Geocryology

A survey of periglacial processes and environments

A. L. Washburn
Quaternary Research Center, University of Washington

A HALSTED PRESS BOOK

John Wiley & Sons
New York

Published in the U.S.A.
by Halsted Press, a Division of
John Wiley & Sons, Inc., New York

Library of Congress Cataloging in Publication Data

Washburn, Albert Lincoln, 1911–
Geocryology.
Edition for 1973 published under title:
Periglacial processes and environments.
'A Halsted Press book.'
Bibliography: p.
Includes index.
1. Frozen ground. 2. Erosion. I. Title.
II. Title: Periglacial processes and environments.
GB641.W37 1978 551.3'8 78-22026

ISBN 0-470-26582-5

Printed in Great Britain

ERRATA

Washburn: Geocryology
A survey of periglacial processes and environments.

The maps on pages 3 and 5 should be inter-changed. The captions relating to the maps (Figures 1.1 and 1.2) are correctly placed.

Page 19, left column, 4th paragraph, 4th line: L_W=80 Cal should read L_W=80 cal.

Page 246, left column, 8th line: beading should read bedding.

Contents

Acknowledgements

The author and publishers wish to thank the following for permission to use or modify copyright material. Figure numbers refer to this book, detailed citations may be found by consulting captions and references. Figures not otherwise credited are by the author.

Academic Press (NY) for Figs. 4.33, 4.34, 4.35, 5.39, 6.18, 6.22, 7.5, 12.5 and 12.6; Akademie Nauk USSR for Fig. 14.16; Alfred-Bentz-Hans for Fig. 5.1; American Association for the Advancement of Science for Fig. 4.14 © 1965; American Geophysical Union for Figs. 3.30 and 3.34; D. M. Anderson for Table 4.1; G. Andreas for Fig. 5.12; Arctic Institute of North America for Figs. 3.29 and 6.21; Association of American Geographers for Figs. 3.3 and 3.18; I Ya Baranov for Fig. 3.9; J. B. Benedict for Figs. 6.18 and 6.22; R. F. Black for Figs. 4.33, 4.34 and 4.35; J. M. Blackwell for Fig. 10.3; J. Brown for Fig. 3.16; R. J. E. Brown for Figs. 3.2, 3.4, 3.14 and 3.27; Butterworths, London for Fig. 4.1 and Table 4.2; K. L. Carey for Figs. 3.19 and 3.20; Ceskoslovenske Academia for Figs. 7.3, 7.4 and 7.5; A. E. Corte for Fig. 4.6 and Table 14.2; T. Czudek for Figs. 12.5 and 12.6; J. Demek for Figs. 12.5 and 12.6; Department of Physical Geography, University of Uppsala for Table 6.3; Dowden, Hutchinson & Ross, Inc. for Table 2.1; Ferd. Duemmlers Verlag for Figs. 5.1, 11.1 and 14.3–14.8; Franz Steiner Verlag, Wiesbaden for Figs. 5.8, 5.20, 5.32, 6.23 and 7.2; B. Frenzel for Figs. 5.8, 5.20, 5.32, 5.54, 6.23, 7.2 and 14.12; Friedr. Vieweg & Sohn, Wiesbaden for Figs. 5.54 and 14.12; Y. Fujii for Tables 3.3 and 3.4; Geo Abstracts Ltd., University of East Anglia for Fig. 6.28; Geological Society of America for Figs. 4.6 and 4.15; Geological Survey of Greenland and Meddelelser om Grømland for Fig. 3.8; Geologiska Föreningen, Stockholm for Fig. 4.13; R. P. Goldthwait for Fig. 5.39, M. A. Gonzalez for Table 14.2; K. Herz for Fig. 5.12; K. Higuchi for Tables 3.3 and 3.4; The Industrial Press (NY) for Figs. 3.1 and 3.32; Japan Association for Quaternary Research for Fig. 5.3; K. Kaiser for Figs. 14.11; Longman Group Ltd for Fig. 14.1 and Table 9.1; J. H. Leeman Ltd for Fig. 14.2; J. R. Mackay for Figs. 3.31, 3.32, 3.36, 4.5, 5.40 and 14.21; Methuen & Co. Ltd for Table 2.1; F. Müller for Fig. 5.50; National Academy of Sciences, Washington DC for Figs. 4.29, 5.21 and 5.22 and 5.23 reproduced from *Permafrost International Conference* (1966), pages 103, 78 and 79 respectively, and Figs. 3.5, 3.7, 3.17, 3.29, 5.52, 5.53 and 14.20 and Table 14.1 from *Permafrost Second International Conference* (1973), pages 38, 74, 225, 10, 82, 81, 90 and 92 respectively; National Research Council, Canada for Figs. 3.2, 4.2, 4.3, 4.23, 5.46 and 12.1 and Tables 3.3, 3.4 and 3.6; T. L. Péwé for Figs. 3.4, 5.52, 7.5 and 14.22; H. Poser for Figs. 14.3–14.8; Regents of the University of Colorado for Fig. 5.40; R. D. Reger for Fig. 7.5; S. Reiger for Table 2.1; J. C. Rogers for Fig. 3.13; Rutgers University Press, © 1977 for Table 2.2; P. V. Sellman for Fig. 3.16; R. P. Sharp for Fig. 5.33; Springer-Verlag, Wian for Fig. 11.1; L. Strömquist for Table 6.3; Tokyo Metropolitan University for Fig. 5.2; University of Chicago Press for Figs. 11.1 and 12.3; University of Toronto Press for Figs. 3.2 and 3.27; US Department of Agriculture, Forest Service for Table 2.3; A. A. Velichko for Figs. 14.13–14.16; Verlag Vandenhoeck & Ruprecht for Figs. 5.13 and 5.25; M. Vigdorchik for Fig. 3.12; E. Watson for Fig. 5.55; John Wiley & Sons (NY) for Fig. 14.1; R. G. B. Williams for Fig. 14.2 and *Zeitschrift für Geomorphologie* for Fig. 14.19.

Preface

This book is intended as a rather comprehensive overview of periglacial processes and their effects, present and past. Geocryology is included in the title to emphasize the pervasive influence of ice and its phase changes in these processes. The book is neither a formal text nor a reference manual but something of both and a guide to the enormous literature. Stress has been laid on the most recent publications, and on documentation by specific page citations for ease of reference. Because of the writer's personal interests and background, the coverage is uneven but he hopes the present edition is an improvement over the first in this respect.

Except as otherwise noted, temperatures are in degrees Celsius and the metric system is used throughout. Equivalent measurements in other units are given in parentheses if the original observations were reported in these units. A conversion table for SI units is given on the inside front cover.

Radiocarbon dates are given in years BP (Before Present, taken to mean before 1950).

The term soil, alone – as distinct from soil horizon, buried soil, or other usage where the context is clear – is used in a general sense as in engineering (cf. Legget, 1973) and does not necessarily imply profile development as in pedology.

In general capitalization, or lack of it, of terms such as Late v. late (Wisconsin, etc.), or Last Glaciation v. last glaciation, follows the author cited. Usages vary, and it should not be assumed that capitalization necessarily implies a formal time frame with well-established boundaries. Pending a universally accepted convention, and unless otherwise indicated, the present writer has adopted the informal lower-case style for his own work.

Much of this book was inspired by Professor Richard Foster Flint of Yale University. As a teacher just prior to World War II, as a fellow officer during that war, as a colleague at Yale many years later and throughout as a friend, Dick Flint encouraged the writer in many ways. An inter-disciplinary Pleistocene seminar at Yale in the early 1960s was also a stimulant in providing a fruitful discussion of periglacial processes and environments.

Professor Emeritus Hugh M. Raup of Harvard University and later of The Johns Hopkins University is another close friend who is responsible for the orientation of this book. As colleagues in fieldwork in Northeast Greenland in the period 1956–66, we had the opportunity to study and discuss many periglacial problems in detail. Numerous facets of this book benefited from this collaboration and some are based on the resulting publications in the *Meddelelser om Grønland.*

The writer is also indebted to Dr Duwayne Anderson, then of the US Army Cold Regions Research and Engineering Laboratory and presently of the State University of New York at Buffalo, and to Professor Larry W. Price of Portland State University for their suggestions relative to the first edition, and to Dr R. J. E. Brown of the National Research Council of Canada, Professor J. Ross Mackay of the University of British Columbia, and to many colleagues the world over whose suggestions benefitted the second. The writer's students at the University of Washington and Dr Chester Burrous of its Periglacial Laboratory at the Quaternary Research Center contributed more than they realized to both editions.

The publishers simplified the writer's task in many ways, and he deeply appreciates the cooperative attitude they exhibited throughout. Mrs Brenda Hall contributed greatly by preparing the index of both editions. Mrs Ruth Hertz, then on the staff of the Quaternary Research Center, typed the draft of the first edition and facilitated it in many other ways. Similarly, the second edition owes its appearance to the excellent work of Ms Patricia Leaming as editorial assistant whose sharp eye was invaluable, and of Ms Anne Swithinbank as library researcher and cartographer, both of the Quaternary Research Center.

Finally neither edition could have been written without Tahoe Washburn's wifely understanding and patient acceptance of too many uncommunicative evenings and weekends.

Conversion Table

Quantity	*SI units*	*Conversion factors*
Area	m^2	1 $cm^2 = 10^{-4}$ m^2 1 hectare $= 10^4$ m^2 1 acre $= 4.047 \times 10^3$ m^2 1 $ft^2 = 9.290 \times 10^{-2}$ m^2
Conductive capacity	$J\ m^{-2}\ s^{-\frac{1}{2}}$	1 cal $cm^{-2}\ s^{-\frac{1}{2}} = 4.187 \times 10^4\ J\ m^{-2}\ s^{-\frac{1}{2}}$
Density	$kg\ m^{-3}$	1 g $cm^{-3} = 10^3\ kg\ m^{-3}$ 1 lb $in^{-3} = 2.768 \times 10^4\ kg\ m^{-3}$
Energy, work, heat	J	1 cal $= 4.187$ J 1 BTU $= 1.055 \times 10^3$ J 1 ft lb $= 1.356$ J
Latent heat	$J\ kg^{-1}$	1 cal $g^{-1} = 4.187 \times 10^3\ J\ kg^{-1}$ 1 BTU $lb^{-1} = 2.326 \times 10^3\ J\ kg^{-1}$
Length	m	1 Å $= 10^{-10}$ m 1 ft $= 0.305$ m
Mass	kg	1 lb $= 0.454$ kg 1 ton $= 9.072 \times 10^2$ kg
Pressure	$N\ m^{-2}$	1 dyne $cm^{-2} = 10^{-1}\ N\ m^{-2}$ 1 lb $in^{-2} = 6.895 \times 10^3\ N\ m^{-2}$ 1 mm Hg (0°C) $= 1.333 \times 10^2\ N\ m^{-2}$ 1 atm $= 1.013 \times 10^5\ N\ M^{-2}$
Shearing stress	$N\ m^{-2}$	1 bar $= 10^5\ N\ m^{-2}$ (= 14.5 lb in^{-2})
Specific heat	$J\ kg^{-1}\ K^{-1}$	1 cal $g^{-1}\ °C^{-1} = 4.187 \times 10^3\ J\ kg^{-1}\ K^{-1}$ 1 BTU $lb^{-1}\ °F^{-1} = 4.187 \times 10^3\ J\ kg^{-1}\ K^{-1}$
Surface tension	$N\ m^{-1}$	1 lb $ft^{-1} = 1.459 \times 10\ N\ m^{-1}$
Temperature	K	°C = K − 273.16 °F = 1.8 (K − 273.16) + 32
Tensile strength	$kg\ m^{-2}$	1 kg $cm^{-2} = 10^4\ kg\ m^{-2}$
Thermal conductivity	$W\ m^{-1}\ K^{-1}$	1 cal $cm^{-1}\ s^{-1}\ °C^{-1} = 4.187 \times 10^2\ W\ m^{-1}\ K^{-1}$
Thermal diffusivity	$m^2\ s^{-1}$	1 $m^2\ h^{-1} = 2.778 \times 10^{-4}\ m^2\ s^{-1}$

1 Introduction

I Definition of geocryology and periglacial: *1 Geocryology; 2 Periglacial.* **II Objectives: III processes:** *1 General; 2 Frost action; 3 Other processes.* **IV Environments:** *1 General; 2 Polar lowlands; 3 Subpolar lowlands; 4 Middle-latitude lowlands; 5 Highlands.* **V References:** *1 General; 2 Frost action*

Geocryology or periglacial research as defined below began with early reports of permafrost in the USSR but the 'father' of periglacial studies was Łoziński in Poland, who in 1909 first discussed the paleoclimatic implications and coined the term periglacial. Intense impetus to development of the subject grew out of the 11th International Geological Congress excursions to Spitsbergen in 1910, whose participants included some of the most prominent geologists of the time (De Geer, 1912*b*). Their reports of the then little-known periglacial features stimulated much interest that has led to an enormous interdisciplinary literature comprising geographic, geologic, engineering and, most recently, environmental interests.

Descriptive, regional surveys and process-oriented studies have emphasized frost wedging, frost cracking, frost heaving, solifluction (gelifluction), and such features as cryoplanation terraces, pingos, and the many forms of patterned ground. Laboratory studies of frost heaving began with Taber in 1929, and since then laboratory research has become increasingly important because of the engineering implications of frost heaving and permafrost. Government research centres established for such studies now include the Centre de Géomorphologie (CNRS) at Caen in France, the Permafrost Institute at Yakutsk in the USSR, and the US Army Cold Regions Research & Engineering Laboratory (CRREL) at Hanover, NH in the United States.

Advances in regional and process studies have also advanced the paleo-environmental aspects of periglacial research, and it is now clear that the most important features pertinent to climatic and other environmental reconstructions are those indicative of permafrost. The temperature significance of these features varies somewhat with their nature, and the approximations now available remain to be refined.

Despite the accomplishments of the past hundred years, numerous aspects of periglacial studies still need to be investigated, many of which bear significantly on practical problems as well as on purely scientific questions.

I Definition of geocryology and periglacial

1 Geocryology

Most definitions of geocryology, although emphasizing frozen ground, do not exclude glaciers, and it is clear from the *English-Russian geocryological dictionary* (Fyodorov and Ivanov, 1974) that the Russian term geokriologiya is highly generalized. As defined by R. J. E. Brown and Kupsch (1974, 13), geocryology is 'The study of earth materials having a temperature below 0°C.' According to the American Geological Institute's *Glossary of geology*, geocryology is 'The study of ice and snow on the Earth, esp. the study of permafrost' (Gary, McAfee, and Wolf, 1972, 290). Permafrost is also emphasized by Poppe and Brown (1976) in equating the term geokriologiya with 'geocryology, permafrost studies', and in some instances (cf. Markuse, 1976, 118) geocryology (Geokriologie) is strictly equated with study of permafrost (Dauerfrostbodenkunde). Cryo-prefixes are increasingly common, especially in Russian, in a complex and sometimes confusing array of terms in the translated literature.

Because the term geocryology is widespread, it is used in the title of this book but in the restricted sense as applying to frozen ground (seasonally frozen ground as well as permafrost) but not to glaciers. It has the advantage of being less of a tongue twister than periglaciology; on the other hand the adjective periglacial, discussed below, is useful because it is widely employed, specifically excludes glaciers, and is less cumbersome than the adjective geocryological. Although geocryologic could have been adopted, periglacial is used in the following to accord with its wide acceptance and with the previous edition of this book.

2 Periglacial

The term periglacial was introduced by Łoziński (1909, 10–18) to designate the climate and the climatically controlled features adjacent to the Pleistocene ice sheets.

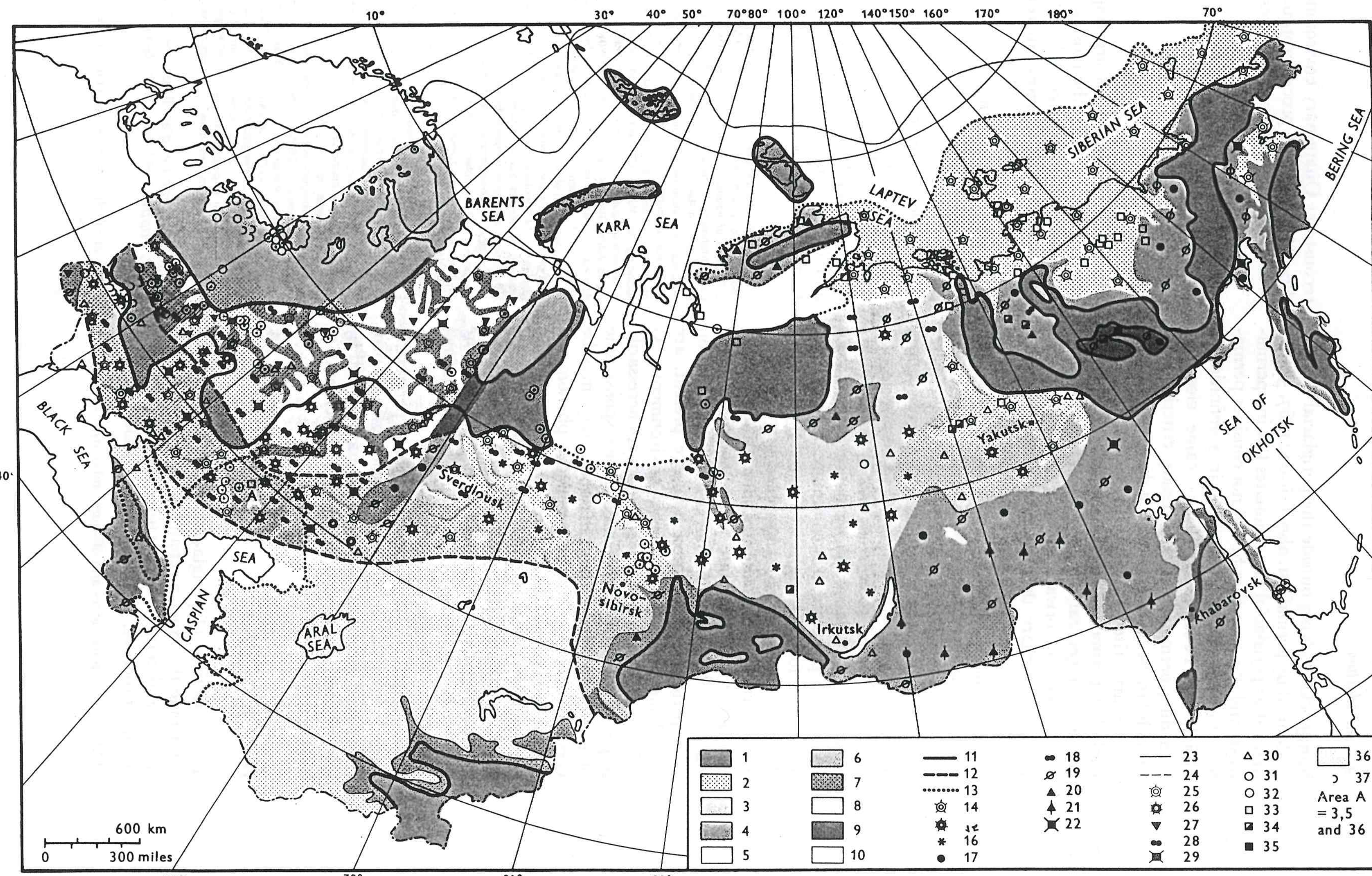
Many investigators have extended the term to designate nonglacial processes and features of cold climates regardless of age and of any proximity to glaciers. As a result there have been varying usages (cf. Butzer, 1964, 105; Dylik, 1964*a*; 1964*b*). The restricted definition is followed in the USSR where it is applied to features bordering former ice sheets in the European section, and adjectives such as geocryologic or cryogenic are employed for contemporary features. As discussed by Jahn (1975, 3–4), this usage presumably reflects the fact that frozen ground was studied in Siberia (where present or former glacial influences are largely absent) long before the term periglacial was proposed. Nevertheless, and despite criticism because of its lack of precision (Linton, 1969), the term is being widely adopted in the extended sense, as here, because of its comprehensiveness and climatic implications (cf., for example, Embleton and King, 1975; French, 1976*a*; Hamelin and Cook, 1967; Jahn, 1970; 1975; Tricart, 1967; and numerous articles in the *Biuletyn Peryglacjalny*).

The term has no generally recognized quantitative parameters, although some rough estimates of precipitation and temperature limits have been given. According to Peltier's (1950, 215, Table 1) estimate, his periglacial morphogenetic region is characterized by an average annual temperature ranging from −15° (5°F) to −1° (30°F) and an average annual rainfall (excluding snow) ranging from 127 mm (5 in) to 1397 mm (55 in). Peltier's (1950, 222, Figure 7) diagram of morphogenetic regions, which has been widely reproduced, is reasonably consistent with these figures for the periglacial region except that the lower limit of rainfall is given as 0 mm (0 in). Peltier also proposed a periglacial erosion cycle dominated by frost action and solifluction as a counterpart to Davis' fluvial ('normal') cycle (W. M. Davis, 1909). However, the concept is highly idealized and fails to take adequate account of running water (cf. French, 1976*a*, 164). According to Lee Wilson's (1968*a*, 723, Figure 9; 1969, 308, Figure 3) morphogenetic classification the periglacial domain has a precipitation range from some 50 to 1250 mm and a temperature range from some −12° (10°F) to 2° (35°F) in variable combinations. French (1976*a*, 5) defined the periglacial domain as including '. . . all areas where the mean annual air temperature is less than +3°C'. However useful for particular purposes, any such definition is arbitrary because the periglacial concept itself is sufficiently broad and imprecise to defy quantification. The diagnostic and necessary criterion is a climate characterized by intense frost action and snow-free ground for part of the year. As stated by Tricart (1967, 9), '*les pays froids sont ceux où l'action géomorphologique de l'eau est commandée par son existence à l'état solide, permanente ou périodique.*'

Tricart (1967, 29, Figure 4; 30) stressed permafrost (whether or not it is in balance with the present climate) as the primary characteristic of the periglacial domain. However, he recognized a distribution of minor periglacial features, such as earth

1 Regions of accumulation with underlying syngenetic permafrost containing ground ice 2 Regions of accumulation with underlying seasonal frost only 3 Regions of equilibrium between processes of accumulation and denudation on syngenetically frozen rocky substrate containing ice wedges 4 Flat regions of equilibrium between processes of accumulation and denudation (on syngenetically frozen rocky substrate) 5 Regions of cryoplanation, stable with respect to accumulation and denudation (on epigenetically frozen rocky substrate) 6 Regions stable with respect to accumulation and denudation (on seasonally frozen rocky substrate) 7 Regions without frozen-ground processes modelling the landscape 8 Regions of dominant denudation with underlying permafrost 9 Regions of dominant denudation with underlying seasonal frost only 10 Glaciers 11 Polygons with ice veins 12 Polygons with ice veins, associated with [agrémentés] thermokarst forms 13 Peat bogs with flat hummocks [buttes gazonées] 14 Baydjarakhs 15 Alases 16 Polygons with soil veins 17 Reduced (deformed) polygons with soil veins – forms infilled with cover loam 18 Forms similar to hummocks [buttes gazonées] with gaps [enfoncements] resulting from degradation 19 Pseudo kames 20 Hummocks [buttes gazonées] – mounds 21 Cryoplanation terraces 22 Nonsorted circles [formes tachetées – médaillons] 23 Sorted polygons [polygons de pierres], circles, and other forms sorted by freezing 24 Solifluction stripes on slopes 25 Stratified icings ['Nalédi'] 26 Seasonal hummocks [buttes gazonées] resulting from soil heaving 27 Perennial hummocks [buttes gazonées] resulting from soil heaving 28 Hummocks [buttes gazonées] due to water migration toward the frozen surface 29 Solifluction forms related to soil flow 30 Present limit of permafrost

1.1 Present and upper Holocene periglacial features in USSR *(after Markov, 1961; Popov, 1961; key translated. For commentary and somewhat different interpretations, cf. Jahn, 1976a, 131, Figure 14, 132–3)*



NOTE: THIS IS FIG. 1.2

hummocks, as lying outside the periglacial domain as defined by permafrost; also Tricart (1967, 56–67) defined as periglacial some climates lacking permafrost, and it is clear that he did not consider permafrost to be a necessary condition in a definition of periglacial. Péwé (1969, 2, 4) came closer to regarding permafrost as a necessary criterion, and French (1976*a*, 2) considered it a diagnostic but not a necessary criterion. Jahn (1975, 11) specifically excluded permafrost as a criterion, although he accepted, in general, a mean annual temperature of $-1°$ as the equatorward boundary of his periglacial zone, a boundary that in many places is in fact applicable to permafrost as noted in the next chapter and recognized by Jahn (1975, 28). Clearly, the periglacial and permafrost environments have much in common but to make them synonymous is overly restrictive, since many features such as gelifluction, frost creep, and several forms of patterned ground that are related to frost action, are commonly regarded as periglacial but are not necessarily associated with permafrost. Furthermore, it is common to speak of former periglacial environments; yet in the present state of our knowledge there are very few criteria by which a former permafrost condition can be proved. Although a cold-climate process, glaciation is not periglacial by definition. Glaciation and 'periglaciation', where both are present, are complementary aspects of cold environments. As noted by Jahn (1975, 21–2), this does not necessarily imply synchroneity in effects, since the response time of glaciers to climate change (Meier, 1965) may be very different than that of periglacial phenomena, especially to cooling. It has been argued that it may take 15 000–30 000 years for an ice sheet to build up (Weertman, 1964), yet the region fronting the eventual border could have been responding to the same climate change by developing permafrost throughout the glacier's advance.

As used in the following, the term periglacial designates primarily terrestrial,[1] non-glacial processes and features of cold climates characterized by intense frost action, regardless of age or proximity to glaciers (Figures 1.1–1.2).

II Objectives

The objectives of periglacial research are to (1) determine the exact mechanism of periglacial processes, (2) determine the environmental significance of the processes, (3) apply the information to reconstruct Quaternary environments, and (4) use these historical and process approaches as an aid in predicting environmental changes.

III Processes

1 General

Many different processes are responsible for periglacial effects, but for the most part these processes are not peculiar to periglacial environments. Rather they are common to many environments that have a climate sufficiently cold to leave physical evidence of its influence. It is the combination and intensity of these processes that characterize periglacial environments.

Given low enough temperatures both the poleward and upper altitudinal limits of most periglacial features, where limits exist, are determined by precipitation blanketing the land with perennial snow and ice; on mountains this limit is the snowline. Permafrost can be an exception, depending on the

1 Regions of maximum glaciation 2 Regions of sediment accumulation corresponding to maximum glaciation (with underlying syngenetic permafrost) 3 Regions of equilibrium relative to processes of sediment accumulation and transportation corresponding to maximum glaciation (with underlying epigenetic permafrost) 4 Regions of dominant transportation processes corresponding to maximum glaciation (with underlying permafrost of destructive origin ['d'origine génétique destructive']) 5 Extension of ocean during period of maximum glaciation 6 Regions of Valdai Glaciation 7 Regions of accumulation corresponding to Valdai Glaciation (with underlying syngenetic permafrost) 8 Regions of equilibrium relative to accumulation and transportation corresponding to Valdai Glaciation (with underlying epigenetic permafrost) 9 Regions of dominant transportation corresponding to Valdai Glaciation (with underlying permafrost of destructive origin ['d'origine génétique destructive']) 10 Regions without permafrost phenomena 11 Maximum limit of glaciation 12 Southern limit of permafrost during maximum glaciation 13 Shoreline during maximum glaciation 14 Fossil ice-wedge polygons ['Polygones a fissures remplies de glace fossile'] developed during maximum glaciation 15 Soil-wedge polygons ['Polygones a fissures remplies de sol'] corresponding to maximum glaciation 16 Hummocks [buttes gazonées] and concave forms corresponding to maximum glaciation 17 Cryoplanation terraces corresponding to maximum glaciation 18 Nonsorted circles ['formes (taches) en médaillon'] corresponding to maximum glaciation 19 Sorted polygons ['Polygones de pierres'], circles, etc., corresponding to maximum glaciation 20 Solifluction stripes ['Bandes de solifluxion sur versant'] corresponding to maximum glaciation 21 Perennial hummocks [buttes gazonées], due to frost heaving, corresponding to maximum glaciation 22 Solifluction forms corresponding to maximum glaciation 23 Limits of Valdai Glaciation 24 Southern limit of permafrost during Valdai Glaciation 25 Fossil ice-wedge polygons ['Polygones à fissures remplies de glace fossile'] developed during Valdai Glaciation 26 Soil-wedged polygons ['Polygones a fissures remplies de sol'] corresponding to Valdai Glaciation 27 Degrading polygons with soil veins (pebbles in cover loam) corresponding to Valdai Glaciation 28 Nonsorted circles ['formes (taches) en médaillon'] corresponding to Valdai Glaciation 29 Solifluction forms corresponding to Valdai Glaciation 30 Discovery site of Upper Paleolithic fauna 31 Periglacial fossil macroflora 32 Periglacial fossil pollen and spores 33 Mammoth remains in permafrost ['tjäle'] 34 Rhinoceros remains in permafrost ['tjäle'] 35 Horse (*Equus caballus*) in permafrost ['tjäle'] 36 Regions of loess ['d'apparition de loess et de formations loessiques'] 37 Regions with ancient continental dunes

1.2 Pleistocene periglacial features in USSR (*after Markov, 1961; Popov, 1961; key translated*)

[1] Marine ice-shove ridges and offshore (submarine or subsea) permafrost are included as periglacial features by the present writer.

(4) use
ı aid in

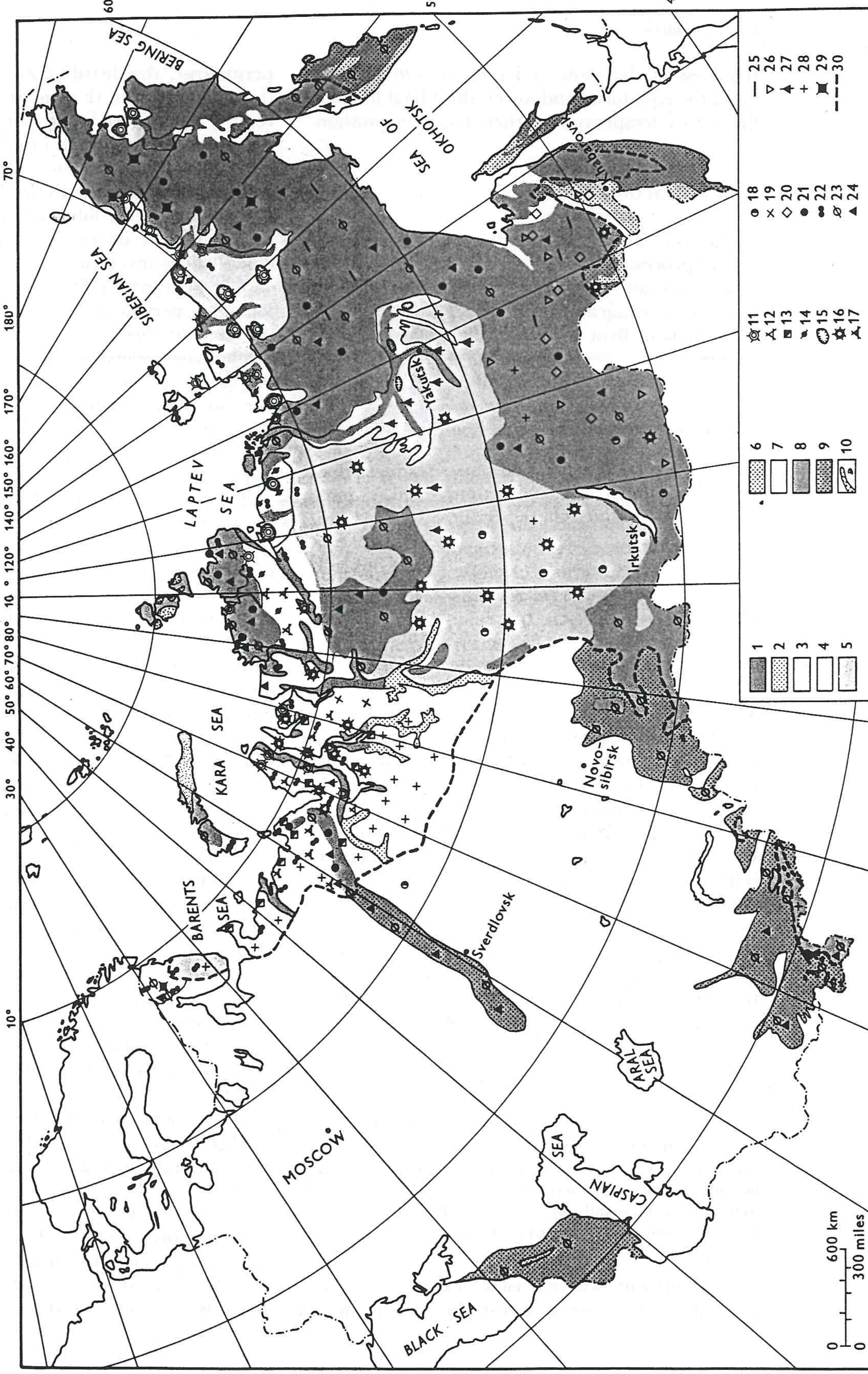
or peri-
rocesses
Rather
at have
vidence
ntensity
al envi-

ne pole-
riglacial
by pre-
ial snow
nowline.
g on the

ılation corre-
permafrost)
nulation and
ng epigenetic
esponding to
in ['d'origine
of maximum
ılation corre-
permafrost)
ı correspond-
9 Regions of
h underlying
10 Regions
12 Southern
ing maximum
nplies de glace
lge polygons
um glaciation
g to maximum
um glaciation
ding to maxi-
, circles, etc.,
andes de soli-
21 Perennial
ling to maxi-
um glaciation
during Valdai
nplies de glace
lygons ['Poly-
ai Glaciation
corresponding
en médaillon']
esponding to
31 Periglacial
nmoth remains
e'] 35 Horse
l'apparition de
tinental dunes

er Markov,

ent writer.



NOTE: THIS IS FIG. 1.1.

thickness of the snow or ice cover. On the other hand the equatorial and lower altitudinal limits are limited by temperature rather than precipitation.

2 Frost action

By far the most widespread and important periglacial process is frost action. Actually, frost action is a 'catch-all' term for a complex of processes involving freezing and thawing including, especially, frost cracking, frost wedging, frost heaving, and frost sorting.

3 Other processes

In addition to frost action, certain aspects of mass-wasting, nivation, fluvial action, lacustrine action, marine action, and wind may produce characteristic periglacial features.

IV Environments

1 General

Many factors make up environment but for periglacial environments the overriding controls are regional climate and topography. Local factors may modify the regional climate and in this and other ways strongly influence processes.

Various periglacial zones based on different criteria reflecting cold climates have been suggested. Büdel (1948) recognized two zones – a frost-debris zone (Frostschuttzone) characterized by barren stony surfaces, and a tundra zone (Tundrenzone) characterized by treeless vegetated surfaces. Corbel (1961, 19, Figure 11) stressed the role of precipitation by outlining 3 arctic periglacial zones based on differences in precipitation. In Jahn's (1975, 11) opinion

> The periglacial zone is unquestionably a climatic zone. Its boundaries are determined neither by winter (frozen ground) nor summer (vegetation) temperatures but rather by the mean annual temperature curve. The annual isotherm of -1°C is perhaps nearest to that limit [outer boundary], although periglacial phenomena often overstep this line, especially where residual Pleistocene permafrost is present in the ground in areas under forest, like in Siberia.

In the present writer's view, except for a few phenomena that are demonstrably dependent on permafrost, the distribution and climatic implications of features that most investigators would readily accept as being periglacial are still too poorly known to permit grouping the features into generally applicable zones. Detailed local zonation studies should eventually lead the way to broader applications. The following types of climate and their influence on processes illustrate some of the complexities involved.

Tricart (1967, 44–67; 1969, 19–27) recognized the following periglacial climates:

A *Cold dry climate with severe winters*

Encompasses elements of the D and E climates of Köppen, discussed later. Characteristics include (*a*) very low winter temperatures, (*b*) short summers, (*c*) permafrost, (*d*) low precipitation, (*e*) violent winds. Consequently there is (*a*) intense freezing, (*b*) reduced or even negligible activity of running water, and (*c*) important wind action.

B *Cold humid climates with severe winters*

(i) *Arctic type:* Corresponds to the most humid parts of Köppen's ET climates, excluding those without marked seasons. Characteristics include (*a*) similar mean temperatures to A but with a tendency for smaller annual range, (*b*) permafrost, (*c*) great climatic irregularities that tend to be masked by mean figures, (*d*) greater humidity with annual precipitation totals almost always exceeding 300 mm, resulting in appreciable snow cover and some rain. Consequently, as compared with A, (*a*) freezing is less intense and less long, (*b*) wind action is reduced by the snow cover, (*c*) running water is more important.

(ii) *Mountain type:* Corresponds to prairie-alpine zone of temperate zone. Characteristics include (*a*) monthly temperature trends similar to B(i) but with higher means, (*b*) precipitation is much greater than in B(i) and tends to inhibit wind action and deep penetration of frost, (*c*) freeze-thaw cycles are less common in the summit areas than in the valleys. Consequently (*a*) frost action is important but permafrost is commonly lacking, (*b*) running water is an important geologic agent, (*c*) wind action is slight.

C *Cold climates with small annual temperature range*

(i) *High-latitude island type:* Characteristics include (*a*) mean annual temperature near 0° with small annual range (generally on the order of 10°), numerous freeze-thaw cycles, (*b*) instability of weather, (*c*) considerable precipitation, generally exceeding 400 mm, which tends to inhibit wind effects. Consequently (*a*) frost action is characterized by many freeze-thaw cycles of short duration and slight penetration into the ground, (*b*) wind action is slight.

(ii) *Low-latitude mountain type:* Characteristics include (*a*) lack of seasonal temperature variations, (*b*) considerable variations in diurnal temperature, far exceeding the seasonal variations, (*c*) high precipitation except in arid mountains such as Puna de Atacama. Consequently (*a*) there is considerable frost action because of the frequent freeze-thaw cycles, (*b*) only slight frost penetration into the ground, (*c*) absence of permafrost, (*d*) lack of wind action in high-precipitation areas.

As stated by J. L. Davies (1969, 13–14), this classification brings out some marked contrasts. Permafrost is characteristic of A, absent in C. Annual temperature cycles are of large amplitude and extend to considerable depth in A; the opposite is true in C. Freeze-thaw cycles are much fewer in A than in C. Running water is much less important in A than in C. In each case conditions in B tend to be intermediate.

In some respects Tricart's classification may be superior to Köppen's but the latter has the advantage of more precise boundaries between categories.

The Köppen classification of climates (Köppen, 1936; Köppen-Geiger, 1954; cf. Strahler, 1969, 224–30) is widely used and is the one on which the following scheme is based for polar, subpolar, and middle-latitude lowland climates and geographic zones. That these climates and zones can be similarly designated arises from the fact that the zones are climatically defined. The addition of highlands is to emphasize the role of altitude in determining environmental factors. Only the most general kind of classification of climates and zones is used here in view of the limited knowledge concerning the distribution and frequency of periglacial processes and features. Gerdel (1969) has cited some of the practical limitations to climatic classifications and has presented a useful general description of the characteristics of cold regions, especially in the Northern Hemisphere.

2 Polar lowlands

In the polar zone, the average temperature of the coldest month is $< -3°$ and of the warmest month $< 10°$. The zone, which is controlled by polar and arctic air masses in the Northern Hemisphere, lies roughly north of lat. 55°N and south of lat. 50°S. It includes an ice cap (Köppen's EF) climate, dominated by arctic and antarctic air masses, in which the average temperature of the warmest month is $< 0°$, and, of more concern to periglacial research, a tundra (Köppen's ET) climate in which the average temperature of this month is $> 0°$.

The zone is characterized by ice caps, bare rock, and/or vegetation of tundra types – mainly grasses, sedges, small flowering plants, and in places herbaceous shrubs.

3 Subpolar lowlands

In the subpolar zone the average temperature of the coldest month is $< -3°$, the temperature of the warmest month is $> 10°$ but there are less than four months above this temperature (Köppen's Dfc, Dfd, Dwc, Dwd climates). The zone is controlled by continental air masses and extends roughly from lat. 50° to 70°N (northern taiga zone) and from lat. 45° to 60°S. Because the climatic gradients vary with longitude there is some latitudinal overlap.

The 10° isotherm for the warmest summer month, which is accepted here as the boundary between polar and subpolar lowlands, tends to coincide with treeline in the Northern Hemisphere. Characteristically, coniferous forest predominates. However, treeline (i.e. the northern limit of scattered trees) can be some $1\frac{1}{2}$ degrees of latitude north of the coniferous forest (cf. Washburn, 1951, 270–1).

4 Middle-latitude lowlands

In the middle-latitude zone the average temperature of the coldest month is $< -3°$ but there are more than four months with average temperature $> 10°$ (Köppen's Dfa, Dfb, Dwa, Dwb climates). This zone is controlled by both polar and tropical air masses and extends roughly from lat. 35° to 60°N.

5 Highlands

A highland climate may differ in important ways from climates primarily controlled by latitude. For instance, highlands are commonly colder than lowlands in the same zone, some highlands have stronger diurnal than seasonal temperature changes, and the orientation of mountain slopes tends to exert a strong climatic influence. Particularly in periglacial environments where topography is a critical factor, it is desirable to make a marked distinction between lowland zones controlled primarily by latitude and highland zones controlled by altitude as well as latitude.

In a general way counterparts to cold lowland zones can be found in highlands, and parallels are frequently cited between arctic and alpine zones and between subarctic and subalpine zones. However, there is considerable difference of opinion regarding zone classifications, many of which are based on biological considerations reflecting climate rather than being based directly on climatic parameters (Löve, 1970). No consistent attempt is made in the following to adopt any particular highland zonation except to recognize that the altitudinal treeline, like the latitudinal treeline, is a critical boundary for certain periglacial processes. Rather, highlands are cited in relation to polar, subpolar, middle-latitude, and low-latitude zones to indicate that certain periglacial processes and features are more common in highlands than in lowlands of the same latitudinal zone, either because of a more rigorous climate in the highlands or because of topography. As an order of magnitude, 1000 m will be considered the minimum altitude difference between a highland and a lowland in the same general region.

V References

1 General

Although periglacial research began to emerge as a recognized field following Łoziński's (1909) work, only recently has the subject been comprehensively treated. The Polish and Russian literature is particularly rich but the present writer has had to forego much of it that has not yet been translated. General references include: Büdel (1977, 37–91), Cailleux and Taylor (1954), J. L. Davies (1969), Embleton and King (1975), French (1976*a*), Jahn (1970; 1975); Kaplina (1965), Kudryavtsev (1978) Lliboutry (1965, 927–1007), Peltier (1950), Poser (1977), L. W. Price (1972), Romanovskiy (1977), Tricart (1963; 1967; 1969), Troll (1944; 1958), Vtyurina (1974; 1976), and the *Biuletyn Peryglacjalny* (1954–). The most detailed updated monograph in English at the research level is the one by Jahn (1975), which is especially valuable because of its discussion of the extensive Russian literature, much of it unavailable in translation. A booklet of permafrost definitions has been compiled by R. J. E. Brown and Kupsch (1974) and an extensively illustrated English and French glossary of periglacial features by Hamelin and Cook (1967). There is also a very useful *English-Russian geocryological dictionary* (Fyodorov and Ivanov, 1974) and *Russian-English glossary of permafrost terms* (Poppe and Brown, 1976).

Regional studies of contemporary periglacial phenomena include Bird (1967, 161–270), Bout (1953), Boyé (1950), Büdel (1960), Kelletat (1969), Jan Lundqvist (1962), Malaurie (1968), Markov and Bodina (1961; 1966), Popov (1961), Schunke (1975*a*), Sekyra (1960), and many others (cf. Poser, 1974*a*; 1977). An album of contemporary permafrost features for use in university courses was compiled by Popov (1973).

A continuing survey and bibliography of field investigations concerning the nature and rate of periglacial processes has been started by the Coordinating Committee for Periglacial Research, a committee of the International Geographical Union's (IGU) Commission on Present-Day Geomorphological Processes (French, 1976*c*).

2 Frost action

Frost action, including permafrost, is at the centre of much periglacial research. The literature here is voluminous and only some of the most comprehensive and helpful bibliographies and references are cited below. (Cf. also references cited under Significance in section on Permafrost in the chapter on Frozen ground.)

Akademiya Nauk SSSR (1973), American Meteorological Society (1953), Andersland and Anderson (1978), Arctic Institute of North America (1953–1975), Beskow (1935; 1947), Canada National Research Council (1978)[2], Carleton University and École Nationale des Ponts et Chausées (1978), Dostovalov and Kudryavtsev (1967), Highway Research Board (1948; 1952*a*; 1952*b*; 1957; 1959; 1962; 1963; 1969; 1970; 1972), Jessberger (1978), T. C.

[2] *Third International Conference on Permafrost – Proceedings*. Volume 1 consists of 139 submitted papers, Volume 2, of 8 review papers, 4 Chinese contributions, and other material. Field guides and 2 volumes of English translations of foreign-language papers complete the publication series. Volume 1 of the *Proceedings* was issued after the present work had been submitted to the publisher but, through the kindness of the authors, provision had been made to cite selected papers, and references were added in proof. The other publications had not yet appeared by the time this note was added.

Johnson *et al.* (1975), Jumikis (1977), Melnikov and Tolstikhin (1974), S. W. Muller (1947), National Academy of Sciences (1973; 1978[3]), National Academy Sciences-National Research Council (1966), Norges Teknisk-Naturvitenskapelige Forskningsråd og Statens Vegvesens Utvalg for Frost i Jord (1970–), Organisation for Economic Co-operation and Development (1973), Popov (1967; 1969–), Protas'yeva (1967; 1975), Terzaghi (1952), US Army Corps of Engineers Cold Regions Research and Engineering Laboratory (1951–; 1973), J. R. Williams (1965), P. J. Williams (1967).

In parallel with the survey of field investigations, noted in the previous section, a continuing survey and bibliography of laboratory investigations of frost action has also been activated by the IGU Co-ordinating Committee for Periglacial Research (Pissart *et al.*, 1976).

[3] *USSR Contribution Permafrost Second International Conference.* This volume, consisting of papers translated from the Russian, appeared after the present work had been submitted to the publisher. Although it has been possible to insert references to several of the translations, a number of pertinent ones had to be omitted.

2 Environmental factors

I Introduction: II Basic factors: *1 Climate: a* Scale, *b* Climatic parameters, *c* Zonal climate, *d* Local climate, *e* Microclimate; *2 Topography; 3 Rock material: a* General, *b* Structure, *c* Mineral composition, *d* Texture, *e* Colour; *4 Time; 5 Human activities.* **III Dependent factors:** *1 Snow cover; 2 Liquid moisture; 3 Vegetation*

I Introduction

The most nearly independent environmental factors influencing the various periglacial processes and the development and distribution of frozen ground are climate, topography, and rock material. Their degree of independence depends on the scale. For instance, zonal climate – a climate determined by the largest-scale factors of latitude, atmospheric circulation systems, and widespread highlands – is an independent factor. However, smaller-scale climatic effects are dependent on superimposed influences, some of which may result in an azonal climate – a climate that is atypical and not truly representative on a zonal basis.

The mutual interaction of climate, topography, and rock material is illustrated by countless examples. Thus topography, through altitude and exposure, modifies zonal climate but climate can modify topography by determining the processes acting on a region – for example, glaciation produces cirques that create variations of exposure and thereby influence the local climate. The nature of a rock influences the effect climate may have on it but climate may determine the kind of rock developed – for example, an evaporite formed in an arid basin.

In contrast to the foregoing factors, snow cover, liquid moisture, and vegetation are always dependent. However, they can be critical controls of periglacial processes, especially frost action, since they can determine whether or not a process is climatically zonal or azonal in its effect.

The following generalized review is to set the stage for later discussion of the environmental implications of periglacial processes. Environmental factors with special reference to permafrost have been reviewed in some detail by R. J. E. Brown (1978) and R. J. E. Brown and Péwé (1973, 72–80).

II Basic factors

1 Climate

a Scale In discussing the influence of climate, it is useful to recognize three scales of climate: zonal climate, local climate, and microclimate. As noted, zonal climate reflects only large-scale effects such as latitude and widespread highlands; it is the critical element to establish when determining past climates and reconstructing climatic changes. Local climate represents the combined influence of zonal climate and local topography. Microclimate is at a still smaller scale in that it incorporates the additional influence of ground-surface characteristics such as vegetation, moisture, and air-earth or air-water interface effects (Geiger, 1965). As indicated below, local and microclimatic influences can be highly significant and must be evaluated in reconstructing past zonal climates from biologic and geologic evidence.

b Climatic parameters The most important climatic parameters controlling periglacial processes are temperature, precipitation, wind, and their seasonal distribution. The past influence of wind may be readily discernible through erosional and depositional effects but temperature and precipitation may interact so that their relative importance may be difficult to determine. For instance, both sufficiently low temperature and sufficiently high moisture are required for glaciation and for certain frost-action effects, but within limits one parameter may substitute for the other. Thus increased snow accumulation and nivation may result from lower summer temperatures and less melting as well as from increased winter snowfall. On the other hand, increased permafrost may be due as much to lower winter snowfall, and hence less insulation of the ground, as to lower air temperatures.

c Zonal climate Present-day climate can be studied and described by quantitative observations that allow for local and microclimatic effects in describing zonal climate. This is much more difficult to do in reconstructing past zonal climates, since the evidence at any one place may be very strongly influenced and in some cases dominated by purely local and microclimatic effects.

d Local climate The extent to which local topography can influence climate is seen in the differing precipitation regime of windward and lee slopes and in the differing temperature regime of north and south slopes of mountains, especially in high latitudes. For instance, on the Jungfrau in Switzerland there were 196 days yr^{-1} with air fluctuations through the freezing point on the south slope whereas on the north slope, 127 m lower, the number was only 22 (Mathys, 1974, 57). Other factors being constant, depth of thawing in the Northern Hemisphere may be 50 to 60 per cent greater on south-facing than on north-facing slopes (Zhestkova *et al.*, 1961, 50; 1969, 8), although the opposite effect has also been observed, presumably because the lower near-surface soil temperatures on the north slope studied resulted in slower evaporation after rainfall (Hannell, 1973, 181). Such differences can strongly affect the nature and distribution of periglacial processes, even in the same valley as described, for example, by P. G. Johnson (1975). A study of Chitistone Pass, Alaska, showed that in this alpine permafrost environment, 'Topographic factors – slope and aspect – explain more of the spatial-temporal variations in active layer thermal regimes than the thermal and radiative properties of specific sites' (Brazel, 1972, 57).

e Microclimate Meteorological observations at the common shelter height (1.7–2.0 m in Europe, 4 ft in the United States) do not adequately indicate the climate at the ground surface, yet this is where some critical processes occur. For instance, the number of freeze-thaw cycles at the surface can be very different from those a little higher in a shelter. Wind velocities are also significantly different. Microclimate can vary within very short distances laterally as well as vertically, and is strongly influenced by differences in distribution of snow and vegetation. In this sense it is more of a dependent than basic factor (cf. below) but the interaction can be complicated. Microclimatic effects in periglacial environments can be very large as illustrated by observations in Antarctica (MacNamara, 1973, 201–4; 220–31). In the Mackenzie Delta, northern Canada, where mean annual air temperatures are −9° to −10°, microclimatic factors can cause the mean annual ground-surface temperature to be 5° to 10° warmer over a distance of about 160 m (M. W. Smith, 1975, 1423).

2 Topography

The way topography can influence climate and thereby, indirectly, process is outlined above. Topography can also exert a direct effect on process. For instance, the configuration and gradient of a hillside can determine whether the dominant mass-wasting process is landsliding as opposed to frost creep or solifluction. A very low-angle slope favours retention of moisture and development of certain forms of patterned ground.

3 Rock material

a General The term rock material as used here covers structure, mineral composition, texture, and colour, and includes both bedrock and unconsolidated material.

Bedrock is essentially an independent factor in relation to periglacial processes, since it is usually a pre-existing, very much older feature. Its influence on the distribution of periglacial features can be critical as illustrated by numerous studies, including Brosche's (1977, 180–8; 1978, 56–61) discussion relative to the Iberian Peninsula, and Graf's (1971, 48–51; 110–13) studies in the Swiss Alps and the Bolivian and Peruvian Andes.

From an engineering viewpoint unconsolidated material derived from bedrock is simply soil. Formerly, for soil to qualify as such in pedologic usage, it had to contain humus but many pedologists now accept mineral soil (i.e. lacking humus) as a true soil (Tedrow, 1977; Ugolini, 1979). Examples of such soil are common in some cold environments. Several classifications of cold-climate soils are in use (Tables 2.1–2.2). Because of its logical (although complex) construction, the soil classification system of the US Department of Agriculture (Soil Survey Staff, 1975) is tending to replace the older Zonal Classification in the United States but there is as yet no universally accepted classification. Tedrow (1977, 267–81) has recently proposed a genetic system for polar soils, based on first order

Table 2.1. Classification of cold-climate soils (*after Rieger, 1974; F. C. Ugolini personal communication, 1977; Ugolini, 1979*)

Cryic Great Subgroup of US Department of Agriculture Soil Classification System[1]

Name	*Description*
Pergelic Cryaquepts	Poorly drained grey or mottled soils with thin (usually less than 20 cm) surface organic mats
Histic Pergelic Cryaquepts	Poorly drained grey or mottled soils with thick surface organic mats
Ruptic-Histic Pergelic Cryaquepts	Poorly drained grey or mottled soils with a thick organic mat in part of each pedon (as in a trough between polygons) and a thin or no mat in other parts; pedons commonly include frost circles
Pergelic Cryaquolls	Poorly drained grey or mottled soils with dark base-rich upper horizons
Pergelic Sphagno-fibrists	Fibrous *sphagnum* moss peat, at least 60 cm thick
Pergelic Cryofibrists	Fibrous sedge peat, at least 40 cm thick
Pergelic Cryohemists	Partially decomposed peat, at least 40 cm thick
Pergelic Cryorthents	Well-drained loamy, clayey or gravelly soils with no developed horizons
Pergelic Cryopsam-ments	Well-drained sandy soils with no developed horizons
Pergelic Cryochrepts	Well-drained brown soils with very thin or no dark upper horizons
Pergelic Cryumbrepts	Well-drained soils with dark, acid upper horizons and, usually, brown subsoil horizons
Pergelic Cryoborolls	Well-drained soils with dark, base-rich upper horizons
Pergelic Cryorthods	Well-drained soils with a thin bleached upper horizon and a brown or reddish brown subsoil horizon in which aluminum, iron and organic carbon have accumulated
Ruptic-Entic Pergelic Cryochrepts, Ruptic-Entic Pergelic Cryumbrepts, etc.	Well-drained soils in which part of each pedon is a frost scar with no recognizable horizons
Lithic Cryorthents, Lithic Cryochrepts, etc.	Well-drained soils with bedrock less than 50 cm deep

Great Soil Groups of Zonal Classification

Name	*Description*
Lithosols	Well-drained soils with shallow bedrock (Lithic Cryorthents)
Regosols	Deep, well-drained soils with no genetic horizons (Pergelic Cryorthents and Cryopsamments)
Polar Desert soils	Well-drained soils of the Polar Desert Zone with no genetic horizons (Pergelic Cryorthents)
Arctic Brown soils	
Normal phase	Well-drained brown soils with or without dark upper horizons (Pergelic Cryochrepts, Cryumbrepts, and Cryoborolls)
Shallow phase	Soils as above with shallow bedrock (Lithic Cryochrepts, Cryumbrepts, and Cryoborolls)
Moderately-well drained phase	Soil gradational between normal phase and Tundra soils, with gleying in lower horizons (Pergelic Cryaquepts)
Rendzinas	Dark soils over shallow calcareous rock (Lithic Cryoborolls)
Podzols or Podzol-like soils	Soils with bleached upper horizons and brown lower horizons (Pergelic Cryorthods)

[1] Five orders of the US Department of Agriculture soil classification system (Soil Survey Staff, 1975) occur in the Arctic: Entisols – soils with little or no evidence of pedogenesis (formative element *–ent*, applied in various combinations); Histosols – soils derived mainly from organic matter (formative element *–ist*); Inceptisols – soils affected by pedogenesis and loss of minerals without appreciable accumulation horizons of clay or mixtures of Al, Fe, and organic C (formative element *–ept*); Mollisols – soils with dark base-rich upper mineral horizon (formative element *–oll*); and Spodosols – soils with an accumulation horizon of amorphous mixtures of Al, Fe, and organic C (formative element *–od*).

Table 2.1. Classification of cold-climate soils—*cont*

Tundra soils Upland Tundra	Relatively drier poorly drained soils with brown and yellow colouration (mottles) in the upper horizon and with thin organic mats (Pergelic Cryaquepts and Cryaquolls)
Meadow Tundra	Wetter soils with dark grey colours and somewhat thicker organic mats (Pergelic Cryaquepts)
Half-Bog soils	Saturated soils with organic mats approximately 15 cm to 30 cm thick (Histic Pergelic Cryaquepts)
Bog soils	Saturated soils with thick organic accumulations (all Histosols)

Cryosolic Order of Canadian Classification

Name	*Description*
Turbic	Mineral soils strongly cryoturbated and generally associated with patterned ground
Orthic Turbic Cryosols	Soils strongly cryoturbated with tongues of intermixed mineral and organic material, imperfectly to moderately well drained (equivalent to Upland tundra and Forest tundra)
Brunisolic Turbic Cryosols	Soils less affected by cryoturbation and having unbroken Bm horizon similar to the Orthics
Regosolic Turbic Cryosols	Soils lacking Bmy or Bm horizons, surface organic horizon may be present but organic intrusions are lacking in subsoil (generally in high Arctic or Alpine areas)
Gleysolic Turbic Cryosols	Poorly drained soils with mottles and low chromas, and Cg or Bg horizons at mineral surface
Static	Mineral soils without strong cryoturbations
Orthic Static Crysols	Soils having a gleyed Bm horizon above the permafrost table
Brunisolic Static Cryosols	Similar to Orthic subgroup but with thicker Bm horizons. (Equivalent to Arctic Brown)
Regosolic Static Cryosols	Except for Ah horizon, soils without pedogenic horizons
Gleysolic Static Cryosols	Poorly drained soils with mottles and low chromas at mineral surface. Peaty layers common at surface.
Organic soils Fibric Organic Cryosols	Soils with organic layers greater than 1 m thick and composed of fibric material in control section
Mesic Organic Cryosols	Soils similar to fibric but organic material is dominated by mesic material in control section
Humic Organic Cryosols	Soils similar to Mesic Organic Cryosols except that organic content is dominated by humic material in control section

Arctic and Tundra Zones Genetic Soil Groups and Subgroups of USSR Soil Classification

Name	*Description*
Arctic soils	Mean July air temperature for A horizon: 2°–6°
Desert-Arctic	Well-drained primitive turfy soils with slight accumulation of humus. Some weak gley in active layer. Associated with polygonal terrain (Regosols? Pergelic Cryorthents)
Sod-Arctic	Well-drained soils developed under sodded tracts (polar steppe) with Al and B horizons. Carbonates may be present in profile (Arctic Brown Soils, Pergelic Cryochrepts, Cryumbrepts, and Cryoborolls)
Tundra soils	Mean July air temperature for A horizon: 6°–10°
Sod Tundra	Well-drained soils of uplands (Pergelic Cryorthents, Cryochrepts, Cryumbrepts, Cryoborolls)
Alluvial Tundra-Sod	Well-drained soils of flood plains (Pergelic Cryorthents)
Gley Tundra	Poorly drained soils (Pergelic Cryaquepts)
Water-logged (moss-gley) Tundra	Poorly drained soils of low positions (Histic Pergelic Cryaquepts)
Bog Tundra	Soils with standing free water (Histosol and Histic Pergelic Cryaquepts)

Table 2.2. Polar soil classification (*after Tedrow, 1977, 267–81, Tables 15-2a—15-2d*)[1]

First order	Tundra soil zone	Polar desert soil zone	Subpolar desert soil zone	Cold desert soil zone (Antarctica)
Second order	Well-drained soils Arctic Brown soil Podzol-like soil Mineral Gley soils Upland Tundra soil Meadow Tundra soil Organic soils Bog soils Other soils Ranker soil Rendzina soil Shungite soil Grumusols Lithosols Regosols Soils of the solifluction slopes	Well-drained soils Polar Desert soil Arctic Brown soil Mineral Gley soils Upland Tundra Meadow Tundra Soils of the hummocky ground Soils of the polar desert-tundra interjacence Organic soils Bog soils Other soils Regosols Lithosols Soils of the solifluction slopes (may be a form of gley soil but usually well drained)	Well-drained soils Polar Desert soil Arctic Brown soil Mineral Gley soils Upland Tundra Meadow Tundra Soils of the hummocky ground Soils of the polar desert-tundra interjacence Organic soils Bog soils Other soils Regosols Lithosols Soils of the solifluction slopes	Ahumic (frigic) soils Ultraxerous Xerous Subxerous Ahumisol Evaporite soils Ornithogenic (Avian) soils Other soils Protoranker Algae peats Hydrothermal soils Regosols (recent soils) Lithosols

[1] The present table differs from the original in capitalizing such terms as Arctic Brown, etc., in conformity with Table 2.1.

soil zones and second order Great Soil Groups, followed by three further orders. His classification (Table 2.2) is thus very similar to the Great Soil Groups of the Zonal Classification but provides for genetic types of the Great Soil Groups to appear in more than one soil zone. Recent surveys of polar soils include Tedrow's (1977) monographic overview based on a lifetime's research, and Walton's (1972) worldwide review of High Arctic soils. Many fine-grained, cold-climate soils are characterized by a vesicular texture as a result of freezing, thawing, and drying (Bunting, 1977).

Soil can be a dependent or independent factor. Since it results from processes acting on bedrock, it is dependent on climate to the extent that the processes are climatically controlled. However, in many places the soil, like bedrock, predates the situation being considered and in this sense is independent. Furthermore there is such a variety of ways in which the unconsolidated material can vary that even though it is a function of one periglacial process it could be considered an independent factor in relation to a different periglacial process.

Rock material is less dependent on topography than vice versa, and in this respect is also essentially an independent factor. Therefore despite intimate interactions between climate, topography, and rock material, it is not circular reasoning to regard rock material as an independent as well as a dependent variable in considering periglacial processes.

As noted above under Microclimate and below under Colour, ground-surface temperatures can be very different from air temperatures. Another illustration of this is provided by observations in an alpine permafrost environment where the rock-surface measured 42°, four times the air temperature 10 cm above the rock (Mathys, 1974, 49, 61, Figure 5).

A convenient quantity by which to compare the ability of materials to adjust to cyclical changes of temperature is known as conductive capacity or contact coefficient, and is defined as $\sqrt{\rho c K}$ (cal cm^{-2} s$^{-\frac{1}{2}}$), where ρ = density, c = mass specific heat, K = thermal conductivity. Rock material can vary considerably in conductive capacity. Thus a cyclical temperature change, other conditions remaining constant, will reach some three times deeper in sandy clay (conductive capacity = 0.037 cal cm^{-2} s$^{-\frac{1}{2}}$) than in dry sand (conductive capacity = 0.011 cal cm^{-2} s$^{-\frac{1}{2}}$) (cf. Gold and Lachenbruch, 1973, 5–6).

b Structure The structure of rock material comprises its gross features: consolidated bedrock or unconsolidated sediments, nature of jointing in bedrock or of fissuring in unconsolidated sediments, or attitude of stratification planes in either case. In many places structure determines the relative importance of processes. For instance, joints and fissures favour ingress of moisture and they localize weathering and subsequent erosion, and the attitude of joints and fissures determines the direction and inclination of the resulting features.

c Mineral composition The mineral composition of rock material strongly influences its reaction to weathering and abrasion and thus to erosion. Bedrock or sediment in a warm humid climate may weather very differently from material of the identical mineral composition in a cold dry climate, and chemically different materials in the same climate may have diverse reactions. In either case the quantitative effect on a given geomorphic process may be very significant. Wind action may produce numerous ventifacts in a limestone region but comparatively few where there are only harder rocks.

d Texture The texture of rock material (i.e. the grain size and arrangement of its mineral constituents or particles) strongly influences its characteristics in any climate, whether the material is bedrock or unconsolidated. In general, fine-grained bedrock is more resistant than coarse grained to weathering in a periglacial environment, other conditions being comparable. On the other hand, fine-grained unconsolidated material may be more subject than coarse to frost action. For instance, as discussed under Freezing process in the chapter on General frost-action processes, it is common engineering practice to regard soils as susceptible to frost heaving if they contain several per cent of particles finer than 0.07 mm (passing the 200 mesh screen, *c.* the upper limit of silt) (Terzaghi, 1952, 14), or (Casagrande criterion) containing more than 3 to 10 per cent particles finer than 0.02 mm (Casagrande, 1932, 169).

Some of the many different ways in which texture can interact with periglacial processes have been discussed by Schunke (1975*a*, 214–26) with specific reference to Iceland. Frost action, in turn, can strongly influence texture by rearranging mineral particles as, for instance, in frost sorting. Another example is the change that can be effected in hydraulic conductivity of soil by freeze-thaw cycles (Benoit, 1975).

e Colour Although colour is a function of mineral composition and texture, it is worth listing separately to emphasize its effect on temperature: namely dark-coloured materials absorb radiant heat and warm their surroundings whereas the opposite is true for light-coloured materials. For instance, thawing of snow has been observed adjacent to dark-coloured objects at air temperatures as low as $-16.5°$ (G. Taylor, 1922, 47) and even $-20°$ (Souchez, 1967, 295).

4 Time

Time is an independent factor that is usually overlooked in periglacial studies. However, in the opinion of P. J. Williams (1961, 346) 'observed variations in density of occurrence of specific fossil frozen ground phenomena are not so much indicative of particular climatic conditions but of length of time during which the features could be formed'. This view is based on the belief that the processes responsible for periglacial phenomena such as solifluction and patterned ground are very slow; for instance, that it would take, say, 12 000 years for solifluction to move materials 10–20 m downslope.

On the other hand, observations in Northeast Greenland indicate that present rates of solifluction on a slope of $10°–14°$ could move materials 9–37 m downslope in 1000 years, the amount depending on the moisture conditions, which varied along the contour. These and other observations on solifluction by various investigators (cf. Washburn, 1967, 93–8, 118) and the rapidity with which some forms of patterned ground can develop (as indicated by occurrences on recently emerged shores and by direct observation), show that more information is required before quantitative inferences can be made regarding the extent to which time affects the processes. In any event, as stressed by Williams, the time factor certainly merits consideration in paleoclimatic reconstructions.

5 Human activities

Human activities are so widespread as to constitute another independent factor. Landscape changes caused by such activities are legion and can so change the natural environment that periglacial processes and features can be destroyed, modified, or enhanced over wide areas. Deforestation is an example that can lead to considerable uncertainty as to whether or not the altitude of some periglacial forms with respect to tree line is climatically determined (cf. Hagedorn, 1977, 232; Höllermann and Poser, 1977, 340). Domestication of grazing animals that tend to disrupt the vegetation is another example (cf. Höllermann, 1977, 250; Kelletat, 1977*b*, 218). The effects of human activities can be sufficiently ancient that their impress on an area is no longer immediately obvious.

III Dependent factors

1 Snow cover and ice cover

The amount and distribution of snow cover are functions of climate and topography. The effect of the latter is particularly important in the distribution of snowdrifts.

The insulating effect of snow cover on the temperature at the snow ground interface can be estimated in cgs units by applying Lachenbruch's (1959, 28; Plate 1) equation

$$\Delta\tilde{\Theta} = \frac{1}{\pi}A^*[1 - A(X)/A^*]$$

where $\Delta\tilde{\Theta}$ = change in mean annual temperature at the ground surface as the result of the snow, A^* = amplitude of mean annual temperature at bare-ground surface in summer and upper snow surface in winter, $A(X)$ = steady amplitude resulting from wave with amplitude A^* passing through snow of thickness X.

By reducing the amplitude and depth of daily and seasonal temperature changes in the ground, snow cover can reduce frost action, even if the thermal properties of the snow are no different than those of the soil, although in many places snow is a good insulator as a result of high porosity. However, the effect of snow cover on heat exchange between the air and ground is complicated. Among other considerations snow, because of decreasing thermal conductivity, tends to be a better insulator the lower the temperature (Pavlov, 1978, 71–2), especially in forested areas of strongly continental climate where loose snow and depth hoar containing large air spaces are common. In windswept tundra and alpine environments having hard-packed snow, depth hoar tends to be lacking, porosity is reduced, and the temperature dependency is much less apparent or absent (S. I. Outcalt, personal communication, 1978; cf. Goodwin and Outcalt, 1975). In areas where permafrost is discontinuous, variations in snow distribution and thickness can be the critical factor controlling its presence or absence (cf. Nicholson and Granberg, 1973). In places variability of snow cover can reverse the effect of exposure in different years (Table 2.3.).

By moderating the ground temperature, snow can protect vegetation from frost action; it can also provide the critical moisture for growth. Thus snow favours vegetation in several ways. However, if lasting too long, snow can also completely inhibit vegetation. In addition, thawing of snow, especially snowdrifts, can control the amount of moisture and thereby the nature of many periglacial processes in an area (cf. Shunke, 1975*a*, 201). Each of the above effects can be critical for a given periglacial process.

Ice is a less effective insulator than an equal thickness of snow but it will prevent a sufficiently deep river or lake from freezing to the bottom and thus inhibit the development of underlying permafrost in areas where it is otherwise present (cf. Gold and Lachenbruch, 1973, 17–18). A glacier with a bottom temperature of 0° will also inhibit development of permafrost but frozen ground would form beneath glaciers whose base was below the PT melting point as in places beneath the Antarctic Ice Sheet (Gow, Ueda, and Garfield, 1968) and Greenland Ice Sheet (B. L. Hansen and Langway, 1966).

2 Liquid moisture

The amount and nature of liquid moisture and its seasonal distribution are functions of climate, topography, and rock material. Climate exerts the large-scale control, topography modifies the large-scale climatic influence, and rock material can determine the amount of moisture entering and remaining in the ground. The texture of the rock material may be diagnostic in this respect and either favour or

Table 2.3. Snow and frost depths on north and south slopes, Coulee Experimental Forest, southwestern Wisconsin (*after Sartz, 1973, 2, Table 1*)

Date	*Snow depth (cm)*		*Frost depth (cm)*		*Remarks*
	North slope	*South slope*	*North slope*	*South slope*	*Snow depth to nearest 5 cm*
25 Feb 1970	25	0	8	11	
25 Feb 1971	55	25	0	0	Snowpack formed before ground froze
10 Mar 1972	45	25	16	23	
1 Mar 1973	5	0	21	12	Little snow throughout winter

inhibit frost action and vegetation, each of which also influences the other in complex ways. Péwé (1974, 42, Figure 3.11) has suggested that a mean annual precipitation between <5 cm and 25 cm (*c.* 15 cm) is a critical minimum for many periglacial processes and features. The nature of the liquid moisture is important because solutes can lower the freezing point appreciably. For instance, the salinity of soils in the Antarctic dry valleys depresses the freezing point sufficiently to prevent freezing and permit movement of pore fluids at depths where the temperature remains about −18° throughout the year (McGinnis, Nakao, and Clark, 1973, 140).

3 Vegetation

Vegetation, like liquid moisture, is a function of climate, topography, and rock material. Some plant species are much more sensitive to climate than others. Zonal climate sets the broad pattern but local climate can dominate a given situation, and there is an intimate interaction and mutual influence between vegetation and microclimate. For instance, the mean annual air temperature above '... unvegetated dry soils can be 1° or 2° warmer than a nearby site underlain by vegetated wet soils' (Ferrians, Kachadoorian, and Greene, 1969, 8). Also, plant remains in the form of peat pose special conditions because of the low thermal conductivity of peat compared with mineral soil. Thus frozen ground can be present in peatland but absent elsewhere (R. J. E. Brown and Williams, 1972). Destruction of a peat or vegetation cover can have far-reaching effects, especially where there is permafrost (Jerry Brown, Rickard, and Vietor, 1969; Rickard and Brown, 1974; Viereck, 1973). Near Fairbanks, Alaska, permafrost in test plots from which the vegetation had been stripped to different degrees in 1946 continued to thaw until at least 1973 without reaching equilibrium (Linell, 1973). Following destruction of vegetation by fire at Inuvik in northern Canada in 1968, thawing continued until at least 1976 (Mackay, 1977*e*).

Aside from its influence on local climate, topography is critical in that the angle of slope affects the continuing contest between the ability of plants to take root and opposing processes such as mass-wasting. Soil characteristics also exert a strong influence on the ability of plants to survive opposing processes.

Thus the especially important ways in which vegetation influences periglacial processes are through its insulating effect and its binding effect on soils. In Kryuchkov's (1976*a*, 38–41; 1976*b*, 33–6) view, the nature and extent of vegetation as primarily determined by climate are the dominant factors controlling some varieties of patterned ground. Their influence on the altitudinal zonation of periglacial features was emphasized by Karrasch (1977, 174). Details of the interaction between vegetation and frost action in soils are complicated but can be of critical importance. The problems are discussed by Balobayev (1964; 1973), Benninghoff (1952; 1966), R. J. E. Brown (1966*a*), Kryuchkov (1976*a*; 1976*b*), Raup (1965; 1969; 1971*a*), and Tyrtikov (1973; 1978) among others.

3 Frozen ground

I Introduction

Frozen ground is central to a consideration of periglacial processes and environments. In particular, knowledge of its nature sets the stage for discussing the various processes collectively known as frost action, considered in the next chapter.

In many respects it is important to distinguish between seasonally frozen ground and permafrost. The latter is especially significant in periglacial studies and is therefore discussed in some detail.

II Seasonally frozen ground

1 General

Seasonally frozen ground '... is ground frozen by low seasonal temperatures and remaining frozen only through the winter' (S. W. Muller, 1947, 221). Some investigators (cf. S. W. Muller, 1947, 6, 213) include the zone of annual freezing and thawing above permafrost (i.e. active layer), others (cf. Black, 1954, 839) restrict or appear to restrict the term to a non-permafrost environment. Because of the ambiguity and difficulty of trying to differentiate between two zones of annual freezing and thawing by a descriptive term equally applicable to both, the present writer uses the term seasonally frozen ground in the broad sense and specifies, where necessary, whether the reference is to a permafrost or non-permafrost environment. In most instances, use of the term active layer (discussed later) for the seasonally frozen ground above permafrost helps to avoid ambiguity. Ground freezing has been estimated to occur over almost half (48 per cent) of the land mass of the Northern Hemisphere, the southern limit being near lat. 40°N and characterized by frost penetration to a depth of about 30.5 cm (1 ft) once in 10 years (Bates and Bilello, 1966, 6). On this basis, the area of seasonally frozen ground as distinct from permafrost constitutes about 26 per cent of the land area, assuming that permafrost, exclusive of mountains, covers 22.35×10^6 km^2 (Table 3.1), or 22 per cent of the total land area in the Northern Hemisphere (100.30×10^6 km^2). Many soil properties and processes are radically changed as the result of ground freezing, with many practical consequences. For instance, permeability is commonly decreased and therefore run-off increased by frozen ground, with vast implications for flooding, water supply, and other hydrologic problems. However, it would be a mistake to assume, as often done, that frozen ground is necessarily impermeable; rather a wide range of conditions is possible (Dingman, 1975, 28–47).

2 Depth of seasonal freezing and thawing

Depth of freezing and thawing is controlled by many of the factors reviewed previously, and is therefore subject to considerable local variation. On a larger scale, there is a clear increase in depth of seasonal freezing with increasing latitude in a non-permafrost environment, the range being from a few millimetres to over 1.8 m (72 in) in the United States (Figure 3.1) and up to a depth of 3 m in Canada (Crawford and Johnston, 1971, 237). In a permafrost environment the trend is towards a decrease in depth with increasing latitude, and the permafrost tends to approach the surface of the ground.

The climatic parameters responsible for deep seasonally frozen ground and for permafrost may be very similar in the transition zone, but the resulting products can be quite different in a non-permafrost or a permafrost environment. Construction problems, for instance, can be many times more difficult in a permafrost environment than where there is seasonally frozen ground without permafrost.

It may not always be practicable to obtain direct measurements of the depths to which seasonal freezing and thawing extend, and several indirect approaches have been suggested. Some of these involve a number of variables, and others are relatively simple yet useful.

For example, the depth of winter freezing can be estimated by the Stefan equation, which in English units has usually the form (Yong and Warkentin, 1975, 391–7)

$$x = \sqrt{\frac{48k_f F}{L}} \qquad (1)$$

where x = depth of freezing (ft), k_f = thermal conductivity of frozen soil (0.9 (BTU h^{-1} ft^{-1} °F^{-1}), F = freezing index (number of F degree days below freezing), and L = latent heat of water in the soil = $L_w w \gamma_d$ (where L_w = latent heat of water [143.4 BTU lb^{-1} water], w = amount of water in soil [in decimal fraction of dry wt. of soil], and γ_d = unit weight of dry soil [in lb ft^{-3}]).

The same equation can be used in metric units by appropriate substitutions, with x = cm, k_f = cal h^{-1} cm^{-1} °C^{-1}, F = C degree days below freezing, and $L_w w \gamma_d$ is of the form where L_w = 80 Cal g^{-1}, ω is decimal fraction of dry wt. of soil, and γ_d is in g cm^{-3}.

Or, following Jumikis (1977, 206–7)

$$\xi = \sqrt{\frac{48K \cdot F}{Q_L}} \qquad (2)$$

where ξ = depth of freezing (m), K = coefficient of thermal conductivity of frozen soil (Cal m^{-1} h^{-1} °C^{-1}), F = surface freezing index (C degree days), Q_L volumetric latent heat of water (Cal m^{-3}).

If the index F is in C degree hours, (2) assumes the form

$$\xi = \sqrt{\frac{2KF}{Q_L}} \qquad (3)$$

By inserting appropriate figures for the thawing period (thermal conductivity of thawed soil, thawing index), the equations can also be used to estimate depth of summer thawing. They apply to both permafrost and non-permafrost environments. A number of refinements and other expressions are available whose complexity increases with the accuracy required (cf. Andersland and Anderson, 1978, 131–49).

The equations cited have several potential sources of error. A serious problem is that the soil properties may be difficult to estimate without drilling and laboratory testing. Also in the absence of ground-surface temperatures, error is introduced by substituting air temperatures without allowing for the effect of snow or vegetation cover and the influence of colour on radiant heating. Lachenbruch's (1959, 28, Plate 1) equation, cited under discussion of Dependent factors in the chapter on Environmental factors, can be applied to evaluate the effect of snow cover. In addition to the mean annual air temperature, this equation also considers the amplitude of the temperature change from summer to winter, which can be a critical variable in evaluating the effect of a disturbance of the ground surface on depth of thawing. 'The larger the amplitude, the worse the thawing problem' (Lachenbruch, 1970*b*, J5). However, the thawing problem is minimal where summer temperatures are low and are above freezing for only a brief period as in the Antarctic dry valleys. Sophisticated computer programs and modeling are being increasingly widely used to evaluate the response of the ground to temperature perturbations (cf. Kliewer, 1973).

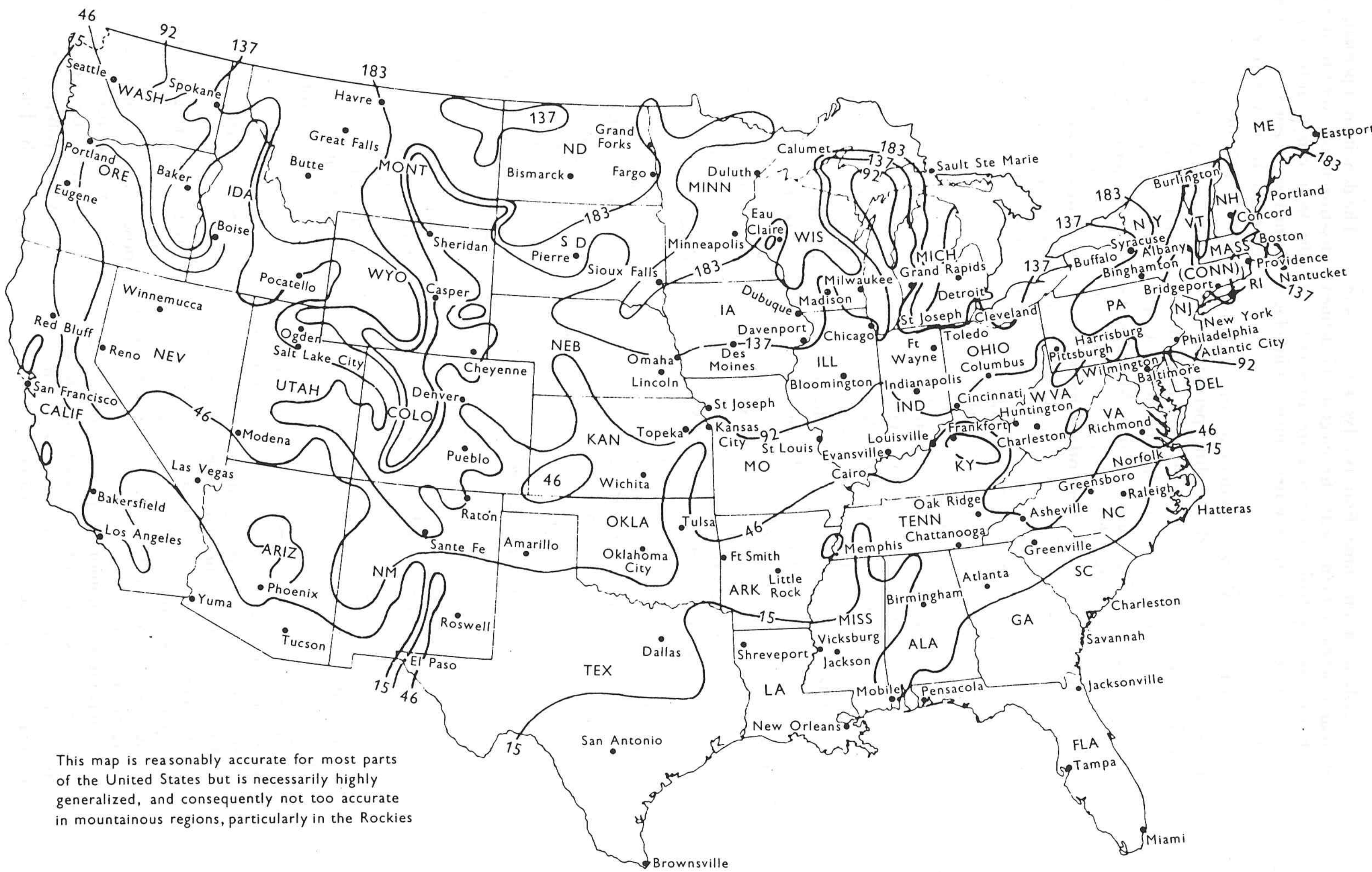
III Permafrost

1 General

The term permafrost, also known as pergelisol (Bryan, 1946, 635, 640) and perennially frozen ground, was first defined by S. W. Muller (1947, 3, cf. 219)

> Permanently frozen ground or permafrost is defined as a thickness of soil or other superficial deposit, or even of bedrock, at a variable depth beneath the surface of the earth in which a temperature below freezing has existed

MAXIMUM DEPTH OF FROST PENETRATION (CM; CONVERTED FROM INCHES)



3.1 Maximum depth of frost penetration in United States (*after Strock and Koral, 1959, 1–102*) continually for a long time (from two [years] to tens of thousands of years). Permanently frozen ground is defined exclusively on the basis of temperature, irrespective of texture, degree of induration, water content, or lithologic character.

The Institut Merzlotovedeniya in Yakutsk, USSR, one of the world's leading permafrost research agencies, specifies three years or more in accordance with the recommendations of the Commission on Terminology of the Institut Merzlotovedeniya im. V. A. Obrucheva (1956, 10; 1960, 6; Corte, 1969*a*, 130). On the other hand, some Canadian workers adopt a minimum period of one year only – i.e. ground remaining frozen throughout at least one summer (R. J. E. Brown, 1967*a*; French, 1976*a*, 47). The present writer follows Muller's original definition, specifying at least two years.

The term permafrost originated from what Muller described as a slip of the tongue that he was encouraged to formalize (S. W. Muller, personal communication, 1959). The term has been widely adopted but defined as perennially frozen rather than permanently frozen ground, since changes in climate and surface conditions can cause rapid thawing of permafrost. Muller's definition in which texture, induration, water content, and lithology are eliminated as factors (cf. also S. W. Muller, 1947, 30) supports the view that 0° is the basic criterion rather than the exact freezing point, which can vary with such factors including salts and pressure. Most workers cite a temperature below 0° in defining permafrost (cf. R. J. E. Brown, 1967*a*; R. J. E. Brown and Kupsch, 1974, 25; Ferrians, 1965), but in practice it is doubtful if ground that rises to 0° for part of the year would be excluded if the temperature rose no higher and its mean annual temperature remained below 0°. In fact a temperature as high as 0° is specifically accepted by the Bol'shaya Sovetskaya Entsiklopediya (Sharbatyan, 1974*a*, 370; 1975, 1) and by French (1976*a*, 49). Amended definitions of permafrost have been suggested. According to Stearns (1966, 1–2)

> The term 'permafrost' is defined as a condition existing below ground surface, irrespective of texture, water content, or geological character, in which:
> *a*. The temperature in the material has remained below 0°C continuously for more than 2 years, and
> *b*. If pore water is present in the material a sufficiently high percentage is frozen to cement the mineral and organic particles.
>
> A temperature definition alone is not considered sufficient, for often a geothermal situation exists in which a frozen, or cemented, state is not obtained even though the temperature of the material is well below 0 C. . . .
>
> The definition of permafrost given above includes both dry-frozen and wet-frozen ground. In the dry-frozen, or dry frost, condition there is very little or no water contained in the pores so that temperature becomes the only criterion. In the wet-frozen condition some cementing ice must be present.

It remains to be seen to what extent Stearns' redefinition is accepted. It is followed by Carey (1973, 8), and from the viewpoint of engineering requirements it has the merit of specifying cementation where there is wet-frozen ground. However, the last two paragraphs pinpoint a problem in that temperature alone is not considered sufficient for wet-frozen ground but is the only criterion for dry-frozen ground. Furthermore cementation is a subjective criterion unless physical parameters defining cementation are established. Muller's definition is more precise and for mapping purposes, at least, the easier to apply, and is followed here.

In an attempt to eliminate ambiguities, van Everdingen (1976) proposed that 'unfrozen' be replaced by nonfrozen, that 'frozen' be used only if ice is present, and that 'cryotic' be adopted to indicate a temperature $<0°$ without implication as to phases in a soil/water system as indicated in Table 3.2. As with Stearns' redefinition of permafrost, van Everdingen's proposal must stand the test of time. The terminological problems addressed are real, and regardless of the terms used it behoves authors to be very specific if there is any possibility of misunderstanding.

Glaciers whose temperature does not reach 0° are permafrost by either definition but they are usually omitted from discussions of permafrost and treated separately (Ferrians and Hobson, 1973, 486). They will be mentioned here only incidentally.

Permafrost in the Soviet Union was mentioned as early as 1642 in Siberian military reports by Glebov and Golovin (Tsytovich, 1966), and was recognized in northern Canada by at least 100 years later (Middleton, 1743, 159; Rich and Johnson, 1949, 67, 71; cf. Legget, 1966, 3).

2 Significance

Civil engineering problems connected with building and highway construction, sewage and waste disposal, water supply, and hot-oil and cold-gas pipelines are magnified in permafrost regions, and much attention has been devoted to permafrost as a result (cf. Alter, 1969*a*; 1969*b*; Anisimova *et al.*, 1973*a*; 1973*b*; 1978; Jerry Brown, Rickard, and Victor,

3.1 Maximum depth of frost penetration in United States (*after Strock and Koral, 1959, 1–102*)

1969; R. J. E. Brown, 1970; Burdick and Johnson, 1977; Carlson, 1977; Cheng, 1975; Cohen, 1973; Corte, 1969*a*; Crawford and Johnston, 1971; Dingman, 1975; Environment Canada, 1974; Ferrians, Kachadoorian, and Greene, 1969; Finn, Yong, and Lee (1978); Gold *et al.*, 1972; Kachadoorian and Ferrians, 1973; Kudryavtsev, 1974; 1978; Kudryavtsev *et al.*, 1977; Lachenbruch, 1970*a*; Linell and Johnston, 1973; Melnikov and Tolstikhin, 1979; Sanger, 1969; Tsytovich, 1975; Vyalov *et al.*, 1962; 1965; 1973*a*; 1973*b*; 1978; Vyalov, Dokuchayev, and Sheynkman, 1976; J. R. Williams, 1970; and J. R. Williams and van Everdingen, 1973. A number of the references noted in the section on References in the first chapter are also highly pertinent.)

The problems are both civilian and military. The military interest that began with World War II led to the establishment of the US Army Snow, Ice, and Permafrost Research Establishment (SIPRE), later renamed the US Army Cold Regions Research and Engineering Laboratory (CRREL), which has been responsible for much of the basic research on permafrost in the United States. The US Geological Survey has also been very active in permafrost research. Yet our knowledge is still inadequate as illustrated by the controversy aroused by the trans-Alaska pipeline system before it was approved – and by the permafrost-related research that has been advocated as a result of the north-south transportation route provided by the system (National Academy of Sciences, 1975). Increasingly, permafrost is becoming recognized as an important factor in Arctic land-use planning and its environmental impact (Andrews, 1978; Jerry Brown, 1973; Jerry Brown and Grave, 1979*a*; 1979*b*; Grave and Sukhodrovskiy, 1978; McVee, 1973; Melnikov, 1977). Not only are contemporary frost action and permafrost of engineering concern but fossil features resulting from past periglacial activity can also be important (Higginbottom and Fookes, 1971).

Aside from the engineering aspects, there are many intriguing scientific questions related to permafrost – its origin, climatic implications, its effect on life (Péwé, 1966*c*), and what it can tell us about past, present, and future environments. Much of the record is there to be read, and part of the task of periglacial research is to interpret that record.

3 Distribution

Because permafrost becomes thinner and breaks up into patches as it merges with seasonally frozen ground where permafrost is lacking, permafrost is commonly classified into continuous permafrost and

Table 3.1. Distribution of permafrost (*after Stearns, 1966, 9–10*)[1]

	Continuous		*Discontinuous*		*Total*	
	km² × 10⁶	mi² × 10⁶	km² × 10⁶	mi² × 10⁶	km² × 10⁶	mi² × 10⁶
Northern Hemisphere	7.64	2.95	14.71	5.68	22.35	8.63
Antarctica	13.21	5.10	—	—	13.21	5.10
Mountains	—	—	2.59[2]	1.00	2.59	1.00
Totals	20.85	8.05	17.30	6.68	38.15	14.73

	Land surface (per cent)	*Permafrost* km² × 10⁶	*Permafrost* mi² × 10⁶
Alaska	80	1.30	0.5
Canada	40–50	3.89–4.92	1.5–1.9
Greenland	99	1.68	0.65
USSR[3]	50	11.14	4.3
		19.04 (max)	7.35 (max)

[1] Few (if any) of the estimates include offshore permafrost.
[2] Regarded as too low a figure by Barsch (1977*d*, 124).
[3] USSR estimates cited in 1976 are similar (Melnikov and Balobayev, 1977). However, the USSR permafrost area having the bulk of visible ice (*c.* 19 000 km³ at depths <50 m) has been estimated by Vtyurin (1973, 17, Table 1; 1978, 164, Table 1) to be only about 7 000 000 km², mainly in Siberia and the Soviet Northeast as shown in three informative sketch maps by Vtyurin.

Table 3.2. Terminology describing ground temperature and water phases in a soil/water system (*after van Everdingen, 1976, 864, Figure 2*)

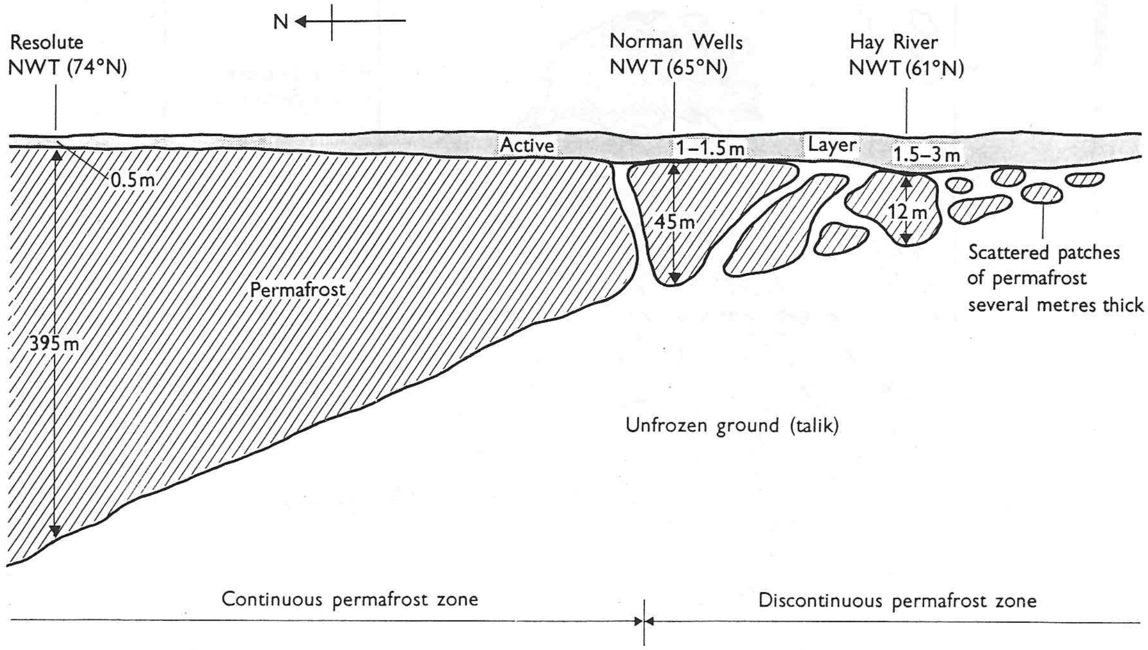
H_2O content→ / Temperature ↓	No H_2O (except chemically bound and adsorbed)		Some H_2O (less than porosity)		Pore spaces filled with H_2O		Containing 'excess' H_2O		Zone descriptions: Phase	Zone descriptions: Temp.
T > 0°C	Dry, noncryotic		Moist or unsaturated, noncryotic		Wet or saturated, noncryotic		---		Nonfrozen	Noncryotic
— 0°C — Cryo point										
T < 0°C	Dry, cryotic	Dry permafrost (1)	Moist or unsaturated, cryotic	Moist permafrost (1)	Wet or saturated, cryotic	Wet permafrost (1)	---	--	Nonfrozen	Cryotic
— Initial freezing point of soil system —										
T < 0°C	Dry, cryotic	Dry permafrost (1)	Ice-poor, partially frozen	Ice-poor permafrost (1)	Partially frozen	Permafrost (1)	Ice-rich, partially frozen	Ice-rich permafrost (1)	Partially frozen	Cryotic
T < 0°C	Dry, cryotic	Dry permafrost (1)	Ice-poor, frozen	Ice-poor permafrost (1)	Frozen	Permafrost (1)	Ice-rich, frozen	Ice-rich permafrost (1)	Frozen	Cryotic

(1) If temperature is perennially below 0°

discontinuous permafrost (Figure 3.2). Some investigators have followed Black (1950, 248–9, Figure 1) and S. W. Muller (1947, 6, after Sumgin) in also recognizing a sporadic permafrost zone, which appears to be convenient for alpine environments (Barsch, 1977*d*; 1978; J. D. Ives, 1974, 166) but perhaps less so elsewhere (R. J. E. Brown, 1967*a*).

The distribution of present-day permafrost in the Northern Hemisphere, disregarding high-altitude occurrences, is illustrated in Figures 3.3–3.11. This distribution correlates, at least in part, with the paths of anticyclonic polar air masses and hence cold, dry winters. A close correlation between the southern extent of the continuous permafrost zone and the southern (winter) position of the Arctic frontal zone is suggested by a map of Bryson's (1966, 266, Figure 33). In general the permafrost realm is characterized by days with temperatures $<0°$ for three fourths of the year, $<-10°$ for half the year, and rarely $>20°$; precipitation is characteristically <100 mm in winter and <300 mm in summer (Velichko, 1975, 100). However, significant departures can occur. For instance, at Mesters Vig, Northeast Greenland, the mean precipitation for the 9 months (Sept.–May) having mean temperatures $<0°$ is about 300 mm rather than <100 mm (Washburn, 1965, 52, Table 2). Stearns (1966, 9–10), following Black (1954, 839–42), concluded that permafrost underlies some 26 per cent of the land surface (including glaciers) of the world, based on figures compiled from various sources (Table 3.1).

Offshore (also known as submarine or subsea) permafrost, reported in Spitsbergen over 50 years ago (Werenskiold, 1922; cf. 1953), is common off some Arctic coasts (Figures 3.3, 3.12–3.13) (F. E. Are, 1976; 1978*a*; 1978*b*; Barnes and Hopkins, 1978, 117–22; Grigor'yev, 1966, 92–119; 1973; 1976, 95–126; 1978; Hobson *et al.*, 1977; Hunter *et al.*, 1976; Judge, 1974; 1975; 1977; Lachenbruch and Marshall, 1977, 1–6; Lewellen, 1974; Mackay, 1972*a*, 19–20; 1972*d*; Molochushkin, 1973; 1978; National Academy of Sciences, 1976; Osterkamp and Harrison, 1976; 1977; Rogers, 1976; 1977; Rogers and Morack, 1977*a*; 1977*b*; 1978; Sellmann *et al.*, 1976; Zhigarev and Plakht, 1974; 1976). It very probably also occurs in the Antarctic. Water temperatures in McMurdo Sound are about $-1.8°$, and on the basis of observed

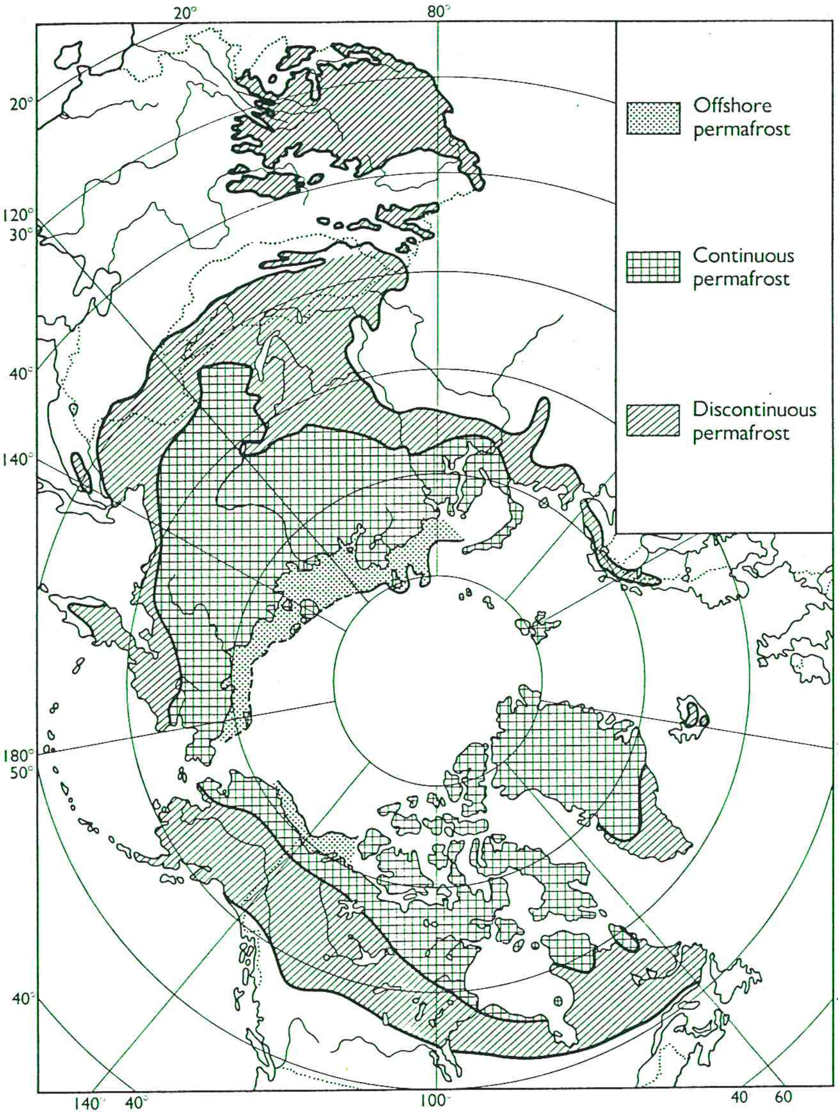


3.2 North-south vertical profile of permafrost in Canada showing decreasing thickness southward and relation to continuous and discontinuous permafrost zones (*after R. J. E. Brown, 1970, 8, Figure 4*)

low pore-water salinity in sediments off New Harbor and compressional velocities of sediments in Terra Nova Bay, the likelihood of ice-cemented permafrost to depths of at least 150 m below sea level in Terra Nova Bay is considered high by L. D. McGinnis (personal communication, 1978). There is now intensive research on Arctic offshore permafrost because of the difficulties it poses for offshore oil development (cf. Hunter *et al.*, 1976). Much of the offshore permafrost probably originated on land and became submerged by postglacial rise of sealevel and coastal erosion. Fossil permafrost beneath water depths of 70–80 m in the Laptev and East Siberian Seas has been reported but without details (Markuse, 1976, 120). Some ice-cemented, offshore permafrost is probably in equilibrium with sea-bottom temperatures (*c.* $-1.3°$ near shore in the Beaufort Sea – Lewellen, 1974, 420, 426; *c.* $-1.5°$ to $-1.8°$ near Little Cornwallis Island in the Queen Elizabeth Islands – Judge, 1974, 432). Ice-bearing offshore permafrost may be widely distributed on the Beaufort Sea continental shelf (Barnes and Hopkins, 1978, 117–22). A comprehensive bibliography of offshore permafrost is being compiled by Vigdorchik (1977*a*; 1977*b*).

The difference in Table 3.1 in the figure for the Northern Hemisphere (22.35×10^6 km^2) and the total for Alaska, Canada, Greenland, and the USSR (19.04×10^6 km^2) is due to the different methods of estimating the figures. If cementation is accepted as a criterion of permafrost, and glaciers are excluded, a major reduction in the estimate for Antarctica and Greenland will apply if it turns out that the ice sheets here are extensively underlain by unfrozen material.[1] Excluding glaciers, Shumskii, Krenke, and Zotikov (1964), cited by Grave (1968*a*, 48; 1968*b*, 2), estimated that 14.1 per cent of the land

[1] Grave (1968*a*, 51–2; 1968*b*, 6) and Zotikov (1963) argued that the central part of the Antarctic is free of frozen ground, and drilling confirms the presence of water at the base of the Ice Sheet near Byrd Station (Gow, Ueda, and Garfield, 1968). However, the estimated PT melting point of $-1.6°$ at the drilling site would still be indicative of permafrost according to Muller's definition. Assuming that the Antarctic Ice Sheet is in a steady state, Budd, Jenssen, and Radok (1970, 301–5; 1971, 117–26) calculated, contrary to Grave and Zotikov, that the central area would be frozen and only relatively small marginal areas would be thawed beneath the ice; their assumption is subject to question but Budd, Jenssen, and Radok concluded that the central area would be characterized by permafrost even if ice gain exceeded ice loss by a factor of two. According to similar calculations, most of the Greenland Ice Sheet has a basal temperature below $-5°$ (W. Budd, personal communication, 1971).

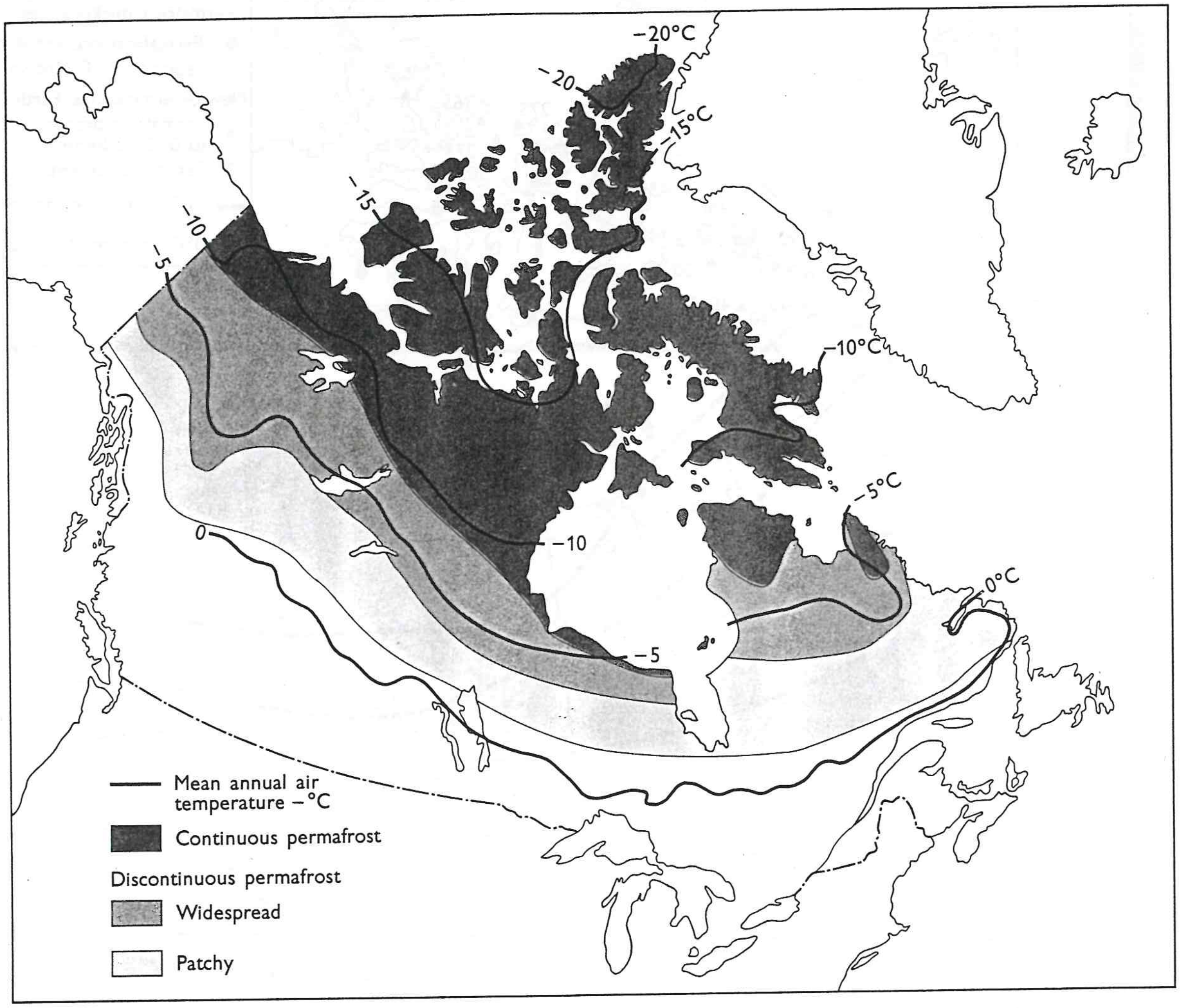


Sources: Alaska, land (T. L. Péwé, personal communication, 1978), offshore (Barnes and Hopkins, 1978); Canada, land (R. J. E. Brown, 1978*b*), offshore (Hunter *et al.*, 1976); Greenland (Weidick, 1968; O. Oleson, personal communication, 1976); Iceland (Thorleifur Einarsson, personal communication, 1966; Priesnitz and Schunke, 1978); Norway (B. J. Andersen, personal communication, 1966; H. Svensson, personal communication, 1966); Svalbard (Liestøl, 1977); Sweden (Rapp and Annersten, 1969); Mongolia (Gravis *et al.*, 1973, 1978); USSR, land (Karpov and Puzanov, 1970); offshore (M. Vigdorchik, personal communication, 1978); China (Academia Sinica, 1975, 20–1). The offshore permafrost extent shown for North America is very probably a minimum, judging from near-shore water temperatures among Canada's arctic islands (J. R. Mackay, personal communication, 1978)

3.3 Permafrost in the Northern Hemisphere (*T. L. Péwé, personal communication, 1978 with sources as indicated*)

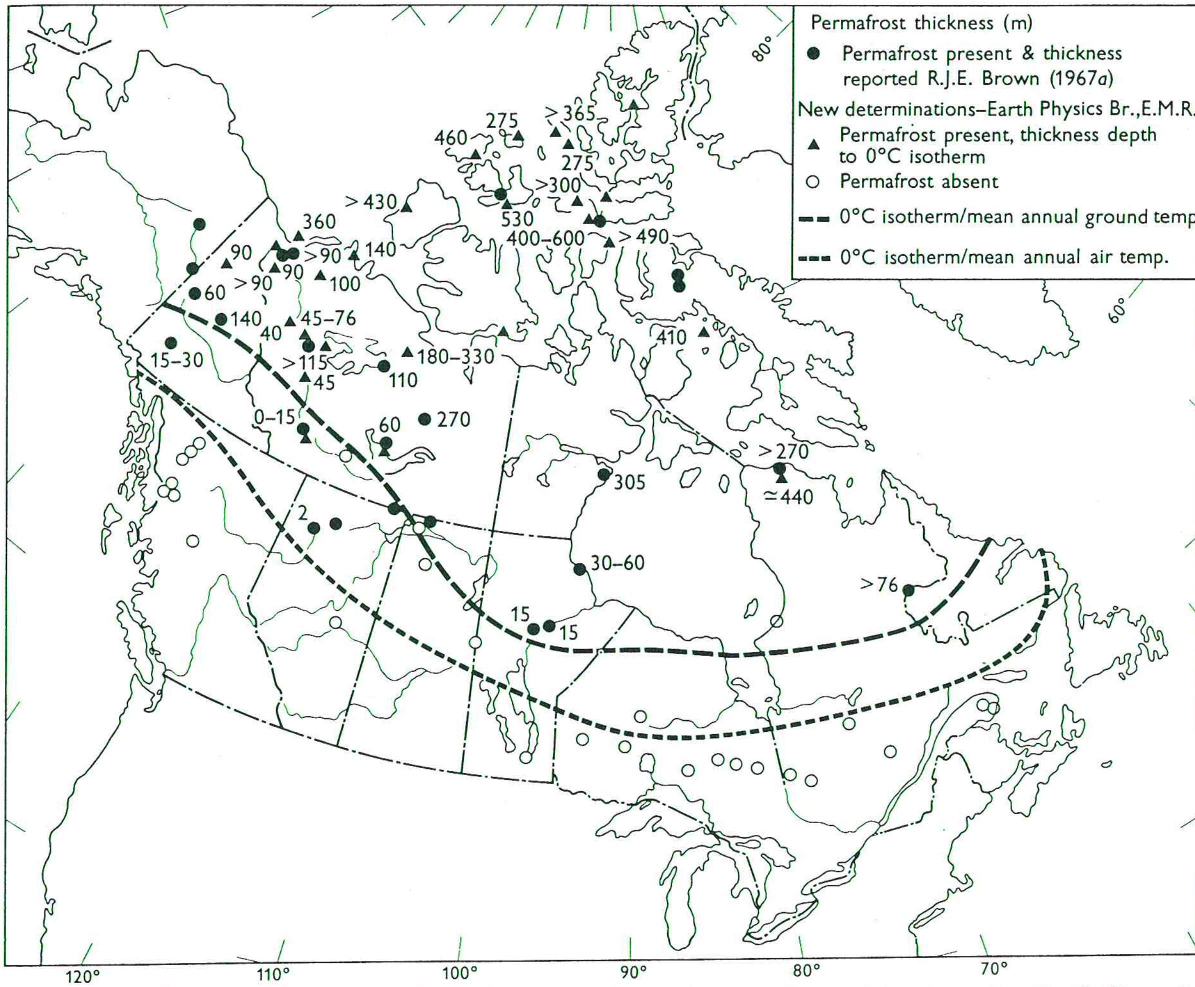
surface is underlain by permafrost. R. J. E. Brown (1970, 7; cf. 27–8) and Péwé (1971; personal communication, 1972) estimated 20 per cent (excluding glaciers), and Péwé in agreement with Stearns (Table 3.1) indicated that permafrost characterizes 80 per cent of Alaska and 50 per cent of Canada. Ferrians and Hobson (1973, 479) agree except for raising the estimate for Alaska to 85 per cent. Jerry Brown (1967, 18, Table 4) estimated that the coastal plain of Arctic Alaska to a depth of 300 m contains 1675 km^3 of ground ice, including pore ice (cf. Péwé, 1975, 49). The volume of ground ice in the USSR (probably excluding pore ice) is calculated to be 1900 km^3 (Vtyurin, 1975; cited by Grave, 1977, 4). These estimates are based on numerous assumptions. According to Grave's earlier report, the world's permafrost comprises 0.83 per cent of the total freshwater ice. Most of the remainder is in the Antarctic Ice Sheet (>90 per cent), the Greenland Ice Sheet, and other glaciers. Together this ice constitutes some 75 per cent of the freshwater resources of the earth. The amount represented by permafrost is very small but its significance far transcends its quantity.

In northern latitudes there is commonly a difference of 1°–6° between the mean annual temperature



3.4 Permafrost in Canada (*after R. J. E. Brown, 1978b*)

of the air and that of the ground surface (Gold and Lachenbruch, 1973, 5). The latter is commonly warmer, and the difference is generally greatest where there is a thick winter snow cover (G. P. Williams and Gold, 1976, 180–3, Figure 3). The mean northward displacement of the ground-surface isotherm relative to the equivalent air temperature in northern Canada is reported to be 3.3° (Judge, 1973*a*, 38) (Figure 3.4). Comparison with climatic maps shows that the southern limit of continuous permafrost is not necessarily parallel to air isotherms. It lies north of the −6° or −8° mean annual air isotherm in Alaska (R. J. E. Brown and Péwé, 1973, 74, Figure 2; Péwé, 1966*a*, 78–9; 1966*b*, 68) (Figure 3.7). In Canada it is commonly north of −8.5° (R. J. E. Brown and Péwé, 1973, 75), although in places discontinuous permafrost is still present at −9.4° (15°F) (R. J. E. Brown, 1969, 52). According to R. J. E. Brown (1967*a*) and Crawford and Johnston (1971, 237), the discontinuous and continuous zones merge where the mean annual air temperature is about −8.3°. This accords with arbitrarily mapping the boundary where the mean annual air temperature corresponds to a mean annual ground temperature of −5° at the depth of zero annual amplitude as discussed below. Probably about −7° approximates the southern boundary of continuous permafrost in the USSR (Baranov, 1959, Figure 24 opposite 201; 1964, 83, Figure 24; Akademiya Nauk SSSR, 1964, 234), but permafrost is absent in the Turukhansk region of USSR at −7° (Shvetsov, 1959, 79; 1964, 5–6; who



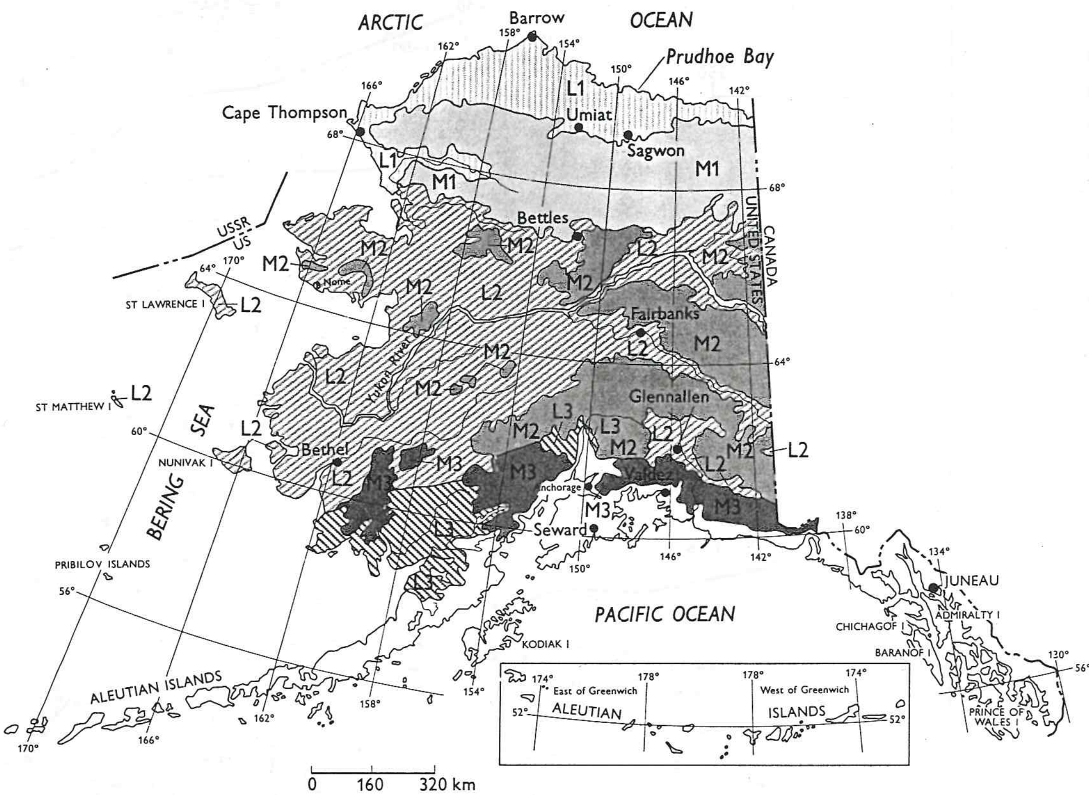
3.5 Permafrost thickness and mean annual air and ground-surface isotherms in Canada (*after Judge, 1973a, 38, Figure 3*)

cited Yachevskii (1889)[2] or −7.4° (Schostakowitsch, 1927, 396). R. J. E. Brown (1967*b*) has pointed out in an excellent review that there are significant differences as well as many similarities in environmental factors and the distribution of permafrost in North America and the USSR.

The distribution of permafrost can also be evaluated with respect to the temperature at the depth of zero annual amplitude – i.e. the depth to which seasonal changes of temperature extend into permafrost – which closely approximates the mean annual ground temperature at the surface (National Academy of Sciences, 1974, 40). Black (1954, 843) noted that permafrost at this depth is generally below −5° in the continuous zone, and according to R. J. E. Brown (1967*a*; 1970, 8) the criterion of −5° at the depth of zero annual amplitude has been adopted in both North America and the USSR as an arbitrary boundary between continuous and discontinuous permafrost. The difficulties of accurately mapping permafrost are discussed by J. D. Ives (1974, 166–71) and Vostokova (1973; 1978).

Permafrost is in a particularly delicate ecological balance in the Subarctic whose southern limit can also be considered as the southern limit of discontinuous permafrost (R. J. E. Brown, 1966*b*). Possibly much of the discontinuous zone is in disequilibrium with the present climate but detailed observations are needed. According to a number of investigators the southern limit of discontinuous

[2] Yachevskii (1889, 346) gave −8° (*c.*) for Turukhansk, which lies in an area of insular permafrost (Akademiya Nauk SSSR, 1964, Plate 234). Kendrew (1941, 212), referring to lack of permafrost at Turukhansk, gave a temperature of 17°F (i.e. also about −8°C).



AREAS WITHIN PERMAFROST REGION

Mountainous areas, generally underlain by bedrock at or near the surface

M1 Underlain by continuous permafrost

M2 Underlain by discontinuous permafrost

Underlain by isolated masses of permafrost

Lowland areas, generally underlain by thick unconsolidated deposits

L1 Underlain by thick permafrost in areas of either fine-grained or coarse-grained deposits

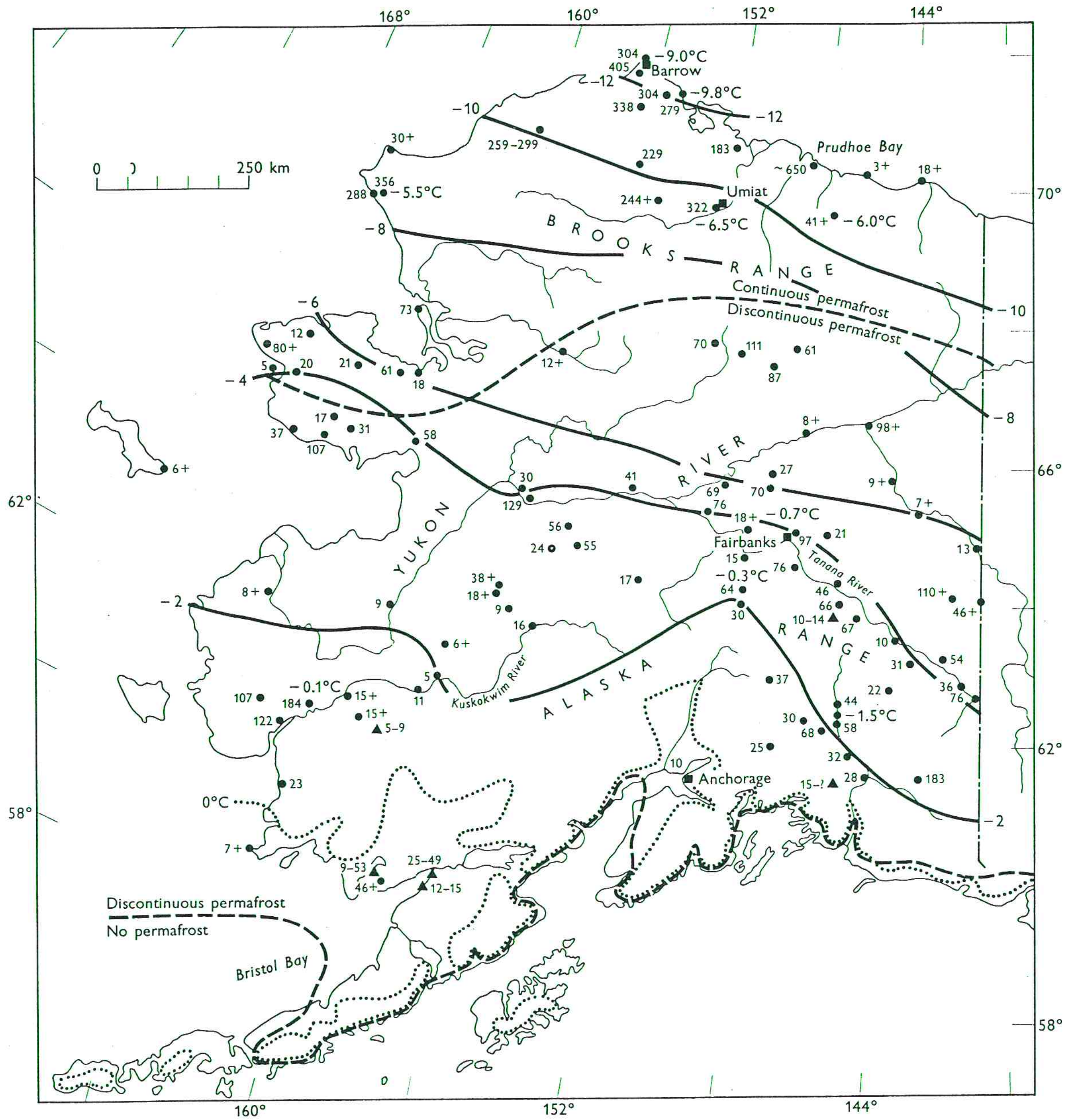
L2 Underlain by moderately thick to thin permafrost in areas of fine-grained deposits, and by discontinuous or isolated masses of permafrost in areas of coarse-grained deposits

L3 Underlain by isolated masses of permafrost in areas of fine-grained deposits, and generally free of permafrost in areas of coarse-grained deposits

AREAS OUTSIDE PERMAFROST REGION

Generally free of permafrost, but a few small isolated masses of permafrost occur at high altitudes, and in lowland areas where ground insulation is high and ground insolation is low, especially near the border of the permafrost region

3.6 Permafrost in Alaska (*after Ferrians, Kachadoorian, and Greene, 1969, 3, Figure 2, cf. Ferrians, 1965*)



PERMAFROST, data in metres

122 ● Thickness, depth to base

18+ ● Minimum thickness, base unknown

12–15 ▲ Relic permafrost, ground frozen between depths shown

−6.5°C ● Temperature of permafrost at 15 to 25 metres depth

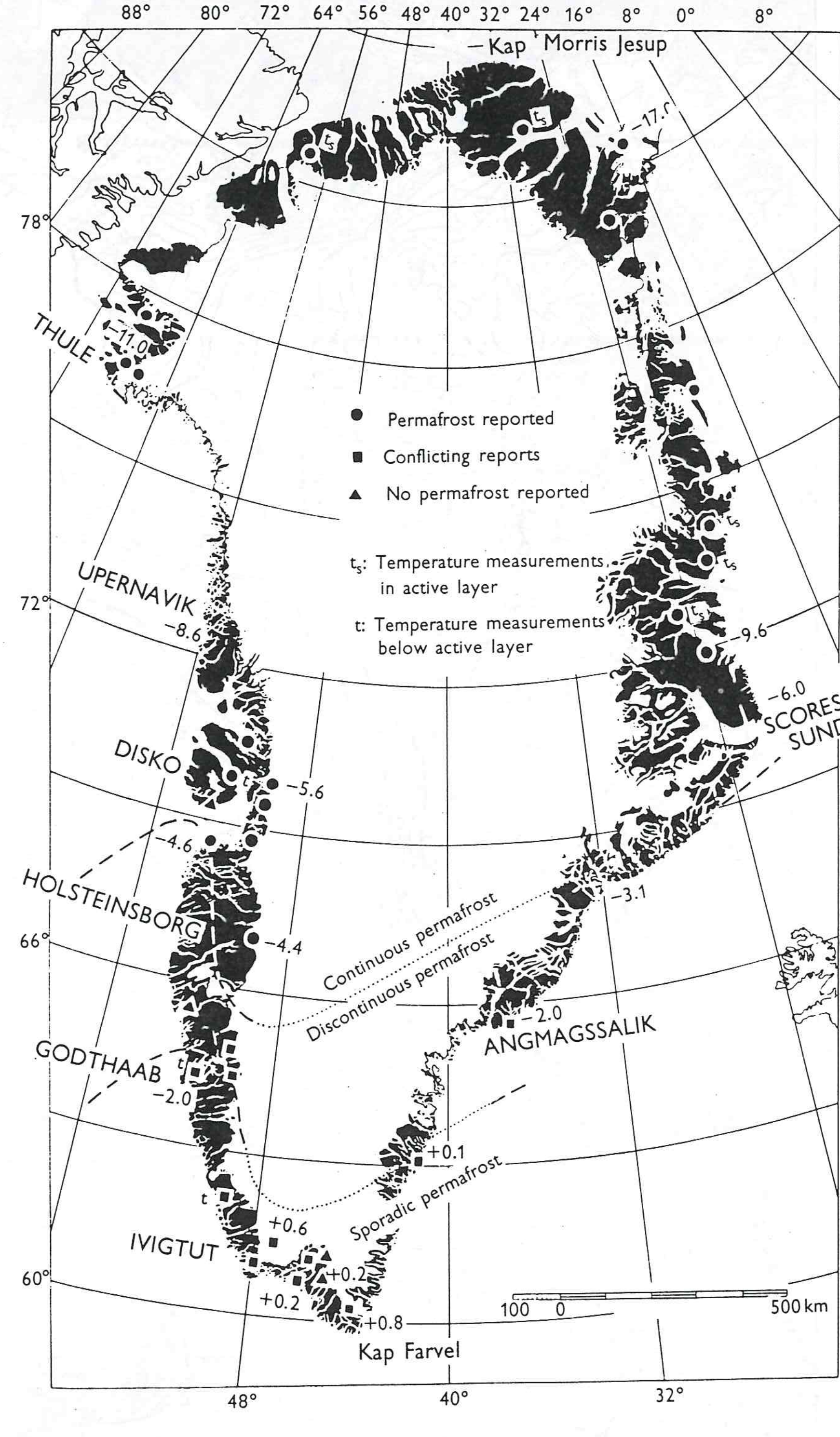
— — — — Permafrost zone boundary

CLIMATE

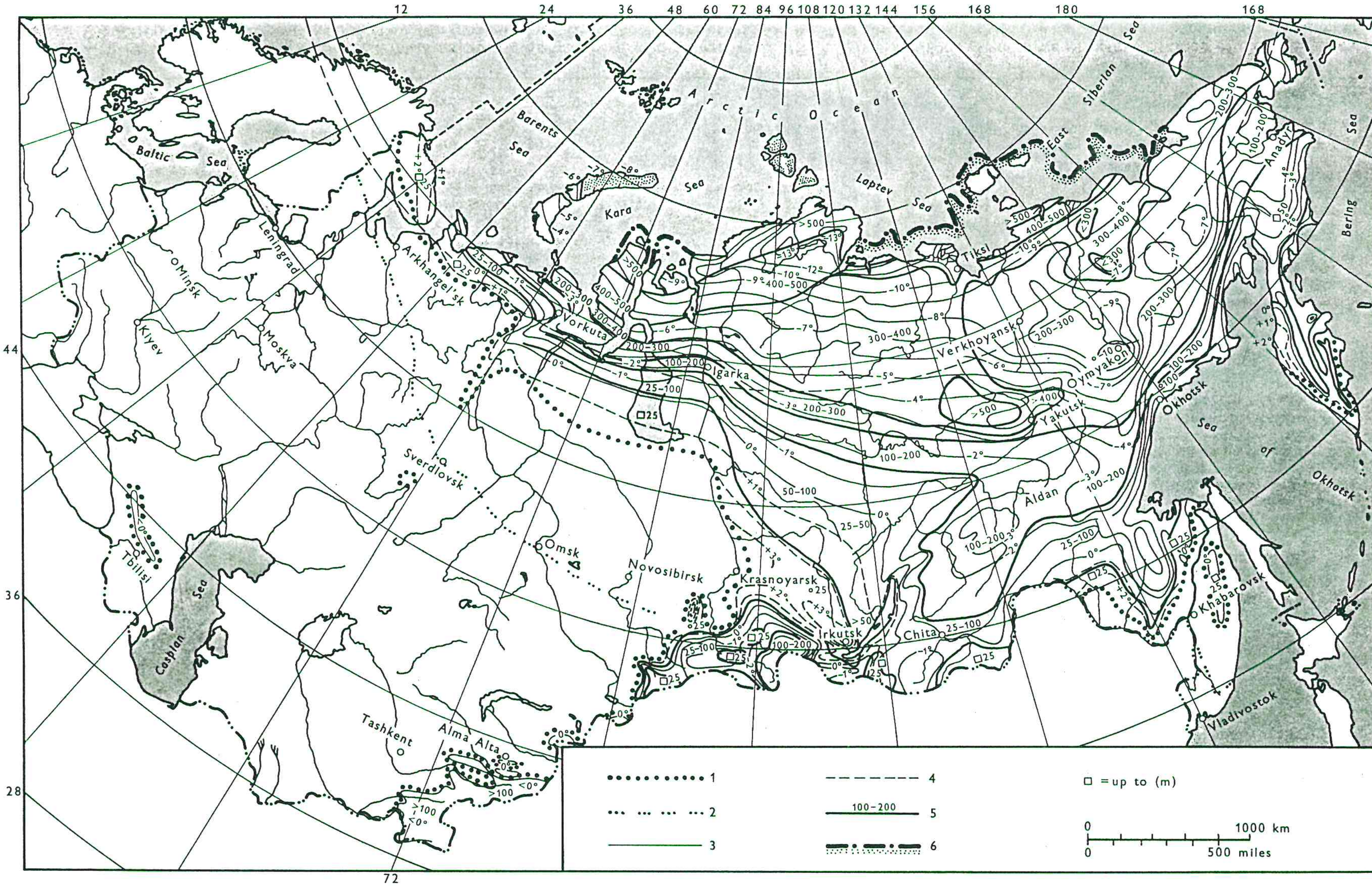
·········· Approximate position of mean annual air temperature isotherm, 0°C

———— Mean annual air temperature, °C

3.7 Permafrost thickness and temperature in relation to mean annual air-temperature isotherms in Alaska (*after R. J. E. Brown and Péwé, 1973, 74, Figure 2*)

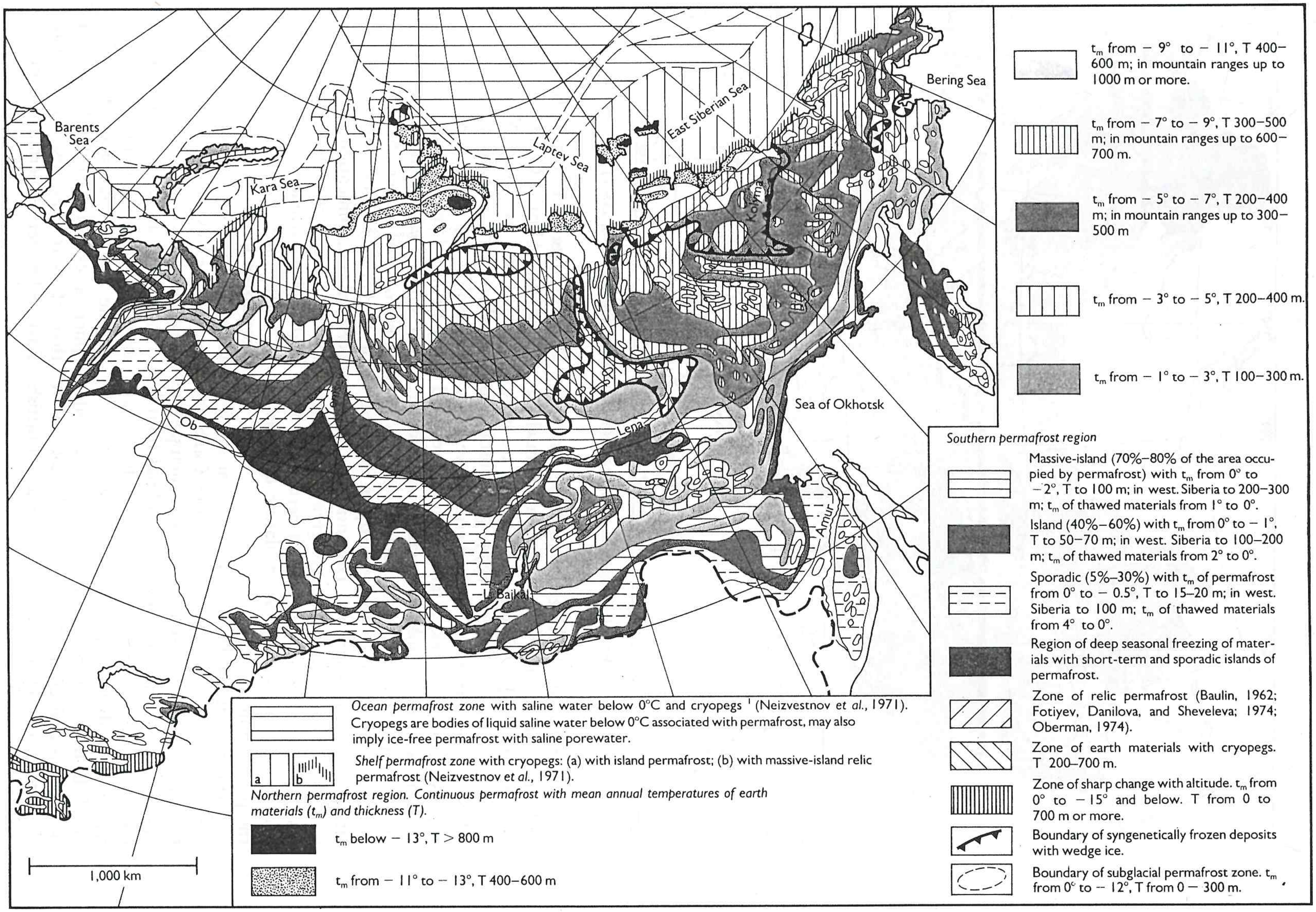


3.8 Permafrost in Greenland in relation to mean annual air temperatures, °C (*after Weidick, 1968, 73, Figure 25*)

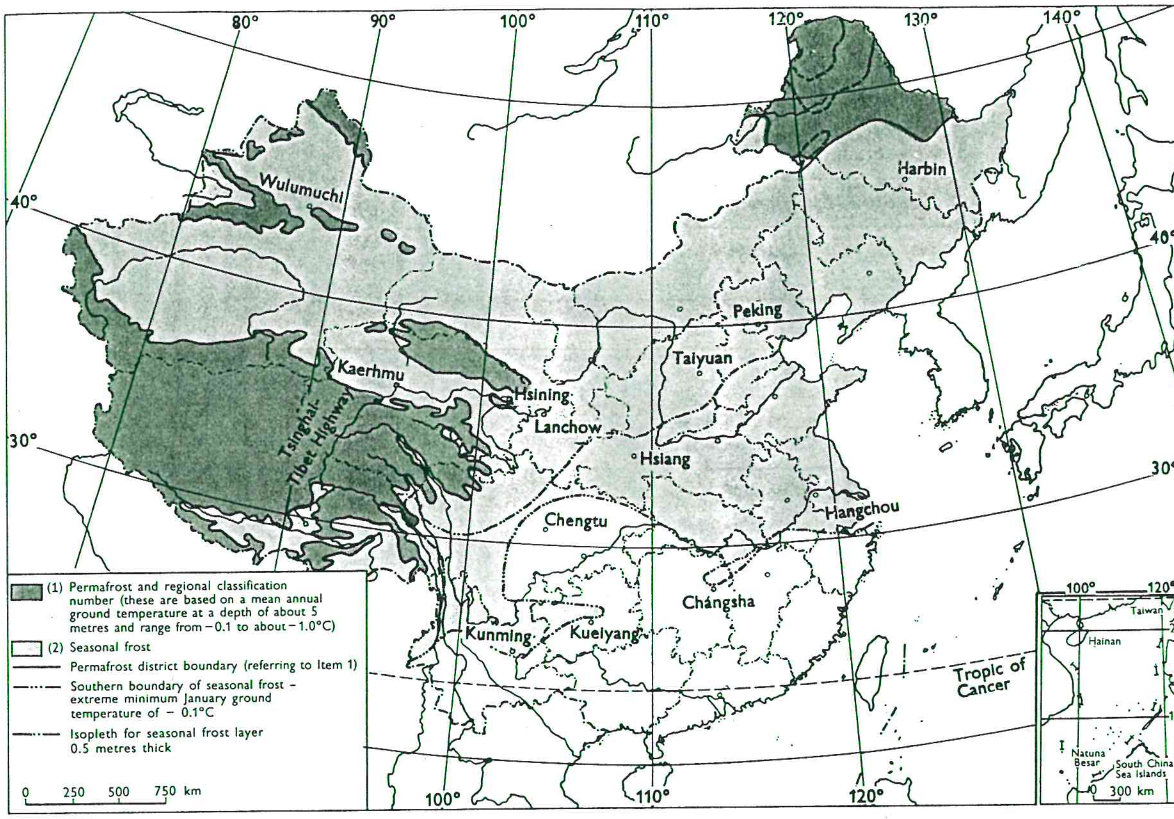


1 Boundary of permafrost area 2 Boundary of zone of frequent pereletoks 3 Minimum ground temperature at level of zero annual amplitude (in mountainous regions, shown for valleys) 4 Soil isotherms at 1–2 m depth under natural conditions 5 Maximum thickness of permafrost (m) 6 Permafrost zone under Arctic Ocean. *Note* Temperatures in °C.

3.9 Permafrost in the Soviet Union (*after Baranov, 1959, Figure 24, opposite 201; cf. Baranov, 1964, 831, Figure 24*)



3.10 Permafrost in the Soviet Union (*after Kudryavtsev, Kondrat'yeva, and Romanovskiy, 1978, Figure 2*). *Note:* A new type of permafrost map of the USSR (scale 1:4 000 000) has been compiled by Popov, Rozenbaum, and Vostokova (1978) but was not received in time to include in this volume. It is a cryolithologic map; that is, it shows the distribution, structure, and origin (type of cryolithogenesis) of various kinds of ice-bearing permafrost.



3.11 Permafrost in the Peoples Republic of China (*after Academia Sinica, 1975, 20–1*)

permafrost in North America coincides roughly with the $-1°$ isotherm for the mean annual air temperature (R. J. E. Brown, 1967*a*; 1970, 21; 1975, 30; Gold and Lachenbruch, 1973, 5); according to others it approximates the 0° isotherm (Péwé, 1975, 47). Some variation is expectable because of regional differences in topography, vegetation, snow cover, and other factors. Not only the amount of snow but also its timing relative to freeze-up is important (R. J. E. Brown, 1975, 29). Discontinuous permafrost in Canada is in the belt of greatest peatland concentration, and the presence of permafrost here is controlled in part by the insulating qualities of the peat (Figure 3.14).

Permafrost in highlands may extend far south of the limit as usually mapped. In the Canadian Cordillera where the southern limit is mapped at valley-bottom levels the lower altitudinal limit is estimated to rise from 1220 m (4000 ft) at lat. 54°30′N to about 2134 m (7000 ft) at lat. 49°N (R. J. E. Brown, 1970, 12). It is present at altitudes of 2655 m and 2691 m on Lookout Mountain on the Continental Divide between Alberta and British Columbia (Scotter, 1975). In the southern Coast Mountains of British Columbia, permafrost has been reported at lat. 49°58′N near Garibaldi Lake (Mathews, 1955, 96), but it is probably relic, or the result of the 'Balch' effect (cf. Balch, 1900, 147–9, 159–61) involving cold air settling where air circulation is restricted, inasmuch as the ground here, consisting of loose cinders, has barely frozen to a depth of 20 cm in winter since 1958 (J. R. Mackay, personal communication, 1978). In the USSR both continuous and discontinuous permafrost, excluding highland occurrences, fails to reach as far south as in North America (Figure 3.3). That discontinuous permafrost in North America may be in disequilibrium was noted above; the same imbalance probably exists in the USSR (cf. Embleton and King, 1975, 32).

Significant amounts of alpine permafrost exist in various parts of the world. The area covered in the middle and low latitudes of the Northern Hemisphere as estimated by Gorbunov (1978, 283) is less than 160×10^4 km², excluding uplands of the Far East and eastern Siberia. As compiled by Fujii and

3.10 Permafrost in the Soviet Union (*after Kudryavtsev, Kondrat'yeva, and Romanovskiy, 1978, Figure 2*). *Note:* A new type of permafrost map of the USSR (scale 1:4 000 000) has been compiled by Popov, Rozenbaum, and Vostokova (1978) but was not received in time to include in this volume. It is a cryolithologic map; that is, it shows the distribution, structure, and origin (type of cryolithogenesis) of various kinds of ice-bearing permafrost.

Table 3.3. Area of alpine permafrost in the middle and low latitudes of the Northern Hemisphere (*after Fujii and Higuchi, 1978, Table 2*)

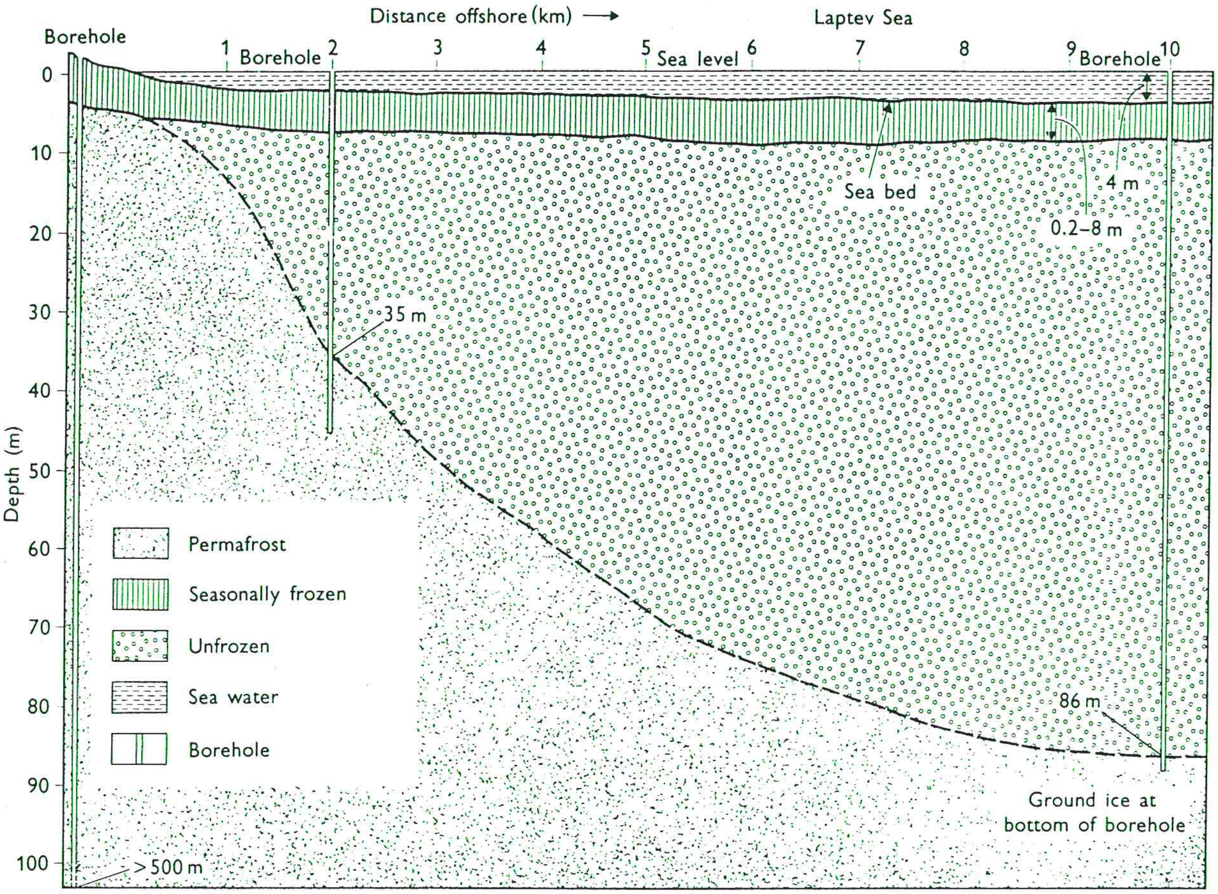
	Location	*Area*	*Total area*
ASIA	Tibetan Plateau Karakoram Mountains	157.8×10^4 km²	
	Himalaya Mountains	10.0×10^4 km²	
	Tien Shan Mountains Pamir	19.1×10^4 km²	
			186.9×10^4 km²
EUROPE	Alps Mountains	0.5×10^4 km²	0.5×10^4 km²
NORTH AMERICA	Rocky Mountains (Canada) Coast Mountains (Canada)	27.8×10^4 km²	
	Rocky Mountains (USA) Sierra Nevada (USA)	17.4×10^4 km²	
			45.2×10^4 km²
NORTHERN HEMISPHERE			232.6×10^4 km²

Higuchi 1978) and shown in Table 3.3, the amount is considerably greater – 232.6×10^4 km². Of this amount 80.4 per cent is in Asia, 19.4 per cent in North America, and only 0.2 per cent in Europe (although in the Swiss Alps permafrost ice is estimated to have a volume comparable to 9 per cent of the glaciers – Barsch, 1978); the lower limit of various occurrences is shown in Table 3.4. Some Canadian examples were noted above. In the United States permafrost has been reported in peat at an altitude of 2957 m (9700 ft) in the Beartooth Mountains of northwestern Wyoming (Pierce,

Table 3.4. Altitude of lower limit of alpine permafrost in the middle and low latitudes of the Northern Hemisphere (*after Fujii and Higuchi, 1978, Table 1*)

Area	*Locality*	*Latitude*	*Longitude*	*Lower limit (m)*	*References*
ROCKY Mts and COAST Mts	Near Cassiar	59°17′N	129°48′W	1370	R. J. E. Brown (1969)
	Beartooth Mts.	44°53′N	109°30′W	<2960	Pierce (1961)
	Niwot Ridge	40° N	106° W	3500	J. D. Ives and Fahey (1971)
	Tesuque Peak	35°47′N	105°47′W	3720	Retzer (1965)
	Mt Elbert	39°07′N	106°27′W	4000	Baranov (1959)
	Mt Whitney	36°35′N	118°17′W	(4420)	Retzer (1965)
MEXICO	Citlaltepetl	18°30′N	97°50′W	4600	Lorenzo (1969)
E. CANADA	Mt Jacques Cartier	49° N	66° W	1270[1]	R. J. E. Brown (1967*a*)
E. USA	Mt Washington	44°15′N	71°20′W		Antevs (1932)
ANDES	Central Chile Andes	33° S	70° W	4000	Lliboutry (1956)
ALPS	Corvatsch Mt	46°25′N	9°50′E	2700	Barsch (1969*b*)
TIEN SHAN	Tien Shan	42° N	78° E	2700	Gorbunov (1967)
HIMALAYA	Mukut Himal	28°45′N	83°30′E	5000	Fujii and Higuchi (1976)
	Khumbu Himal	27°55′N	86°50′E	4900∼5000	Fujii and Higuchi (1976)
	Near Rongbuk Gl.	28°10′N	86°50′E	4900	Hsieh *et al.* (1975)
TIBET	Khulun Shan	31°20′N	91°40′E	4500	Chou and Tu (1963)
	Nienching Tangkula Shan	36°20′N	94°50′E	4200	Chou and Tu (1963)
HAWAII	Mauna Kea	19°30′N	155°40′W	4170	Woodcock (1974)
JAPAN	Mt Fuji	35°21′N	138°44′E	2800∼2900	Fujii and Higuchi (1972)
	Mt Taisetsu	43°40′N	142°55′E	2150	Fukuda and Kinoshita (1974)

[1] The lower limit on Mount Jacques Cartier is now estimated to be at 1100 m. Extrapolated ground temperatures indicate that permafrost is 50–100 m thick at an altitude of 1280 m, just above treeline (*R. J. E. Brown, personal communication, 1979*).

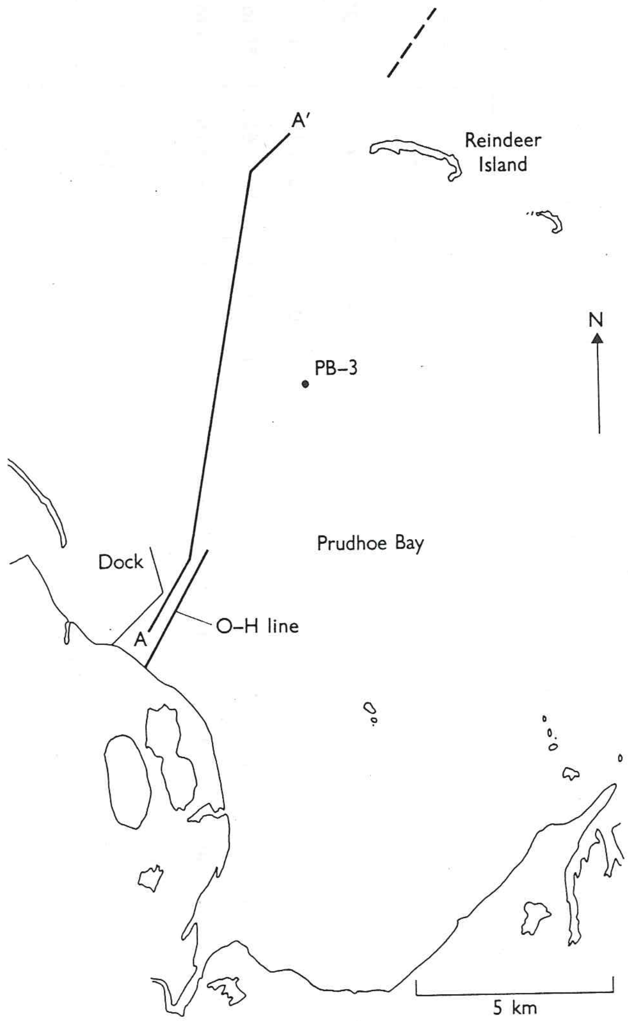


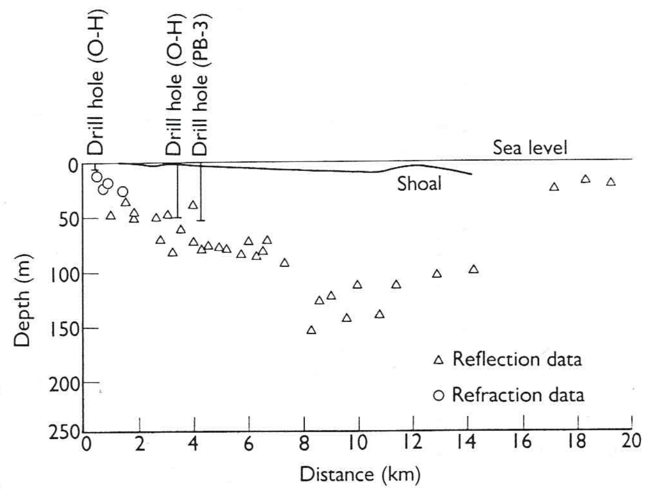
3.12 Offshore permafrost in the eastern Laptev Sea, USSR (*after Vigdorchik, 1977b, Figure 18; cf. Grigor'yev, 1973; 1978; Katasanov and Pudov, 1972; Zhigarev and Plakht, 1974; 1976*)

1961). It occurs in the summit area of Mount Washington (alt. 1917 m) in New Hampshire (R. P. Goldthwait, 1969; John Howe, 1971; Thompson, 1962, 215) where the mean annual temperature is −2.8° (27°F), and it characterizes some areas above treeline in the Rocky Mountains (J. D. Ives, 1973; J. D. Ives and Fahey, 1971). Woodcock (1974) described a curious occurrence at an altitude of 4140 m in the summit area of Mauna Kea (alt. 4206 m), Hawaii. The active layer is about 0.4 m thick and the permafrost extends to a depth of at least 10 m despite a mean annual air temperature of 3.6° at the permafrost altitude. The anomaly is partly explained by a colder microclimate resulting from radiational cooling and entrapment of air, low incidence angle of sunlight, and dry atmosphere. This could account for the top of the permafrost maintaining itself under the present conditions, although the base may be retreating. The permafrost probably originated under a colder climate or is a 'Balch' effect (cf. preceding discussion of the Garibaldi Lake, British Columbia, permafrost occurrence. Both occurrences are in loose cinders). Various permafrost occurrences are known at high altitudes in Europe (Barsch, 1969*b*; 1973; 1977*d*; Furrer and Fitze, 1970; 1973; Haeberli, 1975; 1978), Japan (Fujii and Higuchi, 1972; Fukuda and Kinoshita, 1974) and elsewhere, including even tropical latitudes given high enough altitudes as at Mauna Kea (discussed above), in Africa where 'fossil ice' (ice-cored glacial deposits) and frozen talus occur on Mount Kenya (alt. 5199 m) (17 058 ft) (Baker, 1967, 68), and in Mexico where relict permafrost occurs within an altitude range of 4600–5000 m on Pico de Orizaba (Heine, 1975).

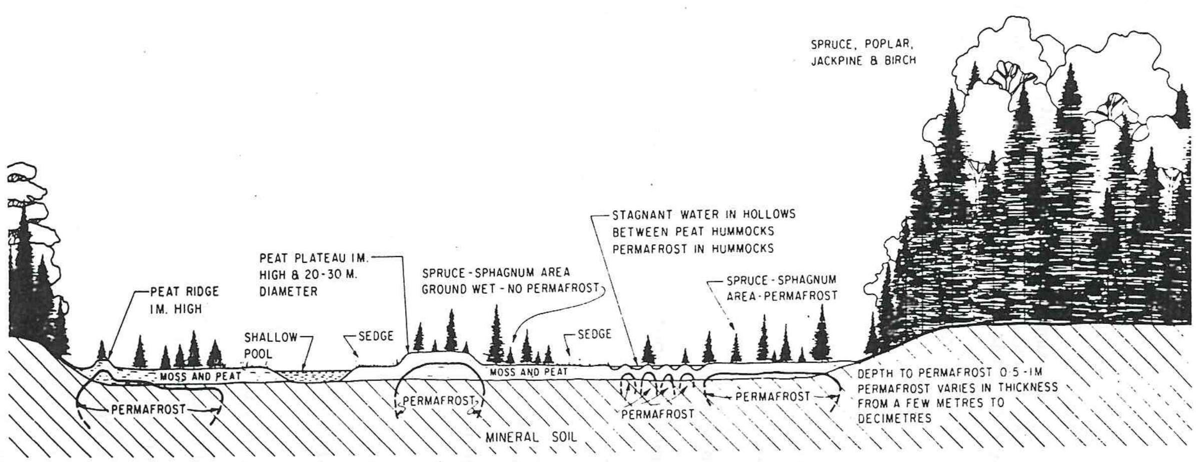
4 Depth and thickness

To date the maximum reported thickness of permafrost is 1400–1450 m in the vicinity of the upper Markha River in Siberia (Grave, 1968*a*; 1968*b*, 3, diagram opposite 4). Bakker (*in* Büdel, 1960, 43) reported *c.* 1000 m in Siberia, and a calculated thickness of 900 m in the Udokan Range, Siberia (Melnikov, 1963). In North America the maximum reported thickness is about 1000 m near Alert, Ellesmere Island, in the Canadian Arctic, as extrapolated from a temperature of $-16.5°$ at the depth of zero annual amplitude (R. J. E. Brown, personal communication, 1979). The depth is on the order of 700 m at Cameron Island (northwest of Bathurst





3.13 Upper surface of offshore ice-bonded permafrost along transect A–A′ as indicated by drilling and seismic data, Prudhoe Bay, Alaska (*after Rogers and Morack, 1977a, 18, Figure 4; 20, Figure 5 [Line 15 omitted]; 1977b; cf. 1978, Figure 2* – sparker data omitted)



3.14 Profile through typical peatland in southern fringe of discontinuous permafrost zone in Canada, showing relation of permafrost to vegetation and topography (*R. J. E. Brown, 1968, 176, Figure 2*)

Table 3.5. Approximate depth to bottom of permafrost at selected places in Northern Hemisphere. Bottom is generally based on 0°C. Source data originally given in °F or in feet are so indicated; feet are converted to nearest metre. Mean annual temperatures are after authors cited unless otherwise indicated. See also Oberman and Kakunov (1978, 144–5, Table 1)

Location	*Depth*		*Mean annual air temperature*		*Ground temperature at depth indicated*			
	M	*Ft*	*°C*	*°F*	*°C*	*°F*	*M*	*Ft*
Alaska								
Barrow[6,8]	204 to 405	670 to 1330	−12.0	—	−6.3	—	179	—
Bethel[6]	13 to 184	42 to 603	−1.8[11]	—	—	—	—	—
Cape Simpson[6,8]	250 to 320	820 to 1050	−12.0	—	−9.9	—	42	—
Cape Thompson[6]	305 to 366	1000 to 1200	—	—	—	—	—	—
Fairbanks[6]	30 to 122	100 to 400	−3.5[11]	—	−7.0	—	25	—
Kotzebue[6]	73	238	−6.2[11]	—	—	—	—	—
Nome[6]	37	120	−3.5[11]	—	—	—	—	—
Northway[6]	63	207	—	—	—	—	—	—
Prudhoe Bay[4]	610	2000	—	—	—	—	—	—
Umiat[6,8]	322	1055	−12.0	—	−7.3	—	21	—
Canada								
Aishihik, YT[1,6]	15 to 30	50 to 100	−4.2	24.5	−2.1	28.3	6	20
Alert, NWT	1000[c]	—	−18.0	−0.4	−16.1	3.0	15 to 18	50 to 60
Asbestos Hill, PQ[1]	>274	>900	−8.3	17.0	−7.2 to −6.7	19.0 to 20.0	15 to 61	50 to 200
Churchill, Man.[1]	30 to 61	100 to 200	−7.2	19.0	−2.5 to −1.7	27.5 to 28.9	8 to 16	25 to 54
Churchill, Man.[6]	0 to 42[9]	0 to 140[9]	—	—	—	—	—	—
Coppermine River, NWT[12]	180 to 370[c]	—	−11.4	—	−7.0	—	14	—
Dawson, YT[1]	61	200	−4.7	23.6	—	—	—	—
Fort Good Hope, NWT[12]	33 to 48[c]	—	−7.6	—	−1.0 to −1.2	—	15	—
Fort McPherson, NWT[12]	90 to 150[c]	—	−8.2	—	−2.6 to −4	—	14	—
Fort Simpson, NWT[1]	12	40	−3.9	25.0	0.7 to 1.9	33.2 to 35.4	0 to 2	0 to 5
Fort Simpson, NWT[12]	0 to 5	—	−3.9	—	1.4 to 2.0	—	15	—

Table 3.5. (continued)

Location	Depth		Mean annual air temperature		Ground temperature at depth indicated			
	M	Ft	°C	°F	°C	°F	M	Ft
Fort Smith, NWT[1]	(unknown)		−3.2	26.2	≃0.0	≃32.0	5	15
Fort Vermilion, Alta.[1]	0	—	−2.1	28.2	4.3 to 4.4	39.8 to 39.9	0 to 2	0 to 5
Hay River, Alta.[6]	2	5	—	—	—	—	—	—
Inuvik, NWT[1]	> 91	> 300			−3.3	26.0	8 to 30	25 to 100
Inuvik, NWT[12]	107[c]	—	−9.6	—	−3.7	—	14	—
Keg River, Alta.[1]	2	5	−0.6	31.0	−0.6 to 0.0	31.0 to 32.0	2	5
Kelsey, Man.[1]	15	50	−3.6	25.5	−0.8 to −0.3	30.5 to 31.5	9	30
Mackenzie Delta, NWT[1]	91	300	−9.1 (Aklavik)	15.6	−4.6 to −3.1	23.8 to 26.5	0 to 30	0 to 100
Mackenzie Delta, NWT[12]	18 to 366[c]	—	−9.1 to −11.3	—	−0.4 to −6.0	—	14	—
Mary River, Baffin I., NWT[1]	(unknown)		−14.3 (Pond Inlet)	6.3	−12.2	10.0	9	30
Milne Inlet, Baffin I., NWT[1]	(unknown)		−14.3 (Pond Inlet)	6.3	−12.2	10.0	15	50
Norman Wells, NWT[1,6]	46 to 61	150 to 200	−6.2	20.8	−3.3 to −1.9	26.0 to 28.5	15 to 30	50 to 100
Port Radium, NWT[1]	107	350	−7.1	19.2	—	—	—	—
Rankin Inlet, NWT[1]	305	1000	−11.6 (Chesterfield Inlet)	11.2	−9.4 to −8.3	15.0 to 17.0	30	100
Resolute, Cornwallis I., NWT[1,8]	396[c]	1300	−16.2	2.8	−4.1 −9.9 −5.6	24.6 25.0 21.9	16 30 135	52 98 443
Schefferville, PQ[1,6]	> 76	> 250	−4.5	23.9	−1.1 to −0.3	30.0 to 31.5	8 to 58	25 to 190
Thompson, Man.[1]	15	50	−3.9	24.9	−0.6 to 0.0	31.0 to 32.0	8	25

Thompson, Man.[1]	15	50	−3.9	24.9	−0.6 to 0.0	31.0 to 32.0	8	25

Table 3.5 (continued)

Tuktoyaktuk and coast, NWT[12]	140[c]	—	−11.9 to −12.8	—	−8.0	—	14	—
Tundra Mines Ltd. NWT[1]	274	900	−8.3	17.0	−1.7	29	99	325
United Keno Hill Mines Ltd., YT[1]	137	450	−4.3 (Elsa)	24.2	−2.2 to −1.5	28.0 to 29.3	30 to 61	100 to 200
Uranium City, Sask.[1,6]	9	30	−4.4	24.0	−0.6 to 0.0	31.0 to 32.0	9	30
Winter Harbour, Melville I., NWT[5]	557	—	−16.0[2]	—	—	—	—	—
Yellowknife, NWT[1,6]	61 to 91	200 to 300	−5.4	22.2	−0.3	31.4	12	40
Yellowknife, NWT[12]	0 to 80	—	−5.4 to −6.2	—	1.8 to −0.5	—	15	—
Greenland Thule[6]	518[10]	1700[10]	—	—	—	—	—	—
Spitsbergen ....[6]	241 to 305	790 to 1000	—	—	—	—	—	—
Braganza Bay, Lowe Sound[7]	320	—	—	—	—	—	—	—
USSR Amderma[8]	500[c]	—	—	—	−4.6 −5.3 −3.8	— — —	20 100 270	— — —
Bykhanay[8]	650[m]	—	−12.0	—	−3.8 −0.2 10.4	— — —	50 500 1000	— — —
Dzhebariki-Khaya[6]	416[c]	—	−11.0	—	−5.5 −3.7	— —	30 170	— —
Kozhevrikova[8]	600[m]	—	−13.0	—	−12.5 −3.6	— —	15 400	— —
Magan[8]	450[c]	—	−10.0	—	−3.0 −0.8 −0.5	— — —	20 300 400	— — —
Markha River, upper reaches[3]	1450 to 1500	—	—	—	—	—	—	—
Mirny[8]	550[m,c]	—	−9.0	—	−2.7 −1.8 0.0	— — —	15 300 550	— — —

Table 3.5. (continued)

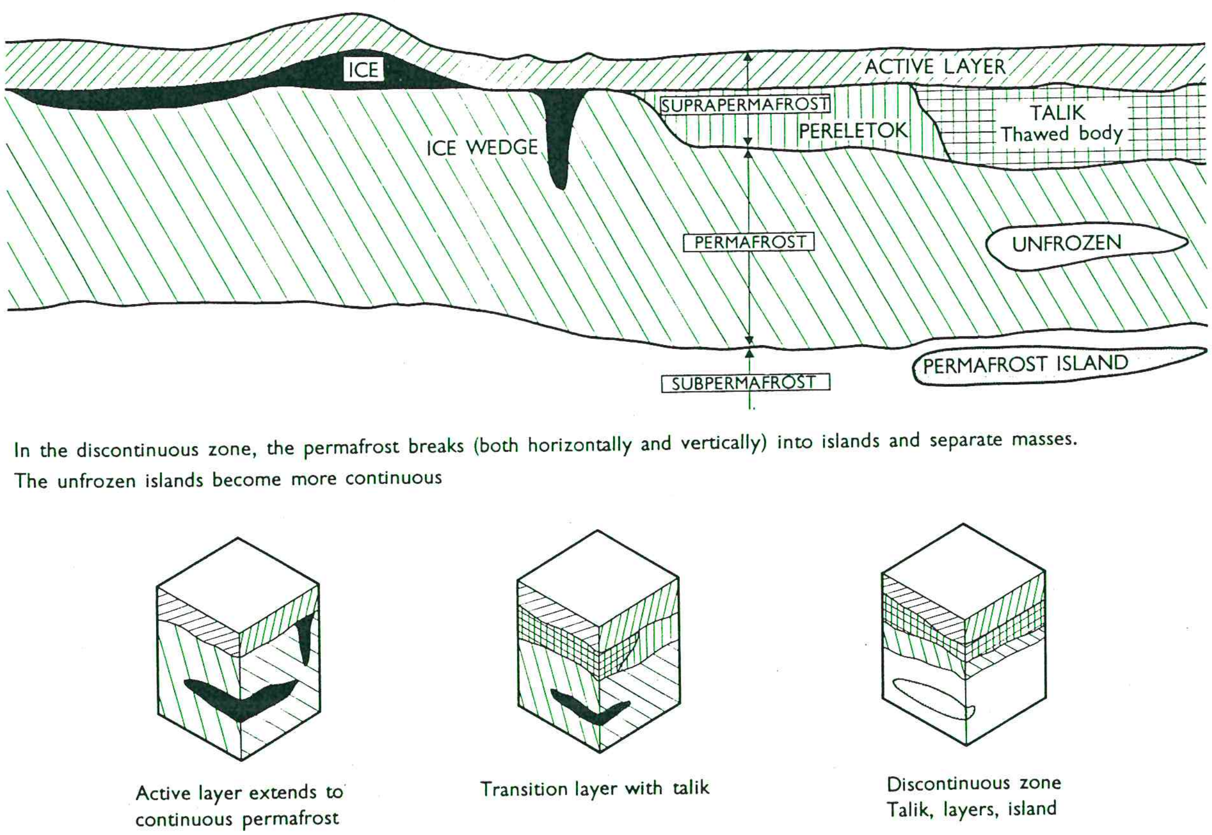
Location	Depth		Mean annual air temperature		Ground temperature at depth indicated			
	M	*Ft*	°C	°F	°C	°F	*M*	*Ft*
Namtsy[8]	560[c]	—	−10.0	—	−2.5 −2.2 −1.6	— — —	20 200 500	— — —
Nordvik[6]	610	2000	—	—	—	—	—	—
Noril'sk[8]	325[m] 400[c]	—	−8.0	—	−7.5 −0.3 0.2	— — —	50 320 330	— — —
Salekhard[8]	350[m] 380[c]	—	−7.0	—	−4.0	—	20	—
Taimyr[6]	305 to 610	1000 to 2000	—	—	—	—	—	—
Tiksi[8]	630[m]	—	−14.0	—	−11.1 −8.8 −3.3	— — —	20 200 500	— — —
Ukodan Range[8]	900[c]	—	−12.0	—	−7.5 −5.0 −3.3	— — —	50 300 500	— — —
Ust'port[8]	425[m] 500[c]	—	−11.0	—	−3.0 −1.2	— —	50 330	— —
Var'yegan[8]	>315	—	—	—	—	—	—	—
Vilyuy River, mouth[8]	420[c]	—	−10.0	—	−2.1 −1.0	— —	20 240	— —
Vorkuta[6]	131	430	—	—	—	—	—	—
Yakutsk[6]	198 to 250	650 to 820	−10.13[11]	—	—	—	—	—

[1] R. J. E. Brown (1970, 10, Table 1).
[2] Embleton and King (1975, 30, Table 2.1).
[3] Grave (1968*a*; 1968*b*, 3, diagram opp. 4).
[4] Howitt (1971); Robert Stonely *in* Lachenbruch (1970*b*, J2–J4).
[5] A. Jessop *in* R. J. E. Brown (1972, 116).
[6] Stearns (1966, 21, Table 1).
[7] Werenskiold (1953, 197).
[8] Yefimov and Dukhin (1966, 94–5; 1968).
[9] Varies with distance from Hudson Bay and Churchill River.
[10] Estimated from temperature gradient and from temperature at 306 m (1005 ft) depth.
[11] US Department of Commerce (1975).
[12] Judge (1973*b*, 125, Table 35; 129–31, Table 36; 138, Figure 17).
[13] R. J. E. Brown (personal communication, 1979). See text.
[c] Calculated.
[m] Measured.

Island), in the Mackenzie Delta area, and in northern Quebec (A. E. Taylor and Judge, 1977, 11–12, Table 2), although climatic considerations indicate thicknesses may be considerably greater in the interior of Baffin Island as well as Ellesmere Island (R. J. E. Brown, 1972, 116–17). In Alaska the greatest reported thickness is at Prudhoe Bay where 610 m (2000 ft) has been cited (cf. Howitt, 1971; Robert Stoneley *in* Lachenbruch, 1970*b*, J2–J4) and 650 m is given in Figure 3.7. Table 3.5 shows approximate thicknesses in the Northern Hemisphere. Regional variations in thickness in Alaska, Canada, and the USSR are shown in Figures 3.7, 3.5, and 3.9, respectively.

In addition to other methods, a number of geophysical approaches, such as seismic surveying (including hammer seismic techniques – cf. Lorenz King, 1977), gravity profiling (for massive ice – Rampton and Walcott, 1974), electrical resistivity (J. Henderson and Hoekstra, 1977; Seguin, 1977), and remote-sensing techniques are being used or tested for delineating the distribution of permafrost and massive ice within permafrost. The determination of permafrost thickness and character is difficult but is being actively attacked (J. L. Davis *et al.*, 1976; Ferrians and Hobson, 1973, 481–90; Mackay, 1975*f*), promising approaches being magnetotelluric sounding (Koziar and Strangway, 1978), and electrical properties (Frolov, 1976) including drawing of isothermal curves from electrical logging (Seguin, 1977). The application of geophysics to permafrost regions continues to expand as reviewed by W. J. Scott, Sellmann, and Hunter (1979).

Permafrost is not limited to the Earth. Martian surface temperatures indicate its presence, and perhaps some of it is ice rich (D. M. Anderson, Gatto, and Ugolini, 1973; Carr and Schaber, 1977; Kuz'min, 1977; Sharp, 1974), although this has been questioned (Black, 1978*a*). Possibly some Martian ground ice occurs as a hydrate of CO_2 (Milton, 1974).



3.15 Some structural features of permafrost (*after Stearns, 1966, 17, Figure 7*)

Table 3.6. Classification of frozen ground (*after Pihlainen and Johnston, 1963, Table 1, 12–17*)

Ice not visible by naked eye
Poorly bonded
Well bonded
Ice visible by naked eye but 2.5 cm (1 in) or less thick
Individual ice crystals or inclusions
Ice coatings on particles
Random or irregularly oriented ice formations
Stratified or distinctly oriented ice formations
Ice greater than 2.5 cm (1 in) thick
Ice with soil inclusions
Ice without soil inclusions

5 Structure

The basic types of frozen ground in a Canadian (Pihlainen and Johnston, 1963) and CRREL classification (Linell and Kaplar, 1966; cf. Andersland and Anderson, 1978, 29–36; R. F. Scott, 1969, 17) are shown in Table 3.6.

Undersaturated, saturated, and supersaturated permafrost (or frozen ground) refer, respectively, to ground in which ice fails to fill all pores, about fills all pores, and exceeds all the pore space (Black, 1953, 127–8). Pore ice (or interstitial ice) is confined to pore spaces, whereas massive ice (large mass of supersaturated ice) commonly weighs 1000 per cent more than any contained soil (Mackay and Black, 1973, 186–7). However, the ice content of peat, even in the absence of ice lenses, can be as much as 2000 per cent of the dry weight (Kinosita *et al.*, 1979, 20, 41). Segregated (or segregation) ice refers primarily to origin rather than distribution of ice in frozen ground, and designates ice masses formed as the result of water being brought to the freezing plane by the freezing process itself (Taber, 1929, 430. Cf. Freezing process in chapter on General frost-action processes). In the CRREL classification above, the term segregated ice rather than ice was used in the original. It was employed in a nongenetic sense and has been omitted by the present writer to avoid confusion, since the genetic usage is dominant and is followed here.

The thickness and orientation of ice accumulations are especially critical aspects of the structure. The amount of ice tends to control the behaviour of permafrost upon thawing, and the amount and orientation of the accumulations may indicate the origin of the ice.

Some characteristic structures of permafrost as determined by certain forms of ice, including ice lenses and ice wedges described below, are shown in Figure 3.15, which also shows certain features described later, including thaw areas (taliks), temporarily frozen areas simulating permafrost (pereletoks), the upper surface of permafrost (permafrost table), the ground above (suprapermafrost layer), and the layer of seasonal freezing and thawing (active layer). The suprafrost layer is not necessarily the same as the active layer because a pereletok or talik may lie between the permafrost table and the active layer.

6 Forms of ice

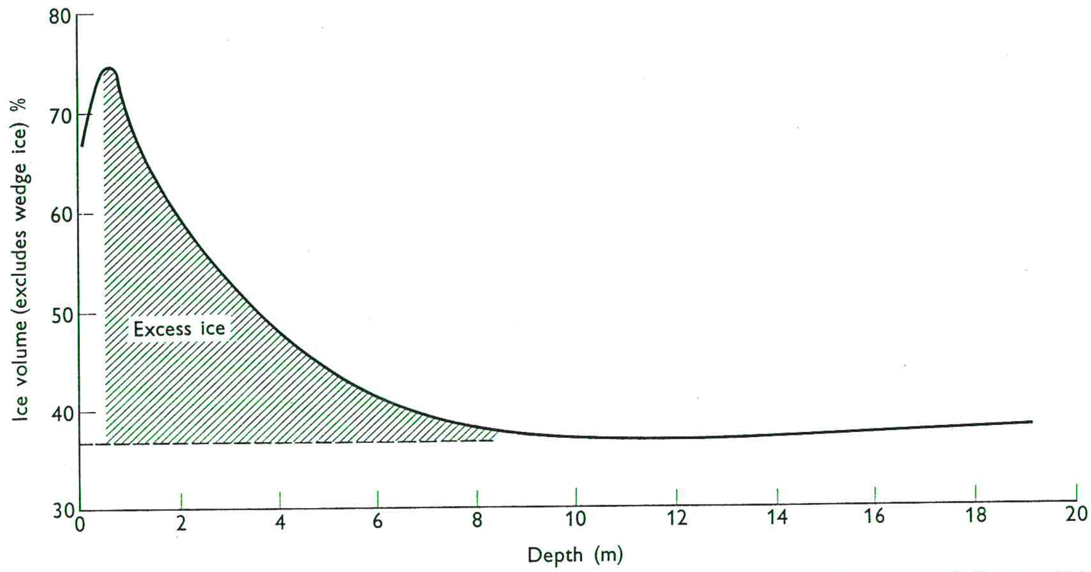
a General Ground ice, defined as ice in the ground regardless of amount or configuration (National Academy of Sciences, 1974, 31),[3] tends to be concentrated in the upper levels of permafrost (Figures 3.15–3.16), but massive ice has been encountered to depths of 45 m or more (Figure 3.17). Compared with ice crystals in seasonally frozen ground, ice crystals in permafrost have been described as being 14–25 times larger except where shearing of the permafrost has caused much smaller crystals by recrystallization (Tsytovich, 1975, 104–5). Ground ice may take many forms, some of the more common being pingo ice, ice lenses and massive ice beds, ice veins, and ice wedges, and more rarely buried glacier ice and buried icings. Overviews include the discussion by Mackay (1972*a*), Mackay and Black (1973), Mackay, Konishchev, and Popov (1979), and the monographs (in Russian) by Vtyurin (1975) and Vtyurina (1974), the last being devoted to seasonally frozen ground. Both the growth and the thawing of ground ice are the direct cause of many characteristic periglacial landforms as stressed by numerous workers (cf. R. J. E. Brown, 1973*a*; Rampton, 1973). A detailed classification of ground ice, based on the origin of the water and on the transfer process involved, is shown in Figure 3.18. Gas hydrates may resemble ice and possibly constitute a gas resource in permafrost and also beneath it (Davidson *et al.*, 1978; Judge, 1977, 103–4; Katz, 1971). It is often difficult to distinguish between different forms of ground ice by the nature of the ice alone, independent of its form and occurrence, but structure, petrofabrics, and geochemistry of the ice can be

[3] There are also other usages. A widely divergent one (Alekseyev, 1973; 1974) includes all ice at air-ground interface and excludes most ground ice as the term is generally employed.

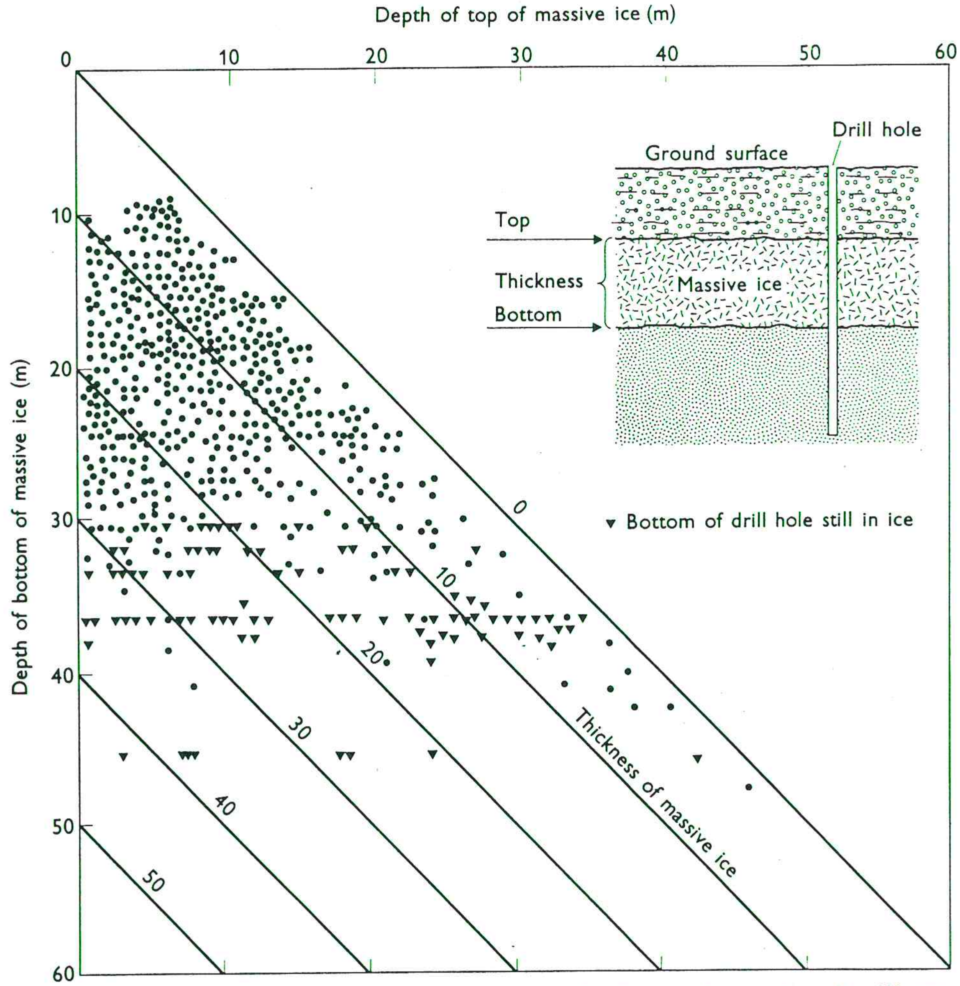
rost as
ng ice
shown
atures
, tem-
perele-
nafrost
layer),
awing
ssarily
tok or
nd the

round
itional
e con-
igures
ntered
npared
id, ice
being
of the
als by
round
mmon
ds, ice
glacier
liscus-
1973),
nd the
) and
onally
awing
harac-
nerous
1973).
ed on
process
es may
source
et al.,
s often
rms of
inde-
icture,
can be

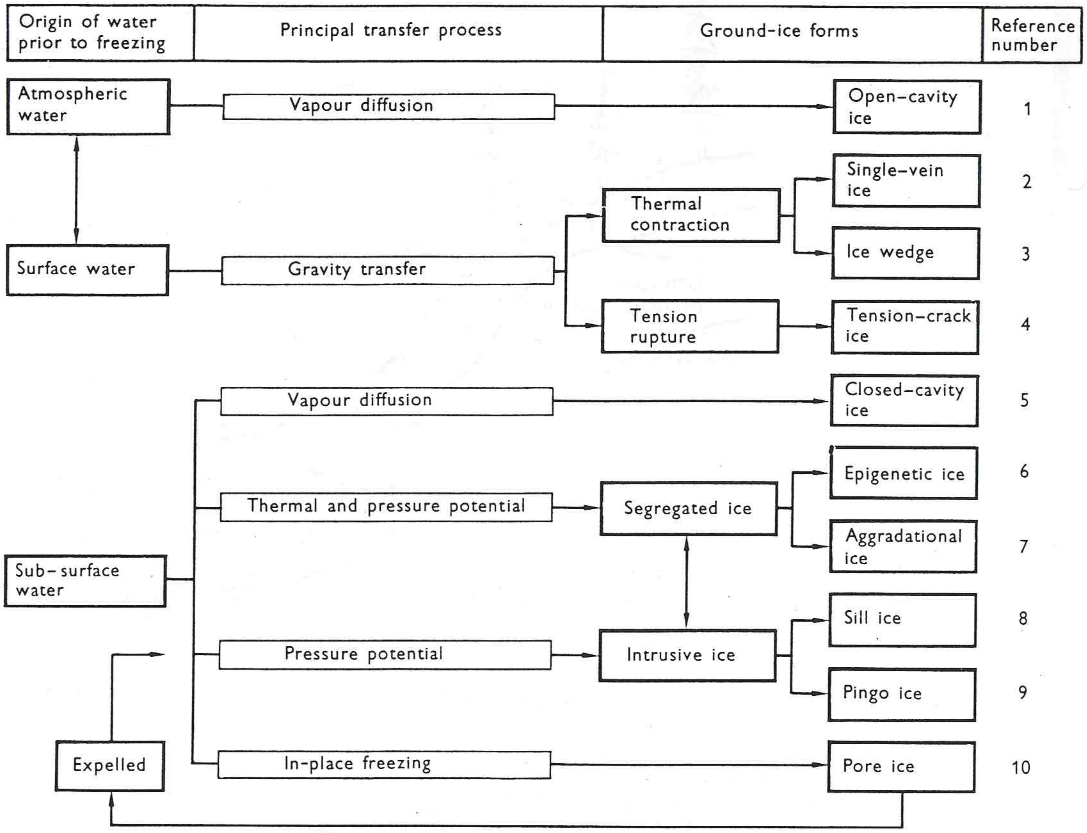
ice and



3.16 Ice volume versus depth based on auguring and coring, Barrow area, Alaska (*after Jerry Brown and Sellmann, 1973, 36, Figure 3*)



3.17 Thicknesses of massive ice, plotted from 560 beds encountered in drill hole logs in Canadian Western Arctic (*after Mackay, 1973a, 225, Figure 4*)



3.18 Classification of ground-ice forms (*after Mackay, 1972a, 4, Figure 2. Courtesy, Assoc. American Geographers*)

diagnostic in places. The fact that elongation of air bubbles in ice develops normal to the cooling surface and follows the heat-flow direction is helpful (Gell, 1974*a*).

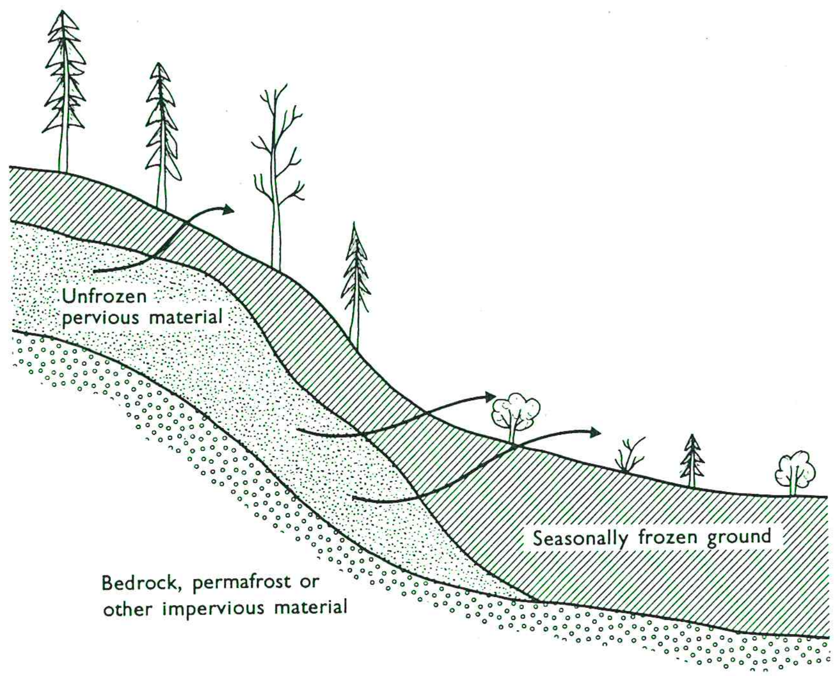
b Glacier ice Glacier ice may become buried under insulating debris and preserved as ground ice. The structure of glacier ice is rarely horizontal over large distances; rather it tends to be folded and faulted as a result of glacier movement. The nature of glacier ice and movement is reviewed in glaciological texts, including Paterson (1969) and Shumskiy (1959; 1964*a*), and further discussion here is omitted as beyond the scope of the present volume.

c Icings An icing (sometimes called Aufeis or naled) is '... a mass of surface ice formed during the winter by successive freezing of sheets of water that may seep from the ground, from a river, or from a spring' (S. W. Muller, 1947, 218) (Figures 3.19–3.21).[4] Like glacier ice, icings may become buried and preserved as ground ice, as reported, for example, near the head of an outwash plain in Spitsbergen (Cegła and Kozarski, 1977). Although descriptions of burial and preservation appear to be rare, icings *per se* are widespread periglacial features as discussed in Chapter 9.

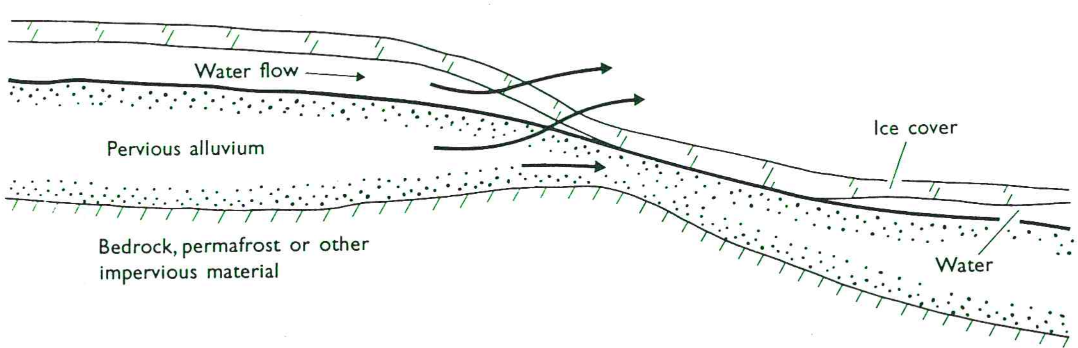
The structure of icings commonly parallels the underlying surface (Shumskiy, 1964*b*, 192–5), and unlike glacier ice it is rarely if ever deformed. The ice ('overflow ice') of river icings is reported to have a higher average extinction coefficient (0.43 cm^{-1}) than normal (clear) river ice (0.092 cm^{-1}) (Wendler, 1970). The fact that most large river icings contain little dirt and overlie clean gravel in valley bottoms helps to distinguish buried icings from beds of massive segregated ice, discussed below, which commonly contain silt bands and occur in silty soil.

d Pingo ice Pingo ice, like buried icings and

[4] It has been suggested that icing be used to designate the process only, and naled the deposit (Harden, Barnes, and Reimnitz, 1977, 28–9) but the need for this is not persuasive. Although often used synonymously, Aufeis as defined in the American Geological Institute's *Glossary of geology* (Gary, McAfee, and Wolf, 1972, 46) refers to river icings only. This would make icing the more general term but it should be noted that the original use of Aufeis was equally general (Middendorff, 1853; 1861, 439–53).



3.19 Development of icing by freezing of groundwater flowing from seepages or springs (*after Carey, 1973, 28, Figure 16a*)

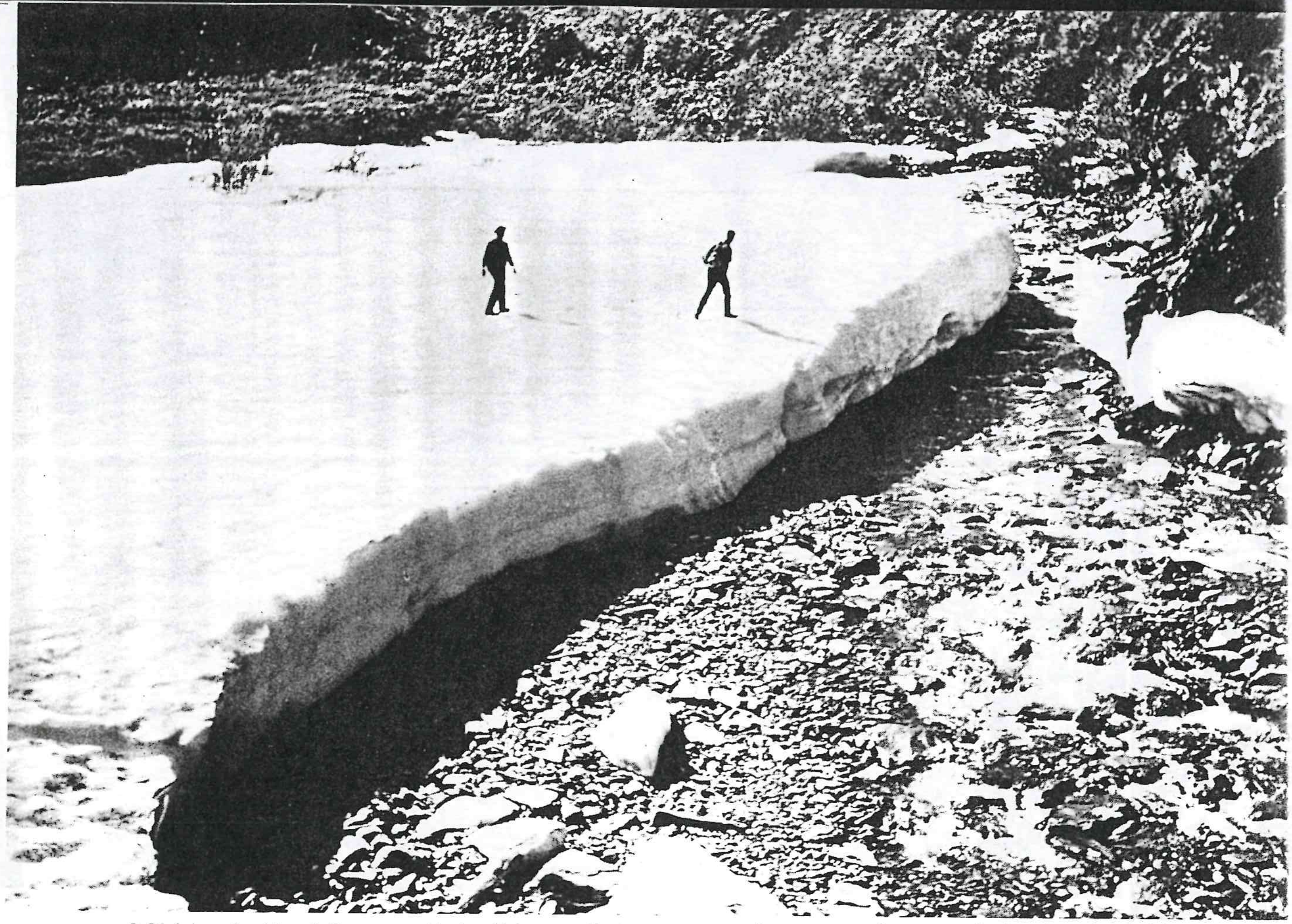


3.20 Development of icing by a river breaking through its ice cover during freeze-up (*after Carey, 1973, 27, Figure 15b*)

glacier ice, is commonly a massive body of fresh-water ice. It is usually formed by (1) progressive, all-sided freezing of a water body or water-rich sediments or (2) injection of groundwater under artesian pressure into permafrost in the manner of a laccolith. In both cases, ice forms the core of a mound and tends to assume a planoconvex shape. Pingos are described in the chapter on Some periglacial forms.

The structure of pingo ice is not well known. Drilling of a pingo in the Mackenzie Delta area of northern Canada revealed ice with no apparent structure; the ice, some 9 m (30 ft) thick, was milky from small air bubbles (Pihlainen, Brown, and Legget, 1956, 1122). An exposure of pingo ice in the same area showed many crystals with diameters of 2.5–5.0 cm (1–2 in), the maximum crystal size noted being 20 cm (8 in); the optic axes had a preferred orientation towards the centre of the pingo's ice core (Mackay and Stager, 1966, 367).[5] Ice cores are also

[5] See also Gell (1978*b*), published while the present volume was in press.



3.21 Icing, Anaktuvuk Pass area, Alaska. Note smooth stone pavement between rougher stream bed and eroded Aufeis. Photo by S. C. Porter

described as having almost vertical banding and (in another case) irregular shear zones and fractures, perhaps reflecting different modes of growth (French, 1975*a*, 459; Pissart and French, 1976, 941–4, Figures 6–7, 9, 11–12).

e Ice lenses and massive ice beds An ice lens is a dominantly horizontal layer of ice. It may be less than a millimetre to tens of metres thick and a few millimetres to hundreds of metres in extent (Figures 3.22–3.24). However, very large lenses (say over 2 m thick and 10 m in shortest diameter and having an ice content of at least 250 per cent on an ice-to-dry-soil weight basis – Mackay, 1973*a*, 223) are better termed massive ice beds (Figure 3.24), or massive icy beds if the amount of contained soil, commonly silt, is large. Some massive ice beds are up to 40 m or more thick (Figure 3.17) and up to 2 km or more in horizontal extent (Mackay, 1973*a*), and some are responsible for prominent topographic rises as in the peculiar 'involuted hills' of the Mackenzie Delta area of northern Canada (French, 1976*a*, 83; Mackay, 1963*a*, 138) and in pingos.

The terms are purely descriptive but many ice lenses and beds of massive ice form *in situ* by segregation of ice during freezing, as discussed under Freezing process in the chapter on General frost-action processes. If the ice is of segregation origin, its structure tends to be parallel to the freezing surface (i.e. usually dominantly horizontal), with any bubbles tending to be elongated and aligned normal to the horizontal layering. Soil fragments broken by freezing show a vertical separation (Mackay, 1971*b*, 411). It should be noted that according to recent Soviet research some frozen

3.22 Small ice lenses in clayey silt, northwest Siberia, USSR. Scale in centimetres. Photo by A. P. Tyrtikov

3.23 Large ice lens, right bank Lena River, 90 km north of Yakutsk, USSR. Lenses here are in bedded sand and up to 3 m thick

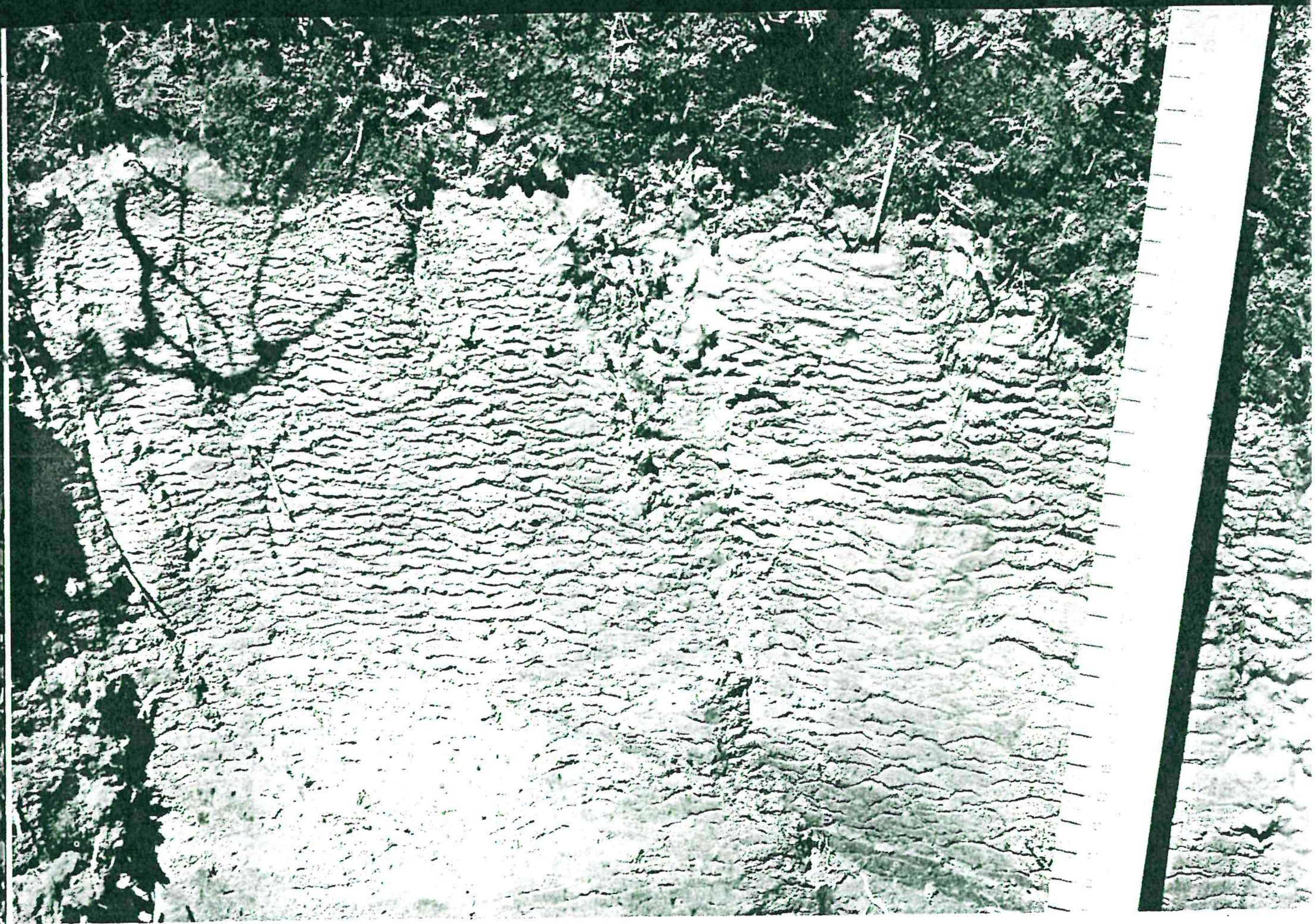
ufeis.

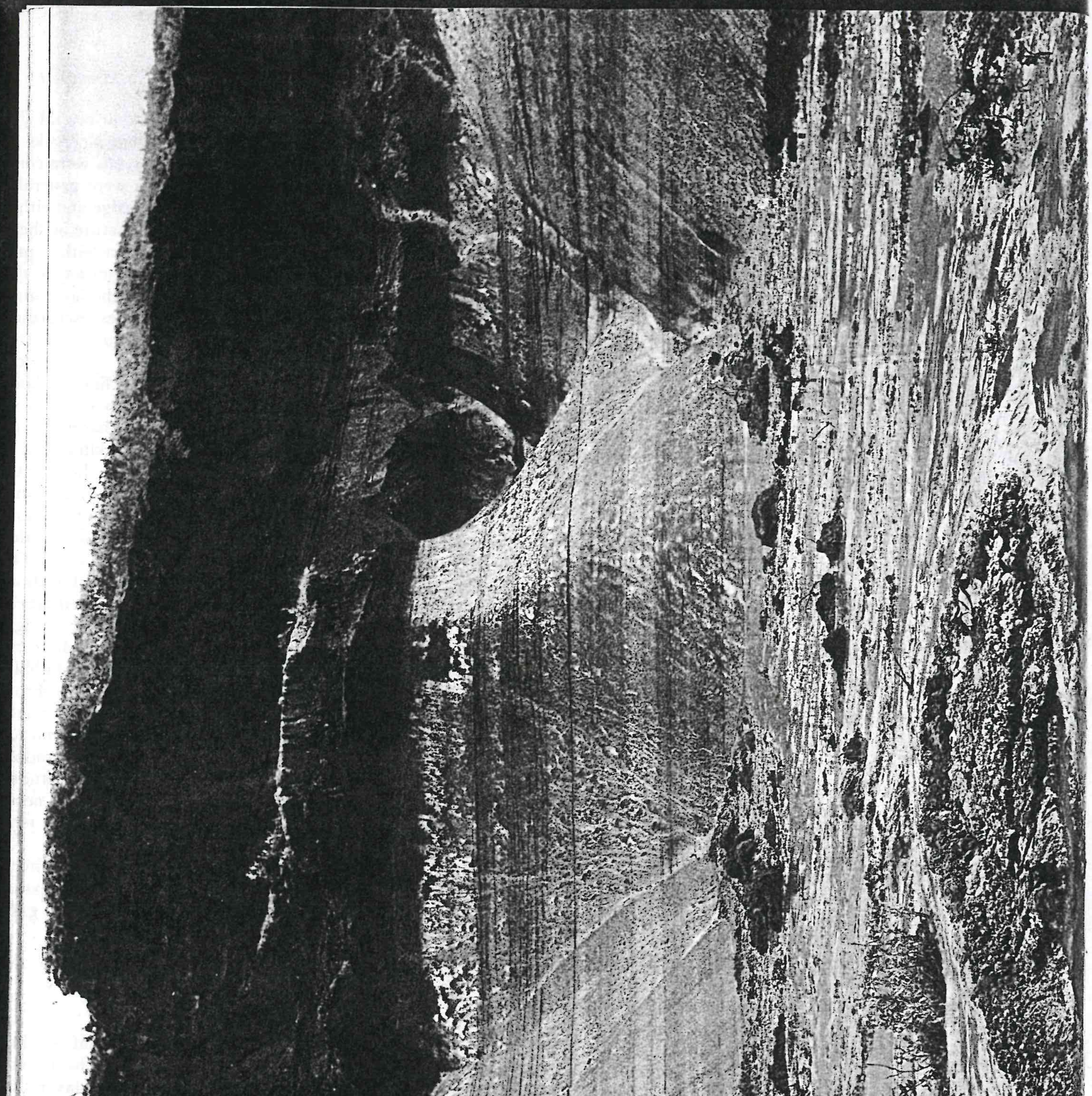
Iackay,

any ice
situ by
d under
l frost-
origin,
reezing
), with
aligned
gments
aration
ed that
frozen

a, USSR.

north of
nd up to





ground textures (fabrics) can change into others with changes in temperature, pressure, and other conditions, including the demonstration that ice lenses can develop from pore ice as a soil warms and thaws (Mackay, Konishchev, and Popov, 1979).

f Ice veins An ice vein, as the term is used by the present author, is a tabular, wedgelike, or irregular body of ice, hair thin to 5 cm (2 in) thick, whose longest dimensions lie in a dominantly vertical plane and may be several metres or more long. The term is purely descriptive. This usage follows that of T. L. Péwé (personal communication, 1970) in stressing the thinness of veins as opposed to ice wedges, discussed below. However, these terms tend to be used interchangeably in translations from the European literature. It should be noted that the term ice vein has also been used for thin (1–10 cm thick) tabular ice masses without regard to orientation (Mackay, 1974*a*, 230–1).

g Ice wedges An ice wedge is '. . . a narrow crack or fissure of the ground filled with ice which may extend below the permafrost table' (S. W. Muller, 1947, 218). As defined, the term is purely descriptive, and this usage is followed here, but to many workers the term has come to imply an origin by frost cracking, discussed later. Ice wedges start as ice veins.

Ice wedges tend to be V-shaped (Figure 3.25) and to have a characteristic, predominantly vertical structure (as opposed to the predominantly horizontal structure of undeformed ice lenses and massive ice beds), imparted by dirt and air bubbles oriented parallel to the wedge edges (Figure 3.26) as a result of the frost-cracking process. The dirt particles and the bubbles tend to be vertically elongated (Black, 1974*a*, 257, 263). The ice has a milky appearance because of many small bubbles less than 0.3 mm in diameter (Gell, 1974*b*). According to Shumskiy (1964*b*, 196–205), the ice crystals tend to be columnar with C axis vertical in the upper part of a wedge and horizontal at depth, but a cataclastic texture may develop as an ice wedge grows.

Ice wedges studied by Black (1954, 844–6; cf. 1963, 265–8; 1974*a*, 251–64) at Barrow, Alaska, had variously shaped ice crystals – equidimensional, prismatic, and irregular – with straight to sutured boundaries. Grain sizes ranged from 0.1 to 100 mm, and Gell's (1974*b*) studies showed that crystal size tends to increase outward from central cracks of wedges. In Black's studies, long crystals were commonly vertical whereas short ones were generally normal to the horizontal axis of a wedge and either horizontal or inclined normal to a fracture or shear plane. The optic axis directions were either predominantly vertical, horizontal and normal to the axial plane of a wedge, or normal to the horizontal axis and inclined to one or both sides. According to Black (1974*a*, 264), the vertical optic axis direction is presumably in response to the vertical temperature gradient, whereas the inclined or horizontal lineations seem to result from rotation or crystal lattices during shearing, but it has also been suggested that freezing of water trickling into a wedge crack may be in response to horizontal temperature gradients (J. R. Mackay, personal communication, 1978). Sampling of surface and buried ice wedges near Fairbanks, Alaska, revealed fabrics similar to those in the Barrow area. The surface wedges had the better developed lineations of the optic axis and the more complicated fabric, whereas recrystallization in the buried wedges caused a more equigranular texture as well as offsetting of silt layers and a tendency for air bubbles to lie along grain boundaries (Black, 1978*b*). From studies of Tyndall figures in ground ice, Péwé (1978) concluded that the optic axis directions in the buried wedge ice he examined were essentially random except along veins and shear zones. In such studies the possibility of recrystallization as a factor needs to be considered if there have been significant temperature changes over a long enough time.

Shape and structure provide unambiguous criteria for well-developed ice wedges, and the ice texture and fabric, and the composition of included gases may assist in more doubtful cases.[6]

7 Thermal regime of permafrost

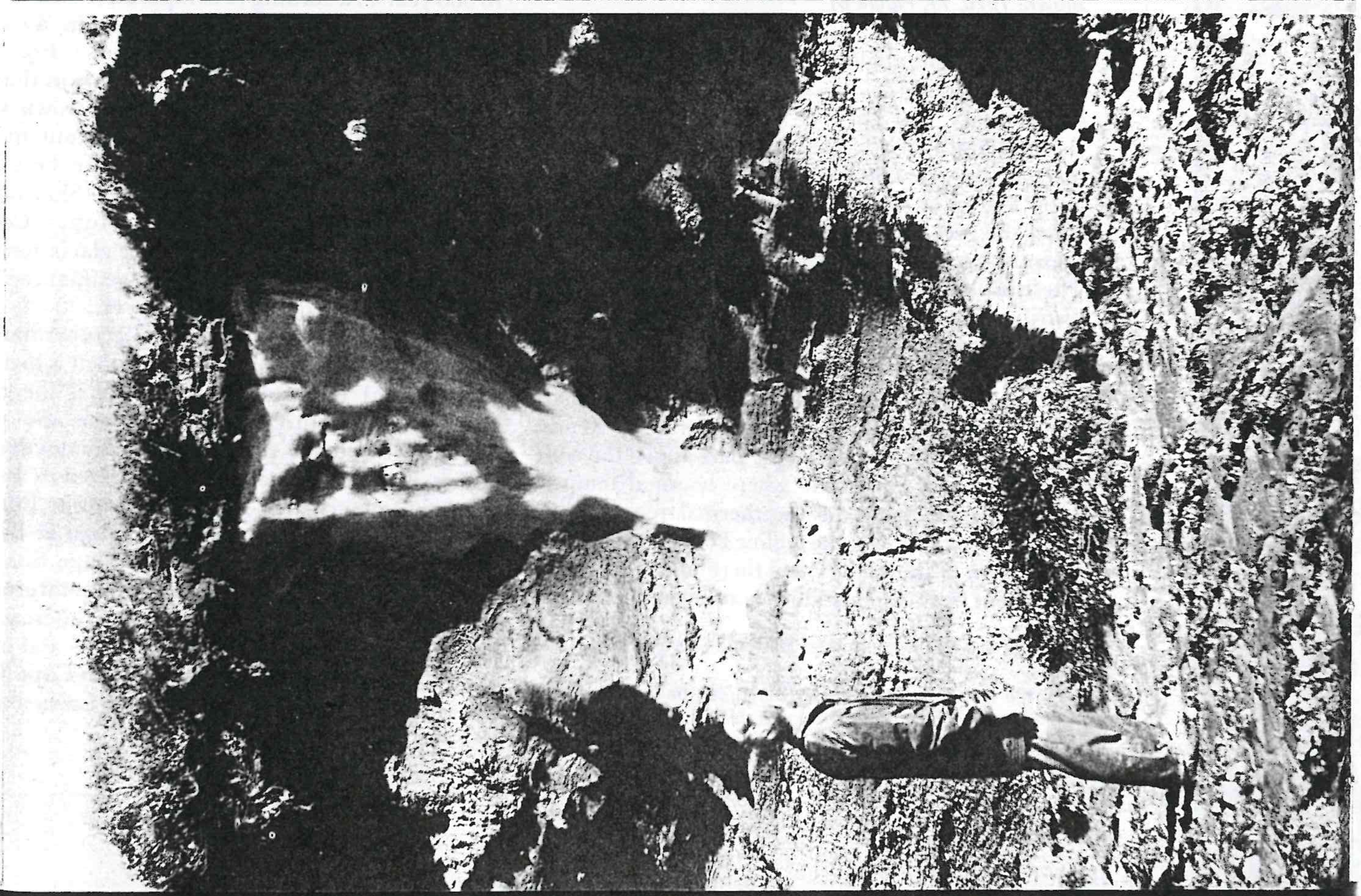
a General The thermal regime of permafrost (Figure 3.27) is dependent on the quantity of heat affecting the permafrost and the overlying layer that freezes and thaws annually – the active layer, described later. The quantity of heat can be expressed as

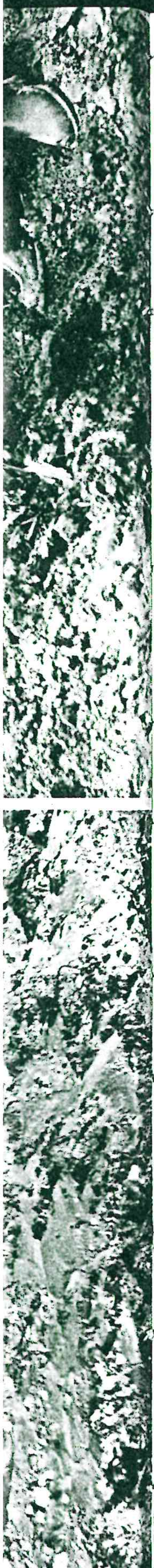
$$Q_h = Ak_h i_g$$

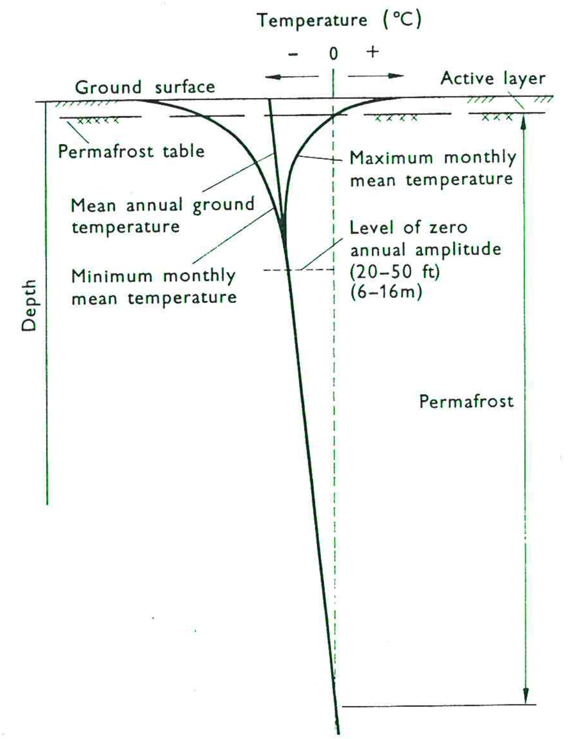
where Q_h = the quantity of heat flowing through

3.24 Massive segregated ice, Stanton, Northwest Territories, Canada. Section is about 7 m high. Photo by J. R. Mackay (*cf. Mackay, 1973a, 224, Figure 1*)

[6] See also Gell (1978*c*), published while the present volume was in press.







3.27 Typical temperature regime in permafrost (*after R. J. E. Brown, 1970, 11, Figure 6*)

an area at right angles to direction of flow per unit of time (cal s^{-1}), A = area (cm^2), k_h = thermal conductivity (cal cm^{-1} s^{-1} $°C^{-1}$), and i_g = geothermal gradient (°C cm^{-1}). The ultimate sources of heat are the earth's interior and the sun (the latter's effect on the earth's surface being some 6000 times the greater – Judge, 1973*a*, 37), and the geothermal gradient expresses their combined effect. These ultimate heat sources are quite stable but the distribution of heat is affected by so many variables that thermal conditions vary widely.

It is convenient to discuss first the relatively stable part of the permafrost, affected by mean annual air temperatures and long-term temperature trends, before discussing the less stable part above the zone of zero annual amplitude where seasonal temperature changes dominate the thermal regime. The many factors affecting the regime here, including the interaction of snow cover, soil thermal conductivity, and latent heat can introduce non-linear effects that can appreciably influence mean annual ground-surface temperatures and the shape of regimen curves such as illustrated in Figure 3.27 (Goodrich, 1978). In many respects it is the regime of the active layer and the upper part of the permafrost that is the most critical for frost-action processes.

b Geothermal heat flow The geothermal heat flow inhibits freezing at the base of stable permafrost, and thaws the base of degrading permafrost. To illustrate (Terzaghi, 1952, 30–1): The rate of geothermal heat flow (q_i) varies somewhat from place to place but approximates 40 cal m^{-2} yr^{-1}. Considering the heat of fusion of ice (80 cal g^{-1} or *c.* 70 cal cm^{-3}), and assuming an ice content of 30 per cent, the heat (q_f) required to melt frozen ground at 0° would be $q_f = 0.3 \times 70$ cal cm^{-3} = 21 cal cm^{-3}. The maximum rate of basal thawing (q_i/q_f) would be

$$q_i/q_f = \frac{40\ \text{cal}}{\text{cm}^2\ \text{yr}} \bigg/ \frac{21\ \text{cal}}{\text{cm}^3} = 2\ \text{cm yr}^{-1}$$

On this basis

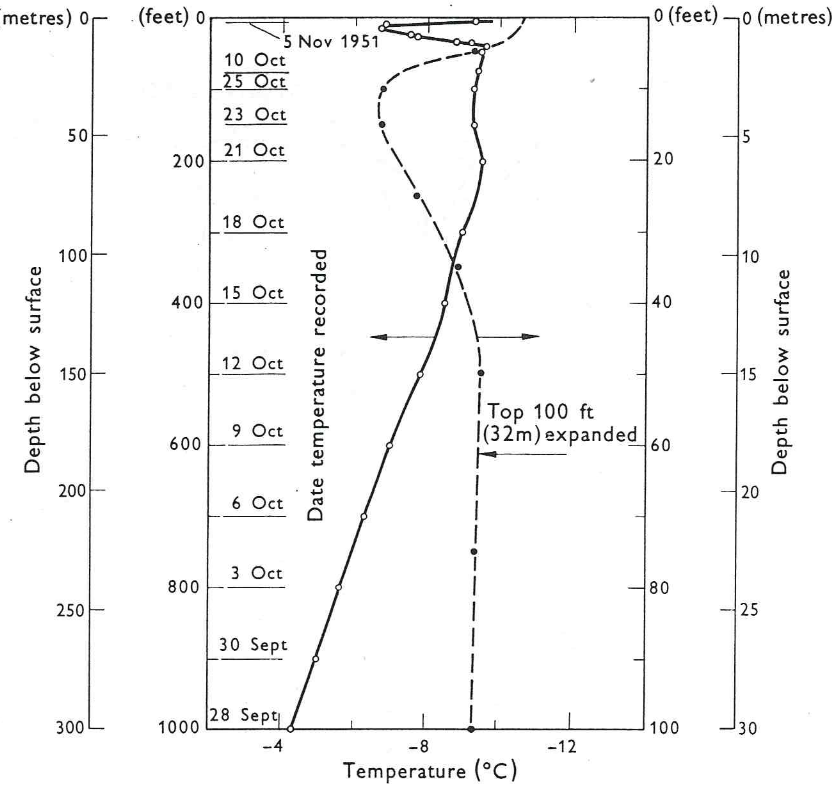
$$t = \frac{H\ (\text{cm})}{2\ (\text{cm yr}^{-1})}$$

where t = time required for thawing from below, and H = thickness of permafrost. If H = 200 m, t = 10 000 years. According to comparative calculations (Terzaghi, 1952, 31), given this situation, a sudden rise in mean annual air temperature of 2° from –1° to 1° would result in less than 100 years in thawing 15 m of permafrost from the surface down while thawing from the base up would amount to only about 2 m. Near-surface heat flow can be significantly perturbed by climatic changes (Sharbatyan, 1974*b*; Sharbatyan and Shumskiy, 1974). Corrections required because of Pleistocene glaciation have been calculated for Canada and generalized in a map by Jessop (1971, 164, Figure 1).

c Geothermal gradient Where the geothermal heat flow is constant, the geothermal gradient is inversely proportional to conductivity. Thus if the mean surface temperature at two places is the same, permafrost extends deeper where conductivity is higher. The effect is startlingly illustrated by Robert Stoneley's observation that permafrost at Prudhoe Bay, Alaska, is 50 per cent thicker than at Barrow and 100 per cent thicker than at Cape Simpson, despite similarity of mean surface temperatures; the increasing thickness was ascribed to increasingly higher proportions of silicious sediments and consequently higher conductivity as between Cape Simpson, Barrow, and Prudhoe Bay (Lachenbruch, 1970*b*, J1).

3.25 Ice wedge near Brakes Bottom, Seward Peninsula, Alaska

3.26 Ice wedge at Shamanskiy Bereg, left bank Lena River, 120 km north of Yakutsk, Yakutia, USSR. Scale given by tape with major intervals in decimetres



3.28 Ground temperatures at Thule, Greenland, September–November 1951 (*after Stearns, 1966, 34, Figure 25*)

Geothermal gradients in permafrost range from about 1° 22 m^{-1} (1°F 40 ft^{-1}) to 1° 60 m^{-1} (1°F 110 ft^{-1}) (Stearns, 1966, 33, Table v). The gradient at Thule, Greenland, which is about 1° 54 m^{-1} (1°F 100 ft^{-1}) is illustrated in Figure 3.28.

The term steep gradient commonly refers to a high °C m^{-1} ratio but gradients are also reported in m °C^{-1}, and depending on the axes chosen in plotting temperature, a high gradient may have a gentle slope in a graph.

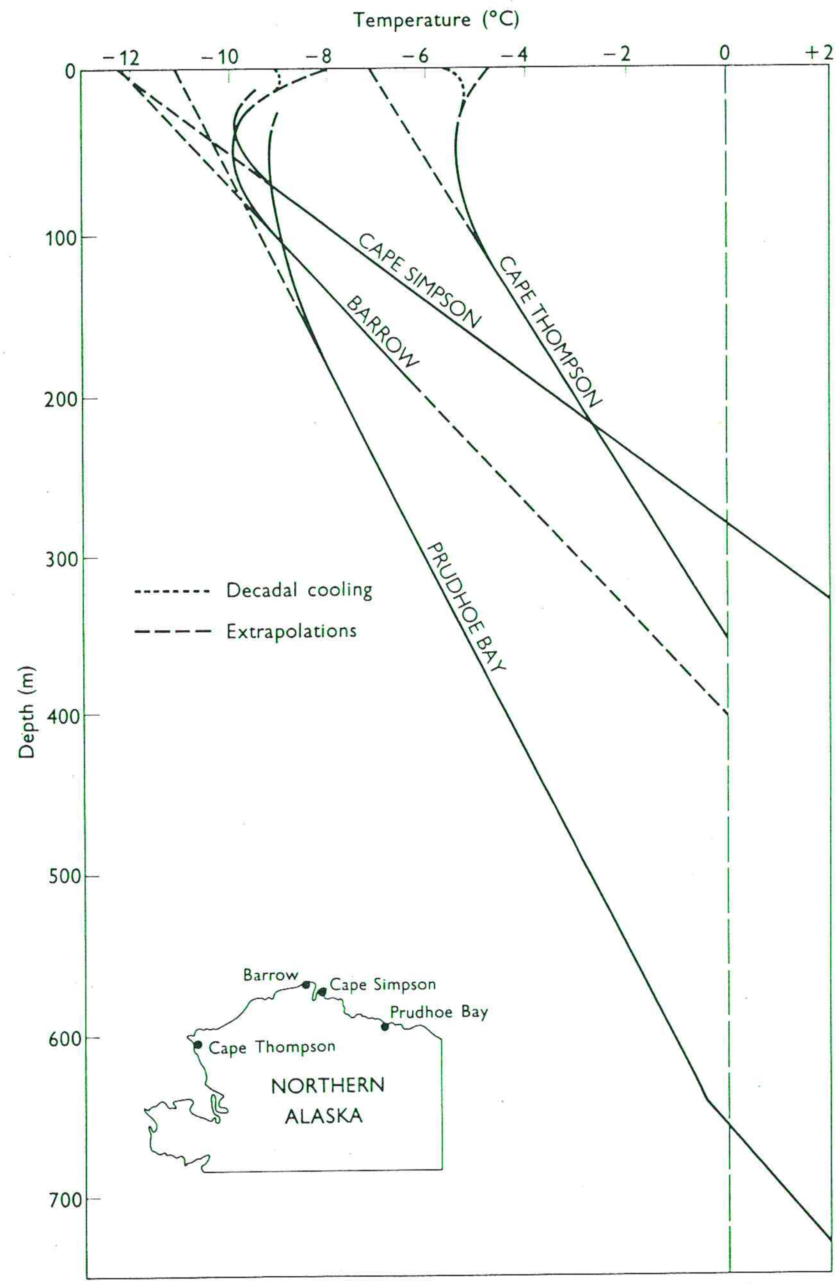
Given sufficient time, the thickness of permafrost is related to temperature and geothermal gradient by the formula (Terzaghi, 1952, 27)

$$H_p = \frac{T}{-i_g}$$

where H_p = thickness (m), T = mean annual ground-surface temperature (°C), and i_g = geothermal gradient (°C m^{-1}).

The geothermal gradient can provide proof of climatic change and indicate its amount rather precisely. For instance, Lachenbruch and Marshall (1969) and Gold and Lachenbruch (1973, 10–15) described temperature profiles from boreholes along the arctic coast of Alaska that show similar pronounced curvatures (Figure 3.29). The curvature is due to a climatic warming following establishment of thermal equilibrium represented by the linear gradient and its extrapolation to the surface. The extrapolated parts of the Barrow and Cape Simpson curves indicate a former mean annual temperature of about −12°. The change to the present mean annual temperature (*c.* −9°) is about +3°. Because repeated observations at the Barrow borehole provided information on the rate of temperature change as a function of depth, Lachenbruch and Marshall were able to calculate that the mean annual *ground–surface* temperature at Barrow must have risen about 4° since the middle of the nineteenth century. They concluded that the one degree difference from the 3° change with respect to the present surface temperature indicated an approximate 1° cooling that had been underway for no longer than a decade or so. Thermal profiles at Cape Simpson and Prudhoe Bay show a similar trend of climatic change.

ɪo–ɪ5)
; along
.r pro-
·vature
.hment
linear
e. The
mpson
:rature
. mean
ecause
le pro-
:hange
arshall
ground–
ı about
·. They
›m the
surface
:ooling
:han a
on and
hange.



3.29 Geothermal gradients in four boreholes in Arctic Alaska. See text (*after Gold and Lachenbruch, 1973, 10, Figure 1, and Lachenbruch and Marshall, 1969, 302, Figure 2*)

Although thermal profiles from boreholes in Spitsbergen also reveal similar curvatures at depths roughly comparable to those in Alaska, the tentative inference was a warm period between 1920 and 1960 rather than one starting earlier (Liestøl, 1977, 11, Figure 3; 12). For the northern USSR, Balobayev (1978) calculated that temperature profiles in disequilibrium permafrost, thawing from below, showed that temperatures 20 000 years ago were 10°–13° colder than now, with permafrost thicknesses reaching 600–800 m. This calculation rests on a number of assumptions.

d Depth of zero annual amplitude The depth of zero annual amplitude – the depth to which seasonal change of temperature can propagate into the ground, when no phase change is involved – increases in direct proportion according to the equation (Terzaghi, 1952, 22)

$$z_1 = \sqrt{12at_1}$$

where z_1 = depth, a = diffusivity, and t_1 = time. Diffusivity (cf. Terzaghi, 1952, 11) is defined as

$$a = \frac{k_h}{C_h \times W_t}$$

where a = diffusivity (cm^2 s^{-1}), k_h = thermal conductivity (cal cm^{-1} s^{-1} °C^{-1}), C_h = heat capacity (cal g^{-1} °C^{-1}), and W_t = unit weight (g cm^{-3}).

Temperatures at the depth of zero annual amplitude can be calculated according to the equation (cf. Grave, 1967)

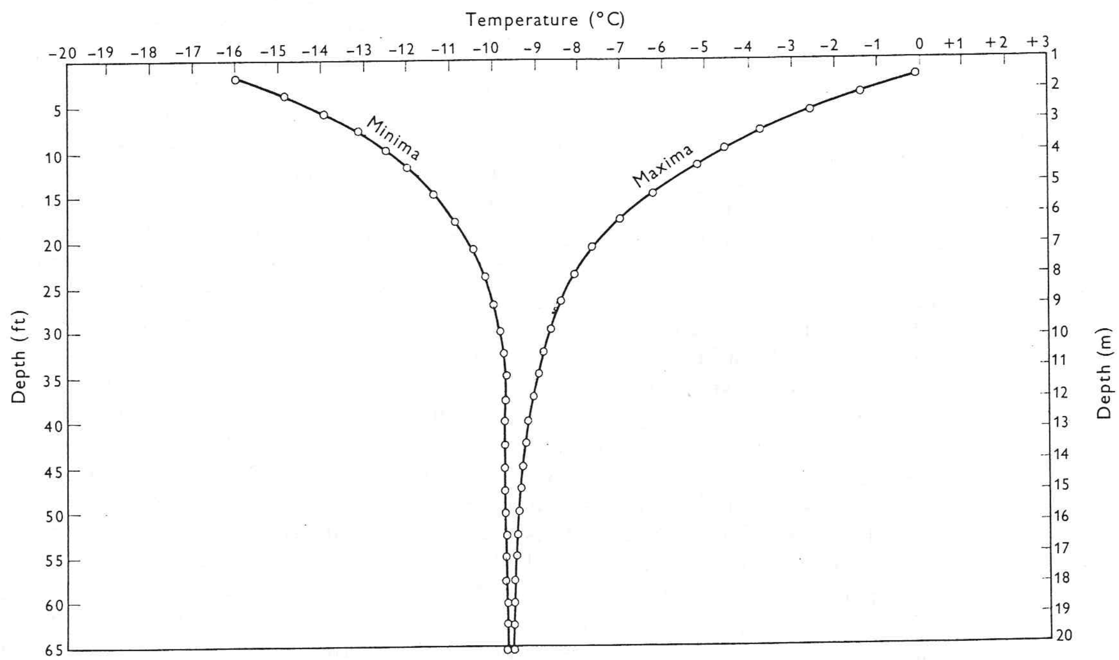
$$t_{TA} = t_b + \frac{A_{cm}}{2}\left(1 - \frac{1}{f}\right)$$

where t_{TA} = temperature (°C) at depth of zero annual amplitude ('bottom of thermoactive layer'), t_b = mean annual air temperature, A_{cm} = annual amplitude of mean monthly air temperatures, and

$$f = e^{+z\sqrt{\frac{\pi}{kT}}}$$

where z = thickness of snow cover in metres,[7] k = thermal diffusivity[7] of snow (in m^2 h^{-1}), and T = period of oscillations equal to one year measured in hours.

The temperature at the depth of zero annual amplitude approximates the mean annual ground-surface temperature (cf. National Academy of Sciences, 1974, 40), and the mean annual temperature at the top of permafrost. Also it is reported



3.30 Amplitude of seasonal temperature changes at Barrow, Alaska, 1949–50 (*after MacCarthy, 1952, 591, Figure 3*)

[7] The metres and the diffusivity (instead of conductivity) are emendations to Grave (1967) (N. A. Grave, personal communication, 1977).

to be generally about 3° (6°F) warmer than the mean annual air temperature (R. J. E. Brown, 1967*a*; 1970, 20). For 38 sites in Canada (of which 20 involved permafrost), the average was 3.3° with a standard deviation of 1.5° (Judge, 1973*b*, 124). However, there is considerable regional variation (1.5°–9.3°) in figures from various parts of the world as compiled by Grave (1967, 1341, Table 1).

The convergence of minimum and maximum temperatures to the level of zero annual amplitude is illustrated in Figure 3.30.

Above the depth of zero annual amplitude, the permafrost is subjected to seasonal fluctuations of temperature. Nevertheless, a mean annual surface temperature can be deduced by upward extrapolation of the geothermal gradient from the level of zero annual amplitude, provided the measured gradient has achieved equilibrium and there are no recent climatic changes. However, to depths of 15 m at any one general location, variations in terrain, snowcover, and vegetation can lead to variations of 2° in mean annual ground temperature (R. J. E. Brown, 1973*b*, 31). As previously noted, it is this upper part of the permafrost and the overlying active layer that are especially critical for frost-action processes and effects.

8 Aggradation and degradation of permafrost

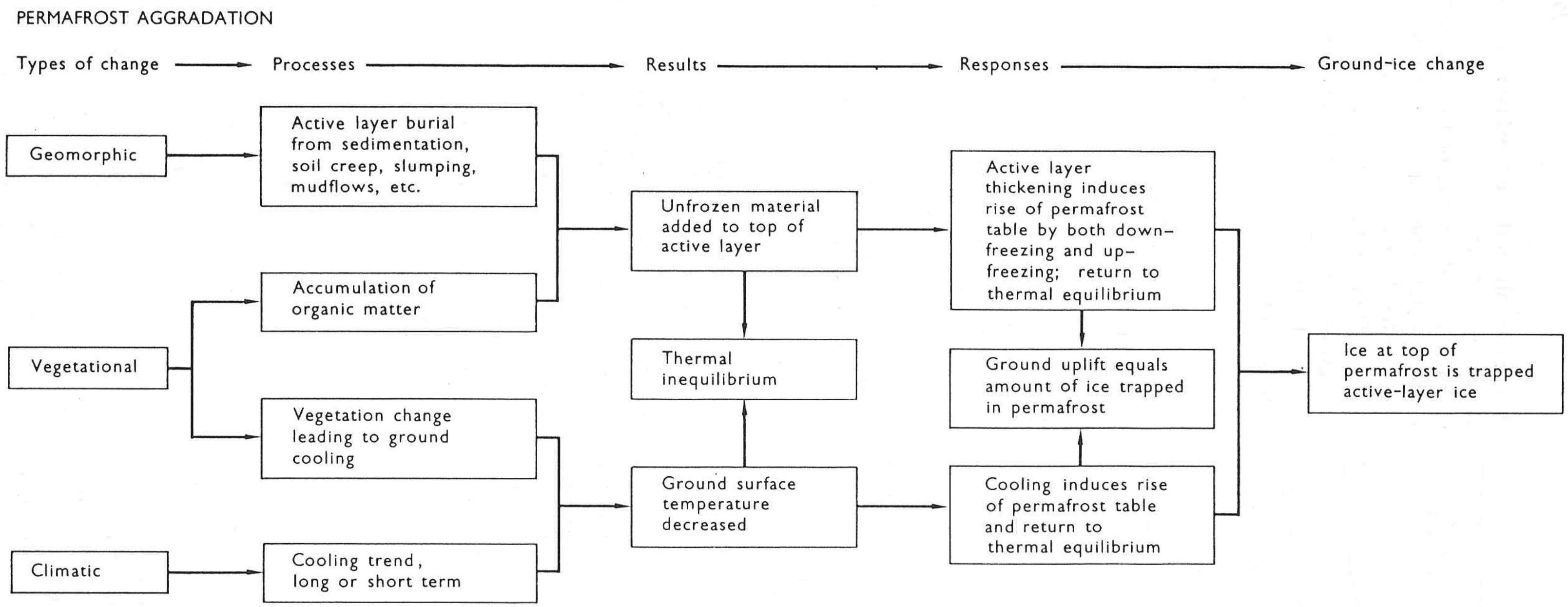
a General There are numerous ways in which permafrost can build up (aggrade), and thaw (degrade), all of which are controlled by the thermal regime. Changes that can lead to aggradation or degradation are climatic, geomorphic, and vegetational, and the ways in which they can affect permafrost are shown in Figures 3.31–3.32, following Mackay (1971*a*). In addition to the geomorphic changes specifically listed in Figures 3.31–3.32, changes induced by shifting shorelines can be very important. Permafrost on the land thins seaward and can become offshore permafrost (Lachenbruch, 1957). Many lakes and rivers that are too deep to freeze to the bottom have extensive taliks beneath them, some piercing through the entire thickness of permafrost (Lachenbruch *et al.*, 1962, 795–9; M. W. Smith, 1976). Consequently, permafrost thickness commonly increases with distance from the water bodies until their influence disappears. In all these situations a shift in the shoreline of the ocean, lake, or river would introduce a thermal anomaly affecting the distribution of permafrost. Detailed studies pertaining to rivers include those by J. R. Williams (1970, 19–20; 26–52) in Alaska, and by M. W. Smith (1976) in the Mackenzie Delta in northern Canada.

b Syngenetic and epigenetic permafrost Aggradational permafrost may be syngenetic or epigenetic depending on whether concurrent sedimentation is present or absent (Baranov and Kudryavtsev, 1966, 100–1). Thus, except for the ice, the rocks and soil of epigenetic permafrost are older than the freezing, whereas those of syngenetic permafrost must be about the same age as the freezing. Epigenetic permafrost may be characterized by a concentration of well-developed segregated ice in the upper horizons because of the water that is drawn upward from considerable depth during freezing, as described under Freezing process in the chapter on General frost-action processes. In syngenetic permafrost the concurrent sedimentation and freezing favour accretion of thin layers near the top of the rising permafrost table. Although this is commonly an ice-rich zone (Mackay, 1972*a*, 10; Pissart, 1975), a large amount of water and generally lower freezing rates in developing epigenetic permafrost permit much thicker layers of segregated ice.

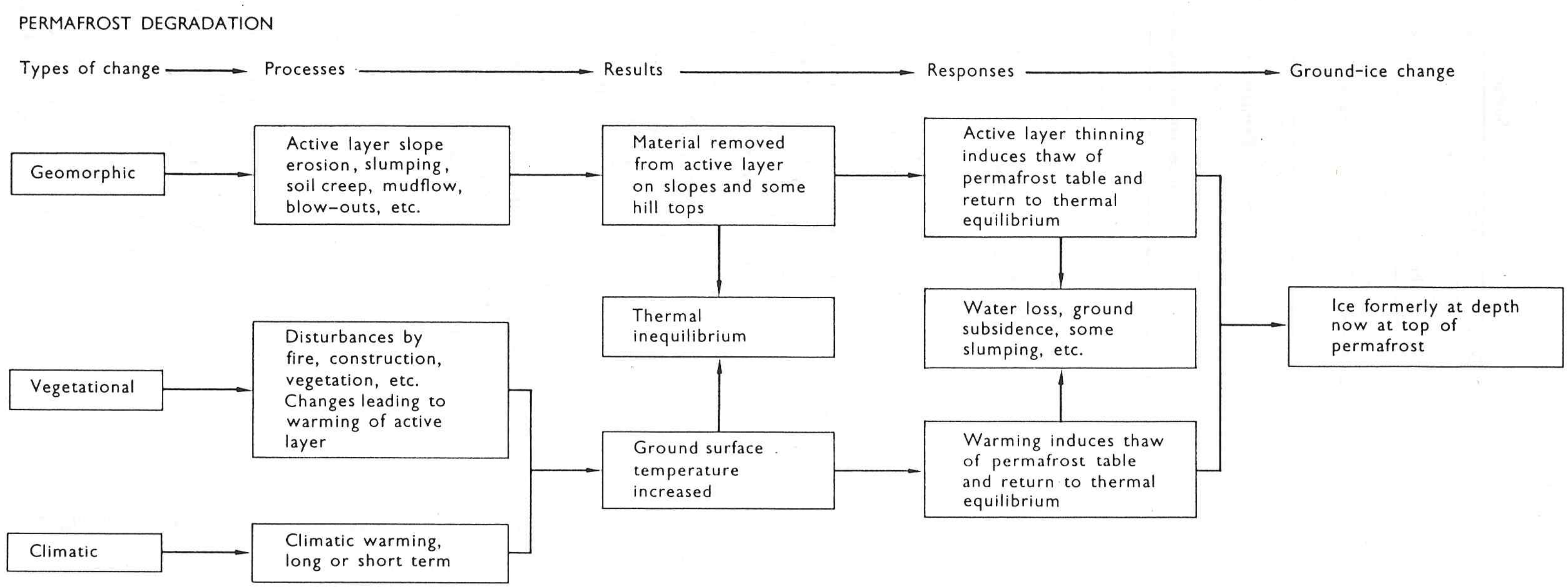
c Thermokarst Thermokarst comprises '... karst-like topographic features produced by the melting of ground-ice and the subsequent settling or caving of the ground' (S. W. Muller, 1947, 223). The term was introduced by Ermolaev in 1932 (French, 1976*a*, 104). Cryokarst (Kryokarst – cf. Schunke, 1977*c*, 45, footnote 1) is synonymous but less common. Thermokarst records degradation of permafrost. The disturbance of the thermal regime may be local such as disruption of insulating vegetation, or more general such as climatic change. There are many cases of well-developed thermokarst features, due to local influences only, in the continuous permafrost zone. Thermokarst resulting from climatic change is more likely to be associated with the discontinuous zone where the thermal balance of permafrost is more delicate, with changes being apparent within 20 years (Thie, 1974); however, by the same token, local disturbances in this zone are also more likely to cause thawing of permafrost. As discussed later the origin of thermokarst must be interpreted with care.

d Taliks A talik is '... a layer of unfrozen ground between the seasonal frozen ground (active layer) and the permafrost. Also applies to an unfrozen layer within the permafrost as well as to the unfrozen ground beneath the permafrost' (S. W. Muller, 1947, 223).[8]

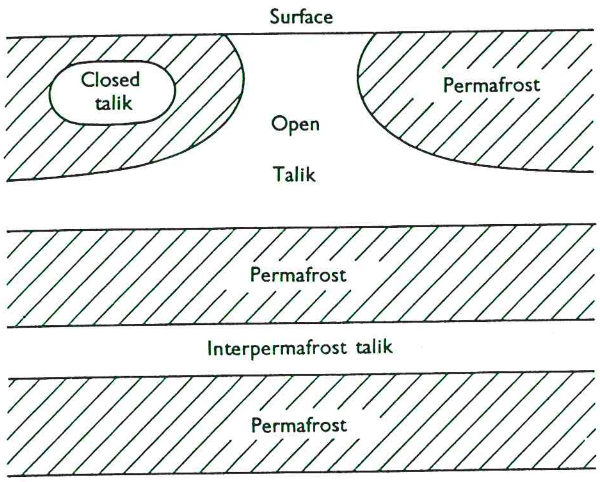
[8] This definition is subject to the criticism cited in the footnote to the definition of permafrost table in the following section.



3.31 Permafrost aggradation (*after Mackay, 1971a, 29, Figure 1*)



3.32 Permafrost degradation (*after Mackay, 1971a, 30, Figure 2*)



3.33 Diagram of open, closed, and interpermafrost talik (*courtesy, Jaromír Demek, 1969*)

Soviet investigators recognize three kinds of talik but contrary to Muller they do not commonly designate the ground below the base of permafrost as talik. The three kinds are open (skvoznoy), closed (zamknuty), and interpermafrost (mezhmerzlotny) talik (Figure 3.33). Taliks within permafrost, even in the continuous zone, are a common feature of the thermal regime and may or may not indicate degradation of permafrost. Some are due to local heat sources, including groundwater circulation, and may be associated with quasi-equilibrium conditions. Other taliks result from a marked change of thermal regime and thus indicate permafrost degradation, in which case thermokarst is likely to be present. Groundwater flowing in open taliks in alluvium beneath frozen rivers can provide the major water supply in some areas of continuous permafrost (cf. Sherman, 1973).

9 Thermal regime of active layer

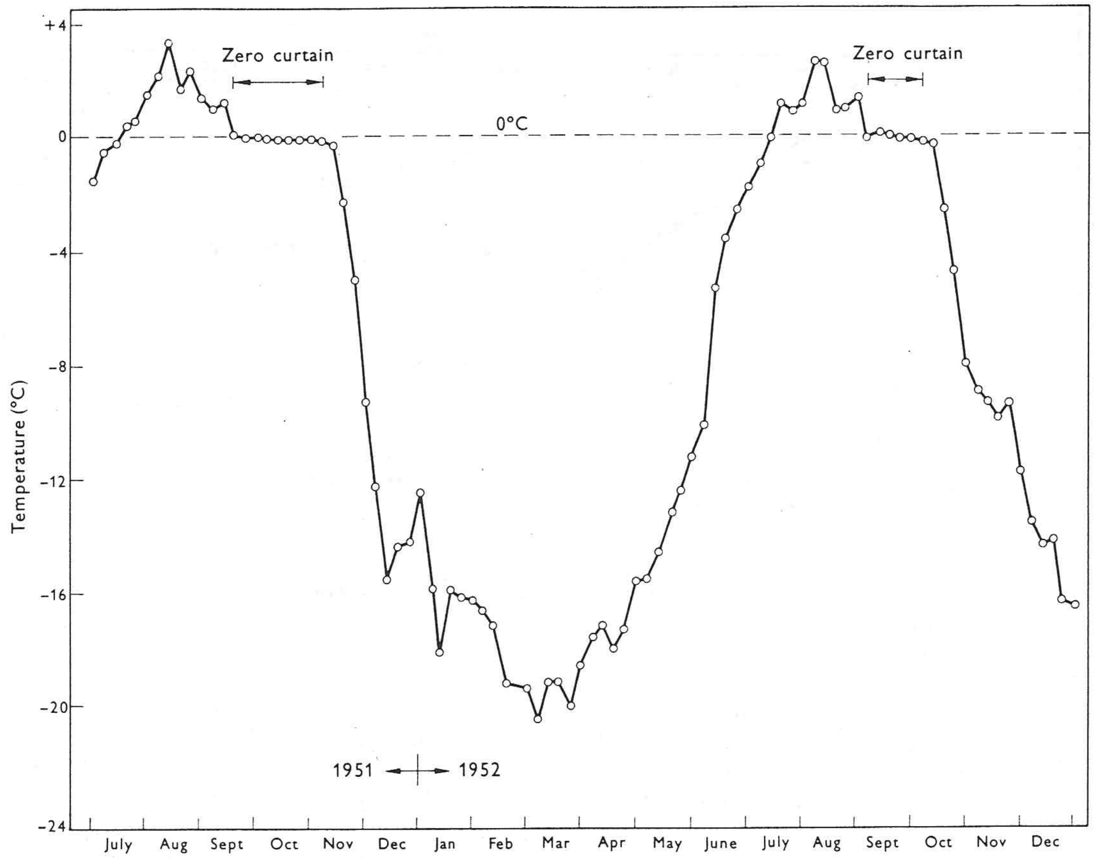
a General The active layer is the 'layer of ground above the permafrost which thaws in the summer and freezes again in the winter' (S. W. Muller, 1947, 213). Association with permafrost is inherent in this definition but some investigators (cf. Péwé, Church, and Andresen, 1969, 7) also apply the term to the layer of ground that freezes and thaws in a non-permafrost environment and a few others have used it in a still different sense (cf. Corte, 1969*a*, 130). However, the term is widely used as originally defined, and the present writer follows this usage. A monograph on the active layer has been authored by Vtyurina (1974).

The thermal regime of the active layer is controlled by the same environmental factors that control the development of frozen ground as discussed earlier. They include the basic factors of climate, topography, material, and time, and the dependent factors of snow and ice cover, moisture, and vegetation. These interact in a complex fashion and determine such regime aspects of the active layer as its thickness, depth to the underlying permafrost table, upward freezing from the permafrost table, the zero-curtain effect, and the structure of the active layer. Environmental factors can cause considerable variation in the thickness of the active layer. A general range would be 15 cm to 5 m (Ferrians and Hobson, 1973, 479) but a range of 3–12 m has been reported from an area as small as 1 km^2, the thickest occurrences being in valley sites where thawing is accelerated by movement of groundwater (Nicholson and Thom, 1973, 161–2). Thickness variations within short distances are common across some beaches (Owens and Harper, 1977). Normally, the active layer is thicker in sands and gravels than in fine-grained soils, other conditions being equal. Nevertheless, the reverse has been reported where the coarse material has permitted meltwater to penetrate rapidly to depth and refreeze, thereby building up an icy layer that delays thawing because of its latent heat of fusion (Semmel, 1969, 42, 43 – footnote 10). The thickness of the active layer in the same spot can vary appreciably from year to year – by up to 71 cm in cases reported by R. J. E. Brown (1978, Table III). A year-to-year variation amounting to 25 per cent of the total depth of the active layer was reported as normal by Nicholson (1978, 430), based on extensive observations in an area of discontinuous permafrost.

Mathematical models of thermal regimes, based on quantitative parameters representative of the controlling factors, are becoming increasingly sophisticated. Simulations derived from a model, utilizing values for a test site at Barrow, Alaska, have shown a remarkable similarity with the observed regime and help to evaluate the relative importance of selected factors (Nakano and Brown, 1972).

b Permafrost table The permafrost table is the 'more or less irregular surface which represents the upper limit of permafrost' (S. W. Muller, 1947, 219). As such it is the dividing surface between the permafrost and the active layer.

Any frozen surface in the active layer as it thaws downward towards the permafrost table is called a



3.34 Time-temperature curve at a depth of 25 cm (10 in) near Barrow, Alaska (*after M. Brewer, 1958, 22, Figure 4*)

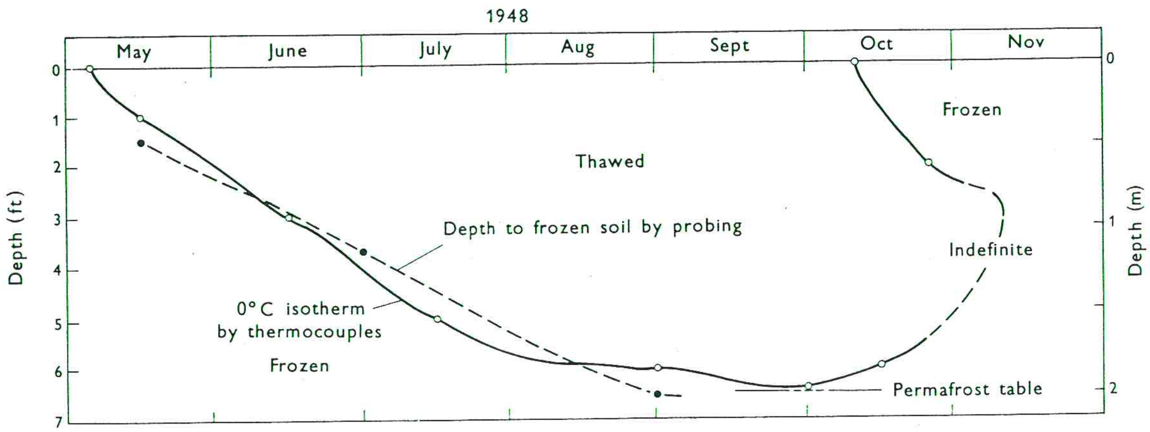
frost table and must not be confused with permafrost or the permafrost table. The term frost table is also applicable to a frozen surface in seasonally frozen ground in a non-permafrost environment.[9]

Other conditions remaining constant, the thermal regime controls the depth to the permafrost table which in the zone of continuous permafrost usually coincides with the thickness of the active layer. However, the position of the permafrost table represents an average condition, and because of short-term fluctuations of regime, such as an unusually cold winter, warm summer, or thick snow cover, the permafrost table need not coincide with the maximum depth of thaw in a given year. Consequently there may be short-term anomalies that can cause a temporary talik at the average position, or a pereletok, defined as '. . . a frozen layer at the base of the active layer which remains unthawed for one or two summers (Russian term meaning "survives over the summer"). Pereletok may easily be mistaken for permafrost' (S. W. Muller, 1947, 219). R. J. E. Brown and Péwé (1973, 72) regarded the distinction as artificial and included pereletok with permafrost even though most definitions of permafrost would exclude a temporarily frozen layer that survived for one summer only.

The permafrost table is a crucial boundary in several ways. It is commonly a rigid surface capable of bearing considerable loads without deforming. It also prevents moisture from seeping downward and it thereby favours a high moisture environment and the formation of ice lenses.

c Zero curtain The zero curtain can be defined as the zone immediately above the permafrost table '. . . where zero temperature (0°C) lasts a considerable period of time (as long as 115 days a year) during

[9] Frost table is synonymous with tjaele, a frequently used term in the European literature. However, as pointed out by Bryan (1951), tjaele has also been erroneously applied to permafrost.



3.35 Freeze-thaw cycle in active layer at Fairbanks, Alaska. Note upward freezing from permafrost table in October (*after Stearns, 1966, 24, Figure 13*)

freezing and thawing of overlying ground' S. W. Muller, 1947, 224).[10]

The zero curtain is caused by the latent heat of fusion of ice (80 cal g^{-1} water), which delays freezing and thawing. The higher the moisture content near the permafrost table, the greater the delay. The zero curtain (Figure 3.34) plays a major role in some frost-action processes and is a notable aspect of the thermal regime of the active layer (cf. Nakano and Brown, 1972, 31–6).

d Upward freezing Because of the thermal regime of permafrost, freezing of the active layer may proceed upward from the permafrost table as well as downward from the ground surface (Black, 1974*a*, 253, Figure 4; Drew *et al.*, 1958; Mackay, 1973*d*; 1974*d*; cf. Meinardus, 1930, 40; Schmertmann and Taylor, 1965, 50; US Army Corps of Engineers, *in* Washburn, 1956*b*, footnote 19, 842; Viereck, 1973, 61). Figures 3.35–3.36 illustrate upward freezing from the permafrost table.[11] According to Mackay's (1973*d*, 392) observations in the Mackenzie Delta area of northern Canada, the greatest upfreezing is where the active layer is thickest; the amount of upfreezing found by Mackay ranged from 2 to 13 cm. In many places upward freezing is shown by bubble patterns in the ice of the active layer (Gell, 1974*a*), and by distinctive ice layering parallel to the bottom of the active layer (Vtyurina, 1974, 33–4), which record the upward-moving as well as the downward-moving freezing front.

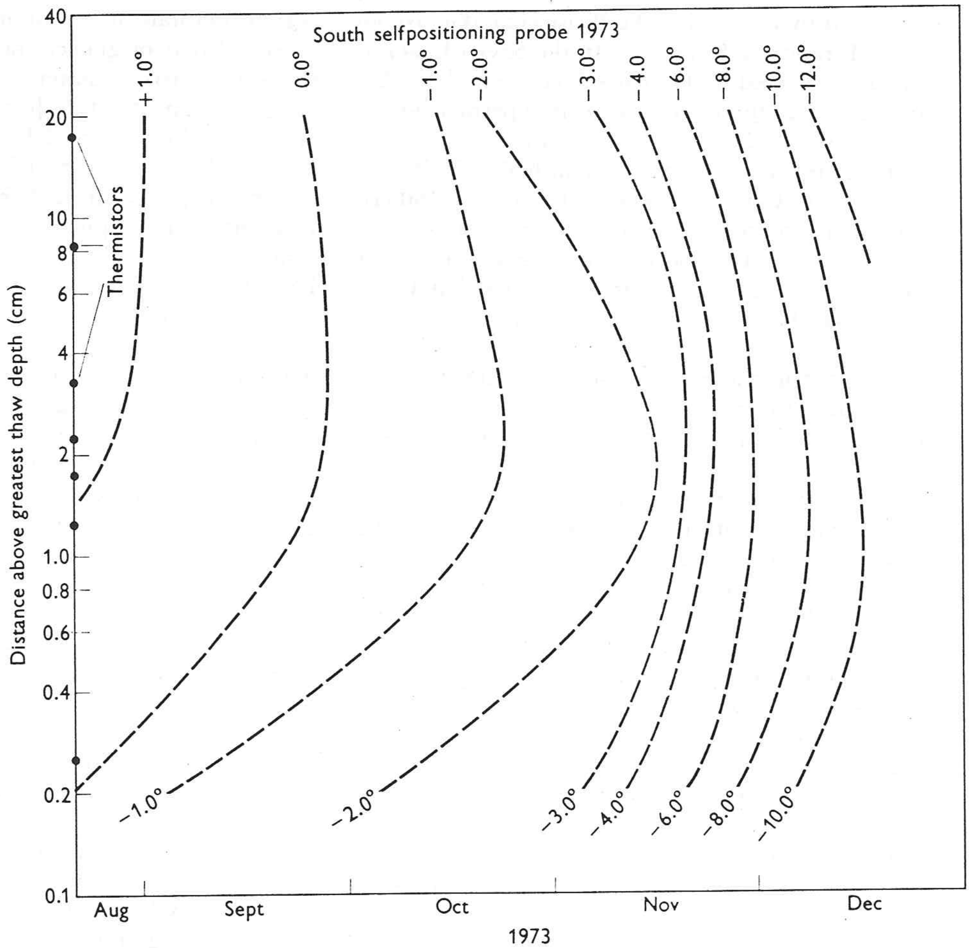
Upward freezing of a frost table in a non-permafrost environment is unlikely. Details as to the prevalence and significance of the process in a permafrost environment are meagre. However, it is probably an important factor in some frost-action processes as discussed later.

e Effect of structure on thermal regime The thermal regime of frozen ground, whether seasonally frozen or permafrost, is strongly affected by structure of permafrost, especially the amount and distribution of the ice. As noted above in discussing the zero curtain, heat flow is influenced by the latent heat of fusion of ice; it is also influenced by the attitude of ice masses (Bakulin, Savel'yev, and Zhukov, 1957, 73–4; 1972, 3). Thus ice lenses and massive ice beds, being predominantly horizontal, delay thawing of underlying ground compared with adjacent ground lacking such ice masses. On the other hand, the predominantly vertical attitude of ice veins and ice wedges promotes vertical heat flow because of the greater thermal conductivity of ice than soil.

f Effect of thermal regime on structure Inversely, the thermal regime also influences the structure of frozen ground. For one thing, there are generally smaller ice masses in the active layer than in permafrost. Also the active layer tends to have a threefold

[10] In this definition, the phrase 'zone immediately above the permafrost table' has been substituted for Muller's (1947, 224) phrase 'layer of ground between active layer and permafrost', because such a layer is inconsistent with Muller's definitions of permafrost table and active layer. The same caveat can be applied to his definition of talik.

[11] As illustrated by Figure 3.35, the depth to frozen soil by probing can differ from the position of the 0° isotherm, especially in fine-grained soil in which probing may penetrate below the isotherm where the soil is not hard frozen due to unfrozen water, discussed later. A depth discrepancy of several decimetres is quite expectable (Mackay, 1977*c*).



3.36 Upfreezing from permafrost table at Garry Island, Northwest Territories, Canada. Compare isotherm trends with similar trend of 0° isotherm for freeze-up period in Figure 3.35 (*after Mackay, 1974d, 251, Figure 3*)

structure consisting of (1) an upper zone with small ice inclusions; (2) a middle zone that is desiccated because of withdrawal of moisture towards freezing fronts, one moving down from the surface, another up from the permafrost table (upward freezing); and (3) a lower zone with a mixture of thin and thicker ice lenses immediately above the permafrost table (Zhestkova *et al.*, 1961, 45; 1969, 5–6).

10 Origin of permafrost

Most permafrost may have originated during the Pleistocene. The evidence for this is that (1) Tissues of woolly mammoths (*Mamuthus primigenius*) and other Pleistocene animals have been preserved in permafrost, indicating presence of permafrost at time of death (Gerasimov and Markov, 1968, 11; Vereshchagin, 1974), and there is no reliable evidence that the woolly mammoth survived into the Holocene (Farrand, 1961), although it appears to have lived as late as the Alleröd interstadial in northern Siberia (Heintz and Garutt, 1965, 76–7). (2) The upper boundary of some permafrost is considerably deeper than the present depth of winter freezing (Gerasimov and Markov, 1968, 12). This evidence merely indicates that some permafrost is not of present-day origin. (3) In places the temperature of permafrost decreases with depth, indicating residual cold (Gerasimov and Markov, 1968, 12). Again, this evidence does not necessarily prove that the residual cold demonstrates Pleistocene permafrost. (4) The thickest permafrost is commonly in areas that remained unglaciated, and therefore not insulated by ice, during the Pleistocene (Gerasimov and Markov, 1968, 14). (5) In the Soviet Union in the area covered by the early Pleistocene Kara Sea transgression, there are two layers of permafrost, the lower one dating from before the transgression (Grave, 1968*a*, 52–3; 1968*b*, i–ii, 6–8). (6) Permafrost in the Central Yakutian Lowland of Siberia has been reported as existing continuously since at least Middle Pleistocene time (Katasonov and Ivanov, 1973, 10–11) and in northern Yakutia since the Lower Pleistocene (Katasonov, 1977). 'According to present-day concepts, the formation of the permafrost in the Soviet Northeast dates from the first half of the Pleistocene and has continued without interruption right up to the present time' (Grigor'yev, 1978, 166). The present writer is not aware of the detailed evidence for continuous permafrost, although the presence of permafrost in the Soviet Union at least as far back as the early Pleistocene is indicated by syngenetic ice-wedge casts in the Olerskaya Formation of the Kolyma Lowland (Arkhangelov and Sher, 1973; 1978). (7) Some permafrost in the Mackenzie Delta area of northern Canada is believed to be of early Wisconsin age or older, since it is glacially deformed but occurs where there is no evidence of glaciation during the last 40 000 ^{14}C years (Mackay, Rampton, and Fyles, 1972). The presence in these sediments, beneath till, of ice-wedge casts with relict wedge ice still present at the bottom is confirming evidence (Mackay, 1976*a*).

In formerly glaciated areas, much permafrost is post-glacial. The argument for this is that: (1) Permafrost would have thawed beneath the thickest Pleistocene ice sheets (Büdel, 1959, 305). The argument is supported by the finding of water beneath the Antarctic Ice Sheet (Gow, Ueda, and Garfield, 1968) but it need not apply to all parts of an ice sheet (Grave, 1968*a*, 51–2; 1968*b*, 6). (2) In polar regions permafrost is forming today in many areas where retreat of glaciers or recent emergence has exposed unfrozen material. Also, taken together, evidence from Alaska (Jerry Brown, 1965, 39–47; Hopkins, MacNeil, and Leopold, 1960, 55; Péwé, 1958; Sellmann, 1967, 16–19) and from Siberia (Jahn, 1975, 29 – citing A. Kudryavtsev) argues for widespread development of postglacial permafrost, some (but not all) of it probably following the Hypsithermal warm interval dated at between about 2500 and 9000 years BP (Flint, 1971, 524).

Clearly the freezing and thawing of permafrost at depth can be consequent on past as well as present environmental conditions. Ground temperatures change particularly slowly where the latent heat of fusion of ice is involved, and permafrost boundaries are correspondingly slow to shift as emphasized by Jessop (1973). Nevertheless, ground-temperature profiles and meteorological records suggest that the southern boundary of continuous permafrost in Canada's Mackenzie Valley moved north as much as 320 km (200 mi) from the late 1800s to the 1940s following an approximate 3° rise in mean annual temperature, then began shifting south in response to a temperature lowering of about 1° (Mackay, 1975*e*). A recent increase in permafrost has also been reported in the USSR (Belopukhova, 1973; 1978).

In summary, (1) the upper part of most continuous permafrost is in balance with the present climate but the base may be slowly thawing, stable, or aggrading, depending on the past and present climate and the geothermal heat flow, (2) most discontinuous permafrost, both at its surface and base, is either out of balance with the present climate or in such delicate equilibrium that the slightest climatic or surface change will have drastic disequilibrium effects, although a number of years may be required for major changes in permafrost boundaries.