1

a steady liquid flux is to be expected, the ice phase may also be conducive to the transport of water by a regelation process. Pore ice appears to be immobile, because its boundaries are stationary, but in reality it may be moving. Molecules of unfrozen water flow through the frozen, porous medium and freeze on to one side of a lens. The latent heat liberated induces thawing on the other side, with a subsequent slow displacement of the ice body towards the melting face. It follows that the presence of ice lenses may not be much of a barrier to flow, even though no direct path exists (Burt and Williams, 1976).

Several procedures have been suggested for the measurement of the apparent hydraulic conductivity of frozen soils. Furthermore, there is a fair degree of agreement of results from the various approaches. Harlan (1973) has advocated the use of hydraulic conductivity values obtained for unsaturated, unfrozen soils, on the assumption that the permeability at a given liquid water content is the same in the frozen as in the partially saturated state. Unfortunately, this analogy breaks down somewhat, as soon as ice lensing occurs in the frozen soil. The amount of unfrozen water present at any given temperature is independent of the total moisture content, but an increase in the latter gives additional ice, which may significantly alter the conditions of flow.

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TEMPERATURE-INDUCED WATER MIGRATION
IN SATURATED FROZEN SOILS

BY

Edmund Perfect

A thesis submitted to the Faculty of Graduate
Studies in partial fulfilment of the requirements
for the degree of Master of Arts.

Department of Geography
Carleton University
Ottawa, Ontario
6th April, 1980.

The undersigned recommend to the Faculty of
Graduate Studies acceptance of the thesis

"TEMPERATURE-INDUCED WATER MIGRATION

IN SATURATED FROZEN SOILS"

submitted by Edmund Perfect, B.Sc., in partial
fulfilment of the degree of Master of Arts.

Thesis Supervisor, Department of
Geography

Supervisor of Graduate Studies,
Department of Geography

Chairman, Department of Geography

Carleton University

April 29, 1980

ABSTRACT

An apparatus has been developed which allows, for the first time, direct measurement of thermally induced water migration in frozen soils. The experimental configuration is similar to that used by Burt and Williams (1976), except the hydraulic gradient is produced by a temperature differential. The frozen soil sample and reservoirs are sandwiched between two cooling plates, controlled by a thermoelectric cooling control system. The water in the reservoirs remains unfrozen because it contains dissolved lactose. When a linear temperature gradient is established across the system, flow normally takes place in the direction of decreasing temperature. Water passes from the 'warm' reservoir into the frozen soil, and there is an outflow into the second, colder reservoir. Flow rates are seen to vary with the magnitude of the temperature (potential) gradient and the permeability of the frozen sample. The results are evaluated in terms of the current theoretical understanding of frozen soils, and the properties of the experiment itself.

PREFACE

This study is concerned with temperature-induced water migrations in frozen soils. The availability of laboratory equipment and expertise in the Geotechnical Science Laboratories of the Department of Geography, Carleton University, made the research possible. Financial assistance was provided by the Earth Physics Branch, Department of Energy, Mines and Resources, Canada.

I would like to express my gratitude for the guidance, criticism and enthusiasm provided by Peter Williams, who introduced me to the field of Geotechnical Science, especially the fundamental effects of freezing and thawing. Thanks are also extended to Ken Torrance and Mike Smith for their suggestions, and constructive criticisms of the thesis draft.

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CHAPTER 1

WATER MIGRATION IN FROZEN SOILS:

THEORETICAL CONSIDERATIONS

1.1 Introduction

The presence of unfrozen water along with ice in frozen ground has been recognised since the early sixties, when calorimetry showed that less heat was involved in freezing and thawing of samples than was expected (eq. Nersesova and Tsyтовich, 1966). Other studies (eg. Koopmans and Miller, 1966) showed the water to have a potential, or equally a Gibbs free energy, relative to pure, bulk water, that was lower by an amount increasing as the temperature decreased. Concurrently, investigations of frost heave were pursued in various laboratories, but without regard for the importance of the characteristics described above, of soils at temperatures down to -1°C or -2°C .

It was not until 1974 that a deliberate attempt was made to determine the hydraulic conductivity of frozen soils, although Hoekstra (1966, 1969) had concluded that some such permeability must exist close to 0°C . Hoekstra had observed the continued growth of an ice lens after it was surrounded by frozen soil. An experimental apparatus was constructed by Burt (1974), which permitted direct measurement of a

hydraulic conductivity coefficient for frozen soils. These observations raised the possibility of design problems for refrigerated pipelines through permafrost. In essence it appeared that the frost heave pressures generated would be much greater than envisaged.

The significance in this context, of the hydraulic conductivity of frozen ground lies in the fact that water may travel into frozen soil, to freeze at temperatures substantially below 0°C . It moves to such regions under the influence of temperature induced potential gradients. These gradients also determine the magnitude of the frost heave pressures generated when the water freezes. The theory is reviewed in this chapter. The measurements of hydraulic conductivity made by Burt (1974) have been quite widely discussed. Whilst they raise very serious questions concerning possible long term heave of chilled gas pipelines, they are not acceptable as providing all the answers. There is an urgent need for further knowledge, to which the present study is intended to contribute.

1.2 Phase Composition of Frozen Soils

It is well established that significant amounts of unfrozen water coexist with ice in soil-water systems at temperatures considerably below 0°C , the normal freezing point of water (Bouyoucos, 1916; Schofield, 1935; Nersesova and Tsytoich, 1966; Williams, 1967, 1976). Several experimental

techniques, the principles of which are well known, permit determination of the amount of unfrozen water occurring over the whole range of negative temperatures of practical importance. Results obtained using the various methods are reasonably consistent; they are also in excellent agreement with predictions based on the thermodynamic relationships discussed in the following section. Examples are presented in Figure 1.1 (the curves were determined by J. Wood using the suction-moisture relationship). For most practical purposes, the unfrozen water content is determined by temperature and soil type. The amount of unfrozen water at any given temperature below 0°C is essentially independent of the amount of ice, and thus of total moisture content. There is a slight dependence on overburden pressure (Williams, 1976).

It is well known that the presence of unfrozen water in frozen soil is due to capillarity and osmotic effects which reduce its freezing point, and to a variety of adsorption effects associated with soil particle surfaces. Anderson and Hoekstra (1965) have shown by x-ray diffraction that most of the unfrozen water is located immediately adjacent to the particle surfaces. The thickness of these interfacial films has been shown to decrease from 5nm or more at 0°C to about 0.9nm at -5.0°C (Anderson, 1968). A clear and precise relationship exists between film thickness (i.e. unfrozen

FIGURE 1.1

UNFROZEN WATER CONTENTS OF FROZEN SOILS AS A
FUNCTION OF TEMPERATURE

2 water content) and the 'suction' associated with the corresponding water content in unfrozen soil (Koopmans and Miller, 1966; Williams, 1967).

1.3 Thermodynamic Relationships: Free Energy and Freezing Point Depression

As the water content of unfrozen soil is progressively reduced, a 'suction' of increasing magnitude develops in the remaining soil water. The dominant cause of this suction is the fall in pore water pressure ascribable to air-water menisci in the soil pores. The phenomenon is described by the well known capillary equation (eg. see Taylor, 1949):

$$H = \frac{2\sigma_{aw}}{r \cdot \rho_w \cdot g} \quad \text{Equation 1}$$

where σ_{aw} = surface tension, air-water interface

ρ_w = density of water

g = gravitational acceleration

and H is the height of a column of water supported by a meniscus of radius, r . H may also be equated with the difference in pressure between the air surrounding the soil and the pore water, $\frac{P_a - P_w}{\rho_w \cdot g}$; while r can be considered the radius of the soil pores in which the air-water menisci reside (Rose, 1966).

The term 'suction' implies a pressure state. At low water contents this is misleading since other forces become more important, especially osmotic and surface adsorption effects. For this reason it is preferable to refer to the

energy state of the soil water in terms of reversible thermodynamics. Application of reversible thermodynamics is limited to systems at equilibrium. This is the case with suction-moisture content relationships, where the soil water is in equilibrium with a given set of forces. Several attempts have been made to apply reversible thermodynamics to soil moisture studies; one of the most impressive examples, with respect to frozen soils is the work of Edlefsen and Anderson (1943).

Within the framework of reversible thermodynamics the Gibbs free energy function is appropriate. At a constant temperature and pressure the energy state of water, in equilibrium with the soil, can be expressed as the difference in absolute Gibbs free energy of the soil water to that of pure water at atmospheric pressure, that is:

$$\Delta G = G_{sw} - G_w \quad \text{Equation 2}$$

where ΔG = the relative Gibbs free energy of the soil water (the term 'potential' as defined by the International Society of Soil Science Bulletin of 1965 is equivalent)

G_{sw} = the partial free energy of the soil water

G_w = the partial free energy of pure, bulk water at the same temperature and pressure.

It is not possible to measure absolute free energy, only differences in this quantity. Depending on how 'suction' is defined or measured it may be exactly equivalent to the relative Gibbs free energy (Aitchison et al., 1966).

The energy state of water in an unfrozen, unsaturated soil resembles that of the unfrozen water (at the corresponding moisture content) in a frozen soil. Regardless of whether the remainder of the pore space is filled with air (unfrozen soil) or ice (frozen soil) the free energies of the water are similar; in the case of a frozen soil, Equation 1 is simply modified for an ice-water interface:

$$P_i - P_w = \frac{2\sigma_{iw}}{r} \quad \text{Equation 3}$$

where σ_{iw} = surface tension, ice-water meniscus.

Thus, the relative Gibbs free energy of the unfrozen water in a freezing soil changes progressively as the quantity of water is diminished by ice formation. A fundamental equation derived from the Clausius-Clapeyron equation relates Gibbs free energy, ΔG , to freezing point depression (Edlefsen and Anderson, 1943):

$$\Delta G = \frac{(T_o - T)\ell}{T} \quad \text{Equation 4}$$

where T = temperature

$T_o - T$ = freezing point depression relative to the
freezing point of pure, bulk water at
atmospheric pressure

ℓ = latent heat of fusion.

The equation as defined, only applies in situations where the ice phase is pure and under atmospheric pressure (Williams, 1976). According to Equation 4 a temperature depression of -0.081°C should induce a negative potential of 10^5Nm^{-2} . It is the change in free energy associated with a reduction in unfrozen water content, that explains the coexistence of ice and water in soils below 0°C ; the free energies of the ice and unfrozen water are in equilibrium.

Continuous films of unfrozen water extend from the frost-line to points in the frozen soil matrix where temperatures are well below zero. Figure 1.2 shows the relative Gibbs free energy of the unfrozen water as a function of temperature. In many soils capillarity is the dominant effect at temperatures down to about -1.5°C (Williams, 1967). When this is the case Equation 3 is most applicable; values for ΔG being essentially the same as those for $P_i - P_w$. Consequently, the potentials generated at such temperatures are normally only a little below the value assigned to pure, bulk water. It is these potentials that are responsible for water movement through the unfrozen zone towards the frost-line, the process that causes primary frost heaving. At colder points within the frozen layer other forces, besides capillarity become important. Here the relative Gibbs free energy of the unfrozen soil water changes in accordance with Equation 4. As indicated by Figure 1.2, the potentials that develop at these points are much lower, in comparison

FIGURE 1.2

RELATIONSHIP BETWEEN TEMPERATURE
AND FREE ENERGY (POTENTIAL) IN
FROZEN SOIL.

with those originating close to the frost-line.¹ It is these potentials that provide the driving force for water transport through already frozen ground, the process responsible for secondary frost heaving.

1.4 Permeability of Frozen Soils

The existence of continuous liquid films in frozen soil suggests a permeability; in other words, water movement is possible. Transport of unfrozen water can take place in both the liquid and vapour phases, in response to pressure, gravitational, temperature, matric, osmotic and electrical gradients (Anderson and Morgenstern, 1973). In unsaturated frozen soils, vapour transport may play a role provided the pore spaces are not blocked by ice. When the ice content of the frozen soil is high, water movement takes place in the unfrozen, interfacial films. At temperatures close to 0°C, pore ice itself, may become permeable on account of liquid filled crystal boundaries in its structure which act as microscopic channels for water flow (Nye and Frank, 1973; Osterkamp, 1973). Moreover, detailed experimental and theoretical work by Miller (1970) and Miller, Loch and Bresler (1975) suggests that water movement in frozen soils is not restricted to the fluid phases. Under conditions in which

¹ Unfrozen water in frozen ground tends to migrate from regions of high potential (low suction) to regions of low potential (high suction).

a steady liquid flux is to be expected, the ice phase may also be conducive to the transport of water by a regelation process. Pore ice appears to be immobile, because its boundaries are stationary, but in reality it may be moving. Molecules of unfrozen water flow through the frozen, porous medium and freeze on to one side of a lens. The latent heat liberated induces thawing on the other side, with a subsequent slow displacement of the ice body towards the melting face. It follows that the presence of ice lenses may not be much of a barrier to flow, even though no direct path exists (Burt and Williams, 1976).

Several procedures have been suggested for the measurement of the apparent hydraulic conductivity of frozen soils. Furthermore, there is a fair degree of agreement of results from the various approaches. Harlan (1973) has advocated the use of hydraulic conductivity values obtained for unsaturated, unfrozen soils, on the assumption that the permeability at a given liquid water content is the same in the frozen as in the partially saturated state. Unfortunately, this analogy breaks down somewhat, as soon as ice lensing occurs in the frozen soil. The amount of unfrozen water present at any given temperature is independent of the total moisture content, but an increase in the latter gives additional ice, which may significantly alter the conditions of flow.

Alternatively, Burt and Williams (1976) developed a permeameter for the direct measurement of water movement in frozen soils. The apparatus is similar to that used to measure the hydraulic conductivity of unfrozen, saturated soils, but with various modifications to accommodate the presence of a frozen sample (Figure 1.3). The experiment is carried out under isothermal conditions, and involves the passage of water from one reservoir into the frozen sample and out into a second reservoir. The water in the reservoirs remains unfrozen because it contains dissolved lactose. The potential gradient is supplied by application of hydrostatic pressure to one reservoir.

The interpretation of hydraulic conductivity coefficients is made difficult by the complex thermodynamic situation in frozen soils. There is little doubt however, that water, in one phase or another, is able to move through frozen soil in significant quantities, particularly at temperatures within a few tenths of 0°C . The latter is to be expected because of the relatively large amounts of unfrozen water at these temperatures. Figure 1.4 shows the apparent hydraulic conductivity-temperature relationships for Castor silt loam and Slims Valley silt. Both curves were determined by the present author using a modified version of the Burt and Williams permeameter. The results are compatible with permeability values for frozen soils

FIGURE 1.3

A SCHEMATIC DIAGRAM OF THE PERMEAMETER USED BY BURT (1974) TO MEASURE THE HYDRAULIC CONDUCTIVITY OF FROZEN SOILS.

FIGURE 1.4

14

HYDRAULIC CONDUCTIVITIES OF FROZEN SOILS AS A
FUNCTION OF TEMPERATURE.
(EXPERIMENTAL DETERMINATIONS).

measured earlier (Burt, 1974). It can be seen that conductivities generally increase as the temperature rises towards 0°C ; commonly observed values being in the range 10^{-8} to $10^{-12} \text{ m} \cdot \text{sec}^{-1}$. The experiment appears to produce conditions of flow for which Darcy's Law is applicable (i.e. a linear relationship between hydraulic gradient and discharge).

1.5 Water Migration in Frozen Soils Induced by a Temperature Gradient

Several authors (Hadley and Eisenstadt, 1955; Dirksen and Miller, 1966; Hoekstra, 1966; Cary and Mayland, 1973; Loch and Kay, 1978) have reported migrations of water in frozen soils in response to imposed thermal gradients. A linear temperature gradient in frozen soil implies a gradient of potential according to the relationship between Equation 4 and Figure 1.2. Thus, a flux of water is to be expected in the direction of decreasing temperature. The amount of moisture migration will also depend upon the apparent hydraulic conductivity of the soil in the frozen state.

(It should be noted that in a system with a temperature gradient, thermodynamic equilibrium, in a precise sense, does not exist; as a result the methods of reversible thermodynamics are not strictly applicable. Taylor and Cary (1964) have attempted to apply irreversible thermodynamics, based mainly on linear phenomenological relationships, to the simultaneous transport of heat and water in non-isothermal systems.

However, the area of irreversible thermodynamics based on non-linear phenomenological relationships has yet to be applied in any detail to studies of soil moisture (E. de Jong, 1967). Consequently, there appears little to be gained from the invocation of irreversible thermodynamics in this context, until the field of non-linear relations has been developed further).

As with unfrozen, partially saturated soils (i.e. with air occupying some pore space), the movement of unfrozen water in frozen soils, in response to potential gradients, necessarily involves concurrent moisture content changes, on account of the substantial variations in conductivity. Consequently a steady state flow is only possible under certain non-linear gradients, and is probably rare. To simplify the mathematical treatment of this process, it is convenient to use a form of Darcy's Law, analogous to the equation used to describe flow in unfrozen, unsaturated soils, for which ready solutions are available. Assuming that transport takes place predominantly in the unfrozen water films, then the flux, q , can be expressed by:

$$q = -K(\theta_{uw}) \frac{d\psi}{d\theta_{uw}} \cdot \frac{d\theta_{uw}}{dx} \quad \text{Equation 5}$$

where K = the apparant hydraulic conductivity of the frozen soil

θ_{uw} = the unfrozen water content

ψ = potential

x = path length of flow

and $q = \frac{Q}{A \cdot t}$ (i.e. the volume of water flowing through a cross-sectional area, A, per time, t).

As defined, the equation only provides an approximation of the flux, q, since it fails to take into account the influence of ice lensing, or indeed, transport within the ice phase itself.

Alternatively, Loch and Kay (1978) have proposed that, under steady state conditions, the hydraulic conductivity-temperature relationship can be integrated over the length of the frozen sample, L, to give an 'overall' hydraulic conductivity, K' , of the form:

$$K' = \frac{L}{\int_{X_1}^{X_2} \frac{1}{K(X)} dx} \quad \text{Equation 6}$$

where X_1 and X_2 = the distance coordinates coinciding with the boundaries of the flow path

$K(X)$ = hydraulic conductivity as a function of the temperature profile.

The present author however, prefers the former approach, since calculation of K' using Equation 6, necessarily involves prior knowledge of $K(X)$ for each soil investigated, which produces a somewhat circular argument.

As unfrozen water accumulates within the frozen ground, its Gibbs free energy tends to rise, such that the ice and water are no longer in equilibrium. As a result, freezing

of the accumulated water will occur. The permeability of frozen soil decreases with temperature as in Figure 1.4, while the potential decreases linearly. Therefore, ice accumulation will be favoured where the permeability is greatest and decreasing rapidly with temperature; that is where the temperatures of the frozen ground are relatively warm. Over long periods of time however, water migration may well result in ice accumulation at much colder temperatures (Williams, 1977).

Water migration within frozen soils is associated with secondary frost heaving of already frozen ground, and the development of high heaving pressures. Mackay et. al. (1979) claim to have measured these effects in the field. The process has important implications for the long term hydrologic regime of permafrost. Thermal gradients normally exist in permafrost due to the equilibrium between internal heat gain with depth, and heat loss from the ground surface. According to Lachenbruch (1968), heat flow from the Earth's interior normally results in an increase of approximately 1°C for every 30-60m. increase in depth (the geothermal gradient), which implies a significant gradient of potential. Thus, one would expect an upward migration of ice and water through the permafrost, toward the colder near-surface temperatures (Harlan, 1974). Over geological time, the limiting effect of low permeability decreases, and considerable ice accumulation can occur. Indeed, this redistribution may be partly responsible for the segregated ice and massive ice bodies commonly observed just below the active layer.

Pipeline construction in permafrost regions must also take into account the volume changes associated with frost heave of already frozen ground. The temperature of a buried, refrigerated gas pipeline must be well below 0°C . Consequently the pipeline will often act as a heat sink, setting up thermal gradients within the permafrost. The resultant potential gradients can be expected to induce water migration through the frozen ground toward the buried pipe. With time, ice accumulation may cause sufficient differential heave around the pipe to rupture it.

1.6 Aims of the Present Study

This study has two principal objectives:

1. To develop an apparatus with which to investigate temperature-induced water migration in saturated, frozen soils. The experiment should allow, for the first time, direct measurement of such movements, and provide data which may be used to estimate apparent hydraulic conductivity coefficients for frozen soils.
2. To evaluate the observations with respect to current theoretical and applied studies of frozen ground, as outlined in the previous sections. Consideration should also be given to the inherent properties of the experiment, which may have significance in relation to values that can be expected in the field, or with alternative test procedures.

CHAPTER 2

EXPERIMENTAL METHOD

2.1 Apparatus Design

This study is concerned with water migration in saturated frozen soils under various temperature gradients. An experiment has been designed which allows, for the first time, direct measurement of such movements. The basic system consists of a frozen soil sample placed between two bodies of water. On application of a temperature gradient across the system, the frozen soil can be expected to act in general, as a kind of pump, inducing water migration in the direction of decreasing temperature. This is in accordance with the relationship between temperature and free energy (Equation 4) illustrated in Figure 1.2. Several important requirements must be met, however, in the construction of an experimental apparatus to investigate this phenomenon:

1. The water in the reservoirs must be at the same temperature as the ends of the frozen soil sample. The juxtaposition of ice and pure, bulk water will however, cause freezing of the latter at any temperature below 0°C . Therefore, the potentials of the water in the reservoirs and the unfrozen soil water must be equalized before a temperature gradient is established across the system.

2. Both ends of the sample must be held at constant temperatures below zero. Because heat flow through the frozen sample may be variable, it is desirable to have the end plates controlled by a device which is able to maintain the required sub-zero temperature with rapid adjustment to fluctuations in cooling load.
3. Heat flow must be uniaxial. If this is not the case, then the temperature and moisture distribution will not be uniform (in cross-section) at any given distance along the sample. It is necessary therefore, to minimize any radial heat exchange between the sample and its surroundings.

The design of the complete experimental apparatus is shown in Figure 2.1 (a scale drawing of the sample cell and end plates is given in Appendix I). In construction it is similar to that used by Burt and Williams (1976) to measure the hydraulic conductivity of frozen soil samples under isothermal conditions. The essential difference is that the hydraulic gradient is induced by a thermal gradient rather than by a pressure differential.

The soil sample is contained in a plexiglass cylinder, 3.175 cm long, with a wall thickness of 1.90 cm and an inside diameter of 5.40 cm. The cylinder and reservoirs are sandwiched between two Aluminum end plates containing Peltier modules. These cooling devices are controlled by a thermo-electric cooling control system which can maintain temperatures

FIGURE 2.1

SCHEMATIC DIAGRAM OF THE COMPLETE EXPERIMENTAL APPARATUS
(SEE APPENDIX I FOR SCALE DRAWING OF THE CELL AND END PLATES)

constant to within $\pm 0.02^{\circ}\text{C}$ under steady state conditions (see Appendix II). The soil sample is held firmly in place by two porous brass plates which are in thermal contact with the thermoelectric cooling plates.

2.2 Preparation of Soil Samples

All samples are prepared as slurries. This is normally done by saturating the soil under vacuum with deaired, deionized water. The sample is then frozen into the plexiglass cylinder. The method of freezing governs the type and pattern of ice lenses. The presence of ice lenses in a sample can be expected to reduce its hydraulic conductivity compared to an unlensed sample, at all temperatures. For example, Burt (1974) found that ice lenses lowered the hydraulic conductivity of a Carleton silt soil by approximately two orders of magnitude, at a temperature of -0.075°C . To minimize moisture redistribution during freezing (the process responsible for ice lensing) the sample is frozen rapidly from all directions in a refrigerator, at a temperature of approximately -18°C . This means that most of the pore water freezes in situ. The temperature of the frozen sample can then be brought into equilibrium with that of the cold plates. Before assembly the ends of the frozen soil sample are scraped with a metal blade to give a plane surface, flush with the plexiglass cylinder. Two soils were used in the experiments, Castor Silt loam and Slims Valley Silt. Grain size distributions for these soils are shown in Figure

2.2. Slims Valley Silt was used in experiments 1, 2 and 6, and Castor Silt loam in experiments 3, 4 and 5.

2.3 Prevention of Freezing in the Reservoirs

Unfrozen water in frozen soil possesses a suction or potential relative to free, pure water at the same temperature due to capillarity, dissolved salts and a variety of adsorption effects (see Section 1.2). Consequently, if pure water was placed in the reservoirs, even if freezing did not occur, there would be a tendency for it to be pulled into the frozen sample. The potentials of the water in the reservoirs, and the unfrozen soil water can be equalized by creating an osmotic potential in the reservoirs. With respect to a pool of pure water at atmospheric pressure, unfrozen water in frozen soil and an aqueous solution of lactose² are both subject to a decrease in Gibbs free energy. Since the potential of the unfrozen water is fixed for any negative temperature, the concentration of lactose can be adjusted so as to produce a corresponding decrease in free energy of the water in the reservoirs. The concentration required to produce the correct osmotic potential is easy to obtain for any given temperature (see Appendix III).

² For a full discussion on why lactose was chosen for use in the reservoirs the reader is referred to Burt (1974), p. 22-25.

FIGURE 2.2

On assembly both reservoirs are normally filled with lactose solution, as with the Burt and Williams permeameter. Since the present experiment is non-isothermal, the concentrations should ideally be different in the two reservoirs. Initially it was tacitly assumed that such an arrangement would result in a counter osmotic potential gradient, which might confound the measurement of thermally induced flow. However, this point of view has changed somewhat during the course of the study. It now seems likely that the opposite is true; a counter osmotic potential gradient should not ensue from different concentrations of lactose in the reservoirs, because of the presence of a thermal gradient.

Discussions with Dr. R. Shigeishi, Dr. J.K. Torrance, and Dr. P.J. Williams indicate that the temperature gradient is critical to this interpretation. If it were to be removed, an osmotic potential gradient would automatically exist across the system, due to the different concentrations of lactose solution in the reservoirs. As it is, the thermal gradient serves to change the potentials in the unfrozen soil water, such that both ends of the frozen sample can be in equilibrium with different concentrations of lactose solution, at different sub-zero temperatures, without setting up an osmotic potential gradient (eg. see Figure 4.1 (C)). Obviously further research is required on this point.

Modification of the test procedure should include the use of two equilibrium concentrations of lactose solution for each temperature gradient investigated, instead of one.

In the experiments reported here the initial assumption was held to be true. For this reason a lactose solution of similar strength was employed in both reservoirs; accuracy at the inflow face being sacrificed in order to achieve a representative rate of outflow. The concentration was adjusted to give a freezing point depression equivalent to the temperature of the coldest cooling plate. It follows that the potentials of the unfrozen soil water and the lactose solution at the 'cold' end are in equilibrium. This procedure however, coupled with the presence of dialysis membranes (see next paragraph) results in a local osmotic potential gradient at the 'warm' end, which could induce flow in the opposite direction to the thermally produced hydraulic gradient. This is discussed in detail in Section 4.1.

The presence of lactose in the end reservoirs creates the possibility of lactose molecules moving into the sample either by molecular diffusion, or through transport in water moving from the inflow reservoir. A dialysis membrane is therefore fitted on each end of the frozen sample to impede entry of the lactose molecules into the unfrozen soil water. Because of their small molecular size however, the lactose molecules are still able, to some extent, to pass through this semi-permeable membrane. The passage of lactose into the sample raises the potential of the unfrozen water, with the result that some pores become ice free (this effect is discussed further in Section 4.2).

The dialysis membranes have very small pores which remain unfrozen even at the lowest temperatures required by the experiment. The lactose solution obviates ice formation in the reservoirs; for a given temperature the lactose solution will be in equilibrium with the unfrozen soil water, and so the reservoirs should remain unfrozen. If however, the temperature were to fall slightly this equilibrium would be upset and ice might form in the reservoirs. In such a situation, the dialysis membranes can be expected to function as phase barriers, preventing the penetration of ice nuclei into the reservoirs. Thus, it is feasible to maintain the solution in a supercooled state until the temperature of the cold plate returns to its equilibrium value.

From a practical point of view the membranes also serve to keep the soil particles in place; the sample is confined by the porous brass plates, but these cannot prevent individual particles from dislodging. The permeability of the dialysis membranes is approximately $10^{-9} \text{m} \cdot \text{sec}^{-1}$ (Burt, 1974). Since the hydraulic conductivity of the frozen soil is only rarely likely to exceed this limiting value, it seems safe to conclude that the membranes do not act as a significant impediment to flow.

2.4 The Use of Supercooled Water

Miller (1976) has suggested that the presence of

solutes confounds interpretation of the dynamics of water transport in frozen soils to an "unknown degree" and that the use of supercooled water might simplify matters. Two experiments (Experiments 5 and 6) were run with the present apparatus in which pure water, previously deionized and deaired was used in the end reservoirs. The dialysis membranes served as effective phase barriers separating soil ice from the supercooled water in the reservoirs. The porous brass plates were replaced by aluminum discs, with machined surfaces and smooth pores. The brass plates have jagged surfaces which could puncture the membranes and thereby permit ice propagation into the reservoirs. The irregularities might also serve as nucleation sites for ice growth. Results from these experiments proved to be somewhat unexpected, since flow appeared to take place in the opposite direction to the hydraulic gradient implied by the temperature profile. See Section 3.5 for a detailed discussion on this topic.

2.5 Establishment of a Linear Temperature Gradient

A gradient of potential is induced by imposing a linear temperature gradient across the frozen sample. This is done by maintaining the two thermoelectric cooling plates at different sub-zero temperatures. The time required to establish an approximately linear temperature gradient was about 12 hours in all experiments. An example is shown in Figure 2.3 (the constant end temperatures represent the temperatures of the thermoelectric cooling plates; deviations from linearity at both ends of the

FIGURE 2.3

APPROACH TO LINEARITY IN THE TEMPERATURE PROFILE.
(EXPERIMENT NO. 4A)

temperature profile are due to the different thermal conductivities of the aluminum end plates and the frozen soil sample). The gradient is increased in a stepwise fashion, with measurements delayed for at least 6 hours, but often over night to allow for re-equilibration. It should be noted that although the temperature profile approaches a steady state situation on a macroscopic scale, there will be continual microscopic perturbations associated with coupled heat and moisture transport.

To reduce radial heat exchange between the frozen sample and air at room temperature, the apparatus is wrapped in a jacket of fibreglass insulation, approximately 5 cm thick. The entire assembly is then placed in a 'Precision' low temperature incubator which maintains a uniform air temperature of between 0 and +1°C. The fact that a nearly linear temperature gradient is achieved suggests that radial heat exchange is relatively small compared with axial heat flow. Continuous monitoring of the temperature profile is provided by five calibrated thermistors (estimated limit of accuracy $\pm 0.01^\circ\text{C}$) located along the sample holder and in the end plates. The tips of the thermistors extend about 0.25 cm into the frozen soil.

2.6 Measurement of Flow

Once a stable, linear temperature gradient is established across the system, water appears to pass from the 'warm' reservoir into the frozen sample, and out into the

second, colder reservoir. Inflow and outflow are measured by timing the movement of menisci along capillary tubes. Readings are taken every thirty minutes for a period of six hours. This procedure may be repeated over a number of days. The observed velocities are converted into a volume using the radii of the capillary tubes (determined to be 0.025 cm).

2.7 Calculation of Hydraulic Conductivity Coefficients

On application of a temperature gradient across the system the frozen soil acts in general as a kind of pump, inducing water migration in the direction of decreasing temperature. Since the temperature profile approaches linearity, at least on a macroscopic scale, the implied gradient of potential must also be linear in nature. This means that we can use the generalized form of Darcy's Law, in conjunction with the implied potential gradient, and the volume of outflow, to calculate an approximate, mean hydraulic conductivity coefficient for the frozen sample. Thus, Equation 5 can be re-written in the form:

$$q = -\bar{K}A \frac{d\psi}{dx} = \frac{-\bar{K} \cdot \Delta h \cdot A}{L} \quad \text{Equation 7}$$

where $\bar{K}$ = mean hydraulic conductivity of the frozen sample

Δh = potential expressed in cms of water head

A = sample cross-sectional area

L = path length of flow

It follows that:

$$\bar{K} = \frac{q \cdot L}{\Delta h \cdot A} \quad \text{Equation 8}$$

The flux, q , is given by the equation:

$$q = \pi r^2 \frac{d}{t} \quad \text{Equation 9}$$

where r = radius of the capillary tube

d = distance travelled by the meniscus in time, t .

Since L , A and r^2 are constants in the design of the apparatus ($L = 3.175$ cm; $A = 22.88$ cm²; and $r^2 = 6.25 \times 10^{-4}$ cm²):

$$\bar{K} \approx \frac{d \cdot c}{t \cdot \Delta h} \quad \text{Equation 10}$$

where c = the product of the constants, which equals 2.725×10^{-4} for the dimensions of the present apparatus. Thus, an estimate of the mean hydraulic conductivity coefficient is given by:

$$\bar{K} = \frac{2.725 \times 10^{-4} \cdot d}{t \cdot \Delta h} \quad \text{Equation 11}$$

It follows that calculation of $\bar{K}$ involves measurement of d , t and Δh . The values of d and t can be obtained directly, while Δh can be deduced from the temperature profile, as in the following example:

In an experiment with a Castor Silt loam soil (see Appendix III, Expt. 4B), the cooling plates at either end of the sample were maintained at temperatures of -0.2 and -0.5°C respectively, giving a temperature gradient of 0.048°C/cm . The mean temperature was -0.35°C . According to Equation 4 (see Section 1.3):

the potential generated at $-0.2^\circ\text{C} \approx 2.469 \times 10^5 \text{ Nm}^{-2}$

and the potential generated at $-0.5^\circ\text{C} \approx 6.173 \times 10^5 \text{ Nm}^{-2}$

The difference between these two values gives the potential, generated over the length of the sample, which equals $3.704 \times 10^5 \text{ Nm}^{-2}$ (equivalent to 3,704 cm of water head).

Given that the capillary meniscus moved 0.25 cm in 1800 seconds, the mean hydraulic conductivity coefficient is:

$$\bar{K} = \frac{2.725 \times 10^{-4} \cdot 0.25}{1800 \cdot 3704} \text{ cms}^{-1} \quad \text{Equation 12}$$

$$\therefore \bar{K} = 1.02 \times 10^{-13} \text{ ms}^{-1}$$

It must be remembered that the hydraulic conductivity is a function of the temperature profile, and that $\bar{K}$ represents an "average" value, over the length of the frozen sample. Moreover, the experiment should not be viewed strictly as a permeability test. In conventional hydraulic conductivity tests the discharge is normally directly proportional to the hydraulic gradient applied; this follows from Equation 7. In the experiments described here however, the applied temperature gradient gives rise to an internal driving force (i.e. the pump analogy). Since a change in the potential gradient necessarily involves an alteration of the temperature profile, there must also be a corresponding change in unfrozen water contents. This, in turn, implies a change in the sample's apparent hydraulic conductivity. Thus, a direct linear relationship between discharge (meniscus velocity) and potential gradient is not to be expected.

CHAPTER 3
EXPERIMENTAL OBSERVATIONS AND
PRELIMINARY EVALUATION

All relevant data collected during the experiments are presented in Appendix IV. The discussion can logically be divided into two separate themes. Firstly, it is convenient to consider the observations in terms of the current theoretical understanding of frozen soils, as outlined earlier. In this chapter a tentative, phenomenological evaluation of the results will be presented, attention being paid to the variation of flow rates with respect to thermal gradients, mean temperatures and soil properties. However, certain properties associated with the experiment itself, produce effects which may modify the observations. This represents the second theme of the discussion, presented in the following chapter.

3.1 Relationship Between Flow Rate and Thermal Gradient

For most tests stable and essentially linear temperature gradients were established across the system, as this served to simplify interpretation of the results. During the approach to linearity in the temperature profile, a steady flux of water was normally observed entering the reservoirs at both ends of the frozen sample. An example

of this phenomenon is illustrated in Figure 3.1. An examination of Figure 2.3 indicates that initially both thermoelectric cooling plates were colder than the interior of the frozen sample. Presumably the sample warms somewhat during assembly of the apparatus. Consequently water tends to leave the frozen soil from both ends, in accordance with the relationship between temperature and free energy (Equation 4). This implies a net loss of water from the soil, until an approximately linear temperature gradient is achieved across the system. The discharge into the 'warm' reservoir may also be due in part, to a counter osmotic potential gradient across the membrane at the 'warm' end of the sample. The extent to which this gradient will influence flow depends on the degree of semi-permeability (resistance to movement of lactose molecules) of the membrane. This effect is considered in more detail in Section 4.1.

One experiment was run using a Slims Valley Silt soil (experiment No. 2A) in which there was no temperature gradient across the sample. Both thermoelectric cooling plates were maintained at a constant temperature of -0.325°C , while the reservoirs contained the same concentration of lactose solution, adjusted to give the appropriate freezing point depression. Once stable temperatures were achieved, cumulative inflow and outflow were observed to be negligible

FIGURE 3.1 CUMULATIVE OUTFLOW DURING APPROACH TO LINEARITY
 IN THE TEMPERATURE PROFILE

(see Appendix IV). Since no driving potential was generated, and the potentials of the lactose solution in the reservoirs, and the unfrozen soil water were in equilibrium, there was no tendency for water to be pulled into or out of the frozen sample.

When a stable, linear temperature gradient is established across the system, a fairly steady flux of water appears to take place in the direction of decreasing temperature. This is consistent with the theory outlined in Section 1.3. Water normally passes from the 'warm' reservoir into the frozen soil, and there is an outflow into the second, colder reservoir. The rate of flow will be governed by the magnitude of the temperature gradient (i.e. the potential for water movement), the mean temperature of the frozen sample (i.e. the apparent hydraulic conductivity), and the soil type. Generally, mean rates of inflow were less than the outflow by approximately an order of magnitude, which implies a net loss of water from the frozen soil sample. In one respect a net gain might be expected since permeability decreases non-linearly with temperature (see Figure 1.4), while the potential decreases linearly. However, certain properties associated with the test procedure itself, produce effects which could account for this observed disparity between rates of inflow and outflow. These are discussed in detail in the following chapter.

For similar soils the flux should be a play-off between two factors: the gradient of potential and the permeability-temperature relationship. In the tests reported here, samples at the same mean temperature, showed a marked increase in flow rates with increasing temperature (potential) gradient. For example, Figures 3.2 and 3.3 show cumulative inflow-outflow over time for two samples (Castor Silt loam and Slims Valley silt) at approximately the same mean temperature (-0.25°C). The temperature gradients were 0.016 and $0.0896^{\circ}\text{C}/\text{cm}$ respectively, giving potential gradients of $1.235 \times 10^5 \text{Nm}^{-2}$ and $6.420 \times 10^5 \text{Nm}^{-2}$. It can be seen that cumulative outflow is well over an order of magnitude greater with the larger temperature (potential) gradient. For comparative purposes cumulative outflow curves from Figures 3.2 and 3.3 are plotted together on a linear scale in Figure 3.4. It should be remembered that the tests were carried out at relatively warm mean temperatures (i.e. in the range 0 to -0.5°C), where the variation of apparent hydraulic conductivity with temperature is considerable. At lower mean temperatures the permeability-temperature relationship levels off; here one would expect the temperature (potential) gradient to become the predominant factor. It is unlikely however, that an increase in the gradient would significantly increase fluxes, due to the limiting effect of the extremely low conductivities below about -0.5°C .

C

FIGURE 3.2 CUMULATIVE INFLOW-OUTFLOW INDUCED BY A
RELATIVELY SMALL TEMPERATURE GRADIENT

FIGURE 3.3 CUMULATIVE INFLOW-OUTFLOW INDUCED BY A
 RELATIVELY LARGE TEMPERATURE GRADIENT

FIGURE 3.4 VARIATION OF CUMULATIVE OUTFLOW WITH
TEMPERATURE GRADIENT, FOR TWO SAMPLES
AT SIMILAR MEAN TEMPERATURES

With a step-wise increase in the temperature gradient, a temporary surge in the amount of outflow was observed. For example, Figure 3.5 shows the effect on cumulative outflow of increasing the temperature gradient from $0.0480^{\circ}\text{C}/\text{cm}$ to $0.0640^{\circ}\text{C}/\text{cm}$. The sample used was ~~a~~ Castor Silt loam soil. Initially, with the stable temperature profile, it can be seen that outflow rates were fairly steady (mean of $1.55 \times 10^{-2} \text{mm}^3/\text{min}$). Following the step change in the temperature gradient there is a distinct increase in the outflow rate (mean of $3.57 \times 10^{-2} \text{mm}^3/\text{min}$). This effect is only temporary, however, and flow rates soon level off, to reach a new steady state (mean of $2.51 \times 10^{-2} \text{mm}^3/\text{min}$) as the temperature profile re-establishes its linearity. It should be noted that flow rates achieved with the new temperature gradient are slightly greater than those measured earlier. The surge was not observed with rates of inflow, although this is not surprising since the step changes were always effected by lowering the temperature of the 'cold' end plate. Concurrently, the reservoirs were flushed with a fresh supply of lactose solution, the concentration of which was adjusted to give a freezing depression equivalent to the new 'cold' end temperature.

3.2 Temperature Dependence of Flow Rates

The same temperature gradient applied across similar soils should induce different rates of flow depending on the mean temperature of the frozen samples. This follows

FIGURE 3.5 CUMULATIVE OUTFLOW DURING A STEP-WISE
INCREASE IN THE TEMPERATURE GRADIENT

directly from the reduction in hydraulic conductivity with decreasing temperature (see Figure 1.4). In other words, a specific gradient of temperature (potential) can be expected to yield a greater flux at temperatures close to 0°C

(i.e. relatively high hydraulic conductivities) than at colder mean temperatures. This is illustrated in Figures 3.6 and 3.7, which show cumulative inflow-outflow over time for similar samples (Slims Valley silt and Castor Silt loam), with mean temperatures of -0.16° and -0.35°C respectively. The temperature gradient in both cases was approximately 0.05°C/cm. This gradient produced a mean outflow rate of $9.846 \times 10^{-2} \text{ mm}^3/\text{min}$ for the warmer sample, whereas for the colder sample it only yielded a mean outflow rate of $1.5546 \times 10^{-2} \text{ mm}^3/\text{min}$. For comparative purposes Figure 3.8 shows cumulative outflow curves from Figures 3.6 and 3.7 plotted together on a linear scale. One would expect this drop-off in flow rates to continue with decreasing temperature, so that permeability becomes the limiting factor, regardless of the size of the temperature induced gradient of potential. Examination of Figure 3.6 reveals that outflow took place from both ends of the sample in experiment No. 2B. This may be due to the existence of a reverse osmotic potential gradient across the membrane at the 'warm' end of the sample. This gradient could be sufficient to locally counteract the temperature induced gradient of potential, with the result that water moves out of the frozen soil into

FIGURE 3.6 CUMULATIVE INFLOW-OUTFLOW INDUCED BY A TEMPERATURE GRADIENT AT A RELATIVELY WARM MEAN TEMPERATURE

FIGURE 3.7 CUMULATIVE INFLOW-OUTFLOW INDUCED BY A TEMPERATURE GRADIENT AT A RELATIVELY COLD MEAN TEMPERATURE

FIGURE 3.8 VARIATION OF CUMULATIVE OUTFLOW WITH MEAN TEMPERATURE, FOR TWO SAMPLES SUBJECTED TO SIMILAR TEMPERATURE GRADIENTS

the 'warm' reservoir. This effect is dealt with in detail in Section 4.1.

3.3 Variation of Flow Rates with Soil Type

One can expect some variation in flow rates as a result of soil type. The unfrozen water content and apparent hydraulic conductivity of a frozen soil are related through pore size distribution, void ratio and surface area. These factors are themselves related to the grain size characteristics of the soil. Consequently, a temperature gradient applied across two different soils, at approximately the same mean temperature should produce different volumes of discharge, depending on the grain size distribution of the soils.

Particle size analyses for the two soils used in the experiments are shown in Figure 2.2. The curves are similar because both Slims Valley silt and Castor silt loam possess a high percentage of grains in the silt size range (2 to 50 μ m).

It is known that silts contain much unfrozen water close to 0°C (see Figure 1.1). This is reflected in the apparent hydraulic conductivity-temperature relationships for both soils, presented in Figure 1.4. If the curves are extrapolated to lower temperatures (i.e. below the range where readings were taken with the Burt and Williams permeameter), then it appears that the samples will have very low permeabilities. This is because most of the unfrozen water

is contained in relatively large pore-spaces, which become frozen just below 0°C . However, a different sample (for example a clay soil) with different grain-size characteristics, and thus a different unfrozen water content-temperature relationship may permit significant flow even at much colder temperatures.

Discussion of the influence of grain-size on unfrozen water content and apparent hydraulic conductivity sheds light on the relationship between soil type and flow rate. However, an explicit comparison is impossible here, since both soils used in the experiments possessed similar, but not identical grain-size distributions; Castor Silt loam having a greater percentage of particles in the clay size range (See Figure 2.2). The reader is referred to Burt (1974) p. 39-48 for a more detailed treatment, albeit with respect to the isothermal situation, of the relationship between particle size and permeability. Further research is needed with the present apparatus to clarify the influence of soil type on thermally induced water movements. This must obviously include the testing of a large number of different soils, for which unfrozen water content-temperature relationships, and grain-size distributions are known.

3.4 Temperature and Apparent Hydraulic Conductivity

The hydraulic conductivity coefficients were derived from measured volumes of outflow, in conjunction with implied gradients of potential, according to the procedure described in Section 2.7. Thus, it is to be expected that the

values will exhibit the same temperature dependence as was observed with the flow rates (see Sections 3.1 and 3.2). Calculated values for $\bar{K}$ are presented in Table 3.1, along with the range of temperatures for which the results are valid. Theoretically, $\bar{K}$ represents the apparent hydraulic conductivity-temperature relationship averaged over the length of the frozen sample. In practice however, $\bar{K}$ will be closer to the limiting, or lowest permeability at the 'cold' end of the sample.³ Assuming that this is the case, then it is most useful to investigate variations in $\bar{K}$ with respect to 'cold' end temperatures.

For tests run on the same sample, it can be seen that conductivities generally decrease with decreasing 'cold' end temperature, although the data do not compare well between samples. It should be remembered that accuracy in hydraulic conductivity measurements is specified in orders of magnitude. This is due to natural variability. No two samples (even of unfrozen soils) can be expected

³ Equations describing the influence of a small layer of low permeability on flow rates (i.e. the funnel analogy) are easily derived from basic considerations (eg. see Williams, P.J., 1964, "Investigation of Plaster of Paris, with a view to its use as a barrier to air in pore water pressure measurements in partially saturated soils", Norwegian Geotechnical Inst., Internal Rept. F97, 62/617).

Table 3.1 Hydraulic Conductivity Coefficients*

Experiment No.		'Cold' end temperature (°C)	'Warm' end temperature (°C)	$\bar{K}$ (ms^{-1})
2B	Slims Valley Silt	-0.300	0.00	3.00×10^{-13}
1B	" "	-0.515	+0.075	2.72×10^{-12}
1A	" "	-0.520	+0.040	3.94×10^{-12}
3A	Castor Silt Loam	-0.300	-0.200	7.04×10^{-13}
3B	" "	-0.400	-0.200	6.27×10^{-14}
4B	" "	-0.500	-0.200	9.73×10^{-14}
4D	" "	-0.600	-0.200	6.44×10^{-14}

*Calculated according to the method outlined in Section 2.7

to give identical values for $\bar{K}$. Add to this the fact that each sample has a different freezing history, then it is hardly surprising that there is not exact agreement between coefficients obtained from different samples. The permeability values presented in Table 3.1 appear to be compatible with data obtained earlier (using the Burt and Williams permeameter), illustrated in Figure 1.4. The curves in Figure 1.4 must be extrapolated to temperatures in the same range as those for the 'cold' end in Table 3.1.

Since a change in temperature profile necessarily involves an alteration of the potential gradient, the experiments cannot be expected to produce a direct linear relationship between discharge (meniscus velocity) and potential gradient. Consequently the form of any hydraulic conductivity-temperature relationship derived with the present apparatus may differ from curves achieved under isothermal conditions. This is simply a function of the nature of the experiment itself. It appears however, that the calculated values for $\bar{K}$, when taken as a whole, represent a fairly good approximation of the limiting permeabilities that can be expected under similar thermal gradients in the field. In conclusion, it should be noted that the values presented in Table 3.1 are only simple averages; for each temperature gradient investigated there will be a range of conductivities. Further research should include transformation of the raw data such that $\bar{K}$ reflects this temperature

dependence more closely. This might be achieved through the use of Loch and Kay's (1978) 'overall' hydraulic conductivity coefficient.

3.5 Observations made with Supercooled Water in the Reservoirs

Supercooled water has been used successfully in the Burt and Williams isothermal permeameter, and the results are essentially identical to those achieved with lactose solution (see Figure 1.4). Two experiments (Expt. 5 and Expt. 6) were run with the present apparatus, in which supercooled water was present in the end reservoirs instead of lactose solution. Here, unlike the Burt and Williams experiment, there is no external driving force, rather the frozen soil, subjected to a temperature gradient, should act as a kind of pump inducing water migration in the direction of decreasing temperature. Initially, results obtained with supercooled water in the reservoirs appear somewhat surprising, since water moved into the frozen sample at the 'cold' end, and outflow took place from the 'warm' end. In other words, water appeared to be moving in the opposite direction to the hydraulic gradient implied by the temperature profile. Cumulative inflow and outflow were essentially equal. For example, a temperature gradient of $0.032^{\circ}\text{C}/\text{cm}$ across a sample of Castor silt loam soil, at a mean temperature of -0.2°C , produced a mean rate of inflow of $6.67 \times 10^{-1} \text{mm}^3/\text{min}$ at the 'cold' end, and a mean outflow rate of $7.93 \times 10^{-1} \text{mm}^3/\text{min}$

at the 'warm' end (see Figure 3.9). There was no apparent decline in flow rates over the duration of the experiment (6 hours).

Further consideration reveals that this phenomenon should not be so surprising. Since supercooled water is not in thermodynamic equilibrium with the unfrozen soil water, it is to be expected that it will be pulled into the frozen sample. The greatest disparity between potentials exists across the 'cold' end of the sample; thus water moves from the 'cold' reservoir into the frozen soil where ice lensing can be expected to take place. On dismantling the apparatus at the end of Expt. 5 (total duration 4 days), an ice lens, approximately 0.2 cm thick was observed at the 'cold' end of the sample. Presumably a frost heave pressure is generated by the lens growth, causing some melting of ice at the 'warm' end. The subsequent expulsion of water follows directly from the rise in pore water pressure. This hypothesis appears to account for the observed reversal in flow. It is likely that the phenomenon will continue until the ice pressure within the confined sample rises to an equilibrium value. The tendency to 'pull' water into the sample from the 'cold' reservoir should then disappear as the potentials of the supercooled water and the unfrozen soil water approach a state of mechanical equilibrium.

FIGURE 3.9 OBSERVED REVERSAL IN CUMULATIVE INFLOW-OUTFLOW
 WITH SUPERCOOLED WATER IN THE RESERVOIRS

One test (Expt. 6A) was conducted using a Slims Valley Silt soil in which there was no temperature gradient. Both thermoelectric cooling plates were maintained at a constant temperature of -0.1°C , while the reservoirs were again filled with supercooled water. According to the explanation developed above, one would expect water to be drawn into the sample from both ends. Figure 3.10 however, shows that this was not the case; water continued to move out of the sample from one end, even though there was no apparent gradient of potential. After a period of 6 hours cumulative inflow was observed to be considerably greater than cumulative outflow (54.98 mm^3 compared to 3.04 mm^3). It may be that a very small temperature gradient can exist across the system (i.e. within the limits of measurement precision). If this is the case, then the implied gradient of potential may produce a similar effect to that described above. Due to the unstable thermodynamic situation within the frozen sample, this phenomenon may even reverse itself periodically, although there should always be a net inflow, until the ice pressure rises to its equilibrium value.

FIGURE 3.10 CUMULATIVE INFLOW-OUTFLOW WITH SUPERCOOLED
WATER IN THE RESERVOIRS, AND NO TEMPERATURE
GRADIENT ACROSS THE SAMPLE

CHAPTER 4

PROPERTIES ASSOCIATED WITH THE
EXPERIMENTAL ARRANGEMENT

The experiment demonstrates the movement of water within saturated frozen soils, in response to applied temperature gradients. The inherent properties of the experiment may have significance in relation to rates of flow which can be expected in the field, or with alternative test procedures. A number of problems result from the experimental design; these are discussed below, with attention being paid to their influence on the results reported in Chapter 3.

4.1 Osmotic Potential Gradients

In most experiments mean rates of inflow were less than the outflow by approximately an order of magnitude (see Appendix IV). This implies a net loss of water from the frozen soil sample. In one respect a net gain might be expected since permeability decreases with temperature, non-linearly as in Figure 1.4, while potential decreases linearly, according to Equation 4. Thus, a net inflow could be explained by ice accumulation occurring within the sample. However, several other factors related to the properties

of the experiment itself, could account for the observed disparity between inflow and outflow rates:

1. For reasons discussed in Section 2.3, the test procedure adopted involved the use of a single concentration of lactose solution in the end reservoirs. This concentration was such that the potentials of the water in the 'cold' reservoir and the adjacent unfrozen soil water were in equilibrium. Since the same concentration of lactose solution was used in the 'warm' reservoir, the resultant local osmotic potential gradient, coupled with the presence of semi-permeable membranes, could be sufficient to partly or wholly counteract the temperature-induced potential gradient at the 'warm' end. Figure 4.1(A) is a schematic representation of the potentials that can be expected to develop in a system with a linear temperature gradient, in which both reservoirs contain similar concentrations of lactose solution. It can be seen that, although the potentials at the 'cold' end are equal, there is a pronounced local osmotic potential gradient across the 'warm' end, equal in magnitude to the total thermal driving potential developed over the whole sample. A decline in observed rates of inflow can be expected on this account. The osmotic potential gradient serves to counteract intake at the 'warm' end, causing an apparent reduction in mean rates of inflow. This is reflected in the cumulative inflow-outflow graphs presented in Chapter 3. These reveal that cumulative inflow is, often one or two orders of magnitude less than cumulative outflow (eg. Expt. 4D), sometimes non-existent

FIGURE 4.1

SCHEMATIC REPRESENTATION OF THE POTENTIALS DEVELOPED WITHIN THE SYSTEM,
ASSUMING 100% SEMI-PERMEABLE MEMBRANES

(eg. Expt. 3B) and rarely manifested as a cumulative outflow (eg. Expt. 2B). Interpretation of the effects of the counter osmotic potential gradient at the 'warm' end is complicated by the fact that the dialysis membranes are not 100% semi-permeable to the lactose molecules (see Section 4.2). The passage of lactose molecules into the soil pores can be expected to weaken the efficiency of the osmotic effect (which depends on the semi-permeability of the membranes), thereby reducing the severity of the counter potential gradient.⁴

Depending on the time it takes for the effects of the osmotic potential gradient to be communicated over the entire length of the frozen sample, the system can be expected to approach an end state, as described in Figure 4.1(B). In this case the thermally induced gradient of potential and the osmotic potential gradient cancel each other out; by definition the two must be exactly equal, since the same concentration of lactose solution used in the 'cold' reservoir, was also employed in the 'warm' reservoir. Thus,

⁴ Lactose in the soil capillaries tends to thaw ice, and suction falls in the unfrozen water (i.e. the total potential increases). However, this additional (osmotic) potential should not affect the hydraulic gradient since semi-permeable membranes, in the sense described above, obviously do not occur in the soil itself.

the total gradient of potential (within the unfrozen soil water) across the length of the sample is equal to zero. As the sample approaches this condition then, one would expect to observe a steady decline in mean rates of inflow and outflow. Eventually flow should stop altogether in accordance with the balance between the two potential gradients. This however, was not the case in any of the tests reported here. For example, Figure 4.2 shows cumulative inflow - outflow measured over the duration of a normal 6 hour experiment (Expt. 1B). The sample used was a Slims Valley Silt soil, subjected to a temperature gradient of $0.0944^{\circ}\text{C}/\text{cm}$, with a mean temperature of -0.220°C . It can be seen that inflow diminishes to zero with time, as might be expected. However, there is no apparent trend towards decreased outflow for the period in which readings were taken. Presumably the decline in observed rates of inflow is attributable to the reverse osmotic potential gradient at the 'warm' end. The absence of a similar trend with respect to rates of outflow requires further consideration. Two tentative and conflicting explanations suggest themselves.

Firstly, because the soil is frozen, the ends of the sample might be expected to be fairly autonomous. As a result, events taking place at one end will probably be transmitted across the entire length of the sample only very slowly. Since no identifiable trend towards decreasing outflow with time was observed, it could be argued that the

FIGURE 4.2 THE DISPARITY BETWEEN CUMULATIVE
INFLOW-OUTFLOW WITH TIME

effects of the reverse osmotic potential gradient are restricted to the 'warm' end. For the duration of the experiments described here (6 hours), it is not unreasonable to assume that the potential gradients developed within the sample remain essentially similar to those illustrated in Figure 4.1(A). If this is the case, then the measured values of outflow should be fairly indicative of flow rates to be expected under field conditions; of course rates of inflow are grossly unrepresentative due to the presence of the local osmotic potential gradient.

Alternatively, it is possible that the effects of the reverse osmotic potential gradient at the 'warm' end are communicated over the length of the sample almost immediately. Once the end state, described in Figure 4.1(B) is reached, then the total driving potential within the unfrozen soil water should approach zero. This condition however, does not hold with respect to the ice phase. By definition, the osmotic potential gradient is only relevant to the liquid phase, and as such its influence on the solid, ice phase can be expected to be minimal. It follows that the ice phase is still subject to a driving potential, in accordance with the temperature gradient. This implies movement of ice within the sample by regelation (see Section 1.4). Thus, it is possible that the observed outflow could be due, almost exclusively to ice movement taking place in the direction of decreasing temperatures (this hypothesis was suggested in

conversation with Dr. R.D. Miller). If this is the case, then the experiment may only be measuring one element (i.e. movement of ice by regelation) of the total flux that can be expected; obviously transport in the unfrozen water films must also be taken into account in any realistic estimate of flow rates induced by thermal gradients in frozen soil.

What in reality is probably happening is some combination of both scenarios. Thus, the counter osmotic potential gradient may be responsible for a small reduction in observed rates of outflow, while at the same time, transport within the ice phase may be emphasized over movement occurring in the films of unfrozen water. Figure 4.1(C) shows the potential gradients that might be expected to develop in a system with two different concentrations (appropriate for the temperatures of the cooling plates) of lactose solution in the reservoirs. The suggested test procedure is described in Section 2.3. As long as the correct temperature gradient is maintained across the sample, no counter osmotic potential gradient should be generated because the unfrozen water at both ends of the sample is in equilibrium with the two different concentrations of lactose solution in the reservoirs. With such an arrangement, one might expect much higher rates of inflow (due to the absence of the local osmotic potential gradient), so that cumulative inflow exceeds cumulative

outflow, the difference being attributable to ice formation within the sample. Outflow rates on the other hand, should be consistent with, or slightly greater than (for reasons discussed in the previous paragraph) the values achieved in the present study.

2. Water flow into the sample results in a local concentration of lactose in the inflow reservoir; the inverse causes dilution in the outflow reservoir. Thus, an additional osmotic potential gradient is generated over the duration of each test (6 hours). This tends to induce flow in the opposite direction to the gradient of potential implied by the temperature profile; a further decline in flow rates can be expected on this account. The effect is not additive since the reservoirs were flushed with a fresh supply of lactose solution at the beginning of each experiment.

Assuming perfect mixing in the reservoirs, the magnitude of the reverse osmotic potential generated can be calculated using the following equation:

$$\text{Osmotic potential} = R.T.(C_1 - C_2) \quad \text{Equation 13}$$

where R = universal gas constant

T = temperature, $^{\circ}\text{K}$

C_1, C_2 = lactose concentrations in the reservoirs, g-mol/L

The effect was investigated for the maximum and minimum rates of flow encountered during the course of the experiments (see Appendix V for full computations). The counter

osmotic potential generated by the maximum rate of flow was equivalent to 761.21 cm of water head, while the minimum rate of flow gave rise to an osmotic head of only 5.39 cm. These values were calculated to be 11.97% and 0.22% respectively, of the thermally induced gradients of potential. It should be remembered that the magnitude of the osmotic potential generated increases with time. Since both cumulative inflow-outflow values were selected at the end of the normal 6 hour test period, the calculations represent the most extreme conditions that can be expected to exist, assuming perfect mixing. It appears that the phenomenon can be ignored with respect to experiments that produced relatively small rates of flow (i.e. those with small temperature gradients and/or 'cold' mean temperatures). However, samples at relatively 'warm' mean temperatures, or with large temperature gradients may suffer a significant reduction in rates of flow. In this case, the phenomenon will serve to accentuate the effects of the reverse osmotic potential gradient discussed earlier.

4.2 Diffusion of Lactose Molecules Through the Dialysis Membranes

Ideally the solute used in the reservoirs should have large enough molecules not to pass through the semi-permeable membranes at the ends of the frozen sample. However, the small size of the lactose molecules means that they are able to pass, to some extent, through almost any membrane,

except those also impermeable to water, which, by definition are unsuitable for this experiment. Gradual movement of lactose molecules into the frozen soil was confirmed by the observation at the end of each experiment of a small thawed layer on both the inflow and outflow faces, normally in the order of 0.3 cm thick, but in extreme cases up to 0.8 cm thick. The problem increases with time, and for this reason tests were usually terminated after a period of two or three days.

The passage of lactose into the soil capillaries adds an osmotic potential which further lowers the freezing point of the unfrozen soil water. Consequently, some melting of ice can be expected to occur, which will result in an increase in the permeability of the thawed part of the sample. According to Burt and Williams (1976) however, the observed agreement when carbowax (Polyethelyene Glycol) and lactose were interchanged in an experiment with Leda clay indicates that the diffusion of the lactose has only a "minimal effect" on flow rates, and thus hydraulic conductivity determinations. This viewpoint is supported by the fact that the present author has repeated Burt and Williams experiment using supercooled water, and the results were essentially the same as those achieved with lactose solution in the reservoirs (see Figure 1.4).

The thawed sections of the sample can be expected to have a much higher permeability compared with that of the frozen soil. However, rates of flow will continue to be determined by the lower conductivities of the frozen portion, regardless of the presence of the unfrozen layers. In calculating hydraulic conductivity values a simple correction must be made for the presence of the thawed layers; the path length of flow L , is taken to be the length of the sample remaining frozen at the end of the experiment.

It could be argued that the observed fluxes are due in part to changes in the sample volume caused by pore ice melting, as a result of lactose infusion. Thawing of ice is associated with a 9% decrease in volume. Progressive contraction of the sample implies a slight tendency to inflow. Experiment No. 2A however, confirms that this was not the case (see Appendix III). The sample used was a Slims Valley Silt soil; both thermoelectric cooling plates were held at the same constant temperature of -0.325°C . Thus, the gradient of potential across the frozen sample was equal to zero. The reservoirs contained lactose solution, the concentration of which was adjusted to give the required freezing point depression. Despite the balance between potentials across the system, lactose continued to move into the sample by molecular diffusion. This was confirmed by the presence of thawed layers on either end of the sample, after running

the experiment for approximately 24 days. Once stable temperatures were achieved, cumulative inflow and outflow were observed to be negligible. In other words there was no tendency for water to move into the sample as a result of volume changes taking place at the thawing, inflow and outflow faces.

Notwithstanding thawed layers at both ends of the sample, it seems reasonable to attribute the outflow associated with applied temperature gradients to moisture migration occurring within the frozen soil. The relationship between temperature and free energy (Equation 4) does not apply above 0°C; consequently thermal gradients in unfrozen soil do not induce the same potential for water movement. It follows that the observed fluxes were most likely due to migration of water within the remaining frozen portion of the sample. Although the passage of lactose molecules into the frozen soil detracts from the theoretical beauty of the experiment, the above evidence seems to indicate that the phenomenon is of little practical import.

4.3 Pressures Generated Within the Sample Container

The possible build-up of frost heave pressures within the sample container must be taken into account, in any interpretation of the fluxes observed. Ice lensing, taking place towards the colder end may give rise to an increase in the frost heave pressures generated within the confined

sample. The subsequent rise in pore water pressures may be sufficient to counteract intake and reduce the observed rates of inflow. Verification of this hypothesis awaits modification of the present apparatus to facilitate measurement of the frost heave pressures generated within the sample container.

4.4 Error Analysis

The calculated rates of flow should be placed into some perspective as to their degree of accuracy. This involves consideration of the accuracy of each individual measurement used in the computation of the flow rates. The length and radius of the plexiglass sample container were known to within 0.0025 cm. Both capillary tubes were determined to have radii of 0.025 cm, with confidence limits of ± 0.002 cm. Similarly, timing the movement of menisci along the capillary tubes using a scale in conjunction with a magnifying glass and stopwatch was also very accurate. Perhaps the greatest source of error lies in the volumes of water contained in the connecting tubes; the following effects may have been responsible for some degree of variation:

1. The flexibility of the 'tygon' tubing used to connect the capillary tubes to the inflow-outflow reservoirs, introduces the possibility of volume changes occurring, other than those associated with water entering or leaving the frozen sample. Expansion or contraction of these tubes, for whatever reason (eg. thermally induced or even accidental

squeezing) will be translated into a positive or negative movement of the menisci in the capillary tubes. Such variance could produce a significant error, especially when one compares the total volume of water contained in the connecting tubes (approximately 9.65 cm^3) with the extremely small volumes of inflow and outflow being measured.

2. A small change in volume may occur as a result of air coming out of solution, in response to an increase in ambient temperature. This phenomenon can be expected to take place where the 'tygon' tubing passes out of the incubator (maintained at a mean temperature of 0°C) into air at room temperature; indeed air bubbles were commonly observed at this junction. The volume changes associated with a build up of air bubbles are likely to be only very small, but they may be sufficient to register in the capillary tubes as a slight positive movement of the menisci. Such movements might serve to accentuate outflow rates and cause an apparent decline in observed rates of inflow.

3. Movement of the menisci may also be influenced by small changes in volume caused by evaporation taking place from the ends of the capillary tubes. This is unlikely however, since the capillary tubes do not open directly to the air, but via a reservoir of water contained in a short section of 'tygon' tubing.

The estimated limit of accuracy of the temperature measurements was $\pm 0.01^\circ\text{C}$. Although the thermoelectric

cooling control system allows for much greater precision ($\pm 0.001^{\circ}\text{C}$), the accuracy of the temperature control was constrained by the tolerance of the thermistors employed. These were calibrated in a refrigerated 'Hotpack' bath circulator, against a mercury in glass thermometer, with a certified accuracy of $\pm 0.01^{\circ}\text{C}$ (N.R.C. calibration report No. APH 2173). Since the same thermistors were used for all the experiments, the relative differences between the recorded temperatures remained constant. The accuracy of the temperature measurements implies confidence limits of approximately $\pm 0.1245 \times 10^5 \text{Nm}^{-2}$ (or 124.5 cm of water head) when interpolating gradients of potential from the temperature profiles. This is a relatively small error (always less than 10%) when one considers the size of the heads induced by the applied thermal gradients. It is impossible to quantify the cumulative error ensuing from all these effects. In all likelihood however, the value will be considerable in absolute terms, but when viewed in the context of orders of magnitude (which this experiment is concerned with), it is probably quite small.

CHAPTER 5

SUMMARY AND CONCLUSIONS

1. Apparatus and procedures for the direct measurement of thermally induced water migration in saturated, frozen soils have been developed. The experimental configuration was similar to that used by Burt and Williams (1976), the fundamental difference being that the hydraulic gradient was created by a gradient of temperature rather than by a pressure differential. The frozen soil sample and reservoirs were sandwiched between two Peltier modules, controlled by a thermoelectric cooling control system. The reservoirs were normally filled with lactose solution, although supercooled water was used for two experiments. When a stable, linear temperature gradient was established across the system, water appeared to pass from the 'warm' reservoir into the frozen sample, and out into the second, colder reservoir.
2. Results obtained from the experiments have been examined with regard to current theoretical and applied studies of frozen ground. It is well known that frozen soils contain significant amounts of unfrozen water coexisting with the ice phase. This water has a potential relative to pure, bulk water, which is lower by an amount increasing as the temperature

decreases. Thus, a temperature gradient in frozen soil implies a flux of water in the direction of decreasing temperature, at a rate dependent on the permeability in the frozen state.

3. Fluxes were seen to vary, regardless of soil type, with the temperature gradient (i.e. the potential for water movement) and the mean temperature of the frozen sample (i.e. the apparent hydraulic conductivity). Observed rates of flow ranged from 10^{-1} to 10^{-3} mm³/min. These values are consistent with conductivity coefficients for frozen soils (at similar temperatures) measured earlier. Two soils were used for all the tests; since these had similar pore size distributions it proved impossible to identify variations in flow rate caused by soil type.

4. With supercooled water in the reservoirs, flow appeared to take place in the opposite direction to the hydraulic gradient implied by the temperature profile. A hypothesis was developed explaining this phenomenon in terms of the frost heave pressures generated within the frozen sample.

5. The data have some shortcomings, which came to light as the experiments progressed. The problem of reverse osmotic potential gradients in particular, complicates analysis of the results. The inherent properties of the test procedure adopted, could have significance with respect

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to rates of flow that might be expected in the field or with alternative experimental arrangements.

6. Although analysis of the results is made difficult by the complex thermodynamic situation, and the number of variables involved, the experiments as a whole demonstrate the migration of water within frozen soils, in response to temperature gradients. The apparatus should not be viewed strictly as a permeameter, rather the frozen soil, subjected to a temperature gradient acts as a kind of pump, inducing water migration in the direction of decreasing temperature. Calculation of hydraulic conductivity coefficients is possible using the Darcy Equation, but the apparatus is perhaps best regarded as providing a novel approach to conventional frost-heaving experiments.

5.1 Suggestions for Further Research

Modification should include monitoring the frost heave pressures generated within the sample container. Volume changes should be measureable, via an expansive, impermeable rubber membrane fitted over the outflow face of the sample, either with a pressure transducer or through displacement of liquid from the outflow reservoir. The application of counter-pressures on the outflow reservoir, sufficient to stop flow, would provide an indication in mechanical terms, of the temperature induced driving forces. Improvement of the test procedure should involve the use of two different

concentrations of lactose solution (corresponding to the temperatures of the cooling plates) in the reservoirs, for each temperature gradient investigated. Clarification of the influence of soil type on flow rates, awaits the testing of a large number of different soils, for which unfrozen water contents and grain size distributions are known.

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APPENDIX I

SCALE DRAWING OF THE EXPERIMENTAL APPARATUS

FIGURE I.1

CROSS-SECTION OF THE EXPERIMENTAL APPARATUS.

APPENDIX II

NOTES ON OPERATION OF THERMOELECTRIC

COOLING CONTROL SYSTEM

The thermoelectric cooling control unit is, in principle, simple in operation: it provides a thermostatic surface, with very accurate temperature control, to better than 0.02°C (Williams, 1968). In practice, however, its operation tends to be more complex, as the thermal properties of soils in contact with the cooling plate interact with the control system. The following procedures provide a brief introduction to the operation of the instrument, and obtain the best results.

(1) A simple 'cooling plate' consists of a small plate of Aluminum fixed to the 'cold' side of a Peltier module, the 'hot' side of which is attached to a heat sink, consisting of a hollow metal block through which tap water circulates.

(2) Mounted on, or preferably within the cold plate should be the control thermistor (to be led to the control unit), and another thermistor, or thermocouple, which is connected to a Multimeter, or other continuously observable temperature measuring device.

(3) Set the "gain" control knob at about the centre of its travel; do likewise with the fine and coarse temperature adjustment knobs. Set "bias" fully counterclockwise.

(4) Turn on the cooling water. Switch on the power of the control unit, the battery switch, and the channel switch; (the control thermistor lead should be inserted in the appropriate sockets, as should the two heavy leads to the module). The plate attached to the module should rapidly cool. (If it heats, reverse the module leads).

(5) High current initially passes, but after a minute or two the current is observed to fall off and then stabilize, by which time the temperature of the plate will be stable.

(6) Observe the current reading; switch off the battery; restore the current reading quickly to its original value, by means of the "bias" knob. Switch battery on again.

(7) The "gain" should be in the highest position possible without a slow oscillation of the current (observable on the meter).

After these adjustments (6 & 7) have been made the plate temperature will be somewhat different (than at 5), but is now controlled with the greatest stability. Make final adjustments to temperature with fine temperature adjustment knob. (In other words, the current supplied to the module has two components: a "carrying" or 'bias' current, which is determined by the setting of the 'bias' knob, and a current supplied through the wheatstone bridge (when the battery switch is on) and which adjusts the current to the module according to any deviation from the desired (set) temperature.)

To obtain a substantially different temperature repeat procedures 3 to 7, with new settings of temperature adjustment knobs. A series of values of temperature versus control settings can be obtained by repeating these procedures. This represents a calibration for the particular cooling plate arrangement.

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Because the temperature obtained, for a given setting of the temperature controls, is in some degree dependent on the thermal properties of the system being cooled, there is no unique calibration. For each experimental setup, the desired temperature is obtained by a final process of trial and error adjustments of the controls. For repetitive freezing of an entirely similar experimental setup, only a very small and rapidly performed adjustment is usually necessary.

Because of the unusual thermal properties of frozen soils, arising from the latent heat of fusion of ice being liberated or absorbed, it may take several hours for uniform temperatures to be obtained throughout a soil sample of average size. This is an effect of the thermal properties of the soil, and not the thermoelectric cooling apparatus. The adjustments described above may then only be carried out after such an interval of time. It should be noted that because the attainment of uniform temperatures throughout a body is a function of the material's thermal diffusivity, this will be achieved more rapidly the shorter the path from the most distant part of the sample to the cooling plate. A long cylinder will thus take substantially longer to cool than a short cylinder of soil on the plate. It is advantageous to have the gain setting as high as possible (see 7, above).

APPENDIX III

FREEZING POINT DEPRESSION, $\Delta^{\circ}\text{C}$, AS

A FUNCTION OF LACTOSE CONCENTRATION

Reference: "CRC Handbook of Chemistry and Physics",
Page D-218.

LACTOSE, $C_{12}H_{22}O_{11} \cdot H_2O$

Molecular weight = 342.30

Formula weight = 360.31

Anhydrous Solute Concentration g/l	Molar Concentration g-mol/l	Freezing Point Depression $\Delta^{\circ}C$
0.50	5.0	0.027
1.00	10.0	0.055
1.50	15.1	0.083
2.00	20.1	0.112
2.50	25.2	0.140
3.00	30.3	0.169
3.50	35.4	0.198
4.00	40.6	0.228
4.50	45.7	0.258
5.00	50.9	0.288
5.50	56.1	0.319
6.00	61.4	0.351
6.50	66.6	0.385
7.00	71.9	0.420
7.50	77.2	0.456
8.00	82.5	0.495
8.50	87.8	No value given
9.00	93.1	" " "
9.50	98.5	" " "
10.00	103.9	" " "

APPENDIX IV

A RECORD OF RESULTS OBTAINED

Note: Prefrozen samples were used throughout. The estimated limit of accuracy for temperature measurements was $\pm 0.01^{\circ}\text{C}$. Thermister code:

T_1 - End Plate temperature ("Cold")

T_2 - Temperature of sample 2.30 cms from T_1

T_3 - Temperature of sample 3.15 cms from T_1

T_4 - Temperature of sample 4.00 cms from T_1

T_5 - End Plate temperature ("warm"), 6.25 cms from T_1

All experiments were conducted using lactose solution in the reservoirs, except for numbers 5 and 6 where supercooled water was used. Flow normally occurred in the direction of decreasing temperature, outflow taking place from the "cold" end (T_1) and inflow at the "warm" end (T_5). Outflow is considered positive, and inflow negative. A positive sign in the inflow column denotes outflow from the "warm" end, and a negative sign in the outflow column denotes inflow at the "cold" end. All moisture contents are expressed as per cent dry weight.

EXPERIMENT NO. 1A, SLIMS VALLEY SILT

Time Interval (Min)	Temperature Gradient (°C)			Meniscus* (Distance traveled, mm)		Cumulative Flow (mm ³)	
	T ₁	T ₂	T ₃	T ₄	T ₅	Inflow (-ve)	Outflow (+ve)
840-870	-0.520/-0.525	-0.365/-0.380	-0.220/-0.235	-0.080/-0.085	+0.040/+0.055	2.0	0.5
870-900	-0.530	-0.390	-0.250	-0.090	+0.055	0.5	2.5
900-930	-0.535	-0.400	-0.260	-0.100	+0.055	0.0	4.0
930-960	-0.540	-0.405	-0.265	-0.105	+0.050	2.5	63.0
960-990	-0.540	-0.405	-0.265	-0.105	+0.050	16.0	72.5
990-1020	-0.540	-0.405	-0.270	-0.100	+0.055	79.0	188.5
1020-1050	-0.515	-0.380	-0.240	-0.075	+0.080	1.0	113.0
1050-1080	-0.515	-0.375	-0.240	-0.075	+0.080	0.0	207.5
1080-1110	-0.510	-0.375	-0.235	-0.070	+0.075	0.5	235.0
1110-1140	-0.515	-0.370	-0.235	-0.070	+0.080	0.0	239.0
1140-1170	-0.510	-0.375	-0.235	-0.070	+0.080	0.0	244.0
1170-1200	-0.505	-0.375	-0.240	-0.070	+0.080	0.0	209.0

*Radius of capillary tube = 0.025 cm

Initial Water Content: 29.01%

Temperature Gradient: 0.0896°C/cm

Mean Temperature: -0.240°C

Inflow Rate (Mean): $8.27 \times 10^{-2} \text{ mm}^3 \text{ min}^{-1}$, for period 840-1500 min

Outflow Rate (Mean): $5.77 \times 10^{-1} \text{ mm}^3 \text{ min}^{-1}$, for period 840-1500 min

EXPERIMENT NO. 1B, SLIMS VALLEY SILT

Time Interval (Min)	Temperature Gradient (°C)			Meniscus* (Distance travel- led, mm)		Cumulative Flow (mm ³)	
	T ₁	T ₂	T ₃	T ₄	T ₅	Inflow (-ve)	Outflow (+ve)
2310-2340	-0.530/ -0.515	-0.400/ -0.370	-0.255/ -0.255	-0.105/ -0.065	+0.035/ +0.075	6.0	228.5
2340-2370	-0.515	-0.375	-0.230	-0.070	+0.080	66.5	168.5
2370-2400	-0.515	-0.380	-0.235	-0.075	+0.080	169.0	213.0
2400-2430	-0.515	-0.380	-0.235	-0.070	+0.075	150.0	134.0
2430-2460	-0.515	-0.380	-0.240	-0.075	+0.075	97.5	13.0
2460-2490	-0.520	-0.380	-0.240	-0.070	+0.085	47.0	161.5
2490-2520	-0.515	-0.390	-0.250	-0.085	+0.070	39.5	296.5
2520-2550	-0.520	-0.390	-0.250	-0.090	+0.070	88.0	148.0
2550-2580	-0.515	-0.385	-0.250	-0.085	+0.070	228.0	160.0
2580-2610	-0.515	-0.385	-0.250	-0.085	+0.070	3.0	231.5
2610-2640	-0.515	-0.385	-0.250	-0.090	+0.070	0.0	122.5
2640-2670	-0.515	-0.385	-0.255	-0.090	+0.070	0.0	118.0
						175.66	391.77

*Radius of capillary tube = 0.025 cm

Initial Water Content: 29.01%

Temperature Gradient: 0.094°C/cm

Mean Temperature: -0.220°C

Inflow Rate (Mean): $2.44 \times 10^{-1} \text{ mm}^3 \text{ min}^{-1}$ for period 2310-3030 minOutflow Rate (Mean): $8.61 \times 10^{-1} \text{ mm}^3 \text{ min}^{-1}$ for period 2310-3030 min

EXPERIMENT NO. 2A, SLIMS VALLEY SILT

Time Interval (Min)	Temperature Gradient (°C)					Meniscus* (Distance travelled, mm)		Cumulative Flow (mm ³)	
	T ₁	T ₂	T ₃	T ₄	T ₅	Inflow (-ve)	Outflow (+ve)	Inflow (-ve)	Outflow (+ve)
1230-1260	-0.320/ -0.320	-0.375/ -0.375	-0.350/ -0.350	-0.315/ -0.315	-0.320/ -0.320	0.0	0.0	0.00	0.00
1260-1290	-0.325	-0.375	-0.355	-0.320	-0.330	0.0	0.0	0.00	0.00
1290-1320	-0.320	-0.380	-0.355	-0.320	-0.325	0.0	0.0	0.00	0.00
1320-1350	-0.325	-0.380	-0.355	-0.320	-0.335	0.0	0.0	0.00	0.00
1350-1380	-0.310	-0.375	-0.355	-0.315	-0.320	0.0	1.0	0.00	0.20
1380-1410	-0.325	-0.380	-0.350	-0.315	-0.330	0.0	0.5	0.00	0.29
1410-1440	-0.325	-0.375	-0.350	-0.315	-0.330	0.0	0.5	0.00	0.39
1440-1470	-0.315	-0.375	-0.350	-0.310	-0.325	0.0	0.5	0.00	0.49
1470-1500	-0.315	-0.370	-0.345	-0.310	-0.320	0.0	1.5	0.00	0.79
1500-1530	-0.320	-0.370	-0.345	-0.310	-0.330	0.0	1.0	0.00	0.98
1530-1560	-0.315	-0.365	-0.345	-0.310	-0.320	0.0	0.5	0.00	1.08
1560-1590	-0.325	-0.370	-0.340	-0.310	-0.330	0.0	0.5	0.00	1.18

*Radius of capillary tube = 0.025 cm

Initial Water Content: 24.29%

Temperature Gradient: 0.00°C/cm

Mean Temperature: 0.3275°C

Inflow Rate (Mean): 0.00 mm³min⁻¹, for period 1050-1770 minOutflow Rate (Mean): 1.64 x 10⁻³ mm³min⁻¹, for period 1050-1770 min

EXPERIMENT NO. 2B, SLIMS VALLEY SILT

Time Interval (Min)	Temperature Gradient (°C)					Meniscus* (Distance traveled, mm)		Cumulative Flow (mm ³)	
	T ₁	T ₂	T ₃	T ₄	T ₅	Inflow (-ve)	Outflow (+ve)	Inflow (-ve)	Outflow (+ve)
2460-2490	-0.300/ -0.325	-0.255/ -0.260	-0.170/ -0.175	-0.080/ -0.085	-0.005/ 0.000	+22.0	3.0	+4.32	0.59
2490-2520	-0.335	-0.265	-0.180	-0.090	-0.010	+10.0	0.5	+6.28	0.69
2520-2550	-0.325	-0.265	-0.185	-0.090	+0.010	+6.5	0.5	+7.56	0.79
2550-2580	-0.320	-0.270	-0.185	-0.090	+0.015	+0.5	2.0	+7.66	1.18
2580-2610	-0.340	-0.270	-0.185	-0.090	+0.015	+4.0	0.5	+8.44	1.28
2610-2640	-0.325	-0.270	-0.185	-0.090	+0.005	+10.0	3.0	+10.41	1.87
2640-2670	-0.320	-0.270	-0.190	-0.090	+0.005	+96.5	1.5	+29.36	2.16
2670-2700	-0.320	-0.270	-0.185	-0.090	0.000	0.0	46.5	+29.36	11.29
2700-2730	-0.315	-0.270	-0.185	-0.090	0.000	+0.5	3.0	+29.46	11.88
2730-2760	-0.320	-0.270	-0.185	-0.090	+0.010	+170.0	44.0	+62.84	20.52
2760-2790	-0.320	-0.270	-0.185	-0.090	+0.015	+1.0	91.5	+63.04	38.49
2790-2820	-0.320	-0.265	-0.185	-0.090	0.000	+1.0	2.5	+63.23	38.98

*Radius of capillary tube = 0.025 cm

Initial Water Content: 24.29%

Temperature Gradient: 0.0512°C/cm

Mean Temperature: -0.160°C

Inflow Rate (Mean): $+2.26 \times 10^{-1} \text{ mm}^3 \text{ min}^{-1}$, for period 2460-3180 minOutflow Rate (Mean): $9.85 \times 10^{-2} \text{ mm}^3 \text{ min}^{-1}$, for period 2460-3180 min

EXPERIMENT NO. 3A, CASTOR SILT LOAM

Time Interval (Min)	Temperature Gradient (°C)					Meniscus* (Distance travelled, mm)		Cumulative Flow (mm ³)	
	T ₁	T ₂	T ₃	T ₄	T ₅	Inflow (-ve)	Outflow (+ve)	Inflow (-ve)	Outflow (+ve)
2460-2490	-0.300/ -0.300	-0.260/ -0.280	-0.210/ -0.220	-0.185/ -0.190	-0.200/ -0.200	0.0	1.0	0.00	0.20
2490-2520	-0.305	-0.280	-0.225	-0.190	-0.205	0.0	0.5	0.00	0.29
2520-2550	-0.300	-0.280	-0.230	-0.195	-0.200	0.5	0.5	0.10	0.39
2550-2580	-0.305	-0.280	-0.230	-0.195	-0.200	1.0	0.0	0.29	0.39
2580-2610	-0.300	-0.285	-0.230	-0.195	-0.200	0.0	8.0	0.29	1.96
2610-2640	-0.305	-0.285	-0.230	-0.195	-0.200	0.5	1.0	0.39	2.16
2640-2670	-0.300	-0.285	-0.230	-0.195	-0.200	0.5	13.5	0.49	4.81
2670-2700	-0.300	-0.285	-0.230	-0.195	-0.195	0.0	0.0	0.49	4.81
2700-2730	-0.300	-0.280	-0.230	-0.195	-0.200	0.5	22.0	0.59	9.13
2730-2760	-0.305	-0.285	-0.230	-0.195	-0.205	0.0	2.0	0.59	9.52
2760-2790	-0.305	-0.280	-0.230	-0.195	-0.205	0.0	10.5	0.59	11.59
2790-2820	-0.300	-0.285	-0.230	-0.195	-0.200	14.0	11.0	3.34	13.75

*Radius of capillary tube = 0.025 cm

Initial Water Content: 47.87%

Temperature Gradient: 0.016°C/cm

Mean Temperature: -0.250°C

Inflow Rate (Mean): $9.27 \times 10^{-3} \text{ mm}^3 \text{ min}^{-1}$ Outflow Rate (Mean): $3.82 \times 10^{-2} \text{ mm}^3 \text{ min}^{-1}$

EXPERIMENT NO. 3B, CASTOR SILT LOAM

Time Interval (Min)	Temperature Gradient (°C)			Meniscus* (Distance traveled, mm)		Cumulative Flow (mm ³)	
	T ₁	T ₂	T ₃	T ₄	T ₅	Inflow (-ve)	Outflow (+ve)
3960-3990	-0.400/ -0.400	-0.355/ -0.355	-0.285/ -0.280	-0.230/ -0.230	-0.200/ -0.200	0.0	2.5
3990-4020	-0.400	-0.360	-0.285	-0.230	-0.195	0.0	6.5
4020-4050	-0.400	-0.355	-0.285	-0.230	-0.200	0.0	1.0
4050-4080	-0.400	-0.355	-0.285	-0.235	-0.205	0.0	1.5
4080-4110	-0.400	-0.355	-0.285	-0.230	-0.200	0.0	1.0
4110-4140	-0.405	-0.355	-0.285	-0.235	-0.210	0.0	0.5
4140-4170	-0.400	-0.355	-0.280	-0.230	-0.200	0.0	3.0
4170-4200	-0.400	-0.355	-0.285	-0.235	-0.200	0.0	1.5
4200-4230	-0.405	-0.360	-0.285	-0.230	-0.205	0.0	3.5
4230-4260	-0.400	-0.355	-0.285	-0.230	-0.200	0.0	2.0
4260-4290	-0.400	-0.355	-0.285	-0.230	-0.200	0.0	1.0
4290-4320	-0.400	-0.355	-0.280	-0.230	-0.205	0.0	1.0

*Radius of capillary tube = 0.025 cm

Initial Water Content: 47.87%

Temperature Gradient: 0.032°C/cm

Mean Temperature: -0.300°C

Inflow Rate (Mean): 0.00 mm³min⁻¹Outflow Rate (Mean): 1.36 x 10⁻² mm³min⁻¹

EXPERIMENT NO. 4A, CASTOR SILT LOAM

Time Interval (Min)	Temperature Gradient (°C)					Meniscus* (Distance travelled, mm)		Cumulative Flow; (mm ³)	
	T ₁	T ₂	T ₃	T ₄	T ₅	Inflow (-ve)	Outflow (+ve)	Inflow (-ve)	Outflow (+ve)
0-30	-0.500/-0.500	-0.125/-0.145	-0.095/-0.115	-0.060/-0.080	-0.205/-0.205	-	-	-	-
30-60	-0.500	-0.165	-0.115	-0.090	-0.205	+389.5	25.5	+76.49	5.01
60-90	-0.500	-0.185	-0.085	-0.050	-0.205	+330.0	48.0	+141.29	14.43
90-120	-0.500	-0.230	-0.105	-0.065	-0.200	+250.0	34.0	+190.39	21.11
120-150	-0.500	-0.265	-0.115	-0.070	-0.200	+188.5	61.0	+227.40	33.09
150-180	-0.500	-0.290	-0.125	-0.075	-0.205	+241.0	12.0	+274.73	35.45
180-210	-0.500	-0.310	-0.140	-0.080	-0.200	+203.0	19.5	+314.60	39.28
210-240	-0.500	-0.325	-0.155	-0.085	-0.200	+5.0	44.0	+315.58	47.92
240-270	-0.505	-0.340	-0.170	-0.095	-0.200	+5.5	5.5	+316.66	49.00
270-300	-0.505	-0.355	-0.185	-0.105	-0.200	+5.0	1.5	+317.64	49.30
300-330	-0.500	-0.360	-0.195	-0.110	-0.200	+290.5	6.0	+374.68	50.47
330-360	-0.500	-0.370	-0.210	-0.125	-0.205	+118.0	108.5	+397.86	71.78

*Radius of capillary tube = 0.025 cm

Initial Water Content: 41.33%
 Temperature Gradient: 0.048°C/cm (approach to linearity)
 Mean Temperature: -
 Inflow Rate (Mean): 1.21 mm³min⁻¹
 Outflow Rate (Mean): 2.18 x 10⁻¹ mm³min⁻¹

EXPERIMENT NO. 4B, CASTOR SILT LOAM

Time Interval (Min)	Temperature Gradient (°C)			Meniscus* (Distance travelled, mm)		Cumulative Flow (mm ³)	
	T ₁	T ₂	T ₃	T ₄	T ₅	Inflow (-ve)	Outflow (+ve)
1050-1080	-0.500/-0.500	-0.445/-0.445	-0.325/-0.325	-0.235/-0.235	-0.200/-0.205	0.0	3.0
1080-1110	-0.500	-0.440	-0.325	-0.235	-0.200	0.0	2.5
1110-1140	-0.505	-0.445	-0.325	-0.235	-0.205	0.5	1.5
1140-1170	-0.500	-0.445	-0.325	-0.235	-0.200	0.0	3.0
1170-1200	-0.500	-0.445	-0.325	-0.235	-0.200	0.0	2.5
1200-1230	-0.500	-0.440	-0.325	-0.235	-0.205	0.0	0.5
1230-1260	-0.505	-0.445	-0.325	-0.235	-0.205	+0.5	1.0
1260-1290	-0.500	-0.445	-0.325	-0.235	-0.205	0.0	1.0
1290-1320	-0.505	-0.440	-0.325	-0.235	-0.205	0.0	2.5
1320-1350	-0.505	-0.440	-0.325	-0.235	-0.200	0.0	3.5
1350-1380	-0.500	-0.445	-0.325	-0.235	-0.200	0.0	1.0
1380-1410	-0.500	-0.440	-0.325	-0.235	-0.200	0.0	6.5
						0.0	5.60

*Radius of capillary tube = 0.025 cm

Initial Water Content: 41.33%

Temperature Gradient: 0.048°C/cm

Mean Temperature: -0.350°C

Inflow Rate (Mean): 0.00 mm³min⁻¹Outflow Rate (Mean): 1.55 x 10⁻² mm³min⁻¹

EXPERIMENT NO. 4C, CASTOR SILT LOAM

Time Interval (Min)	Temperature Gradient (°C)			Meniscus* (Distance travelled, mm)		Cumulative Flow (mm ³)	
	T ₁	T ₂	T ₃	T ₄	T ₅	Inflow (-ve)	Outflow (+ve)
1440-1470	-0.600/-0.600	-0.480/-0.495	-0.340/-0.355	-0.240/-0.255	-0.205/-0.200	+12.0	16.5
1470-1500	-0.600	-0.505	-0.360	-0.255	-0.200	+7.5	13.0
1500-1530	-0.600	-0.505	-0.365	-0.260	-0.205	+0.5	16.0
1530-1560	-0.600	-0.505	-0.370	-0.265	-0.200	+1.5	2.5
1560-1590	-0.600	-0.505	-0.370	-0.265	-0.200	+0.5	5.0
1590-1620	-0.600	-0.505	-0.370	-0.265	-0.200	+0.5	4.5
1620-1650	-0.600	-0.505	-0.370	-0.265	-0.200	+0.5	2.5
1650-1680	-0.600	-0.510	-0.370	-0.265	-0.200	+0.5	0.5
1680-1710	-0.605	-0.510	-0.370	-0.265	-0.205	0.5	3.0
1710-1740	-0.600	-0.505	-0.370	-0.265	-0.200	0.0	0.5
1740-1770	-0.600	-0.510	-0.370	-0.265	-0.200	0.0	0.0
1770-1800	-0.600	-0.505	-0.370	-0.265	-0.200	0.0	1.5

*Radius of capillary tube 0.025 cm

Initial Water Content: 41.33%

Temperature Gradient: 0.064°C/cm (approach to linearity)

Mean Temperature: -

Inflow Rate (Mean): +1.25 x 10⁻² mm³min⁻¹Outflow Rate (Mean): 3.57 x 10⁻² mm³min⁻¹

EXPERIMENT NO. 4D, CASTOR SILT LOAM

Time Interval (Min)	Temperature Gradient (°C)				Meniscus* (Distance travel- led, mm)		Cumulative Flow (mm ³)	
	T ₁	T ₂	T ₃	T ₄	T ₅	Inflow (-ve) Outflow (+ve)	Inflow (-ve) Outflow (+ve)	
2490-2520	-0.600/ -0.600	-0.510/ -0.510	-0.370/ -0.370	-0.265/ -0.265	-0.200/ -0.200	0.5 0.5	0.10 0.10	0.29 0.29
2520-2550	-0.600	-0.510	-0.370	-0.265	-0.205	0.0	0.10	0.88
2550-2580	-0.600	-0.505	-0.370	-0.265	-0.205	0.5	0.20	2.06
2580-2610	-0.600	-0.505	-0.375	-0.265	-0.200	0.5	0.10	2.55
2610-2640	-0.600	-0.510	-0.370	-0.265	-0.205	0.0	0.10	3.34
2640-2670	-0.605	-0.510	-0.375	-0.265	-0.205	0.0	0.10	3.53
2670-2700	-0.605	-0.515	-0.375	-0.265	-0.200	0.5	0.20	3.93
2700-2730	-0.600	-0.510	-0.375	-0.265	-0.200	1.0	0.39	6.48
2730-2760	-0.605	-0.510	-0.370	-0.265	-0.200	0.0	0.39	6.97
2760-2790	-0.600	-0.510	-0.370	-0.265	-0.200	0.0	0.39	7.36
2790-2820	-0.600	-0.510	-0.370	-0.265	-0.200	0.0	0.39	8.64
2820-2850	-0.600	-0.510	-0.370	-0.265	-0.200	0.0	0.39	9.03

*Radius of capillary tube = 0.025, cm

Initial Water Content: 41.33%
 Temperature Gradient: 0.064°C/cm
 Mean Temperature: -0.400°C
 Inflow Rate (Mean): 1.09 x 10⁻³ mm³min⁻¹
 Outflow Rate (Mean): 2.51 x 10⁻² mm³min⁻¹

EXPERIMENT NO. 5A, CASTOR SILT LOAM
(Supercooled Water in Reservoirs)

Time Interval (Min)	Temperature Gradient (°C)			Meniscus* (Distance traveled, mm)		Cumulative Flow (mm ³)	
	T ₁	T ₂	T ₃	T ₄	T ₅	Inflow (-ve)	Outflow (+ve)
4290-4320	-0.305/ -0.300	-0.200/ -0.210	-0.135/ -0.135	-0.090/ -0.085	-0.105/ -0.095	+169.0	+172.0
4320-4350	-0.305	-0.215	-0.135	-0.090	-0.100	+51.5	-49.5
4350-4380	-0.300	-0.220	-0.140	-0.090	-0.095	+111.5	-138.9
4380-4410	-0.300	-0.220	-0.140	-0.090	-0.095	+137.0	-131.0
4410-4440	-0.300	-0.225	-0.140	-0.090	-0.095	+137.5	-131.0
4440-4470	-0.300	-0.230	-0.145	-0.090	-0.095	+157.0	-68.5
4470-4500	-0.305	-0.230	-0.145	-0.090	-0.105	+124.0	-101.0
4500-4530	-0.300	-0.230	-0.150	-0.090	-0.100	+30.0	-109.0
4530-4560	-0.300	-0.235	-0.155	-0.090	-0.105	+176.0	-115.5
4560-4590	-0.305	-0.240	-0.155	-0.095	-0.105	+148.5	-106.0
4590-4620	-0.300	-0.240	-0.155	-0.095	-0.100	+140.5	-15.0
4620-4650	-0.305	-0.245	-0.160	-0.095	-0.105	+96.0	-102.0

*Radius of capillary tube = 0.025 cm

Initial Water Content: 43.78%
 Temperature Gradient: 0.032°C/cm
 Mean Temperature: -0.200°C
 Inflow Rate (Mean): $+7.93 \times 10^{-1} \text{ mm}^3 \text{ min}^{-1}$
 Outflow Rate (Mean): $-6.67 \times 10^{-1} \text{ mm}^3 \text{ min}^{-1}$

EXPERIMENT NO. 5B, CASTOR SILT LOAM
(Supercooled Water in Reservoirs)

Time Interval (Min)	Temperature Gradient (°C)					Meniscus* (Distance travel- led, mm)		Cumulative Flow (mm ³)	
	T ₁	T ₂	T ₃	T ₄	T ₅	Inflow (-ve)	Outflow (+ve)	Inflow (-ve)	Outflow (+ve)
5370-5400	-0.300/ -0.300	-0.215/ -0.230	-0.140/ -0.155	-0.080/ -0.090	-0.100/ -0.100	+51.0	1.0	+10.02	0.20
5400-5430	-0.310	-0.255	-0.175	-0.105	-0.105	+93.5	-4.0	+28.38	-0.59
5430-5460	-0.300	-0.265	-0.185	-0.110	-0.100	+99.5	-31.0	+47.92	-6.68
5460-5490	-0.305	-0.265	-0.185	-0.115	-0.105	+91.5	-65.0	+65.88	-19.44
5490-5520	-0.290	-0.250	-0.170	-0.105	-0.095	+68.5	-71.5	+79.34	-33.48
5520-5550	-0.295	-0.250	-0.175	-0.105	-0.105	+69.5	-125.5	+92.98	-58.13
5550-5580	-0.300	-0.260	-0.185	-0.115	-0.105	+97.0	-1.0	+112.03	-58.32
5580-5610	-0.305	-0.270	-0.185	-0.115	-0.105	+16.5	-8.5	+115.27	-59.99
5610-5640	-0.300	-0.260	-0.185	-0.110	-0.105	+56.5	-22.5	+126.37	-64.41
5640-5670	-0.305	-0.260	-0.185	-0.115	-0.105	+64.0	-50.0	+138.94	-74.23
5670-5700	-0.295	-0.270	-0.190	-0.115	-0.095	+71.0	-58.0	+152.88	-85.62
5700-5730	-0.290	-0.255	-0.180	-0.110	-0.095	+60.5	-97.5	+164.76	-104.77

*Radius of capillary tube 0.025 cm

Initial Water Content: 43.78%
 Temperature Gradient: 0.032°C/cm
 Mean Temperature: -0.200°C
 Inflow Rate (Mean): +4.58 x 10⁻¹ mm³min⁻¹
 Outflow Rate (Mean): -2.91 x 10⁻¹ mm³min⁻¹

EXPERIMENT NO. 6A, SLIMS VALLEY SILT
(Supercooled Water in Reservoirs)

Time Interval (Min)	Temperature Gradient (°C)				Meniscus* (Distance traveled, mm)		Cumulative Flow (mm ³)	
	T ₁	T ₂	T ₃	T ₄	T ₅	Inflow (-ve) Outflow (+ve)	Inflow Outflow (-ve) (+ve)	
1320-1350	-0.095/ -0.095	-0.075/ -0.075	-0.085/ -0.085	-0.065/ -0.065	-0.100/ -0.100	+0.5 -0.5	+0.10 -0.10	
1350-1380	-0.100	-0.075	-0.085	-0.065	-0.100	0.0 -79.5	+0.10 -15.71	
1380-1410	-0.100	-0.080	-0.085	-0.065	-0.095	0.0 -2.5	+0.10 -16.20	
1410-1440	-0.100	-0.080	-0.085	-0.065	-0.100	+3.0 -1.0	+0.69 -16.40	
1440-1470	-0.085	-0.080	-0.085	-0.065	-0.100	+1.0 0.0	+0.88 -16.40	
1470-1500	-0.100	-0.080	-0.085	-0.070	-0.095	+6.0 -0.5	+2.06 -16.49	
1500-1530	-0.100	-0.090	-0.110	-0.075	-0.095	+3.5 -2.0	+2.75 -16.89	
1530-1560	-0.100	-0.085	-0.105	-0.090	-0.100	+3.5 -62.0	+3.44 -29.10	
1560-1590	-0.105	-0.080	-0.090	-0.070	-0.095	0.0 -60.5	+3.44 -40.94	
1590-1620	-0.100	-0.085	-0.085	-0.065	-0.100	0.0 -24.0	+3.44 -45.65	
1620-1650	-0.095	-0.085	-0.085	-0.065	-0.100	2.0 -47.7	+3.04 -54.98	
1650-1680	-0.100	-0.085	-0.090	-0.070	-0.100	0.0 0.0	+3.04 -54.98	

*Radius of capillary tube = 0.025 cm

Initial Water Content: 29.92%
Temperature Gradient: 0.00°C/cm
Mean Temperature: -0.100°C
Inflow Rate (Mean): $+8.44 \times 10^{-3} \text{ mm}^3 \text{ min}^{-1}$
Outflow Rate (Mean): $-1.15 \times 10^{-1} \text{ mm}^3 \text{ min}^{-1}$

EXPERIMENT NO. 6B, SLIMS VALLEY SILT
(Supercooled Water in Reservoirs)

Time Interval (Min)	Temperature Gradient (°C)			Meniscus* (Distance travel- led, mm)		Cumulative Flow (mm ³)	
	T ₁	T ₂	T ₃	T ₄	T ₅	Inflow (-ve)	Outflow (+ve)
8310-8340	-0.200/ -0.205	-0.175/ -0.175	-0.165/ -0.170	-0.120/ -0.125	-0.100/ -0.105	+ 3.5	-0.10
8340-8370	-0.200	-0.160	-0.150	-0.105	-0.105	0.0	-0.59
8370-8400	-0.200	-0.160	-0.145	-0.105	-0.095	+48.5	0.10
8400-8430	-0.205	-0.175	-0.160	-0.115	-0.100	0.0	0.10
8430-8460	-0.205	-0.175	-0.165	-0.125	-0.100	+ 1.0	0.10
8460-8490	-0.205	-0.180	-0.175	-0.125	-0.100	0.0	0.00
8490-8520	-0.200	-0.165	-0.145	-0.105	-0.100	0.0	-0.79
8520-8550	-0.205	-0.175	-0.160	-0.115	-0.100	0.0	-0.79
8550-8580	-0.205	-0.180	-0.170	-0.125	-0.105	1.0	-0.88
8580-8610	-0.205	-0.180	-0.170	-0.125	-0.105	0.0	-2.75
8610-8640	-0.200	-0.165	-0.155	-0.110	-0.095	0.5	-4.12
8640-8670	-0.200	-0.170	-0.160	-0.120	-0.100	0.0	-5.20

*Radius of capillary tube = 0.025 cm

Initial Water Content: 29.92%

Temperature Gradient: 0.016°C/cm

Mean Temperature: -0.150°C

Inflow Rate (Mean): +2.02 x 10⁻² mm³min⁻¹, for period 8310-8790 min

Outflow Rate (Mean): -2.67 x 10⁻² mm³min⁻¹, for period 8310-8790 min

APPENDIX V

CALCULATION OF THE REVERSE OSMOTIC GRADIENT,

ASSUMING PERFECT MIXING

(i) For the worst situation (i.e. the maximum outflow observed)

Expt. No. 1B

$$\text{Osmotic Head} = R.T.(C_1 - C_2)$$

where R = universal gas constant

$$T = T^{\circ}K$$

C_1, C_2 = reservoir lactose solution concentrations,
g-mol-L

$$\begin{aligned} \text{Volume of water passing into the soil over 6 hours} \\ = 0.176 \text{ cm}^3 \end{aligned}$$

$$\begin{aligned} \text{Volume of water leaving the soil over 6 hours} \\ = 0.392 \text{ cm}^3 \end{aligned}$$

$$\text{Volume of end reservoirs} \approx 4.0 \text{ cm}^3$$

$$\text{Temperature of 'cold' reservoir} = -0.515^{\circ}C$$

$$\text{Temperature of 'warm' reservoir} = +0.075^{\circ}C$$

$$\text{Lactose concentration in both reservoirs} \quad 82.5 \text{ g/L}$$

$$\therefore \text{ in } 4.0 \text{ cm}^3 \approx 0.33 \text{ gm of lactose}$$

Following passage of water:

$$\begin{aligned} \text{Inflow reservoir concentration} &= \frac{(0.33 \times 100)}{4 - 0.176} \\ &= 8.629\% \end{aligned}$$

$$\begin{aligned} \text{Outflow reservoir concentration} &= \frac{(0.33 \times 100)}{4 + 0.392} \\ &= 7.514\% \end{aligned}$$

$$\text{Inflow concentration} \approx 0.246980152 \text{ g-mol-L}$$

$$\text{Outflow concentration} \approx 0.214104776 \text{ g-mol-L}$$

$$\therefore \text{Difference in Concentration} = 3.288 \times 10^{-2}$$

$$\left[\begin{array}{l} \text{Mean Temperature} \\ = -0.220^{\circ}\text{C} \\ = 272.93\text{K} \end{array} \right]$$

$$\begin{aligned} \text{Osmotic head} &= RT(C_1 - C_2) = (33.4 \times 2.54) \\ &\quad \times 3.288 \times 10^{-2} \times 272.93 \\ &\approx 761.21 \text{ cm water head} \end{aligned}$$

Water head deduced from the temperature gradient = 6,358 cm

$\therefore$ the reverse osmotic potential gradient = 11.79% of the total thermal driving gradient.

(ii) For the best situation (i.e. the minimum outflow observed)

Expt. No. 3B

Using similar calculations

Volume of water passing into the soil over 6 hours
= 0.00 cm³

Volume of water leaving the soil over 6 hours
= 4.909×10^{-3} cm³

Volume of end reservoirs ≈ 4.0 cm³

Temperature of 'cold' reservoir = -0.4°C

$$\left[\begin{array}{l} \text{Mean Temperature} \\ = -0.300^{\circ}\text{C} \\ = 272.85\text{K} \end{array} \right]$$

Lactose concentration in both reservoirs ≈ 66.6 g/L

$\therefore$ in 4.0 cm³ ≈ 0.266 gm of lactose

Following passage of water:

$$\text{Inflow reservoir concentration} = \frac{(0.266 \times 100)}{4 - 0.0} = 6.66\%$$

$$\begin{aligned} \text{Outflow reservoir concentration} &= \frac{(0.266 \times 100)}{4 + 4.909 \times 10^{-3}} \\ &= 6.652\% \end{aligned}$$

Inflow concentration ≈ 0.18986842 g-mol-L

Outflow concentration ≈ 0.189635663 g-mol-L

∴ Difference in concentration = 2.328×10^{-6}

$$\begin{aligned}\text{Osmotic head} &= RT(C_1 - C_2) = (33.4 \times 2.54) \times 2.328 \\ &\quad \times 10^{-4} \times 272.85 \\ &\approx 5.39 \text{ cm water head}\end{aligned}$$

Water head deduced from the temperature gradient = 2,469 cm

∴ the reverse osmotic potential gradient = 0.22%
of the total thermal driving potential.

END

2 3 1 1 2 1 8 1 1

FIN