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MAPPING POPULATION DENSITY IN CANADA, WITH PARTICULAR REFERENCE TO PLATE 48, ATLAS OF CANADA, 1957*

R. T. Gajda

ABSTRACT: In the compilation of Plate 48 of the *Atlas of Canada* an attempt was made to present a locationally accurate density-of-population-map, insofar as the map scale, the available time and the statistical data would permit. The problem, made difficult by the very uneven distribution of the population, was resolved by using a modified dasymetric-choropleth method. In this technique the basic data for a statistical division were accepted, but uninhabited areas were isolated, and specific densities for settled areas were located within each division. The inhabited areas were subdivided into tracts of different densities with homogeneous characteristics, by means of a detailed analysis of topographic maps on airphotos, and by interviews with persons familiar with particular localities. The result was an accurate picture of population distribution and density. Before a decision was made to use the above method a number of experiments were carried out using various methods and techniques for presenting population distribution and density, such as dot distributions, and the isopleth and choropleth methods.

RÉSUMÉ: On s'est efforcé dans la préparation de la planche 48 de l'Atlas du Canada de représenter avec précision la densité de la population. Pour ce faire, il a fallu tenir compte de l'échelle de la carte ainsi que des données statistiques et du temps disponibles. Au départ, la répartition inégale de la population au pays a créé certaines difficultés que l'on a réussi à surmonter en modifiant quelque peu la méthode *choroplèthe dasymétrique*. Bien que l'on ait accepté comme telles les données se rapportant aux divisions statistiques, on a dû établir une distinction entre les régions inhabitées et les densités de population à l'intérieur de chaque division, dans le cas des secteurs habités. L'examen détaillé des cartes topographiques et des photographies aériennes de même que les renseignements obtenus sur certaines localités particulières ont permis de classer les régions habitées en zones de densités différentes mais affichant des caractéristiques homogènes. Grâce à cette méthode, il a été possible de représenter avec concision la répartition et la densité de la population. Au préalable, on a mis à l'essai d'autres méthodes et techniques utilisées dans la représentation de ces phénomènes. Parmi celles-ci, on note la méthode par les points ainsi que les méthodes *isoplèthe* et *choroplèthe*.

INTRODUCTION

Many problems are inherent in the mapping of population density. The perfect population-density map would be a dot map, with one dot representing the location of each individual. Such a map, however, is cartographically impossible, except for a very small area at a very large scale. As a result, if dots are used to show population density, it is necessary to

*Based on a paper presented at the 9th Annual Meeting of the Canadian Association of Geographers, at Saskatoon, 1959. Original research was conducted by the writer; application of the cartographical method to the compilation of Plate 48 was directed by J. J. Lefebvre, head of the Mapping Section, Census Division, Dominion Bureau of Statistics, assisted by Miss P. Beland. Additional research on population density mapping was also carried out by the writer with the assistance of Miss D. Ostrowska after publication of the Atlas of Canada.

MS submitted May, 1961.

employ dots of varying values, thus intoducing a locational error where, for example, population groups are numerically smaller than the dot value.

The purpose of this paper is to explain fhe final method adopted for the compilation of Plate 48 of the *Atlas of Canada*, to outline some of the problems involved, and to present the results of experiments carried out using various methods and techniques for depicting population densities in Canada. In compiling Plate 48 an attempt was made to create a locationally accurate density-of-population-map, insofar as the available data and the scale of the map would permit. Although it is not claimed that the method used allowed the optimum possible accuracy to be obtained, nevertheless, it is felt that this plate is the best attempt yet published in Canada to give a synoptic picture of the population density in the settled part of the country.

THE PROBLEM OF MAPPING POPULATION DENSITY IN CANADA

Following census years, the Dominion Bureau of Statistics publishes highly generalized maps of population density on the basis of census divisions, and dot maps showing the distribution of population. Some of these maps, prepared by the choropleth method and based on large or medium sized administrative divisions, are not realistic in locational factors, and thus their practical use is limited for comparative studies, either geographic or economic.

Characteristics of Population Distribution

To realize the difficulties in mapping population density in Canada, it is essential to visualize the irregular distribution of the settlements. Fairly continuous settlement runs in a narrow band along the Canada-United States boundary, with occasional branches and islands radiating towards a sparsely populated or practically empty northland. Because of these contrasts and because of the semi-nomadic character of the northern population, many cartographers have avoided showing the density in the territories lying north of the 10 provinces, or they have assigned a meaningless density of less than one inhabitant per square mile to the entire area. Even in the 10 provinces the real density of population is difficult to portray because of the unevenness of its distribution. Briefly, the pattern of settlement in Canada is as follows. In the Maritimes and in Newfoundland, a nucleated type of settlement stretches along the coastline. The major centres of the Maritimes are separated from the Quebec centres by the sparsely populated highlands of the Gaspé Peninsula and the thinly settled south shores of the Gulf of St. Lawrence. The interior contains either very few inhabitants or is entirely uninhabited. Farther west, settlement extends along the International Boundary zone, but even here, in this narrow strip, there are great contrasts in occupancy. The sparsely peopled country of the Frontenac axis cuts into the well-settled St. Lawrence lowlands and the densely populated regions of southern Ontario north of the Great Lakes, the unsettled Canadian Shield separates southern Ontario from the concentration of population centred on Winnipeg. The settled belt in the Prairies widens and swings slightly to the north, avoiding semi-arid southern Saskatchewan. In the Rocky Mountain region, the settlement is limited to the more open valleys, while on the west coast it focuses on the densely populated southwest segment of the Vancouver-Victoria area. The general pattern portrays a strip-like distribution of population and a mosaic of densities located within 200 miles of the International Boundary. Even within relatively densely populated areas, the distribution is patchy and there is hardly any populated area that lies more than 50 miles from wild and almost empty land.

North beyond the zone of continuous settlement, occasional concentrations occur, such as the Lake St. John area, the Clay Belt, or the Peace River district. These settled areas are connected to the main zone by widely dispersed settlements along the roads and railways. Still farther north beyond the 55th parallel, the country is truly sparsely peopled. In this part of Canada, permanent settlement is limited mostly to small mining centres or service outposts. There are also large tracts of land visited periodically by a partly nomadic, indigenous, population often not associated with any of the larger settlements. It follows that this wide dispersion and diversification of the distribution of population in Canada creates a difficult problem in devising a satisfactory method for mapping population densities. This problem is connected directly also with the choice of the unit to which statistical data are applied.

The Problem of the Area Unit

A population-density map may represent either a rough approximation or a fairly accurate presentation of density according to the area unit chosen for detailed analytical work. The proper selection of the area unit has to be determined by the purpose of the map, the accuracy desired and by the available sources of information. For the purpose of compiling Plate 48, the smallest administrative units were used, and these were further subdivided into smaller tracts, particularly in sparsely inhabited areas, in order to obtain a more realistic picture of distribution of the population and, consequently, its density. The procedure followed is described below.

THE MODIFIED DASYMETRIC-CHOROPLETH METHOD

The above title is given to the method used in compiling Plate 48. It must be admitted that the title is rather unwieldy; nevertheless it indicates the genesis of the method. Interpreted, the term means 'a method of delineating areas of specific densities'. The word 'choropleth' (choro—area; plethron—measure) connotes value-representation by area, as opposed to lineal representation of values, as in the case of isopleths, or punctiform representation of value, as in the case of dots. 'Dasymetric' simply means 'measurement of density' and in the context may be interpreted as meaning 'determination of specific densities'. The term as applied to population-density mapping was apparently first used by Russian cartographers to indicate a method of choropleth mapping which attempted to get away from the inaccuracies introduced in mapping by simple statistical divisions.

To 1925, a total of 43 colored dasymetric maps had been prepared by the Research Division of the Soils and Subsoils Institute in Moscow at the scale of 10 versts to 1 inch or 1:420,000, and published under the direction of P. Semenov-Tien-Shansky. The area covered by these sheets includes the strips north and south of Moscow, a large portion of Ukraine and Byeloruthenia and several sheets of Siberia. The density of population in square versts is arranged according to the following classification: Marsh, sands, uninhabited, less than 10, 20, 30, 40, 50, 60, 70, 80, 90 ,100, 200, 500, 1,000 and more than 1,000. The dasymetric-choropleth method was also used successfully by J. K. Wright in his population-density maps of Cape Cod (Wright, 1950).

In the preparation of simple choropleth maps of population density the area of the statistical division for which population figures are available is used to derive the ratio of persons per unit area and the resultant density for the whole statistical division is then plotted. Tables can depict all the information given by a simple choropleth map. If density within the statistical division is heterogeneous, locational accuracy is lost because the simple choropleth method inevitably indicates a homogeneous density throughout the statistical division.

In applying the dasymetric-choropleth method, the basic data for the statistical division is used, but uninhabited areas and specific densities are located within the statistical division. If, for example, there are towns within the statistical division, and their populations are known, the town areas are delineated, measured, and then density-calculated. The town populations are then subtracted from the population figure for the whole statistical division, and the rural population density is thus derived. Large-scale topographical maps frequently indicate gradations in population density by the distribution of buildings plotted on the map. In such cases areas of homogeneous density can be judged and delineated. Information can also be obtained from airphotos, land-use maps, interviews, etc. The population assigned to any one delineated area must take into account the numbers for all other areas delineated within the statistical division, so that the total will tally with published figures for the whole statistical division. As these estimates are made on large-scale maps, inaccuracies in delineation are of little consequence after reduction, and in many cases the margin of error is absorbed in the density groupings that are necessary for cartographical representation.

In Canada there are a number of different types of statistical divisions for which population figures are published, or are otherwise available. The standard statistical series are for census divisions as delineated by the Dominion Bureau of Statistics. In provinces that have multi-tier systems of government, census divisions usually conform to counties; where counties are subdivided into municipal townships, or equivalent government areas, population figures are available for those subdivisions. In provinces which are not subdivided for purposes of government, census divisions are arbitrarily determined with varying relevance to demographic realities. The census is taken by enumeration areas, small areas which on the average can be covered by one enumerator in about three weeks. Enumeration areas aggregate into census divisions. Population statistics by enumeration areas are not published, but were made available for the compilation of Plate 48. Population figures are published for all urban centres incorporated and unincorporated.

In compiling Plate 48 it was known in advance that the publication scale would be 1:5,000,000. This scale determined that the smallest area that could be represented as a pattern on the map, taking draughting and printing limitations into account, would be about 42 square miles. The degree of necessary generalization implied by this scale made it unfeasible, considering that time available for compilation was limited, to compile for areas smaller than Ontario townships. In areas where there are no townships, enumeration areas were combined into units of suitable size. The chosen unit areas were transcribed onto maps of several scales varying from 3.95 to 13 miles to the inch, the selection of scale depending on the maps available for different parts of the country. The first compilations, derived from inspection of the largest scale topographical sheets available, were made on these sheets.

The first stage of compilation was to delimit uninhabited areas (Figure 1) and to subtract them from the over-all area of the chosen units. This, incidentally was the first departure from straight choropleth mapping by statistical divisions. An area was deemed uninhabited if it was considered to be unoccupied during at least part of the year. Areas regularly entered by a migrant population, such as logging camps, were considered occupied. Areas whose only known visitors were prospectors were considered uninhabited. Since far northern areas are not covered on Plate 48, the problems posed by areas periodically or occasionally visited by nomadic peoples, and by scientific stations with seasonally varying populations, did not have to be faced. Most areas in which there is either agriculture or some other significant economic activity are covered by large-scale topographical maps, so that the boundaries of uninhabited land, according to the criterion accepted, could thus be determined quite accurately. The work-areas of farms and ranches were considered inhabited, but no attempt was made to delineate small enclaves of unoccupied lands such as badlands, swamps, etc., within otherwise occupied land.

When uninhabited areas were isolated, the process of delineating areas estimated as having homogeneous densities was begun (Figure 2). The inhabited areas within each unit (smallest administrative unit or enumeration area, less uninhabited parts of this unit) were further subdivided into tracts of different densities with homogeneous characteristics. This work was done by detailed analysis of large-scale topographical maps coupled with the use of all geographical information with regard to the distribution of population within each unit. In addition, interviews were conducted with persons familiar with areas under consideration and corrections applied accordingly. In effect, redistribution of the population within the smallest units was made, and this resulted in a realistic picture of the population distribution and density (Figure 3). In order to provide an objective method for the selection of class intervals a frequency distribution graph was constructed by plotting density of population on the y axis and the number of divisional units on the x axis. On the basis of this graph, class intervals were established and a tentative characteristic density classification was chosen. These density groupings were then plotted on a fresh set of compilation sheets and the patterns transcribed to a plotting sheet at publication scale. As the number of units totalled over 11,000, it was thus possible to present a discrete pattern of population densities. Experience showed that on reduction to the publication scale many of these patterns were eliminated by generalization. It was, therefore, necessary to make allowances for the disappearance of many separate patterns, and still have the implied population of any particular census division tally with published figures. This generalization was carried out before the final transcription was made to publishing scale.

The map study that accompanied this work made it evident that many urban centres less than 42 square miles in area were too small to be represented on the map by a separate pattern. It was decided therefore to consider the population of urban centres of 5,000 population and less as rural population dispersed over an area of not more than 49 square miles. The procedures and techniques described above represented a departure from the straight dasymetric-choropleth method, and the term 'modified' was introduced into the title to account for this departure.

Appraisal of the Modified Dasymetric-choropleth Method

The various experiments conducted prior to the compilation of this density-of-population-map indicate that the technique has many merits, and is of value for geographical studies of population as it depicts a density pattern very close to reality. The greatest value of the map is that it gives an accurate representation of the sharp transitions that exist in Canada between uninhabited and inhabited areas, and between densely and thinly populated regions. It is in this respect that choropleth mapping by large statistical divisions that straddle regions of greatly different demographic characteristics is inadequate for mapping population densities in Canada.

However, there are also certain disadvantages:

(1) In outlining the smallest area units of population densities, certain subjective judgments have to be made. This demands, on the part of the compiler, a very good knowledge of the geography of the area in question.

(2) The estimation of certain values of the distribution of the population and consequently the densities are in proportion to the skill of the compiler.

(3) The technique is time-consuming, as area measurements of a great number of tracts of different densities are necessary.

(4) Finally, it is difficult to adjust this method to the representation of population densities in certain specific parts of Canada, particularly in northern regions. These regions constitute a specific cartographical problem as far as the mapping of population density is concerned.

EXPERIMENTS WITH OTHER TECHNIQUES

Previous to the final adoption of the modified dasymetric-choropleth method for compiling Plate 48, experiments were carried out with several other techniques for depicting population densities in Canada. Some of these were tested for the province of New Brunswick as an example; others were tried on maps of the whole country. Below are presented some selected experiments conducted over the area of the province of New Brunswick.

Dot Method

In considering this system, a study was made of a published map of Canada showing the distribution of population by the dot method. In addition, three maps of New Brunswick were prepared, using (a) dots representing 100 people distributed evenly over enumeration areas (Figure 4), (b) same dots distributed evenly over enumeration areas known to be inhabited (Figure 5), and (c) dots placed only in the most prominent concentrations (Figure 6).

The maps showed the distribution of the population but not the density. In spite of the fact that on large-scale maps dot maps can depict the distribution of population very close to the actual distribution, it was considered that this method was unsatisfactory for use in the *Atlas of Canada.*

Relative density, of course, may be deduced from the concentrations of dots but direct quantitative measurements cannot be made easily from the map.

Owing to the demographically diversified areas of Canada, it is difficult to determine the most suitable size of dots to be used or to give them quantitative values. If, for example, a dot is given too small a value, a very large number of dots are required; and this gives the impression of greater density than exists, and may also cause unwanted blending of dots. Again, if the dot is given too high a value, then a distorted picture of relative sparseness results, and the map may give an erroneous impression of density.

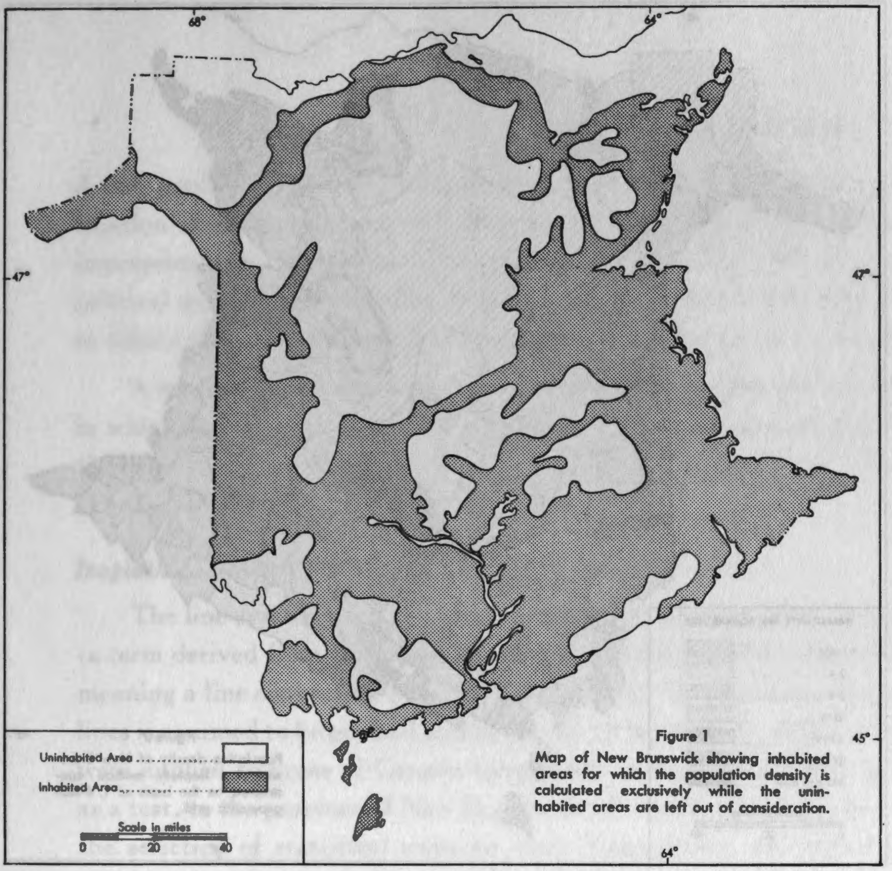

Figure 1
Map of New Brunswick showing inhabited areas for which the population density is calculated exclusively while the uninhabited areas are left out of consideration.

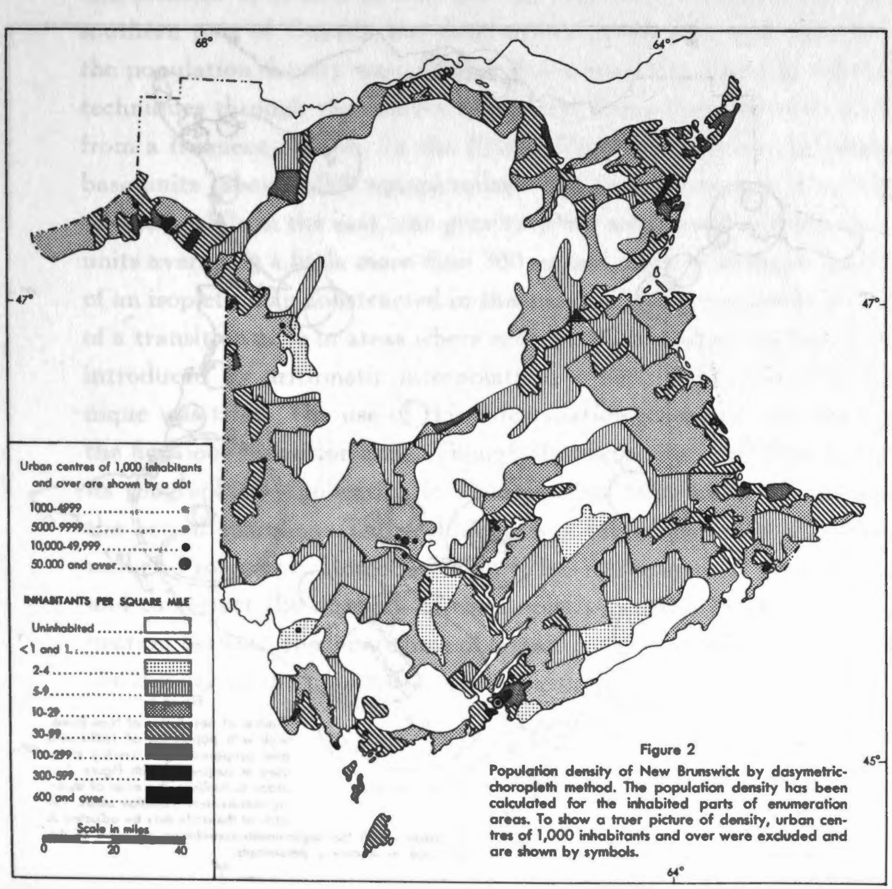

Figure 2
Population density of New Brunswick by dasymetric-choropleth method. The population density has been calculated for the inhabited parts of enumeration areas. To show a truer picture of density, urban centres of 1,000 inhabitants and over were excluded and are shown by symbols.

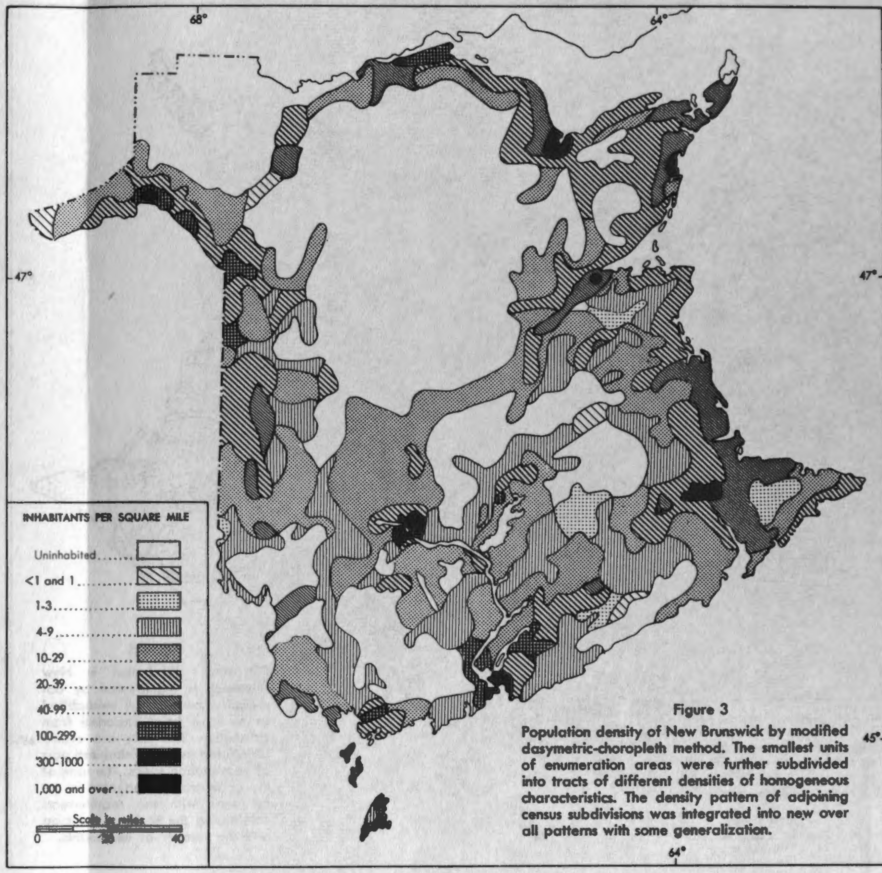

Figure 3
Population density of New Brunswick by modified dasymetric-choropleth method. The smallest units of enumeration areas were further subdivided into tracts of different densities of homogeneous characteristics. The density pattern of adjoining census subdivisions was integrated into new over all patterns with some generalization.

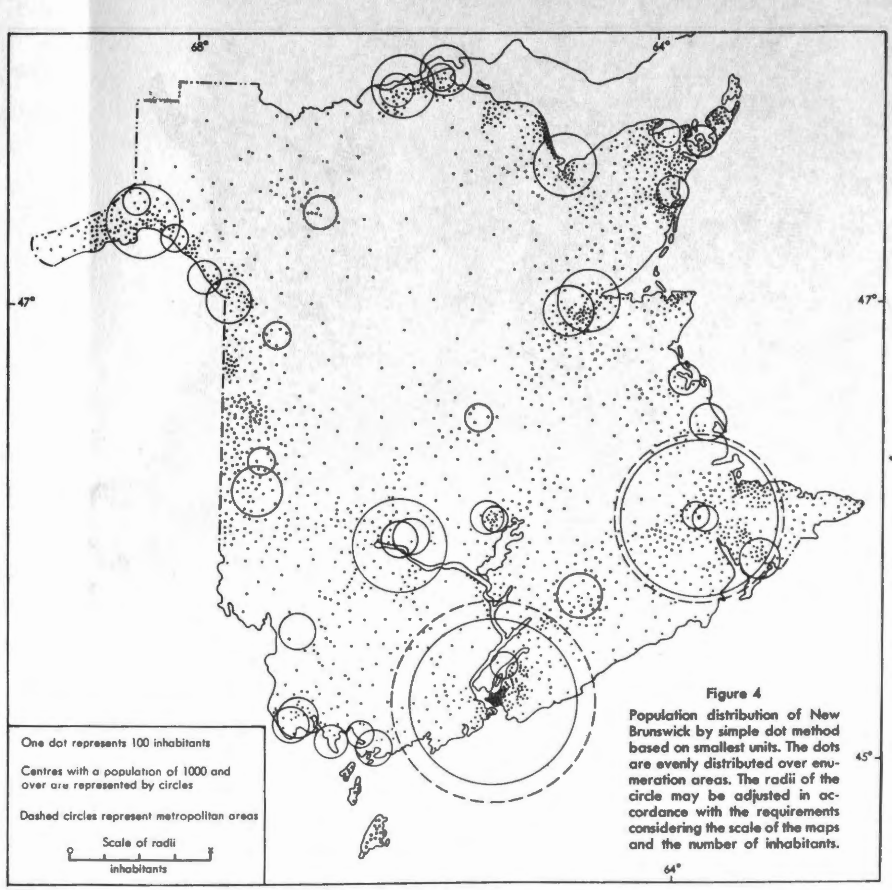

Figure 4
Population distribution of New Brunswick by simple dot method based on smallest units. The dots are evenly distributed over enumeration areas. The radii of the circle may be adjusted in accordance with the requirements considering the scale of the maps and the number of inhabitants.

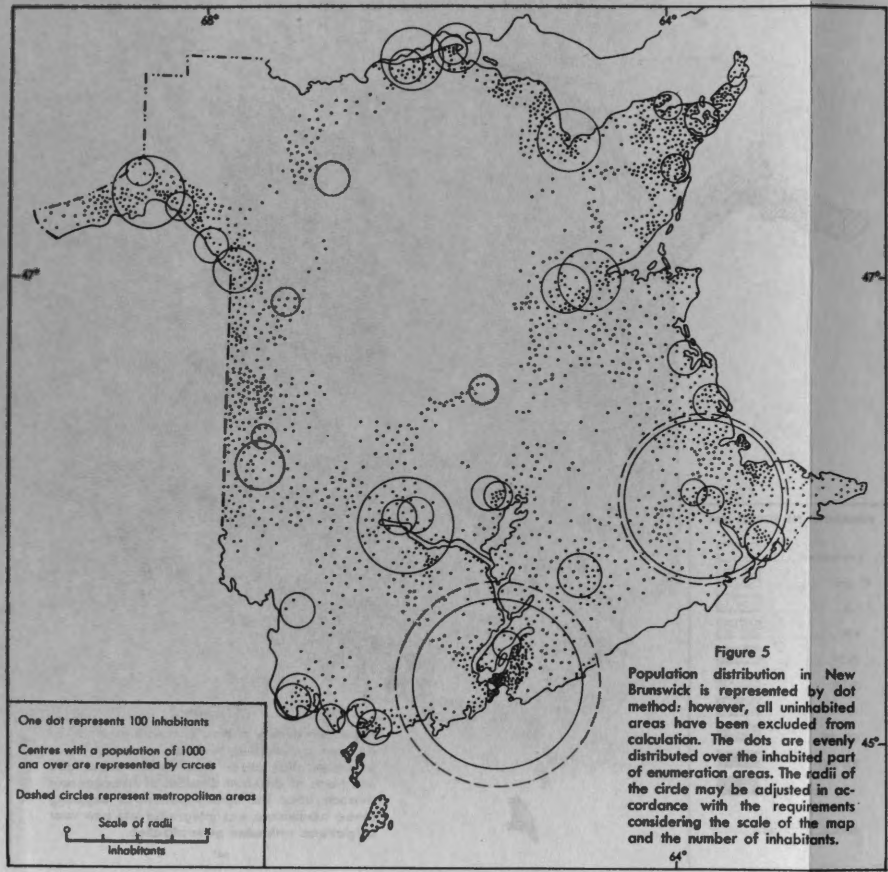

Figure 5

Population distribution in New Brunswick is represented by dot method: however, all uninhabited areas have been excluded from calculation. The dots are evenly distributed over the inhabited part of enumeration areas. The radii of the circle may be adjusted in accordance with the requirements considering the scale of the map and the number of inhabitants.

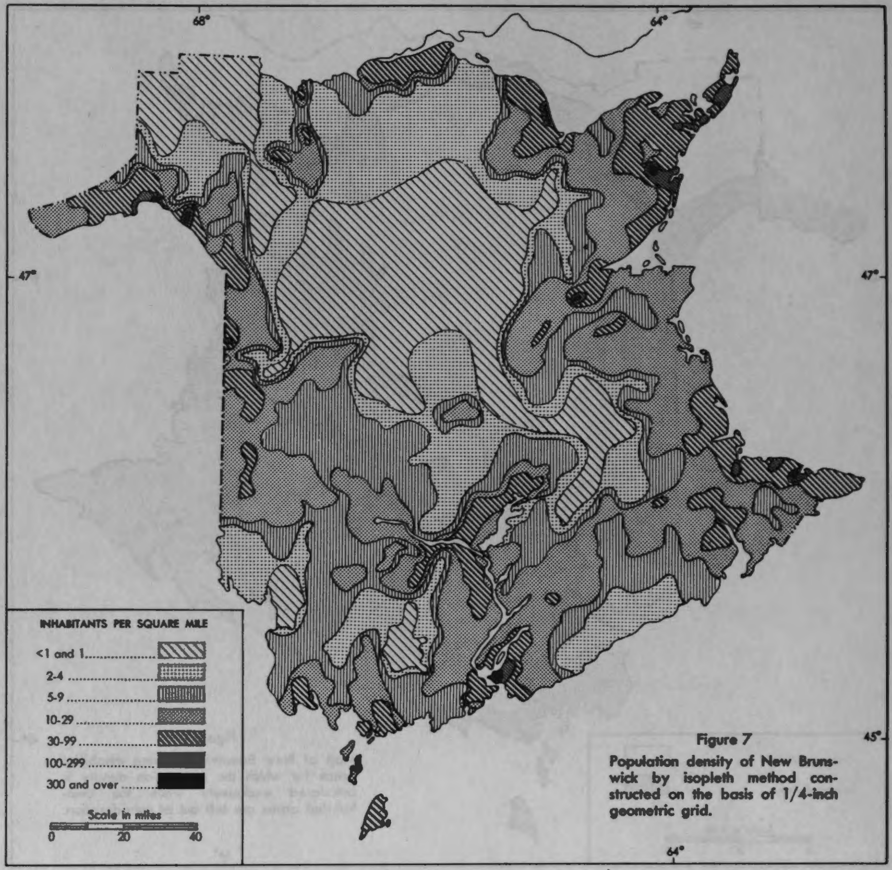

Figure 7

Population density of New Brunswick by isopleth method constructed on the basis of 1/4-inch geometric grid.

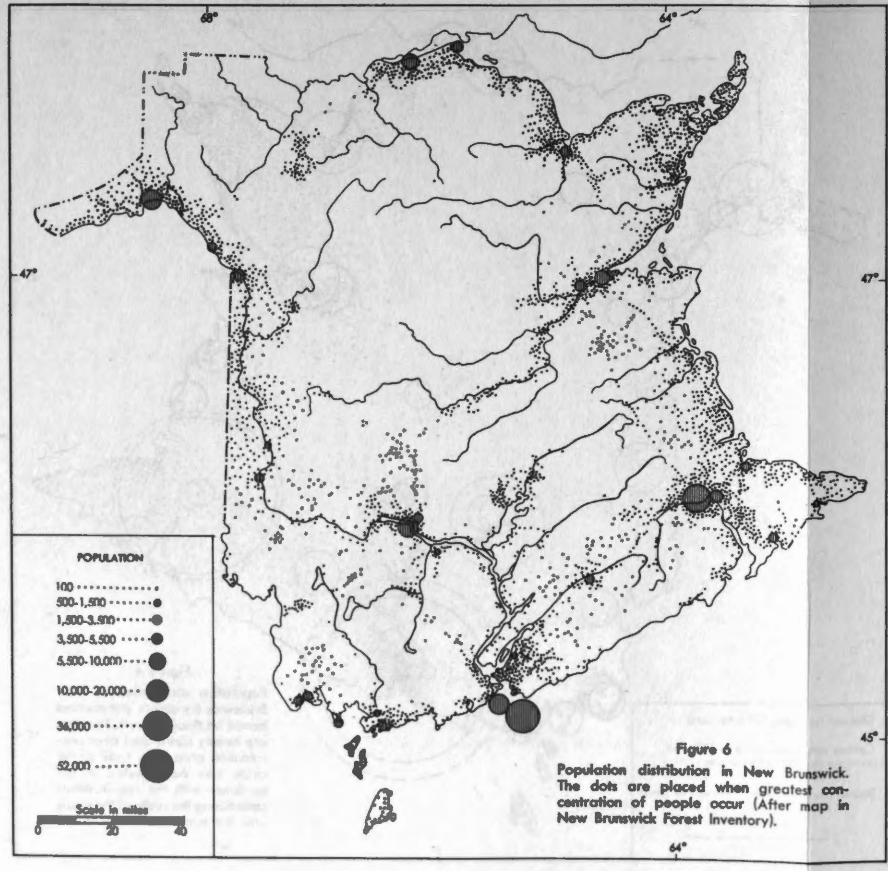

Figure 6

Population distribution in New Brunswick. The dots are placed when greatest concentration of people occur (After map in New Brunswick Forest Inventory).

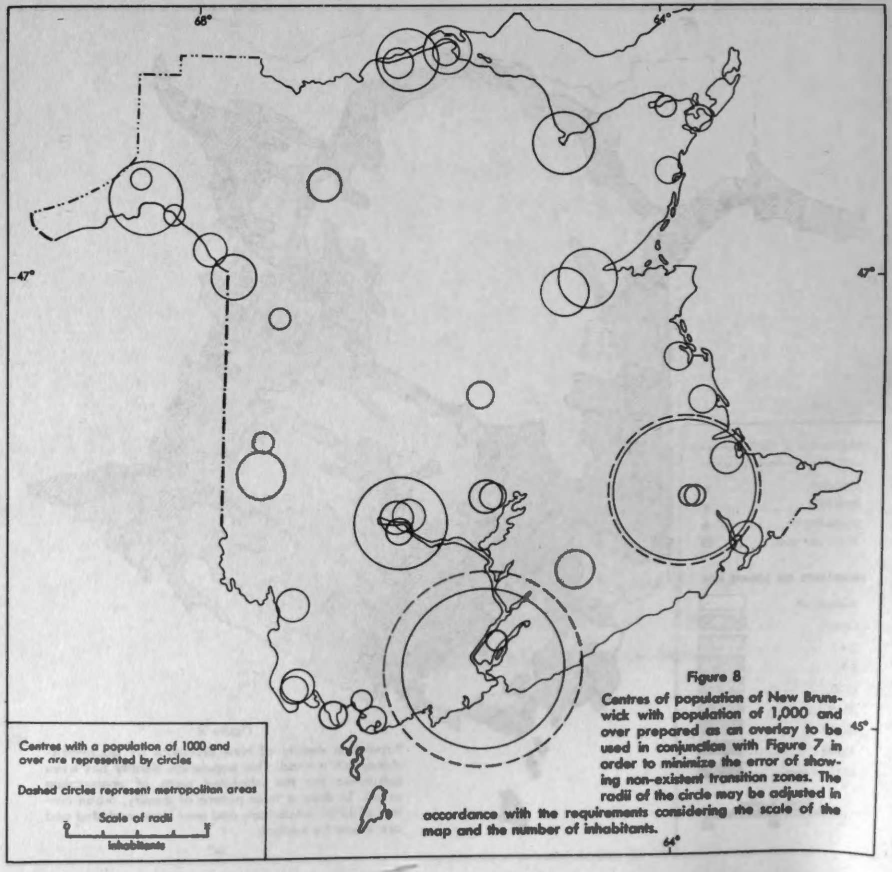

Figure 8

Centres of population of New Brunswick with population of 1,000 and over prepared as an overlay to be used in conjunction with Figure 7 in order to minimize the error of showing non-existent transition zones. The radii of the circle may be adjusted in accordance with the requirements considering the scale of the map and the number of inhabitants.

A dot map of small scale, such as used in atlases, cannot show the exact location of the population and, therefore, the method is limited to visual impression only. Furthermore, the placing of dots within a region or within political units requires detailed geographical knowledge of the area in order to depict the real situation. Furthermore, the method is time-consuming.

A number of articles have been published concerning the dot method in which the theoretical as well as practical aspects are considered. Of particular interest are those by Mackay (1949, 1953), Barnes and Robinson (1940), de Geer (1922), and Geddes (1957).

Isopleth Method

The line symbol used in population-density maps is called an isopleth, (a term derived from the Greek words *isos*–equal and *plethron*–measure, and meaning a line connecting points of equal value.) The transition along these lines is assumed to be gradual and not abrupt. Experiments with this method were applied to areas of Canada lying south of the 55th parallel and also, as a test, to the province of New Brunswick. The preliminary work included the selection of statistical units for their homogeneity, the establishment and location of control points, and the selection of isopleth intervals. The southern part of Canada was divided into about 750 area units for which the population density was calculated and isopleths drawn by interpolation techniques through the centres of gravity. Class intervals were established from a frequency graph. In the Prairie Provinces, due to the large size of base units (about 1,000 square miles), two or more centres of gravity were selected, while in the east, one gravity point was considered satisfactory for units averaging a little more than 300 square miles. The main disadvantage of an isopleth map constructed in this manner is the inevitable introduction of a transition zone in areas where none exists in reality. To lessen the error introduced by arithmetic interpolation, a logarithmic interpolation technique was tried. The use of this interpolation, however, did not eliminate the fictitious transition zone. Though the method is mathematically sound, its geographical application is dubious since the results do not reproduce the actual situation. The technique was introduced by a European geographer, Professor E. Romer, in order to lessen the error of gradual transition and to correct the distance of the isopleth interval. Since there is no connection between the logarithms of certain numbers and the distribution and the density of the population, the information on population is forced into a mathematical assumption that is hardly objective. Italian cartographers* have used the so-called corrected isopleths placed, not according to mathematical interpolation, but by subjective judgment based on knowledge of the real situation. This method, however, introduced an error. If corrected isopleths are drawn, then the line between any two control points should follow a path *subjectively chosen* by the cartographer, who, knowing the population distribution attempts to draw the isopleth where it should be in reality, thus avoiding unnecessary transition zones. In areas with a broad change of density, the isopleths are merely brought very close together and give the impression of a very sharp transition. These corrected isopleths however, are superior to both the standard and the logarithmic ones. In general, the use of this technique in Canadian mapping is greatly restricted due to the diversified pattern of population distribution and densities across the country.

In constructing the isopleth map for the province of New Brunswick another modification of the isopleth technique was attempted (Figure 7). Density data was analysed for enumeration districts. As these districts were of very uneven size, isopleths drawn through their respective geometrical centres would produce inaccurate pictures of the real population density. It was also found that deducing the centre of gravity of each enumeration district for the whole province would be impractical as the 1-inch-to-1-mile topographical sheets would have to be used, and more than 100 would be needed to cover the whole province. In an effort to improve accuracy, the following procedure was applied. First, a 5-inch grid was superimposed on the base map (scale 70 miles to 1 inch). Each corner of each grid square was marked with the density of the area in which it fell. Lines were then plotted through these points according to a predetermined isopleth interval. The interval was chosen subjectively to bring out the more sparsely and the more densely populated areas, as New Brunswick is a province in which towns with a dense population border on areas that are practically uninhabited. To minimize the error of showing non-existent transition zones, all centres with a population of 1,000 and over were excluded and the population of these centres was shown by means of proportional circles (Figure 8). The method, however, was found to be unsatisfactory for New Brunswick because of great variability in the size of the smallest units for which population figures are available, and because of the great contrasts in population density over small areas.

*This technique was used in "Atlantic Fisico-Economico D'Ittalia." Consociuzione Turistica Italiana, Milano 1939, plate 25, Densita di Popolazione.

Appraisal of Isopleth Method

As a result of these experiments, it was concluded that the method is best suited to areas with smooth gradations of density over large areas. Since the isopleths join points of equal density, it is necessary to have large areas of approximately homogeneous density whose value can be represented reasonably accurately by points judiciously located within the areas of homogeneous density. Although there are areas in Canada where such conditions exist, there are also many areas in which the density varies radically over such short distances that innumerable closed lines would be required to represent the density. Moreover, the isopleth method implies that where more than one isopleth intervenes between areas of different densities there is a transition zone of progressively increasing or decreasing density. In many areas of Canada the transition, as for example, between densely inhabited and uninhabited areas, is so abrupt that the representation of a transition zone is erroneous. Although the isopleth method permits the interpolation of values directly from the map, and refinement could still be made by using a greater number of control points, nevertheless, because of numerous disadvantages, the application of this method to population-density mapping in Canada appears unsatisfactory. The main disadvantages noted on the basis of experiments conducted are as follows:

(1) This method distorts the actual density because of the intoduction of transition areas which are seldom present and very often are non-existent. The abrupt contrasts of population density where a small town borders an uninhabited region cannot be shown. For this reason, the characteristic pattern is greatly disrupted.

(2) Where densities are based upon average values, it is obvious that parts of each statistical division must have both higher and lower values than the average. Therefore, unless adjustments are made, isopleth maps tend to lower high-density concentrations, and to raise low-density concentrations. The procedure for adjusting these discrepancies is described in detail by Brooks and Carruthers (1953).

(3) It is impossible to show the areas that are actually settled and that are uninhabited. Instead, a low-density value results over entire uninhabited regions.

(4) The method is not adaptable to show greatly variable elements in the density of population in Canada and it requires a good geographical knowledge of the area under consideration.

It should be noted that some of the disadvantages of the isopleth method are also inherent in the dasymetric technique.

Valuable information regarding isopleth methods, particularly concerning the steps to be followed in the construction of these maps, in the selection of control points, the placing of isopleths and the selection of class intervals is included in various papers by Alexander and Zahorchak (1943), Mackay (1951, 1955) and Bollinger (1930).

Choropleth Method

In the choropleth method the classification is applied to the area of a certain unit over which the density is assumed to be uniform. By the choropleth method, certain predetermined areas as units of reference can be used. They may be either geographically defined regions, a network of geometrical figures, or simply administrative units. The choropleth method assumes that the density of population does not vary within the unit of reference. The resulting pattern does not show the true density at every point, but merely the over-all density of a particular unit. However, the method is widely used because of the ease of obtaining population statistics for administrative units from government census reports. It is possible to include both rural and urban populations in choropleth maps. However, the method gives a very disproportionate picture of real density, especially on large-scale maps. Therefore, it is preferable to show only the rural population by choropleths and to mark urban concentrations by other methods, e.g., by differently colored dots, each corresponding to a certain number of people.

Different types of units may be chosen for calculating densities by this method, resulting in various types of choropleth maps. However, for these experiments, only the administrative-choropleth method was tested. By this method, the densities were calculated directly from census reports for a specific class of statistical unit. In the map of New Brunswick (Figure 9), large administrative units (counties) produced an unsatisfactory choropleth map considered unsuitable even for small-scale density-of-population maps. In compiling Figure 10, medium-size administrative units (parishes) were used, but only a rough picture of population densities was obtained. A more satisfactory map (Figure 11) was produced when the smallest statistical units (enumeration areas) were considered. However, significant patterns in population densities still could not be properly emphasized.

In provinces where there are no townships or municipalities, such as in Newfoundland, Nova Scotia, and British Columbia, census enumeration

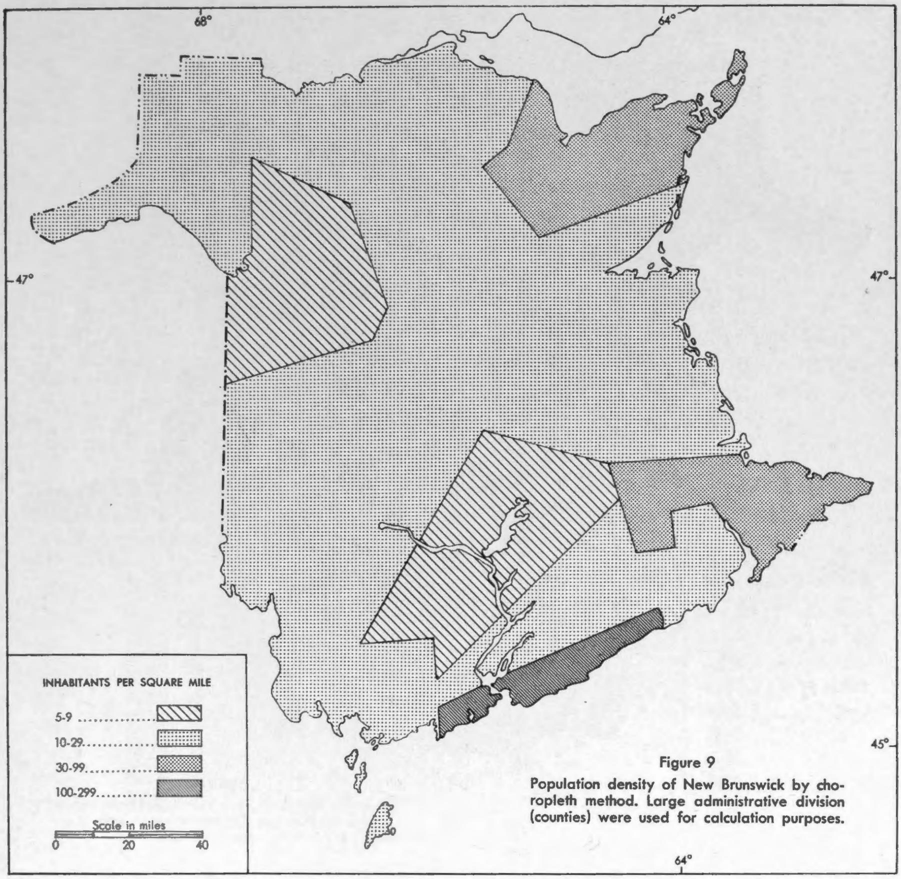

Figure 9
Population density of New Brunswick by choropleth method. Large administrative division (counties) were used for calculation purposes.

areas were grouped to form units of about the same size as the administrative subdivisions in other parts of Canada. The same technique was followed in the territories.

The population densities compiled by these small units depicted a more realistic picture of the distribution of the rural population. There was, however, the problem of the non-rural population living in agglomerations within the boundaries of rural municipalities. This population, of course, could not be spread over the municipality and, therefore, the population of urban centres with less than 5,000 persons was added to the municipality in which the urban centre was located. In such cases, however, the municipalities were subdivided so that the area affected was not larger than 50 square miles. Cities and towns with populations over 5,000 were shown separately by different symbols. It should be mentioned that metropolitan areas and major urban areas were shown by the general outline of their

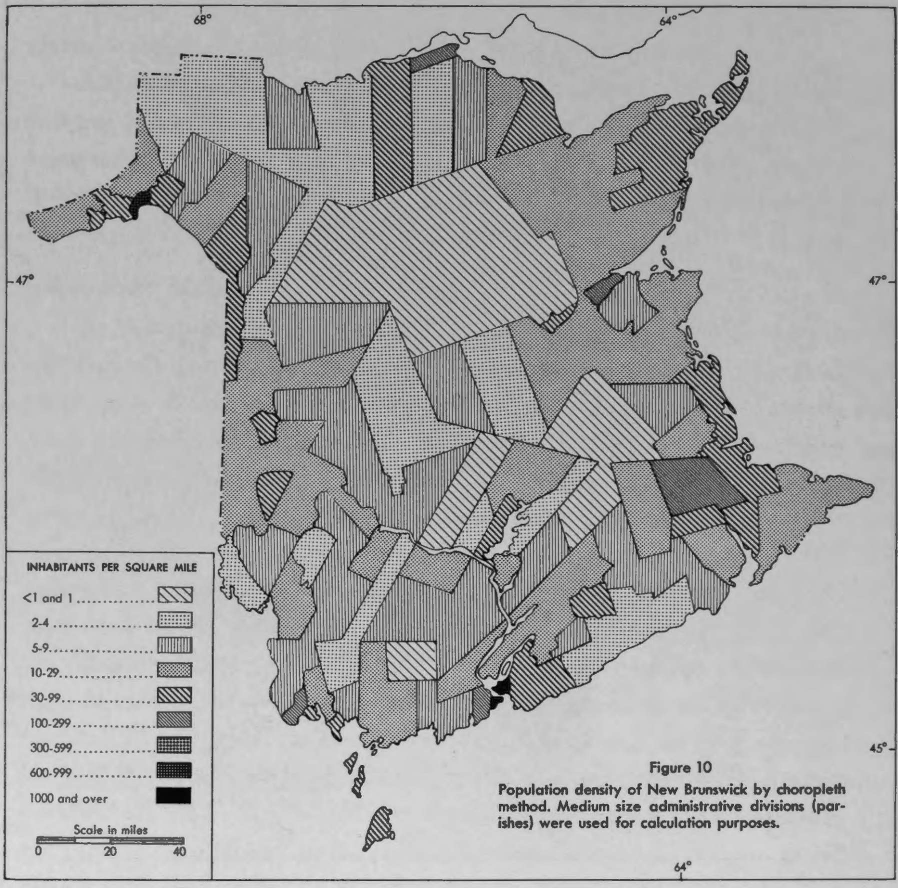

Figure 10
Population density of New Brunswick by choropleth method. Medium size administrative divisions (parishes) were used for calculation purposes.

municipal limits. The major urban areas published by the census are cities of 25,000 persons and over, with a fringe of villages, small towns or heavily populated suburban areas. Population statistics for Indian Reserves are shown as separate items in census data, and various other sources of information had to be consulted in order to determine the exact localization of Indian population. These population figures were always added to the unit in which they occurred in cases when the area of the reserve was too small to form a separate unit.

Appraisal of Choropleth Method

As a result of experiments with the administrative-choropleth method as applied to Canada, it was found that one advantage of this method is the ease of determining various densities and the relative simplicity of constructing the map. As long as the basic limitation, i.e., the impression of

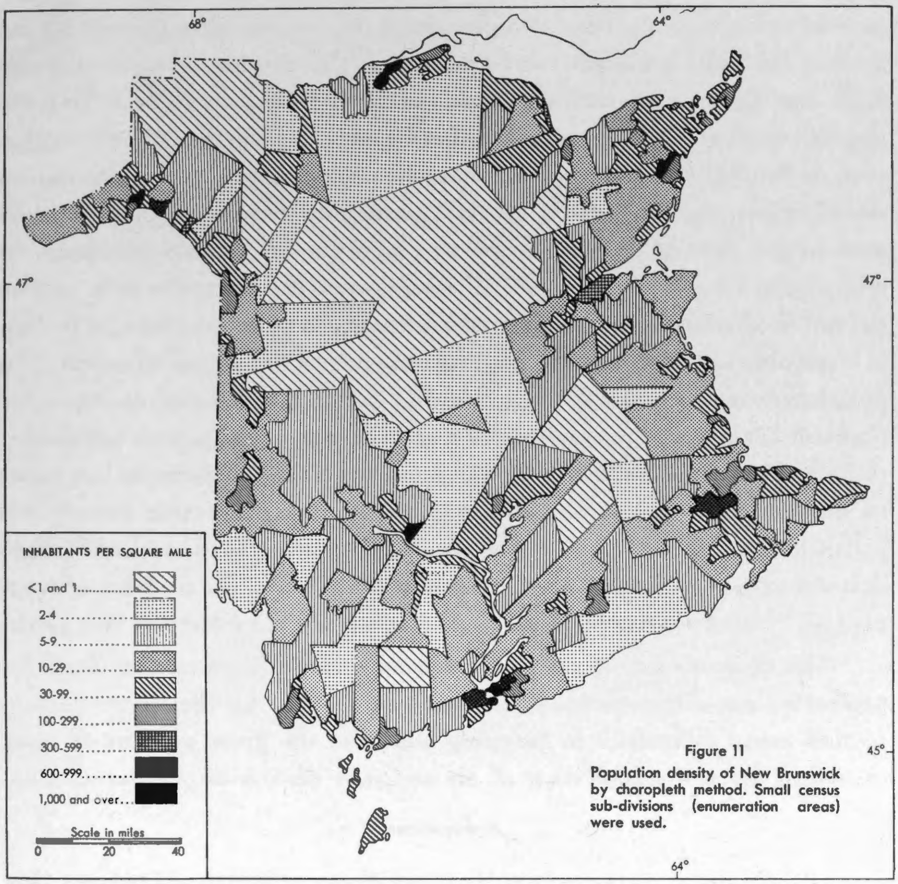

Figure 11
Population density of New Brunswick by choropleth method. Small census sub-divisions (enumeration areas) were used.

uniformity in any one unit, is kept in mind, these maps may be used to show population densities with relative accuracy, provided that small administrative units are used for the calculation of densities. However, it was still considered that the disadvantages of this method were still to numerous to allow its use in the *Atlas of Canada*. The main disadvantages are as follows: (1) The method fails to portray the real and significant geographical population pattern, even though small administrative units are chosen. (2) The representation of the true population pattern can never be achieved as the population is very rarely, if ever, uniformly distributed in a given administrative unit.

SUMMARY

There appears to be no ideal solution for presenting the density of population in Canada by a single cartographic method, but by combining several techniques a generalized picture of the pattern of settlement at the scale of the *Atlas of Canada* can be produced. The dasymetric method shares both the advantages and the disadvantages of the choropleth and the isopleth methods. However, for small-scale mapping, particularly of a region such as Southern Canada with its inherent problems in mapping population distributions, the dasymetric method, particularly in its modified form as used in the *Atlas of Canada*, appears to be the most suitable technique. In addition to the relative simplicity of compilation, it is comparatively easy to determine the area of the unit for which density is to be calculated. Further, it is possible to check how the map corresponds to the actual situation. The population density in Northern Canada presents a special cartographic problem because of the existence of a semi-nomadic, indigenous population (Gajda, 1960) that is not located in any of the larger settlements but exists in small family groups. In order to portray this population certain adjustments to the method would be required; however, as the area of Plate 48 does not extend far enough north to include these areas, the problem of mapping population density in Northern Canada is not considered in this paper.

The experiments on various cartographic techniques and methods for presenting population density that were carried out by the writer demonstrated many difficulties in mapping, and also the great amount of work necessary for the compilation of an accurate density-of-population-map.

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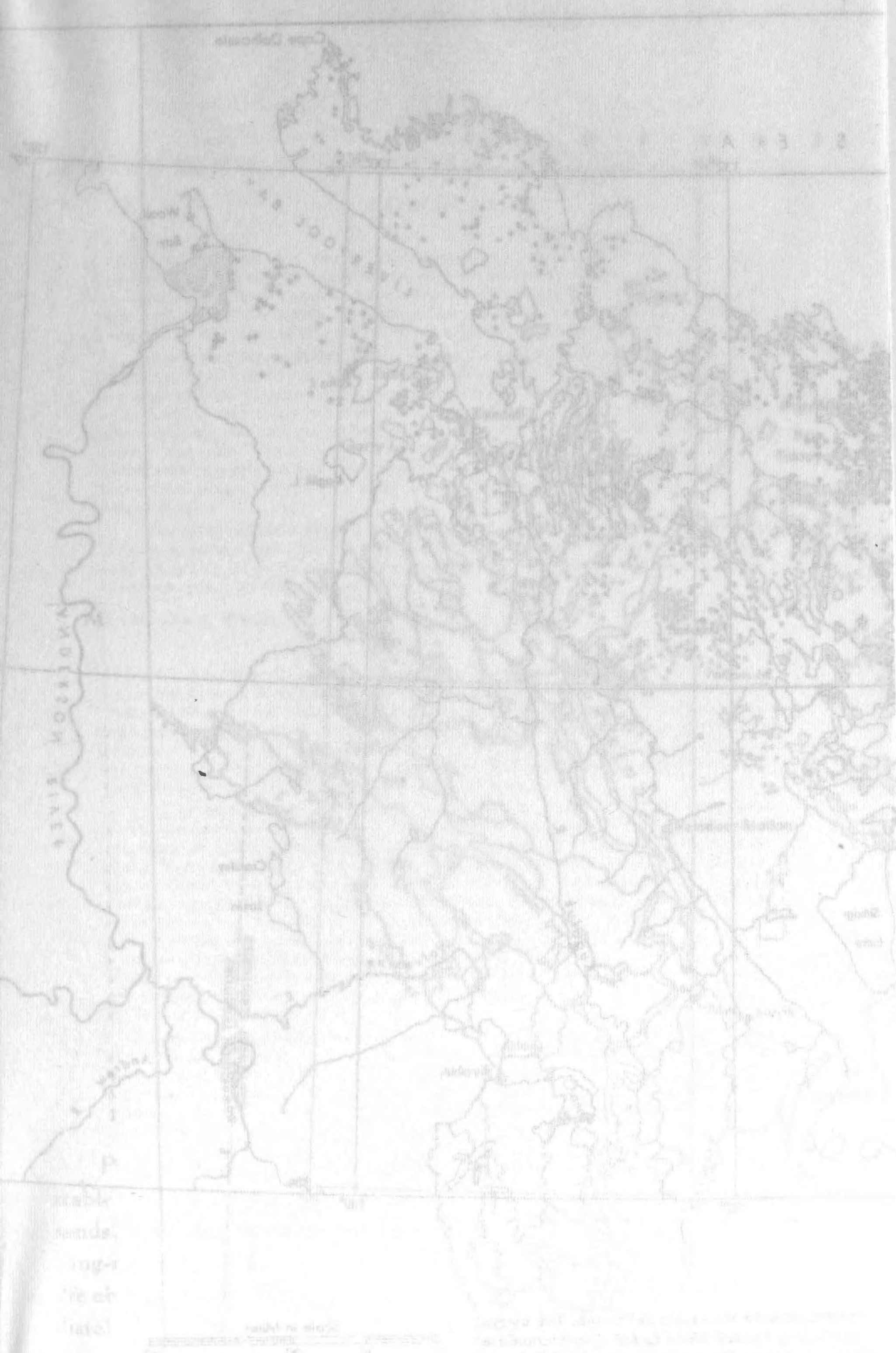
Cape Dalhousie
Scale in Miles

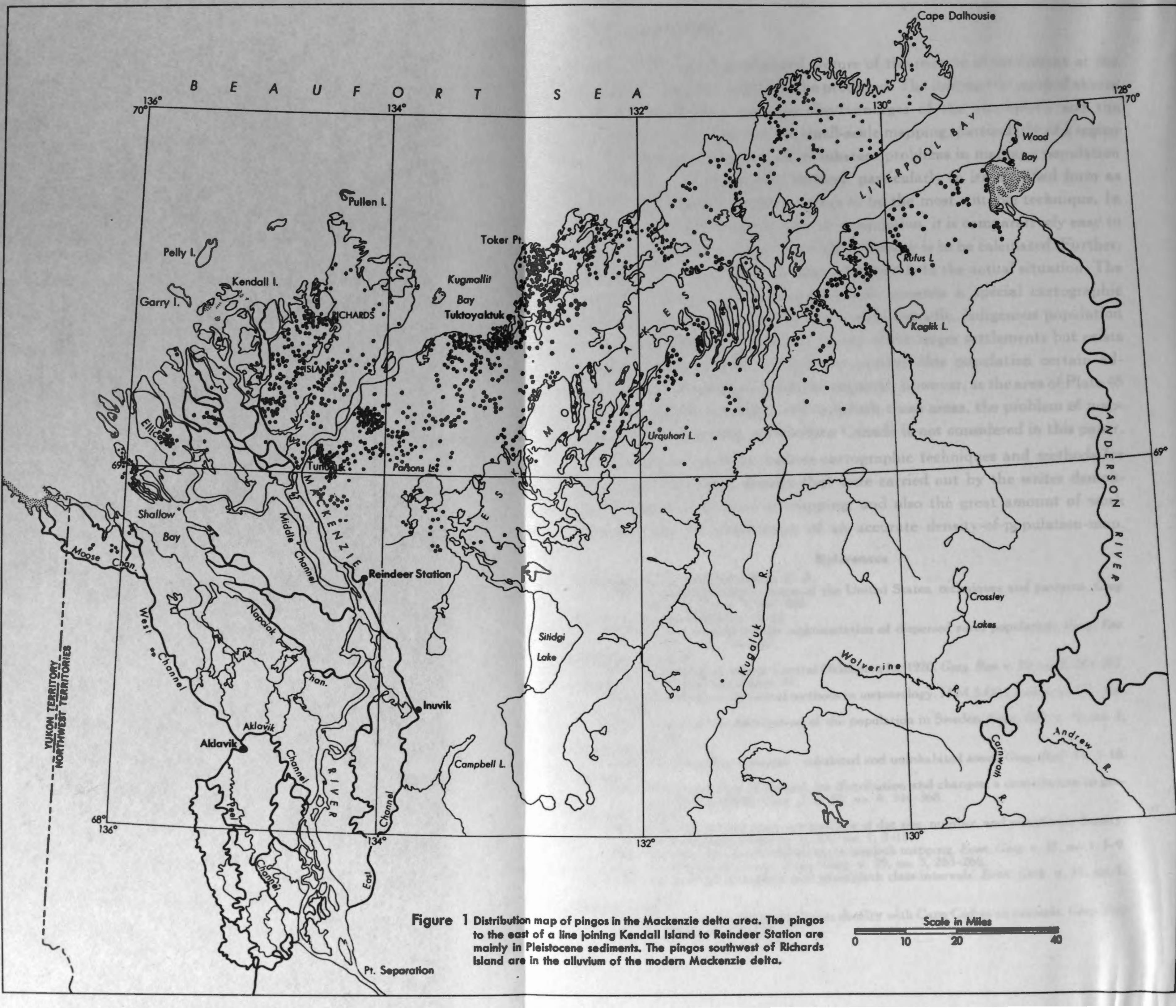

Figure 1 Distribution map of pingos in the Mackenzie delta area. The pingos to the east of a line joining Kendall Island to Reindeer Station are mainly in Pleistocene sediments. The pingos southwest of Richards Island are in the alluvium of the modern Mackenzie delta.

PINGOS OF THE PLEISTOCENE MACKENZIE DELTA AREA

J. Ross Mackay

ABSTRACT: Pingos are stable intrapermafrost ice-cored hills. Most were formed over a thousand years ago as an indirect result of the shoaling of lakes by geomorphic and climatic processes. As the lakes gradually shoaled and lake ice froze to the bottom in winter, an impermeable permafrost cover was formed, ab initio, over the unfrozen saturated sediments beneath.

Downward aggradation of the newly formed permafrost cover was predominantly in fine to medium sands which were not susceptible to extensive ice segregation. Consequently, the pressure of expelled pore water, trapped in a closed system, was relieved by an upward doming of the thinnest part of the overlying permafrost cover, which normally coincided with the deepest part of the lake. Water in the dome froze to form the pingo ice-core. Calculations based on ground and lake temperatures and the nature of the sediments show that pingo growth was slow, probably involving tens of years for the larger pingos.

The large pingos reported in the literature as occurring in the Mackenzie delta are, in fact, not in the modern delta but in Pleistocene sediments to the east. A group of small pingos, which have not been described in detail do, however, occur in the modern delta.

MS submitted, March, 1962.

RÉSUMÉ: Les pingos sont des buttes formées de lentilles de glace qui s'étant individualisées à l'intérieur du pergélisol s'y sont maintenues par la suite. Pour la plupart, leur formation remonte à au-delà d'un siècle et sont dues indirectement à l'asséchement des lacs par les agents climatiques et morphologiques. Au cours du processus d'asséchement, les lacs gèlent complètement en hiver et, au début, amènent la formation d'une couche superficielle de pergélisol au-dessus des sédiments saturés et non gelés.

La progression en profondeur de la couche superficielle de pergélisol s'effectue dans des matériaux sableux dont la granulométrie varie de fine à moyenne et qui ne sont pas susceptibles de former de nombreuses lentilles de glace. Par conséquent, la pression exercée par l'eau capillaire retenue dans ce circuit fermé donne naissance à un soulèvement en forme de dôme, là où la couche superficielle du pergélisol est la plus mince. Cette couche coïncide d'ordinaire avec la partie la plus profonde du lac. L'eau à l'intérieur du dôme gèle et forme alors une lentille de glace qui constitue le noyau du pingo. Les mesures de température prises sur le sol et l'eau du lac ainsi que les renseignements recueillis sur la nature des sédiments démontrent que la formation de ces pingos est très lente; dans le cas des pingos plus considérables, la période de formation serait de 10 ans.

Il est souvent fait mention dans les écrits scientifiques des gros pingos du delta du Mackenzie; leur formation ne remonte pas à l'époque du delta actuel, mais bien à l'époque pléistocène, tout comme d'ailleurs les sédiments situés plus à l'ouest. On rencontre dans le delta récent un réseau de petits pingos dont la description détaillée n'apparaît pas dans la présente étude.

Pingos are ice-cored hills, typically conical in shape. They are relatively stable intrapermafrost features, normally enduring from hundreds to thousands of years, thus differing from the smaller seasonal frost blisters or icing-mounds which form above the permafrost surface in winter. Pingos are abundant and widely distributed in the Mackenzie delta area in: (a) the distal part of the modern Mackenzie delta, and (b) the area of Pleistocene deposits to the east (Figure 1). The pingos in the two areas differ in age, size, and details of formation. This paper is concerned primarily with the older and larger pingos of the Pleistocene area; it describes their distribution and characteristics, discusses theoretical and practical aspects of their origin, and estimates their age.

TERMINOLOGY

The Eskimo word pingo, applied locally by natives to conical mounds in the Mackenzie delta area, was proposed by Porsild (1938) as a technical term for ice-cored hills. The term is now widely adopted (e.g. Muller, 1947, p. 220; American Geological Institute, Glossary of Geology and Related Sciences, 1957, p. 221). The Russian word *bulgunniakh* is synonymous with pingo (Grave, 1956). Some writers use hydrolaccolith and pingo interchangeably (Cederstrom and others, 1953, p. 8) but to do so is undesirable, because of the genetic connotation involved. As Müller (1959, p. 118) stresses, pingos are intrapermafrost mounds, lasting for many years. Therefore, annual mounds, such as frost blisters and icing-mounds, which grow in the active layer in winter, by definition are not pingos.

At the present time, two genetic types of pingos are generally recognized. Leffingwell (1919, pp. 153–155) has attributed the origin of mounds (pingos), which he studied in Northern Alaska, to hydraulic pressure that bowed up the overlying strata. Porsild (1938) has pointed out that the typical pingos of the Mackenzie delta area are different from the hydraulic type of Northern Alaska and similar mounds in East Greenland. Müller (1959), in his comprehensive study of both kinds of pingos, classifies the above types into the "open system" or East Greenland type and the "closed system" or Mackenzie type. The pingos of the Pleistocene Mackenzie delta region are of the closed system type, but some pingos do not fit neatly into either category.

DISTRIBUTION

There are over 1,400 pingos in the Mackenzie delta area (Figure 1) between the west side of the delta, at the Yukon-Northwest Territories boundary, and Nicholson Peninsula on the east. The area involved is a northeast-southwest coastal zone some 200 miles* (320 km) long and 50 miles (80 km) wide. The pingos occurring along the Yukon coast, west of

*Note: The cgs system has been used for computations and the presentation of data for the sake of efficiency and simplicity. Insofar as almost all measurements and graphs are concerned, data shown in metres can be read as yards without significant loss of accuracy the Mackenzie delta, may not exceed ten. To the east of Nicholson Peninsula, there are very few pingos (Mackay, 1958a, pp. 54–56).

The largest group of pingos, numbering over 1,350, lies in a belt extending east from Richards Island through Tuktoyaktuk Peninsula to Cape Dalhousie; this group includes those pingos on the south side of the Eskimo Lakes. Stager (1956) has plotted the locations of about 1,380 pingos in this area, and the writer about 1,350, the slight discrepancy arising from the definition of size used to delimit the smallest pingo. As the figures should be considered a minimum, rather than a maximum, the total number of pingos is 1,350 to 1,400. and occurs in an area of Pleistocene sands and silts with rolling relief. A second group of pingos, numbering more than 70, occurs on the seaward part of the modern Mackenzie delta below storm level, with their bases usually within 1.5 metres of sea level. These pingos are within 15 miles (24 km) of the coast and lie southwest of Richards Island. A third group of about 10 pingos is found on an older part of the modern Mackenzie delta near the lower course of West Channel. There are several pingo-like hills along the eastern slopes of the Richardson Mountains at 68°29'N and near Mount Goodenough (Fraser, 1956), and one or more pingos in the southern part of the Port Brabant map-sheet of the National Topographic System. This paper deals with the first and largest group of pingos, with only passing reference to other groups.

CHARACTERISTICS

Occurrence in Depressions

The most distinctive terrain characteristic of the pingos is their occurrence in flat low-lying areas. With very few exceptions, the low areas are present or former lake basins or channels with poor drainage. Every pingo seen in the field has been in a depression, and every mound, identifiable from airphotos as a pingo, has also been in a depression (*see* Frost, 1950, p. 34 for Alaskan pingos). Typically, the flats are broken up into tundra polygons, normally in the depressed centre stage of development. High-centred peaty tundra polygons are most abundant in the drier land around the bases of the pingos. A few pingos protrude as islands, usually near the centres of shallow lakes. No pingo has been observed rising as an island in a deep lake, such as a lake more than 3 metres deep. About 56 per cent (Stager, 1956, p. 16) occur as islands or are partially surrounded by water. The remaining 44 per cent are surrounded by poorly drained tundra polygon flats. A small percentage of the pingos, possibly less than 0.1 per cent, are in abandoned drainage channels. Several are in modern floodplains, for example, of Holmes (Peters) Creek, but these may be of the East Greenland type. No feature, which unequivocally could be identified as a pingo, has been observed either in a large lake (e.g. 1 mile or 1.5 km in diameter) surrounded by water considerably deeper than the winter thickness of ice, nor on top of a positive relief feature, such as a mesa-like tableland. The nearly perfect correlation between pingos and present or former lake basins and channels has obvious implications on their origin.

The correlation between areas of pingo concentration with completely or almost completely drained lakes is borne out by their irregular distribution. Most drained or partially drained lakes are found in areas where a river system, even though poorly integrated, has developed in postglacial time, or else where coastal recession has initiated drainage. Areas with interior or no visible surface drainage tend to have few old lakebeds.

On Richards Island, the main concentration of pingos is on the west half where drainage is most complete (Figure 1). The eastern and north-eastern part has numerous relatively deep lakes with few streams, and consequently few old lakebeds or shallow lakes. On the east side of East

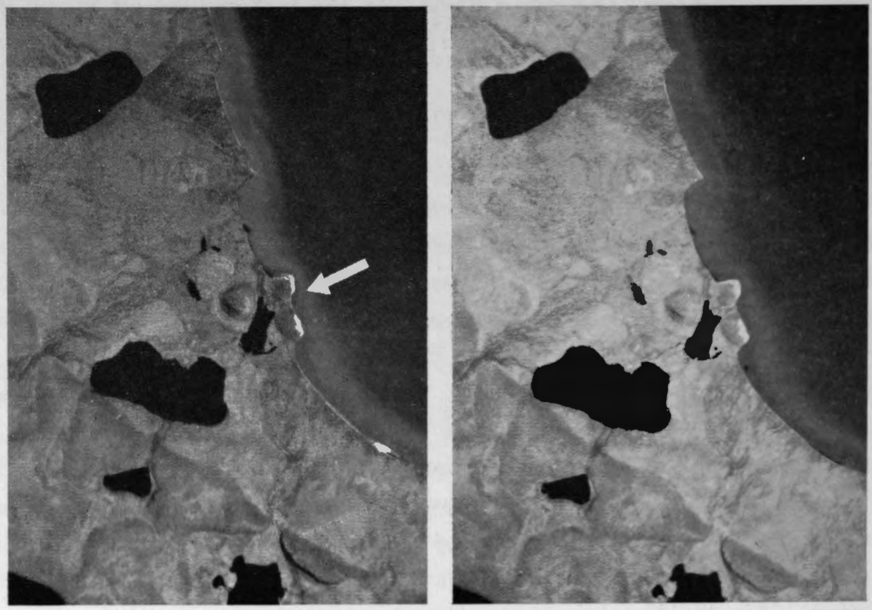

Figure 2 The stereo pair shows four pingos on the north shore of the Eskimo Lakes at 69° 02′ N, 133° 12′ W. The three large pingos, two of which are wave cut, exceed 50 feet in height. A tiny pingo arises on the edge of the lake in the center of the group. (RCAF photos A 12918 – 81 and 82)

Channel, between Tununuk and Holmes (Peters) Creek, the concentrations are related to stream drainage. The high density of pingos, about 10 per 10 square miles (26 square km), between Kittigazuit and Toker Point, is in a coastal region where cliff recession and stream action have partially drained large areas. Tuktoyaktuk Peninsula, east of Toker Point, has a pingo density averaging less than 1 pingo per 10 square miles. Much of the region is low and flat, with either a very poorly integrated drainage system, or no visible surface drainage. Thus, the inferred conditions suitable for pingo growth are largely absent. Pingo concentrations along the north and south coasts of the Eskimo Lakes are closely associated with shore-cliff recession and stream action.

There is a general relationship between the density of depressions "suitable" for pingo growth and the number of pingos present. The data in Figure 3 has been computed and plotted from a random sample of 25 air-photos. The area studied on each sample photo was about 30 square miles (78 square km), the total sampled area covering about 750 square miles (1950 square km). For each 30 square miles, the limits of all depressions containing pingos were outlined, the boundaries following as closely as possible the estimated extent of the depressions which generated the pingos. Then all similarly appearing areas without pingos on the air photos were marked. As shown in Figure 3, the 25 points can be represented reasonably well—for the range of data—by the line

$$y = 10^{.071x - .283} \tag{1}$$

where y is the pingo density per 10 square miles and x is the depression density per 10 square miles. (Note: unit areas of 10 square miles are used to avoid, as far as possible, "fractional" pingo densities of less than one pingo per unit area). Although a certain degree of subjectivity was unavoidably involved in the sampling procedure, the results suggest that the probability of finding a nearby pingo, if one pingo is selected at random, increases non-linearly with the depression density or number of "suitable" pingo depressions. For example, where the number of depressions considered suitable for pingo growth averages 5 per 10 square miles, only about one depression out of 5 (20 per cent) will usually contain a pingo. On the other hand, if there are 20 suitable depressions per 10 square miles, about 13 of the 20 depressions (65 per cent) will usually contain pingos. This implies unusually favorable conditions for the growth of pingo fields in certain areas.

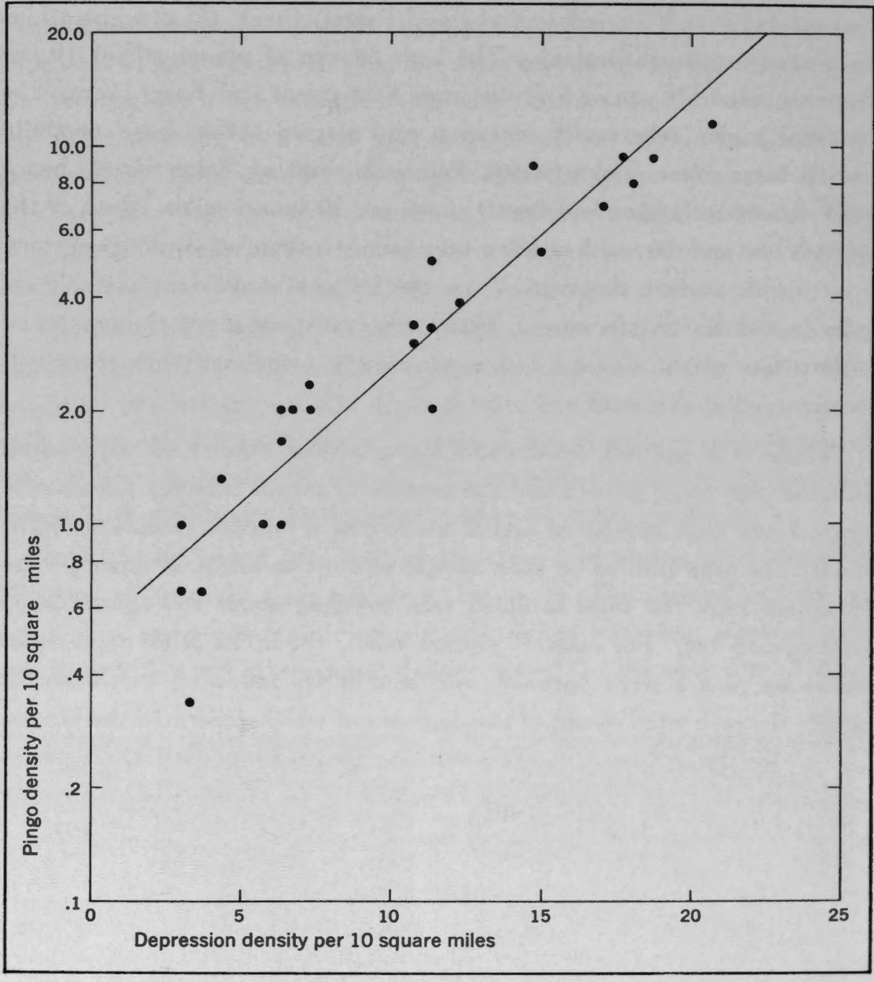

Figure 3 Scatter diagram for a random sample of 25 air photos to show the relation between the actual pingo density per 10 square miles (26 square km) and the number of depressions, considered suitable for pingo growth, per 10 square miles.

SHAPE

The typical pingo rises as a solitary hill, ranging from round through elliptical to oval in ground plan. Basal diameters vary from about 30 to 600 metres and heights from about 3 to 45 metres. The highest pingos have intermediate sized ground diameters of about 150 to 200 metres. As diameters increase beyond 150 to 200 metres, altitudes tend to decrease, so that the broadest and longest pingos become merely large bulge-like swellings or protuberances.

In plan, the outline of a pingo tends to be simple and smooth, without crenulations or indentations. Oval shapes are most common, but many are elliptical or nearly circular. The rare crenulated outlines are usually the result of the coalescence of two or more round-to-oval pingos. Of the remaining irregular shapes, most are straight to sinuous barrows, with lengths several times their breadths. By way of contrast, in the modern Mackenzie delta, some pingos are so long and sinuous that they resemble eskers.

In vertical cross-section, most pingos are asymmetric. Even the classic and most photographed pingo, Ibyuk pingo southwest of Tuktoyaktuk, is symmetrical when viewed from only one direction; the variation in slope between opposite sides approaches 20°. The steepest pingo slopes rarely exceed 45°. Such steep slopes are usually present on only one side of a pingo, so that spoon-shaped asymmetric pingos predominate over plug shapes. The gentlest slopes, which may be only several degrees, are found on small incipient pingos but most commonly on large bulge-like swellings with diameters of more than 300 metres. The bulge-like swellings suggest that lateral growth has taken place at the expense of vertical growth.

The side slopes of pingos are either straight or convex upward. No pingos with concave sides, resembling those of classic volcanoes, have been observed. Most of them—76 per cent according to Stager (1956, p. 16)—have smooth rounded summits. Most summits of the largest pingos are ruptured, with star-shaped craters surrounded by cuestas which seemingly open out as rupturing progresses, like the petals of a budding flower. Rupturing of pingo summits frequently results in exposure of the ice-core to melting with partial or total collapse in prospect.

OVERBURDEN THICKNESS

The thickness of the sediments over the ice-core varies considerably and appears to be directly related to the size of the pingo, with small pingos having the thinnest cover of sediments, large ones the thickest. However, the evidence is, at present, based upon estimated and known overburden cover of only a few pingos. For example, on the east side of McKinley Bay, a violent storm (in the summer of 1955) cleaned off the face of a partially sectioned pingo to expose the ice-core (Figure 4). About three-fifths of the pingo had previously been cut away. The original was probably oval, about 90 metres long and 7 to 10 metres high. The cut face showed an ice-core of clear, white to bluish bubbly ice, with crystals 2.5 to 4.0 cm in diameter, beneath 1.2 to 1.5 metres of brownish sand. The sand at the contact with

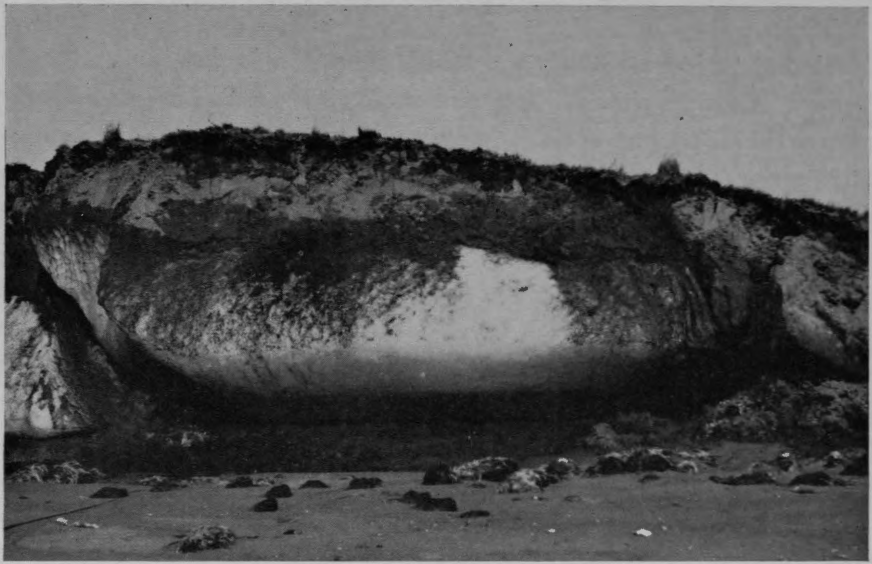

Figure 4 Wave-cut pingo at McKinley Bay, 69° 59′ 30″ N, 131° 01′ 40″ W taken on August 23, 1955. (Geographical Branch photo)

the ice was dark grey. Part of the overlying sands were lacustrine, as fresh-water shells and peat were present. Sections in a nearby larger collapsed pingo at Point Atkinson suggest an overburden thickness of 3 to 6 metres, with the original height of the pingo being 13 to 15 metres. A collapsed pingo on the southwest side of Hendrickson Island has an annular ridge up to 12 metres high, surrounding a lake about 30 by 75 metres in size. At least 1.5 metres of peat over sand is exposed at the top of the rim. The overburden thickness is estimated at 4.5 metres or more. Ibyuk pingo, near Tuktoyaktuk, one of the largest known, is about 40 metres high, and the overburden is 14 metres thick (Müller, 1959, p. 79). A nearby smaller 9-metre-high pingo has an overburden of 4.5 metres (Müller, 1959, p. 84). A ruptured pingo, 30 metres high, has been drilled on one side to a vertical depth of about 8 metres or about 7 metres normal to the surface (Pihlainen and others, 1956) without reaching ice. As ice in the center of the crater occurs 11 metres below the original surface, the overburden is about 9 metres thick. In the modern Mackenzie delta, pingo ice was encountered 1.5 metres below the summit of a pingo 4.5 metres high.

The preceding examples give data on overburden thicknesses which are known with reasonable accuracy. The overburden above the pingo ice, for the five examples cited, is about one-half to one-third of the pingo height (*see Foundations of Geocryology*, 1959, Vol. 1, p. 292). The greatest variation in cover thickness, as determined from collapsed pingos, is in irregularly shaped pingos, or those with asymmetrically located ice-cores.

TYPE OF SEDIMENTS

All of the pingos are believed to have developed in Pleistocene interglacial sands and silts locally veneered with glacial drift and post-glacial sands, silts, and organic matter rarely totalling more than 3 to 6 metres in thickness (Mackay, 1956a, 1956b, 1957, 1958b). Inasmuch as most pingos have grown in lake depressions, organic matter interbedded with fine-grained lake sediments is typical of the top metre or so of a pingo. The lake sediments may overlie postglacial material or glacial drift, some of which is reworked.

At a variable depth, which is estimated at about 3 to 6 metres, the underlying sediments are believed to be predominantly of sands to an undetermined depth. Although no stratigraphic section has been observed directly beneath a pingo ice-core, many tens of miles of coastal sections in which pingos occur have been examined in the field, there being every reason to believe that the sediments observed are identical with those beneath nearby pingos. Over 30 'samples' of the sediments from Richards Island to Nicholson Peninsula have been tested for grain size. With very few exceptions, the sediments are at least 95 per cent in the sand range, most specimens testing 99 per cent sand. The sand is predominantly fine to medium sand (0.1 to 0.5 mm). The silt fraction (<0.05 mm) rarely exceeds one or two per cent. A few silty laminae are locally present; silty clays may occur at depth. However, pingos appear to have developed in sandy material which is too coarse grained to be susceptible to extensive ice lens formation. The thick (e.g. 2 to 6 metres) horizontal ice sheets which are widely distributed in the same region as the pingos have not grown in the sands. Most have developed in situ in silty material near the surface of the ground, with the top of the ice lens at, or just beneath, the base of the active layer. A few ice lenses have grown at a depth of 3, 6 or more metres as a sandwich filling between sandy beds. In either case, the ice lenses appear to have developed in what was initially a thin silty bed, the silty material becoming dispersed through the ice lens during its growth. The grain-size distribution of the soil in the ice, or that released by melting, usually has well over 25 per cent in the clay-silt range.

PINGO ICE-CORES

Drilling records and exposures of naturally sectioned and collapsed pingos suggest that their ice-cores generally resemble the gross outer shapes of the pingos themselves but that they have somewhat steeper sides. The bottom of the pingo ice is inferred to lie close to, or below, the level of the flat terrain around the pingo base.

Much information on the size and shape of pingo ice-cores has been obtained from a study of collapsed pingos (Figure 5). When pingo ice melts, usually through summit rupturing of the protective cover but also through lateral cutting along a bluff, the overburden cover, along with slumped side material, in-fills part of the depression left by melting of the core. The bottom of the collapsed central depression—which usually contains a lake—does not represent the bottom of the pingo ice-core but stands at least as much above it as there is infilling by overburden material. The bottoms of lakes or ponds in the collapsed pingos are not often appreciably above the bases of the pingos; in some, the bottoms may be 2, 3 or more metres below the surrounding flat terrain. When allowance is made for slumped and collapsed material, pingo ice-cores are inferred to bottom below the level of the adjacent terrain. This is in agreement with the observation by Müller (1959, p. 84) for the Sitiyok pingo near Tuktoyaktuk and the drawing of a pingo in the USSR (*Foundations of Geocryology*, 1959, Vol. 1, p. 292). The collapse pattern of ice-cores indicates that they are often plug-like or spoon-shaped (like an egg sectioned through the long axis) with opposed steep and gentle sides.

The shape of the central depression left by melting would reflect that of the ice-core if it could be assumed that the overburden cover were of uniform thickness and all the ice had melted, none being preserved beneath collapsed material. An examination of collapsed pingos appearing on air photographs, together with field observations, shows that the overburden thickness may vary in any one pingo. This means that the outer shape of the pingo must normally differ from that of its core.

There seems to be a good possibility that some pingos have more than one ice-core. Pingos, apparently compound in form, have been observed with one collapsed ice-core (for example, RCAF photo A 12760–20) adjacent to an isolated, uncollapsed portion. The physical appearance of such a compound pingo strongly suggests two separate ice-cores; some may have three cores.*

*Gravity measurements have been made by the British American Oil Company, of a pingo 65 feet high and 1,000 feet in diameter at 69°20′N, 133°26′W. The gravity observations, corrected for elevation, Bouguer effect, latitude, terrain and regional dip, give a preferred density of approximately 1.0. As the density of ice is about 0.9, this suggests a high percentage of ice in the pingo. However, as the amount of ground-ice in the area and underlying the pingo is unknown, it would not be realistic to estimate the exact size of the ice-core.

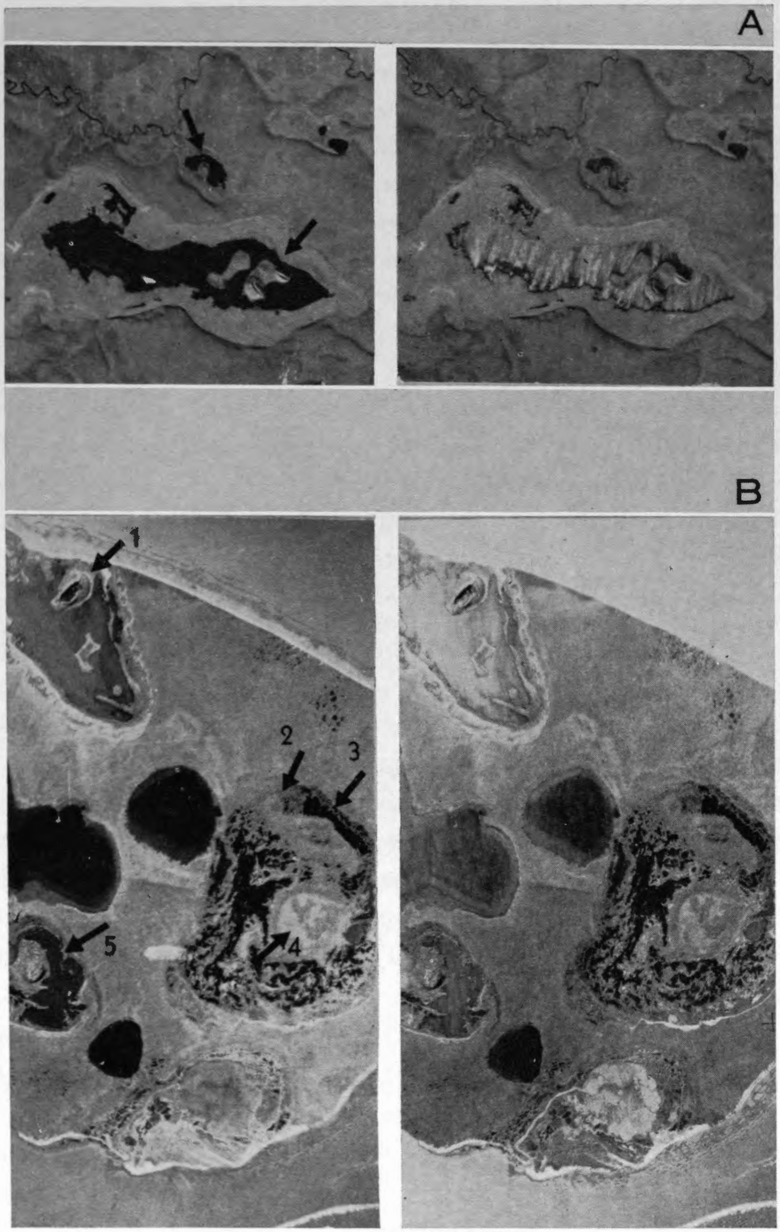

Figure 5A A group of complex pingos with ruptured summits occur in a shallow partially drained lake to the northeast of Tuktoyaktuk at 69° 37′ N, 132° 24′ W. A small pingo rises in another partially drained lake. (RCAF photos A 12847 – 335 and 336)

B The pingos are on Hendrickson Island (69° 30′ N, 133° 35′ W) whose 15-foot bluffs may be used as a scale. Pingo 1 shows a collapsed center from melting of the ice-core. The bottom of the lake in the pingo stands above sea level. A photo of the pingo (Porsild, 1938) taken more than 20 years before the air photos shows a ground view similar to the present. Pingos 2, 3 and 5 show excellent tundra polygons. Pingo 4 is of the low bulge type. (RCAF photos A 12918 – 250 and 251)

ORIGIN OF PINGOS

Only two basic theories have been seriously proposed for the origin of the pingos in the study area. The early Arctic explorer Richardson (1851) who observed the pingos in 1826 and again in 1848, considered them to be a peculiar type of sand hill, but his theory was presented more in the nature of a comment than as a reasoned theory. Gussow (1954) has suggested that pingos are piercement domes which have formed as a result of a geostatic load on thick buried ice of Pleistocene origin. The theory does not seem to be tenable (Mackay, 1958a, pp. 54–56). For example, all evidence indicates that the ice-cores terminate a short distance beneath the pingos. The overburden of some pingos contains glacial drift and wood dated at more than 33,000 years, and therefore it is highly unlikely that relic glacier ice could have been preserved in buried form for such a long period in the area. In addition the close association of pingo size, shape, and overburden thickness with present or recently drained lakes, all show that pingos are superficial features formed in response to local conditions.

The most generally accepted origin for pingos has been proposed by Porsild (1938, p. 55) who states they "... were formed by local upheaval due to expansion following the progressive downward freezing of a body or lens of water or semi-fluid mud or silt enclosed between bedrock and the frozen surface soil, much in the way in which the cork of a bottle filled with water is pushed up by the expansion of the water when freezing." Porsild points out that the bottoms of the larger lakes are unfrozen, but will freeze when a lake becomes shallow enough to freeze solid in winter. The rapid infilling of lakes which formed a "closed system" is attributed primarily to sedimentation.

Müller (1959) in his detailed analysis of the Mackenzie "closed system" type of pingo accepts Porsild's basic theory, including the shoaling of lakes resulting from infilling. The growth of a pingo is depicted as follows: (1) a relatively large (e.g. more than 300 metres in diameter) and deep (e.g. more than 30 metres) lake will have an unfrozen central portion surrounded by permafrost; (2) gradual infilling of the lake by organic matter and sediments simultaneously reduces the lake size and depth; (3) permafrost forms, ab initio, at the bottom of the lake concurrent with infilling, to form an impermeable cap on the unfrozen material beneath; (4) the inward growth of permafrost at the sides of the unfrozen central core creates a hydrostatic pressure through volume expansion on freezing; (5) the superfluous water is trapped in a "closed system" created by the impermeable capping of permafrost at the lake bottom, the advancing permafrost surface at the sides, and saturated soil at depth; and (6) the expelled water, taking the path of least resistance in a closed system, forms an ice-core near the geometric lake center.

A consideration of the origin of pingos as proposed by Porsild, and elaborated upon by Müller, requires a knowledge of permafrost conditions beneath lakes.

Depth of Unfrozen Ground Beneath Lakes

If the depth of a lake exceeds that of the winter ice thickness, the subjacent bottom sediments in the deeper portions will be unfrozen, irrespective of the presence or absence of permafrost at greater depth, because the thermal capacity of the unfrozen pool of water will prevent them from freezing. The thickness of lake ice in the Mackenzie delta area is estimated to range from about 1 to 1.5 metres in the south to 1.5 to 2 metres along the coast, with considerable local variation depending on factors such as the severity of the winter and the nature of the snow cover. Therefore, lakes deeper than 1.5 to 2 metres, depending upon local conditions, may be expected to have unfrozen bottom sediments.

The depth of unfrozen ground beneath a winter lake pool depends on factors such as: lake area, shape, depth, bottom configuration, and water temperature; the mean ground temperature and its past thermal history; the nature of the soils and disturbances induced by nearby lakes.

The only record for the depth of permafrost beneath lakes in the Mackenzie delta area has come from drilling operations of the National Research Council (Johnston and Brown, 1961) some 5 miles (8 km) southwest of Inuvik. The drill holes are in the modern alluvial delta where disturbances are introduced by adjacent bodies of water and the processes of delta sedimentation. Nevertheless, in a shallow lake some 275 metres in diameter, the centre was unfrozen to a depth of at least 73 metres, the permafrost surface plunging at a high angle from the lake edge towards the centre. Away from the lake, permafrost at least 79 metres thick was encountered. Brewer (1958 b) states that in the Point Barrow area of Alaska lakes having a diameter of about 750 metres or more, and a depth of at least 2 metres, may be expected to have unfrozen basins beneath them to a depth of at least 60 metres. In the USSR and elsewhere, unfrozen areas are known to exist beneath large lakes and rivers.

Ground Temperatures

There are no continuous long-term ground temperature records for the pingo area. North of Inuvik, a pingo at 69°02'N and 134°25'W was drilled in 1954 and thermocouples installed to a depth of 7.5 metres (Pihlainen and others, 1956). Ground temperatures at 7.5 metres were −4.1°C for August 29, 1954 and August 15, 1955. At the Ibyuk (Crater Summit) pingo near Tuktoyaktuk the temperature on July 1, 1955, at a depth of 12 metres in the ice-core was −4.4°C and that at nearby Sitiyok pingo on June 30, 1955, at 11 metres was −8.2°C (Müller, 1959, pp. 94–95). As the level of zero annual amplitude for seasonal temperature disturbances may be over 15 metres, the surface areas of pingos are not plane, and the diffusivity of pingo ice is higher than that for adjacent soils, the ground temperatures cannot be representative of normal horizontal ground conditions. However, they do permit rough estimates of ground temperatures at the base of the active layer.

At Inuvik, the mean ground temperature at a depth of 14 metres for three years of record is about −3.4°C (Brown, 1960, p. 170). However, disturbance by drilling might have affected the results slightly but probably not enough to alter appreciably their magnitude. As mean ground temperatures tend to increase with depth because of geothermal effects, the mean temperature at the top of permafrost at Inuvik might be as low as −5°C and farther north in the area with pingos the temperatures might be as low as −6°C or −7°C.

Mean Lake Temperatures

The water in arctic lakes of relatively shallow depths, such as 2 to 5 metres, is probably isothermal both vertically and horizontally during most of the ice-free period of July to September. This has been observed for the lakes in the Point Barrow area, and judging from similar conditions, there should be no appreciable summer stratification in the Mackenzie delta area. At Point Barrow, where mean annual air temperatures are as low or lower than in the Mackenzie delta pingo area, summer lake temperatures reach about 13°C (Brewer, 1958a; Livingstone and others, 1958). From September 7, 1954 to September 6, 1955, the mean annual temperature at the bottom of a lake near Point Barrow, at a point 108 metres from shore and beneath 2.4 metres of water was 1.79°C. (Brewer, 1958a, p. 284). If similar conditions are applicable to the Mackenzie delta region, mean monthly water temperatures for the ice-free months of July to September might be expected to range from slightly over 0°C to 10°C or more. As the water temperatures in an unfrozen pool must be 0°C or above during the winter months, the mean annual water temperature of an unfrozen pool can hardly be less than about 2°C.

Three-Dimensional Heat Conduction Beneath Lakes

In view of the paucity of data on permafrost beneath lakes, a theoretical approach may give information on permafrost surfaces. In the ensuing discussion, it should be apparent that certain simplifying assumptions must obviously be made to keep the study within bounds. Theoretical aspects of the general problem of three-dimensional heat conduction in a semi-infinite medium, disturbed by surface effects, are treated by Carslaw and Jaeger (1960, pp. 255–296), Birch (1950), Lachenbruch (1957; 1959) and others. In particular, Lachenbruch (1957) has discussed theoretical aspects of three-dimensional heat conduction in permafrost beneath heated buildings. The same theory may be applied to lakes, by treating them as heated buildings, and thus the temperature disturbances in the ground beneath and around them may be estimated. In Lachenbruch's study, the ground is assumed to be homogeneous, although the effects of periodic heat flow in a stratified medium have also been examined (Lachenbruch, 1959). The effects of latent heat are neglected, but this has no relevancy if only steady state or equilibrium conditions are being investigated. Seasonal disturbances may also be estimated but may be omitted in a study of pingo formation.

The symbols, given below, are mainly based upon Lachenbruch (1957):

Θ = temperature in °C

t = time since initiation of temperature disturbance in lake U

α = thermal diffusivity (cm²/sec)

Ω = solid angle (steradians)

U = finite region of circular lake of plane $z=0$

ρ,λ = polar co-ordinates in plane $z=0$

r $=[(x-x')^2+(y-y')^2+z^2]^{\frac{1}{2}}=[\rho^2+z^2]^{\frac{1}{2}}$

T_G = mean annual temperature ($T_G<0$) of undisturbed permafrost region outside U of plane $z=0$

T_L = mean annual temperature ($T_L>0$) of the lake (U)

T = temperature under the centre of a circular lake

B = constant temperature difference between a point inside the disturbed region (the lake, U) of plane $z=0$ and the mean annual ground temperature of the undisturbed region outside the lake; i.e. $B=T_L-T_G$

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I =geothermal gradient in cm/°C

R =radius of circular lake

In applying the heated building theory to ground and sub-lakebottom temperatures, let us consider the disturbance induced by placing a circular lake U of radius R on a semi-infinite flat area of horizontal terrain. As lake depths are negligible compared to the radii, the lake bottom can be taken as flush with the adjacent ground. This simplified approach is justified by the relative shallowness of the lakes in comparison with their radii and other measures arising in the computations.

The principal disturbance (Θ) is defined as the temperature disturbance resulting in the ground ($z>0$) from a constant temperature difference (B) between a point inside the disturbed area of the lake and the mean annual temperature of the undisturbed region outside of the lake.

The principal disturbance (Θ) is given by Lachenbruch (1957, p. 55, equation 26) as:

$$\Theta(x,y,z,t)=\frac{B}{2\pi}\int\int\left\{\frac{r}{\sqrt{\pi\alpha t}}e^{\frac{-r^2}{4\alpha t}}+erfc\frac{r}{2\sqrt{\alpha t}}\right\}d\Omega \tag{2}$$

where $d\Omega$ is the element of solid angle subtended by $dx'dy'$ at the field point (x,y,z); (*see* Birch, 1950, p. 586, equation 8).

Integration of equation (2) gives:

$$\Theta(x,y,z,t)=-\frac{Bz}{2\pi}\int\left[\frac{1}{r}erfc\frac{r}{2\sqrt{\alpha t}}\right]_{r_1}^{r_2}d\lambda \tag{3}$$

In the study of pingo formation, attention focuses upon the depth to permafrost, if it occurs, beneath the centre of the lake. The principal disturbance under the vertex is therefore

$$\begin{aligned}\Theta(0,0,z,t)&=-\frac{Bz}{2\pi}\int d\lambda\left[\frac{1}{r}erfc\frac{r}{2\sqrt{\alpha t}}\right]_{z}^{\sqrt{z^2+R^2}}\\&=\frac{B\lambda}{2\pi}\left[erfc\frac{z}{2\sqrt{\alpha t}}-\frac{z}{\sqrt{z^2+R^2}}erfc\frac{\sqrt{z^2+R^2}}{2\sqrt{\alpha t}}\right]\end{aligned} \tag{4}$$

which is Lachenbruch's equation 29 (*see* Carslaw and Jaeger, 1960, p. 264) for the disturbance caused by a sector of central angle λ at depth z and time t beneath the lake centre. For a circular region (Lachenbruch equation 30)

$$\Theta(0,0,z,t)=B\left[erfc\frac{z}{2\sqrt{\alpha t}}-\frac{z}{\sqrt{z^2+R^2}}erfc\frac{\sqrt{z^2+R^2}}{2\sqrt{\alpha t}}\right] \tag{5}$$

Under steady state or equilibrium conditions, equation 5 becomes

$$\underset{t\to\infty}{\Theta(0,0,z,t)} = B\left[1-\frac{z}{\sqrt{z^2+R^2}}\right] = (T_L-T_G)\left[1-\frac{z}{\sqrt{z^2+R^2}}\right] \tag{6}$$

It is interesting to note that equation 6 is completely independent of the physical constants of the ground, such as the thermal diffusivity and water content, so long as steady state conditions exist.

The temperature (T) under the centre of a circular lake is then the sum of the undisturbed ground temperature (T_G), the principal disturbance (equation 6), and the increase of temperature from geothermal effects, viz.

$$T = T_G + (T_L-T_G)\left[1-\frac{z}{\sqrt{z^2+R^2}}\right] + \frac{z}{I} \tag{7}$$

The lakes under consideration, however, freeze to the bottom in the shallower areas each winter. As the centres of lakes are usually the deepest parts, those lakes which do not freeze to the bottom in winter will usually have an outer annulus of ice adhering to the bottom, and an inner pool of water. Assuming circular uniform lakes, if the outer radius of the lake is R_2 and that for the unfrozen pool in winter is R_1, then the frozen annulus or ring of ice which adheres to the bottom each winter is of width R_2-R_1. As B is the mean annual temperature difference between a point inside the disturbed region (i.e. the lake) and that outside, $\frac{B}{2}$ can be used as a reasonable approximation for the difference between the mean annual temperature of the annulus and the undisturbed ground beyond. The principal disturbance (Θ) at depth z beneath the centre of the lake is, therefore, the sum of the disturbances due to the annulus of width R_2-R_1, and that of the unfrozen central portion of radius R_1, that is

$$\begin{aligned}\Theta(0,0,z,t) = {} & B\left[\mathit{erfc}\frac{z}{2\sqrt{\alpha t}} - \frac{z}{\sqrt{z^2+R_1^2}}\mathit{erfc}\frac{\sqrt{z^2+R_1^2}}{2\sqrt{\alpha t}}\right] + \\ & + \frac{B}{2}\left[\frac{z}{\sqrt{z^2+R_1^2}}\mathit{erfc}\frac{\sqrt{z^2+R_1^2}}{2\sqrt{\alpha t}} - \frac{z}{\sqrt{z^2+R_2^2}}\mathit{erfc}\frac{\sqrt{z^2+R_2^2}}{2\sqrt{\alpha t}}\right]\end{aligned} \tag{8}$$

(*see* Birch, 1950, p. 588, equation 18; Lachenbruch, 1957, pp. 56–57, equations 30 and 33). Under steady state conditions, equation 8 becomes

$$\underset{t\to\infty}{\Theta(o,o,z,t)} = \frac{B}{2}\left[2-\frac{z}{\sqrt{z^2+R_1^2}}-\frac{z}{\sqrt{z^2+R_2^2}}\right] \tag{9}$$

The ground temperature under the centre of the lake is then (*see* equation 7)

$$T = T_G + \frac{(T_L - T_G)}{2}\left[2 - \frac{z}{\sqrt{z^2+R_1^2}} - \frac{z}{\sqrt{z^2+R_2^2}}\right] + \frac{z}{I} \qquad (10)$$

In the heated building problem, the disturbances are relatively shallow, because of the comparatively small sizes of buildings, so geothermal effects may be neglected. However, in the case of pingo formation, where depths of 100 metres or more may be involved, geothermal effects must be taken into consideration. Assuming that the geothermal gradient is constant with depth, the increase in temperature at depth z is $\frac{z}{I}$ so the undisturbed ground temperature (T_U) at depth z in °C is

$$T_U = T_G + \frac{z}{I} \qquad (11)$$

The lower boundary of permafrost or the 0°C surface in the undisturbed area is then reached where T_U is zero or $z = -T_G I$.

Under steady state conditions, in a circular lake, the principal disturbance beneath the lake centre is given by equation 6, and the undisturbed temperature, with geothermal effects included, by equation 11. If there is a permafrost surface beneath the centre of the lake—in which case there must be an upper and lower surface—then the temperature is 0°C. Therefore, a permafrost surface occurs at a depth z where the sum of the undisturbed temperature (equation 11) and that of the principal disturbance induced by the lake (equation 6) are zero

$$B\left[1 - \frac{z}{\sqrt{z^2+R^2}}\right] + T_G + \frac{z}{I} = 0 \qquad (12)$$

Equation 12 is not a single valued function of z; therefore it is unsuited for direct determination of the top and bottom of permafrost and the equation may not be satisfied by a real value of z, as would be the case in a lake too large for permafrost to occur beneath the centre under the specified conditions.

It is evident that as a lake increases in radius from small to large, a critical radius will be reached beyond which no permafrost will be present beneath the lake centre. The radius of a lake with the top (bottom) of permafrost at specified z is obtainable from equation 12 by rearranging terms

$$R = \left[\frac{(IBz)^2}{(IB + IT_G + z)^2} - z^2 \right]^{\frac{1}{2}} \tag{13}$$

If permafrost exists beneath the centre of a lake, and the lake radius is gradually increased, with steady state conditions prevailing, a critical radius will be reached when the permafrost lens beneath the lake 'opens up', so that all the material directly beneath the lake centre is above 0°C. The critical radius is reached when $\frac{dR}{dz}$ for equation 13 is zero

$$\frac{z}{R} \left[\frac{I^3B^2(B + T_G) - (IB + IT_G + z)^3}{(IB + IT_G + z)^3} \right] = 0 \tag{14}$$

As neither z nor R can be zero, therefore

$$I^3B^2(B + T_G) - (IB + IT_G + z)^3 = 0 \tag{15}$$

and

$$z = I[(B^2T_L)^{\frac{1}{3}} - T_L]. \tag{16}$$

Equation 16 gives the maximum depth to permafrost, at the critical radius where permafrost 'opens' beneath the centre of a lake under steady state conditions. It is interesting to note that neither R nor the physical constants of the soil enter into equation 16. The maximum depth to permafrost depends only upon the interrelationships among the geothermal gradient, mean temperature of the lake, and mean temperature of the undisturbed ground. Once z is determined from equation 16, substitution into equation 13 gives the required radius of the lake under steady state conditions.

In Figure 6, equation 6 is applied for a lake 150 metres in radius. For example, if the mean annual ground temperature (T_G) is taken as −7°C and the mean annual lake temperature (T_L) as 3°C, then the value of B is 10°C. Under the assumed conditions for the computations, the warming effect of the lake under steady state conditions at a depth of 100 metres is about 4.5°C and at 200 metres about 2°C. Thus, temporarily omitting geothermal effects, the ground temperature at a depth of 100 metres in the centre of the lake would be −2.5°C, and at 200 metres −5°C. A geothermal gradient of 1°C per 44 metres has been added to the data shown in Figure 6 to produce Figure 7. The result can be obtained graphically merely by adding the geothermal rise in temperature, at any depth, to the values plotted in Figure 6. By using the same ground and lake temperatures as given above, the sum of the warming effect of the lake and the geothermal effects for a B of 10°C is 7°C at a depth of 80 metres and again at 240

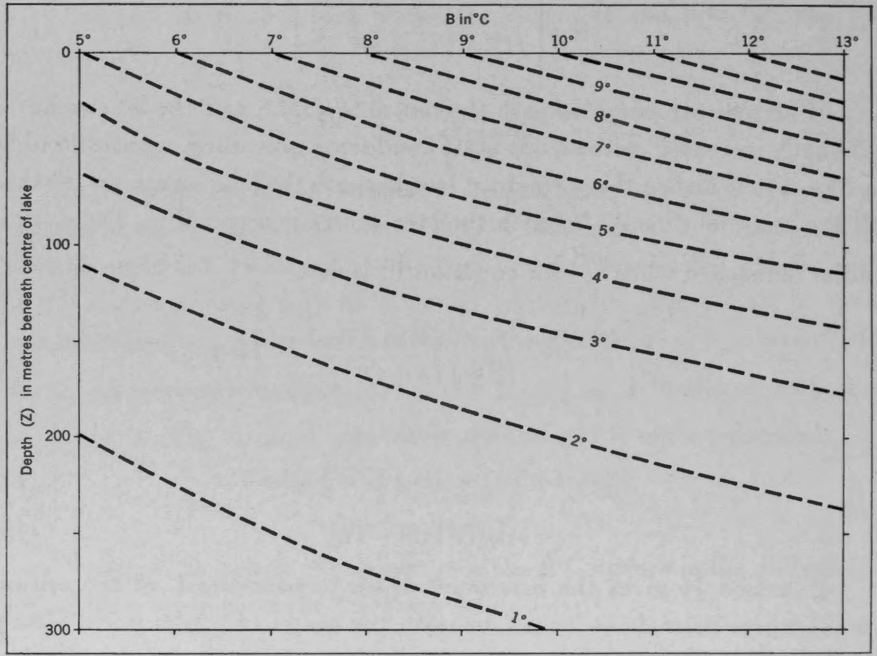

Figure 6 Nomogram for the disturbance under the vertex of a circular lake of 150 metre radius.

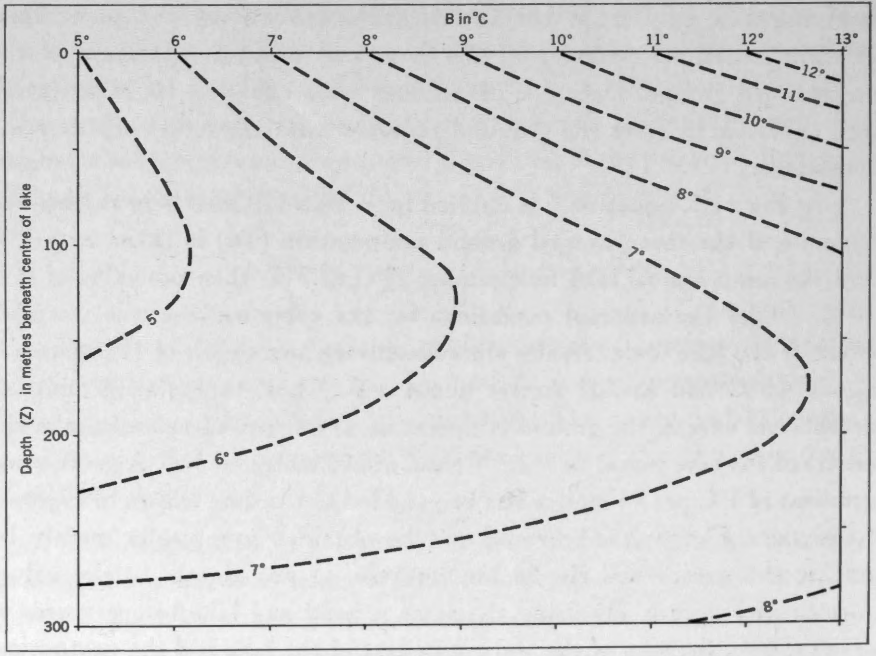

Figure 7 Nomogram for the disturbance under the vertex of a circular lake of 150-metre radius with a geothermal gradient of 1° C per 44 metres.

metres. As the undisturbed ground temperature is taken at $-7°C$, this means that the top 0°C isotherm (top of permafrost) would be encountered at 80 metres and the lower 0°C isotherm (bottom of permafrost) at 240 metres. However, if the mean ground temperature (T_G) were $-5°C$ and the lake temperature (T_L) 2°C, with B of 7°C, then no permafrost would lie beneath the lake centre.

Figure 8 is plotted from data derived from equation 16, the diagonal dashed lines showing the maximum depth to permafrost under given temperatures and geothermal gradients. Again, using B of 10°C, T_G of $-7°C$, T_L of 3°C, and I of 44 metres per 1°C, Figure 8 shows that the maximum depth to permafrost beneath the lake centre is about 170 metres; permafrost 'opens out' with any increase in lake size after that.

Figure 9 has been plotted for B of 12°C, T_G of $-10°C$, T_L of 2°C, and I of 44 metres per 1°C. The data used are approximately that of conditions at Point Barrow, Alaska, the temperatures of land and water being lower than might reasonably be expected in the pingo area under discussion, but possibly no colder than during some postglacial periods. The curve is plotted from equations 13 and 16. Thus, with a lake of 200 metres radius, the top and bottom of permafrost beneath the centre of the lake would be expected to occur at about 60 and 375 metres, respectively. A lake of 300 metres in radius would have a thin permafrost 'lens' in the centre, a lake of 350 metres in radius, no permafrost.

Figures 10, 11 and 12 have been computed for equation 9 for depths (z) of 100, 200, and 300 metres for various combinations of R_1 and R_2, the isolines being in units of B. Thus, in Figure 10, using R_1 of 150 metres and R_2 of 200 metres, the disturbance at a depth of 100 metres is about .50B, from Figure 11 at 200 metres, about .25B, and from Figure 12 at 300 metres about .135B. If a B of 10°C, T_G of $-7°C$, and T_L of 3°C are used as before, the disturbances at 100, 200, and 300 metres are about 5.0°C, 2.5°C, and 1.35°C. These values may be compared with those from Figure 6, viz. 4.5°C, 2.0°C, and 1.0°C. The slightly greater disturbances of about .5°C, derived from Figures 10, 11 and 12 are the steady state effects of the 50-metre annulus which freezes seasonally to the bottom. Figures 10, 11 and 12 make it possible to estimate steady state disturbances beneath lakes of varying sizes with varying seasonally frozen annuli. Geothermal effects can be added to the warming disturbance of the lake to obtain an estimated ground temperature.

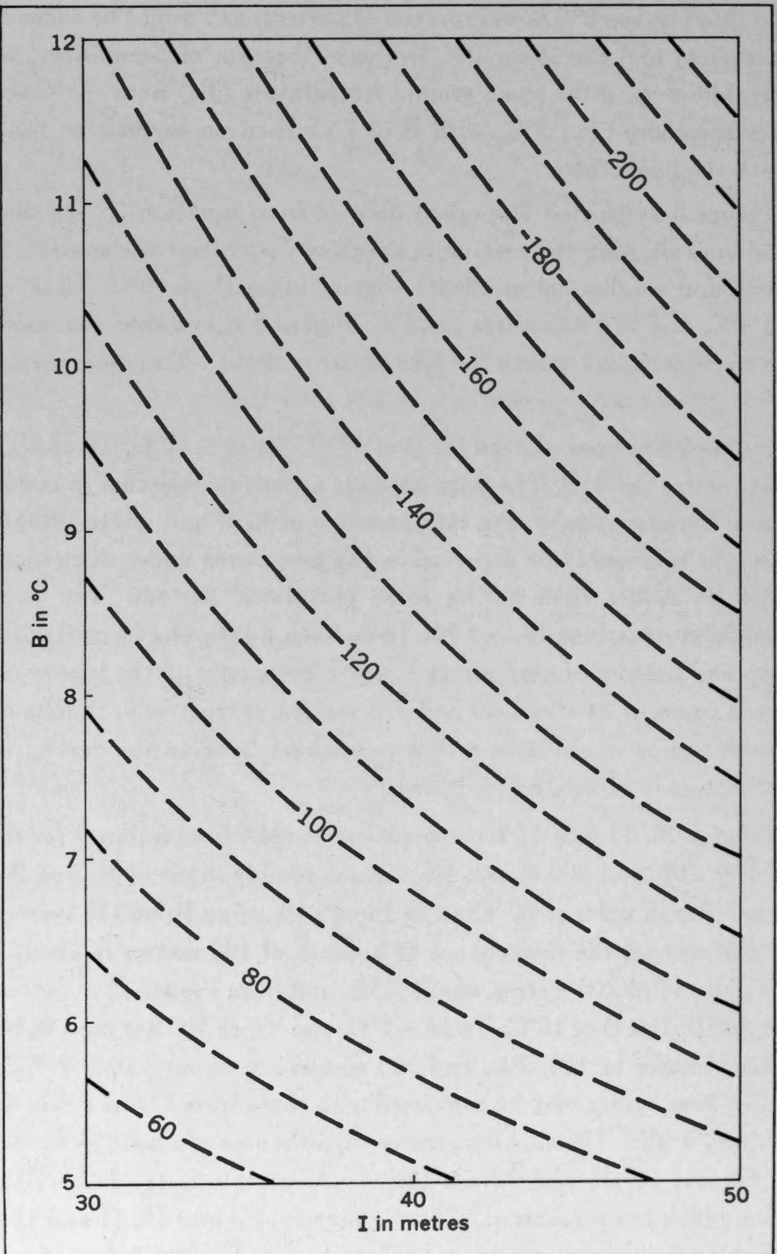

Figure 8. Nomogram for the maximum depth to permafrost beneath the vertex of a circular lake under given temperatures and geothermal gradients.

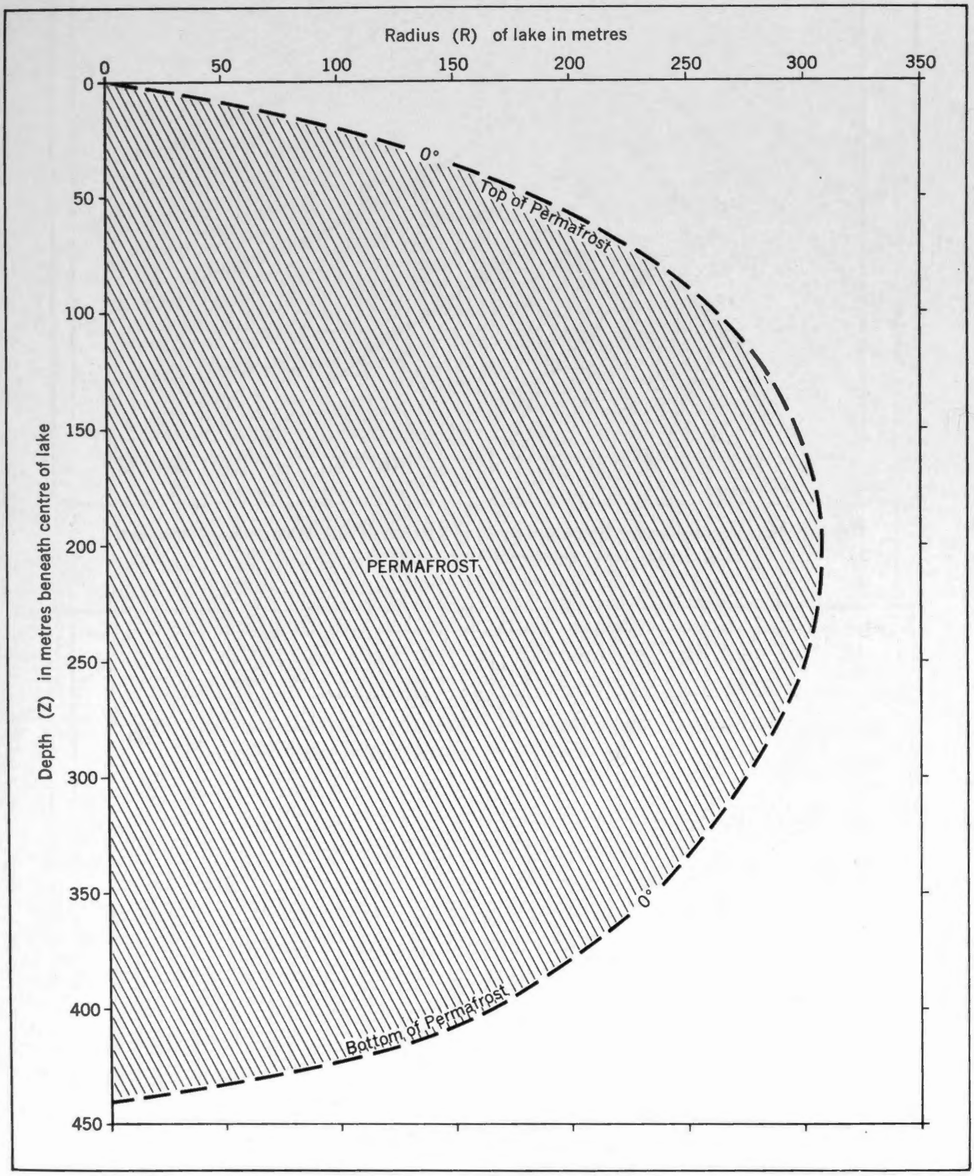

Figure 9. Inferred position of top and bottom of permafrost (under steady state conditions) beneath the centre of a lake of given radius for a ground temperature of $-10°C$, lake temperature of $2°C$, and geothermal gradient of $1°C$ per 44 metres.

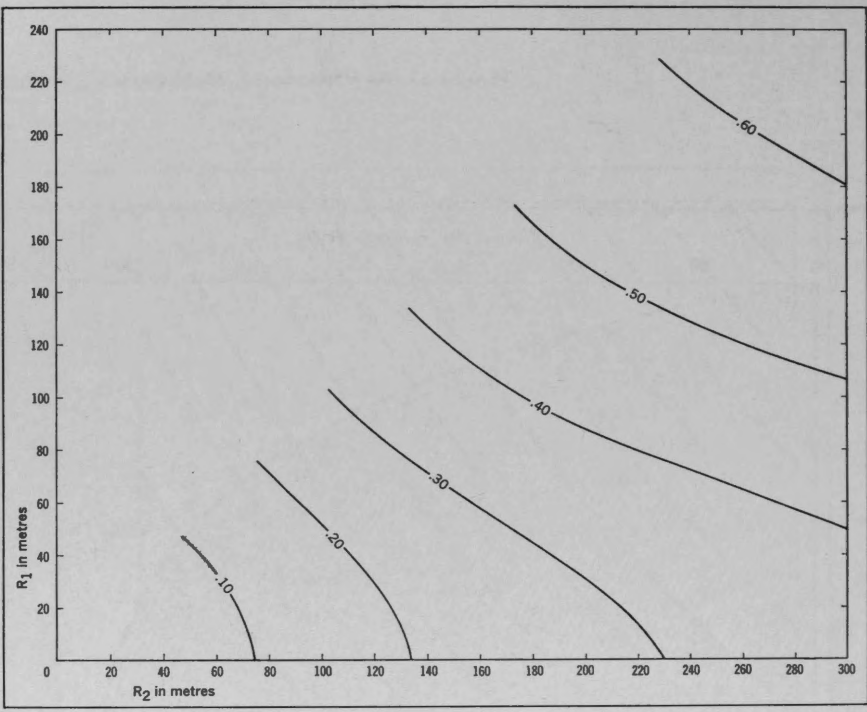

Figure 10. Nomogram for the disturbance, at a depth of 100 metres beneath the vertex of a circular lake, expressed in units of B.

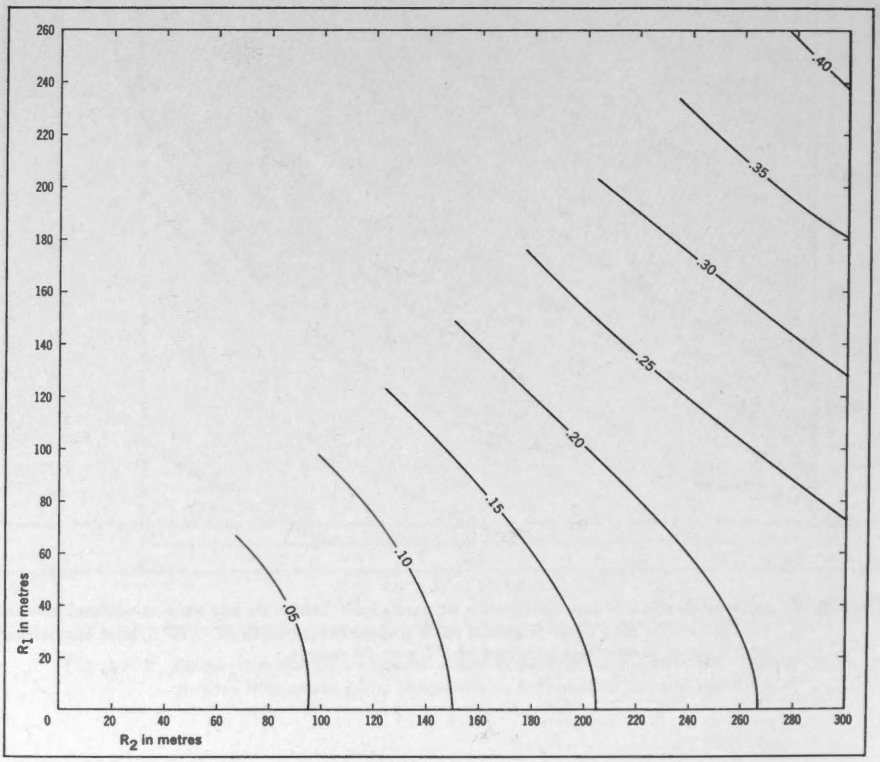

Figure 11. Nomogram for the disturbance, at a depth of 200 metres beneath the vertex of a circular lake, expressed in units of B.

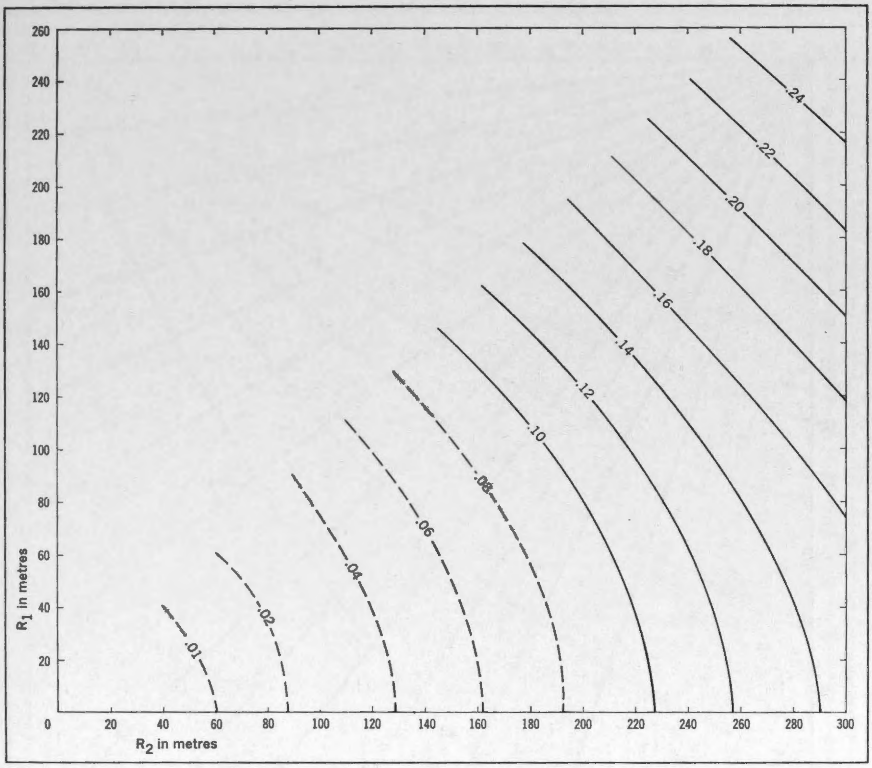

Figure 12. Nomogram for the disturbance, at a depth of 300 metres beneath the vertex of a circular lake, expressed in units of B.

In the preceding discussion, steady state conditions have been assumed. However, it seems most unlikely that such an equilibrium stage existed when pingos formed, not just because of post-glacial climatic and geomorphic changes, but also because of the relatively short span of time involved. Figure 13 shows steady state conditions (equation 6) scaled in terms of units of B, whereas Figure 14 shows the approach to equilibrium conditions (equation 5) at the end of 1,000 years, assuming a diffusivity of $\alpha = 0.01$ cm^2/sec, latent heat being neglected. A comparison of the two graphs shows that the differences between curves for the same units of B are relatively minor. At the end of 1,000 years, dry soil beneath a lake should be close to equilibrium temperatures, other things being equal.

So far, the discussion of the distribution of permafrost beneath lakes has been based upon lakes of circular shape. The method, however, is also applicable to lakes of irregular shape. Such lakes can be subdivided into a series of assumed uniform sectors which radiate from a vertex beneath

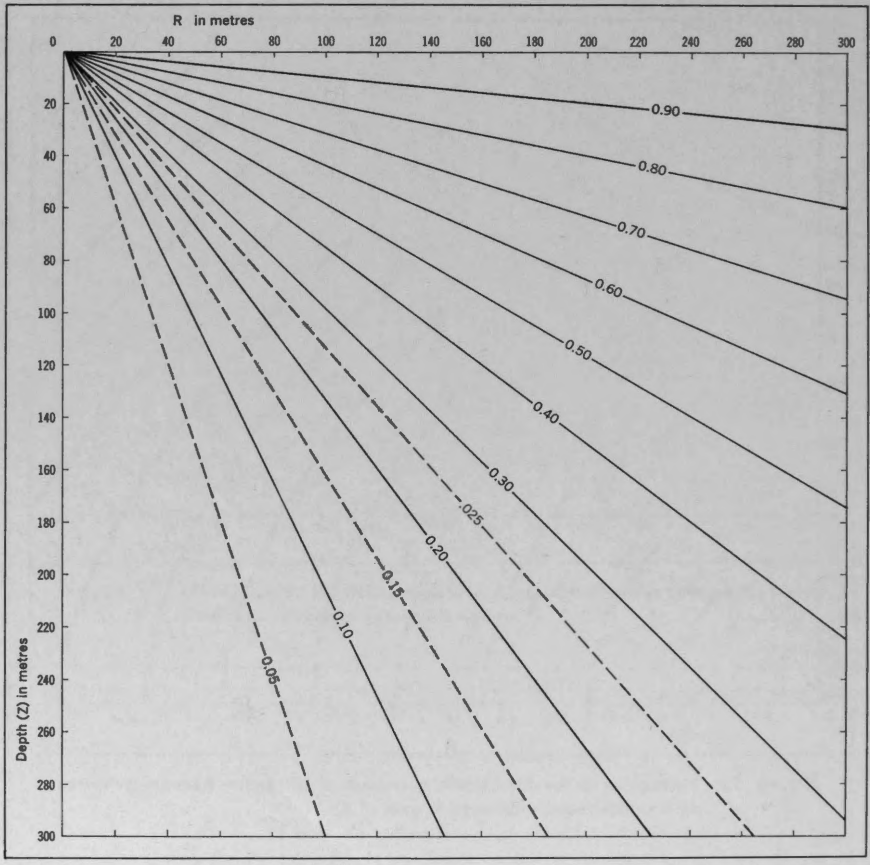

Figure 13. Nomogram for the disturbance (expressed in units of B) beneath the vertex of a circular lake under steady state conditions, assuming no frozen annulus.

which the temperature conditions are required. A sector of central angle λ would then contribute a fractional disturbance approximately equal to $\frac{\lambda}{2\pi}$ of the total disturbance of a lake with properties similar to that of the sector. Thus, each sector may be evaluated separately, and then the results summed, using equation 8 for transient conditions and equation 9 for steady state conditions. Graphical methods can also be applied (*see* Birch, 1950; Lachenbruch, 1957). Figures 10, 11, and 12 may be used to determine approximately the disturbances at depths of 100, 200, and 300 metres. For example, as previously given with R_1 of 150 metres and R_2 of 200 metres, the disturbance at a depth of 100 metres (from Figure 10) is about .50B. For a sector with central angle $\lambda=36°$, the disturbance would be $.50B\frac{\lambda}{2\pi}$ or

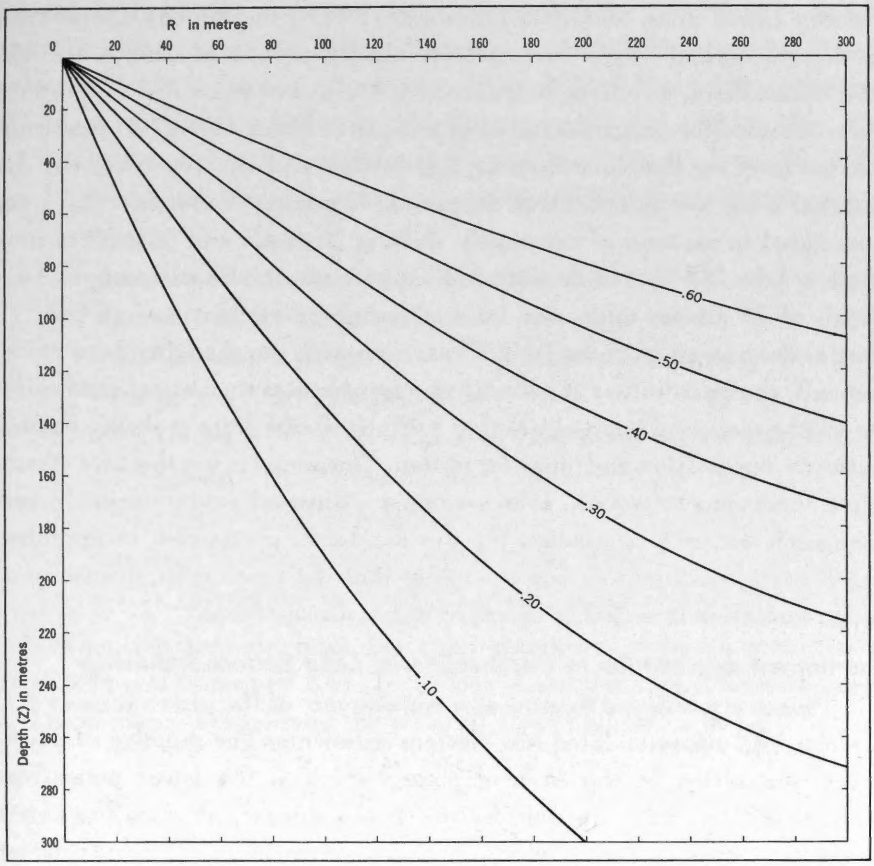

Figure 14. Nomogram for the disturbance (expressed in units of B) beneath the vertex of a circular lake at the end of 1,000 years assuming a dry soil of diffusivity of 0.01 cm^2/sec, and no frozen annulus.

.05B. The values determined for the other sectors would be similarly determined, and the results added.

Because information on ground temperatures at the time of glacial retreat is unknown, it seems pointless to attempt to analyze transient effects in detail. For example, the ground surface beneath the ice may have been unfrozen, or it may have been frozen to a depth of tens of metres. If the ground were unfrozen, then freezing would have to be considered; if frozen, then thawing beneath a deep lake would take place. Latent heat effects would also be involved, its role being to delay thawing or freezing, depending upon temperature conditions.

Deglaciation of the pingo area along the coast at Tuktoyaktuk took place prior to 12,000 years ago (Müller, 1962) and in the eastern part of the

Eskimo Lakes prior to 11,000 years ago (I-483), the information coming from radiocarbon dates for material near the tops of pingos. In the Mackenzie delta, a radiocarbon date of 6,900 $\pm$ 110 years (GSC -54) has been obtained for organic material at a depth of about 32 metres below present sea level for bore hole number 3 (Johnston and Brown, 1961). In the interval since the material was deposited, 38 metres of sediment have accumulated in an area of constantly shifting channels and lakes. Yet in a shallow lake 275 metres in diameter, no permafrost was encountered to a depth of 73 metres under the lake, although permafrost was at least 79 metres deep away from the lake. Thus, even with the changing delta environment, the distribution of permafrost approximates theoretical conditions. As will be discussed later, a period of 5,000 to 10,000 years probably elapsed between deglaciation and pingo formation. The period is too short for steady state conditions to pertain, even assuming a constant environment. Nevertheless, it seems long enough for the depths to permafrost to approach equilibrium conditions closely enough so that the geomorphic discussion of pingo formation is scaled to the right order of magnitude.

Permafrost Aggradation in Unconsolidated Lake Bottom Sediments

Pingos are believed to grow as a consequence of the penetration of permafrost into unconsolidated lake-bottom sediments. The rapidity of downward penetration of the freezing plane, which is the lower permafrost surface, will depend upon such factors as the climate, physical properties of the sediments, topography, vegetation cover, presence of nearby water bodies, and so forth.

In the present discussion, only the thickness of bottom sediment corresponding to the cover over pingo ice-cores is considered. The bottom sediments, as exposed along wave-cut bluffs and in pingos, are of granular materials, predominantly in the sand range.

Prior to downward penetration of permafrost in unfrozen lake bottom sediments, all the sediments would be saturated. Therefore, the latent heat of fusion would be important in delaying the downward penetration of the freezing plane, although it has no effect upon the equilibrium temperature profile. To simplify the problem, first assume that the unfrozen sediment, at the instant that permafrost commences to develop in the lake centre, is at 0°C for a depth of 10 or 15 metres. This does not seem to be an unreasonable assumption, because the mean bottom temperatures for a depth of 10 or 15 metres would normally be close to the mean lake water temperature which, at commencement of permafrost, would be about 0°C.

If the preceding assumptions are valid, then the general approach to the rate of freezing may be derived from theoretical studies involving change of state in the formation of ice, viz. Stefan's and Neumann's solutions (Carslaw and Jaeger, 1960, pp. 282–286; Ingersoll and others, 1954, pp. 190–199). The approach used below is adapted from that by Terzaghi (1952, pp. 39–44) for the rate of thawing beneath heated buildings. All units are in the cgs system.

T_T = mean surface temperature at the top of permafrost
η = porosity of the frozen ground
Υ_s = 2.65 gm cm^{-3} average unit weight of the soil particles
Υ_i = 0.9 gm cm^{-3}, unit weight of ice
c_h = 0.2 cal gm^{-1} (°C)$^{-1}$, average heat capacity of the soil particles
c_i = 0.5 cal gm^{-1} (°C)$^{-1}$, average heat capacity of the ice
L = 80 cal gm^{-1}, heat of fusion of ice
k_s = average thermal conductivity of frozen ground
k_i = 5.3×10^{-3} cal cm^{-1} sec^{-1} (°C)$^{-1}$, average thermal conductivity of ice

When the ground freezes, one cubic centimeter of frozen ground contains 0.9η gms of ice and $2.65\ (1-\eta)$ gms of soil particles. The latent heat of fusion released in freezing the water in 1 cc of ground is in c.g.s. units

$$q_e = L\eta\Upsilon_i = 72\eta \tag{17}$$

The quantity of heat given off in the cooling of 1 cc of frozen ground by 1°C is about

$$q_t = \Upsilon_s(1-\eta)c_s + \Upsilon_i\eta c_i \tag{18}$$

As the top of the permafrost, which is the lake bottom, is at temperature T_T and the bottom at zero, the mean temperature of the permafrost layer is approximately $.5T_T$. When permafrost is of thickness z the total quantity of heat which has been released is about

$$Q = (q_e + 0.5T_{T_{qt}})z = Az \tag{19}$$

If the permafrost layer is thin enough so that the temperature gradient can be considered as uniform between the top of the permafrost (bottom of the lake) and the bottom of the permafrost, the quantity of heat flowing upward, in time *dt*, will be

$$q = -\frac{k_s T_T}{z} dt$$

but this must equal the average rate of heat released (Q) when the permafrost increases in thickness by dz. As

$$dQ = A\,dz$$

therefore

$$dQ = q$$

and

$$A\,dz = -\frac{k_s T_T}{z} dt$$

hence

$$z = \left(\frac{-2tk_s T_T}{A} \right)^{\frac{1}{2}} \tag{20}$$

If the heat capacity of the soil particles and ice in the permafrost layer are neglected, equation 20 is essentially Stefan's approximate solution for the freezing of ice (Ingersoll and others, 1954, p. 198). Unfortunately, the type of derivation given by equation 20 is not satisfactory for thick layers of frozen ground (Redozubov, 1946, translation of 1954, pp. 9–10). However, the approach may give comparative information on rates of freezing of relatively thin permafrost layers, of thickness relevant to the origin of pingos.

Equation 20 may be further simplified. The value of A (from equations 17, 18 and 19) is

$$A = 72\eta + 0.5T_T(0.53 - 0.08\eta) \tag{21}$$

The range of porosities encountered in the sands is unlikely to exceed $.30 \pm .15$ and T_T is unlikely to exceed greatly $-10°C$, a ground temperature considerably colder than present temperatures. As a rough approximation, therefore,

$$A \sim 72\eta \tag{22}$$

because the second term in equation 21 is relatively small in comparison to the first. Therefore

$$z \sim \left(\frac{-2tk_s T_T}{72\eta} \right)^{\frac{1}{2}} \tag{23}$$

The thermal conductivity of frozen soil is highly variable, but ranges are from about 3.0×10^{-3} for frozen organic silty clay, 5.0×10^{-3} for a saturated frozen clay, 5.3×10^{-3} for pure ice, 6.0×10^{-3} for an icy silt, to 9.0×10^{-3} cgs. units for a dense frozen sand (Lachenbruch, 1959, p. 10; Terzaghi, 1952, p. 9). As most of the material above the ice is of sand, the conductivity of the soil may range from about 6.0×10^{-3} to 9.0×10^{-3} cgs units.

From Equation 23, it is evident that the depths of penetration of permafrost in two soils for the same time and temperature are about proportional to the square root of the conductivity divided by porosity for each soil. A substitution of the range of values which might be expected for ice and frozen soil indicates that the thickness of ice (z_i) which might develop would be from about one-third to two-thirds that of frozen sand (z_s) under similar temperature and time conditions

$$z_i \sim (\tfrac{1}{3} \text{ to } \tfrac{2}{3}) z_s \tag{24}$$

and a saturated soil with a high porosity would freeze much more slowly than a soil with low porosity.

Under actual field conditions, three factors would seem to operate in combination to produce a dome-shaped lower permafrost surface at the site of pingo growth during the aggradation of permafrost in a lake basin.

Firstly: if in an otherwise uniform, flat, homogeneous lake bottom area, ice formation commenced at one spot during the downward penetration of permafrost, equation 24 shows that the penetration of the lower permafrost surface would be slowed at the site of ice segregation so a dome-shaped lower surface would tend to result. Thus, under conditions of uniform thickness of permafrost, local segregation of ground ice would have a tendency towards warping the lower permafrost surface into the form of a dome.

Secondly: most lakes are deepest towards the centre, where pingo growth normally occurs, and shallowest near the shores. Thus, if the water gradually shoals, and permafrost commences to aggrade in an unfrozen lake basin, permafrost will have had the longest time to grow near the shore, the shortest in the centre, thus contributing towards a dome-shaped lower surface.

Thirdly: a pingo protruding above the lake bottom alters the original flat lake bottom topography. The exact effects are impossible to specify, since ground temperatures, position of the freezing plane, size of pingo, snow cover, etc., are involved. However, in general, the formation of a mound would be expected to result in an upward bending of isotherm surfaces beneath the mound; approximate positions may be obtained by estimating the bending of the isotherms from relaxation methods of mathematics.

Consequently, as a result of the three factors discussed above, the penetration of permafrost in a lake basin should be uneven. A dome-shaped lower permafrost surface should form in most basins, the arch of the dome being near the lake centre (Figure 15B). If, in addition, ice segregation commenced in the top of the dome, the shape should be accentuated. The relevancy of the dome-shape to pingo formation is discussed later.

Hydrostatic Pressure

The generally accepted theory for pingo formation depends upon updoming by hydrostatic pressure. The hydrostatic pressure may result from a hydraulic head associated with slopes (East Greenland type) or pressure resulting from volume expansion of water on freezing as permafrost advances into a closed unfrozen pocket (Mackenzie type).

Laboratory and field experiments carried out on the freezing of saturated sands shows that excess pore water is forced to recede before constantly growing ice crystals of an advancing freezing plane (Balduzzi, 1959; Kosmachev, 1953; Petrov, 1934). The surplus water is pressed into the unfrozen material, thus building up a hydrostatic pressure. The minimum permeability required for excess water to be squeezed out has been estimated at about 0.1 metre per day (*Foundations of Geocryology*, Vol. 11, 1959, pp. 33–35). The soils in which the pingos have grown are non-cohesive and of a grain size suitable for the rapid expulsion of pore water.

If V is the volume of unfrozen material in a closed system and η the average porosity, then the volume of expelled pore water (V_e) cannot exceed about a tenth the volume of ice.

$$V_e < 0.1\ \eta V$$

The amount of expelled pore water could be considerably less than $0.1\eta V$, from such causes as local ice lensing in thin silty laminae and soil grains being pushed apart by some in situ freezing.

Development of a Closed System

The development of a closed system whereby expelled pore water is trapped under pressure would seem to require only the growth of a continuous permafrost seal on the lake bottom (Figure 15 B). Expelled pore water could, therefore, not escape upward through the impermeable permafrost seal; it could not go sideways because of permafrost extending out from the shore beneath the unfrozen sediments; and it could not escape downwards, because of either the presence of permafrost, saturated soils, or both. There is no need to have the unfrozen material, beneath the surface cover, completely enclosed by permafrost in order to entrap expelled pore water, because the unfrozen sediments below the lake would already be saturated so there would be no easy downward escape for expelled pore water. It would be most difficult for expelled water to escape at depth beneath the permafrost of adjacent areas and out to sea, especially since geomorphic evidence suggests relatively impermeable sediments beneath the sands.

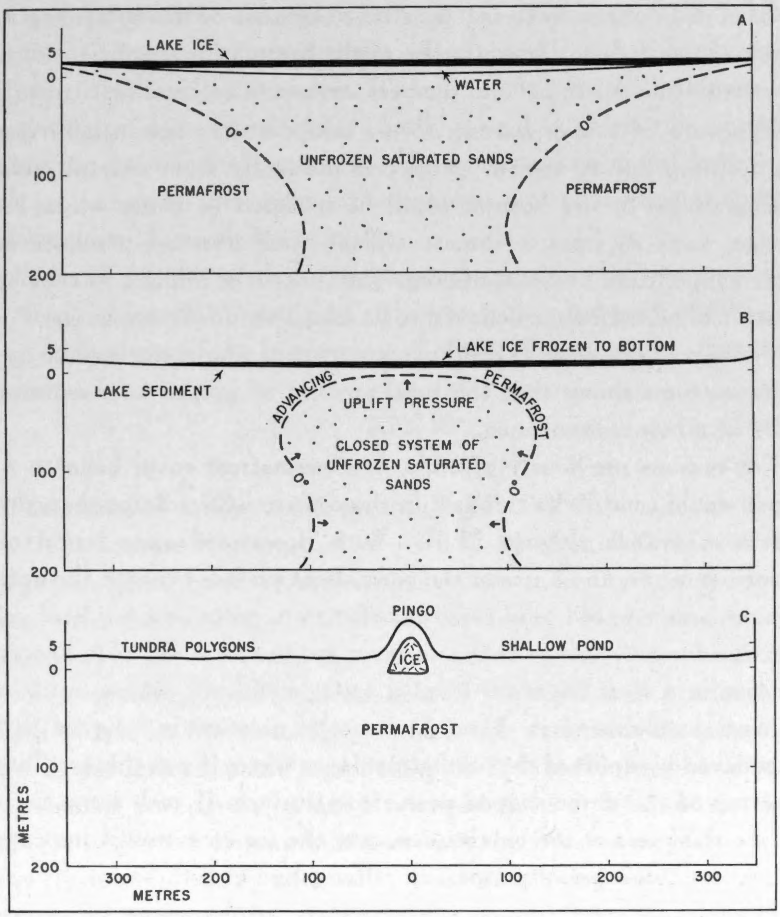

Figure 15. In diagrams A, B and C a vertical exaggeration of five has been used for the height above zero in order to show the lake ice and the open pool of water. Figure 15 A. A broad shallow lake has an open pool of water in winter with a frozen annulus around it. No permafrost lies beneath the centre of the lake. Figure 15 B. Prolonged shoaling has caused the lake ice to freeze to the bottom in winter and has induced downward aggradation of permafrost. Infilling has raised the lake bottom slightly. The deepest part of the lake, which has the thinnest permafrost, is later domed up to relieve the hydrostatic pressure. Figure 15 C. The pingo ice-core, being within permafrost, is a stable feature. The old lake bottom is occupied by tundra polygons and shallow ponds. Because of scale changes in the diagram, the volume of the ice-core should not be construed as showing a direct relationship to the initial volume of unfrozen material.

Working Hypothesis of Pingo Formation

Pingos are believed to have formed primarily in lake basins with unfrozen central portions which subsequently acquired a surface layer of permafrost to form a closed system. The initiation of permafrost along a lake bottom would seem to require either exposure of the bottom to air, as by drainage, or freezing of winter ice to the bottom for a sufficient duration each year so that mean bottom temperatures were below 0°C (Figure 15 B).

Exposure of a lake bottom to air temperatures has usually resulted from draining due to stream erosion or indirectly from coastal recession. Freezing of ice to the bottom could be initiated in many ways, besides drainage, such as from a climate change with lessened precipitation or greater evaporation or from infilling. The process of infilling in causing the formation of permafrost is believed to be relatively unimportant (*see* Porsild, 1938; Müller, 1959, p. 99). Field examination of numerous coastal sections of lake bottoms shows that the total amount of infilled lake sediments is usually of minor consequence.

For reasons previously given, a new permafrost cover beneath a lake bottom would tend to be thinnest in the centre, with a dome-shaped lower permafrost surface (Figure 15 B). With downward aggradation of the permafrost cover, and a rise of the permafrost surface beneath the unfrozen sediment, water would tend to be expelled from pores as a result of volume expansion due to freezing. As the unfrozen sediments would be in a saturated condition in a closed system, surplus water could not escape until 'room' was made available to it. Thus, hydrostatic pressure is believed to have been relieved by uplift of the permafrost layer where it was thinnest, namely at the top of the dome-shaped permafrost surface. If such were the case, then the thickness of the overburden over the ice-core would indicate the approximate thickness of permafrost when pingo growth began.

In the larger lake basins, the central portions could be completely unfrozen at all depths, although permafrost would surround the unfrozen centre (Figure 15 A). If a permafrost surface cover were to develop, the unfrozen portion would be shaped like an hourglass. Inward migration of the permafrost neck of the hourglass could produce two closed systems. Therefore, pore water could be expelled from a number of 'different' permafrost surfaces.

Lakes with irregular bottoms might have more than one deep pool, so that more than one dome-shaped lower permafrost surface might develop. Multiple pingos are believed to have grown in such lakes. Likewise, if the deeper pools were off-centre, the pingos would likewise be off-centre.

As shown in the theoretical discussion of permafrost, the rate of formation might be taken as proportional to the square root of time (equation 23). Thus the rate of downward aggradation of permafrost would be relatively rapid when perma frost was thin (e.g. .5 metres) and the temperature gradient high, but it would be relatively slow, when permafrost had extended down to greater depths, e.g. 50 metres. Therefore the expulsion of pore water, from downward freezing, would gradually slow down from this cause alone. In addition, the volume of unfrozen sediment would normally decrease with depth, so that equal increments of permafrost growth might not contribute equal amounts of expelled pore water.

As an illustration of the probable slowness of permafrost growth, consider the freezing of 15 metres of sediment with a porosity of 30 per cent and a conductivity of the frozen ground of 9×10^{-3} cgs. units. Equation 23 becomes

$$t \backsim \frac{2.5\times10^{9}}{-T_T}$$

For a ground surface temperature (T_T) of $-5°C$, it is about 16 years; for $-1°C$, t is about 80 years; and for $-0.1°C$, t is about 800 years. Although the figures cannot be more than approximate ones, they do suggest that during prolonged shoaling of a lake, whether through geomorphic or climatic causes, downward penetration of permafrost will be slow, because mean lake bottom temperatures will oscillate around 0°C and then gradually drop below it as lake ice freezes for longer and longer periods to the lake bottom. (Note: in the derivation and use of equation 23, above, the ground beneath the advancing permafrost surface is assumed to be at 0°C, and geothermal effects are neglected. It may be shown that the relative magnitudes of the results are not appreciably affected by these simplifying assumptions, except for temperatures near 0°C).

With the development of permafrost on the lake bottom the permafrost surface beneath the unfrozen sediments would gradually rise in response to lowered temperatures. However, it is clear that the temperature gradient is much greater in the initial stages of permafrost aggradation, in the lake bottom layer of permafrost, and therefore its contribution to the expulsion of pore water should normally be the greater.

Pingo ice-cores seem, in general, to be of nearly pure ice, quite unlike the dirty tabular ice sheets which may occur nearby. The implication is that pingo ice forms from the freezing of a pool of water. However, there is no reason to believe that pools of water of the sizes of the ice-cores ever existed at one time, except, perhaps, in the smallest pingos. Rather, pingo ice-cores are believed to have grown slowly over many years or decades by the constant freezing of a replenished pool.

The rate of pingo growth cannot be rapid, except in very special cases. There are reports from Russia (Suslov, 1961, p. 144) of a growth of 1.5 to 2 metres for small pingos in 20 years with larger pingos reputed to develop in several dozen years, but no verifiable reports are available for the Mackenzie area. However, a consideration of the volume of pingo ice can lead to an estimate of the volume of unfrozen material required to produce the pingo, and hence to the downward penetration of permafrost. For example, if the ice-core in Ibyuk pingo is assumed to be a right cone about 40 metres high with a base 70 metres in radius, the volume would be approximately 200,000 cubic metres. If this represented a 10-per-cent-volume expansion of soil with 30-per-cent porosity, the required volume of unfrozen material would have been about 7,000,000 cubic metres. Assume that the unfrozen material was in a circular lake and that it was in the shape of a right cone with a slope of 45°. If so, the radius and depth of the cone would have been 190 metres. The freezing of such an unfrozen cone would have taken many decades. If the radius of the cone was taken as 300 metres, the depth of the cone would have been 75 metres. Again, freezing to such a depth, to which should be added the frozen layer which existed prior to the development of doming, would be long.

In Figure 16, the graph shows the volume of unfrozen sediment, assuming a 30-per-cent porosity and 10-per-cent-volume expansion, to form conical pingo ice-cores of given radii, where the radii are equal to the heights. Large pingo ice-cores involve very large volumes of unfrozen sediments. In

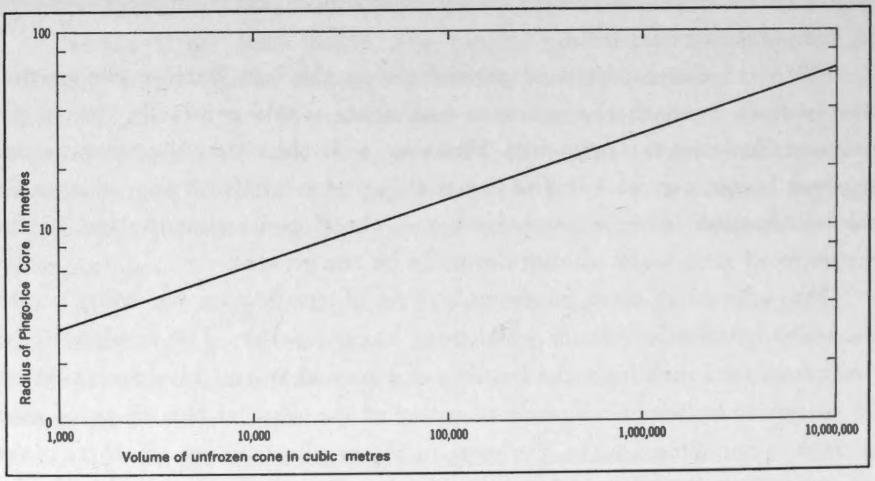

Figure 16. The graph shows the volume of sediment, with 30-per-cent porosity and 10-per-cent-volume expansion, required to form conical ice-cores with radii equal to the heights.

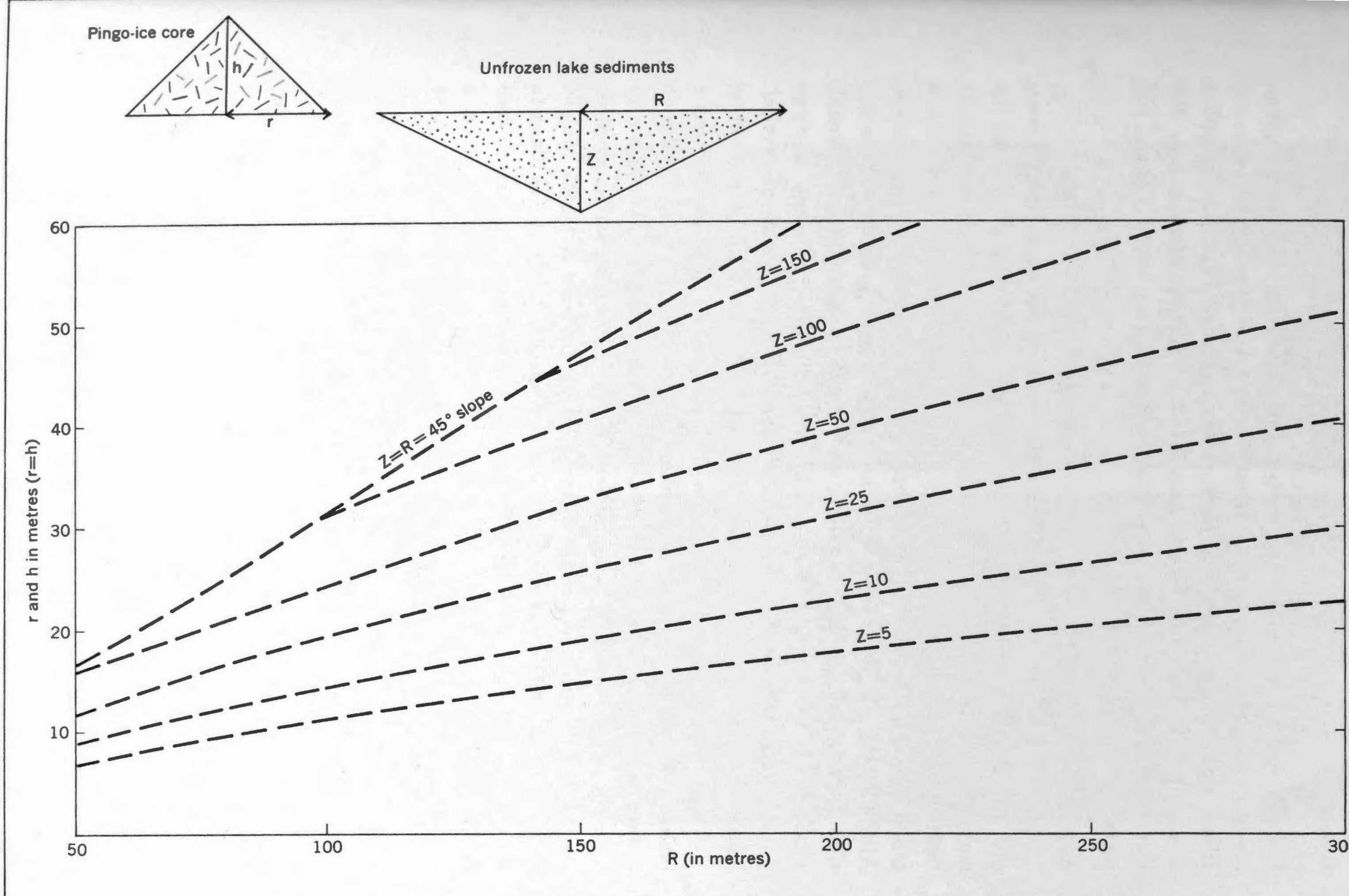

Figure 17. Nomograph for determining the dimensions of a conical volume of unfrozen sediment, given a 30-per-cent porosity and 10-per-cent-volume expansion, required to form conical ice-cores of radii equal to the heights.

Figure 17, the graph shows the relative dimensions between conical pingo ice-cores and conical unfrozen sediments, given a 30-per-cent porosity and 10-per-cent-volume expansion. Thus, a pingo ice-core 30 metres in radius and 30 metres high might have been formed from an unfrozen cone 300 metres in radius and 10 metres deep, or 200 and 25 metres, or 100 and 100 metres.

Age of Pingos

Most of the pingos are probably hundreds, if not thousands of years old. As long as permafrost is thicker than the height of a pingo, and the surface cover remains intact, a pingo should be a relatively permanent feature of the postglacial landscape. Under present climatic conditions, a pingo should collapse only through breaching of the protective overburden. Although the Eskimo names for some pingos suggest growth, as "the one that is growing" or "the poor thing that is getting to be a pingo" (Porsild, 1938, p. 52) there are, so far as is known, no reliable accounts of pingo growth either witnessed by Eskimos, or traceable back to a specific period. The pingo appearing in Richardson's book of 1851 appears unchanged today.

The vegetation cover of pingos with humus, old willows, and other plants attest to an age of at least several hundred years. The accumulation of humus is very slow, so that the presence of a turf mat at the surface may indicate an age of several thousand years. Tundra polygons, with ice-wedges exceeding 2 metres in width, are common features of large and small pingos. In general, the pattern of tundra polygons is continuous from the pingo to the surrounding area, showing that formation of the polygonal pattern had taken place prior to pingo growth. Studies of ice-wedge growth suggest a rate of about 0.5 to 1 metre per thousand years (Black, 1952) for the Point Barrow area and if rates of growth in the Mackenzie area are approximately the same, the widths of the larger ice-wedges point to ages of at least several thousand years. Some pingos are surrounded by high-centred peaty polygons whose peat feathers out against the pingo sides. The feathering out of the peat, as shown in sectioned pingos, suggests that the peat accumulated after the pingo had grown. Two radiocarbon dates for peat deposits, one near Inuvik (GSC—25) and the other for the mid-Eskimo Lakes area (GSC—16) give an average rate of growth of about .3 to .5 metres per thousand years. On this basis, a typical sectioned pingo in the Eskimo Lakes (69°32'N,131°41'W) is three to five thousand years old. Similar estimates can be made for many other pingos. Along the Arctic coast and along that of the Eskimo Lakes, coastal recession has cut back

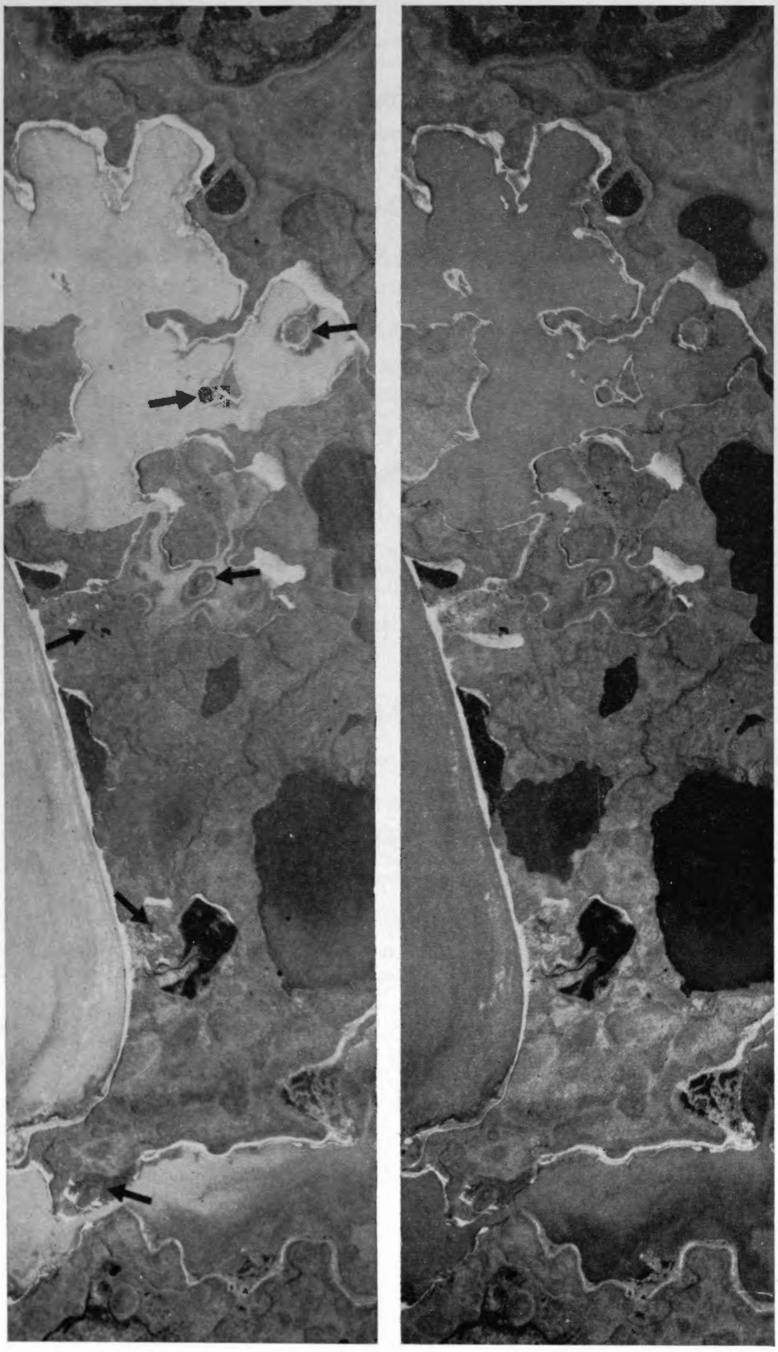

Figure 18. The stereo pair shows the characteristic association of pingos with depressions, some of which are completely infilled with tundra polygons, others in lakes connecting with the sea. The water bodies are all shoal. The hills, with ridges, have extensive ground ice in the form of ice wedges and sheets. The general location is southwest of Tuktoyaktuk at 69° 23'N, 133° 15'W. (RCAF photos A 12918—98 and 99)

into numerous lakes which contain pingos. As the pingos must have formed prior to cliff recession, which may amount to several thousand feet, again an old pingo age is suggested. Müller (1962) places the minimum ages of two pingos near Tuktoyaktuk at 7,000 and 4,000 years.

The assembled evidence from vegetation types, humus thickness, widths of ice-wedges, peat accumulation, and cliff recession all suggest an age of a few thousand years for most of the pingos. Such an age would be compatible with a period of cooling following the postglacial thermal maximum, or even earlier growth. To the east, at 64°19'N, 102°41'W, Craig (1959) has described a pingo whose formation may have been due to the marked cooling of climate following the postglacial thermal maximum. The growth of some pingos in the USSR have also been ascribed to the same period of cooling (Grave, 1956). Although pingos in the Pleistocene Mackenzie delta area might be growing today, all the evidence indicates that most of them have been stable landscape features for many hundreds of years.

SUMMARY

The pingos of the Pleistocene Mackenzie delta area number about 1,350 to 1,400. They occur typically in present or former lake basins and are most numerous in areas of drained lakes. The overburden thicknesses above the pingo ice-cores are estimated at about one-half to one-third the pingo heights. Pingo ice-cores are inferred to bottom slightly below the adjacent terrain. The pingos have grown in fine to medium sands with usually less than one or two per cent of silt and clay fractions.

Analyses of the depth of unfrozen ground beneath lakes, mean lake and ground temperatures, and three-dimensional heat conduction beneath lakes give estimates for the extent of permafrost beneath circular lakes under varying environmental conditions. As lakes shoal, either from climatic or geomorphic changes, permafrost would tend to be thinnest in the deepest parts of the lake, where open pools would remain the longest. Once a continuous permafrost cover extended over the lake bottoms, hydrostatic pressures of pore water expelled from encroaching permafrost surfaces are believed to have domed up the permafrost, in the thinnest places, to form pingos.

Pingo growth could be fast, but most of the larger pingos probably took tens of years to reach stability. Most of them are probably many hundreds to several thousands of years old.

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FREEZE-THAW CYCLES AT RESOLUTE, N.W.T.

Frank A. Cook and V. G. Raiche

ABSTRACT: Research workers in many fields have been interested in freeze-thaw cycles in the active layer. They are a phenomenon, however, on which few quantitative data exist, particularly for far northern regions. This paper presents soil-temperature data from Resolute, N.W.T., for 1960, based on continuous Speedomax records from five levels: air, ground surface, and at depths of 2.5, 10 and 20 cm.

The number of cycles is surprisingly small. On the ground, for example, where most were recorded, there were only 23 in the range 28° to 32° F, 18 in the range 28° to 34° F, and 7 in the range 25° to 35° F. Most occurred in May and June. At depths of 2.5 cm and below, only the annual cycle was recorded. However, crossings of the freezing line of short duration and small amplitude were more numerous; with 194 occurring in the standard screen, and 170 on the ground surface during the summer. On one day alone the temperature crossed the freezing line 13 times on the ground surface.

Problems in assessing the significance of cycles are discussed, and it is concluded that assumptions still widely held that they are a vigorous process in arctic regions are in need of revision. Probably it is impossible to define their parameters. The effectiveness of fluctuations across the freezing line depends, among other things, on thermal conductivity and water content of the soil, and on speed of changes in temperature and their duration. Each situation is likely a special case, and without universal application.

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RÉSUMÉ: Le problème des cycles de gel-dégel dans le mollisol a attiré l'attention de nombreux chercheurs en divers domaines. Toutefois, on ne possède, pour les régions arctiques notamment, que peu de données quantitatives sur le sujet. La présente communication rapporte des données sur la température du sol à Resolute, T.N.-O. pour l'année 1960, telle qu'enregistrée au moyen d'un thermographe de type Speedomax à cinq niveaux différents, soit dans l'air, à la surface du sol et à des profondeurs de 2.5, 10 et 20 cm.

Le nombres de cycles est, contrairement à toute attente, peu élevé. A la surface du sol, par exemple, on en a noté que 23 d'une amplitude allant de 28° à 32°F., 18 de 28° à 34°F. et 7 de 25° à 35°F. La plupart de ces cycles se présentent en mai et en juin. A 2.5 cm ou plus de profondeur, on observe un cycle annuel seulement. Toutefois, les alternances gel-dégel de faible amplitude et de courte durée sont fréquentes: 194 dans l'abri Stevenson et 170 à la surface du sol. On en a observé jusqu'à 13 en un même jour à la surface du sol.

Les auteurs traitent des problèmes inhérents à l'évaluation de l'importance des cycles de gel-dégel et mettent en doute la théorie selon laquelle ces cycles sont un agent morphologique très efficace dans les régions arctiques. Il serait probablement impossible de définir leurs paramètres. L'efficacité des alternances gel-dégel au point de vue morphologique dépend, entre autres, d'une part de la conductibilité et de l'aquosité du sol et, d'autre part, de la rapidité et de la fréquence des variations de températures. Les facteurs variables sont différents pour chaque cas de sorte qu'il est probablement impossible d'en arriver à des conclusions dont l'application soit universelle.

The problem of freeze-thaw cycles in the active layer above the permafrost is an important one to research workers in many fields. Engineers are concerned with special constructional problems created by frost action. Botanists, biologists and ecologists have a vital interest in soil-temperature fluctuations as they are often of greater ecological importance than air temperatures. Meteorologists and climatologists are giving increasing attention to the study of microclimatology and near-surface conditions. Geomorphologists consider freeze-thaw cycles an important factor in the mechanical weathering of bedrock and soils.

Freeze-thaw cycles, however, are a phenomenon about which few quantitative data exist, especially in far northern regions. No universally accepted definition of the term 'freeze-thaw cycle' has been developed, resulting in some confusion, as in most cases the term is used without definite parameters being stated. In addition to the problem of definition the study of the significance of freeze-thaw cycles is complicated by a number of other factors, among which the most important is their duration. Furthermore, different soils cool at different rates depending on heat conductivity and moisture content and thus freeze at different rates. The effect on soil of surface temperature remaining for a considerable time slightly below freezing may be the same as that of a much more pronounced drop in temperature of only short duration.

Another factor of importance is that the freezing of soil does not necessarily commence at 32°F, and thus the temperature of the soil is not necessarily an accurate index of the extent of freezing. Laboratory experiments (Bouyoucous, 1920, 1921; Rowland and others, 1955) have indicated that soil does not freeze until temperature is below 32°F.

In 1959, the Geographical Branch instituted a program of research at Resolute, N.W.T., into problems of periglacial geomorphology (Cook, 1960). One phase of the project was the study of near-surface soil temperatures. This paper presents temperature data collected for the period June 1, 1959 to September 30, 1961, inclusive, and relates them to freeze-thaw cycles. The problems involved in assessing their significance in mechanical weathering and formation of structural soils are also discussed.

PREVIOUS WORK

Troll (1943), in a discussion of frost-change frequency in air and ground climates of the earth, discussed freeze-thaw action at high mountain stations in Peru, as well as at selected stations in Europe and Asia. These data were tabulated on the basis of ice days (days on which the temperature never rose above the freezing point), frost days (on which the temperature crossed the freezing point either up or down), and frost-free days (on which the minimum temperature did not fall below the freezing point). He also discussed frost-change days, *frost-change* being defined as each crossing of the freezing line whether from freezing to thawing, or vice versa. More than one fluctuation across the freezing line may occur on any frost-change day. The number of frost changes to the number of frost-change days is called the frost-change density (*see* Table 4).

In North America, Russell (1943) mapped freeze-thaw cycles in continental United States, based on daily maximum and minimum temperature records from standard screens at 863 stations over an 18-year period. He considered that a freeze-thaw cycle would be represented by a fall in temperature from 32° to 28°F followed by a rise to 32°F. He reasoned that a fall in temperature to 28°F in the standard screen should indicate an effective freeze at the ground whereas a thaw would be recorded when the temperature rose to 32°F, as air temperatures lag behind ground temperatures.

Visher (1945) in presenting his "map of geologic interest" showed freeze-thaw cycles as the average number of days per year when the temperature fluctuated between freezing and thawing, based on the difference between the annual number of nights with frost and the annual number of days with temperatures below freezing. He later (1946) published a map showing the potential number of freeze-thaw frequencies as indicated by the number of days per year when the daily mean temperature was between 20°F and 50°F with a normal daily range of 25 degrees, the average for the United States. Under these conditions the daily maxima or minima could rise or fall to 32°F.

Fraser (1958) analysed maximum and minimum temperatures for each of 42 Canadian stations for varying periods ranging from 3 to 5 years, and published a map of the distribution of annual freeze-thaw cycles in Canada, based on a range of 6 degrees from 34° to 28°F. He also noted that D. W. Boyd used a range of 10 degrees, from 25° to 35°F in some unpublished work on Canadian stations. The data used in this paper have been analysed at the five depths of the three ranges cited above; 28° to 32°F, 28° to 34°F, and 25° to 35°F.

EQUIPMENT AND LOCATION

The instrument used at Resolute was a Leeds and Northrup Type G Speedomax recorder, which gives a continuous record of temperatures at each connection point every four minutes. Thermocouples of standard copper and constantin 18-gauge nylon were encased in a ½-inch long, ⅛-inch diameter copper tube, and installed at two remote sites. The first site was about 60 metres north of the Dept. of Transoprt ionosphere station at an elevation of 12.5 metres above sea level and 210 metres inland from the shore of Resolute bay. This site, representative of the raised beaches in the area has limestone bedrock overlain with shattered rock and gravel to a depth of approximately 2 metres. Thermocouples were installed at three levels; air (in a standard Stevenson screen), on the shattered rock surface, and at a depth of 2.5 cm. The thermocouple on the shattered rock surface was approximately the color of the rock. It was placed on the surface, exposed to insolation, precipitation, wind and other climatic elements, as would be the rock surface, and should respond to them in the same manner. The second site was approximately 15 metres to the southeast, and here thermocouples were installed in a pocket of clay, at depths of 10 and 20 cm. This area is representative of the centres of many patterned ground types common at Resolute, such as sorted circles and polygons.

The active layer is approximately .6 metre at Resolute. Thawing generally begins in late May or early June, and the ground is completely frozen over again by mid-September. Ground frost may occur on any day during the summer, although sometimes it is not recorded in the Stevenson screen, and is confined to the top few centimetres of the ground.

FREQUENCY OF FREEZE-THAW CYCLES

Table 1 shows that freeze-thaw cycles are twice as numerous at the ground surface as in the Stevenson screen (Table 1). This is logical as the ground surface is subject to the effect of incoming solar radiation in the day, and, to a lesser degree, outgoing radiation at the time of low sun at night. Ranges are small, however, as there is continuous daylight at Resolute from early May to early August, a period of relatively weak air circulation in the Eastern Arctic.

Table 1

Freeze-thaw cycles, Resolute, N.W.T. May 1-September 30, 1960

Depth	Range 28° to 32° F	Range 28° to 34° F	Range 25° to 35° F	Ice days	Frost-free days	Frost-change days
Air (Stevenson screen)	15	9	3	57	59	37
Ground surface	23	18	7	43	66	44
2.5 cm	1	1	0	58	82	13
10 cm	0	0	0	69	76	8
20 cm	0	0	0	70	75	9

The number of freeze-thaw cycles is surprisingly small. On the ground surface, for example, where most cycles were recorded, there were 23 cycles in the range 28° to 32°F, 18 cycles in the range 28° to 34°F, and 7 cycles in the range 25° to 35°F. The average duration of these cycles was 17 hours, 19 hours and 38 hours, respectively.

The number of ice, frost-change and frost-free days (Troll, 1943) in the period studied is also shown in Table 1, where it is seen that the number of frost-change days is considerably in excess of the number of cycles recorded. This is because many of the frost-change days experienced smaller cycles than the ranges given in the table. As shown in Table 5, the actual number of crossings of the freezing point during the period was much greater because frost-change days are days recording one or more crossings of the freezing line, whereas in reality on a number of days there were several crossings of the freezing line.

Nearly all freeze-thaw cycles occur in May and June (Table 2) when daily maximum temperatures are crossing the freezing line as summer temperatures are approached, and the daily minimum temperature has not yet risen to 32°F. Cycles are recorded on the ground surface early in May, and up to 2 weeks before they are recorded in the Stevenson screen. In the autumn both daily maximum and minimum temperatures drop so rapidly that few freeze-thaw cycles occur.

Table 2

Freeze-thaw cycles, Resolute, N.W.T., 1960, 28° to 32°F.

Depth	May	June	July	Aug.	Sept.	Total
Air (Stevenson screen)..........	4	10	—	—	1	15
Ground surface................	14	7	—	—	2	23
2.5 cm.......................	—	1	—	—	—	1

At a depth of 2.5 cm only one cycle was recorded, whereas at depths of 10 and 20 cm, respectively, no freeze-thaw cycle was recorded in the summer months. However, at all depths, an annual cycle is recorded when the temperature crosses the freezing point in spring as the ground thaws, and crosses it again in the fall when the ground freezes.

The number of freeze-thaw cycles in the range 28° to 32°F tabulated from meteorological records, and based on ambient daily maximum and

Table 3

Freeze-thaw cycles, Resolute, N.W.T. air temperatures, 1948-59

Year	Range 28° to 32°F	Range 28° to 34°F	Range 25° to 35°F	Ice days	Frost-change days	Frost-free days
1948	20	14	2	261	50	55
1949	17	5	0	260	47	58
1950	12	5	2	270	36	59
1951	15	12	4	260	41	64
1952	10	5	0	271	35	60
1953	12	5	2	257	61	47
1954	8	2	0	267	37	61
1955	12	3	1	280	46	39
1956	6	3	1	295	19	52
1957	16	8	3	274	34	57
1958	4	2	1	267	17	81
1959	13	5	1	275	31	59
Average	12.1	5.5	1.4	269.8	37.8	57.5

minimum air temperatures for the period 1948–59, inclusive, is given in Table 3. It is seen that 20 cycles were the most recorded in any one year, and that during 1958 only 4 cycles were recorded in that range. In no year were more than 4 cycles recorded in the range 25° to 35°F, and in three individual years, no cycle in this range occurred.

Ice days, frost-change days and frost-free days are also given for the 12-year period; ice days range from 280 to 251 days, frost-change days from 61 to 19 days, and frost-free days from 81 to 39 days.

Table 4

Crossing of the freezing line and frost-change density, Resolute, N.W.T.—September, 1960

Depths	Air	Surface	2.5 cm	10 cm	20 cm
May					
No. of crossings	42	83	—	—	—
Frost-change density	1.4	2.7	—	—	—
June					
No. of crossings	113	52	15	7	3
Frost-change density	3.8	1.7	.5	.2	.1
July					
No. of crossings	21	14	—	—	—
Frost-change density	.7	.5	—	—	—
August					
No. of crossings	12	8	—	—	—
Frost-change density	.4	.3	—	—	—
September					
No. of crossings	6	13	.1	17	5
Frost-change density	.2	.4	—	.6	.2
Total—May to September					
No. of crossings	194	170	16	24	8
Frost-change density	1.3	1.1	.1	.2	—

Although few freeze-thaw cycles of any magnitude occur at Resolute, a continuous record reveals a great number of crossings of the freezing line. In 1960, for example, the air temperature crossed the freezing line 194 times, and 170 times on the ground surface. As with freeze-thaw cycles, these crossings were more numerous in the spring; again, they were demonstrated at ground surface earlier than in the air. These data may be used to calculate an alternative freeze-thaw cycle to the three previously studied, and based on crossings of the freezing line, a drop below 32°F, followed by a rise above it, or vice versa (Table 4).

Table 5

Annual cycle (or range)—Resolute, N.W.T.—1960

	Maximum	Minimum	Range
Air (Stevenson screen)	55.0°F	−47.0°F	102.0°F
Ground surface	80.0°F	−47.0°F	127.0°F
2.5 cm	57.5°F	−42.0°F	99.5°F
10 cm	59.0°F	−23.0°F	82.0°F
20 cm	57.5°F	−24.0°F	81.5°F

The annual cycle (or range) at the five levels is given in Table 5, and, as would be expected, it is greater at ground surface than in the air, or at depths below ground surface. Although not shown in Table 5, even of greater interest, is the daily range of temperatures included in the annual range. At Resolute, at each of the levels, over 70 per cent of all daily maximum temperatures are below freezing. This condition is not experienced elsewhere in the world, apart from in the arctic or on high mountainous regions, and demonstrates the very low temperatures throughout the year in the area. These extremely low temperatures are further illustrated in Table 6 which gives the number of days of maximum temperature below certain given temperatures. For example, the ground surface experiences more than 100 days per year with temperatures registering more than 40 degrees of frost. This table, although not associated with freeze-thaw cycles, gives new data on the duration of freezing temperatures at Resolute. This prolonged intense annual cycle is perhaps of more significance in mechanical weathering than freeze-thaw cycles of shorter duration.

Table 6

Number of days maximum temperature, Resolute, N.W.T.—1960

Depth	over 32°	32° or less	25°F	20°F	15°F	10°F	5°F	0°F
Air	102	264	229	209	192	180	162	145
Surface	114	252	222	211	197	189	175	158
2.5 cm	96	270	245	230	216	202	180	168
20 cm	84	282	265	259	243	229	202	171

Depth	−5°F	−10°F	−15°F	−20°F	−25°F	−30°F	−35°F
Air	117	91	69	40	31	8	1
Surface	133	111	89	58	36	18	3
2.5 cm	146	124	93	75	50	9	1
20 cm	145	110	41	—	—	—	—

DISCUSSION

There has been a widespread feeling among workers in the field of periglacial geomorphology that freeze-thaw cycles are more numerous in northern and mountain summit areas than elsewhere, and thus account in great part for mechanical weathering in those areas. Recent literature suggests that this may not be the case. Dahl (1955), for example, in discussing the origin of mountain-top detritus in Scandinavia, states that available meteorological observations do not confirm existing beliefs that shifts around the freezing point are more frequent in arctic and high alpine localities than in the valleys and plains of Northern Europe.

Conover (1960) reported that at three stations in the vicinity of Barrow, Alaska, the surface thermometers showed a total of 6 to 14 thaws and freezes combined, compared with a figure of 92 at Blue Hill, Milton, Mass., for one spring period, or approximately 180 per season. Although no definition of "thaws and freezes" was given, the results compare favorably with Fraser's (1958) average of 15 cycles per year for the five stations in the Queen Elizabeth Islands (an average which includes an anomalously high average of 22 cycles per year for Alert on the northern coast of Ellesmere Island). Fraser also showed that the annual frequency of freeze-thaw cycles, based on a range of from 32° to 28°F, decreases from south to north. Black (1954) observed few diurnal thaws and freezes on the Alaskan coastal plain.

The data obtained by the writer at Resolute show that freeze-thaw cycles are fewer in number at that station than in southern regions. Freeze-thaw cycles recorded on the standard screen were 15 in 1960 for the range

28° to 32°F. (Table 1), and averaged 12.1 cycles per year over the 12-year period 1948-59 (Table 3), a total in line with Fraser's results for the Queen Elizabeth Islands. Surface temperatures during 1960 showed 23 cycles in this range, which is higher than Conover's results from Barrow, although as noted above no definition of that cycle was given. If, however, Conover's total was based solely on crossing the freezing line and back, the Resolute total of 85 (Table 4) is considerably higher.

At Resolute, in the case of cycles based on crossing the freezing line, totals were higher for the air temperature (97 in the air as compared with 85 on the ground). However, the opposite is true if cycles in the range 28° to 32°F. are considered, where a total of 15 cycles in the air is compared with 23 cycles on the ground. This shows that more minor fluctuations around the freezing point occur in the air than are recorded on the surface, whereas cycles on the ground, although fewer in number, tend to have a larger range. This fact should be taken into consideration when using air temperature to estimate freeze-thaw cycles on the ground.

It is also seen that freeze-thaw cycles in the range 25° to 35°F are very few, and that they, like other cycles with a range of 4 degrees or more, are confined to the top few centimetres of ground. Many workers contend that the freeze-thaw cycles investigated above have considerable effect on mechanical weathering beneath the surface, and on the formation of types of structural soil, such as patterned ground. These structures are three dimensional, with sortation at depth as well as on the surface. It appears that at Resolute, at least, an area where patterned ground is especially well developed, the hypothesis does not hold, as freeze-thaw cycles do not exist at depth.

On the other hand, crossing of the freezing line is common (Table 4) and may have some significance. Conover (1960) advanced the theory that frost effects within the ground at the top of the frozen layer are more important than a daily cycle of freezing and thawing, ". . . A small amount of thawing and freezing occurs at this level daily as the diurnal temperature wave reaches it. The amount of thaw and freeze is largest immediately after the surface thaws, and diminishes as the depth of the active layer increases, due to the absorption of the diurnal heat wave . . . This change is not necessarily detectable by thermometric means because the liquid and solid states may both remain at 32°F."

DURATION OF FREEZE-THAW CYCLES

The duration of freeze-thaw cycles is a factor which, although it deserves careful consideration, has seldom been studied, perhaps because most calculations have been based on daily maxima and minima, rather than on continuous records where many more cycles are recorded, and their exact duration ascertained.

Chang (1958) pointed out that for ground to freeze something more is required than a short dip to sub-freezing temperature. There is a duration factor as well. The effect of temperature lingering barely below freezing is the same as a much more pronounced drop of short duration. Ground will become frozen in 4 days when subjected to a surface temperature of −1.1°C (30.2°F). Petit (1877) found the duration of slight sub-freezing temperatures required to freeze ground were 160 minutes for sand, 190 for clay and 266 for peat.

Rapp (1960) suggested that it might be appropriate to distinguish between the following types of frost cycles, classified according to duration:

(a) Short frost cycles, less than diurnal, with several cycles possible in one day.

(b) Daily frost cycles, longer than (a) with freezing at night, and thawing the following day.

(c) Frost cycles of several days, such as freezing in a cold period of several days duration.

(d) Annual frost cycle; freezing in winter, with thawing the following summer.

(e) Frost cycles of several years' duration, with thawing during any particularly warm summer.

The results from Resolute show that frost cycles of short duration (type a) are limited in that area to the shallowest layer of ground. An analysis of duration also showed that 29 per cent of the cycles (at the surface) were of 10 minutes duration or less, 20 per cent were of 10 to 30 minutes duration and 18 lasted from 30 minutes to 1 hour. These short-term cycles, amounting to 67 per cent of the total, were of a temperature range of 3 degrees or less, and would have little or no effect on surface freezing. Rapp's suggestion that several cycles could possibly occur in one day is in accord with the writer's conclusions which show that, on May 18th, 1960, the temperature crossed the freezing line 13 times (Figure 1), producing 6 freeze-thaw cycles of short duration and limited range.

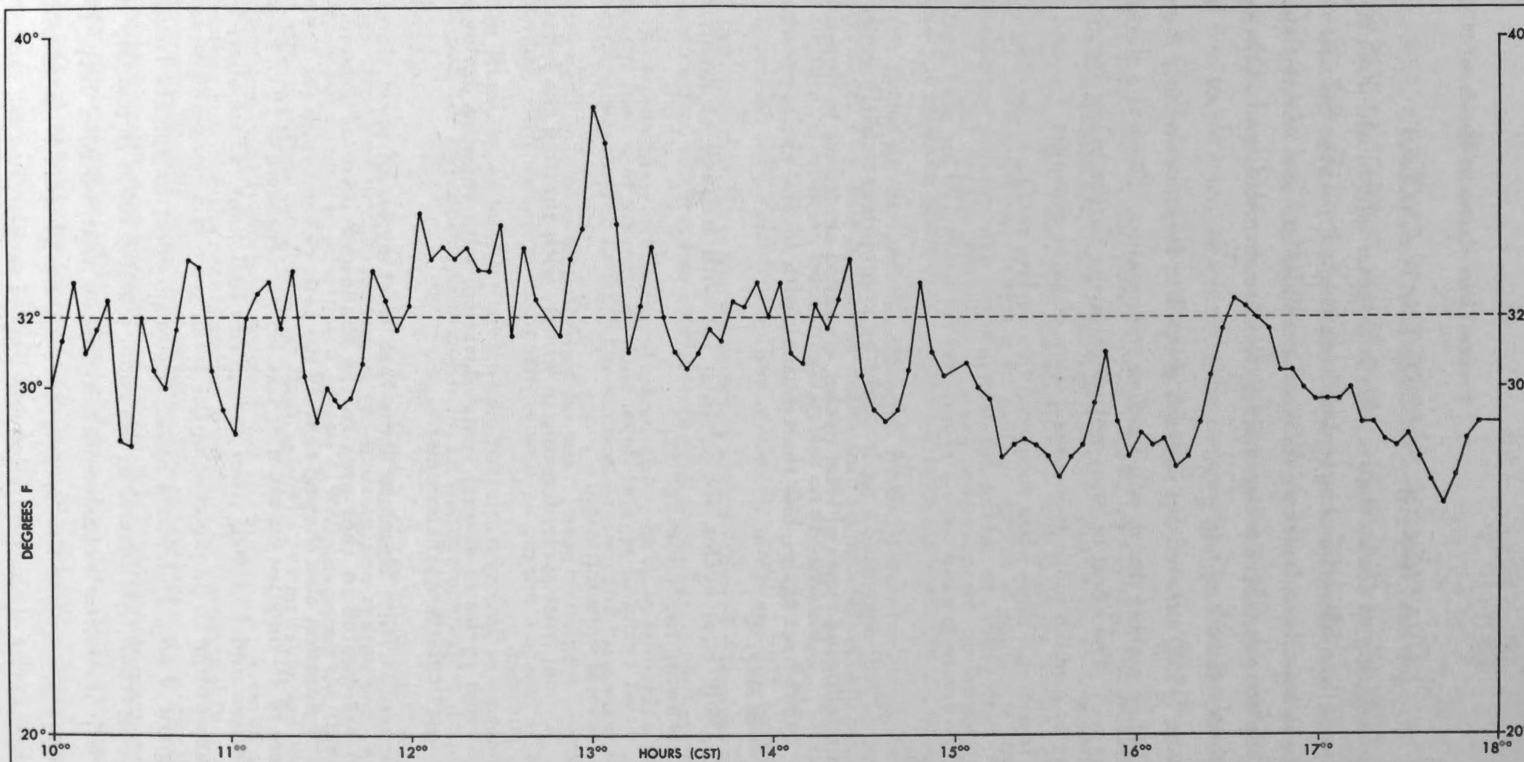

Figure 1. Crossing of 32°F at ground surface, Resolute, N.W.T. May 18, 1960.

The effect of the annual range of temperature on mechanical weathering and the formation of soil patterns is still a controversial topic. However, the wide range of temperature at Resolute, and the long periods of extreme low temperature (Table 4) no doubt are of considerable importance. For example (Cook, 1956) postulated that mud circles could partly be explained by a variation of the cryostatic hypothesis (Washburn, 1950, 1956). In that case it was reasoned that the cryostatic hypothesis would help explain sortation within the circles, not viewed so much as debris squeezed between downward freezing ground and the permafrost table, but as between horizontal pressure exerted from freezing regolith inward against the unfrozen debris in the plug. The extreme low temperatures recorded at Resolute will result in expansion of the plug when it freezes, as it usually has a high water content especially if the freeze-back is preceded by rain and/or wet snow. This freezing would result in internal sorting as the freezing expanding plug reinforces the pressure already being exerted on it by the already frozen regolith.

Finally, the critical temperature point in defining freeze-thaw cycles has not yet been agreed upon. Chang (1958) has pointed out that the temperature of the soil is not in itself an accurate index of the extent of freezing. As previously stated, the laboratory experiments of Rowland and others (1955) and Bouyoucos (1920, 1921) indicate that soil freezing commences at a temperature below freezing and is a prolonged process.

The reasons for the supercooling of soils before freezing is twofold. In the first place, the absorption of the soil water by the particles creates forces which require energy to overcome, this energy resulting in the lowered freezing point. The absorptive forces increase as the grain size of the particles decreases. The second reason is the presence of salts in the water. Bodman and Day (1937) demonstrated that the salts of colloids retard freezing. Usually the lower the water content of a soil the more concentrated the solution; therefore the freezing point is depressed with a decrease in soil moisture.

A voluminous literature, too numerous to cite, has grown up in recent years on the freezing rates of various soils. Among interested research agencies have been the Snow, Ice and Permafrost Research Establishment of the U.S. Corps of Engineers, the Highway Research Board of the United States National Research Council and the Division of Building Research of the National Research Council of Canada. It is evident, then, that different degrees of supercooling are needed under field conditions for the freezing of different types of soils. The implication is that probably no given range of temperature will adequately describe a universally effective freeze-thaw cycle. As different soils have different rates of thermal conductivity, and as such other factors as water content and speed of change in temperature will determine the effectiveness of the cycle, each case must be considered separately; each set of variables is subject to a different freeze-thaw cycle, although the same condition may be encountered elsewhere.

Observations made at Resolute on 39 evenings of frost during 4 summers showed that water on the ground began to freeze at temperatures between 32° and 31°F, and that the surface of the ground with no apparent free water began to firm at about 30°F, although several hours freezing temperature at slightly below 32° produced the same results. On the other hand, at depths of 10 and 20 cm, a temperature of about 25°F was needed to produce frozen ground and ice crystals, although, as previously stated, a longer period at a higher temperature would probably produce the same result. Insufficient observations were made of thawing to permit an estimate of the critical temperature for the reverse process, but it would appear that under field conditions at Resolute a temperature of 25°F would be a realistic lower limit for a freeze-thaw cycle of short duration beneath the surface, whereas 28°F could be effective at the surface or immediately below it. It should be noted, however, that both duration of the freezing temperatures and water content of the soil would change the critical temperatures defining an effective freeze-thaw cycle.

CONCLUSIONS

The Resolute temperature data show that the whole concept of mechanical weathering of the mantle due to frequent oscillations around the freezing line is in need of revision. It should be emphasized that there are no cycles, apart from the annual cycle, at depths below a few centimeters and it follows that assumptions still widely held among some geographers and geologists that frost cycles are a vigorous process producing frost splitting at depth in arctic countries today are not valid.

Probably it is impossible to define the parameters of a universally acceptable freeze-thaw cycle. The effectiveness of fluctuations across the freezing line depends, among other things, on thermal conductivity of the soil, water content and speed of the change in temperature, and each situation is likely a special case.

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SIMPLE TRANSVERSE NIVATION HOLLOWS AT RESOLUTE, N.W.T.

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ABSTRACT: Simple transverse nivation hollows are being formed on slopes to the northeast of Resolute Bay, N.W.T. In this area snow banks persist until late summer and in some cases are semi-permanent to permanent features. This paper presents quantitative data on angles and dimensions of 8 hollows. It was found that regardless of over-all dimensions, the slope of the backwall is generally between 20 and 23 degrees, and that of the floor between 5 and 7 degrees. The angle formed by both planes is approximately 165 degrees. The hollows appear to be developing according to a physical law.

RÉSUMÉ: Dans la région de Resolute, T.N.-O. un grand nombre de bancs de neige ne disparaissent que très tard à l'été et certains ont même un caractère semi-permanent. La nivation, processus très efficace en cette contrée, sculpte à même la couverture de calcaire gélivé des creux en forme de cirque de dimensions variables, dits niches de nivation. Une étude détaillé révèle que, indépendamment des dimensions, l'inclinaison de la paroi arrière de ces niches est de l'ordre de 20 à 23 degrés, celle du replat, de 5 à 7 degrés. L'angle formé par ces deux plans et appelé ici angle de nivation est approximativement de 165 degrés. L'évolution morphologique de ces niches semble être guidée par certaines lois de la physique.

Nivation is an important erosive process in northern regions, although as (Cook, 1962), pointed out in a recent paper, it is a process on which few data, either qualitative or quantitative, exist. It has been defined as "... the process by which quiescent névé effects the disintegration of rocks, and the destruction of some landforms, and the formation of others ..." (Ekblaw, 1918). One of the prominent features formed by nivation is a form called nivation cirque or hollow. Nivation hollows are shallow, cirque-like basins found on slopes and upland surfaces in regions that are now, or have been, under the influence of a periglacial climate. When active they are occupied by semi-permanent or permanent snow banks which never grow into glaciers, (Lewis, 1939; McCabe, 1939), when inactive, the snow banks are lacking (Henderson, 1957). Nivation hollows appear to be produced by rillwash and solifluction beneath snow banks. They lack the steep headwalls associated with true cirques, and their floors generally show movement of materials by solifluction. Lewis (1939) recognized three types related to different classes of snow patches "... First, transverse ... whose major axes lie transverse to lines of drainage, secondly, longitudinal ... elongated downhill ... and thirdly, circular snow-patches which, though transitional between the other two in ground plan, are frequently of far greater thickness ..."

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In the Resolute area, N.W.T., (lat. 74°51'N, long. 94°45'W) nivation hollows are often a striking feature of the landscape. Although all three types noted by Lewis may be found, the majority are of a simple transverse form that often occur up to the summit of hill slopes that average between 600 and 700 feet above sea level in the Resolute area. Large forms, several hundred feet long are found at the base of hills that trend in a north-south direction, or along the foot of hills encircling wider valleys and lakes. Some of these larger hollows have developed into complex forms.

No previous work has been published on the development of nivation hollows under active periglacial conditions in the Canadian Arctic. The present paper presents both qualitative and quantitative data on 8 simple nivation hollows studied in some detail in the course of field work carried out in 1960 as part of the Geographical Branch program of research on periglacial phenomena and processes.

GEOMORPHOLOGY

The study area is underlain by relatively undisturbed dolomitic limestone of Devonian age. Maximum elevations are less than 700 feet with local relief from 300 to 400 feet. In the immediate vicinity of Resolute Bay, raised beaches are found up to 300 feet above sea level. The hills behind the beaches have gently rolling plateau-like summits with a local relief generally less than 10 feet. The summits of these upland areas are composed of fairly homogeneous frost-shattered material with a high fine content, and of polygonal structure. The hill slopes separating the raised beaches from the upland surfaces consist almost entirely of frost-shattered angular limestone blocks, some very large, resting approximately at the angle of repose. There is very little fine material on the slopes, although an occasional debris island occurs. Slopes up to 39° were measured, although a range of 15° to 25° is more typical. In the wider valleys, slopes tend to be asymmetric, the northern-facing slope steeper than the southern-facing slope which may be because the aspect of the latter favors increased solar radiation resulting in maximum thawing of the permafrost, and accelerated microsolifluction.

The mantle of frost-shattered material is everywhere very thin, and the active layer above permafrost less than 3 feet. Rill wash is not an important factor in the area as annual rainfall is less than 3 inches with individual rainfalls of very low intensity (Cook, 1961). Some water was heard flowing down slopes beneath the surface along the permafrost table. Vegetation is practically non-existent.

Simple Transverse Nivation Hollows at Resolute, N.W.T.

The simple transverse nivation hollows studied are on the northerly-facing slope of Signal Hill, facing Resolute Bay, at elevations ranging from 300 to 650 feet. They are individual forms (Figure 1) separated by distances ranging from tens to hundreds of feet.

Figure 1. Simple transverse nivation hollows at Resolute, N.W.T.

SIMPLE TRANSVERSE NIVATION HOLLOWS

The most striking feature of the simple transverse nivation hollows is their similarity. In most respects measurements fell within a very close range, suggesting that the hollows were in approximately the same stage of development, and responding to the same geomorphic process.

The initiation of a nivation hollow is not well understood, their initial development apparently depending on an accumulation of snow which persists after snow on exposed surrounding slopes has disappeared. The most likely sites are small declivities in northerly-facing slopes subject to minimum insolation, or on slopes with aspects favorable to the accumulation of drifting snow. In far northerly latitudes however, semi-permanent snow banks or lingering snow accumulations may occur on southerly slopes. The development of nivation hollows, therefore, would seem to be dependent

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upon these snow accumulations, which in spring facilitate the two complementary denudation processes, nivation and solifluction.

It is apparent that once the process starts it is accumulative, because the snow remaining in the hollow prolongs the period of freeze-thaw action. Congelifraction and comminution of material enlarges the hollow. In the next annual cycle the enlarged hollow retains a greater amount of snow, again prolonging the period of frost action and further enlarging the hollow. When this process has progressed far enough a permanent snow-patch should result. Some of the hollows studied retained snow until nearly the end of the season; in other nearby areas a few permanent snow-banks were observed. The frost-action should be most active at the contact zone of the snow-patch with the back or side walls of the hollow and air temperature, and the comminuted material carried by rill wash or solifluction and deposited on the floor of the hollow. The amount of frost action under the snow-patch is unknown, although recent studies suggest that it is slight (Battle and Lewis, 1951; Thompson, 1953; Thompson and Bonnlander, 1956; Battle, 1960).

Measurements of angles and distances between numerous points in 8 simple transverse nivation hollows on one slope are given in Table 1. The key to these points is given in Figure 2. The simple nivation hollows studied varied in length from 42 to 212 feet; and in width from 37 to 192 feet. The floors of the hollows sloped downhill at angles ranging from $3\frac{1}{2}°$ to 7°, and some showed a strike ranging up to 4°. The over-all slope of the hill ranged from 15° to 18°, whereas the slopes of the backwalls of the hollows ranged from $19\frac{1}{2}$ to $23\frac{1}{2}°$. One interesting observation was that the angle formed by the plane of the floor and the plane of the backwall in every case was approximately 165 degrees.

There was a striking difference in the texture of the material between the floor of the nivation hollow, and the backwall and surrounding slopes. The backwall and surrounding slopes were composed of medium- to large-sized congelifracts with little evidence of fine material. On the other hand, the floor of the nivation hollow resembled a solifluction lobe. There was a considerable concentration of fines, and evidence of solifluction. In some cases, sorted circles, and miniature polygons were observed, and solifluction stripes were common. The fine material in several nivation hollows was very wet, and although no tests were made, appeared to be near the plastic limit.

Snow in the hollows studied persisted until late summer, six or eight weeks after it had completely disappeared elsewhere. Little run-off or rill wash was observed, although in one or two cases incipient runnels were noted

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Table 1

Measurements of slopes and angles, simple transverse nivation hollows Resolute, N.W.T.—1960

No. of hollow	Slope of hill	Slopes: Angles in degrees							Strike of floor	Slopes: Length in feet										
		AB	BC	AC	MN	RS	GF	KJ		AB	BC	DE	AD	AE	MN	RS	BG	BK	AM	AR
1		6	21½	15½	6	6	26	28½	0°	37	56	98	45	45	36	34	18	20	34	30
2	15°	6	20	15	6½	5½	20	19½	0°	56	70	42	50	50	38	43	11	14	16	16
3		3½	20	14	3½	3½	20	19	0	44	65	66	43	42	28	30	17	15	14	12
4		5	23½	9½	4½	6	22½	22½	ED 1	70	88	120	73	86	35	43	35	26	21	30
5		5½	21½	9	7	5½	18	20	DE 4	190	51	170	164	151	112	157	39	24	49	35
6	18°	6	20	14½	6	8	20	21	DE 1	70	86	212	82	165	56	36	63	56	50	48
7		7	22½	15	9	11	21	21	ED 2	41	50	90	52	37	16	19	17	11	20	24
8		4	19½	11	5	5	24	22	DE 2	40	30	165	81	60	41	43	35	28	30	25
Average		5.4	21.6	12.9	6	6.2	20.4	20.7	1.2	68.5	62	123.7	73.4	79.5	45.2	50.6	29.4	24.4	28.2	27.5

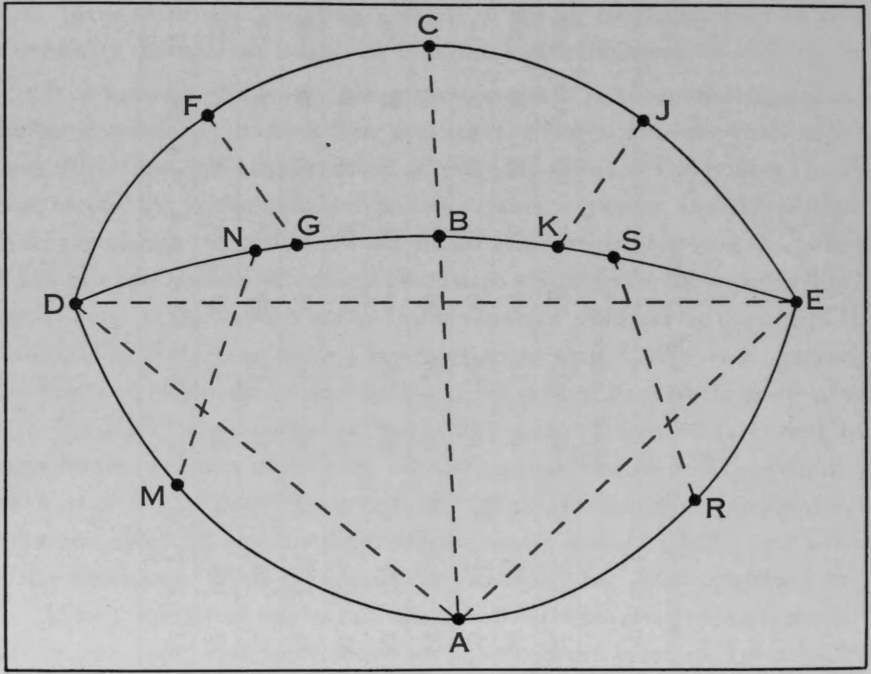

Figure 2. Measurements of nivation hollow, Resolute, N.W.T. DCE—Top of backwall; DBE—Base of backwall; DAE—Front of floor.

as well as standing pools of water that were usually ice-covered at night. It appeared that melting snow, percolating through the snow patch and into the soil below, was seeping downslope as part of the solifluction process, or moving downslope on top of the permafrost table.

In summary, although the initiation of nivation hollows is little understood, it seems that once they become established they develop according to physical laws. Nivation hollows are forming today in periglacial and other regions where semi-permanent snow banks allow the dual processes of nivation and solifluction to operate in already established depressions or hollows. In the Resolute area although all forms of nivation hollows are present, simple transverse forms are the most common.

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THE GEOMORPHOLOGY OF THE MEDICINE HAT AREA

R. Common

ABSTRACT: It is considered that a preglacial plains landscape in S. E. Alberta has been modified by glacial deposition and drainage arrangement. Most of the minor landforms, however, are the result of distinctive processes which occurred during deglaciation of the area. Of these processes, those associated with stagnating ice and glacial melt-water were especially important, for the variety of features they produced.

Sufficient variation occurs within the landforms to have influenced both the general development of soils and vegetation, as well as the present pattern of land utilization in the area.

RÉSUMÉ: L'auteur est d'avis que la déposition de matériaux glaciaires et la mise en place du réseau de drainage ont apporté des modifications au relief de plaine préglaciaire dans le Sud-Est de l'Alberta. Quant aux formes mineures du relief, elles seraient dues à des processus distincts mis en oeuvre durant la déglaciation de la région. Les processus liés à la présence de glace morte et à l'eau de fonte glaciaire sont sans contredit parmi les plus importants, surtout lorsque l'on considère la grande variété de formes sculptées par ceux-ci.

Ces formes sont assez nombreuses dans la région pour avoir exercé une influence sur le développement des sols et de la végétation, de même que sur les caractéristiques de l'utilisation actuelle des terres.

The area under consideration lies in that part of Alberta where prevailing subhumid conditions have been reported as exerting considerable influence on the whole landscape and its uses (Putnam, 1951, p. 62; Kendrew, 1955, p. 149–172; Porsild, 1958, p. 70). The city of Medicine Hat lies in a locality where fluvial and aeolian features have been partially superimposed upon more extensive glacio-lacustrine landforms. To the northeast, around Bowmanton, strongly-grained till physiography occurs, in contrast to the rolling relief of the glacial deposits near Bowell. To the southwest, in the vicinity of Sevenpersons, lies a shallow trough of Quaternary deposits, while still farther south of a composite valley of meltwater origin lying above 2500 feet, contrasting terrain rises into the Cypress Hills.

The region was visited by Palliser and several others of the early explorers. However, the geologist McConnell (1885) prepared the first systematic account of the country about the Cypress Hills. Following the Township Surveyor's Report (1886), and during this century, other geologists have studied these plains, although most of their accounts have been concerned with bedrock, economic and structural geology (Dowling, 1917; Dowling and others, 1919; Williams and Dyer, 1930; Russell and Landes, 1940). A soil map and text (Wyatt and Newton, 1926) did provide some

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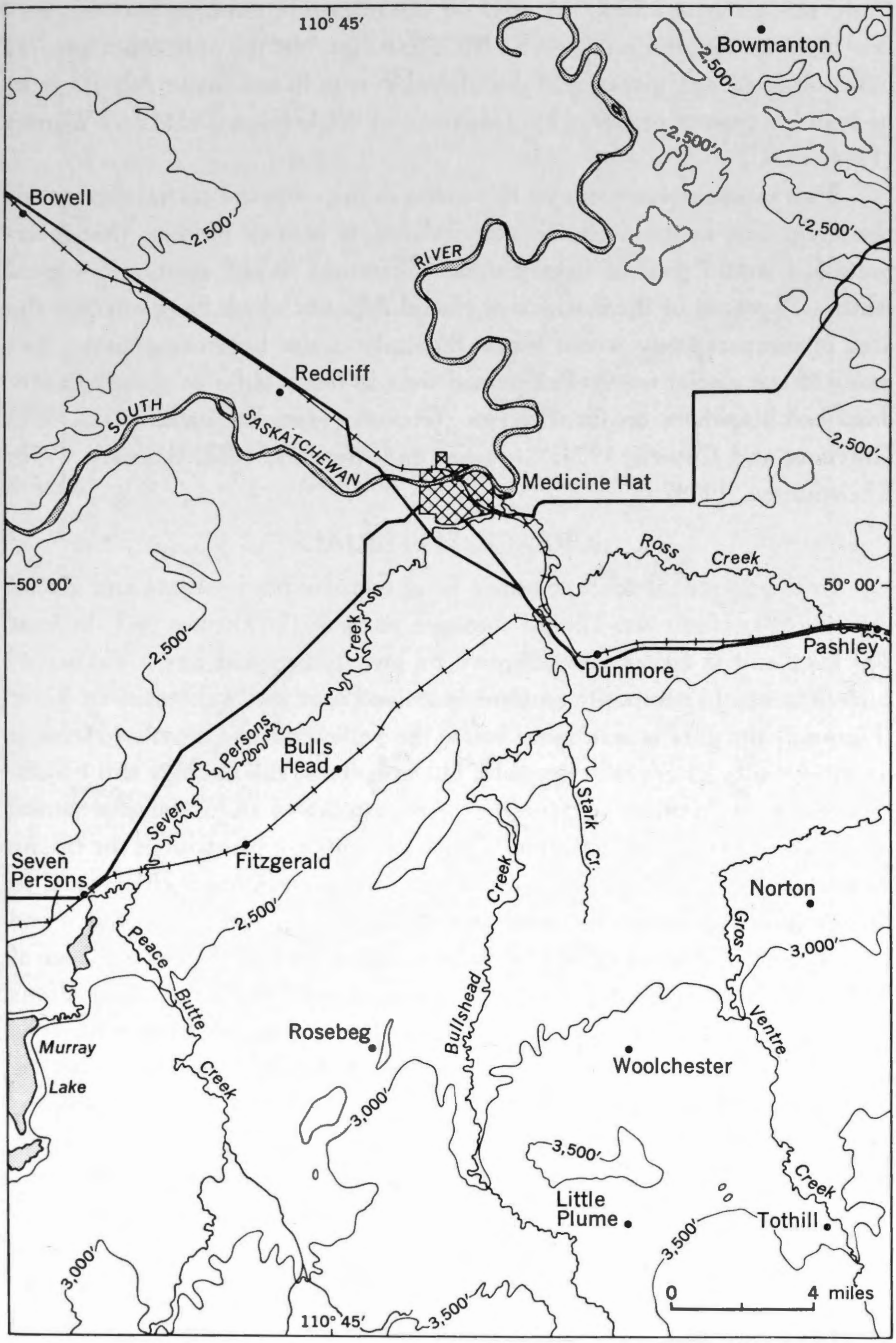


Figure 1. HB Location map showing the area under review.

fresh, though generalized, material on the inter-relationships between surface geology and the soils around Medicine Hat, but the only other general information about glacial and postglacial events in southeast Alberta is to be found in reports prepared by Johnson and Wickenden (1931) and Warren (1944, 1954).

Two seasons were spent by the writer in mapping the glacial deposits of this area, but, as the evidence accumulated, it became obvious that interpretation would depend largely upon differences in the geomorphological features, because of the distinctive glacial deposits which have affected the area in comparatively recent times. Similarly it also became apparent that many of the glacial features examined were in fact similar to those recently described elsewhere on the Prairies (Gravenor and Bayrock, 1955, 1956; Gravenor and Kupsch, 1959; Gravenor and Meneley, 1958; Bayrock, 1958; Christiansen, 1959).

SURFACE MATERIALS

In the present landscape minor relief features predominate and glacial deposits are ubiquitous. The till possesses variable thicknesses up to at least 200 feet, and is either yellow-brown or grey-brown and has a variety of surface forms. In composite sections examined near the Saskatchewan River (Figure 2) the grey is invariably below the yellow till and separated from it by sandy, silty or gravelly deposits, but in spite of this there is still insufficient evidence to prove or disprove the occurrence of an interstadial period in this area. Again, although small rock fragments are plentiful in the till, no distinctive indicators are to be found amongst the erratics of either local or Keewatin origin. Even the Athabasca sandstone fragments (Gravenor and Bayrock, 1955) present could have been redistributed by the last advance of ice if it had come into this area from the northwest. Since glacial striations and similar ice-eroded features are also absent on the country rocks it is difficult to decide whether the predominant ice-flow direction was from the northwest or north-northeast. However, from the available evidence of washboard moraines and other ice disintegration features in the till (Gravenor and Kupsch, 1959) it is now suggested that while the major advance came from the northwest, minor stirring took place in stagnating ice during deglaciation and caused sporadic movement from the north-northeast. There is no evidence to suggest that the upper slopes or plateau surface of the Cypress Hills were covered by ice although there are abundant indications that the ice did wrap itself around the western end of this high ground as it flowed into Montana.

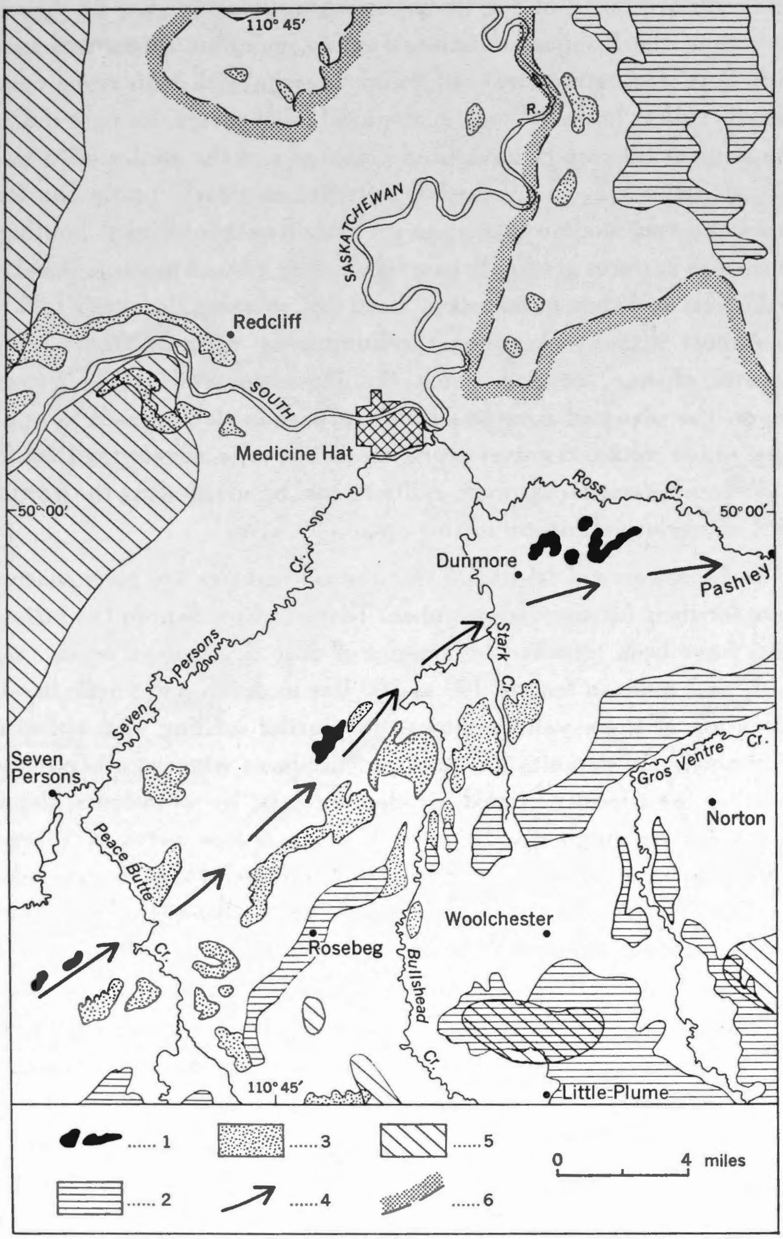


Figure 2. Till deposit features of the Medicine Hat area. (1) moraine remnants, (2) hummocky till, (3) planated till, (4) Robinson-Dunmore spillway line, (5) gray till, (6) trenched till deposits.

Features of the Till Deposits

The surface forms of the till are most unusual (Figure 2). Along the north side of the Robinson-Dunmore valley, morainic ridges and mounds provide both distinctive local relief and deposits with high stone content. Across the valley, however, only gentle local relief occurs, for here the swells on the original till surface have been planated and the swales filled in with a thin, stoneless clay. Such distinctive features clearly imply the former presence of a very shallow lake at an ice margin in this vicinity. Southwards these smooth surfaces gradually pass into rolling ground moraine (local relief 10 to 25 feet) and then hummocky "dead ice" moraine (local relief 30 to 50 feet). Almost without exception the hummocky till zone fronts a glacial melt-water channel beyond which the sequence tends to be repeated. However, the planated zone has a counterpart in thin spreads of cobbles, pebbles, sands, silts and gravel overlying till deposits, suggesting that somewhat different glacio-lacustrine conditions had occurred along the ice margin when it was relatively static in this area.

Northeastwards of Medicine Hat the till features are particularly distinctive for their furrowed appearance. Towards Bowmanton the yellow till deposits have been trenched by a series of deep dry valleys which may be as much as 7 miles in length, 100 to 200 feet in depth and $\frac{1}{4}$ mile in width. Examination of these valleys shows that partial infilling and obliteration by hummocky till deposits has occurred in places whereas other stretches have either been locally eroded, or else veneered by wind-borne deposits, since these were originally formed. The interfluves between these dry valleys are either planated or else smoothly rounded. However, at and near their crests finer fractions of the overburden have apparently been removed by washing and/or deflation. The dry valley sides, though steep, are stable and frequently cut by disused gullies, although fragments of benches or terraces are often noteworthy features towards their eastern margin. Bedrock exposures on the valley sides are uncommon and appear only near the South Saskatchewan River. In places, the inter-linkage of adjacent valleys by small spillways suggests that overflow of water was possible, and certainly the bevelled crests already mentioned imply that the valleys could have been full to the brim and overspilling with melt-water.

Features of Other Surface Materials

Glacio-lacustrine deposits of well-bedded clays, silts and sands also cover extensive stretches of the area (Figure 3). The smooth-surfaced and

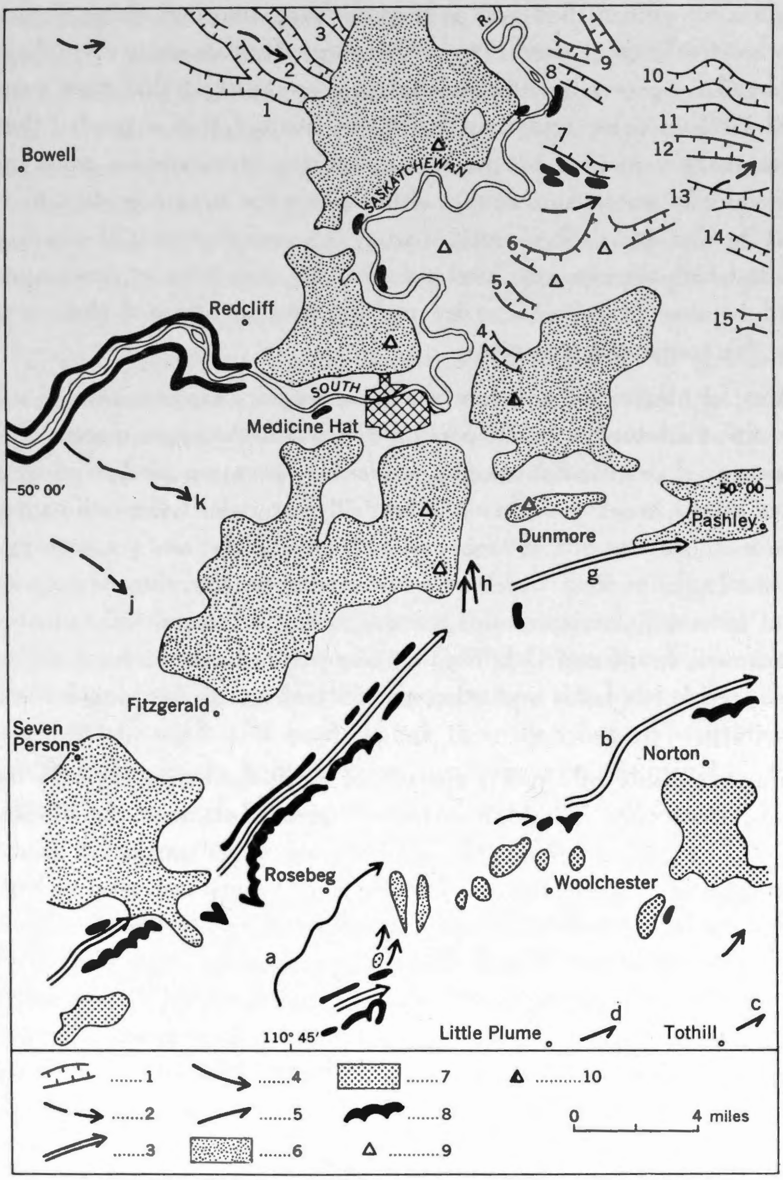


Figure 3. Surface features other than till deposits. (1) ice-walled melt-water channels, (2) overflow melt-water channels, (3) ice margin melt-water channels of large proportions, (4) small ice margin melt-water channels, (5) partially obliterated marginal channels, (6) glacio-lacustrine deposits of fine material, (7) glacio-lacustrine deposits of generally coarse material, (8) bedrock outcrops, (9) blown sand, (10) loess pockets.

gently sloping terrain about Sevenpersons and Medicine Hat, which is underlain by undisturbed lake sediments, suggests open water deposition under relatively quiet conditions. Since these deposits show disturbed sections but have generally not been eroded, it seems likely that they were not laid directly upon the underlying till floor. Instead, it is suggested that, as the ice margin receded northeastward in the Sevenpersons embayment, proglacial lake waters spread over either a thin ice margin or detached and buried ice blocks, in the vicinity of what is now Fitzgerald. The lacustrine deposits which then accumulated over this ice were later let down and disturbed because of subsequent ice melting and so attained their present distinctive forms and structures.

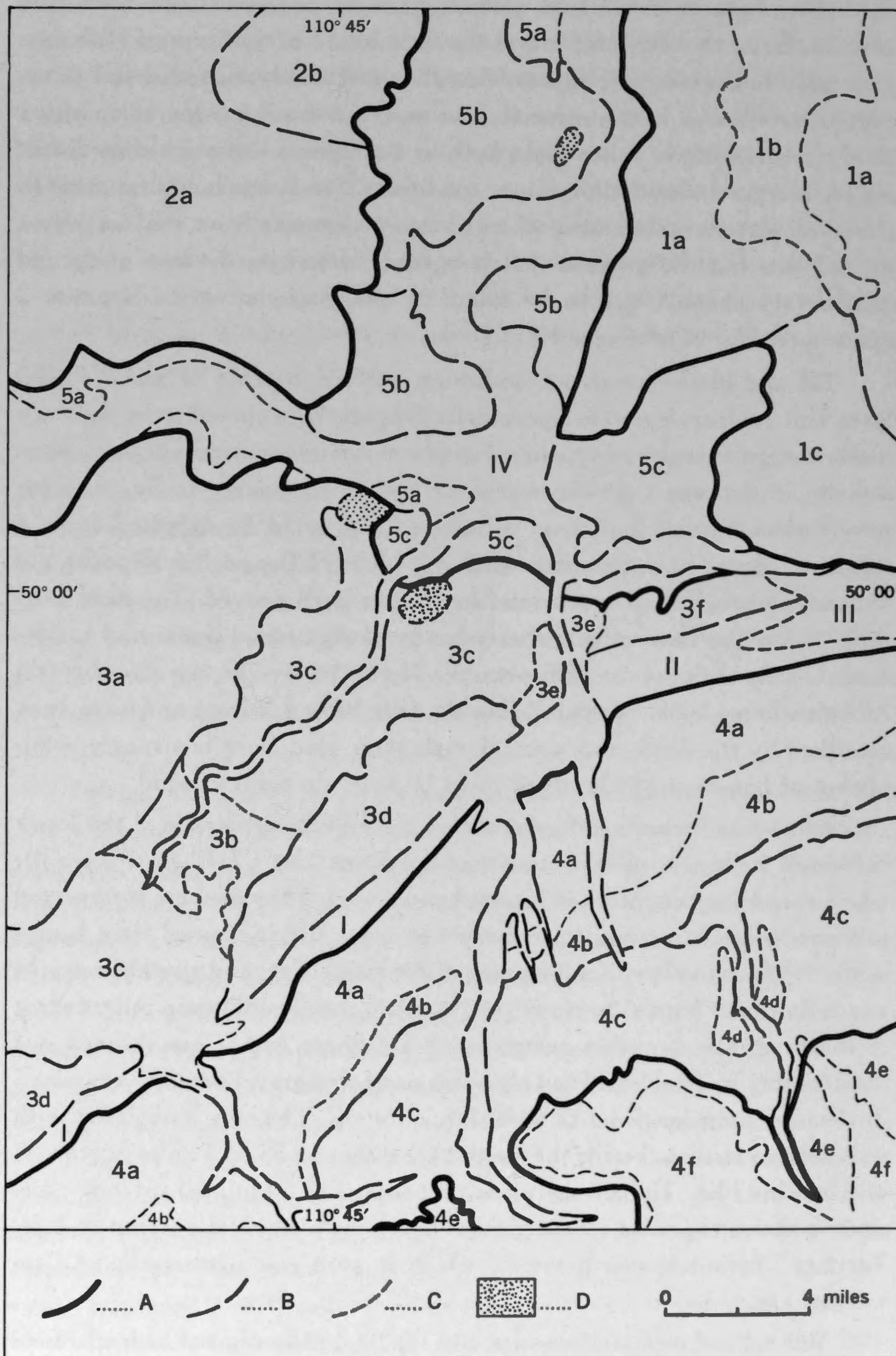
East of Medicine Hat other lacustrine deposits are preserved in a shallow basin of subdued relief. Inspection shows that the upper margin of these deposits is not at a constant height, nor is there always a good physiographic feature at the break between lake and till materials. Lake sediments cap the river cliffs at about 2,450 feet east of Medicine Hat and yet, as they also form the banks of Ross Creek at a lower level nearby, they are obviously related to vestigial remnants of the dry valley system already mentioned (i.e. channels 4, 5, 6 and 15 in Figure 3), as well as to the series of elongated and aligned kettle holes and other partially obscured dead-ice features in the vicinity.

Farther afield only minor spreads of glacio-lacustrine deposits occur. The stoneless clay found on planated interfluves near the Robinson-Dunmore valley has already been mentioned. It is considered that this composite valley ('g' in Figure 3) was a large ice-margin spillway and the end member of a series which began to the southwest. (*See*, for example, Forty-mile Coulee, Chin Coulee and Etzikom Coulee on Sheet 72E, "Foremost", National Topographical Series, 1:250,000, 1958). Periodically, the volume of water in this channel appears to have been great, for, in overspilling southwards, shallow lakes were produced while overflow northwards cut small melt-water channels. In contrast, the Roseberg melt-water channel ('a' in Figure 3) is graded into, and associated with, a ribbon of smooth-surfaced silt, clay, and sand deposits lying along the middle stretches of Bullshead Creek. Clearly the ice margin and local relief together were here responsible for the formation of both the melt-water channel and the small, proglacial lake. Finally, the deposition of the older and higher spreads of silt, sands and gravels upon till, for example south of Norton and near Woolchester, should be considered. With one exception, all these sites are distinctive because of the high percentage of stained quartzite cobbles they contain. Since these cobbles are of the type found in the Cypress Hills conglomerate to the south, it appears that they first had been eroded and transported by streams from unglaciated or early ice-freed terrain. Deposition then appears to have taken place both at and upon a static or slowly receding ice margin under shallow water conditions. Small, similar deltas occur in the Sevenpersons embayment where overflow channels from the northwest graded out, but the greatest deltaic spread, as well as the line of an old distributary channel, are to be found in the glacio-lacustrine deposits 2 miles northwest of Medicine Hat.

Till and glacio-lacustrine landforms, surface patches of glacio-fluvial loam and aeolian deposits appear quite frequently, although true outwash materials are conspicuously rare. The glacio-fluvial loam occurs only occasionally in deposits that are extensive and thick enough to be mapped; nevertheless its distribution is widespread and is to be expected from a surface of dirty, disintegrating, dead ice. Most of the aeolian deposits are of sand, although small pockets of loess have been proved. The sand normally takes the form of thin sheets but low longitudinal dunes and anomalous U-shaped dunes are also present. The latter are known elsewhere in Alberta where similar proglacial lake deposits have dried out and have been reworked by the wind, and, even though their plan form is strongly reminiscent of barchans, the horns of these U-dunes do point upwind.

Kame-like deposits form distinctive mounds on either side of the lower Bullshead valley, 4 miles southeast of Medicine Hat ('3 E' on Figure 4), where they have been worked for sand and gravel. They are here interpreted as kame-moraine features. In the same vicinity a striking gravel train begins in the Robinson valley, 8 miles south of Medicine Hat, and stretches northwards for about 5 miles towards the Ross-Bullshead confluence. An offshoot of this train also branches eastwards at Dunmore, but it has thinned out considerably by Pashley. The only other extensive gravel deposits examined in the area are considered to be of fluvial origin. They are associated with well-defined terraces beside the South Saskatchewan River 4 miles northwest of Medicine Hat. The northernmost exposure has additional interest, too, since it shows the more recent gravels overlying a residual outcrop of late-Tertiary "Saskatchewan gravels", which in turn rest upon rocks of Cretaceous age.

Alluvial and colluvial deposits, like the limited sections of bedrock, tend to be restricted to incised valleys only. Nevertheless, the type and thickness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of bedrock exposed in the present valleys and old melt-water channels does affect the details of local relief, because of differential erosion rates. On the other hand, the prevalence of gentle dips scarcely suggests that complicated bedrock structures influence the present surface relief.

The overburden exhibits unusual features such as the distinctive dry valleys in the Bowmanton area, which do not appear to lie amongst a set of parallel marginal moraines. Furthermore since the regional bedrock slope rises to the north-northeast in this locality and the ice is considered to have thinned out towards the northwest it is difficult to imagine these valleys as resulting from a set of ice-margin spillways. Ice-tunnel valleys are also rejected as a possible origin for the features because of their large size. However, some of them may have begun as ice-tunnel valleys and continued to develop after ice roofs had collapsed. By elimination, then, the possibility of their origin as ice-walled channels should be considered.

The incising of a melt-water channel series through an ice sheet whose margin thinned eastwards does offer a solution for the dimensions of the valleys and their general orientation. Parallelism in the series, however, implies some additional control. Sub-parallel fracturing or fissuring of the ice alone might be envisaged in the initial location of the channels, but once the ice disintegration features described by Gravenor and Kupsch (1959) are considered, alternative possibilities arise. For example, if ice disintegration ridges, produced either by squeezing or ablation, were to appear as surface

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Figure 4. Geomorphological units of the Medicine Hat area on regional (A), locality (B), and area (C) bases. Till islands in the former proglacial lake are stippled (D). This diagram indicates the variety of land forms which occur in the area.

Detailed Key:

1a furrowed till
1b hummocky till
1c modified hummocky till

2a rolling till
2b furrowed till

3a rolling till
3b broken minor relief of till and glacio-lacustrine material
3c glacio-lacustrine deposits with smooth outlines
3d till marginal footing capped with moraine remnants in places
3e kame moraine
3f morainic remnants

4a bevelled and gently sloped till
4b rolling and hummocky till
4c rolling till
4d hummocky till
4e rolling till
4f hummocky till

5a riverside terraces and benches
5b glacio-lacustrine surfaces with various degrees of subsequent modification
5c glacio-lacustrine areas

I & II large marginal melt-water channels
III pashley lake flats
IV Saskatchewan tract

Figure 5 A bevelled till surface west of Stark Creek and 9 miles SSW of Medicine Hat. Frequently the farms are located in rather inaccessible creeks—the farmers apparently being prepared to tolerate the high local temperatures of summer and the risk of rattlesnakes in return for wind shelter and more reliable water supplies.

features on a wasting and stagnating ice-sheet surface, and particularly near its margin, they would at once pond, and then direct the flow of surface melt-water. Adjacent ponds of water on the ice would first become aligned and then interlinked. With continued melting the volume of ice-surface water would increase and, in favored cases, lead to the formation of melt-water channels. Where these channels became entrenched into the stagnating ice, along predetermined lines, their melt-waters could remove either surface and englacial material, or these together with basal deposits, which would otherwise have been left as overburden by ablation. Persistence and eventual preservation of channels so formed would obviously be fortuitous, depending upon local conditions during and after late-glacial time. Thus the line of a disused ice-walled channel could be completely lost if the ice were sufficiently reactivated to erase it. Alternatively only the channel-floor deposits would be preserved, to form an esker-like feature in the present landscape. In those areas where the bedrock surface is undulatory, the superimposed channel might well be preserved as a crisp rock-cut spillway in one locality and yet completely lost or indifferently preserved nearby in different material (Diagrams A and B). Localised stirring of ice walls would also be reflected in the ultimate form of these channels.

An ice-walled origin for these dry valleys is therefore suggested, since it affords an acceptable explanation for (a) the irregular stretches of thalweg seen in some of these till-floored valleys, (b) hummocky till deposits, which appear to have been extruded into stretches of the valley floors, sides and interfluves, and (c) aligned kettle holes, which are interpreted as incompletely obliterated channel lines; as well as their size, orientation and variations.

Melt-water channels, apparently of different proportions and origin, appear on rising ground south of the composite Robinson-Dunmore valley (Figure 3). They developed along relatively slow moving or briefly static ice margins and, where they were associated with proglacial lakes, distinctive local relief and stratigraphy are invariably present. The largest of these marginal channels is represented by the Robinson-Dunmore valley, which is over 18 miles long with a floor up to $\frac{1}{2}$ mile wide. It is, in fact, an enormous marginal footing, but, because of its depth and position, the northern side of this channel forms the scarp of a cuesta bounding the nearby Sevenpersons embayment. Both the Robinson and Dunmore sections contain low corroms, the most important of which diverts the Ross Creek westwards from Irvine into the South Saskatchewan River catchment. In spite of its size, however, this marginal footing does not possess the unusual features of the dry valleys around Bowmanton.

While the melt-water channels lying north of Sevenpersons are also of modest proportions, they, too, are genetically different. These spillways, of the overflow type ('j' and 'k' in Figure 3), cross the low divide between the South Saskatchewan and Sevenpersons catchment basins, and indicate that this watershed once served to separate waning ice and impounded melt-water. It is inferred, therefore that after the ice had receded from the line of the Robinson valley, and had cleared the upper stretches of the Sevenpersons catchment basin, it parted over the line of the present divide. Ice-margin lake waters then migrated down the centre of the Sevenpersons trough, although periodically these were fed by overflow waters from another proglacial lake to the northwest. (Field observations showed these overflow channels to be associated with distinctive minor relief features and stratigraphy at their intakes and outlets). It also seems probable that the present course of the South Saskatchewan River was determined by the end member of this overflow series, as the river now crosses the same preglacial ridge as the channels to the west of Redcliff.

At about the same time that these overflow channels were being cut, another small proglacial lake east and southeast of Medicine Hat also must have formed. Its waters at first flowed eastwards into the Dunmore channel at Pashley, but later drained westwards into the main lake of the area, after the ice front had receded north of Medicine Hat. The difficulties in interpreting the history of this lake have been suggested earlier in the text.

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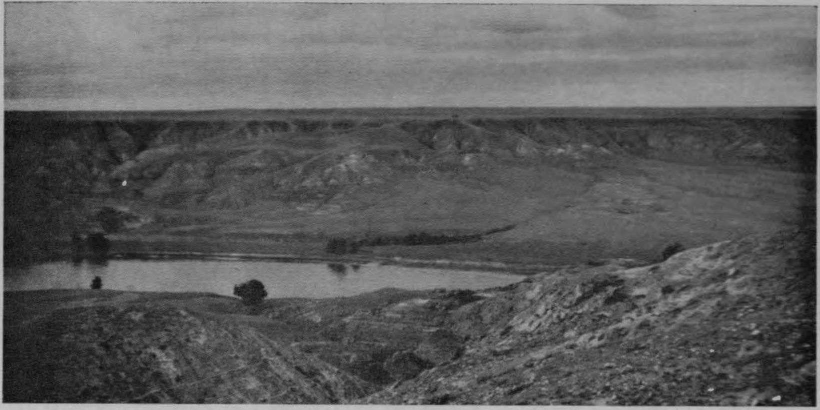

Figure 6 Looking northward over the incised South Saskatchewan valley, 1 mile SW of Redcliff. On the east the coalescing fans beneath a retreating scarp of bedrock and overburden produce a valley-side scene which is distinctly different from that to the west where gullying by running water is the more dominant process.

The mechanics of dead-ice melting and the behaviour of glacial melt-water in extensive areas of disintegrating, stagnant ice are still imperfectly understood and it should not be assumed that this small lake had to be interconnected with the Sevenpersons area.

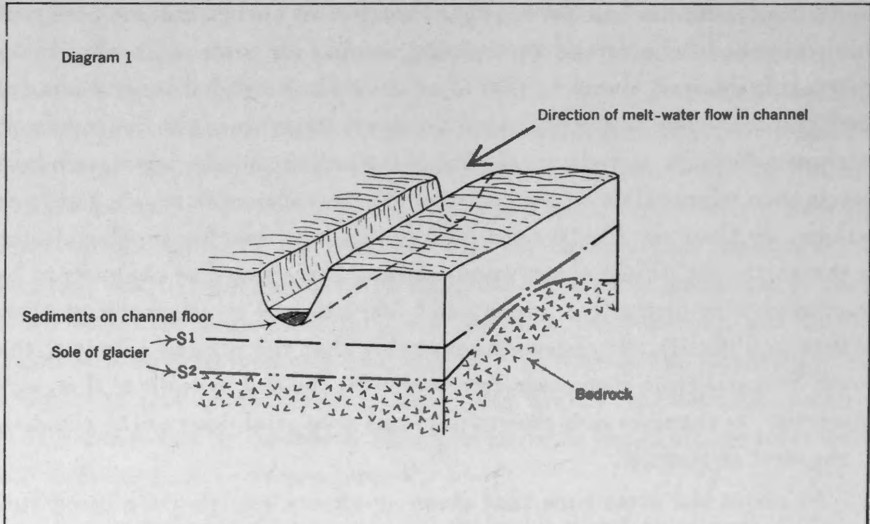

Diagram 1. It is assumed that melt-water is flowing in an ice-walled channel cut wholly in stagnant but relatively clean ice and that sediments accumulate on the floor of the channel. If the channel is abandoned and melting is continued then the deposits on the channel floor could be let down onto the underlying surface. In this process the proximity of the glacier's sole (S) will also influence the form of the superimposed deposits i.e. the amount of disturbance when the channel deposits are lowered to S1 should be less than to S2.

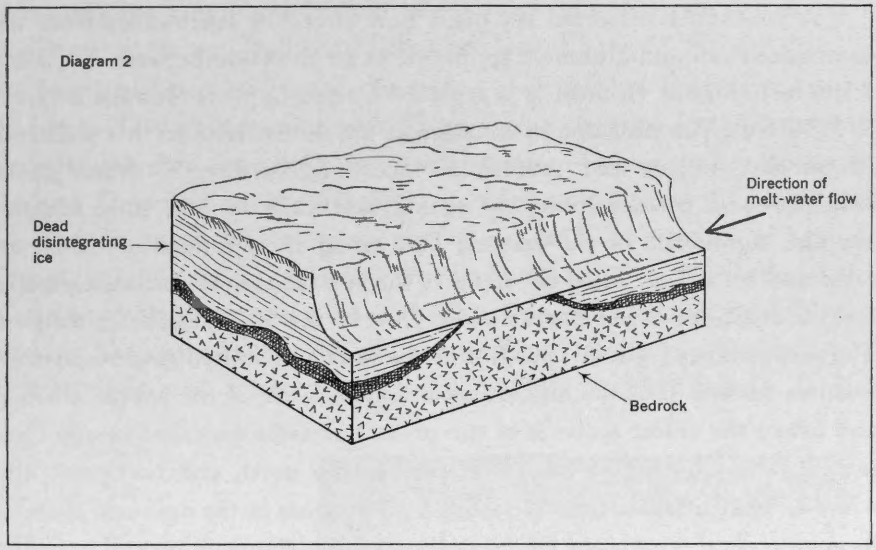

Diagram 2. **A hypothetical longitudinal section along an ice-walled channel which is assumed to be cut partially in dirty ice (A) and partially through bedrock. It will be apparent that the final form of the channel can be influenced by a number of factors e.g.—(i) If the ice stirs on the channel sides varying degrees of modification can occur. (ii) The dirt content and thickness of A will also affect the floor character of the channel.**

It is considered that this area was first traversed by ice-walled channels similar to those found near Bowmanton. Later, either a localized ice readvance occurred or dirty, decaying ice warmed sufficiently to stir and obliterate most of the channels before the ice finally stagnated. These movements, therefore, blocked any free flow of melt-water eastwards and ponded sufficient water to produce a small lake. Water from this lake first overflowed eastwards along the line of the upper Ross Creek to Pashley, but later the changed location of the ice margin allowed drainage in the opposite direction.

REGIONAL GEOMORPHOLOGY

Both Dawson (1882) and McConnell (1885) suggested that one of the results of glaciation in this part of the Prairies was the levelling out of relief differences by infilling valleys and eroding ridges of the pre-existing landscape. They also recognized that in this area the course of the South Saskatchewan River was not within its preglacial bed. Field mapping and the available subsurface data now support the view that glaciation has modified a preglacial plains landscape, but chiefly by deposition and changes to the former drainage pattern.

A youthfully dissected till plain now stretches southwards from the composite Robinson-Dunmore spillways, at an elevation between 2,000 and 2,500 feet (Figure 4), until it is replaced by the Cypress foothills at 3,000 to 3,200 feet. The pronounced zonation of the minor relief on this plain into hummocky, rolling and smooth surfaces results directly from glacial influences and, besides giving the countryside some variety, these features are also significant to the farmer, because of their influence upon local relief and soils. Into these till deposits the surface run-off initiated parallel and dendritic drainage patterns which grew by ingrafting as the ice receded. The disposition of glacial melt-water channels and limited glacio-lacustrine features confirm that the highest ground stood clear of the ice margin first and hence the oldest sections of the present streams now lie towards their source. The lowering of the ice margin to the north and northwest also offers a rational explanation for some derangements in the drainage pattern, because several portions of the former melt-water channels are still in use by present streams, for example, the eastward deflection of the Gros Ventre Creek near Norton. Yet it should not be imagined that the ice retreat was without minor oscillations, for, in places, small lobes readvanced or the ice warmed sufficiently to stir and then stagnate, erasing all or parts of recently cut margin spillways.

In the Sevenpersons embayment the present landscape is almost wholly of depositional origin, with the axial stretches dominated by lacustrine deposits and with more variable till surfaces on the flanks. The Sevenpersons Creek meanders northeastward over this ground but is clearly misfitted, and has its principal left bank tributaries lying in old glacial melt-water channels.

The city of Medicine Hat is located upon fluvial terraces near a water node formed by the abrupt bend of the South Saskatchewan River and the convergence of Sevenpersons and Ross creeks. These terraces begin only 4 miles upstream from the city, replacing the high-level benches at a point where the river emerges from a rock-girt channel, but they soon become significant features in the riverside scene until Rapid Narrows is reached 4 miles north of the city where they form a most impressive terrace flight, on the right bank, at about 2,300, 2,250, 2,150 and 2,000 feet. Sufficient evidence is also available to indicate that an old buried channel runs north from Medicine Hat but west of the present river course. The same or another channel runs southwestwards in the Sevenpersons embayment and apparently has a number of minor offshoots on its south side, for example into Peace Butte and Bullshead Creek catchment basins.*

North of Medicine Hat the lacustrine plain is now most extensive on the left bank of the Saskatchewan River, for on the right bank blown sand deposits, dissected river cliffs, and fluvial terraces hide or replace lacustrine features. This plain has a surface of low relief with yellow till underlying the lacustrine material, but dunes and ridges, sand "blow outs" and gullies diversify the scene in several places. Nine miles north-northeast of Medicine Hat an interesting inlier of till occurs and this was probably an island in the old lake, for its upper surface was first water-trimmed and then its margins were later eroded when the lake level had fallen.

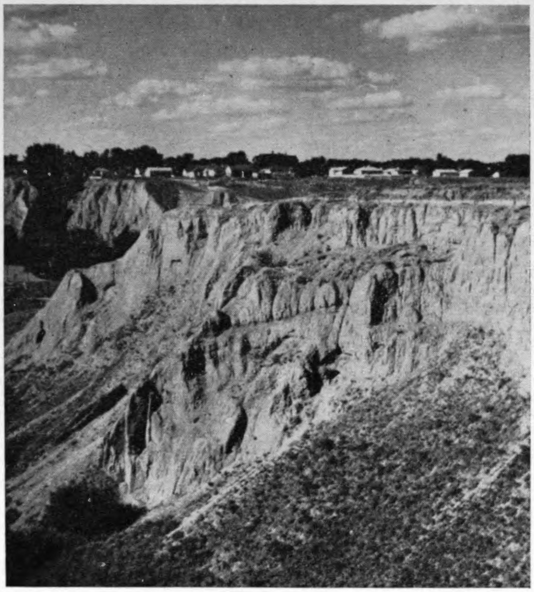

Figure 7

Kinpark, Medicine Hat looking north towards Central Park with its tall cottonwood trees. The overburden is incised 120 to 150 feet and the tendency for the till to fissure is displayed on its faces. In the foreground the lower till is separated from the upper by an oblique cleft developed along the line of weakness formed by soft silt and sand deposits. Above the till the white band is owing to salt accumulation at the base of the lacustrine silts. In the background a fairly recent gully produced by slope failure has produced the shaded cleft in the photograph.

The nature and origin of the furrowed till deposits lying nearby have already been discussed at length and hence only a few additional comments are needed about this area. The ice-walled channels are considered to predate the formation of the main lake, but, because they are probably related to former proglacial lakes in the adjacent Many Island and Vale districts (to the east), a complete interpretation of all these features is not yet possible. From a practical standpoint these parallel furrows are also distinctive

*An assured supply of cool groundwater has now been proved in this buried channel and water is being drawn at the rate of 1.7 million gallons per day for domestic and industrial purposes.

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because of the rigid limitations which they impose upon the potential usefulness of the land and their nuisance value to lines of transportation.

Still more variety to the landscape is afforded by the rolling and undulating till surfaces which are on the fringe of a "dead-ice" zone west of Redcliff. South of the Trans-Canada Highway fluvial and lacustrine features which border the river quickly rise into rolling plains relief, in which closed disintegration ridges sporadically appear, while north of the highway a low, broad-based ridge gradually declines eastwards to a dissected margin. Here the shallow valleys of small non-perennial creeks are accompanied by

Figure 8. 10 miles N. of Medicine Hat. The incised, meandering course of the South Saskatchewan River lies approximately at the centre of the photograph. On the left bank the lacustrine plain has been veneered by aeolian deposits which have "blown" in places recently. The ovoid feature is a till inlier which was modified by erosion from the glacio-lacustrine waters. On the right bank most of the glacial ice-walled channels have been eroded by more recent surface run-off while to the east "dead ice" hummocky terrain is clearly displayed.

another series of ice-walled channels which apparently grade east-southeast into the former Medicine Hat lake waters.

APPRAISAL

Surface and groundwater resources are of vital importance in this semi-arid area where moisture deficiency is generally reflected in the occurrence of drought-tolerant vegetal species such as blue grama-grass, spear-grass, sagebrush and cactus. It is apparent that sufficient variation occurs within the landforms to influence the general development of vegetation and soil, favoring the growth of salicones in saline depressions and wild rose and chokecherry in the sandier areas and treelined stretches along the main river.

The South Saskatchewan River and Ross Creek are the two most reliable streams within the area but even the former has a widely ranging discharge in any one year or period of years (i.e. from 145,000 cfs to 360 cfs at Medicine Hat). Surface discharge from the other streams and creeks tends to be concentrated into the period from 1st March to 1st June and results chiefly from spring thaw and local cloudbursts in summer.

The occurrence of springs in the area has long been known and there can be little doubt that the distribution of known springs and shallow wells was formerly an important factor in the orientation of overland trails. Those springs present are mostly of the contact type although examples of pocket and dimple varieties are also considered to be present. Their role is now subordinate, however as farm supplies are mostly obtained from wells drilled or dug in superficial deposits or bedrock. This reliance upon local groundwater sources appears to be justified by the nature and succession of the superficial deposits, for at most sites the chances of drilling into water-bearing deposits at less than 200 feet from the surface are fair to good (Table 1). The distribution of sorted sand and gravel deposits is also convenient for commercial purposes, and there are several large gravel pits in the area.

Table 1

Water-bearing deposits

Recent	Blown sand, fluvial sands and gravels—in places up to 40 feet thick
Pleistocene	Lacustrine silt, clays, sands and gravels—in places up to 35 feet thick Yellow-brown till of varying thickness Silt, sand and gravels, in places up to 30 feet thick Grey-brown till of varying thickness
Pliocene	Saskatchewan gravels—in places up to 15 feet thick

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Periodically the local climatic conditions produce marked water deficiencies which naturally increases the susceptibility of the finer deposts to wind erosion. This tendency is increased because of the natural vegetation cover, the occurrence of extensive areas where the superficial deposits have high silt or sand contents and the open nature of the terrain. There can be little doubt that misuse of the land in comparatively recent years has been responsible for some of the present aeolian deposits, but it also seems likely that natural conditions were favorable for wind erosion and deposition at an earlier time (following deglaciation). Excessive amounts of water can be received just as unpredicably from chinooks or thunderstorms, so that overloading, and then failures, occur on steep slopes whilst ephemeral discharges take place in the creeks and gullies. Two distinctive types of slopes and gullies develop along the incised valleys and in the foothills, the one essentially due to the action of running water and the other primarily the result of overloading and slope failure under gravity, with subsequent grading by both wind and water.

Christiansen (1959), dealing with slope failure and slope development around Swift Current, noted that much depended upon the competence of strata underlying the overburden in stream valleys . . . "When downcutting by the stream penetrates the drift-bedrock contact, the confining pressure is reduced sufficiently to cause the plastic shales to move laterally into the stream valley. The movement of the shale places the overlying competent drift in a state of tension and fractures are formed." These fractures eventually result in slope failure and valley widening by slumping. Russell and Landes (1940) likewise, in their discussion of butte features, recognize the importance of slumping in slope development on the fringes of the Cypress Hills. Such an explanation applies to some of the slope failures in this area where soft clay, coal, shale and sandstone materials have been exposed, but in other localities additional factors appear to be involved. On the banks of the South Saskatchewan River south of Redcliff, for example, although the exposure of overburden upon bedrock are identical, the lacerated valley side physiography produced by steeply graded gullies abruptly changes into that of a less dissected, but more imposing, scarp. The latter is frequently fronted by coalescing fans either upon or of graded slump deposits. Further downstream, near Medicine Hat, examples of mixed slopes developed wholly in superficial deposits can be seen. Here the upper terraces have serrated margins which result from gullying by running water. Many of these gullies now appear to be virtually senile but their form, slopes and outlines differ from those produced in the more recent gullies which have been produced by slope failure on the south side of the city. Even more impressive than these features are the fresh "catsteps" and "terracettes" about Hill Road, Medicine Hat. The former are composed of overburden blocks 30 feet thick and 50 to 60 feet wide which have broken loose and moved downhill, often breaking up into smaller slivers 5 to 6 feet in width. The latter are features up to 600 feet in length, 6 to 14 feet thick and 2 to 8 feet wide, which bulge downhill but display only slight rotational effects due to movement. Again, in the lower, incised stretches of Seven-persons Creek another variation in slope development seems to depend largely upon the weathering characteristics of the overburden. In places the till has a tendency to fissure, and these lines of weakness open out faster, often resulting in erosional remnants such as hoodoo forms.

In his conclusions on slope development Schumm (1956, p. 597) wrote . . . "Vegetation and thick soil horizons aid infiltration which promotes creep and the development of the rounded humid cycle landforms characterized by declining slope retreat in later stages. In arid regions sparse vegetation and meagre soils aid rapid run off and the formation of steep, parallel retreating slopes and pediments." While these statements apply generally to the Medicine Hat area the lithology, climate and vegetation here are distinctive enough to produce mixed slope development. Realization that both these slope types can occur side by side is of more than academic interest as their occurrence could affect lines of communication and property in several localities here and elsewhere in the Prairie Provinces.

CONCLUSIONS

Although the succession and geomorphology of the superficial deposits are comparable to those already described elsewhere in Alberta and Saskatchewan, similarity of lithology does not justify correlation yet. The dominant role played by stagnating ice in fashioning the present landscape is most impressive and points up the need for still more information about "dead-ice" processes. Recent and forthcoming literature on the subject emanating from the Prairie Provinces may lead to a reappraisal of the work by Carruthers (1939) on glacial drifts, 15 to 20 years ago. In the light of new information, alternative interpretations might be possible for some puzzling features of the superficial deposits which so readily, in the recent past, have been assigned to periglacial or solifluction processes.

Recent concepts in the Prairies also support the doubts expressed by Sissons (1958) about the widespread occurrence of proglacial lakes in Great Britain. Furthermore the presence of two genetically different types of slope, often in close proximity, in both Alberta and Saskatchewan invalidates the concept that natural occurrences can be arbitrarily regimented.

ACKNOWLEDGMENTS

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ÉTUDE STATISTIQUE SUR LES STRIES GLACIAIRES DE LA RÉGION DE FORT-CHURCHILL

Michel Brochu

RÉSUMÉ: Cette note présente les résultats d'une étude sur les principales caractéristiques dimensionnelles des stries glaciaires observées sur des affleurements de quartzite de 1m² dans la région de Fort-Churchill, Manitoba. Les caractéristiques étudiées sont: la concentration des stries au m linéaire et au m², leur longueur, leur largeur et leur profondeur. Les valeurs médianes, maximums et minimums ont été calculées à partir des mesures effectuées.

ABSTRACT: This note presents the results of a study of glacial striae on quartzite outcrops in the Fort Churchill area, Manitoba. Striae number, length, and depth were measured for unit areas of 1 square metre. The median, maximum values are given for the squares studied.

INTRODUCTION

La plupart des travaux géologiques qui portent sur des régions glaciées au Pléistocène, tant en Europe qu'en Amérique du Nord, et où affleure la roche en place, signalent la présence de stries glaciaires. Or, pour autant que nous sachions, les chercheurs ont surtout porté leur attention sur la direction des stries pour en déduire le sens de la progression des inlandsis, des calottes glaciaires ou des glaciers de plateau ou de vallée.

Ainsi, il n'y a que très peu de données disponibles sur la longueur et la largeur médianes, minimums et maximums ainsi que sur la concentration, au m² ou au m linéaire, des stries glaciaires dans une région donnée. La présence, dans la région immédiate de Fort-Churchill, de très beaux affleurements de quartzite gris azoïques, sur lesquels sont admirablement conservées de nombreuses stries glaciaires, m'a permis de faire des mesures sur les caractéristiques les moins étudiées de celles-ci: longueur, largeur et concentration au m² ou au m linéaire.

MÉTHODES

Les mesures qui suivent ont été effectuées sur un vaste affleurement de quartzite de plusieurs centaines de m², sis à 2km à l'Est de Fort-Churchill. Cet affleurement est pratiquement dépourvu de végétation et laisse apparaître une multitude de stries à sa surface qui, à toute évidence, est demeurée pratiquement intouchée par le gel ou par la dissolution chimique, depuis la dernière glaciation.

MS submitted March, 1962.

Dans cette zone, l'auteur a choisi au hasard quatre carrés de 1 m de côté, espacés de 3 ou 4m chacun, et les a circonscrits à la craie sur la roche en place et sur des surfaces horizontales, ou encore, là où la pente était inférieure à 10°. Ces carrés ont été tracés de façon à ce que deux des côtés soient perpendiculaires à l'orientation générale des stries.

On a ensuite procédé aux mesures suivantes: calcul de la concentration des stries à l'intérieur de chaque carré; calcul du nombre de stries interceptées par un ou deux des côtés du carré, afin de déterminer la concentration au mètre linéaire; mesure de la longueur et de la largeur de chaque strie interceptée par un des côtés du carré, de même que de la strie de longueur maximum à l'intérieur de chaque carré. Ces séries de mesures ont servi à l'établissement des valeurs médianes.

Les observations faites à partir de la méthode des carrés ont donné des résultats qui répondent aux exigences d'un échantillonnage au hasard.

OBSERVATIONS DE TERRAIN

A—Concentration des stries:

1° Concentration au m²: pour les 4 carrés étudiés, la concentration des stries est respectivement de 12, 19, 35 et 46, ce qui fait une médiane de 27 au m² avec un maximum de 46 et un minimum de 12 (tableau I).

Dans ce nombre, on a compté toutes les stries comprises entièrement à l'intérieur de chaque carré, de même que celles recoupées par une des lignes à la craie délimitant les carrés.

2° Concentration au mètre linéaire: il s'agit du nombre de stries recoupées par l'un des côtés des carrés tel qu'indiqué au tableau I.

Tableau 1

Concentration des stries dans des carrés de 1 M de côté à Fort-Churchill

	Carré 1	Carré 2		Carré 3		Carré 4	
		(Côté est)	(Côté ouest)	(Côté est)	(Côté ouest)	(Côté est)	(Côté ouest)
Concentration au m²	12		46		35		19
Concentration au mètre linéaire	10	5	8	5	12	8	5

La concentration au mètre linéaire s'établit entre 5 et 12 avec une médiane de 8, soit une strie environ tous les 12 cm.

B—Longueur des stries:

Le tableau II indique la valeur de chacune des stries mesurées. Les médianes s'échelonnent de 12,6 à 66,6 cm tandis que la médiane des médianes est de 17,5 cm.

Les longueurs minimums mesurées sont comprises entre 3,0 et 6,4 cm avec une médiane de 4,8 cm.

Tableau 2

Limites dimensionnelles des stries glaciaires à Fort-Churchill

Longueur des stries recoupées par des côtés de 1 m

	Carré 1	Carré 2		Carré 3		Carré 4	
		(Côté est)	(Côté ouest)	(Côté est)	(Côté ouest)	(Côté est	(Côté ouest)
	18,5	11,4	7,2	24,1	25,4	66,6	34,8
	6,3	7,8	17,4	48,4	33,1	8,4	13,2
	14	31,9	3,0	19,9	30,3	99,3	12,6
	3,6	3,6	4,2	35,5	9,9	201,0	6,2
	17,5	14,1	19,8	15,9	33,4	160,2	5,7
	4,8		14,1		7,3	31,8	
	12,4		10,6		26,7	47,7	
	32,4		6,4		73,2	27,6	
	28,7				10,8		
	24,3				9		
					14		
					6,4		
Longueur médiane....	17,5	12,4		19,6		34,9	
Longueur maximum..	32,4	114		73,2		201	
Longueur minimum...	3,6	3,0		6,4		5,7	

Les longueurs maximums se situent entre 72,6 et 278,6 cm, cette dernière étant la plus longue strie mesurée.

C—Largeur des stries

Cette dimension a été mesurée pour deux côtés de deux carrés seulement. Les valeurs obtenues permettent d'établir que la largeur médiane est de 0,9 à 1,8 pour les deux carrés étudiés, que les largeurs maximums sont de 4 à 4,5 cm et la largeur maximum absolue de 4,5; les largeurs minimums sont de 0,2 à 0,5 cm. (tableau III).

D—Profondeur des stries

Toutes les stries observées ont une profondeur de 1 à 2 mm et atteignent, très souvent, une fraction de millimètre. La plupart des stries sont constituées de dépressions et de cannelures très fines à la surface de la roche; elles sont si peu profondes, si peu marquées, que, très souvent, elles sont plus perceptibles au toucher que par les yeux.

Tableau 3

Largeur des stries dans les quartzites gris de Fort-Churchill

Carré 3		Carré 4	
(Côté est)	(Côté ouest)	(Côté est)	(Côté ouest)
0,8	0,6	3,6	0,6
1,2	1,3	0,5	0,6
2,9	4,5	2,6	4
0,9	0,4	1,7	1
0,3	2,2	1,8	0,7
	0,5	2	
	0,3	2,7	
	0,6		
	0,9		
	0,6		
	1,4		
	0,2		
Largeur médiane: 0,9 Largeur maximum: 4,5 Largeur minimum: 0,2		Largeur médiane: 1,8 Largeur maximum: 4 Largeur minimum: 0,5	

INTERPRÉTATION

La conservation presque parfaite des stries sur les quartzites de la région de Fort-Churchill est un indice de l'extraordinaire résistance de ce type de roche aux effets de la gélivation et aussi, bien entendu, de l'altération chimique.

La présente note est, semble-t-il, la première étude à présenter des données statistiques sur les caractéristiques dimensionnelles des stries glaciaires. En conséquence, les mesures, analysées plus haut, bien qu'effectuées dans une zone localisée et sur un seul type de roches, donnent une première approximation numérique des dimensions des stries glaciaires. Ces données pourront servir de base de comparaison à des mesures à venir; elles seront du plus grand intérêt puisque dans les calcaires non métamorphisés, ou dans les grès, par exemple, dont la dureté est nettement inférieure à celle des quartzites, la longueur, la largeur et la profondeur maximums et médianes des stries peuvent avoir des valeurs sensiblement différentes de celles trouvées pour les quartzites.

Cette première étude n'aura de véritable intérêt en autant qu'elle sera suivie d'autres travaux détaillés portant sur la plus grande variété possible de roches cristallines, sédimentaires et métamorphiques.

MAP NOTES—FICHES CARTOGRAPHIQUES

CANADA. 1:2,000,000. Dept. Mines & Tech. Surv., *Surveys and Mapping Br.*, Ottawa, 1959. (MCR 5) Price $1.00 per sheet.

This general-purpose map of Canada printed in 10 colors covers the whole country to the polar region on 6 sheets, and is designed for use as a wall-map or as a general reference map. At the mapping scale of 1:2,000,000, one inch equals approximately 32 miles. Each sheet measures 52" by 37½", with the result that the 6 sheets mounted produce a map measuring 9'4½" by 8'8". The map is constructed on a Lambert Conformal Conic projection with standard parallels at 49° N and 77° N. For the polar region north of 80° N a modified Polyconic projection has been used in the construction.

The map was compiled basically from sheets of the National Topographic System at 1:1,000,000, with additional information from 1:500,000-series maps or larger scale topographic maps for selected areas. Administrative boundaries shown include counties, townships and ranges—information that is not included in the 1:1,000,000 N.T.S. maps. The boundaries of provincial parks are shown and the parks named. The number of populated places included exceeds the density shown on the 1:1,000,000 maps, which are fundamentally base maps for aeronautical information. Five categories are included, from settlements under 500 persons to cities over 100,000. The latter are drawn graphically according to generalized city boundaries; the remaining categories are shown by symbols and variations in type face. Railways are delineated, and the network of main highways is now included. The only indication of relief is by a sparse scattering of spot elevations, and the inclusion of glaciers by stylized drawing.

Both English and French editions are in press.

(B.V.G.)

CANADA, COMMUNICATIONS. Dept. Mines & Tech. Surv., *Geog. Br.*, Ottawa, 1962.

Railways 1:5,000,000

Major Roads 1:5,000,000

Air Lines 1:10,000,000

These 3 maps are printed in black and white on the bases used for Plates 83, 85 and 87 of the *Atlas of Canada*, 1957. They provide revised information on railway, road and airline networks but are not designed to supercede the atlas plates. The Railways map shows existing lines, single and double, and proposed lines, but does not differentiate the operating companies as does the atlas plate. The map was revised to December 1961 from information supplied by the Department of Transport. The Major Roads map lists 4 categories—paved, other, under construction, and approved for construction, whereas *Atlas of Canada* Plate 85 differentiates the Trans-Canada Highway, paved highway (4 lanes), paved highway, gravel highway, and important unimproved roads. The revised map does not detail route numbers as does the atlas plate. The Air Lines map shows scheduled, non-scheduled and foreign air lines and by means of initial letters differentiates 4 Canadian air lines and 10 foreign air lines. The map was compiled from a blue-line map, published periodically under the title "Airline routes licensed by the Air Transport Board". This map is corrected to August, 1961. The new map shows a greater density of air lines than does Plate 87 of the atlas; it demanded some generalizations over southern areas of the country, however, because of scale limitation.

(B.V.G.)

ATLANTIC PROVINCES. 1:2,000,000. Dept. Mines & Tech. Surv., *Surveys and Mapping Br.*, Ottawa, 1961. (MCR 77) Price 25 cents.

This map is reproduced without revision directly from the map "Canada" at the same scale, with the addition of sheet lines. It is planned to produce the maps as a set of separate sheets covering Quebec, Ontario, Prairie Provinces, and British Columbia.

(B.V.G.)

LAND-USE SERIES: NIAGARA PENINSULA 1:50,000, Dept. Mines & Tech. Surv., *Geog. Br.*, Ottawa.

Dunnville	30L/13E	published 1961
Welland	30L/14W	" 1962
Grimsby	30M/4E	" 1962
Niagara	30M/3W	" 1962
Brantford	40P/1E	" 1962

These five multicolored land-use maps are the first of eleven to be published for the Niagara Peninsula.

The land use is superimposed on N.T.S. 1:50,000 base maps. There are twenty land use categories: 5 urban; 7 agriculture; 6 grassland-woodland; 1 unproductive; and 1 swamp and marsh. Special boundaries such as Indian Reserves and National Parks are also indicated.

(R.F.W.)

LAND-USE SERIES: NOVA SCOTIA 1:250,000, Dept. Mines & Tech. Surv., *Geog. Br.*, Ottawa.

Halifax	11D	published 1961
Truro	11E	" 1961
Shelburne	20 O-P	" 1961
Sydney	11 K-J-N	" 1962
Canso	11 F-C	" 1962

With the publication of these five multicolored land-use maps most of Nova Scotia is completed.

The land use is superimposed on N.T.S. 1:250,000 base maps. There are sixteen land-use categories: 3 urban; 6 agriculture; 5 grassland-woodland; 1 unproductive; and 1 swamp and marsh. Special boundaries such as community pastures are also indicated.

(R.F.W.)

BOOK NOTES—FICHES BIBLIOGRAPHIQUES

PROCEEDINGS OF THE SECOND NATIONAL NORTHERN DEVELOPMENT CONFERENCE. The Alberta and Northwest Chamber of Mines and Resources, and the Edmonton Chamber of Commerce, Edmonton, 1961.215 pages, tables, graphs, maps, photos. Price $5.00.

The Conference was held in Edmonton, September 13th to 15th, 1961, and developed the theme, "Canada's New Role in Resource Development".

Speakers from private enterprise and from several levels of government participated in the conference. The proceedings were organized into four sections, "Resources", "Transportation", "Finance", and "Marketing, Processing, and Manufacturing".

The papers on resources covered metallic and non-metallic minerals, petroleum, forest resources, and power. Papers on railways and trucking, airlines, shipping, and pipelines were presented in the transportation section. In addition, a paper on the development of the Hovercraft in Great Britain was included under transportation. Three reports on the financing of northern development were given, and four addresses were presented under the section: on marketing, processing and manufacturing.

Included in the published *Proceedings* are several papers presented as introductory, dinner, and luncheon addresses at the conference. The keynote address, "Canada from the Atlantic to the Pacific to the Arctic", was presented by the Hon. W. G. Dinsdale, Minister, Dept. Northern Affairs and National Resources. Cyrus Eaton gave the dinner address entitled, "The North—Keystone of Canada's Second Century". An Eskimo, Abraham A. Okpik, formerly of the Mackenzie River delta, presented a luncheon address on the topic, "The Hunter—A Changing World—Problems of Adaptation".

The volume concludes with a record of the general summary discussion, and an outline of the resolutions of the conference.

(J.W.M.)

THE HUMAN ECOLOGY AND SOCIAL AND ECONOMIC CHANGE IN THE COMMUNITY OF TUKTOYAKTUK, N.W.T. J.D. Ferguson. Northern Coordination and Research Centre (NCRC-61-2) *Dept. Northern Affairs and National Resources*, Ottawa, 1961, 80 p., 33 photos, 2 maps, 1 graph, mimeo.

This report is based on anthropological field work carried out at Tuktoyaktuk, N.W.T. in 1957 and is divided into two parts, the first dealing with the human geography and economy of the area, and the second analyzing the social organization of both the Eskimo, and white communities.

(G.F.)

THE CARIBOU ESKIMOS OF ESKIMO POINT. J.W. Van Stone and W. Oswalt Northern Coordination and Research Centre (NCRC-59-2) *Dept. Northern Affairs and National Resources*, Ottawa, (No date, circa 1959) 33 p. mimeo.

A social study of the Eskimos at Eskimo Point on the west coast of Hudson Bay, this work presents an account of the social structure, economy and influences affecting the life of the community in recent years, and secondly an analysis of Government policies towards the Eskimo together with the author's recommendations for community development.

(G.F.)

THE ESKIMO COMMUNITY AT PORT HARRISON, QUE. William E. Willmott Northern Coordination and Research Centre (NCRC-61-1) *Dept. Northern Affairs and National Resources*, Ottawa, 1961. 197 p., 11 tables, 9 figs. 22 photographs, 9 maps, mimeo.

A detailed study of the social and economic life of the Port Harrison area Eskimos based on research carried out in 1958. The report examines the process of social change in the settlement and surrounding camps, and includes chapters on the Eskimo family, kinship, recreation, religion, education, health and on the economics of the community, with detailed descriptions and statistical summaries given in appendices.

(G.F.)

SOILS IN CANADA, GEOLOGICAL, PEDOLOGICAL AND ENGINEERING STUDIES. Univ. Toronto Press in co-operation with the Royal Society of Canada (ed. R.F. Legget). *Roy. Soc. Can.* Special Publication, No. 3. 1961. 229 p.

The seventeen papers in this volume were presented as a symposium at the 1960 meeting of Royal Society of Canada in Kingston. Seven treat the geology of soil-parent materials, six discuss soils from several pedological aspects and four deal with the significance of geological and pedological phenomena in engineering studies. The thirteen papers on geology and pedology are most useful to geographers.

Most of the papers provide general treatment of their subjects, giving concise and over-all views; each also includes an extensive bibliography.

(J.H.L.)

ONTARIO SURVEY: ECONOMIC AND SOCIAL ASPECTS. Ontario, *Department of Economics*, Toronto 1961. Ninth edition.

This economic survey of the province of Ontario is the first revision to be published since 1957 and shows improvement both in format and content. Part 1, a new addition to the report, concerns the history of the provincial government and the growth of the province, and describes the organization and function of each government department. A final section reviews Ontario's finances since 1944 and its position with respect to federal-provincial fiscal policies. Part 2 concerns the economy as a whole and provides information on population, physiography, natural resources, transportation, communications, price levels and other factors. This part of the report is well documented with tables and graphs, and illustrated by maps and photographs. Part 3 includes detailed statistical data by economic regions using the county or district as the basic unit.

(N.J.S.)

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