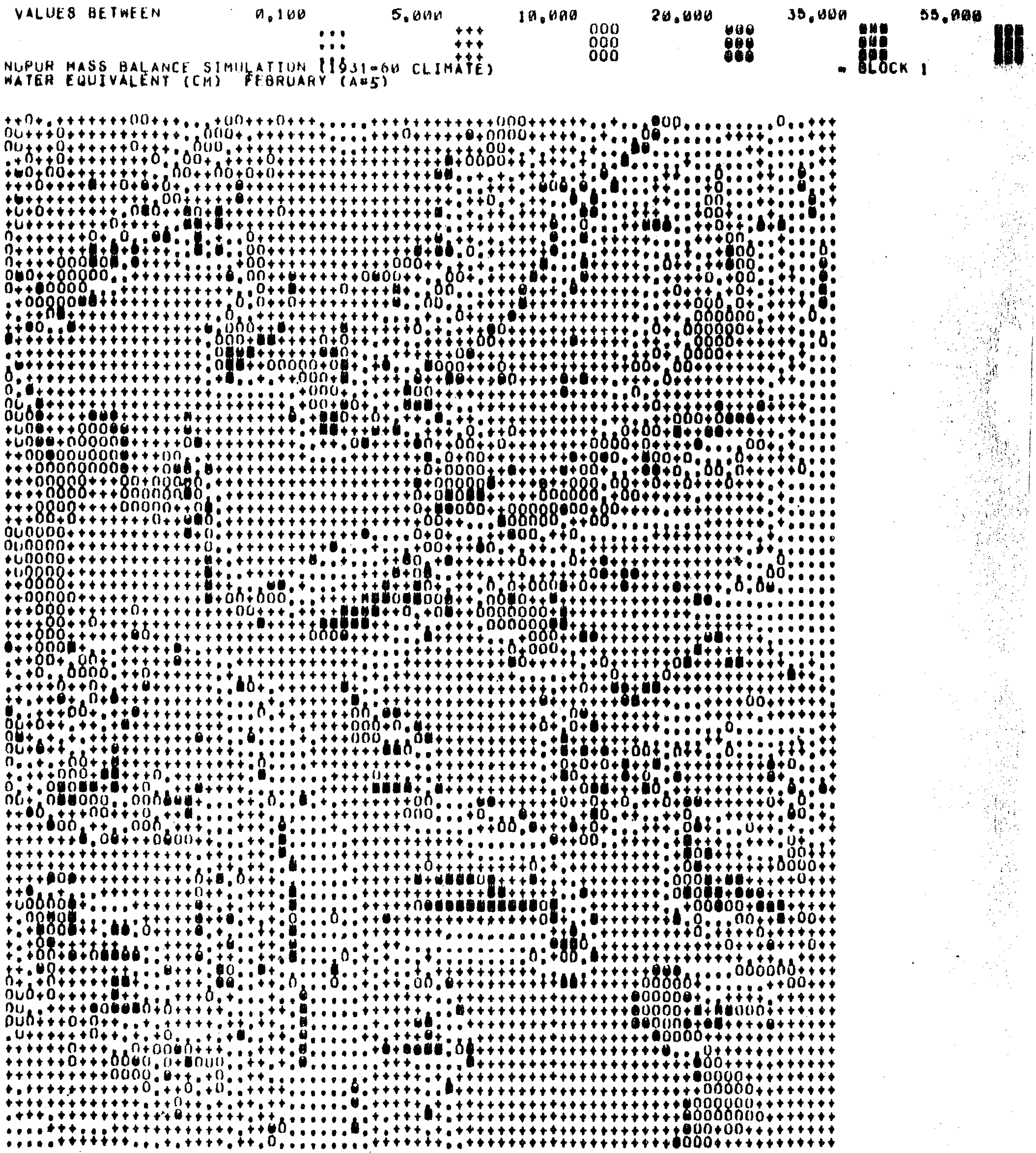


FIGURE 9.5
Nupur Mass Balance Simulation (1931-60): February (A=5)

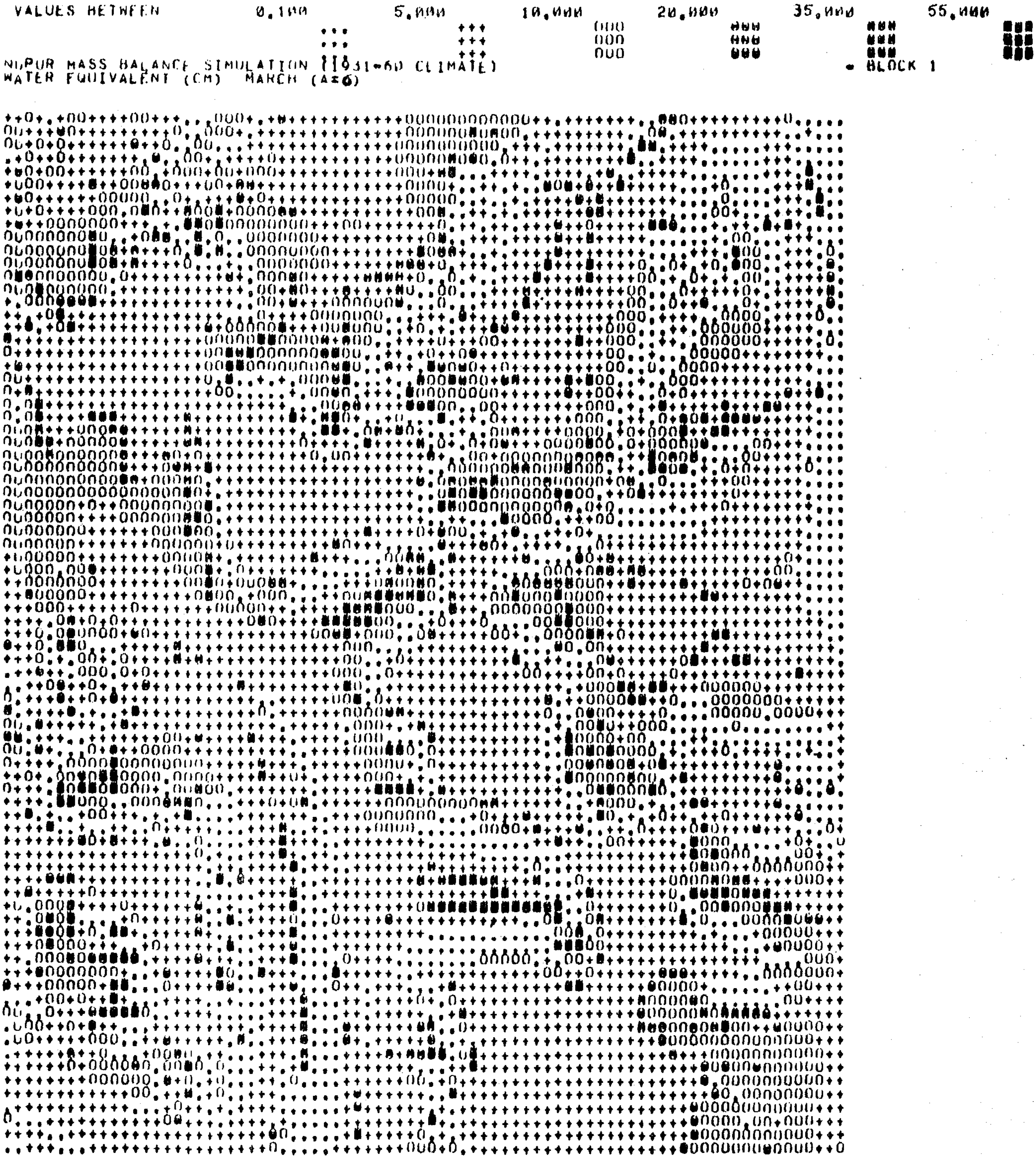


FIGURE 9.6
Nupur Mass Balance Simulation (1931-60): March (A=6)

VALUES BETWEEN 0.100 5.000 10.000 20.000 35.000 55.000

NUPUR MASS BALANCE SIMULATION (1931-60 CLIMATE) - BLOCK 1
WATER EQUIVALENT (CM) APRIL (A=7)

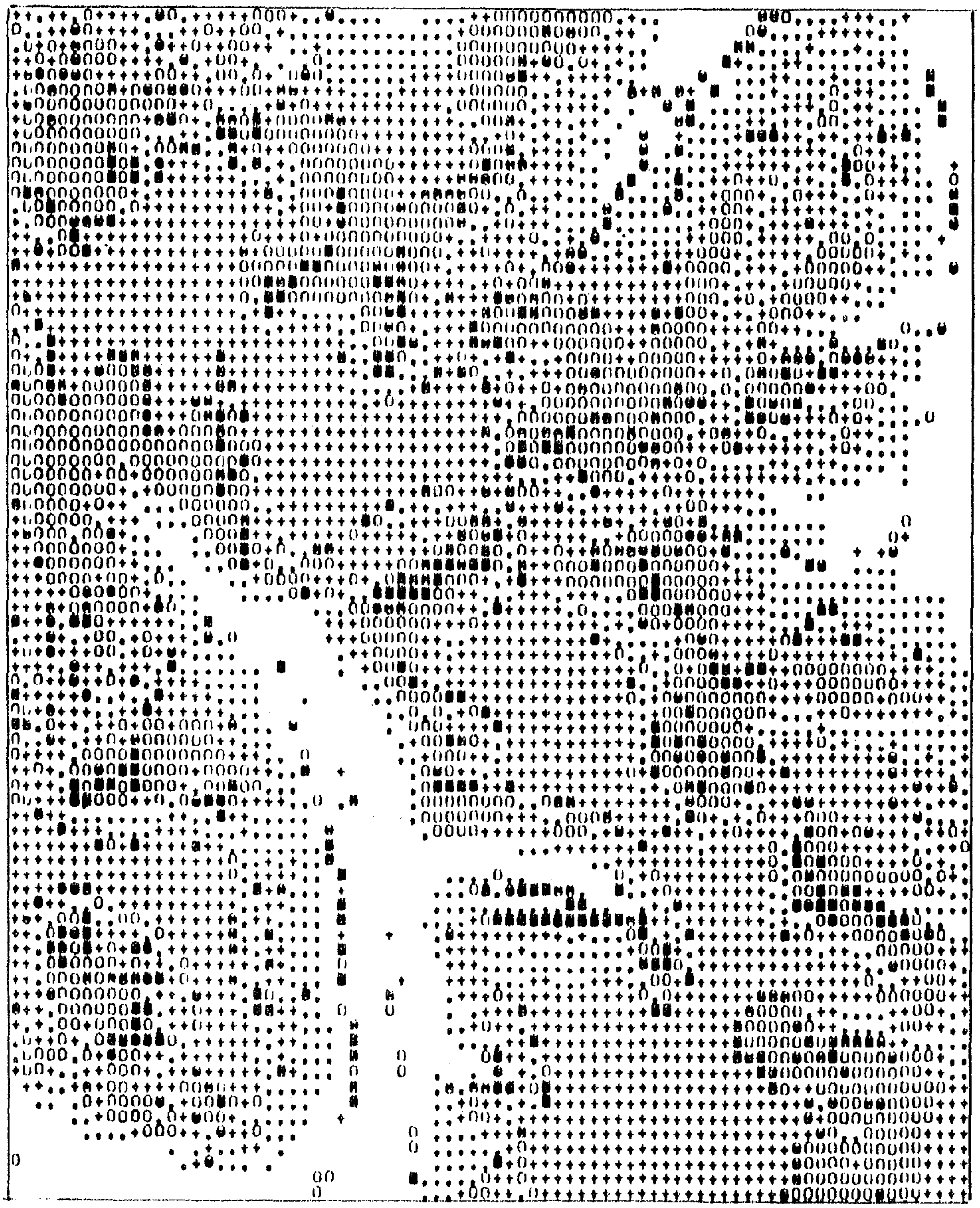


FIGURE 9.7
Nupur Mass Balance Simulation (1931-60): April (A=7)

VALUES BETWEEN 0.100 5.000 10.000 20.000 35.000 55.000

NUPUR MASS BALANCE SIMULATION (1931-60 CLIMATE) - BLOCK 1
WATER EQUIVALENT (CM) MAY (A=8)

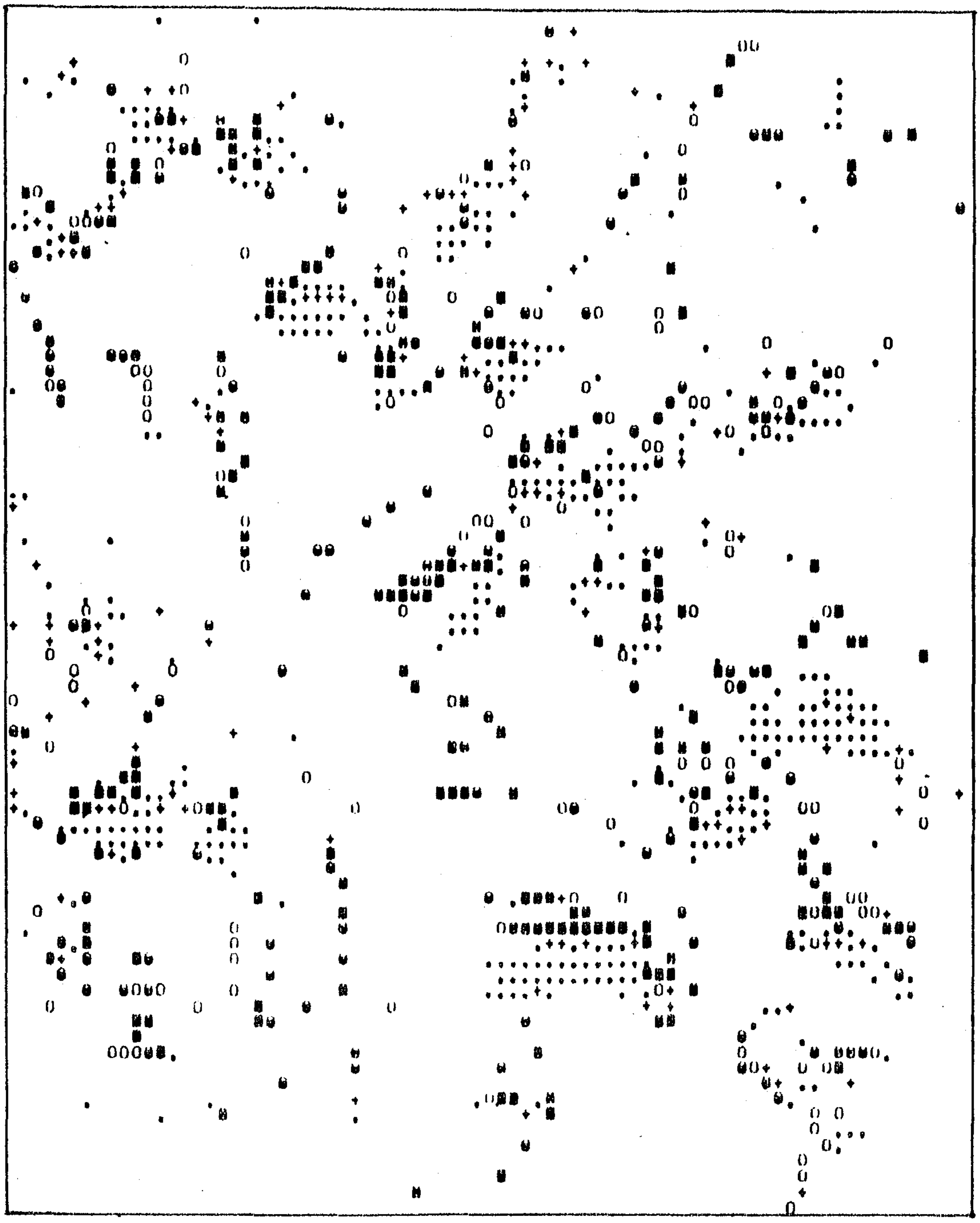


FIGURE 9.8
Nupur Mass Balance Simulation (1931-60): May (A=8)

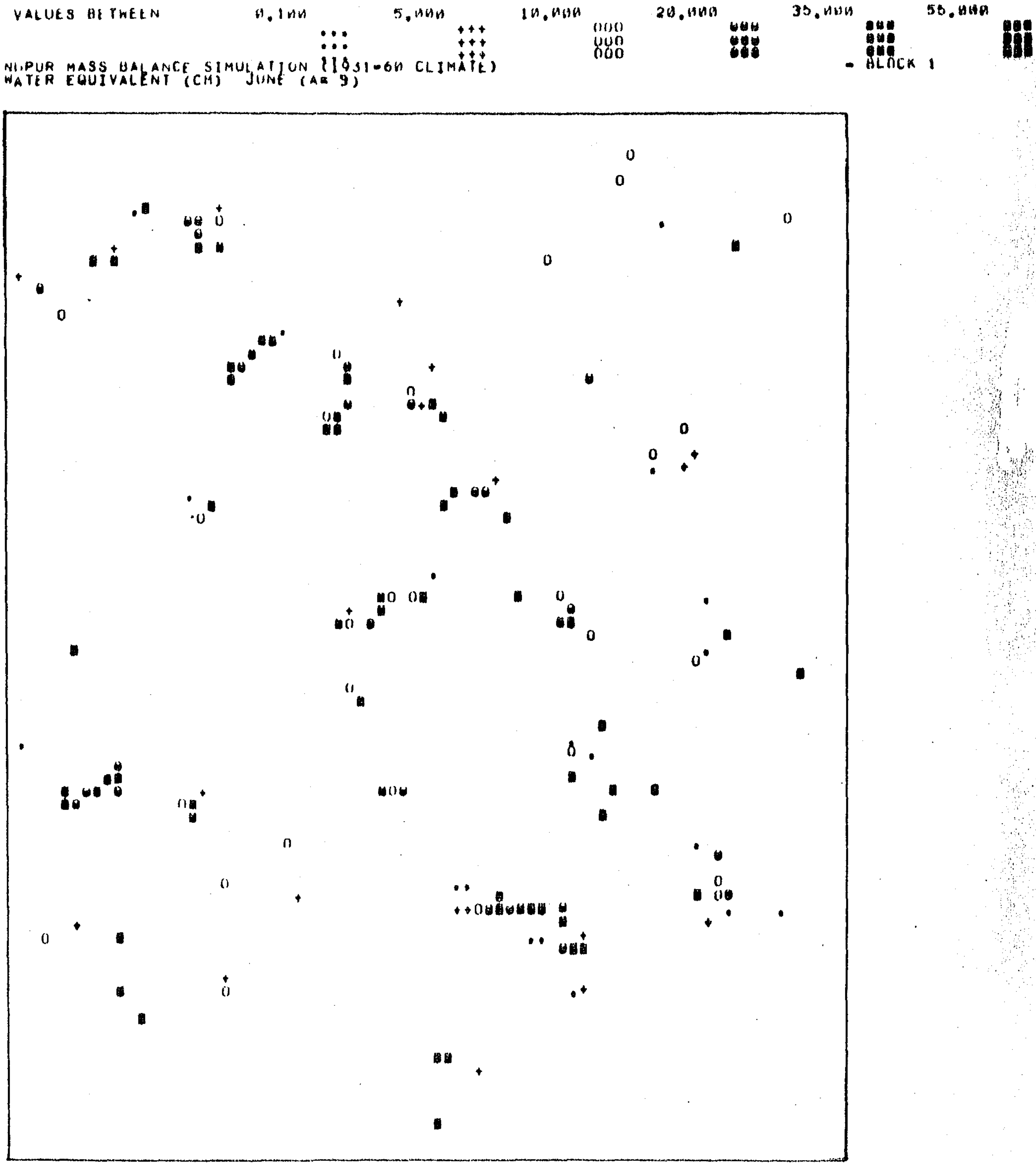


FIGURE 9.9
Nupur Mass Balance Simulation (1931-60): June (A=9)

VALUES BETWEEN 0.100 5.000 10.000 20.000 35.000 55.000

NUPUR MASS BALANCE SIMULATION (1931-60 CLIMATE) - BLOCK 1
WATER EQUIVALENT (CM) JULY (A=10)

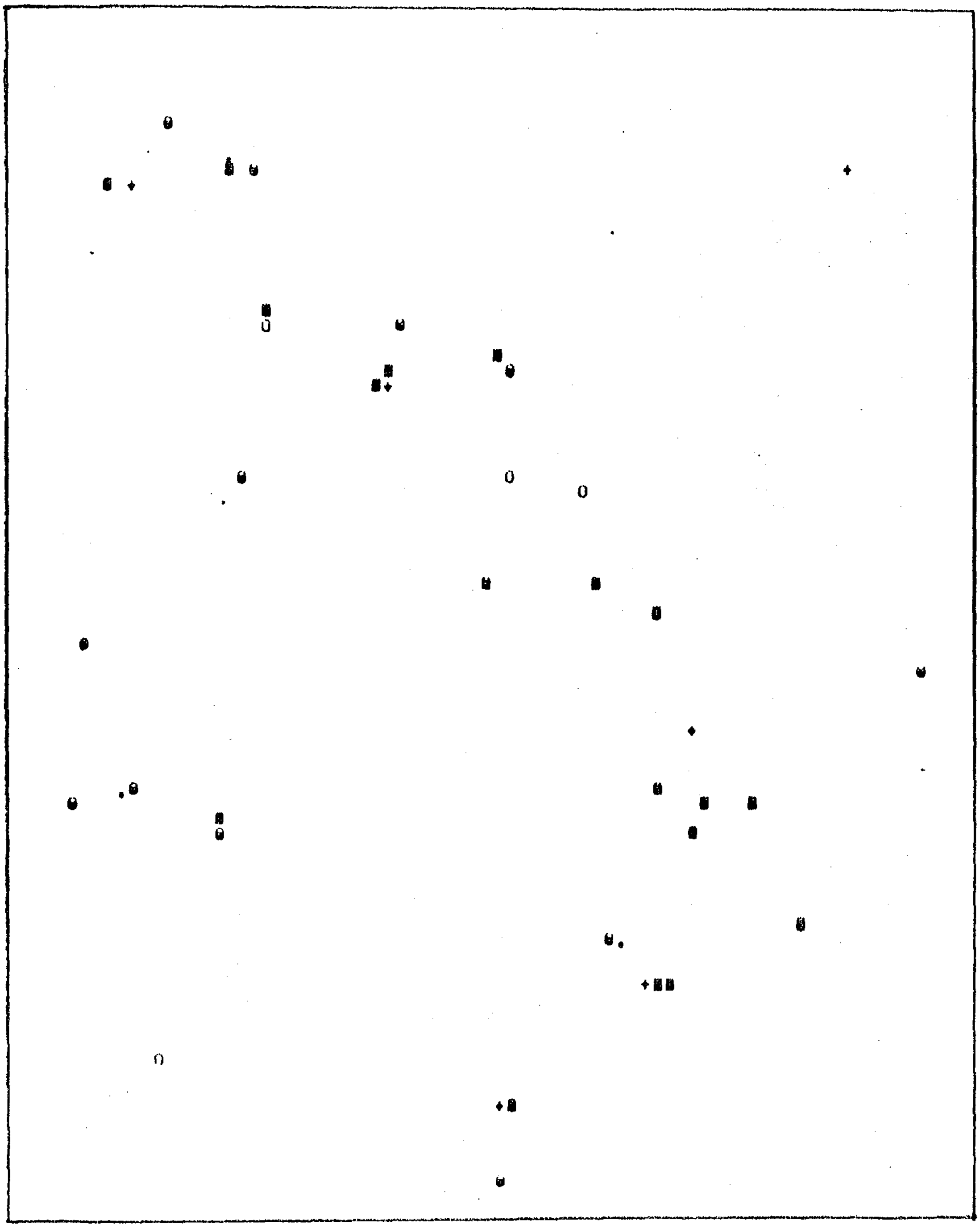


FIGURE 9.10
Nupur Mass Balance Simulation (1931-60): July (A=10)

VALUES BETWEEN 0.100 5.000 10.000 20.000 35.000 55.000

NUPUR MASS BALANCE SIMULATION (1931-60 CLIMATE) - BLOCK 1
WATER EQUIVALENT (CM) AUGUST (A=11)

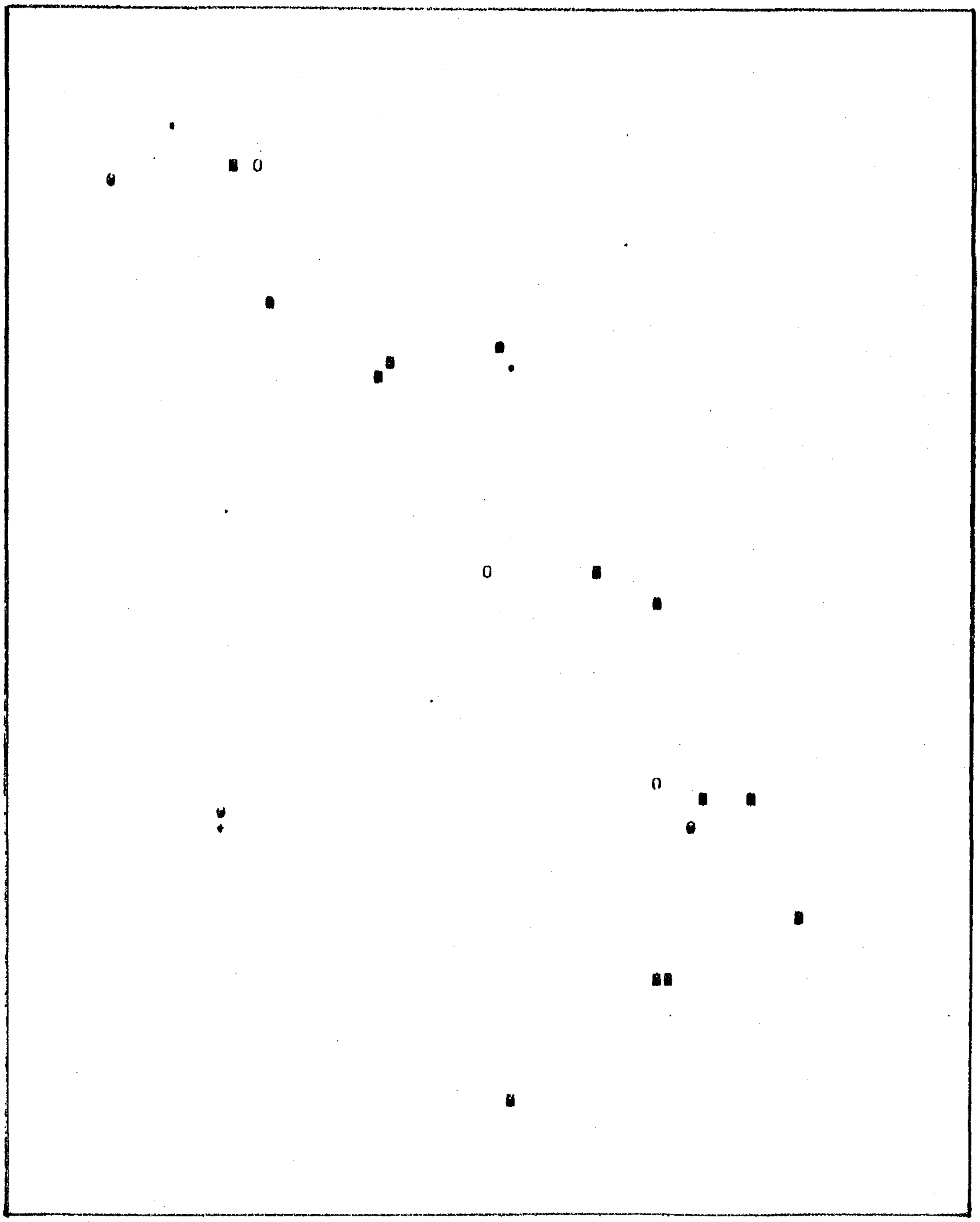


FIGURE 9.11
Nupur Mass Balance Simulation (1931-60): August (A=11)

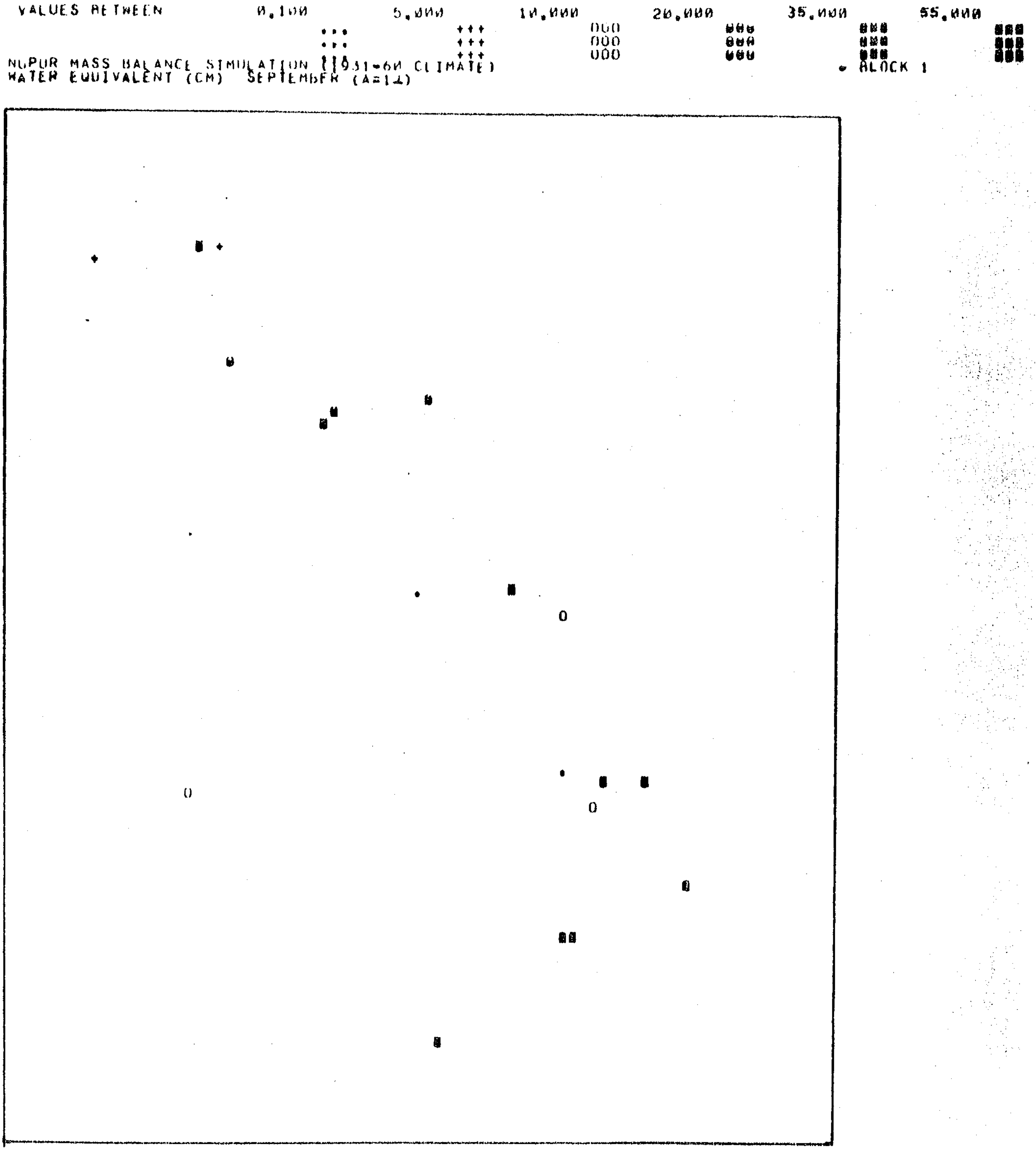


FIGURE 9.12
Nupur Mass Balance Simulation (1931-60): September (A=12)

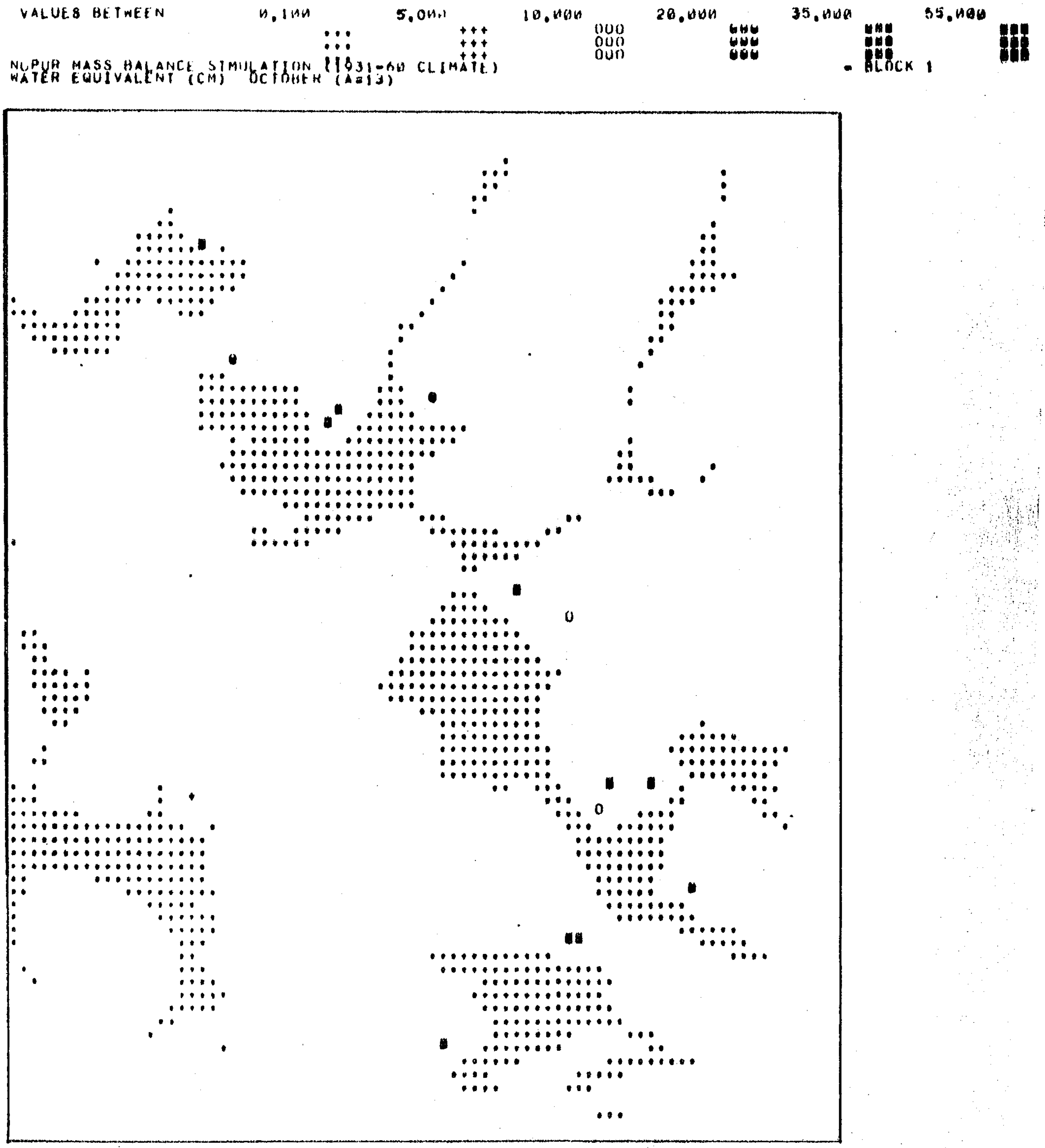


FIGURE 9.13
Nupur Mass Balance Simulation (1931-60): October (A=13)

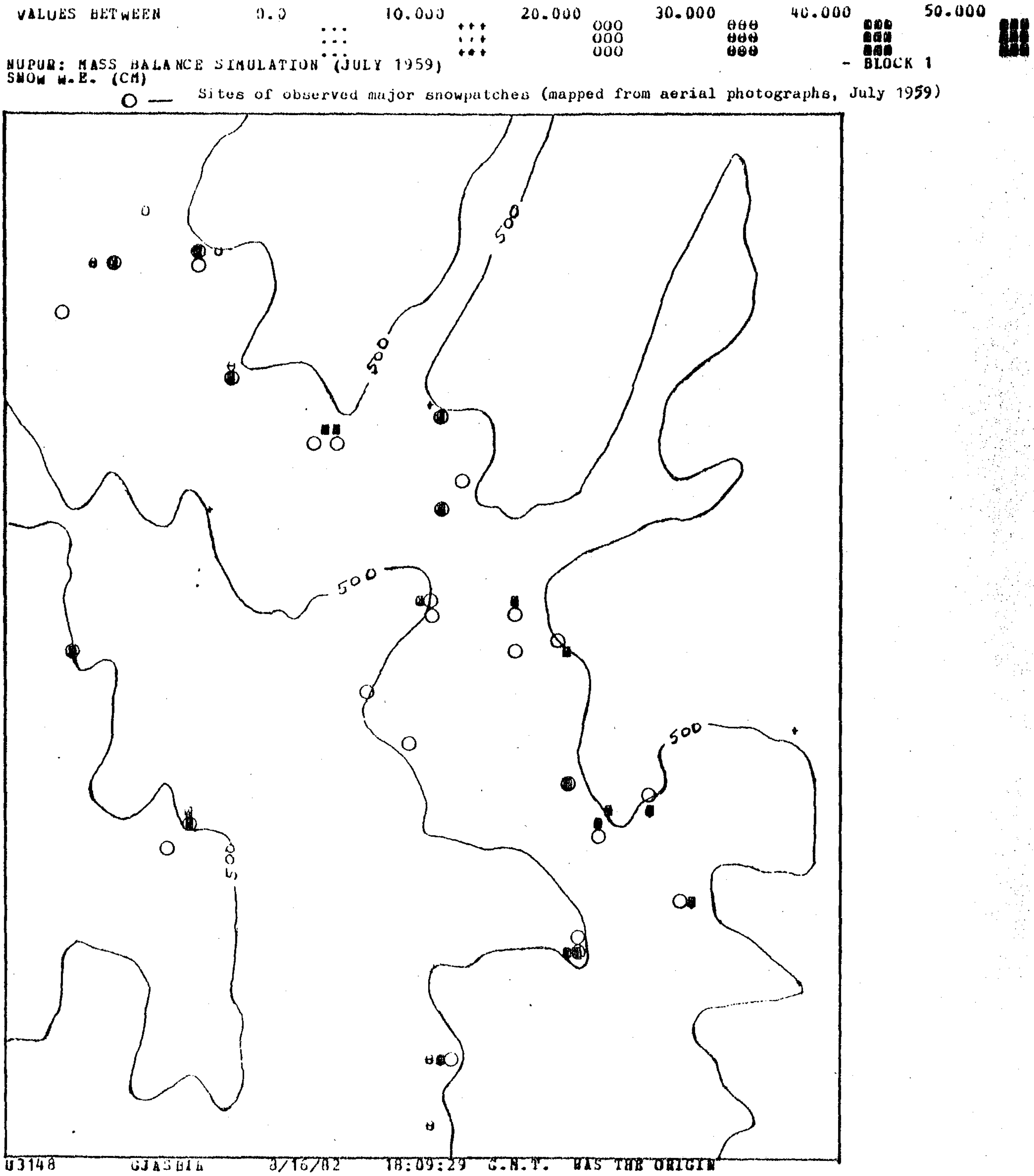


FIGURE 9.14
Comparison of Nupur Mass Balance Simulation Results (for July 1959) with the Distribution of Major Snowpatches in the Nupur Matrix (mapped from aerial photographs, July 1959).

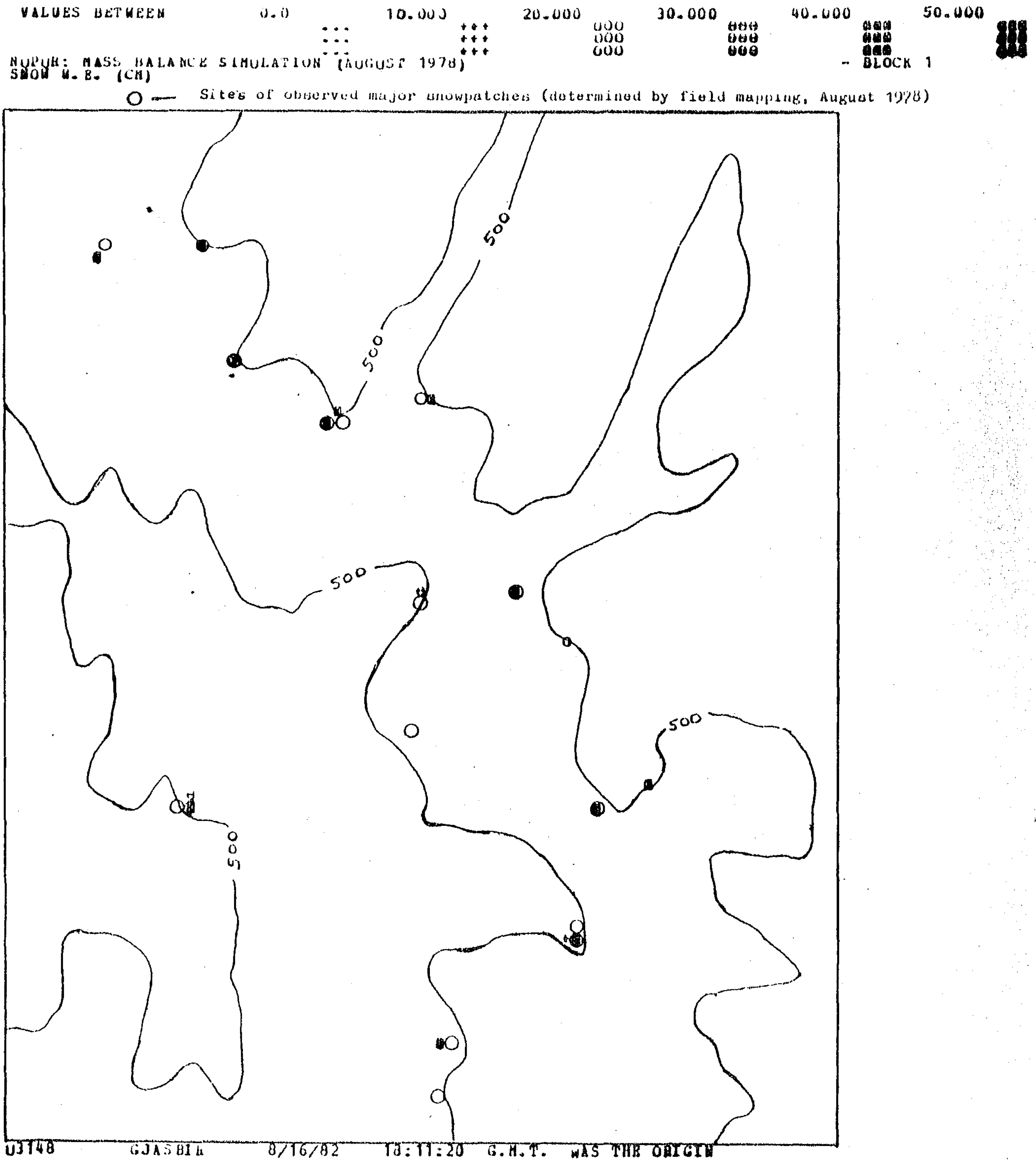


FIGURE 9.15
Comparison of Nupur Mass Balance Simulation Results (for August 1978) with the Distribution of Major Snowpatches in the Nupur Matrix (determined by field mapping, August 1978).

FIGURE 9.16
Local Glacier Distribution in the Thvera Matrix

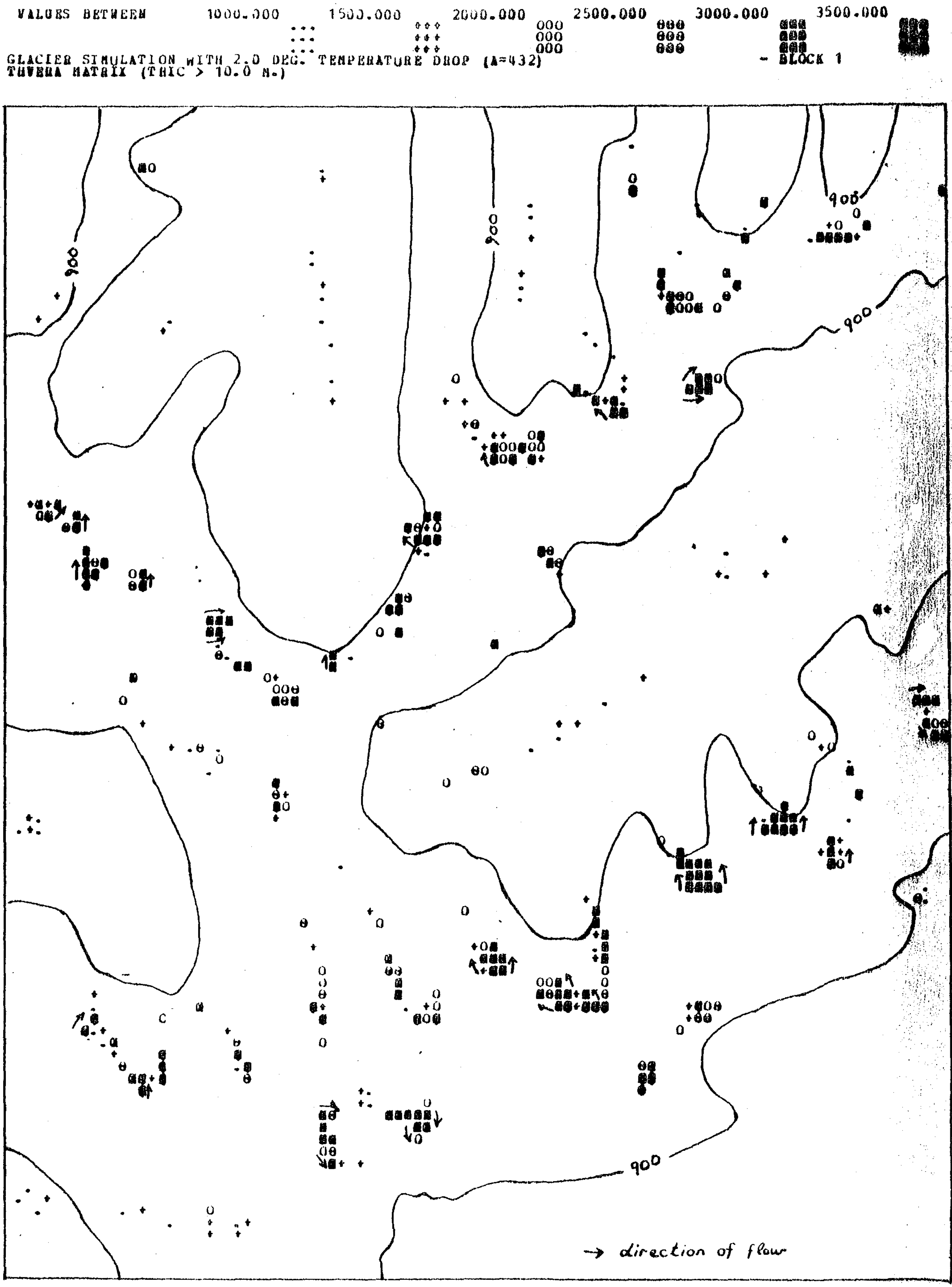


FIGURE 9.17
Glacier Simulation in the Thvera Area (with 2^{o}C temperature drop)

VALUES BETWEEN 0.0 5.000 10.000 15.000 20.000 25.000

PERCENTAGE REDUCTION IN DIRECT S.W. RADIATION DUE TO TOPOGRAPHIC SHADING. - BLOCK 1
THVERA MATRIX (JULY : SCANNING FOR A DISTANCE OF 4 KM.)

FIGURE 9.18
Percentage Reduction of July Solar Radiation Due to Topographic Shading in the Thvera Matrix

VALUES BETWEEN 1000.000 1500.000 2000.000 2500.000 3000.000 3500.000

NUPUR GLACIER BALANCE SIMULATION (2 DEG. TEMP DROP) - BLOCK 1
ICE THICKNESS (CM). A=360

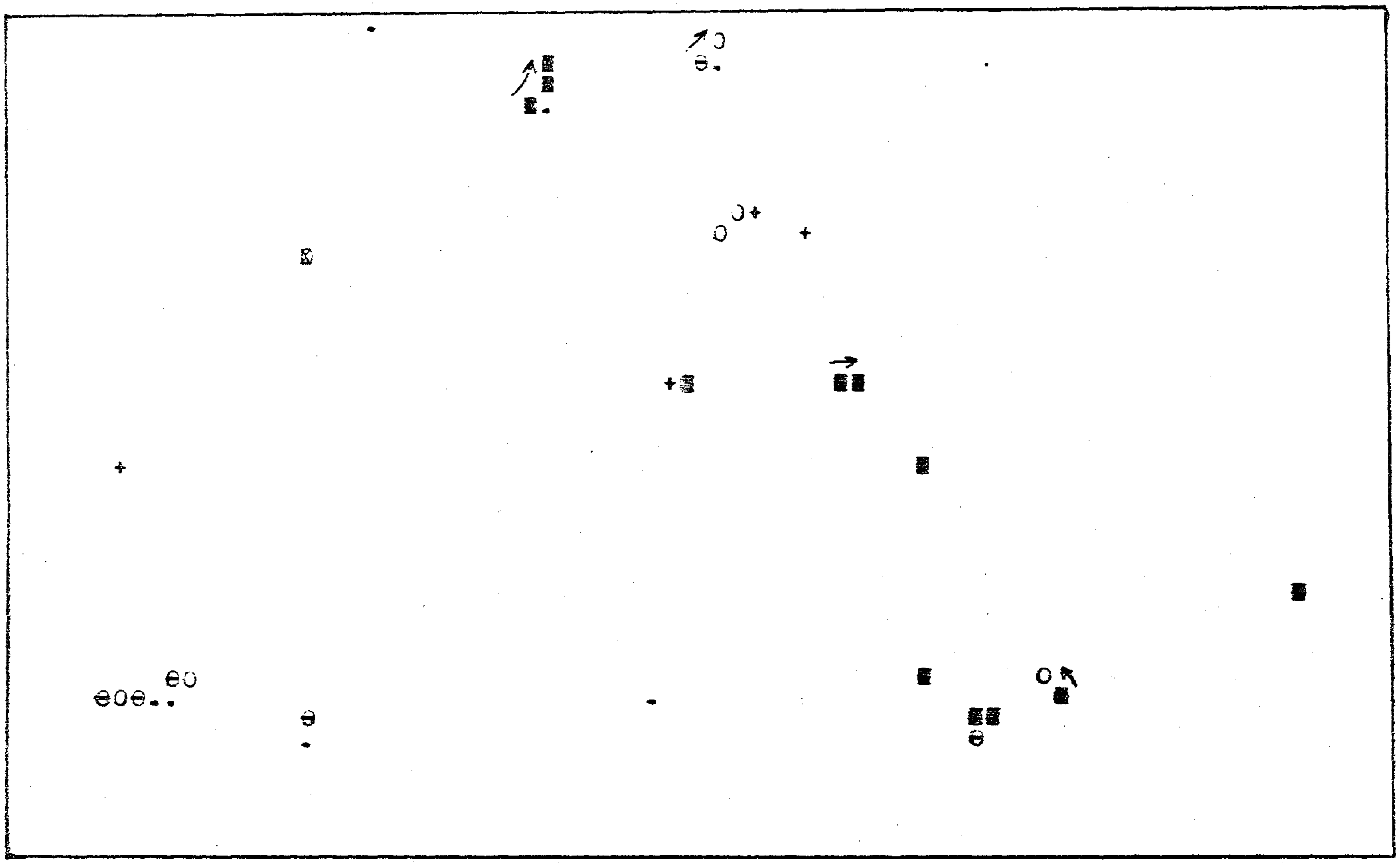


FIGURE 9.19
Central Nupur Glacier Balance Simulation ($2^{o}C$ temperature drop)

VALUES BETWEEN 1000.000 1500.000 2000.000 2500.000 3000.000 3500.000

NUPUR GLACIER BALANCE SIMULATION (2 DEG. 'WINTER TEMP' DROP) - BLOCK 1
ICE THICKNESS (CM). A=360

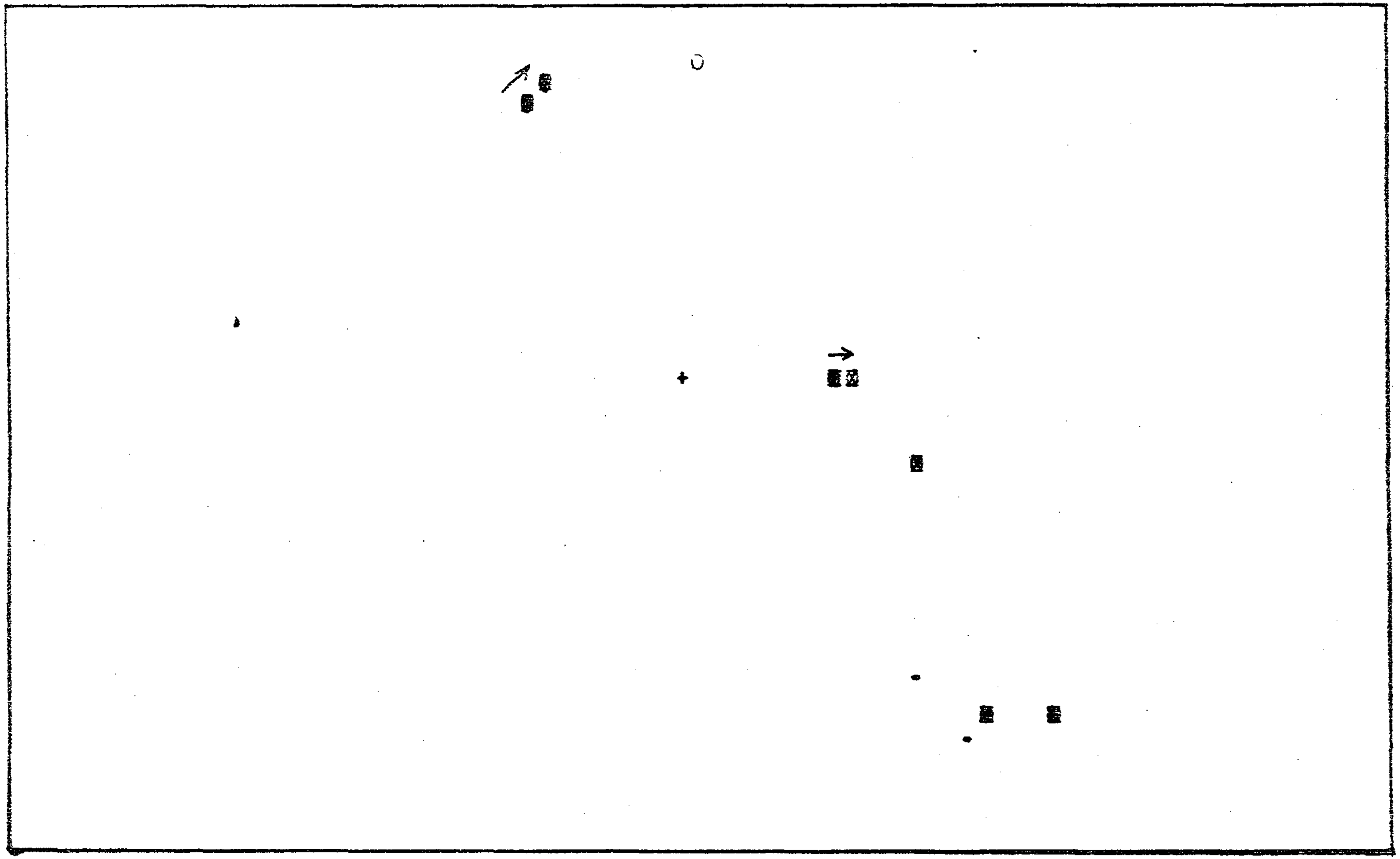


FIGURE 9.20
Central Nupur Glacier Balance Simulation (2^{o} 'winter temperature' drop)

VALUES BETWEEN 1000.000 1500.000 2000.000 2500.000 3000.000 3500.000

NUPUR GLACIER BALANCE SIMULATION (2 DEG. 'SUMMER TEMP' DROP) - BLOCK 1
ICE THICKNESS (CM). A=360

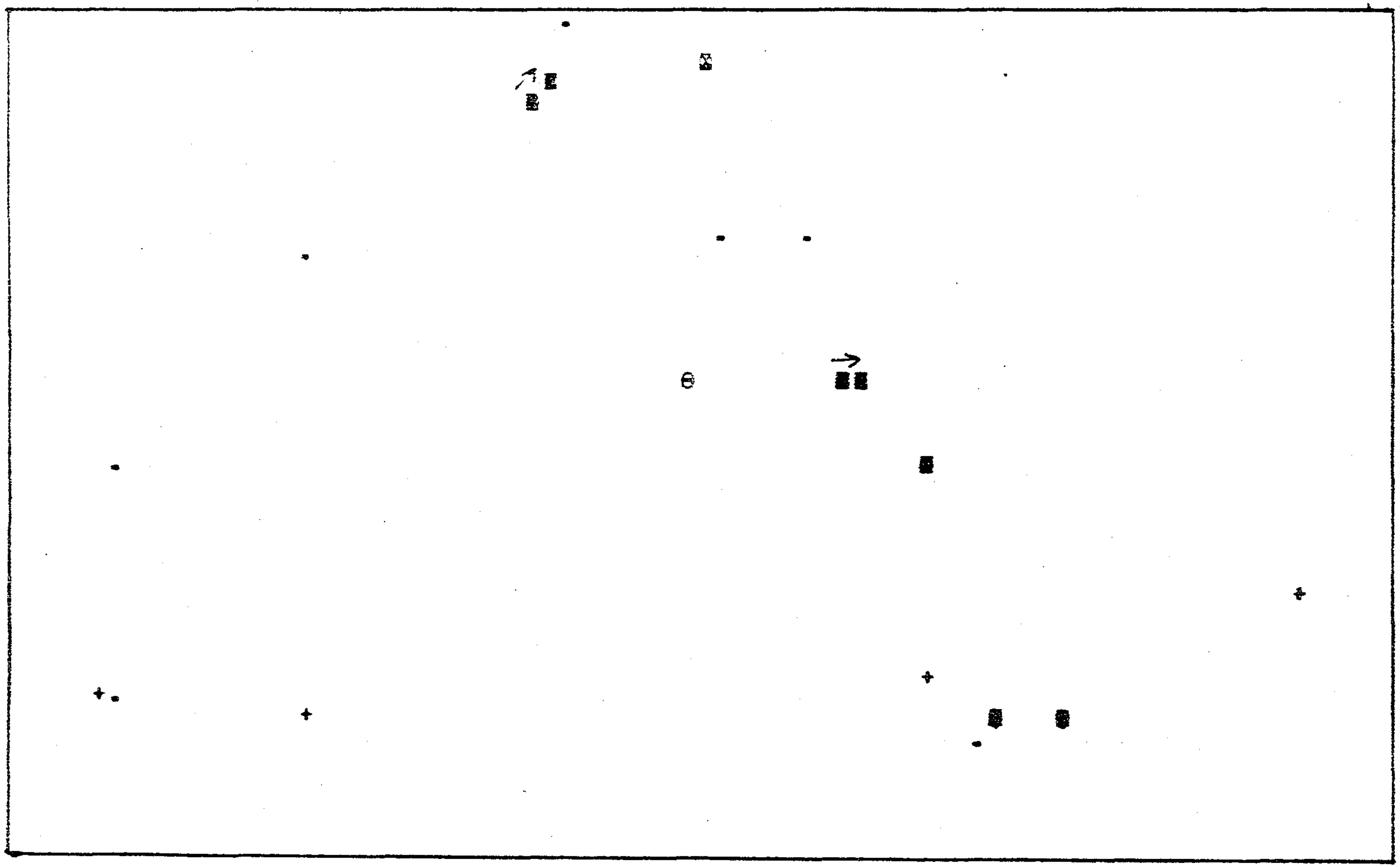


TABLE 9.21
Central Nupur Glacier Balance Simulation (2^{o} 'summer temperature' drop)

VALUES BETWEEN 1000.000 1500.000 2000.000 2500.000 3000.000 3500.000

NUPUR GLACIER BALANCE SIMULATION (50% PREC. INCREASE) - BLOCK 1
ICE THICKNESS (CM). A=360

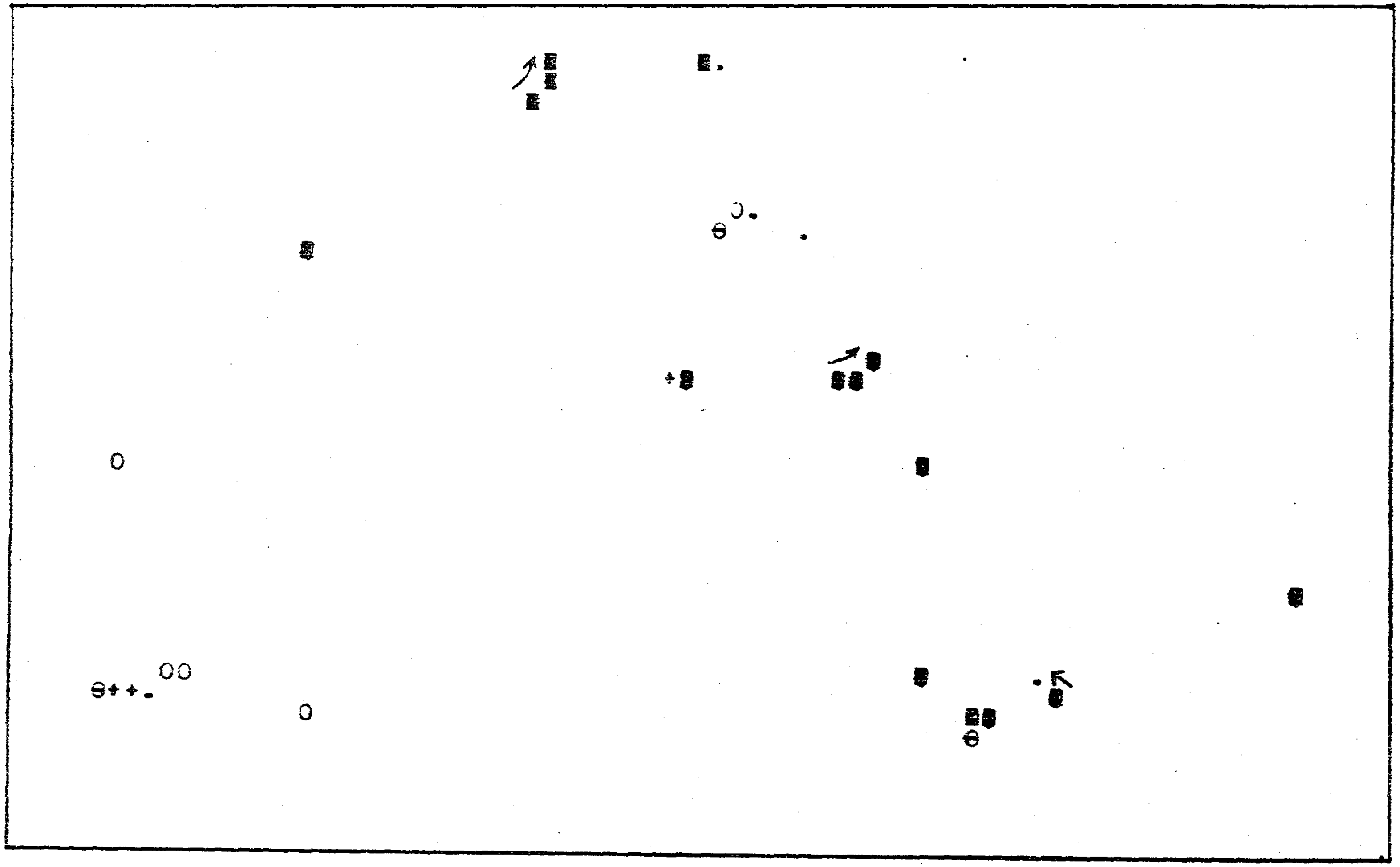


FIGURE 9.22
Central Nupur Glacier Balance Simulation (50% precipitation increase)

VALUES BETWEEN 1000.000 1500.000 2000.000 2500.000 3000.000 3500.000

NUPUR GLACIER BALANCE SIMULATION (100% PREC. INCREASE) - BLOCK 1
ICE THICKNESS (CM). A=360

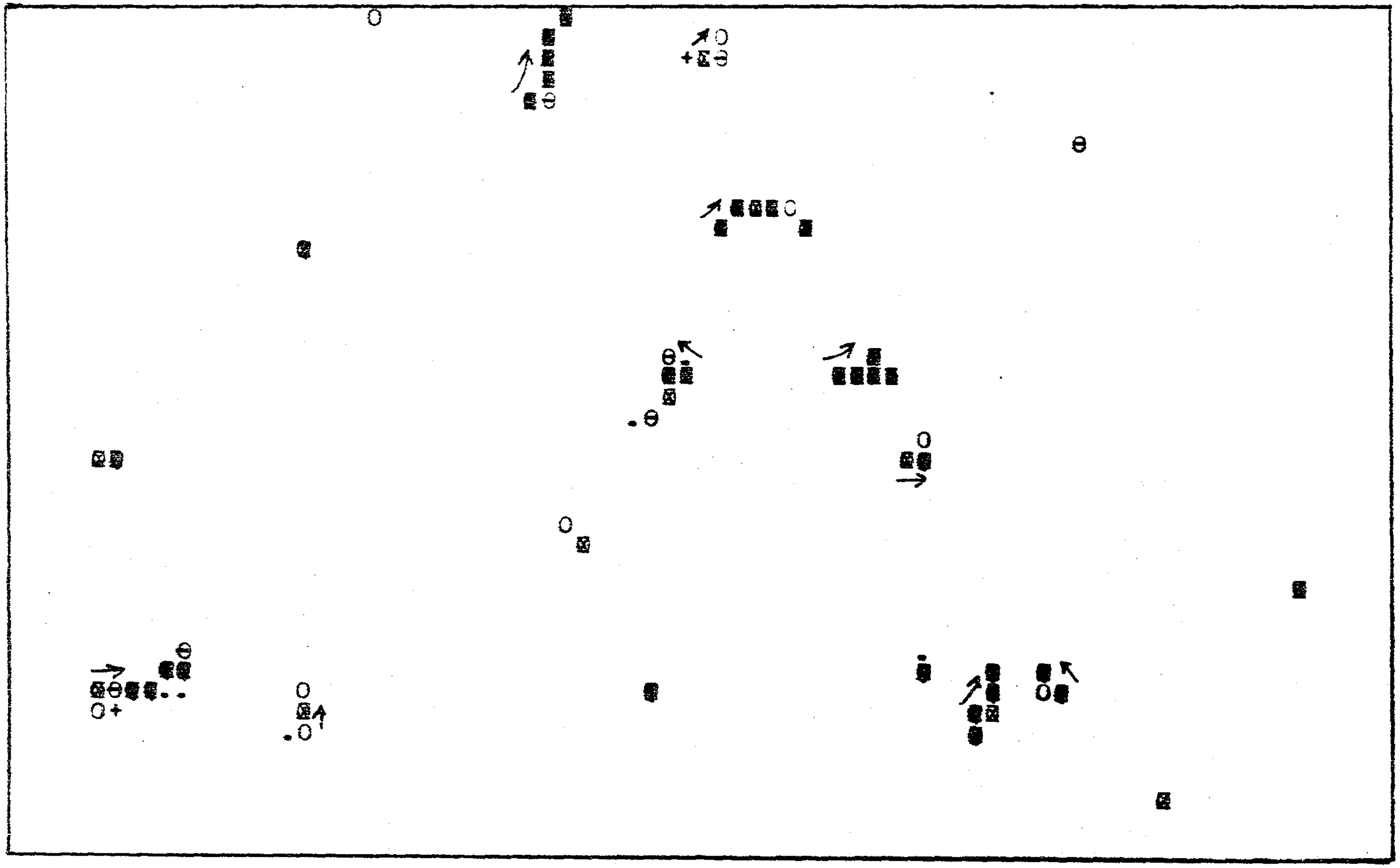


FIGURE 9.23
Central Nupur Glacier Balance Simulation (100% precipitation increase)

VALUES BETWEEN 1000.000 1500.000 2000.000 2500.000 3000.000 3500.000

NUPUR GLACIER BALANCE SIMULATION (4 DEG. TEMP DROP) - BLOCK 1
ICE THICKNESS (CM). A=360

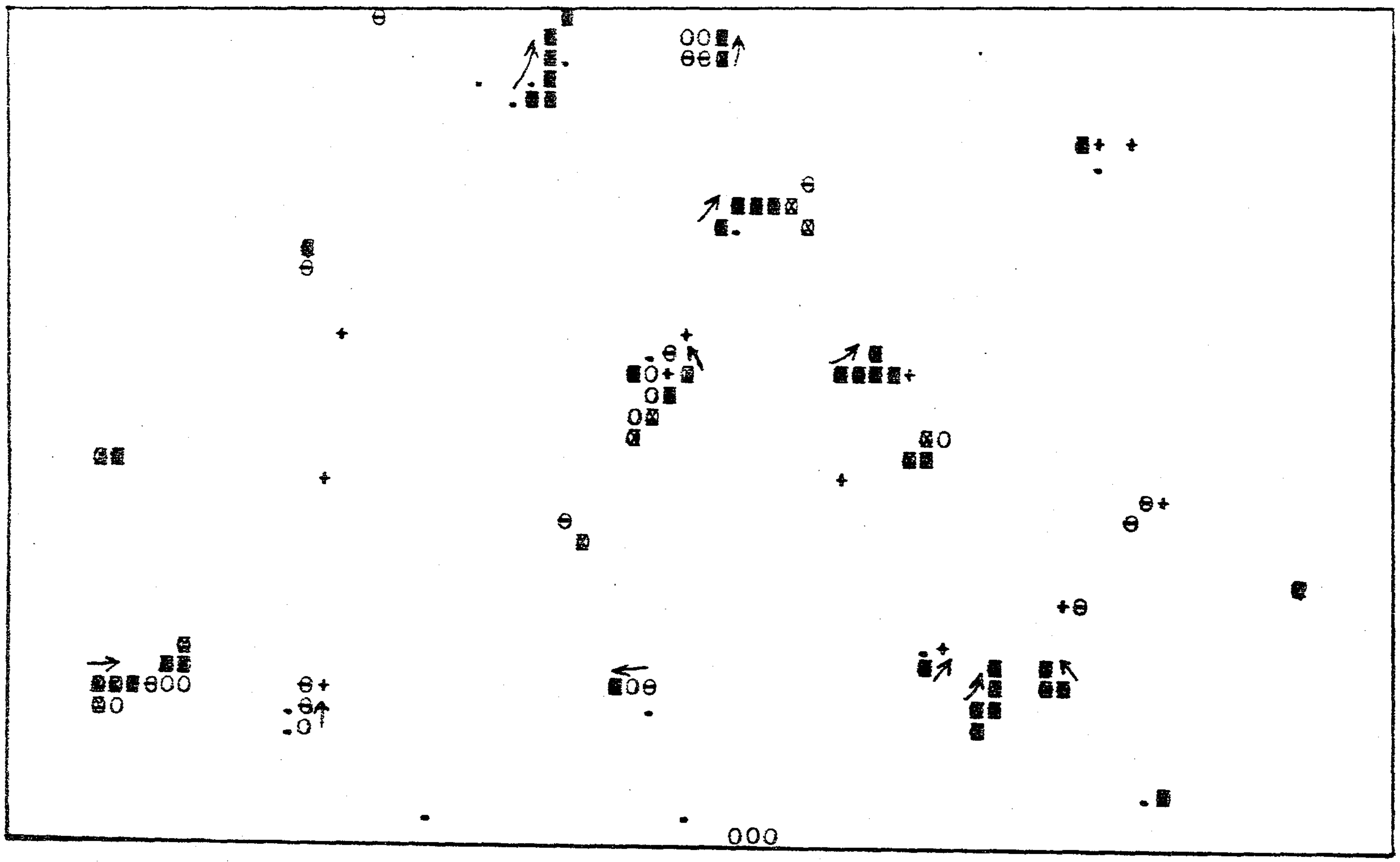


FIGURE 9.24
Central Nupur Glacier Balance Simulation (4° temperature drop)

VALUES BETWEEN 1000.000 1500.000 2000.000 2500.000 3000.000 3500.000

NUPUR GLACIER BALANCE SIMULATION(2 DEG. TEMP DROP + 50% PREC INCREASE) - BLOCK 1
ICE THICKNESS (CM). A=360

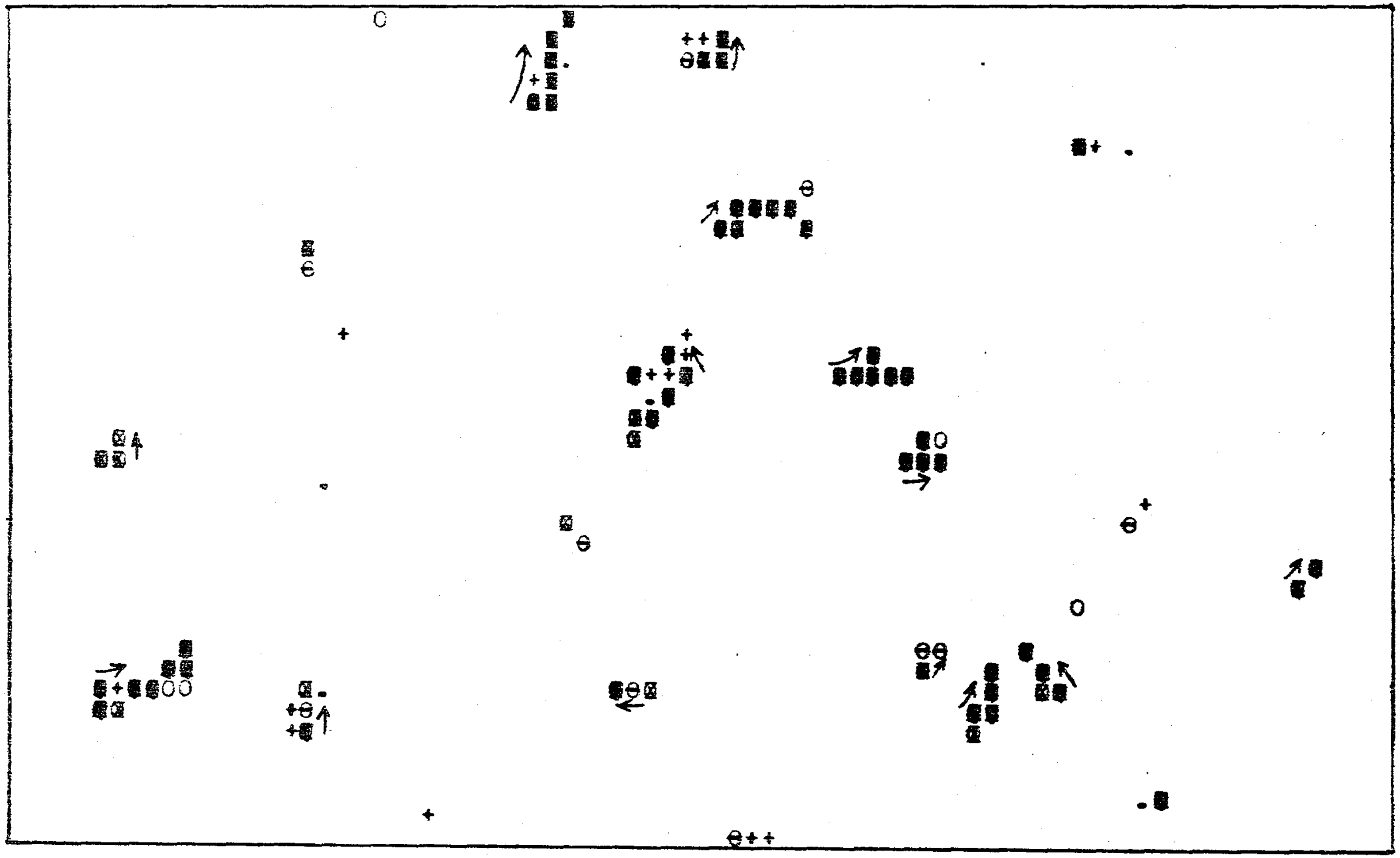


FIGURE 9.25
Central Nupur Glacier Balance Simulation (2° temperature drop + 50% precipitation increase)

CHAPTER X

CONCLUSIONS AND SUGGESTIONS FOR FURTHER STUDY

This study has been concerned with an examination of the manner in which local topographical and climatological factors interact to control glacier balance. An examination of the factors affecting local glacier distribution and a review of glacier-climate studies, enabled the identification of the significant interactions between topographic, climate and glaciologic factors (Fig. 3.11). This was followed with the development of a deterministic simulation model which, with given inputs of local topography and monthly climate, locates sites of positive mass balance and cirque glacier generation in mountainous ranges of temperate areas.

Space was represented in the mass balance model by use of a square-grid approach in which a regular array of interpolated altitudes serves as the basic input. Topographic parameters in the form of point altitude, gradient, aspect, profile and plan convexity values were generated for each grid intersection of the altitude matrix by means of a terrain-analysis program. The correspondence between calculated and measured values of these topographic parameters was shown to be close (sec. 5.4). The accurate determination of these topographic parameters is important in view of their critical role in controlling mass balance processes.

Following the definition of the topographic input, point values of local climatic variables (air temperature, snowfall, wind speed and direction) were estimated at each grid point from postulated relations between regional climate and local topographic parameters. The paucity of existing climatic data (which are often unrepresentative and sometimes completely lacking) led to the use of rather crude estimations and trends from what data that were available. These empirically-derived relations need further examination. In particular there is a need to set up field studies to examine the variation of local climatic variables over a range of different altitudes and slopes of differing aspects and gradients. The findings of these field investigations should then be combined with theoretical deductive analyses to produce more general models of local climatic variability.

The use of a strictly physical approach for modelling certain mass balance processes was also prevented by a lack of quantitative understanding about these processes. For example, the complexity of the aerodynamic processes controlling snow drifting and avalanching defies the development of a generalised physical-based mathematical model. Instead it was found necessary to adopt a more statistical approach, based on empirical relationships, which incorporated a number of assumptions regarding the manner in which snow is transported during wind drifting and avalanching. Further work should be aimed at testing the validity of these assumptions by comparing the simulated results with those obtained by detailed and continuous monitoring of snow transport in mountainous regions. In this respect, recent work on the measurement of wind speeds over mountain crests and the amount of drift-snow deposited in lee slopes (e.g. Föhn, 1980) should prove useful for understanding the processes governing the rate and spatial distribution of drift accumulation and avalanche formation.

Comprehensive modelling of glacial material melting was also made difficult by the fact that most of the reported energy balance studies over melting snow and ice surfaces are of short duration and refer to limited sections of isolated glaciers. There is a paucity of information regarding the effect of large-scale energy balance processes, such as the horizontal advection of heat from ice-free areas. Clearly, "if the thorough process of glacier melt is to be fully grasped, glaciologists must examine not only the glaciers themselves, but the surrounding areas as well" (Ohmura, 1971 p.37).

In the process of developing a model framework for examining mass balance variations, then, this simulation exercise has also highlighted some of the areas in which further refinement and testing of model routines with more comprehensive and detailed data is necessary. For the present, some parts of the mass balance model (e.g. the computation of solar radiation receipt) inevitably are more detailed than others which are less amenable to mathematical treatment (e.g. snow drifting and avalanching).

The mass balance model was developed and tested with reference to two areas in northern Iceland, viz. Nupur and Thvera. However, given the process-based approach, it should be suitable for the simulation of local glacier balance in any 'temperate' mountainous region. While the detailed verification of the various assumptions and simplifications used in the model has been constrained by the lack of suitable field data, the limited validation procedures that were possible, show the simulated mass balance results to compare fairly closely with the observed distribution of snowpatches and sites of local glacier generation in the Icelandic study areas (sec. 9.2).

The simulated snow cover distribution in the Nupur area was found to be sensitive to changes in the snow drift routine, especially regarding the thickness of snow necessary to cover the minor roughness on the plateau remnants. Loss of solar radiation receipt by topographic shading was found to be an important factor in the Thvera area. In both areas, the success of the modelling approach reveals the important role of topographic parameters in determining spatial variations of local glacier mass balance. However, there is a further need to examine the relevance (for local glacier balance determination) of topographic parameters calculated from altitude matrices with differing grid intervals than the 100 m. used in this study.

The mass balance model has been restricted to simulating the initial sites of local glacier generation. To simulate complete glacier forms would require the incorporation of routines for determining ice flow and glacier erosion as a function of glacier geometry, mass balance, thermal regime and the underlying rock type. Furthermore, with the modelling of large ice masses, the problem of how far the glacier modifies its own meteorological environment will need to be taken into account.

The application of the mass balance model was illustrated by means of a number of simulation experiments in which the effect of changing climatic inputs on glacier balance was examined (sec. 9.3). These simulation experiments reveal the glaciological sensitivity of marginally glacierised areas (such as northern Iceland) to small changes in climatic conditions. Such a simulation exercise should prove to be of value to the glacial geomorphologist concerned with assessing the effects of particular climatic conditions on past, present, or future patterns of local mountain glacierisation. Thus climatic interpretations based on indirect evidence (e.g. cirque floor distributions) can be put on a sounder footing by simulating the actual process of local glacier development.

The mass balance model also permits the relative effects of the various mass balance processes in different areas of glacierisation to be examined. A final possible application of the mass balance model is in contributing to the long-term planning of water resources, by enabling the effect of any postulated climatic trend on glacier mass balance to be analysed; and for the appropriate corrective action (e.g. by slowing down glacier melting, through albedo modifications, in the event of an expected increase in air temperature) to be taken.

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APPENDIX A

LISTING OF PROGRAM GSP1

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C   GSP1
C   PROGRAM TO SIMULATE MONTHLY GLACIER MASS BALANCE (SNOW,FIRN,ICE) AT
C   EACH GRID INTERSECTION OF AN ALTITUDE MATRIX(100 M. GRID SPACING).
C   GSP1 (FOR A 40*80 RECTANGULAR MATRIX).
C
C   JASBIR S. GILL (DURHAM) 1979
C
C   PROGRAM GSP1 COMPRISED OF MAIN ROUTINE + 19 SUBROUTINES
C     SUBROUTINES : MTEMP          MAVAL
C                   MSNOW          AVDIR
C                   MWIND          MELTPR
C                   WEXP           SWRAD
C                   TPGW           GLSR
C                   MDAVAL         SHADE
C                   PLAV           LWRAD
C                   MDRIFT         TBHEAT
C                   DEP            FLOW
C                   DRWIND
C
C*********************************************************************
C
C   MAIN ROUTINE
C     THE MAIN ROUTINE PROVIDES OVERALL CONTROL OF THE COMPUTATIONAL
C   PROCEDURES SIMULATING GLACIER BALANCE.
C
C   COMMON/DIMENSION BLOCKS PROVIDED FOR THE FOLLOWING-
C
C     TOPOGRAPHY          SNOW/ICE BALANCE        CLIMATE (MONTHLY)
C     ----------          ----------------        -----------------
C
C     ALTITUDE (ALT)      WATER EQUIVALENT     INITIAL SNOW COVER (SNOW)
C     ASPECT (ASP)        ----------------     -AFTER DRIFTING (SNOW1)
C     SLOPE (SLOP)          SNOW (WSNOW)       -AFTER AVALANCHING (SNOW2)
C  PROFILE CONVEXITY        FIRN (WFIRN)          TEMPERATURE (TEMP)
C       (PROFC)             ICE (WICE)            WIND (WIND) (WRED)
C    PLAN CONVEXITY         TOTAL (TWEQ)
C       (PLANC)          TOTAL DEPTH (TDEPTH)        ENERGY BALANCE
C                                                    --------------
C     POTENTIAL MELT          DENSITY             GLOBAL RAD. (GRBL)
C     --------------          -------             LONG WAVE RAD. (XLWR)
C        PERIODS            SNOW (DSNOW)         TOTAL RAD. BALANCE (TRBL)
C        -------            FIRN (DFIRN)          SENSIBLE HEAT (SBHT)
C        (DMHRS)            ICE (DICE)            LATENT HEAT (TLHT)
C        (TEMPML)                              NET ENERGY BALANCE (ENGBL)
C        (DAYSML)             FLOW
C                             ----
C                      THICKNESS OF ICE (THIC)
C                 CRITICAL THICKNESS FOR FLOW (CRIT)
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C
C 'CO' DENOTES AN OPTIONAL PROCEDURE.
C
C  THE FOLLOWING LOGICAL I/O UNITS ARE USED IN THIS PROGRAM-
C  '1'  TO READ IN TOPOGRAPHIC INPUT (IN FILE TPG1).
C  '5'  TO READ IN PERIOD OF SIMULATION (FROM TERMINAL OR CARDS).
C  '2'  TO READ IN INITIAL ACCUMULATION COVER DATA (IN FILE RESO).
C  '3'  TO PRINT MAP SHOWING POINTS WITH SNOW/FIRN/ICE (IN FILE SMAP).
C  '6'  TO WRITE MESSAGES PROVIDING A CHECK ON THE SIMULATION RUN (ON
C       TERMINAL OR PRINTED).
C  '7'  TO WRITE VALUES OF MASS BALANCE PARAMETERS (AT SPECIFIEDED
C       INTERVALS) FOR MAPPING (IN FILE MBMAP).
C  '4'  TO WRITE NET MASS BALANCE VALUES AFTER THE END OF A SIMULATION
C       RUN (IN FILE RES12).
C
C
      COMMON ALT(40,80),ASP(40,80),SLOP(40,80),PROFC(40,80),PLANC(40,80)
     1,WSNOW(40,80),WFIRN(40,80),WICE(40,80)
      DIMENSION DSNOW(40,80),DFIRN(40,80),DICE(40,80)
      DIMENSION TDEPTH(40,80),TWEQ(40,80)
      DIMENSION SNOW(40,80),TEMP(40,80),SNOW1(40,80),SNOW2(40,80)
      DIMENSION WIND(40,80),WRED(40,80)
      DIMENSION GRBL(40,80),XLWR(40,80),TRBL(40,80)
      DIMENSION SBHT(40,80),TLHT(40,80),ENGBL(40,80)
      DIMENSION DMHRS(40,80),TEMPML(40,80),DAYSML(40,80)
      DIMENSION CRIT(40,80),THIC(40,80)
C
CO  DIMENSION FOR MAP SHOWING POINTS WITH SNOW/FIRN/ICE AFTER EVERY
CO  MONTH'S SIMULATION (OPTIONAL).
CO    DATA K0/1H /,K1/1H+/,K2/1H*/,K3/1H./
CO    DIMENSION KOR(80)
C
C  READ IN NUMBER OF MONTH IN WHICH SIMULATION IS TO BE STARTED (A1) AND
C  ENDED (A2).
      READ(5,100)A1,A2
  100 FORMAT(2F3.0)
C
C  READ IN INPUT VALUES OF W.E.,DENSITY, AND DEPTH TO DEFINE INITIAL
C  ACCUMULATION COVER (USING PROGRAM SIMINP1).
C  NOTE: X IS THE ACTUAL CALENDAR MONTH (1-12) IN WHICH SIMULATION IS
C    STARTED.
      READ(2) X
      READ(2) ((DSNOW(I,J),I=1,40),J=1,80)
      READ(2) ((DFIRN(I,J),I=1,40),J=1,80)
      READ(2) ((DICE(I,J),I=1,40),J=1,80)
      READ(2) ((WSNOW(I,J),I=1,40),J=1,80)
      READ(2) ((WFIRN(I,J),I=1,40),J=1,80)
      READ(2) ((WICE(I,J),I=1,40),J=1,80)
      READ(2) ((TDEPTH(I,J),I=1,40),J=1,80)
      READ(2) ((TWEQ(I,J),I=1,40),J=1,80)
C
C  READ IN BASIC INPUT TOPOGRAPHICAL DATA FROM TERRAIN ANALYSIS PROGRAM.
      READ(1) ((ALT(I,J),I=1,40),J=1,80)
      READ(1) ((ASP(I,J),I=1,40),J=1,80)
      READ(1) ((SLOP(I,J),I=1,40),J=1,80)
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      READ(1) ((PROFC(I,J),I=1,40),J=1,80)
      READ(1) ((PLANC(I,J),I=1,40),J=1,80)
C
C  DETERMINE "EXPOSURE" AT EACH POINT FOR EFFECT ON WIND SPEED
C  CALCULATIONS USING SUBROUTINE WEXP (RESULTS IN MATRIX WRED).
      CALL WEXP(WRED)
C
C  DETERMINE "CRITICAL THICKNESS" OF ICE AT WHICH FLOW WILL BEGIN.
C  MATRIX CRIT(I,J)
      DO 10 I=1,40
      DO 10 J=1,80
      CRIT(I,J)=0.0
      SL=SLOP(I,J)*1745.3E-5
      IF(SL.EQ.0.0)GO TO 11
      GO TO 12
   11 CRIT(I,J)=999999.0
      GO TO 10
   12 CRIT(I,J)=1.06E6/(0.85*980.665*SIN(SL))
   10 CONTINUE
C
C  A IS THE COUNTER FOR THE NUMBER OF MONTHS THAT HAVE BEEN SIMULATED.
C  X DENOTES THE PARTICULAR CALENDAR MONTH BEING SIMULATED.
C  SIMULATION RUN IS STARTED IN OCTOBER. (AT THE BEGINNING OF THE
C  ACCUMULATION SEASON).
      A=A1
    2 X=X+1.0
      A=A+1.0
      X=AMOD(X-1.0,12.0)+1.0
C
CO TO KEEP A CHECK OF THE PARTICULAR MONTH FOR WHICH SIMULATION IS IN
CO PROGRESS WRITE OUT THE SIMULATION MONTH.
      WRITE(6,1000)A,X
 1000 FORMAT(1X,16HSIMULATION MONTH,1X,F4.0,1X,F4.0)
C
CO OPTIONAL WRITE STATEMENTS INSERTED AFTER EACH MAJOR SECTION OF THE
CO SIMULATION HAS BEEN COMPLETED IN ORDER TO KEEP TRACK OF THE
CC COMPUTATIONS.
CO    WRITE(6,1001)
 1001 FORMAT(1X,61HSYSTEM INITIALISED,INPUT DATA READ,BEGIN CLIMATE COMP
     1UTATIONS)
C
C  CALCULATE MONTHLY TEMP AT EACH POINT USING SUBROUTINE MTEMP.
      CALL MTEMP(TEMP,X)
CO    WRITE(6,1002)
 1002 FORMAT(1X,27HTEMP CALCULATIONS COMPLETED)
C
C  CALCULATE MONTHLY SNOW AT EACH POINT USING SUBROUTINE MSNOW.
      CALL MSNOW(TEMP,SNOW,X)
CO    WRITE(6,1003)
 1003 FORMAT(1X,27HSNOW CALCULATIONS COMPLETED)
C
C  CALCULATE MONTHLY WIND SPEED AT EACH POINT USING SUBROUTINE MWIND.
      CALL MWIND(X,WIND,WRED)
CO    WRITE(6,1004)
 1004 FORMAT(1X,27HWIND CALCULATIONS COMPLETED)
```

```
C
C  CALCULATE AVALANCHING ON TO PLATEAU PRIOR TO DRIFTING USING
C  SUBROUTINE MDAVAL.
      CALL MDAVAL(X,SNOW,TDEPTH)
CO    WRITE(6,1005)
 1005 FORMAT(1X,36HAVALANCHING ON TO PLATEAU DETERMINED)
C
C  CALCULATE MONTHLY DRIFT AT EACH POINT USING SUBROUTINE MDRIFT.
      CALL MDRIFT(TEMP,SNOW,SNOW1,TDEPTH)
CO    WRITE(6,1006)
 1006 FORMAT(1X,28HDRIFT CALCULATIONS COMPLETED)
C
C  CALCULATE MONTHLY AVALANCHING AT EACH POINT USING SUBROUTINE MAVAL.
      CALL MAVAL(SNOW1,SNOW2,TDEPTH)
CO    WRITE(6,1007)
 1007 FORMAT(1X,34HAVALANCHING CALCULATIONS COMPLETED)
C
C
C  TO FIND W.E.(CM) OF CURRENT MONTHS SNOW, TAKING ACCOUNT OF SETTLING,
C  DRIFTING, AND AVALANCHING ON SNOW DENSITY.
C  DENSITY ASSUMPTIONS FOR ACCUMULATION STATE AS FOLLOWS :
C  P2(AVAL SNOW) 0.40, P(SETTLED SNOW) 0.25, P1(DRIFTED SNOW) 0.30
C  SNOW1(I,J) GIVES THE W.E. AT EACH POINT.
C  SNOW(I,J) GIVES THE DENSITY AT EACH POINT WEIGHTED FOR W.E.
C  FIRST FIND THE PROPORTIONS OF SNOWFALL,DRIFT,AND AVALANCHING AT EACH
C  POINT.
      DO 13 I=1,40
      DO 13 J=1,80
      P2=0.0
      P1=0.0
      P=0.0
      IF(SNOW2(I,J).EQ.0.0.AND.SNOW1(I,J).EQ.0.0.AND.SNOW(I,J).EQ.0.0)
     1GO TO 14
      IF(SNOW2(I,J).EQ.SNOW1(I,J).AND.SNOW1(I,J).EQ.SNOW(I,J)) GO TO 15
      IF(SNOW2(I,J).EQ.SNOW1(I,J).AND.SNOW1(I,J).LT.SNOW(I,J).AND.SNOW1(
     1I,J).EQ.(30.0-TDEPTH(I,J))) GO TO 15
      IF(SNOW2(I,J).EQ.SNOW1(I,J).AND.SNOW1(I,J).GT.SNOW(I,J)) GO TO 16
      IF(SNOW2(I,J).GT.SNOW1(I,J).AND.SNOW1(I,J).GT.SNOW(I,J)) GO TO 17
      IF(SNOW2(I,J).GT.SNOW1(I,J).AND.SNOW1(I,J).EQ.SNOW(I,J)) GO TO 18
      IF(SNOW2(I,J).LT.SNOW1(I,J).AND.SNOW2(I,J).EQ.(10.0-TDEPTH(I,J)))
     1GO TO 15
      IF(SNOW1(I,J).EQ.(30.0-TDEPTH(I,J)).AND.SNOW2(I,J).GT.SNOW1(I,J).A
     1ND.SNOW1(I,J).LT.SNOW(I,J)) GO TO 15
C  IN CASE NONE OF THESE CONDITIONS APPLY WRITE AN ERROR MESSAGE.
      WRITE(6,200)I,J
  200 FORMAT(1X,39HSOMETHING WRONG IN CURRENT SNOW DENSITY,2I4)
      GO TO 13
   14 SNOW1(I,J)=0.0
      SNOW(I,J)=0.0
      GO TO 13
   15 P=SNOW2(I,J)
      GO TO 19
   16 P=SNOW(I,J)
      P1=SNOW1(I,J)-SNOW(I,J)
      GO TO 19
```

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   17 P=SNCW(I,J)
      P1=SNOW1(I,J)-SNOW(I,J)
      P2=SNOW2(I,J)-SNOW1(I,J)
      GO TC 19
   18 P=SNCW(I,J)
      P2=SNOW2(I,J)-SNOW1(I,J)
C  NCW FIND THE AV. W.E. AND DENSITY
   19 SNOW1(I,J)=P2*0.40+P*0.25+P1*0.30
      SNOW(I,J)=SNOW1(I,J)/(P2+P+P1)
   13 CONTINUE
CO    WRITE(6,1008)
 1008 FORMAT(1X,46HCURRENT MONTH'S DENSITY CALCULATIONS COMPLETED)
C
C  TO FIND BALANCE OF SNCW, FIRN, AND ICE AT END OF EACH MONTH, WITH
C  ADJUSTMENT FOR CHANGES IN DENSITY AND CONSEQUENT CHANGES IN STATE OF
C  GLACIAL MATERIAL.
C  SNOW (DEN <0.58), FIRN (DEN 0.58-0.85), ICE (DEN >0.85).
C  DED IS THE MONTHLY INCREASE IN SNOW AND FIRN, TAKEN AS -
C    0.02(SNOW) 0.01(FIRN) GM/CM3 DURING THE WINTER MONTHS (TEMP<0) AND
C    0.05(SNOW) 0.04(FIRN) GM/CM3 DURING THE SUMMER MONTHS.
      DO 20 I=1,40
      DO 20 J=1,80
      DED=0.01
      IF(TEMP(I,J).GT.0.0) DED=0.04
      IF((DFIRN(I,J)+DED).LT.0.85) GC TO 21
C  CONVERSION OF FIRN INTO ICE . DICE WEIGHTED FOR W.E.
      DICE(I,J)=((DFIRN(I,J)+DED)*WFIRN(I,J)+DICE(I,J)*WICE(I,J))/(WICE
     1(I,J)+WFIRN(I,J))
      WICE(I,J)=WICE(I,J)+WFIRN(I,J)
      WFIRN(I,J)=0.0
      DFIRN(I,J)=0.0
   21 CONTINUE
C  DENSIFICATION OF FIRN.
      IF(DFIRN(I,J).EQ.0.0) GO TO 22
      DFIRN(I,J)=DFIRN(I,J)+CED
   22 CED=0.02
      IF(TEMP(I,J).GE.0.0) DED=0.05
      IF((DSNOW(I,J)+DEC).LT.0.58) GC TO 23
C  CONVERSION OF SNOW INTC FIRN.
      DFIRN(I,J)=((DSNOW(I,J)+DED)*WSNCW(I,J)+DFIRN(I,J)*WFIRN(I,J))/(W
     1NOW(I,J)+WFIRN(I,J))
      WFIRN(I,J)=WFIRN(I,J)+WSNOW(I,J)
      WSNOW(I,J)=0.0
      DSNOW(I,J)=0.0
   23 CONTINUE
C  DENSIFICATION OF SNOW.
      IF(WSNOW(I,J)+SNOW1(I,J).EQ.0.0) GO TC 24
      DSNOW(I,J)=((DSNOW(I,J)+DED)*WSNCW(I,J)+SNCW(I,J)*SNOW1(I,J))/(WS
     1OW(I,J)+SNOW1(I,J))
      WSNOW(I,J)=WSNOW(I,J)+SNOW1(I,J)
      GO TC 20
   24 DSNOW(I,J)=0.0
      WSNOW(I,J)=0.0
   20 CONTINUE
CO    WRITE(6,1009)
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 1009 FORMAT(1X,34HSNOW/FIRN DENSIFICATION DETERMINED)
C
C   DETERMINATION OF THE POTENTIAL MELT PERIODS.
C   DMHRS(I,J),TEMPML(I,J),DAYSML(I,J)
      CALL MELTPR(X,TEMP,DMHRS,TEMPML,DAYSML)
CO    WRITE(6,1010)
 1010 FORMAT(1X,23HMELT PERIODS DETERMINED)
C
C   DETERMINATION OF ENERGY BALANCE (RAD+TURBULENT HEAT).
C   FIRST CALCULATE RAD BALANCE CONSISTING OF SWRAD AND LWRAD BALANCE.
      CALL SWRAD(X,GRBL,DMHRS,TEMPML,DAYSML)
CO    WRITE(6,1011)
 1011 FORMAT(1X,23HS.W. RADIATION COMPUTED)
C
      CALL LWRAD(X,XLWR,DMHRS,TEMPML,DAYSML)
CO    WRITE(6,1012)
 1012 FORMAT(1X,23HL.W. RADIATION COMPUTED)
C
C   SUM SWRAD BAL AND LWRAD BAL TO FIND TOTAL RADIATION BALANCE AT EACH
C   POINT (TRBL) IN LANGLETS/MONTH.
      DO 25 I=1,40
      DO 25 J=1,80
      TRBL(I,J)=0.0
      IF(GRBL(I,J).EQ.-1.0) GO TO 25
      TRBL(I,J)=GRBL(I,J)+XLWR(I,J)
      IF(TRBL(I,J).LT.0.0) TRBL(I,J)=0.0
   25 CONTINUE
C
C   CALCULATE TURBULENT HEAT BALANCE (SENSIBLE HEAT + LATENT HEAT)
      CALL TBHEAT(X,WIND,DMHRS,TEMPML,DAYSML,SBHT,TLHT)
CO    WRITE(6,1013)
 1013 FORMAT(1X,47HSENSIBLE AND LATENT HEAT CALCULATIONS COMPLETED)
C
C   CALCULATE NET ENERGY BALANCE (ENGBL) BY SUMMING TRBL WITH SENSIBLE
C   (SBHT) AND LATENT TLHT) HEAT BALANCES.
      DO 26 I=1,40
      DO 26 J=1,80
      ENGBL(I,J)=0.0
      IF(SBHT(I,J).EQ.-1.0) GO TO 26
      ENGBL(I,J)=TRBL(I,J)+SBHT(I,J)+TLHT(I,J)
   26 CONTINUE
C
C   ADJUSMENT FOR MONTHLY MELT IF ENGBL IS POSITIVE
      DO 27 I=1,40
      DO 27 J=1,80
      IF(WSNOW(I,J).EQ.0.0) GO TO 28
      T=WSNOW(I,J)
      WSNOW(I,J)=WSNOW(I,J)-ENGBL(I,J)/80.0
      IF(WSNOW(I,J).LT.0.0) GO TO 29
      GO TO 30
   29 WSNOW(I,J)=0.0
      DSNOW(I,J)=0.0
   30 ENGBL(I,J)=ENGBL(I,J)-T*80.0
      IF(ENGBL(I,J).LE.0.0) GO TO 27
   28 IF(WFIRN(I,J).EQ.0.0) GO TO 31
```

```
      T=WFIRN(I,J)
      WFIRN(I,J)=WFIRN(I,J)-ENGBL(I,J)/80.0
      IF(WFIRN(I,J).LT.0.0) GO TO 32
      GO TO 33
   32 WFIRN(I,J)=0.0
      DFIRN(I,J)=0.0
   33 ENGBL(I,J)=ENGBL(I,J)-T*80.0
      IF(ENGBL(I,J).LE.0.0) GO TO 27
   31 IF(WICE(I,J).EQ.0.0) GO TO 27
      WICE(I,J)=WICE(I,J)-ENGBL(I,J)/80.0
      IF(WICE(I,J).LT.0.0) GO TO 34
      GO TO 27
   34 WICE(I,J)=0.0
      DICE(I,J)=0.0
   27 CONTINUE
CO    WRITE(6,1014)
 1014 FORMAT(1X,50HMELT AND RESULTING MASS BALANCE CHANGES DETERMINED)
C
C  MAP SHOWING AT WHICH POINTS THERE IS SNOW(*):FIRN(+):ICE( ):NO DEP(.)
CO    WRITE(3,300) A,X
CO300 FORMAT(1X,16HSIMULATION MONTH,1X,F4.0,1X,F4.0)
CO    DO 35 I=1,40
CO    DO 36 J=1,80
CO    KOR(J)=K3
CO    IF(WSNOW(I,J).GT.0.0) KOR(J)=K2
CO    IF(WFIRN(I,J).GT.0.0) KOR(J)=K1
CO    IF(WICE(I,J).GT.0.0) KOR(J)=K0
CO 36 CONTINUE
CO 35 WRITE(3,400) (KOR(J),J=1,80)
CO400 FORMAT(1X,80A1)
C
C  CALL FLOW ROUTINE
C  AT YEARLY INTERVALS (AT END OF ABLATION SEASON) COMPUTE IF THE ICE
C  THICKNESS (THIC(I,J)) IS GREATER THAN 2*CRITICAL THICKNESS NEEDED
C  FOR FLOW (CRIT(I,J)) . IF SO THEN CALL FLOW ROUTINE.
      IF(AMOD(A,12.0).EQ.0.0) GO TO 37
      GO TO 38
   37 DO 39 I=1,40
      DO 39 J=1,80
      THIC(I,J)=0.0
      IF(DICE(I,J).EQ.0.0) GO TO 39
      THIC(I,J)=WICE(I,J)/DICE(I,J)
      IF(THIC(I,J).LE.2.0*CRIT(I,J)) GO TO 39
      CALL FLOW(I,J,CRIT,THIC,DSNOW,DFIRN,DICE)
   39 CONTINUE
CO    WRITE(6,1015)
 1015 FORMAT(1X,15HFLOW DETERMINED)
C
C  CALCULATE AND WRITE OUT TOTAL W.E. AND DEPTH (DETERMINED FROM W.E.
C  AND DENSITY ASSUMPTIONS) AT EACH POINT AFTER EACH MONTHS SIMULATION.
   38 DO 40 I=1,40
      DO 40 J=1,80
      TWEQ(I,J)=WSNOW(I,J)+WFIRN(I,J)+WICE(I,J)
      IF(DSNOW(I,J).EQ.0.0) GO TO 41
      XX=WSNOW(I,J)/DSNOW(I,J)
```

```
      GO TC 42
   41 XX=0.0
   42 IF(DFIRN(I,J).EQ.0.0) GO TO 43
      Y=WFIRN(I,J)/DFIRN(I,J)
      GO TC 44
   43 Y=0.0
   44 IF(DICE(I,J).EQ.0.0) GC TO 45
      Z=WICE(I,J)/DICE(I,J)
      GO TC 46
   45 Z=0.0
   46 TDEPTH(I,J)=XX+Y+Z
   40 CONTINUE
C
C   AT YEARLY (OR OTHER SPECIFIED) INTERVALS WRITE OUT NET MASS BALANCE
C   VALUES AT EACH POINT.
      IF(AMOD(A,12.0).NE.0.0) GO TC 49
      DO 47 I=1,40
      DO 47 J=1,80
      IF(TWEQ(I,J).GE.50.0) GO TC 48
      GO TC 47
   48 WRITE(7,500)I,J,A,X,TWEQ(I,J),WSNOW(I,J),WFIRN(I,J),WICE(I,J),
     1THIC(I,J),CRIT(I,J)
  500 FORMAT(2I4,2X,8F10.2)
   47 CONTINUE.
C
C   CHECK IF END OF SIMULATION PERIOD.
   49 IF(A.LT.A2) GO TO 2
C
C   WRITE THE VALUES OF X,W.E.,DENSITY,AND DEPTH IN FILE RES12,
C   THESE WILL FORM THE INPUT FOR THE NEXT SIMULATION RUN.
      WRITE(4) X
      WRITE(4) ((DSNOW(I,J),I=1,40),J=1,80)
      WRITE(4) ((DFIRN(I,J),I=1,40),J=1,80)
      WRITE(4) ((DICE(I,J),I=1,40),J=1,80)
      WRITE(4) ((WSNOW(I,J),I=1,40),J=1,80)
      WRITE(4) ((WFIRN(I,J),I=1,40),J=1,80)
      WRITE(4) ((WICE(I,J),I=1,40),J=1,80)
      WRITE(4) ((TDEPTH(I,J),I=1,40),J=1,80)
      WRITE(4) ((TWEQ(I,J),I=1,40),J=1,80)
      STOP
      END
C
C
C
C   TO OBTAIN MONTHLY VALLES CF CLIMATIC PARAMETERS  AT EACH GRID INT.
C MTEMP
C   SUBROUTINE FOR MONTHLY TEMP
      SUBROLTINE MTEMP(TEMP,X)
      DIMENSION TEMP(40,80),BASET(12),TLRT(12)
      COMMCN ALT(40,80),ASP(40,80),SLOP(40,80),PROFC(40,80),PLANC(40,80
      DATA BASET/-1.5, 1.6,-0.5,0.6,5.4,8.5,10.5,9.3,7.3,3.9,1.8,-0.1/
      DATA TLRT/0.53,0.53,0.61,0.62,0.70,0.68,0.61,0.65,0.63,0.58,0.56,
     10.54/
C   BASE TEMP (30-YEAR MONTHLY AVERAGE)
      NN=X
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```
      T=BASET(NN)
C  MONTHLY LAPSE RATE(DEG.C/100 M.)
      Y=TLRT(NN)
C  MONTHLY LAPSE RATE (DEG C/M)
      Y=Y/100.0
C  LINEAR VARIATION OF TEMP WITH ALT(BASE CONSTANT C)
      C=T+20.0*Y
C  DETERMINE TEMP AT ANY ALT USING ALT-TEMP RELATION
      DO 1 I=1,40
      DO 1 J=1,80
      TEMP(I,J)=0.0
    1 TEMP(I,J)=-Y*ALT(I,J)+C
C  CORRECTION FOR ASPECT WITH DIFFERENT SEASONS
      DO 2 I=1,40
      DO 3 J=1,80
      IF(SLOP(I,J).LT.20.0) GO TO 3
C  CORRECTION FOR SEASON (SON MAM JJA DJF)
      IF(X.GE.9.AND.X.LE.11) CORR=0.7
      IF(X.GE.3.AND.X.LE.5) CORR=0.5
      IF(X.GE.6.AND.X.LE.8) CORR=0.4
      IF(X.EQ.12) CORR=0.6
      IF(X.GE.1.AND.X.LE.2) CORR=0.6
C  CONSIDERATION OF ASPECT (E S W N)
      IF(ASP(I,J).GE.45.00.AND.ASP(I,J).LT.135.0) GO TO 4
      IF(ASP(I,J).GE.135.00.AND.ASP(I,J).LT.225.0) GO TO 5
      IF(ASP(I,J).GE.225.00.AND.ASP(I,J).LT.315.0) GO TO 6
      TEMP(I,J)=TEMP(I,J)-CORR
      GO TO 3
    4 TEMP(I,J)=TEMP(I,J)+0.1-CORR
      GO TO 3
    5 TEMP(I,J)=TEMP(I,J)-0.2+CORR
      GO TO 3
    6 TEMP(I,J)=TEMP(I,J)-0.3+CORR
    3 CONTINUE
    2 CONTINUE
      RETURN
      END
C
C
C MSNOW
C  SUBROUTINE FOR MONTHLY SNOW
      SUBROUTINE MSNOW(TEMP,SNOW,X)
      DIMENSION SNOW(40,80),TEMP(40,80),AP(40,80),CAP(40,80)
      DIMENSION PERPR(12)
      COMMON ALT(40,80),ASP(40,80),SLOP(40,80),PROFC(40,80),PLANC(40,80
      DATA PERPR/13.0,13.3,8.3,5.9,3.2,3.0,3.5,3.9,9.6,12.2,12.7,11.4/
C  ANN PREC AT EACH POINT USING PREC-ALT RELATIONSHIP
C  AP=ANNUAL PREC (30-YEAR AV) CAP=AP CORRECTED FOR ASPECT
      DO 1 I=1,40
      DO 1 J=1,80
      SNOW(I,J)=0.0
    1 AP(I,J)=1.32*ALT(I,J)+967.6
C  VARIATION OF PREC WITH ASPECT (NE E SE S SW W NW N)
      DO 2 I=1,40
      DO 3 J=1,80
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      IF(SLCP(I,J).LT.20.0) GO TO 4
      IF (ASP(I,J).GE.22.5.AND.ASP(I,J).LT.67.5) GO TO 5
      IF (ASP(I,J).GE.67.5.AND.ASP(I,J).LT.112.5) GO TO 6
      IF (ASP(I,J).GE.112.5.AND.ASP(I,J).LT.157.5) GO TO 7
      IF (ASP(I,J).GE.157.5.AND.ASP(I,J).LT.202.5) GO TO 8
      IF (ASP(I,J).GE.202.5.AND.ASP(I,J).LT.247.5) GO TO 9
      IF (ASP(I,J).GE.247.5.AND.ASP(I,J).LT.292.5) GO TO 10
      IF (ASP(I,J).GE.292.5.AND.ASP(I,J).LT.337.5) GO TO 11
      IF (ASP(I,J).GE.337.5.AND.ASP(I,J).LT.360.0) GO TO 12
      IF (ASP(I,J).GE.0.00.AND.ASP(I,J).LT.22.5) GO TO 12
    5 CALL TPGW(I,J,-1,1,KTP)
      IF(KTP.EQ.1) GO TO 4
      CAP(I,J)=AP(I,J)+AP(I,J)*0.089
      GO TO 3
    6 CALL TPGW(I,J,0,1,KTP)
      IF(KTP.EQ.1) GO TO 4
      CAP(I,J)=AP(I,J)+AP(I,J)*0.015
      GO TO 3
    7 CALL TPGW(I,J,1,1,KTP)
      IF(KTP.EQ.1) GO TO 4
      CAP(I,J)=AP(I,J)-AP(I,J)*0.011
      GO TO 3
    8 CALL TPGW(I,J,1,0,KTP)
      IF(KTP.EQ.1) GO TO 4
      CAP(I,J)=AP(I,J)-AP(I,J)*0.041
      GO TO 3
    9 CALL TPGW(I,J,1,-1,KTP)
      IF(KTP.EQ.1) GO TO 4
      CAP(I,J)=AP(I,J)+AP(I,J)*0.051
      GO TO 3
   10 CALL TPGW(I,J,0, 1,KTP)
      IF(KTP.EQ.1) GO TO 4
      CAP(I,J)=AP(I,J)-AP(I,J)*0.015
      GO TO 3
   11 CALL TPGW(I,J,-1,-1,KTP)
      IF(KTP.EQ.1) GO TO 4
      CAP(I,J)=AP(I,J)-AP(I,J)*0.029
      GO TO 3
   12 CALL TPGW(I,J,-1,0,KTP)
      IF(KTP.EQ.1) GO TO 4
      CAP(I,J)=AP(I,J)-AP(I,J)*0.059
      GO TO 3
    4 CAP(I,J)=AP(I,J)
    3 CONTINUE
    2 CONTINUE
C     MONTHLY PREC(AP) IN MM USING CURVE TO DETERMINE PERCENTAGE OF ANNUAL
C     PREC FALLING IN ANY GIVEN MONTH (P).
      NN=X
      P=PERPR(NN)
      DO 13 I=1,40
      DO 13 J=1,80
      AP(I,J)=0.0
      AP(I,J)=CAP(I,J)*P/100.0
   13 CONTINUE
C     PERCENTAGE OF MONTHLY SOLID PREC (PSP)
```

```
      DO 14 I=1,40
      DO 14 J=1,80
      PSP=33.246-5.565*TEMP(I,J)
C   MONTHLY SNOW USING RELATICN FCR SOLID PREC
      SNOW(I,J)=AP(I,J)*PSP/100.0
      IF(SNOW(I,J).LT.0.0) SNOW(I,J)=0.0
C MONTHLY SNOW(CM)
      SNOW(I,J)=SNOW(I,J)/10.0
   14 CONTINUE
      RETURN
      END
C
C
C MWIND
C   SUBROUTINE FOR MONTHLY WINDSPEEC
C    (FOR USE IN TURBULENT HEAT COMPUTATIONS)
      SUBROUTINE MWIND(X,WINC,WRED)
      DIMENSION WIND(40,80),BASEW(12),WRED(40,80)
      COMMON ALT(40,80),ASP(40,80),SLOP(40,80),PROFC(40,80),PLANC(40,80)
      DATA BASEW/3.3,3.3,2.9,2.4,2.4,2.2,1.7,2.0,2.4,2.6,2.7,3.1/
C   MONTHLY WIND SPEED AT BASE STATION (THORUSTADIR (ALT 20M)) M/S
      NN=X
      W=BASEW(NN)
C   DETERMINATION OF MONTHLY WIND SPEEC AT EACH POINT USING MEAN
C   AEROLOGICAL WIND SPEEC - ALTITUDE RELATICN (SLCPE=0.00142).
C   FIRST NEEC TO DETERMINE CONSTANT(C) OF THE LINEAR REL.  C WILL BE
C   DIFFERENT FOR EACH MONTH CEPENDING CN THE MONTHLY BASE WIND SPEED.
      C=W-(0.00142*20.0)
      DO 1 I=1,40
      DO 1 J=1,80
      WIND(I,J)=0.00142*(ALT(I,J))+C
    1 CONTINUE
C   NOW IN VALLEYS, CIRQUES, AND OTHER SHELTERED CONCAVITIES THE MEAN
C   WIND SPEED WILL BE RECLCEC.
C   PERCENTAGE REDUCTION GIVEN BY MATRIX WRED(I,J).
      DO 2 I=1,40
      DO 2 J=1,80
      WIND(I,J)=WIND(I,J)-WIND(I,J)*WRED(I,J)
    2 CONTINUE
      RETURN
      END
C
C
C WEXP
C   SUBROUTINE FOR MEAN WIND SPEED MOCIFICATION DUE TO 'EXPOSURE' AS
C   DETERMINEC BY SUBROUTINE TPGW.
C   MEAN WIND SPEED REDUCTION WEIGHTEC ACCCRDING TO FREQUENCY DIST. OF
C    WINDS FROM DIFFERENT DIRECTICNS   (N,NE,E,SE,S,SW,W,NW)
      SUBROUTINE WEXP(WRED)
      COMMON ALT(40,80),ASP(40,80),SLOP(40,80),PROFC(40,80),PLANC(40,80)
      DIMENSION WRED(40,80)
      DO 1 I=1,40
      DO 1 J=1,80
      WRED(I,J)=0.0
      CALL TPGW(I,J,-1,0,KTP)
```

```
      IF(KTP.EQ.1) WRED(I,J)=0.06
      CALL TPGW(I,J,-1,1,KTP)
      IF(KTP.EQ.1) WRED(I,J)=WRED(I,J)+0.23
      CALL TPGW(I,J,0,1,KTP)
      IF(KTP.EQ.1) WRED(I,J)=WRED(I,J)+0.16
      CALL TPGW(I,J,1,1,KTP)
      IF(KTP.EQ.1) WRED(I,J)=WRED(I,J)+0.12
      CALL TPGW(I,J,1,0,KTP)
      IF(KTP.EQ.1) WRED(I,J)=WRED(I,J)+0.10
      CALL TPGW(I,J,1, 1,KTP)
      IF(KTP.EQ.1) WRED(I,J)=WRED(I,J)+0.14
      CALL TPGW(I,J,0, 1,KTP)
      IF(KTP.EQ.1) WRED(I,J)=WRED(I,J)+0.08
      CALL TPGW(I,J,-1,-1,KTP)
      IF(KTP.EQ.1) WRED(I,J)=WRED(I,J)+0.11
      WRED(I,J)=WRED(I,J)*0.60
    1 CONTINUE
      RETURN
      END
C
C
C TPGW
C  SUBROUTINE FOR THE DETERMINATION CF 'EXPCSURE' DUE TJ SURROUNDING
C   TOPOGRAPHY.
      SUBROUTINE TPGW(II,JJ,ID,JD,KTP)
      COMMON ALT(40,80),ASP(40,80),SLOP(40,80),PROFC(40,80),PLANC(40,80
      KTP=0
      K=0
      IF=II
      JF=JJ
    1 K=K+1
      IF=IF+ID
      IF(IF.LT.1.OR.IF.GT.40) RETURN
      JF=JF+JD
      IF(JF.LT.1.OR.JF.GT.80) RETURN
C  FOR A GIVEN DIRECTION OF WIND FLOW CONSIDER EACH ADJACENT POINT (FOR
C  1 KM.) IN THAT DIRECTICN TO SEE IF ANY OF THE NEIGHBOURING POINTS
C  HAVE GREATER ALTITUDE THAN THE POINT IN QUESTICN.
      IF(ALT(IF,JF).GT.ALT(II,JJ)+100.0) GO TO 2
      IF(K.GT.10) RETURN
      GO TC 1
    2 KTP=1
      RETURN
      END
C
C
C MDAVAL
C  SUBROUTINE FOR AVALANCHING CN TC PLATEAU PRIOR TO DRIFT
C
C  FIRST NEED TO FIND OUT IF THERE ARE ANY PCINTS HIGHER THAN THE
C  PLATEAU WHICH SATISFY THE CONCITIONS FCR AVALANCHING. NEXT SEE IF TH
C  DEPOSITION AREA FOR THESE AVALANCHING POINTS LIES ON THE PLATEAJ.
C  IF IT DOES CHANGE SNOW CCVER - SNCW(I,J) PRIOR TO DRIFT ACCORDINGLY.
C
       SUBRCUTINE MDAVAL(X,SNCW,TDEPTH)
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      COMMON ALT(40,80),ASP(40,80),SLCP(40,80),PROFC(40,80),PLANC(40,80)
      DIMENSION SNOW(40,80),TDEPTH(40,80),DAVAL(40,80)
C   POINTS FROM WHICH AVAL WILL OCCUR GIVEN VALUE 0.0
C   POINTS ON PLATEAU WHERE AVAL SNCW MAY BE DEPOSITED GIVEN VALUE -2.0
C   POINTS NOT RELEVANT IN THIS CASE GIVEN VALUE -1.0
      DO 1 I=1,40
      DO 1 J=1,80
      DAVAL(I,J)=-1.0
      IF(ALT(I,J).GT.580.0.AND.SLOP(I,J).GE.45.0.AND.(TDEPTH(I,J)+SNOW(I
     1,J).GT.10.0)) DAVAL(I,J)=0.0
      IF(ALT(I,J).GT.580.C.AND.SLOP(I,J).GE.35.0.AND.(TDEPTH(I,J)+SNOW(I
     1,J).GT.30.0))DAVAL(I,J)=0.0
      IF(ALT(I,J).GT.580.0.AND.SLOP(I,J).GE.25.0.AND.(TDEPTH(I,J)+SNOW(I
     1,J).GT.50.0))DAVAL(I,J)=0.0
      IF(ALT(I,J).GE.580.0.AND.SLOP(I,J).LE.20.0) DAVAL(I,J)=-2.0
    1 CONTINUE
C   NOW CONSIDERING ONLY POINTS WHICH WILL AVALANCHE(0.0) FIND OUT IF ANY
C   POINT ON THE PLATEAU WILL RECEIVE SNOW AND CHANGE SNOW(I,J)
      OUTSN=0.0
      PER=0.80
      DO 2 I=1,40
      DO 2 J=1,80
      IF(DAVAL(I,J).LT.0.0) GO TO 2
      CALL PLAV(X,I,J,SNCW,DAVAL,PER,OUTSN,TDEPTH)
    2 CONTINUE
      RETURN
      END
C
C
C PLAV
C   SUBROUTINE PLAV: USED TO FIND OUT THE DIRECTION IN WHICH AVALANCHING
C   IS GOING TO OCCUR. FOR EACH AVALANCHING POINT THE SUCCESSIVE POINT OF
C   AVALANCHING IS FOUND UNTIL A PCINT SATISFYING CONDITICNS OF
C   DEPOSITION OR THE LIMITS CF AREA ARE REACHED. FROM THE ORIGINAL POINT
C   SNOW AVALANCHED IS GIVEN BY (PER). THE MATRIX SNOW IS CHANGED
C   TO TAKE ACCOUNT OF AVALANCHED SNOW RECEIVED AT THE PLATEAU.
C   OUTSN IS THE VARIABLE ACCUMULATING THE AMOUNT OF SNOW AVALANCHED OUT
C   THE AREA.
C
      SUBROUTINE PLAV(X,II,JJ,SNCW,DAVAL,PER,CUTSN,TDEPTH)
      COMMON ALT(40,80),ASP(40,80),SLOP(40,80),PROFC(40,80),PLANC(40,80)
      DIMENSION SNOW(40,80),CAVAL(40,80),AVSNOW(40,80)
      DIMENSION TDEPTH(40,80)
      DO 1 I=1,40
      DO 1 J=1,80
      AVSNOW(I,J)=0.0
    1 CONTINUE
      IB=II
      JB=JJ
    2 CONTINUE
C   CONDITICN IF LIMIT CF AREA REACHED
      IF(IB.LT.1.OR.IB.GT.40.OR.JB.LT.1.CR.JB.GT.80) GO TO 3
C   CONDITICN IF DEPOSITION PCINT REACHED
      IF(DAVAL(IB,JB).EQ.-2.0) GO TO 4
C   IF LEVEL BELOW PLATEAU REACHEC IGNORE POINT
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      IF(ALT(IB,JB).LT.560.0) GO TC 5
C  EACH ASPECT IS CONSIDERED IN TURN IN THE FOLLOWING ORDER
C      NE,E,SE,S,SW,W,NW,N.
      IF(ASP(IB,JB).GE.22.5.AND.ASP(IB,JB).LT.67.5) GO TO 6
      IF(ASP(IB,JB).GE.67.5.AND.ASP(IB,JB).LT.112.5) GO TO 7
      IF(ASP(IB,JB).GE.112.5.AND.ASP(IB,JB).LT.157.5) GO TO 8
      IF(ASP(IB,JB).GE.157.5.AND.ASP(IB,JB).LT.202.5) GO TO 9
      IF(ASP(IB,JB).GE.202.5.AND.ASP(IB,JB).LT.247.5) GO TO 10
      IF(ASP(IB,JB).GE.247.5.AND.ASP(IB,JB).LT.292.5) GO TO 11
      IF(ASP(IB,JB).GE.292.5.AND.ASP(IB,JB).LT.337.5) GO TO 12
      IF(ASP(IB,JB).GE.337.5.AND.ASP(IB,JB).LT.360.0) GO TO 13
      IF(ASP(IB,JB).GE.0.0.AND.ASP(IB,JB).LT.22.5) GO TO 13
    6 IB=IB-1
      JB=JB+1
      GO TC 2
    7 IB=IB
      JB=JB+1
      GO TC 2
    8 IB=IB+1
      JB=JB+1
      GO TO 2
    9 IB=IB+1
      JB=JB
      GO TO 2
   10 IB=IB+1
      JB=JB-1
      GO TC 2
   11 IB=IB
      JB=JB-1
      GO TC 2
   12 IB=IB-1
      JB=JB-1
      GO TC 2
   13 IB=IB-1
      JB=JB
      GO TC 2
C   FIND OUT THE AMOUNT OF SNCW AVALANCHEC OUT CF AREA
    3 OUTSN=OUTSN+PER*(SNCW(II,JJ)-(10.0-TDEPTH(II,JJ)))
      SNOW(II,JJ)=10.0-TDEPTH(II,JJ)
      RETURN
C   FIND OUT AMOUNT OF SNOW AVALANCHED AT EACH POINT OF DEPOSITION
    4 AVSNOW(IB,JB)=AVSNCW(IB,JB)+PER*(SNOW(II,JJ)-(10.)-TDEPTH(II,JJ)))
      SNOW(II,JJ)=10.0-TCEPTH(II,JJ)
      SNOW(IB,JB)=AVSNOW(IB,JB)+SNCW(IB,JB)
CO    WRITE(4,100)X,II,JJ,IB,JB
  100 FORMAT(F4.1,2I4,10X,2I4)
    5 RETURN
      END
C
C
C MDRIFT
C  SUBROUTINE FOR SNOW CRIFT
      SUBROUTINE MDRIFT(TEMP,SNOW,SNCW1,TDEPTH)
      COMMON ALT(40,80),ASP(40,80),SLOP(40,80),PROFC(40,80),PLANC(40,80)
      DIMENSION TEMP(40,80),SNOW(40,80),DRIFT(40,80),SNOW1(40,80)
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      DIMENSION TDEPTH(40,80)
C
CO  DRIFT SITUATION MAP DIMENSIONS
CO    DATA K0/1H0/,K1/1H*/,K2/1H./
CO    DIMENSION KOR(80)
C
C   POINTS FROM WHICH DRIFT WILL OCCUR GIVEN VALUE(-1.0). ALL OTHER
C   POINTS GIVEN VALUE(-2.0) I.E. NOT IN QUESTION AS FAR AS DRIFT IS
C   CONCERNED.
      DO 1 I=1,40
      DO 1 J=1,80
      DRIFT(I,J)=-2.0
      IF(TEMP(I,J).LE.0.0.AND.ALT(I,J).GE.580.0.AND.SLOP(I,J).LE.20.0.AN
     1D.(TDEPTH(I,J)+SNOW(I,J)).GT.30.0) DRIFT(I,J)=-1.0
    1 CONTINUE
C   NOW TO DETERMINE POINTS TO WHICH DRIFT IS GOING TO TAKE PLACE,
C   CALL SUBROUTINE(DEP) FIRST IGNORING THE POINTS WITH VALUE(-1.0)
      DO 2 I=1,40
      DO 2 J=1,80
      IF(DRIFT(I,J).EQ.-1.0) GO TO 2
      CALL DEP(DRIFT,I,J)
    2 CONTINUE
C
CO  DRIFT SITUATION MAP TO SHOW POINTS FROM WHICH DRIFT IS TAKING
CO  PLACE (*), WHERE IT IS DRIFTING TO (0), AND POINTS NOT RELEVANT
CO  FOR DRIFT COMPUTATIONS (.).
CO    DO 3 I=1,40
CO    DO 4 J=1,80
CO    IF(DRIFT(I,J).EQ.-1.0) KOR(J)=K1
CO    IF(DRIFT(I,J).EQ.-2.0) KOR(J)=K2
CO    IF(DRIFT(I,J).EQ.0.0)  KOR(J)=K0
CO  4 CONTINUE
CO  3 WRITE(3,100) (KOR(J),J=1,80)
  100 FORMAT(1X,80A1)
C
C   NOW CONSIDERING EACH POINT WHICH IS ON THE PLATEAU EDGE  DRIFT(I,J)=0
C   COMPUTE THE AMOUNT OF SNOW WHICH WILL BE DRIFTED TO IT FROM THE
C   POINTS IN DIRECT LINE WITH IT IN 8 DIFFERENT DIRECTIONS(IN TURN).
C   ORDER IN WHICH WIND DIRECTION IS DEALT  N,NE,E,SE,S,SW,W,NW
      DO 5 I=1,40
      DO 5 J=1,80
      IF(DRIFT(I,J).EQ.-2.0) GO TO 5
      IF(DRIFT(I,J).EQ.-1.0) GO TO 5
C   SLOST IS THE PERC. OF SNOW ASSUMED LOST BY BEING BLOWN OFF PLATEAU
      SLOST=0.2
      CALL DRWIND(I,J,-1,0,DRIFT,0.03,SLOST,SNOW,TDEPTH)
      CALL DRWIND(I,J,-1,+1,DRIFT,0.28,SLOST,SNOW,TDEPTH)
      CALL DRWIND(I,J,0,+1,DRIFT,0.17,SLOST,SNOW,TDEPTH)
      CALL DRWIND(I,J,+1,+1,DRIFT,0.15,SLOST,SNOW,TDEPTH)
      CALL DRWIND(I,J,+1,0,DRIFT,0.09,SLOST,SNOW,TDEPTH)
      CALL DRWIND(I,J,+1,-1,DRIFT,0.15,SLOST,SNOW,TDEPTH)
      CALL DRWIND(I,J,0,-1,DRIFT,0.07,SLOST,SNOW,TDEPTH)
      CALL DRWIND(I,J,-1,-1,DRIFT,0.06,SLOST,SNOW,TDEPTH)
    5 CONTINUE
C
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CO   TO CHECK DRIFT COMPUTATICNS ANC CCNSIDERATION UF BOUNDARY EFFECTS.
CO   TOTAL SNOW DRIFTED OFF PLATEAU (TPL)
CO   AMOUNT OF DRIFT DEPOISTEC AT PLATEAU EDGE (TCR)
CO      TPL=0.0
CO      TDR=0.0
CO      DO 6 I=1,40
CO      DO 6 J=1,80
CO      IF(DRIFT(I,J).EQ.-2.0) GO TO 6
CO      IF(DRIFT(I,J).EQ.-1.0) GO TO 7
CO      TDR=TDR+DRIFT(I,J)
CO      GO TC 6
CC    7 TPL=TPL+(SNOW(I,J)-(30.0-TDEPTH(I,J)))*(1.0-SLOST)
CO    6 CONTINUE
CO   WRITE VALUES OF TPL,TCR
CO      WRITE(2,200) TPL,TDR
  200 FORMAT(1H ,2F10.2)
C
C    TO DETERMINE THE SNOW COVER AFTER DRIFTING AT EACH POINT (SNOW1(I,J))
        DO 8 I=1,40
        DO 8 J=1,80
        IF(DRIFT(I,J).EQ.-2.0) SNOW1(I,J)=SNOW(I,J)
        IF(DRIFT(I,J).EQ.-1.0) SNOW1(I,J)=30.0-TDEPTH(I,J)
        IF(DRIFT(I,J).GE.0.0) SNOW1(I,J)=SNOW(I,J)+DRIFT(I,J)
      8 CONTINUE
        RETURN
        END
C
C
C DEP
C   SUBROUTINE (DEP) TO DETERMINE PCINTS WHICH WILL RECEIVE DRIFT
C   FOR THIS TO BE THE CASE THESE PCINTS MUST HAVE A NEIGHBOURING POINT
C   WITH VALUE (-1.0). THESE DRIFT DEP. POINTS ARE GIVEN THE VALUE (0.0).
C
        SUBROUTINE DEP(DRIFT,I,J)
        DIMENSION DRIFT(40,80)
        II=I+1
        IF(II.GT.40) GO TC 1
        IF(DRIFT(II,J).EQ.-1.0) GO TC 2
        JJ=J+1
        IF(JJ.GT.80) GO TC 3
        IF(DRIFT(II,JJ).EQ.-1.0) GC TO 2
      3 JJ=J-1
        IF(JJ.LT.1) GO TO 1
        IF(DRIFT(II,JJ).EQ.-1.0) GC TO 2
      1 CONTINUE
        II=I-1
        IF(II.LT.1) GO TO 4
        IF(DRIFT(II,J).EQ.-1.0) GO TC 2
        JJ=J+1
        IF(JJ.GT.80) GO TO 5
        IF(DRIFT(II,JJ).EQ.-1.0) GO TO 2
      5 JJ=J-1
        IF(JJ.LT.1) GO TO 4
        IF(DRIFT(II,JJ).EQ.-1.0) GC TO 2
      4 CONTINUE
```

```
      JJ=J+1
      IF(JJ.GT.80) GO TC 6
      IF(DRIFT(I,JJ).EQ.-1.0) GO TC 2
    6 JJ=J-1
      IF(JJ.LT.1) RETURN
      IF(DRIFT(I,JJ).EQ.-1.0) GO TC 2
      RETURN
    2 DRIFT(I,J)=0.0
      RETURN
      END
C
C
C DRWIND
C   SUBROUTINE (DRWIND) CALCULATES AMOUNT OF SNOW DRIFTED TO EACH POINT
C   ON THE PLATEAU EDGE FRCM THE DIFFERENT DIRECTIONS. SLOST IS THE % OF
C   SNOW ASSUMED LOST BY BEING BLOWN OFF PLATEAU EDGE (20 % IN THIS CASE)
C   ID,JD,ARE THE INCREMENTS FOR I,J TO CHANGE WIND DIRECTION .
C   PER IS THE % OF SNOW DRIFTEC WITH A PARTICULAR WIND DIRECTION.
C   30.0 CM OF SNOW IS SUBTRACTED FROM EACH POINT AS THE INITIAL AMOUNT
C   NECESSARY BEFORE DRIFT CCCURS.
C
      SUBROUTINE DRWIND(II,JJ,ID,JD,DRIFT,PER,SLOST,SNOW,TDEPTH)
      COMMON ALT(40,80),ASP(40,80),SLOP(40,80),PROFC(40,80),PLANC(40,80)
      DIMENSION DRIFT(40,80),SNOW(40,80)
      DIMENSION TDEPTH(40,80)
      IF=II
      JF=JJ
    1 IF=IF+ID
C   IF LIMIT OF AREA REACHED RETURN
      IF(IF.LT.1.OR.IF.GT.40) RETURN
      JF=JF+JD
      IF(JF.LT.1.OR.JF.GT.80) RETURN
C   FOR A GIVEN DIRECTION THE ADJACENT POINT TO THE POINT RECEIVING
C   DRIFT (I.E. 0.0) IS SUCCESSIVELY CONSIDERED AND THE AMOUNT OF DRIFT
C   CONTINUALLY ADDED TO (I,J=0.0) UNTIL THERE IS NO SUCH NEIGHBOURING
C   POINT OR THE LIMIT OF THE AREA IS REACHEC.
      IF(DRIFT(IF,JF).NE.-1.0) RETURN
      DRIFT(II,JJ)=DRIFT(II,JJ)+(SNOW(IF,JF)-(30.0-TDEPTH(IF,JF)))
     1*(1.0-SLOST)*PER
      GO TC 1
      END
C
C
C MAVAL
C   SUBROUTINE FOR SNOW AVALANCHING
      SUBROUTINE MAVAL(SNOW1,SNOW2,TDEPTH)
      COMMON ALT(40,80),ASP(40,80),SLOP(40,80),PROFC(40,80),PLANC(40,80)
      DIMENSION SNOW1(40,80),SNOW2(40,80),AVAL(40,80)
      DIMENSION TDEPTH(40,80)
C
C   DIMENSICN FOR MAP OF AVALANCHES
CO    DATA K0/1H0/,K1/1H./,K2/1H*/
CO    DIMENSION KOR(80)
C   DEFINE MATRIX IN THE FCLLCWING MANNER
C   -2.0 FOR POINTS WHICH SATISFY THE CCNDITIONS FOR AVAL SNOW DEPOSITIO
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C   -1.0 FOR POINTS WHICH ARE NOT REVELANT FCR AVALANCHING
C    0.0 FOR POINTS FROM WHICH SNOW WILL BE AVALANCHED
C   SET PER TO THE % OF SNCW WHICH WILL BE AVALANCHED FROM POINTS (0.0)
C   PER INITIALLY SET TO 80%
        PER=0.8
        DO 1 I=1,40
        DO 1 J=1,80
        IF(SLOP(I,J).LE.15.0.AND.ALT(I,J).LE.580.0) GO TO 2
        IF(PROFC(I,J).LE.-2.0.AND.PROFC(I,J).GE.-30.0.AND.
       1PLANC(I,J).LE.-15.0.ANC.PLANC(I,J).GE.-100.0.AND.
       1ALT(I,J).LT.580.0) GO TO 2
        IF(SLOP(I,J).GE.45.0.AND.(TDEPTH(I,J)+SNOW1(I,J)).GT.10.0)
       1GO TC 3
        IF(SLOP(I,J).GE.35.0.AND.(TDEPTH(I,J)+SNOW1(I,J)).GT.30.0)
       1GO TC 3
        IF(SLOP(I,J).GE.25.0.AND.(TDEPTH(I,J)+SNOW1(I,J)).GT.50.0)
       1GO TO 3
        AVAL(I,J)=-1.0
        GO TC 1
      2 AVAL(I,J)=-2.0
        GO TC 1
      3 AVAL(I,J)=0.0
      1 CONTINUE
C
C   MAP SHOWING FROM WHICH PCINTS SNOW WILL BE AVALANCHED(*);DEPOSITED(J)
C   NOT AFFECTED(.).
CO      DO 4 I=1,40
CO      DO 5 J=1,80
CO      IF(AVAL(I,J).EQ.-1.0) KOR(J)=K1
CO      IF(AVAL(I,J).EQ.-2.0) KOR(J)=K0
CO      IF(AVAL(I,J).EQ.0.0)  KOR(J)=K2
CO    5 CONTINUE
CO    4 WRITE(3,100) (KOR(J),J=1,80)
  100 FORMAT(1X,80A1)
C
C   NCW CONSIDERING ONLY PCINTS WHICH WILL AVALANCHE (AVAL(I,J) EQ 0.0)
C   FIND OUT (USING SUBROUTINE AVDIR) HCW MUCH AND TO WHICH POINTS SNOW
C   WILL BE AVALANCHED (SNOW2) AND ALSO HOW MUCH SNOW IS AVALANCHED OUT
C   OF THE AREA (OUTSN).
C
        OUTSN=0.0
        DO 6 I=1,40
        DO 6 J=1,80
      6 SNOW2(I,J)=0.0
        DO 7 I=1,40
        DO 7 J=1,80
        IF(AVAL(I,J).LT.0.0) GC TO 7
        CALL AVDIR(I,J,SNOW1,SNOW2,AVAL,PER,OUTSN,TDEPTH)
      7 CONTINUE
C   DETERMINATION OF THE SNOW COVER AT EACH GRID-POINT AFTER DRIFTING AND
C   AVALANCHING (SNOW2)
        DO 8 I=1,40
        DO 8 J=1,80
        IF(AVAL(I,J).EQ.0.0) GC TO 9
        IF(AVAL(I,J).EQ.-1.0) GO TC 10
```

```
      IF(AVAL(I,J).EQ.-2.0) GO TC 11
      GO TO 8
    9 SNOW2(I,J)=10.0-TDEPTH(I,J)
      GO TC 8
   10 SNOW2(I,J)=SNOW1(I,J)
      GO TO 8
   11 SNOW2(I,J)=SNOW2(I,J)+SNOW1(I,J)
    8 CONTINUE
C
C  TO CHECK THAT AMOUNTS OF SNOW AVALANCHED ARE BEING CORRECTLY COMPUTED
C  SUM UP ALL THE SNOW AVALANCHED IN VARIABLE (AMAV) AND ALL THE SNOW
C  DEPOSITED AFTER AVALANCHING (AMDP)
C  THEN AMAV SHOULD EQUAL AMDP+OUTSN
CO    AMAV=0.0
CO    AMDP=0.0
CO    DO 12 I=1,40
CO    DO 12 J=1,80
CO    IF(AVAL(I,J).EQ.-1.0) GO TC 12
CO    IF(AVAL(I,J).EQ.-2.0) AMDP=AMDP+(SNOW2(I,J)-SNOW1(I,J))
CO    IF(AVAL(I,J).EQ.0.0)AMAV=AMAV+PER*(SNOW1(I,J)-(10.0-TDEPTH(I,J)))
CO 12 CONTINUE
CC    WRITE(3,200) AMAV,OUTSN,AMDP
  200 FORMAT(1H ,3F10.2)
      RETURN
      END
C
C
C AVDIR
C  SUBROUTINE AVDIR: USED TO FIND THE DIRECTION IN WHICH AVALANCHING
C  IS GOING TO OCCUR. FOR EACH AVALANCHING POINT THE SUCCESSIVE POINT OF
C  AVALANCHING FOUND UNTIL A POINT SATISFYING CONDITIONS OF DEPOSITION
C  OR THE LIMITS OF THE AREA ARE REACHED. FROM ORIGINAL POINT THE % OF
C  SNOW AVALANCHED IS GIVEN BY (PER). A NEW MATRIX (SNOW2) ACCUMULATES
C  SNOW RECEIVED FROM AVALANCHING.
C  OUTSN IS THE VARIABLE ACCUMULATING THE AMOUNT OF SNOW AVALANCHED OUT
C  OF THE AREA.
C
      SUBROUTINE AVDIR(II,JJ,SNOW1,SNOW2,AVAL,PER,OUTSN,TDEPTH)
      COMMON ALT(40,80),ASP(40,80),SLOP(40,80),PROFC(40,80),PLANC(40,80)
      DIMENSION SNOW1(40,80),SNOW2(40,80),AVAL(40,80)
      DIMENSION TDEPTH(40,80)
      IB=II
      JB=JJ
    1 CONTINUE
C  CONDITION IF LIMIT OF AREA REACHED
      IF(IB.LT.1.OR.IB.GT.40.OR.JB.LT.1.OR.JB.GT.80) GO TO 2
C  CONDITION IF DEPOSITION POINT REACHED
      IF(AVAL(IB,JB).EQ.-2.0) GO TC 3
C  EACH ASPECT IS CONSIDERED IN TURN IN THE FOLLOWING ORDER
C     NE,E,SE,S,SW,W,NW,N.
      IF(ASP(IB,JB).GE.22.5.AND.ASP(IB,JB).LT.67.5) GO TO 4
      IF(ASP(IB,JB).GE.67.5.AND.ASP(IB,JB).LT.112.5) GO TO 5
      IF(ASP(IB,JB).GE.112.5.AND.ASP(IB,JB).LT.157.5) GO TO 6
      IF(ASP(IB,JB).GE.157.5.AND.ASP(IB,JB).LT.202.5) GO TO 7
      IF(ASP(IB,JB).GE.202.5.AND.ASP(IB,JB).LT.247.5) GO TO 8
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      IF(ASP(IB,JB).GE.247.5.AND.ASP(IB,JB).LT.292.5) GJ TO 9
      IF(ASP(IB,JB).GE.292.5.AND.ASP(IB,JB).LT.337.5) GJ TO 10
      IF(ASP(IB,JB).GE.337.5.AND.ASP(IB,JB).LT.360.0) GJ TO 11
      IF(ASP(IB,JB).GE.0.0.AND.ASP(IB,JB).LT.22.5) GO TJ 11
    4 IB=IB-1
      JB=JB+1
      GO TC 1
    5 IB=IB
      JB=JB+1
      GO TC 1
    6 IB=IB+1
      JB=JB+1
      GO TC 1
    7 IB=IB+1
      JB=JB
      GO TC 1
    8 IB=IB+1
      JB=JB-1
      GO TC 1
    9 IB=IB
      JB=JB-1
      GO TC 1
   10 IB=IB-1
      JB=JB-1
      GO TC 1
   11 IB=IB-1
      JB=JB
      GO TC 1
C  FIND OUT THE AMOUNT OF SNCW AVALANCHEC CUT CF AREA
    2 OUTSN=OUTSN+PER*(SNOW1(II,JJ)-(10.0-TDEPTH(II,JJ)))
      RETURN
C  FIND OUT AMOUNT OF SNCW AVALANCHEC AT EACH POINT OF DEPOSITION
    3 SNOW2(IB,JB)=SNOW2(IB,JB)+PER*(SNOW1(II,JJ)-(10.0-TDEPTH(II,JJ)))
      RETURN
      END
C
C
C  ABLATION OF GLACIER MATERIAL (THROUGH ENERGY BALANCE COMPUTATIJNS).
C
C    THE ENERGY BALANCE IS ONLY CCMPUTED FOR "MELT PERIJDS" WHEN THE
C  AIR TEMP(I,J) IS > 0 CEG.
C  PRIOR TO ENERGY BAL. CCMPUTATIONS NEED TC WORK OUT -
C  (1) NO. OF HRS. FOR WHICH TEMP IS >0.0 DURING THE DAY (DMHRS)
C  (2) THE MELT TEMP FOR THIS DAILY PERIOD WHEN TEMP IS >0.0 (TEMPML)
C  (3) NO. OF DAYS IN THE MONTH MEAN TEMP EXPECTED TO BE >0.0 (DAYSML)
C  THESE VARIABLES ARE CALCULATED BY THE SUBROUTINE MELTPR
C
C
C MELTPR
C  SUBROUTINE FOR WORKING OUT THE POTENTIAL MELT PERIODS
C
      SUBROUTINE MELTPR(X,TEMP,DMHRS,TEMPML,DAYSML)
      DIMENSION TEMP(40,80),DAYSML(40,80),DMHRS(40,80),TEMPML(40,30)
      DIMENSION DAYS(12),CTAMP(12)
      DATA DAYS/31.0,28.0,31.0,30.0,31.0,30.0,31.0,31.0,30.0,31.0,30.0,3
```

```
     11.0/
      CATA DTAMP/0.2,0.9,1.3,2.1,2.9,3.3,3.0,2.7,2.0,1.5,0.4,0.2/
C
C  MCNTHLY AMPLITUDE OF CIURNAL TEMPERATURE (AMP) DEG C.
      NN=X
      PI=3.14159
      AMP=DTAMP(NN)
      AMP=AMP/2.0
C
C  FIND THE NO. OF DAYS WITH FROST AND SO THE NO. OF POTENTIALMELTING
C  DAYS. DAYSML(I,J).
      DO 1 I=1,40
      DO 1 J=1,80
      DAYSML(I,J)=0.0
      DMHRS(I,J)=0.0
      TEMPML(I,J)=-1.0
      IF(TEMP(I,J).LE.-AMP) GO TC 1
      FDAYS=-2.42*TEMP(I,J)+21.5
      IF(FCAYS.LT.0.0) FCAYS=0.0
      IF(FCAYS.GT.DAYS(NN)) FDAYS=CAYS(NN)
      DAYSML(I,J)=DAYS(NN)-FDAYS
      IF(TEMP(I,J).GE.AMP)  GO TO 2
      GO TC 3 .
    2 DMHRS(I,J)=24.0
      TEMPML(I,J)=TEMP(I,J)
      GO TC 1
C
C  DETERMINE THE DAILY POTENTIAL MELTING PERIOD (HRS) · TIME FOR WHICH
C  MEAN TEMP IS >0.0 DEG C.  DMHRS(I,J)
C  FIRST FIND ACTUAL TIMES WHEN TEMP GCES ABOVE 0.0 DEG. (T1,T2)
C          THE MEAN TEMP DURING PERIODS WHEN POTENTIAL MELT MAY OCCUR
C  DETERMINED BY INTEGRATING THAT PART OF THE DIURNAL TEMP CURVE WHICH
C  LIES ABOVE 0.0 DEG.
C
    3 T1=ARCOS(TEMP(I,J)/AMP)
      T2=2.0*PI-T1
      TEMPML(I,J)=TEMP(I,J)+(AMP*SIN(T1)/(PI-T1))
      IF(TEMPML(I,J).EQ.0.0) GO TC 4
      GO TO 5
    4 DAYSML(I,J)=0.0
      TEMPML(I,J)=-1.0
      GO TC 1
    5 T1=((T1*12.0)/PI)+4.0
      T2=((T2*12.0)/PI)+4.0
      DMHRS(I,J)=T2-T1
    1 CONTINUE
      RETURN
      END
C
C
C SWRAD
C  SUBROUTINE SWRAD FOR CETERMING THE SHORT WAVE RAD BALANCE CONSISTING
C  OF DIRECT AND DIFFUSE RACIATICN CCMFONENTS.
C
      SUBROUTINE SWRAD(X,GRBL,DMHRS,TEMPML,DAYSML)
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      COMMCN ALT(40,80),ASP(40,80),SLOP(40,80),PROFC(40,80),PLANC(40,80)
     1,WSNOW(40,80),WFIRN(40,80),WICE(40,80)
C   DIMENSICN AND PROVIDE DATA FOR THE FOLLOWING:
C   GRBL - VARIABLE FOR CCMPUTED S.W.RAC BALANCE
C   DEC8,DEC24   CONTAINING THE DECLINATICN ANGLES (DEGREES) (+VE WHEN
C    SUN IS N OF EQUATOR -VE WHEN S) FOR THE 8TH AND 24TH OF EACH MONTH.
C    THE RAC BL BEING CONSIDEC FOR THESE TWO DAYS IN EVERY MONTH IN
C    ORDER TO REDUCE THE NUMBER OF COMPUTATICNS.
C   DURS8,DURS24 -CONTAINING THE DURATICN OF SUNLIGHT (IN MINUTES) OVER
C    WHICH THE BALANCE WILL BE COMPUTED.
C   TRS8,TRS24 -CONTAINING THE MEAN AZIMUTH-PATH TRANSMISSIVITY OF THE
C    ATMOSPHERE FOR THE 1ST AND 15TH OF EACH MONTH.
C   CLOUD - MEAN MONTHLY CLOUCINESS (IN TENTHS)
C   TEMP - MEAN MONTHLY TEMP
C
      DIMENSION GRBL(40,80),TEMP(40,80),DUR8(12),DUR24(12),DAYS(12),D
     1EC8(12),DEC24(12),CLOUD(12),TRS8(12),TRS24(12)
      DIMENSION DMHRS(40,80),TEMPML(40,80),DAYSML(40,80)
      DATA DEC8/-22.34,-15.23,-5.19,6.90,16.87,22.77,22.57,16.38,6.00,-5
     1.59, 16.34,-22.63/
      DATA DEC24/-19.38,-9.74,1.11,12.59,20.62,23.43,20.03,11.37,-0.14,-
     111.48,-20.21,-23.42/
      DATA DUR8/264.0,460.0,652.0,863.0,1073.0,1282.0,1261.0,1048.0,834
     1.0,634.0,424.0,245.0/
      DATA DUR24/359.0,570.0,761.0,974.0,1186.0,1319.0,1153.0,937.0,727.
     10,526.0,319.0,216.0/
      DATA TRS8/0.936,0.900,0.869,0.837,0.813,0.798,0.794,0.798,0.811,0.
     1831,0.868,0.935/
      DATA TRS24/0.918,0.882,0.852,0.823,0.804,0.795,0.795,0.804,0.821,0
     1.847,0.898,0.994/
      DATA CLOUD/0.86,0.86,0.85,0.80,0.71,0.69,0.74,0.73,0.80,0.83,0.88,
     10.89/
C
C   X DENOTES THE MONTH
      NN=X
C   XK IS THE EMPRICAL COEFF USED AS THE CLOUD CORRECTION FOR GLOBAL
C      RADIATION (VARIES WITH LATITUDE)
      XK=0.45
      XK=(1.0-XK)*CLOUD(NN)
C   START COMPUTATION FOR GLOBAL SOLAR RADIATICN (USING SUBROUTINE GLSR)
C   TO CALCULATE THE BALANCE OF DIRECT AND DIFFUSE SOLAR RAD. IGNORE
C   POINTS NOT NEEDED FOR ENERGY BALANCE (IE THOSE WITH TEMPML LT 0.0).
C   FIRST CONSIDER THE RADIATION BALANCE ON THE 8TH DAY OF EACH MONTH.
      DO 1 I=1,40
      DO 1 J=1,80
      GRBL(I,J)=-1.0
      IF(TEMPML(I,J).LE.0.0) GO TO 1
C   CHOOSE ALBEDO VALUE ACCORDING TO SURFACE MAT. ADJUST FOR CLOUDINESS.
C   IF NO SNOW/FIRN/ICE AT POINT DONT CARRY OUT ENERGY COMPUTATIONS
      AL=0.50+0.1*CLOUD(NN)
      R=0.10
      IF(WSNOW(I,J).GT.0.0) GO TO 2
      AL=0.40+0.1*CLOUD(NN)
      R=0.0917
      IF(WFIRN(I,J).GT.0.0) GO TO 2
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      AL=0.30+0.1*CLOUD(NN)
      R=0.0857
      IF(WICE(I,J).GT.0.0) GO TO 2
      GO TO 1
    2 CONTINUE
C   MEAN LATITUDE OF AREA IS 65.96
      XLAT=65.96
      H=ALT(I,J)
      AZ=ASP(I,J)
      GR=SLOP(I,J)
      DEC=DEC1(NN)
      DUR=DUR1(NN)
      TRS=TRS1(NN)
C   IF MELT PERIOD IS SHORTER THAN THAT FOR DAYLIGHT, COMPUTE RADIATION
C   BALANCE ONLY FOR MELT PERIOD GIVEN BY DMHRS(I,J).
      DMHR=DMHRS(I,J)*60.0
      IF(DUR.GT.DMHR) DUR=DMHR
      CALL GLSR(I,J,XLAT,DEC,TRS,H,AZ,GR,DUR,GL,AL,R,XK)
C
C   GL IS THE TOTAL GLOBAL RADIATION ABSORBED BY THE SURFACE (AFTER BEING
C   CORRECTED FOR CLOUD, ALBEDO AND ALTITUDE)
C   SUM GL FOR THE 1ST HALF OF THE MONTH (NO. OF MELT DAYS / 2) TO FIND
C   TOTAL GLOBAL RAD FOR THE FIRST PART OF THE MONTH.
      GRBL(I,J)=GL*(DAYSML(I,J)/2.0)
C NOW DO THE SAME COMPUTATIONS TO  FIND GLOBAL RAD FOR THE REST OF THE
C   MONTH USING  VALUE OBTAINED FOR THE 24TH OF EACH MONTH
      DEC=DEC15(NN)
      DUR=DUR15(NN)
      TRS=TRS15(NN)
      DMHR=DMHRS(I,J)*60.0
      IF(DUR.GT.DMHR) DUR=DMHR
      CALL GLSR(I,J,XLAT,DEC,TRS,H,AZ,GR,DUR,GL,AL,R,XK)
C
C   SUM THE TOTAL GLOBAL RD BL BOTH FOR THE 1ST AND 2ND PART OF MONTH.
      GRBL(I,J)=GRBL(I,J)+(DAYSML(I,J)/2.0)*GL
    1 CONTINUE
      RETURN
      END
C
C
C GLSR
C   SUBROUTINE GLSR IS USED TO CALCULATE DAILY (USING INCREMENTS OF 60
C   MINUTES) GLOBAL RADIATION (DIRECT +DIFFUSE) ABSORBED BY THE GLACIAL
C   MATERIAL ON A SLOPE OF GIVEN ASPECT AND GRADIENT.
C
      SUBROUTINE GLSR(I,J,F,FO,G,H,A,B,DUR,TQ,AL,R,XK)
C   SUBSCRIPT F=LAT,FO=DEC,G=TRS,A=ASP,B=SLOP,DUR=DURSUN,Q=DIRECT,SJ=DIFF
      TQ=0.0
C   CONVERT DEGREES TO RADIANS FOR CALCULATION
      F=F*1745.3E-5
      FO=FO*1745.3E-5
      A=A*1745.3E-5
      B=B*1745.3E-5
C   ESTABLISH SINES
      SF=SIN(F)
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      SFO=SIN(FO)
C  ESTABLISH COSINES
      CF=CCS(F)
      CFO=COS(FO)
C  SELECTICN OF SOLAR CCNSTANT (SOL)
      IF(FO.GT.0.411)GO TC 1
      IF(FC.LE.0.411)SOL=1.S4
      IF(FC.LT.0.341)SOL=1.97
      IF(FO.LT.0.205)SOL=1.SS
      IF(FO.LT.0.068)SOL=2.02
      IF(FO.LT.-0.068)SCL=2.04
      IF(FO.LT.-0.205)SOL=2.06
      IF(FO.LT.-0.341)SOL=2.07
      IF(FC.LT.-0.411)GO TO 1
C INITIATE VALUES
      Q=0.0
      SQ=0.0
      X=-CCS(A)*SIN(B)
      Y=SIN(B)*SIN(A)
      Z=COS(B)
      T1=(X*SF+Z*CF)*CFO
      T2=(-X*CF+Z*SF)*SFO
C SET HOUR ANGLE TO SUNRISE  I.E. DURS X0.25(CONVERSION TO HOUR ANGLE)
C  DIVEDED BY 2 AND CONVERTED TO RADIANS. HOUR ANGLE IS -VE BEFORE SOLAR
C  NOON ANC +VC AFTER NOCN
      W=-DUR*0.0021812
C INCREMENT HOUR ANGLE BY 60 MINUTES
    2 W=W+26179.8E-5
C DETERMINE COSINE OF SUNS ZENITH ANGLE AND TEST WHETHER SUN IS ABOVE
C  HORIZON
      QD=CFC*CF*COS(W)+SFC*SF
      IF(QD.LE.0.0) GO TC 3
C  DETERMINE COSINE OF ANGLE BETWEEN THE SOLAR BEAM AND THE NORMAL TO
C  THE SLOPE
      Q2=-Y*SIN(W)*CFO+T1*CCS(W)
      QT=Q2+T2
C  CHECK THAT SLOPE IS NOT IN SHADOW
C     IF(QT.LE.0.0) GO TC 3
C  ***********************************************************************
CO                     SUBROUTINE SHADE (OPTIONAL)
CO CHECK THAT POINT(I,J) IS NOT IN SHADOW DUE TO SURROUNDING TOPOGRAPHY
CO FIRST DETERMINE SUNS ZENITH ANGLE (SALT) NOTE: QD GIVES COS(SALT)
CO     IF(QC.GT.1.0) QD=1.0
CO     SALT=ARCOS(QD)
CO NOW DETERMINE SUNS AZIMUTH ANGLE (SAZI)
CO     AZ=(SF*COS(SALT)-SFC)/(CF*SIN(SALT))
CO     IF(AZ.GT.1.0) AZ=1.0
CO     SAZI=ARCOS(AZ)
CO     IF(W.LT.0.0) SAZI=-SAZI
CO     CALL SHADE(I,J,SALT,SAZI,KAP)
CO IF KAP IS RETURNED AS 1, POINT(I,J) IS IN SHADOW -SO SKIP COMPUTATION
CO OF DIRECT S.W. RADIATION
CO     IF(KAP.EQ.1) GO TO 4
CO     GO TC 5
CO   4 CQ=0.0
```

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C     ************************************************************************
C     DETERMINE OPTICAL AIR MASS(Q1),SECANT APPROXIMATION(0 TO 70 DEGREES)
    5 Q1=ABS(1.0/(QD+1.0E-12))
      IF(Q1.LT.2.9) GO TO 6
C     DETERMINE OPTICAL AIR MASS FOR 70 TO 90 DEGREES.(SMITHSONIAN TABLES).
      IF(Q1.GE.114.6) QQ=30.00
      IF(Q1.LT.114.6) QQ=26.96
      IF(Q1.LT.38.20) QQ=19.79
      IF(Q1.LT.22.93) QQ=15.36
      IF(Q1.LT.16.38) QQ=12.44
      IF(Q1.LT.12.74) QQ=10.39
      IF(Q1.LT.10.43) QQ= 8.90
      IF(Q1.LT. 8.84) QQ= 7.77
      IF(Q1.LT. 7.66) QQ= 6.88
      IF(Q1.LT. 6.76) QQ= 6.18
      IF(Q1.LT. 6.06) QQ= 5.60
      IF(Q1.LT. 5.49) QQ= 5.12
      IF(Q1.LT. 5.02) QQ= 4.72
      IF(Q1.LT. 4.62) QQ= 4.37
      IF(Q1.LT. 4.28) QQ= 4.07
      IF(Q1.LT. 3.99) QQ=3.82
      IF(Q1.LT. 3.74) QQ=3.59
      IF(Q1.LT. 3.52) QQ=3.39
      IF(Q1.LT. 3.33) QQ=3.21
      IF(Q1.LT. 3.15) QQ=3.05
      IF(Q1.LT. 2.99) QQ=2.90
      Q1=QQ
CO    IF(KAP.EQ.1) GO TO 7
C
C     DETERMINE SIXTY MINUTE VALUE OF DIRECT RAD (Q)
    6 CQ=60.0*SOL*(G**Q1)*QT
      IF(CQ.LT.0.0) CQ=0.0
C     DETERMINE SIXTY MINUTE VALUE OF DIFFUSE RAD (SQ)
    7 SQ=60.0*0.5*COS(B/2.0)*COS(B/2.0)*SOL*QD*(0.91-G**Q1)
C     SUM Q AND SQ FOR SIXTY MINUTE VALUE OF GLOBAL RADIATION (QQ)
      QQ=CQ+SQ
C     INCREASE OF SOLAR RADIATION WITH ALTITUDE
      QQ=QQ+(1.0E-4*QQ*H)
C     CORRECT THIS CLEAR SKY GLOBAL RADIATION FOR CLOUDINESS
      QQ=QQ*(1.0-XK)
C     CORRECT FOR ALBEDO (DETERMINED AT HOURLY INTERVALS USING RELATION
C     DEVELOPED WITH SUNS ALTITUDE) TO FIND ACTUAL RADIATION ABSORBED BY
C     SURFACE.
C     FIRST FIND SOLAR ALTITUDE (IN DEG.)
      ZA=ARCOS(QD)*57.2958
      ZA=90.0-ZA
C     NOW FIND ALBEDO
      C=-ALOG(R)/30.0
      ALB=((1-AL)*EXP(-C*ZA))+AL
C     NOW FIND ABSORBED GLOBAL RADIATION
      QQ=QQ*(1.0-ALB)
C     ADD TO TOTAL
      TQ=TQ+QQ
    3 CONTINUE
C     STOP COMPUTATION IF SUNSET REACHED
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      IF(W.LT.DUR*0.0021812) GO TC 2
      RETURN
C  WRITE OUT ERROR MESSAGE IF DECLINATION INCORRECT
    1 WRITE(6,100)
  100 FORMAT(53HOSELECTICN SKIPPED BECAUSE OF UNALLOWABLE DECLINATION)
      RETURN
      END
C
C
C SHADE
C  SHADE ROUTINE (OPTIONAL)
C  SUBROUTINE TO CHECK IF PCINT(I,J) WILL BE IN SHADOW DUE TO
C  SURROUNDING TOPOGRAPHY.
C  IF PARAMETER KAP IS RETURNED WITH A VALUE OF 1 , POINT IS IN SHADOW.
C  SEARCH MADE ALONG THE LINE OF SUNS AZIMUTH FOR AN ELEVATION GREAT
C  ENOUGH TO OBSCURE THE SUN. ELEVATIONS OBTAINED BY INTERPOLATION OF
C  GRID POINT ELEVATIONS AT THE INTERSECTIONS OF THE SUNS AZIMUTH WITH
C  THE GRID LINES.
C  SCREENING BY HORIZON CHECKED FOR A DISTANCE OF 1 KM
C
      SUBROUTINE SHADE(II,JJ,SALT,SAZI,KAP)
      LOGICAL KL1,KL2
      COMMON ALT(40,80),ASP(40,80),SLOP(40,80),PROFC(40,80),PLANC(40,80)
      KL1=.FALSE.
      KL2=.FALSE.
      H=100.0
      KAP=0
      SALT=1.5705-SALT
      DO 1 K=1,10
      IF(KL1) GO TO 2
      C=K
      IC=II+K
      IF(ABS(SAZI).GT.1.5705) IC=II-K
      IF(IC.LE.1.OR.IC.GE.40) GO TC 1
      A=C*TAN(SAZI)
      IF(SAZI.GE.1.5707963) A=ABS(A)
      IF(SAZI.LT.-1.5707963) A=-A
      IF(SAZI.GE.0.0) B=A+1.0
      IF(SAZI.LT.0.0) B=A-1.0
      JA=A
      JB=B
      IF((JJ+JA).GE.80.OR.(JJ+JB).GE.80) GO TO 3
      IF((JJ+JA).LE.1.OR.(JJ+JB).LE.1) GO TO 3
      IF((JJ-JA).LE.1.OR.(JJ-JB).LE.1) GC TC 3
      IF((JJ-JA).GE.80.OR.(JJ-JB).GE.80) GO TO 3
      IF(JA.LE.10.AND.JA.GE.-10.ANC.JB.LE.10.AND.JB.GE.-10) GO TO 4
    3 KL1=.TRUE.
      GO TO 2
C  EI ELEVATION AT INTERSECTION OF SUNS AZIMUTH WITH I'TH HORIZONTAL
C  LINE
    4 EI=(A-JA)*ALT(IC,JJ-JB)+(JB-A)*ALT(IC,JJ-JA)
      EI=ABS(EI)
C  DISTANCE FROM POINT(II,JJ) TO PCINT AT INTERSECTION OF I AXIS
      IF(COS(SAZI).EQ.0.0) GC TO 5
      GO TC 6
```

```
    5 DISTI=0.0
      GO TO 7
    6 DISTI=ABS(C/COS(SAZI))*H
C   IF EI > SUNS ALTITUDE POINT(II,JJ) IS IN SHADOW
    7 IF(EI-ALT(II,JJ).GT.DISTI*TAN(SALT)) GO TO 8
    2 IF(KL2) GO TO 1
      F=K
      JF=JJ-K
      IF(SAZI.LT.0.0) JF=JJ+K
      IF(JF.LE.1.OR.JF.GE.80) GO TO 1
      IF(TAN(SAZI).EQ.0.0) GO TO 9
      GO TO 10
    9 D=0.0
      GO TO 11
   10 D=F/TAN(SAZI)
   11 IF(SAZI.LT.0.0.AND.SAZI.GE.-1.5707963) D=ABS(D)
      IF(SAZI.LT.-1.5707963) D=-D
      IF(ABS(SAZI).LT.1.5705) E=D+1.0
      IF(ABS(SAZI).GE.1.5705) E=D-1.0
      ID=D
      IE=E
      IF((II+ID).GE.40.OR.(II+IE).GE.40) GO TO 12
      IF((II+ID).LE.1.OR.(II+IE).LE.1) GO TO 12
      IF((II-ID).LE.1.OR.(II-IE).LE.1) GO TO 12
      IF((II-ID).GE.40.OR.(II-IE).GE.40) GO TO 12
      IF(ID.LE.10.AND.ID.GE.-10.AND.IE.LE.10.AND.IE.GE.-10) GO TO 13
   12 KL2=.TRUE.
      GO TO 1
C   EJ  ELEVATION AT INTERSECTION OF SUNS AZIMUTH WITH J'TH VERTICAL LINE
   13 EJ=(D-ID)*ALT(II+IE,JF)+(IE-D)*ALT(II+ID,JF)
      EJ=ABS(EJ)
C   DISTANCE FROM POINT(II,JJ) TO POINT AT INTERSECTION OF J AXIS
      IF(SIN(SAZI).EQ.0.0) GO TO 14
      GO TO 15
   14 DISTJ=0.0
      GO TO 16
   15 DISTJ=ABS(F/SIN(SAZI))*H
C   IF EJ > SUNS ALTITUDE, POINT (II,JJ) IS IN SHADOW
   16 IF(EJ-ALT(II,JJ).GT.DISTJ*TAN(SALT)) GO TO 8
    1 CONTINUE
      RETURN
    8 KAP=1
      RETURN
      END
C
C
C LWRAD
C   SUBROUTINE FOR LONGWAVE RADIATION BALANCE
C     (CORRECTED FOR SLOPE AND CLOUD)
C
      SUBROUTINE LWRAD(X,XLWR,DMHRS,TEMPML,DAYSML)
      COMMON ALT(40,80),ASP(40,80),SLOP(40,80),PROFC(40,80),PLANC(40,80)
     1,WSNOW(40,80),WFIRN(40,80),WICE(40,80)
      DIMENSION XLWR(40,80),CLCUD(12)
      DIMENSION DMHRS(40,80),TEMPML(40,80),DAYSML(40,80)
```

```
      DATA CLOUD/0.86,0.86,0.85,0.80,0.71,0.69,0.74,0.78,0.80,0.88,0.88,
     10.89/
C  TEMP OF SNOW (TS) TAKEN AS 0.0 AT ALL TIMES (273 DEG KELVIN)
      TS=273.0
      NN=X
C  STEFAN BOLTZMAN CONSTANT IS 4.92E-9 LY/HR/KE-4
      SBCON=4.92E-9
C  EMPIRICAL CLOUD PARAMETER DETERMINING EFFECTIVE NET LW RAD UNDER
C  CLOUDY CONDITIONS TAKEN AS 1.40 (VARIES WITH CLOUD TYPE AND HEIGHT)
      CLK=1.40
C  COMPUTE DAILY LW RAD BALANCE,REMEMBERING TO CONVERT TEMP VALUES INTO
C  DEGREES KELVIN
      DO 1 I=1,40
      DO 1 J=1,80
      XLWR(I,J)=-1.0
      IF(TEMPML(I,J).LE.0.0) GO TO 1
C  DEFINE EMISSIVITY VALUE
C  IF NO SNOW/FIRN/ICE AT POINT DONT CARRY OUT ENERGY COMPUTATIONS
      EMIS=1.0
      IF(WSNOW(I,J).GT.0.0) GO TO 2
      IF(WFIRN(I,J).GT.0.0) GO TO 2
      IF(WICE(I,J).GT.0.0) GO TO 2
      GO TO 1
    2 CONTINUE
      XX=EMIS*SBCON
      YY=XX*(1.0-CLK*CLOUD(NN)*CLOUD(NN))
      T=TEMPML(I,J)+273.0
C  REMEMBER TO CONVERT SLOPE ANGLE TO RADIANS
      G=(SLOP(I,J)*1745.3E-5)/2.0
      XLWR(I,J)=YY*(9.35E-6*T**6*COS(G)**2-T**4)+(4*XX*T**3*(T-TS))
C  MULTIPLY XLWR BY NO. OF HOURS PER DAY TEMP EXPECTED TO BE > 0.0 DEG.
      XLWR(I,J)=XLWR(I,J)*DMHRS(I,J)
C  MULTIPLY XLWR BY NO. OF MELT DAYS IN MONTH TO FIND MONTHLY L.W. RAD
C  BALANCE.
      XLWR(I,J)=XLWR(I,J)*DAYSML(I,J)
    1 CONTINUE
      RETURN
      END
C
C
C TBHEAT
C  SUBROUTINE TBHEAT TO DETERMINE THE NET ENERGY EXCHANGE AT SNOW/ICE
C  SURFACE DUE TO THE PROCESSES OF SENSIBLE HEAT (SBHT) AND LATENT HEAT
C  (TLHT) EXCHANGE.
C
      SUBROUTINE TBHEAT(X,WIND,DMHRS,TEMPML,DAYSML,SBHT,TLHT)
      COMMON ALT(40,80),ASP(40,80),SLOP(40,80),PROFC(40,80),PLANC(40,80)
     1,WSNOW(40,80),WFIRN(40,80),WICE(40,80)
      DIMENSION WIND(40,80),PERCW(12),RLHUM(12),BASET(12),PRES(12)
      DIMENSION DMHRS(40,80),TEMPML(40,80),DAYSML(40,80)
      DIMENSION SBHT(40,80),TLHT(40,80)
C  PERCW IS THE % OF TIME FOR WHICH THERE IS NO WIND (CALM)
      DATA PERCW/0.13,0.14,0.17,0.17,0.15,0.15,0.18,0.19,0.17,0.16,0.13,
     10.13/
C  RLHUM IS THE MONTHLY MEAN RELATIVE HUMIDITY AT THE BASE STATION
```

```
      DATA RLHUM/78.0,78.0,76.0,79.0,79.0,80.0,82.0,83.0,80.0,80.0,79.0,
     178.0/
C   PRES IS THE MONTHLY MEAN ATMOSPHERIC PRESSURE AT THE BASE STATION
      DATA PRES/1000.4,1005.7,1008.0,1009.9,1015.4,1012.2,1009.9,1009.1,
     11006.1,1003.5,1002.5,999.9/
      DATA BASET/-1.5,-1.6,-0.5,0.8,5.4,8.5,10.5,9.3,7.3,3.9,1.8,-0.1/
C   X DENOTES THE MONTH
      NN=X
      BT=BASET(NN)
C   SPECIFIC HEAT OF DRY AIR AT CONSTANT PRESSURE (SH) CAL/GM/DEG
      SH=0.24
C   VON KARMANS CONSTANT (VK)
      VK=0.4
C   ESTIMATION OF HEIGHT AT WHICH MEASUREMENTS OF TEMP AND WIND WERE MADE
C   Z(2)CM
      Z2=200.0
C   TEMPERATURE OF SURFACE (TO) TAKEN AS 273.0 DEG K.
      TO=273.0
C   FIRST DETERMINE THOSE POINTS FOR WHICH THE MEAN MONTHLY TEMPML IS
C    >0.0, IGNORING THE REST(-1.0)
C   THEN TBHEAT CAN BE DETERMINED AT EACH POINT FROM THE PREVAILING
C    TEMP AND WIND CONDITIONS.
C   REMEMBER TO CONVERT TEMP(I,J) FROM DEG C. TO DEG K. AND WIND(I,J)
C    FROM M/S TO CM/HR.
C   ALSO FIND MEAN PRESSURE AND DENSITY OF AIR AT EACH POINT
      DO 1 I=1,40
      DO 1 J=1,80
      SBHT(I,J)=-1.0
      TLHT(I,J)=0.0
      IF(TEMPML(I,J).LE.0.0) GO TO 1
C   CHOOSE SURFACE ROUGHNESS FACTOR (ZO) ACCORDING TO SURFACE MATERIAL
C   IF NO SNOW/FIRN/ICE AT POINT DONT CARRY OUT ENERGY COMPUTATIONS
      ZO=0.4
      IF(WSNOW(I,J).GT.0.0) GO TO 2
      ZO=0.25
      IF(WFIRN(I,J).GT.0.0) GO TO 2
      ZO=0.1
      IF(WICE(I,J).GT.0.0) GO TO 2
      GO TO 1
    2 CONTINUE
C   CALCULATE PRESSURE (PR)
      PR=PRES(NN)*(1.0-(0.0065*ALT(I,J))/288.0)**5.256
C   TEMP IN DEG. KELVIN
      TK=TEMPML(I,J)+273.0
C   SATURATION VAPOUR PRESSURE (SVP) MB
      SVP=6.11*10.0**((7.5*BT)/(BT+237.3))
C   CALCULATION OF ACTUAL VAPOUR PRESSURE (VP) MB
      VP=(RLHUM(NN)*SVP)/100.0
C   VARIATION OF VP WITH ALTITUDE
      VPZ=VP*10.0**(-0.1587*(ALT(I,J)/1000.0))
C   CALCULATION OF VIRTUAL TEMP(VT)
      VT=TEMPML(I,J)/(1.0-0.379*(VPZ/PR))
C   AIR DENSITY (AD) GM/CM 3
      AD=3.4838E-4*(PR/(VT+273.0))
C   CONVERSION OF WIND SPEED TO CM/HR
```

```
      WNS=WIND(I,J)*3.6E05
C
C   CALCULATION OF EDDY DIFFUSIVITY (D) FOR TRANSFER OF SENSIBLE AND
C   LATENT HEAT.
      D=(VK**2*WNS)/(ALOG(Z2/Z0))**2
C   CORRECTION OF D FOR STABILITY (DC)
C   FIRST COMPUTE RICHARDSON NUMBER (RI)
C   ACC. DUE TO GRAVITY=1.271E10 CM/HR 2
      RI=(1.271E10*Z2*(TK-T0))/(TK*WNS**2)
      DC=D/(1.0+10.0*RI)
C
C   COMPUTATION OF SENSIBLE HEAT TRANSFER TO SNOW/ICE SURFACE (LY/HR)
      SBHT(I,J)=AD*SH*DC*TEMPML(I,J)
C
C   COMPUTATION OF LATENT HEAT EXCHANGE (+VE IF DIRECTED TOWARDS SURFACE
C   (CONDENSATION); -VE IF DIRECTED AWAY FROM SURFACE (EVAPORATION);
C   DEPENDING ON THE VALUE OF VAPOUR PRESSURE ABOVE THE SNOW/ICE SURFACE.
C   LATENT HEAT OF VAPORIZATION=597.3 CAL/GM
C   VAPOUR PRESSURE OF MELTING SNOW/ICE SURFACE TAKEN AS 6.11 MB
      TLHT(I,J)=597.3*AD*DC*(0.622/PR)*(VPZ-6.11)
C
C   NOW DETERMINE THE DAILY PERIOD OVER WHICH TO SUM SBHT AND TLHT
C   EXCHANGE. THIS WILL BE LIMITED BY NO. OF HRS FOR WHICH AIR TEMP
C   IS > 0.0 (GIVEN BY TEMPML) AND BY THE TIME FOR WHICH THERE IS WIND
C   BLOWING (GIVEN BY PERCW) , WHICHEVER IS THE SHORTER.
      TT=(1.0-PERCW(NN))*24.0
      IF(TT.LT.DMHRS(I,J)) DMHRS(I,J)=TT
      SBHT(I,J)=SBHT(I,J)*DMHRS(I,J)
      TLHT(I,J)=TLHT(I,J)*DMHRS(I,J)
C   NOW FIND THE MONTHLY EXCHANGE OF TURBULENT HEAT FLUXES BY MULTPLYING
C   THE DAILY TOTALS BY THE NO. OF DAYS IN THE MONTH MEAN TEMP IS
C   EXPECTED TO BE > 0.0 (DAYSML).
      SBHT(I,J)=SBHT(I,J)*DAYSML(I,J)
      TLHT(I,J)=TLHT(I,J)*DAYSML(I,J)
    1 CONTINUE
      RETURN
      END
C
C
C FLOW
C   SUBROUTINE FLOW
C   THIS IS CALLED WHEN THE ICE THICKNESS THIC(I,J) AT ANY POINT BECOMES
C   GREATER THAN 2*CRITICAL ICE THICKNESS (CRIT(I,J)) NECESSARY FOR FLOW
C   TO COMMENCE.
C   CONSIDERING THE POINT FROM WHICH ICE IS TO FLOW (II,JJ), THE ROUTINE
C   FIRST FINDS OUT THE DIRECTION OF FLOW (GIVEN BY THE ASPECT OF MAX.
C   SLOPE AT (II,JJ).
C   HALF THE TOTAL W.E. OF ICE,FIRN AND SNOW AT (II,JJ) IS THEN
C   TRANSFERRED TO THE ADJACENT POINT IN THE DIRECTION OF FLOW (IB,JB).
C   THE W.E. , DENSITY AND ICE THICKNESS AT (IB,JB) ARE RECALCULATED
C   TAKING INTO ACCOUNT THE AMOUNT OF ACCUMULATION TRANSFERRED AND THAT
C   (IF ANY) ALREADY ACCUMULATED THERE.
C   IF THE ICE THICKNESS AT POINT (IB,JB IS IN TURN FOUND TO EXCEED
C   2*CRIT (IB,JB) THEN HALF THE TOTAL W.E. IS TRANSFERRED TO THE NEXT
C   POINT IN THE FLOW DIRECTION AT (IB,JB). THIS PROCESS IS CONTINUED
```

```
C   UNTIL THE ICE THICKNESS AT THE GRID SQUARE RECEIVING ACCUMULATION
C   DOES NOT EXCEED 2*CRIT.
C
C
       SUBROUTINE FLOW(II,JJ,CRIT,THIC,DSNOW,DFIRN,DICE)
       COMMON ALT(40,80),ASP(40,80),SLOP(40,80),PROFC(40,80),PLANC(40,80)
      1,WSNOW(40,80),WFIRN(40,80),WICE(40,80)
       DIMENSION CRIT(40,80),THIC(40,80)
       DIMENSION DSNOW(40,80),DFIRN(40,80),DICE(40,80)
     9 IB=II
       JB=JJ
C   DIRECTION OF FLOW
C   EACH ASPECT IS CONSIDERED IN TURN IN THE FOLLOWING ORDER
C      NE,E,SE,S,SW,W,NW,N.
       IF(ASP(IB,JB).GE.22.5.AND.ASP(IB,JB).LT.67.5) GO TO 1
       IF(ASP(IB,JB).GE.67.5.AND.ASP(IB,JB).LT.112.5) GO TO 2
       IF(ASP(IB,JB).GE.112.5.AND.ASP(IB,JB).LT.157.5) GO TO 3
       IF(ASP(IB,JB).GE.157.5.AND.ASP(IB,JB).LT.202.5) GO TO 4
       IF(ASP(IB,JB).GE.202.5.AND.ASP(IB,JB).LT.247.5) GO TO 5
       IF(ASP(IB,JB).GE.247.5.AND.ASP(IB,JB).LT.292.5) GO TO 6
       IF(ASP(IB,JB).GE.292.5.AND.ASP(IB,JB).LT.337.5) GO TO 7
       IF(ASP(IB,JB).GE.337.5.AND.ASP(IB,JB).LT.360.0) GO TO 8
       IF(ASP(IB,JB).GE.0.0.AND.ASP(IB,JB).LT.22.5) GO TO 8
     1 IB=IB-1
       JB=JB+1
       GO TO 10
     2 IB=IB
       JB=JB+1
       GO TO 10
     3 IB=IB+1
       JB=JB+1
       GO TO 10
     4 IB=IB+1
       JB=JB
       GO TO 10
     5 IB=IB+1
       JB=JB-1
       GO TO 10
     6 IB=IB
       JB=JB-1
       GO TO 10
     7 IB=IB-1
       JB=JB-1
       GO TO 10
     8 IB=IB-1
       JB=JB
C   CONDITION IF LIMIT OF AREA REACHED
    10 IF(IB.LT.1.OR.IB.GT.40.OR.JB.LT.1.OR.JB.GT.80) GO TO 11
C   TRANSFER OF W.E.
       DICE(IB,JB)=(DICE(IB,JB)*WICE(IB,JB)+DICE(II,JJ)*(WICE(II,JJ)/2.0)
      1)/(WICE(IB,JB)+WICE(II,JJ)/2.0)
       WICE(IB,JB)=WICE(IB,JB)+WICE(II,JJ)/2.0
       IF((WFIRN(IB,JB)+WFIRN(II,JJ)).EQ.0.0) GO TO 12
       GO TO 13
    12 DFIRN(IB,JB)=0.0
```

APPENDIX 8 (CONT.)

(2) LISTING OF PROGRAM SIMINP1

```
C   SIMINP1
C   PROGRAM TO READ IN INITIAL INPUT VALUES OF MASS BALANCE VARIATION
C   INTO SIMULATION PROGRAM GSP1.
C   VALUES READ INTO FILE RESO.
C
C   X - THE CALENDAR MONTH FOR WHICH THE MASS BALANCE VALUES ARE PROVIDED
C       AT EACH GRID INTERSECTION.
C   DSNOW, DFIRN, DICE - DENSITY OF SNOW, FIRN, ICE.
C   WSNOW, WFIRN, WICE - WATER EQUIVALENT OF SNOW, FIRN, ICE.
C   TDEPTH, TWEQ - TOTAL DEPTH AND WATER EQIIVALENT OF ACCUMULATION.
C
C
C
      DIMENSION DSNOW(40,80),DFIRN(40,80),DICE(40,80)
      DIMENSION WSNOW(40,80),WFIRN(40,80),WICE(40,80)
      DIMENSION TDEPTH(40,80),TWEQ(40,80)
C   TO START SIMULATION IN SEPTEMBER (AT THE END OF THE ABLATION SEASON)
C   X NEEDS TO BE 8.0 (SINCE IN GSP1 X=X+1.0)
      X=9.0
      DO 1 I=1,40
      DO 1 J=1,80
      DSNOW(I,J)=0.0
      DFIRN(I,J)=0.0
      DICE(I,J)=0.0
      WSNOW(I,J)=0.0
      WFIRN(I,J)=0.0
      WICE(I,J)=0.0
      TDEPTH(I,J)=0.0
      TWEQ(I,J)=0.0
    1 CONTINUE
C   WRITE THESE INITIAL VALUES INTO INPUT FILE CURR1
      WRITE(2) X
      WRITE(2) ((DSNOW(I,J),I=1,40),J=1,80)
      WRITE(2) ((DFIRN(I,J),I=1,40),J=1,80)
      WRITE(2) ((DICE(I,J),I=1,40),J=1,80)
      WRITE(2) ((WSNOW(I,J),I=1,40),J=1,80)
      WRITE(2) ((WFIRN(I,J),I=1,40),J=1,80)
      WRITE(2) ((WICE(I,J),I=1,40),J=1,80)
      WRITE(2) ((TDEPTH(I,J),I=1,40),J=1,80)
      WRITE(2) ((TWEQ(I,J),I=1,40),J=1,80)
      STOP
      END
```

APPENDIX C

(1) Sample Input Into Terrain Analysis Program (OMY8)

1.	Data header (20A4)	NUPUR (N.W. ICELAND)
2.	Name of file containing altitude data	ALTNUP
3.	Name of file for detailed results	-RESULTS
4.	Name of file for calculated values	TPGNUP
5.	Name of file for limits	-LIMITS
6.	Grid size in metres (F6·3)	1ØØ Ø
7.	Scaling factor to convert altitude data to metres	1·Ø
8.	Number of columns in altitude matrix (I4)	ΔΔ82
9.	Format for altitude data	(1X, 20F4·Ø)

APPENDIX C (cont.)

(2) Sample Input Requirements for Line Printer Mapping Program (OYSM)

1.	Data header (20A4)	NUPUR (N.W. ICELAND)
2.	Subtitle (20A4)	PROFILE CONVEXITY (DEGREES/100 METRES)
3.	Name of File	TPGNUP
4.	Number of columns in matrix of data (I4)	ΔΔ8Ø
5.	Format of data	(2I4, 32X, F1Ø·2)
6.	Class boundaries for different densities (6F8·3)	ΔΔ-20ΔΔΔΔΔ-10ΔΔΔ ....ΔΔΔ 20

(2) <u>Thvera Matrix</u>

(showing distribution of glacierised areas)

(1) Nupur Matrix

(3) Central Nupur Matrix