THE UNIVERSITY OF MANITOBA

SOIL TEMPERATURE ESTIMATION FROM METEOROLOGICAL MEASUREMENTS

by

AB REIMER

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ABSTRACT

Soil temperature is important in agriculture, ecology and engineering. The measurement of soil temperature is expensive, time consuming and, under certain conditions, it may be almost impossible (permafrost). It seemed timely and appropriate, therefore, to seek an efficient prediction method based on readily available meteorological information.

Field soil temperature measurements were analysed to determine the soil temperature climate for a five year period in a well drained old field under forage and zero tillage conditions at the Whiteshell Nuclear Research Establishment (WNRE) at Pinawa, Manitoba. Concurrent meteorological measurements were similarly analysed to identify significant relationships, with a view to developing realistic and efficient soil temperature prediction techniques.

A literature search on soil temperature modelling techniques was conducted. The Gibbs free energy approach to soil heat and water transport was studied, as well as the non-equilibrium thermodynamics of heat transfer due to a concentration gradient (thermo-osmosis) and mass transfer due to a thermal gradient (thermo-filtration).

Emperical simulation was used as an expedient solution to the soil temperature prediction problem. Monthly mean soil surface temperatures were estimated for summer and winter months

from regression equations with meteorological predictors.

Daily mean soil surface temperatures were predicted from regression equations with meteorological predictors combined with Fourier-series best fit seasonal curves. Daily mean subsoil temperatures at 10 cm were estimated from predicted surface temperatures by applying an appropriate damping factor. The standard deviation of the difference between predicted and observed was generally less than 1°C ($p=99\%$) for daily and monthly estimates.

A good estimate of the seasonal subsoil temperature cycle at 10 to 200 cm was found from a periodic function with damping and phase parameters. The explained variance of this function was 95% or more. With appropriate assumptions regarding soil thermal properties and mean annual soil temperature, accurate results were obtained quickly and economically.

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1. INTRODUCTION

Soil temperature is important in plant growth, and in related areas such as soil fertility and microbial activity. Soil water movement, plant transpiration and surface evaporation are strongly related not only to soil temperature but also to temperature gradients. Soil temperature measurement is not only time consuming and expensive, but it can also be very difficult under certain conditions (saturated or frozen soils). Therefore the need is clear for an economical and useful soil temperature prediction method based on readily available information.

Ouellete (1973) has pointed out that although there are numerous air temperature measuring stations in Canada (1920), the number of soil temperature measuring sites is much smaller (58). Therefore a soil temperature prediction model based on air temperatures and other meteorological measurements can only broaden this information base. Such a prediction scheme based on careful analysis of adequate data can expand soil temperature information, both in detail and in extent, to cover most regions in Canada.

A sequence of soil temperatures, like air temperatures, is dominated by two periodicities, namely the annual cycle and the diurnal cycle, at least in this mid-latitude climate.

The annual cycle is usually confined to the top ten meters of soil and the diurnal cycle is contained in the top one meter of soil. Short period and somewhat irregular fluctuations lasting anywhere from a few seconds to a few hours often occur in daytime in the top few millimeters of soil under certain weather conditions. Periodicities of the order of a few days corresponding to synoptic scale meteorological features are important in a daily temperature prediction scheme and are considered here. Temperature fluctuations of a few years and longer associated with natural, climatic or geophysical changes, and the heat flux through the Earth's crust due to the geothermal gradient will not be considered directly. However all periodicities contribute to the actually measured cycle of soil temperatures and so are implicit in emperical studies even though the individual contribution may be small and may not be considered explicitly.

The mean annual soil temperature cycle at the Whiteshell Nuclear Research Establishment (WNRE) at Pinawa, Manitoba (1968 to 1972) has an amplitude of about 11^{o}C at the surface (Figures 1-3). This falls to zero in about the first ten meters or less of soil. The rate at which the surface temperature wave is damped with depth depends on the thermal properties of the soil.

The mean diurnal cycle of soil temperature at the surface has an amplitude of 4 to 5^{o}C in summer (Figure 4). This decreases exponentially with depth, and becomes neglegible in the first meter or less of soil. In winter, under snow cover, the diurnal temperature amplitude at the soil surface is insignificant for most practical purposes, amounting to a couple of tenths of a degree Celsius in January (see Figure 6).

In the present study Fourier analysis is used to obtain expressions for these long term average annual and diurnal cycles. Departures from these values, related to meteorological factors, are expressed by regression equations. These expressions are then applied in combination or separately to predict daily and monthly surface and subsoil temperatures and annual subsoil temperatures waves.

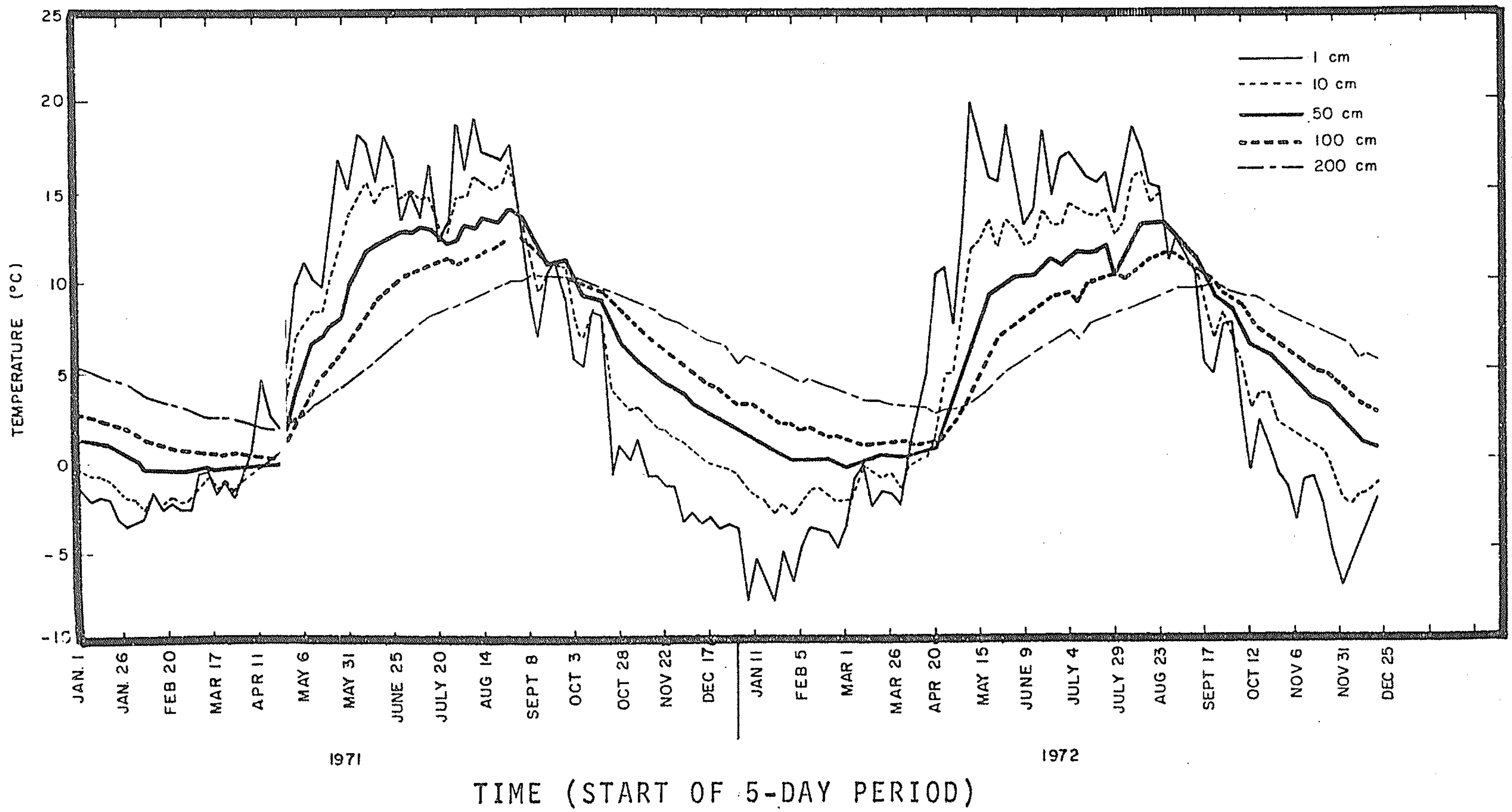


Figure 1. Observed five-day average soil temperatures at 1, 10, 50, 100 and 200 cm under grass (zero tillage) (snow) at WNRE.

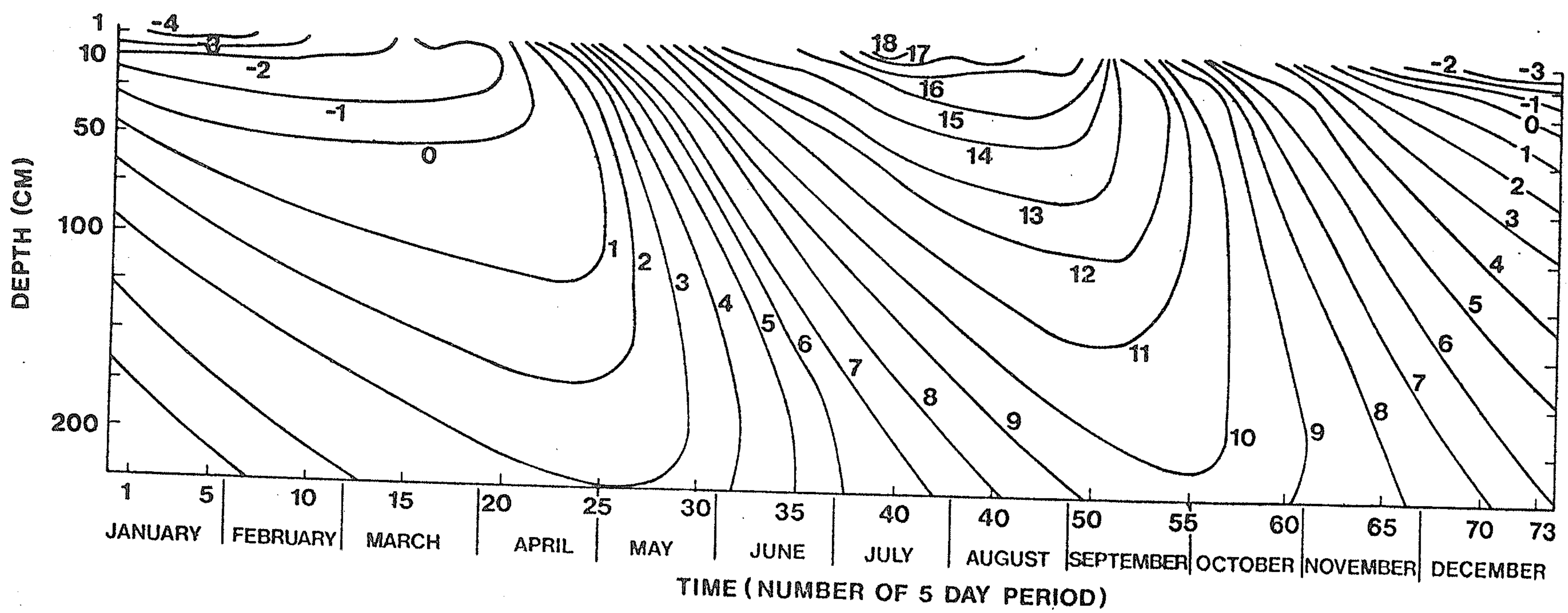


Figure 2. Five year mean (1968-1972) vertical time section of observed soil temperatures under grass (snow) in a well drained old field under zero tillage conditions at WNRE.

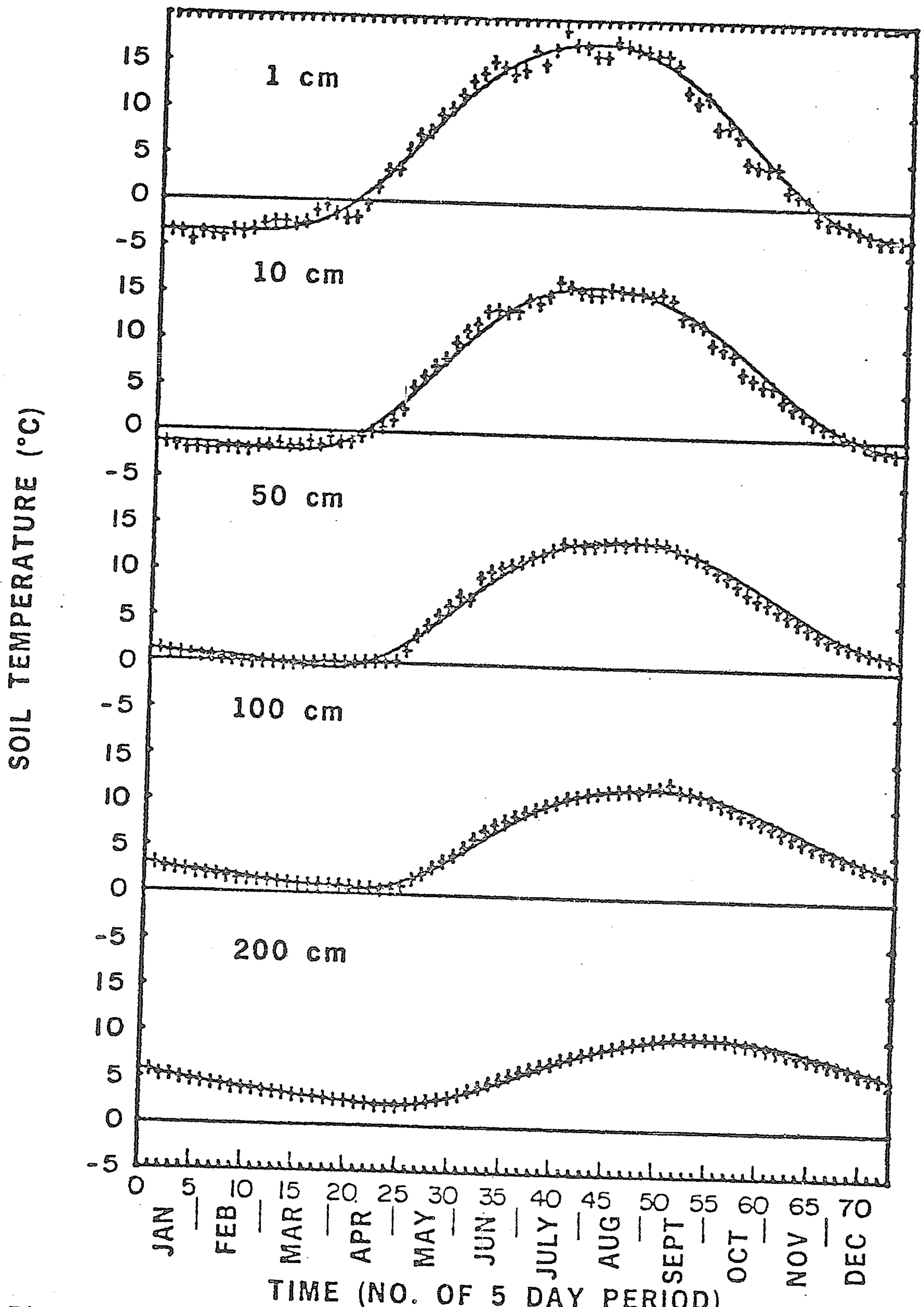


Figure 3. Observed (+) average annual soil temperature waves at five levels at WNRE (1968-1972) and best-fit Fourier-series curves (——).

2. LITERATURE REVIEW

2.1 Phenomenological Relations

Field soil temperatures vary continually, particularly at the surface. This results in varying temperature gradients. Such temperature gradients give rise to heat flux and also water flux. Simultaneous flow of two or more entities such as heat and moisture leads to interference effects and the production of essentially new flow phenomena. Many such effects have been observed and can be delineated by the so called phenomenological relations.

Temperature gradients can cause heat transfer in soil by conduction, convection and radiation, in order of decreasing importance.

Radiation heat transfer is negligible in water saturated soil and in soil where the air filled voids are water vapour saturated. In dry soils where the pore space air is unsaturated, long wave radiative heat transfer may be significant. Near the soil surface, of course, short and long wave radiation can participate in the overall heat budget (Rosenberg 1974), particularly under dry conditions and with strong solar insolation. Radiative heat flux in soil will not be dealt with specifically here, but only as it is expressed implicitly in measured soil temperatures and in heat flux transactions in general.

Convection is generally negligible except during irrigation, heavy rainfall or strong wind conditions (Scotter and Raats 1969). Conduction is the dominant heat transfer process in solid and liquid fractions of soil.

Bouyoucos (1915) obtained evidence of moisture movement in soil in response to a thermal gradient, and he found that net transfer was from warm to cold regions. Moisture flow can be in the liquid or vapour phase or both and sometimes in opposite directions. Bouyoucos (1915) introduced the air gap in soil columns to distinguish between liquid and vapour flow.

Liquid phase flow was quantified by using a tracer such as chloride ion, either natural or added. Gurr, Marshall and Hutton (1952) assumed that Cl^- movement is due solely to liquid water convection. They found a net transfer of water from warm to cold regions and a net transfer of Cl^- from cold to warm. This was rationalized as water vapour moving from warm to cold regions and liquid water moving down a soil water potential gradient from cold to warm. Taylor and Cavazza (1954), Jackson et al (1965), Weeks et al (1968) and Jackson, Rose and Penman (1965) have reported similar results with various Cl^- salts and this phenomenon in closed soil columns has been referred to as a circular convection process (Gurr et al 1952, Chang 1976) or a vapour-liquid circulatory flow (Jackson, Rose and Penman 1965).

Coupled flow of heat and moisture in soils was also studied by Philip and de Vries (1957), Cary and Taylor (1962), Rose (1968a,b), Dirksen (1969), Cassel, Nielsen and Biggar (1969), Jury and Miller (1974), Kay and Groenevelt (1974) and Chang and Cho (1974). The gradient of molecular vibration i.e., "hot" and "cold" molecules produced by a thermal gradient gives rise to thermal selection at, for example, a water-ice interface. Three such selection sites are recognized: (1) the liquid water-air interface, (2) the liquid water-matrix interface and (3) the liquid water-ice interface. Surface tension is implicit in the liquid pressure caused locally by temperature gradients.

A consistent theory of non-equilibrium phenomena incorporating both Onsager's reciprocity theorem and an explicit entropy term was formalized in the 1940's by Meixner (1941) and Prigogine (1947) (in de Groot and Mazur 1963).

In irreversible thermodynamics three assumptions or postulates must be recognized. One is the assumption of 'local equilibrium', which states that on a small or micro-scale the relationships of equilibrium thermodynamics apply. The second assumption states that if entropy production can be written as

$$\Phi = \sum_i G_i F_i \qquad (1)$$

where G and F are the fluxes and forces in question then the fluxes G_i are linear homogeneous functions of the forces F_i, so that

$$G_i = \sum_j L_{ij} F_j \qquad (2)$$

where the phenomenological coefficients L_{ij} are independent of the forces. Assumption three states that if one and two are true the matrix of phenomenological coefficients can be put in symmetric form, i.e.,

$$L_{ij} = L_{ji} \tag{3}$$

This is Onsager's reciprocal relation. The validity or usefulness of this relation has been questioned, perhaps mainly because the proper choice of fluxes and forces largely determines whether Onsager's relation is satisfied (Raats 1975, Nerpin 1975). Furthermore the above assumptions also prescribe that the domain of non-linear relations and higher order symmetries cannot be treated by this approach (Srivastava and Abrol 1966).

For the flow of water and heat (energy) equations (2) and (3) can now be written:

$$\begin{aligned} j_w &= L_{ww}\Delta\psi/T + L_{wq}\Delta T/T^2 \\ j_q &= L_{qw}\Delta\psi/T + L_{qq}\Delta T/T^2 \\ L_{wq} &= L_{qw} \end{aligned} \tag{4}$$

where w and q are water and heat respectively, ψ is water potential, T is temperature and L are the cross-linked coefficients, and the symbol Δ has the meaning of gradient in equation (4) only (de Groot and Mazur 1963).

In completely frozen soil columns both water and solute move from warm to cold regions (Cary and Mayland 1972) and water movement was found to be comparatively rapid (Dirksen and Miller 1966, Hoekstra 1966). Liquid water movement as a continuous unfrozen water film was hypothesized. The thickness of such a film decreases with temperature (Anderson et al 1973).

The rapidity of flow in a frozen soil may be related to the fact that the potential of ice is independent of the presence of soil (Hoekstra 1966, Groenevelt and Kay 1974) and no equilibrium moisture content can be reached in contrast to the steady state that is reached in unfrozen soil.

This unlimited capacity of frozen soil to take up water is likely to account for the very dry region often found next to frozen soil which is amenable to vapour flow. Liquid flow in response to a water potential gradient is likely to bring water to this dry region (Dirksen and Miller 1966) from the warm or unfrozen side.

2.2 Water Vapour Transport

Kay and Groenevelt (1974), Groenevelt and Kay (1974) and Raats (1975) have dealt theoretically with the question of water transport as a result of a temperature gradient (thermo-osmosis) and heat transport as a result of a water potential gradient (thermo-filtration) in both frozen and unfrozen porous systems, with assumptions concerning gravity, solute, streaming potential, rigid matrix and pressure.

They showed from theoretical considerations that in unfrozen systems liquid water moves from cold to warm regions and water vapour moves from warm to cold regions. For vapour movement in response to a thermal gradient (thermo-osmosis) from the generalized Clapeyron equations and the Gibbs-Duhem relation

$$\nabla\tilde{p} = (Hv/\bar{V}v)(\nabla T/T) \qquad (5)$$

where $\overline{V}_v$ is partial specific volume of water vapour, $\nabla\tilde{p}$ is the vapour pressure gradient caused by the temperature gradient ∇T ($^oK\ cm^{-1}$), H_v is partial specific latent heat (erg/g) of vaporization and T is temperature (^oK). Vapour flux caused by this vapour pressure gradient is

$$j_v = -k_v\nabla\tilde{p} \tag{6}$$

where k_v is vapour conductivity and, in terms of liquid water, this is

$$j_l = -k_v(H_v/\overline{V}_v)(\nabla T/T) \tag{7}$$

Referring to the phenomenological equation (4)

$$j_l = -L_{wq}\nabla T/T - L_{ww}\overline{V}_l^e\nabla\overline{p} \tag{8}$$

where L_{wq} and L_{ww} are the phenomenological coefficients of water transport due to a temperature gradient and water transport due to a concentration gradient respectively, $\overline{V}_l^e$ is extramatric partial specific volume of liquid water and $\nabla\overline{p}$ is the presure gradient in the extramatric liquid water. When $\nabla\overline{p} = 0$

$$j_l = -L_{wq}\nabla T/T$$

and from equation (7) the cross-coefficient is

$$L_{wq} = k_vH_v/\overline{V}_v \quad (g\ cm^{-1}s^{-1}) \tag{9}$$

and since this coefficient is positive, flow is down the temperature gradient, i.e., from warm to cold. In this and following developments the notation of Groenevelt and Kay (1974) is used, which is, $\sim$, $-$, $\wedge$ for vapour, liquid and ice respectively.

In frozen soil for a temperature gradient without an ice pressure gradient ($\nabla\hat{p} = 0$) by a similar argument,

$$\nabla\tilde{p} = (H_s/\overline{V}_v)(\nabla T/T) \qquad (10)$$

where H_s is partial specific latent heat of sublimation. The vapour pressure gradient associated with the thermal gradient will cause vapour flux and, expressed in terms of ice, we have

$$\hat{L}_{wq} = (k_v/\overline{V}_v)H_s \,(g\ cm^{-1}s^{-1}) \qquad (11)$$

For vapour phase flow in the air filled pore space, which is usually assumed continuous, k_v is often estimated from (Kay and Groenevelt 1974)

$$k_v = \tilde{\epsilon}K_V^o/\lambda RT \qquad (12)$$

where $\tilde{\epsilon}$ is air filled porosity

K_V^o is diffusion coefficient of water vapour in air ($=0.24\ cm^2s^{-1}$)

λ is tortuosity

R is the gas constant for water vapour ($4.62 \times 10^6 erg\ g^{-1}K^{-1}$)

T is temperature (^oK)

k_v is apparent vapour conductivity coefficient.

For unsaturated soil liquid water is usually visualized as rings or islands of water between soil particles (Philip and de Vries 1957). Water vapour may condense on such a water island on the cold side of a pore. The pressure of surface tension and wetting will then adjust the island to accommodate the additional water. To continue the process water may then evaporate from the down gradient side of the island into the next pore, and so on. This adjustment may take place more rapidly than the diffusion flux of the vapour, thus suggesting an acceleration or enhancement term. This may be one possible explanation for the often observed very rapid movement of water vapour, greatly exceeding Fickian diffusion, in soil.

It is a well known fact that in a mixture of gases if a temperature gradient exists, a concentration gradient must also exist (Chapman and Dootson 1917, in Grew and Ibbs 1952).

$$\partial n_1/\partial \mathbf{r} = -k_T(1/T)(\partial T/\partial \mathbf{r}) \qquad (13)$$

for a binary mixture where n_1 is the concentration of molecular species 1, T is temperature in ^oK, k_T is the thermal diffusion ratio and $\mathbf{r}$ is the vector distance (in the notation of Grew and Ibbs 1952 $\partial/\partial\mathbf{r} = \nabla$). If k_T is assumed constant, integration will give

$$n_1 - n_1' = k_T \ln (T'/T) \qquad (14)$$

where n_1 and n_1' are concentrations at temperature T and T' respectively. This difference in composition was first demonstrated by Chapman and Dootson (1917). The value n_1-n_1' is called the separation.

The degree of separation and thermal diffusion ratio depend in a complex manner on

(1) ratio of masses and diameters of the species

(2) nature of forces between like and unlike molecules and

(3) relative proportions of the species, n_1 and n_2.

The value of k_T, the thermal diffusion ratio, can be calculated for a mixture of molecules of a given mass and size in known proportions if they are assumed to interact as rigid elastic spheres and k_T is maximum for this assumption. Therefore comparison of measured $k_{T(m)}$ vs calculated $k_{T(c)}$ thermal diffusion ratios is a measure of the nature of the actual interaction between molecules. For example, the $k_{T(m)}/k_{T(c)}$ for a hydrogen-nitrogen mixture is 0.6 and for a nitrogen-carbon dioxide mixture it is 0.3. The separation difference observed here is partly due to the difference in mass ratios and partly due to the "softness" of the nitrogen-carbon dioxide interaction.

The elementary theories of thermal diffusion state that the heavier component (bigger) of a mixture should diffuse down the temperature gradient if $\nu>5$ (Jost 1960). This is seen from an expression for transfer of momentum for a binary mixture

$$\overline{(m_2 - m_1)v^{(\nu-5)/(\nu-1)}} \tag{15}$$

where m_1 and m_2 are momenta of molecules 1 and 2, v is relative velocity, ν is the exponent of the repulsive force and the bar indicates averaging over all velocities.

Thus thermal diffusion vanishes if $\nu=5$ (Frankel 1940 in Grew and Ibbs 1952) and is reversed if $\nu<5$. This is in accordance with the exact theory for Maxwelliam molecules (Jost 1960). Since $\nu>5$ for a nitrogen-water mixture, water molecules will diffuse to the hot region. If this is the case for an air-water mixture, it is immediately seen that thermal diffusion is opposite in direction to the observed flux of water vapour in soils under a temperature gradient.

2.3 Liquid Water Transport

A theoretical treatment of liquid water movement involving the heat of wetting effect demands a distinction between free (extramatric) water and bound (matric) water. There is no distinction between extramatric and matric water in the solid state (ice) or in the vapour state. However liquid matric water must be regarded as having many states, each one a function of the distance from the solid surface.

For liquid phase flow (Groenevelt and Kay 1974) in response to a thermal gradient (thermo-osmosis), assuming the extramatric liquid pressure gradient is zero ($\nabla \bar{p}^e = 0$), the temperature gradient is accompanied by a local matric liquid pressure gradient (see equation 5),

$$\nabla \bar{p} = (\bar{H}_w / \bar{V}_1) \nabla T / T \tag{16}$$

where $\bar{H}_w$, the partial specific heat of wetting, is a function of distance from the solid matrix. Now assuming that the liquid is in horizontal layers of thickness b ($0 \leq \xi \leq b$) and flow is assumed laminar, steady state velocity caused by $\nabla \bar{p}$ is governed by

$$\nabla \bar{p} = -d/d\xi \left[\eta(\xi) \, (d/d\xi) \, v_1 \right] \tag{17}$$

where η is viscosity and v_1 is local liquid velocity. Integrating twice

$$v_1 = \nabla T/T \int_0^{\xi'} 1/\eta \int_{\xi''}^{b} \bar{H}_w / \bar{V}_1 \quad d\xi' \, d\xi'' \tag{18}$$

Flux through a liquid film can be obtained by integration of $v_1/\bar{V}_1$ from $\xi = 0$ to $\xi = b$ giving

$$j_1 = (\nabla T/T) \, \epsilon/\lambda b \int_0^b 1/\bar{V}_1 \int_0^{\xi'} 1/\eta \int_{\xi''}^{b} \bar{H}_w / \bar{V}_1 \quad d\xi' d\xi'' \, d\xi \tag{19}$$

This equation (19) estimates macroscopic liquid flux if multiplied by $\epsilon/\lambda b$ to allow for the effects of water-filled porosity (ϵ) and tortuosity (λ). Then from equations (4) and (19), as in section (2.2)

$$L_{wq} = -\epsilon/\lambda b \int_0^b 1/\overline{V}_1 \int_0^{\xi'} 1/\eta \int_{\xi''}^b \overline{H}_w/\overline{V}_1 \quad d\xi' d\xi'' \, d\xi \qquad (20)$$

This coefficient is negative for positive $\overline{H}_w$, thus flow is from cold to warm.

The terms $1/\overline{V}_1$, $1/\eta$ and $\overline{H}_w/\overline{V}_1$ as functions of ξ for matric water must be known, however, to evaluate this coefficient. Since these functions have not been rigorously defined to date, other methods will be used in this study.

If the viscosity (η) of matric water is greater than the viscosity of free water, expected flow will be less than that for extramatric water. This effect could be a maximum in clay soils near saturation. Phillips and Brown (1968) found the self-diffusion coefficients for tritiated water increased linearly as the number of water layers on each mineral surface increased. However, they found the diffusion coefficients for montmorillonite with about 5 layers to be the same ($\sim 6\text{x}10^{-6}$ cm^2s^{-1}) as kaolinite with about 22 layers. Their explanation is that a longer pathlength of diffusing water molecules in the kaolinite and a smaller relative mobility of the diffusing water molecules in the montmorillonite counteract each other. This apparent paradox may relieve the problem of identifying water flow rates in clays of different texture. Nakayama and Jackson (1963) found the "apparent" diffusion coefficient of tritiated water in loam soil to be nearly constant for water content (θ) of 40 to 10%, rising to a very sharp peak at 4% water of about 3.5 times the 40 to 10% value ($1.5 \times 10^{-5} cm^2/s$). Corey and Horton (1968) found the diffusion rates of ^2H, ^3H and ^{18}O tagged water nearly equal in water-saturated acidic kaolinite (Vaucluse) soil.

Recent evidence suggests that heat of wetting is the only cause of coupled flow in the liquid phase (Groenevelt and Kay 1974, Jury and Miller 1974). If the value of $(\partial\psi/\partial T)_\theta$ for the range of moisture contents (θ) in question can be measured or estimated for the soil in question, L_{wq} can be estimated from (Jury and Miller 1974);

$$L_{wq} \cong -k_1(\partial\psi/\partial T)_\theta \cong -k_1 \; \Delta\psi/\Delta T \qquad (21)$$

If $\Delta\theta/\Delta T$ and $\Delta\psi/\Delta\theta$ are known, $\Delta\psi/\Delta T$ can be found from

$$\Delta\psi/\Delta T = (\Delta\theta/\Delta T)(\Delta\psi/\Delta\theta) \qquad (22)$$

The unsaturated hydraulic conductivity, k_1, can be reliably estimated for most soils (Shaykewich 1970).

The cross-coefficients (L_{wq} and L_{qw}) described above together with the ordinary coefficients (L_{ww} and L_{qq}) via the phenomenological equations provide a way of estimating heat and water transfer through the soil. Moisture transfer is required at each step to recalculate the physical properties and the same is true of temperature.

Liquid flow in frozen soil again involves three integrations to obtain the velocity term and to account for the multi-phase water hypothesis to give

$$\hat{L}_{wq} = L_{wq}(\bar{H}_w) + L_{wq}(H_f) \qquad (23)$$

where H_f is partial specific heat of fusion. The $L_{wq}(H_f)$ may be expressed by $H_f L_{ww}$ (g $cm^{-1}s^{-1}$).

Because heat of fusion is homogenous for all liquid water in porous media

$$j_1 = -L_{ww}V_i^e \nabla p = -L_{ww}H_f(\nabla T/T) \qquad (24)$$

and as above

$$L_{wq}(H_f) = H_f L_{ww}(g\ cm^{-1}s^{-1}) \qquad (25)$$

Both vapour and liquid in terms of ice flow in frozen soil is from warm to cold. Liquid phase heat of fusion flow seems to dominate. Heat of wetting is generally not significant (<1%) in frozen soil.

In a similar way the coefficients for heat flux due to water concentration gradients (thermofiltration, L_{qw}) can be obtained. Groenevelt and Kay (1974) found the corresponding coefficients of thermo-osmosis and thermo-filtration to be equal, which is in agreement with Onsager's relation (equation 3). The key to this relation, however, is largely dependent on the proper choice of fluxes and forces and/or the proper transformations. Raats (1975) has illustrated this point with the theorem of Meixner, and it is evident that for certain applications the Onsager relations may not always be satisfied. However, further developments, especially for systems that are too far from equilibrium, may facilitate successful application of non-linear phenomenological laws (de Jong 1967, Srivastava and Abrol 1966).

2.4 Heat Transfer in Soil

From the phenomenological laws (equation 2,4) it is seen that in the absence of a water potential gradient ($\Delta\psi$=0) the 'ordinary' equation for heat transfer is obtained, which is an approximation:

$$J_q = -k_q \Delta T/\Delta z \qquad (26)$$

where J_q is heat flux density, k_q is thermal conductivity (cal cm^{-1} s^{-1} $^{o}C^{-1}$) and $\Delta T/\Delta z$ is the one dimensional thermal gradient in the vertical or z direction. Thus this equation is a special case of the phenomenological laws (equation 4) when $\Delta\psi=0$. It is valid for saturated, frozen or totally dry soils where there is no movement of water due to a temperature gradient.

This is the steady-state equation (26) for heat conduction in solids for which homogeneous and isotropic conditions are usually assumed. Field soils generally are not homogeneous and there is evidence that they may not always be isotropic, especially clay soils. Furthermore, steady-state does not occur very often in field soils, particularly near the surface.

The continuity equation for one dimensional heat conduction (Taylor and Ashcroft 1972) is

$$\rho c_p \ \partial T/\partial t = -\partial J_q/\partial z \qquad (27)$$

Combining equations (26) and (27) in the differential form gives

$$\partial T/\partial t = 1/\rho c_p \ \partial/\partial z \ (k_q \ \partial T/\partial z) \qquad (28)$$

and if k_q is independent of temperature we have

$$\partial T/\partial t = K \ \partial^2 T/\partial z^2 \qquad (29)$$

where ρ is bulk density, c_p is gravimetric specific heat for a soil system and $K = k_q/\rho c_p$. This is the non-steady-state differential form of the heat transfer equation often attributed to Fourier.

Of the thermal properties implicit in K, the term ρc_p, often called the volumetric heat capacity C_v, can be calculated by the method of de Vries (1966),

$$C_v = C_s X_s + C_w X_w + C_a X_a \tag{30}$$

where C_s, C_w and C_a are volumetric specific heats and X_s, X_w and X_a are volume fractions of solid, water and air respectively. Soil thermal conductivity (k_q) is also a function of solid, water and air fractions. Soil texture was at one time thought to affect conductivity, but when soil water is expressed in potential terms the effect of texture largely disappears, or at least is minimized. The density and gravimetric specific heat of individual soil particles vary somewhat, depending on the mineral type and crystal structure. However, these properties, taken on a macroscopic scale, which is large compared to particle size, tend to lose their individual identity in the soil aggregate, with the net result that moisture content emerges as a strong factor in the definition of soil thermal properties. In the development of a soil model it is important, therefore, to include soil moisture in each heat transfer calculation, as will be shown later.

3. MODEL DEVELOPMENT

3.1 Soil Temperature

A periodic function f(x) with a period of 2π(rad) can be represented by a trigonometric series,

$$f(x) = a_o + \sum_{n=1}^{\infty}(a_n \cos nx + b_n \sin nx) \tag{31}$$

The right hand side is called a Fourier series and the Fourier coefficients (a_o, a_n, b_n) can be calculated from the Euler formulas:

$$\begin{aligned} a_o &= 1/2\pi \int_{-\pi}^{\pi} f(x)\,dx \\ a_n &= 1/\pi \int_{-\pi}^{\pi} f(x) \cos nx\,dx \\ b_n &= 1/\pi \int_{-\pi}^{\pi} f(x) \sin nx\,dx \end{aligned} \tag{32}$$

Such a series, if it converges, can be used to represent diurnal and annual soil and air temperature fluctuations (Kreyszig 1968).

At the soil surface (z = 0) the temperature as a function of time F(t) can be represented by:

$$F(0,t) = a_m + \sum_{n=1}^{\infty}\left[a_n \cos(2\pi/P)nt + b_n \sin(2\pi/P)nt\right] \tag{33}$$

where a_m is the mean temperature, n is the number of the harmonic, t is time, and P is period. The angular frequency is $2\pi/P$. This equation can be put into an equivalent form that is more convenient for subsequent development.

$$F(0,t) = a_m + \sum_{n=1}^{\infty} A_{on} \sin((2\pi/P)nt + \phi_n) \tag{34}$$

where the relations between constants are as follows:

$$a_n = A_{on}\sin\phi_n$$

$$b_n = A_{on}\cos\phi_n$$

$$A_{on}^2 = a_n^2 + b_n^2$$

$$\phi_n = \mathrm{ARCTAN}(a_n/b_n)$$

For application to a semi-infinite solid equation (33) is the boundary condition for equation (29) at $z = 0$, and at infinite depth ($z = \infty$) the temperature is assumed constant and equal to a_m. Since the surface temperature is assumed to be a periodic function for all time, no separate initial condition is required (for $t = -\infty$).

From section 2.4 (Heat Transfer in Soil) we have equation (29),

$$\partial T/\partial t = K(\partial^2 T/\partial_z{}^2)$$

the heat conduction equation valid for one dimensional heat flow in a homogeneous medium. Van Wijk and de Vries (1966) have obtained the following expressions:

$$F(z,t) = a_m + A_o \exp(-z/D)\sin(2\pi/P)t - z/D) \tag{35}$$

and

$$D = (2k/C\omega)^{\frac{1}{2}} = (2K/\omega)^{\frac{1}{2}} \tag{36}$$

where k is thermal conductivity and C is volumetric heat capacity. The amplitude A_o is multiplied by a factor $\exp(-z/D)$ and there is a phase shift of $-z/D$. The damping depth, D, is a function of diffusivity (K) and period ($1/\omega$). The solution of Eq. (35) satisfies (29) when D is given by (36) if the negative root is excluded. A more general solution of (29) is a superposition of expressions like (35) for periodic variations that are not sine functions.

$$F(z,t) = a_m + \sum_{n=1}^{\infty} A_{on}\exp(-z\sqrt{n}/D)\sin((2\pi/P)nt + \phi_n - z\sqrt{n}/D) \tag{37}$$

The boundary conditions of (29) are satisfied at $z = 0$ and $z = \infty$ by F from (37). Such an expression may be used to model soil temperature at any depth as amplitude change and phase shift are accommodated (See Figure 16).

The smoothed subsoil temperature regime can be expressed if the physical properties of thermal conductivity, specific heat, density or damping depth are known for the diurnal or annual period. Conversely, if the subsoil temperature regime is known from experimental measurement, the damping depth and physical properties may be calculated. In this way a smoothed value of D over a given depth and time interval can be obtained.

For a periodic temperature variation at the surface with a frequency $\omega = 2\pi/P$, equation (35) can give diffusivity for the soil layer from z_1 to z_2 (for details see van Wijk 1966), thus:

$$K = \pi/P\left(\frac{z_1 - z_2}{l_n(A_1/A_2)}\right)^2 \qquad (38)$$

$$K = \pi/P\left(\frac{z_1 - z_2}{(\phi_2 - \phi_1)}\right)^2 \qquad (39)$$

where P is the period of the fundamental harmonic
z_1, z_2 are measurement levels (cm)
A_1, A_2 are amplitudes (^oC)
ϕ_1, ϕ_2 are phase constants (rad)
K is thermal diffusivity ($cm^2 s^{-1}$).

This method is valid for higher harmonics if required. If the soil is homogeneous, the temperature gradient is one dimensional and the surface variation is sinusoidal, (38) and (39) should give the same result (Figure 3,4, Table 13). Diffusivities were calculated with five-day average values since daily measurements contain too much random variation.

Fourier coefficients were obtained with the IBM subroutine FORIT (Ralston and Wilf 1960).

As seen in the diurnal and annual cycles (Figure 3,4), the amount of scatter about the best fit Fourier-series curve decreases with depth. Thus, the best potential for improved temperature estimation is in the soil layers nearest the surface. Since the top layers are also the most important for many ecological and physical applications, predicted departure from the long-term at the surface must be utilized.

Equations (35) to (39) contain frequency (ω) or period ($1/\omega$) terms, and the identification of periodicities other than the obvious diurnal or annual cycle requires harmonic or spectrum analysis. Meteorological spectra have been obtained by Taylor (1938), Sutton (1953), Pasquill (1962), Panofsky and Brier (1968) and Lumley and Panofsky (1964). Improved methods have recently been described by Schickedanz and Bowen (1977) and Rikiishi (1976). In the present study a CDC 6600 computer with the IMSL (1977) program FTFREQ was used (see Appendix 1).

3.2 Soil Temperature Simulation

Soil temperature simulation, to be successful, must be based on an understanding of long term temperature fluctuations, i.e., the soil temperature climate.

Stoller and Wax (1973), McDole and Fosberg (1974a, b), Leger and Millette (1975), Hay (1976) and Wildsmith (1976) have described annual and diurnal fluctuations at various depths and under a variety of conditions, for a particular location or in a particular area. Ouellet (1973) and Ouellet and Desjardins (1975) have analysed temperature records for numerous locations in Canada and have produced what might be called a soil temperature climatology for most agricultural areas in Canada.

Bonham and Fye (1970) used emperical methods to model soil temperature, because they found heat transfer equations 'inappropriate'. They argued that these equations are valid only if the soil is homogeneous, and if the physical properties are contant in time and space. In cultivated soils, where they worked, these conditions are not met and so the heat-transfer equations are not strictly valid for practical application. Hasfurther and Burman (1974) used an emperical method based on air temperature to predict the soil surface temperature, and a one-dimensional heat flow equation to calculate subsoil temperature (assuming homogeneous and constant conditions).

Theoretical methods have been described by Lettau (1962) Wierenga and de Wit (1970), Goudrian and Waggoner (1972), Hadas and Fuchs (1973) and Palagin (1976) which allow for variation of physical properties. Palagin assumes that all meteorological characteristics reduce to a surface boundary condition. Wierenga and de Wit found good results in wet soil but significant error in dry soil, with the greatest error in the late morning, that is, during the strongest temperature gradients.

Multi-layer models using numerical (integration) techniques can become unstable under certain conditions and may oscillate. Significant truncation error may be produced after a large number of iterations, if they converge too slowly (Parlange 1971, Philip 1975). Simplified theoretical models have been devised to avoid these problems (MacKinnon 1976) but generally there is some loss of accuracy and generality.

For the sake of economy in computer and programming time, a different approach can be taken. The surface boundary condition can be defined in terms of equation (33) as forcing function, with Fourier coefficients based on long-term average annual or diurnal temperature cycles. Departures from these long-term mean values are then related to meteorological variables by statistical regression analysis. For monthly mean values, five years of data does not provide a sufficiently large sample size to give stable and truely representative regression coefficients (Walpole 1968). Climatologically speaking, the only remedy is to collect more data. However, from a numerical or statistical point of view, this situation can be improved by pooling the data for adjacent months with the month in question (Table 1, 2 and 3). This was reasonably successful.

Monthly mean soil surface temperatures can thus be predicted from the forcing function and regression equations (Table 4 to 9).

Daily soil surface temperature estimation is dependent on the monthly mean. The annual forcing function is used to define the shape of the trend line for the month and the mean is adjusted to the estimate above. Daily departures from this mean based on regression equations from daily meteorological and soil measurements produce the final predicted value for any particular day (Figures 7 to 14).

Daily meteorological observations contain what may be called random fluctuations, at least, for the scale of this study, and therefore, some smoothing is indicated. Weighted smoothing functions (.25, .50, .25) are used (Panofsky and Brier 1968) on daily air temperatures and daily hours of bright sunshine (low-pass filtering).

Surface soil temperature prediction in winter, with snow on the ground, poses the problem of dealing with an intervening layer between the atmosphere and the soil. Technically this requires the introduction of a two layer model, with complete delineation of the physical and physico-chemical properties of this layer, snow. An abbreviated method was developed to accommodate the snow layer by applying the damping and phase shift factors directly to the input data of the model. This requires a knowledge of the periodicity of the particular fluctuation in question. Apart from annual and diurnal fluctuations, a somewhat less regular but nevertheless identifiable cycle of 2 to 4 days was found by spectrum analysis of meteorological variables, as described earlier. This is in general agreement with the findings of Misra (1971), Reimer et al (1974), Tilley and McBean (1973). The damping factor

$$\exp(-z/D)$$

where

$$D = (2K/\omega)^{\frac{1}{2}}$$

was used with $1/\omega$ the period set equal to two days, with good results. Periodicities of a few days, which are associated with synoptic scale meteorological events, seem to be the only significant short-term control of soil temperature under snow since the diurnal cycle appears to be almost non-existent (Figure 6).

Subsoil temperatures may be estimated in a similar way by using the appropriate forcing function based on equation (37) and by applying a departure from the mean as for surface soil, but with proper corrections for damping and phase shift, for daily values (see Figure 15). Once a good surface annual soil temperature wave has been established equation (37) can also be used to estimate subsoil annual temperature waves with appropriate damping and phase shift values from estimated diffusivities (K) or damping depths (D) (Figure 16).

4. METHODS

4.1 Site

The experimental site is in a well drained old field that was farmed briefly 15 or 20 years ago. It is on flat terrain about 500 meters from the Winnipeg River (Reimer 1966) (s20,T14,Rge 11E). The soil was classified as a Whitemouth Series clay loam by Smith et al (1967). Under the System of Soil Classification for Canada (1974) the soil would be classed as an orthic dark grey Luvisol. Soil samples from the experimental site show a particle size distribution of 20,30 and 50% sand, silt and clay respectively. Organic matter content is about 10% in the surface layer (18cm) and near zero below that level. Vegetation consists of meadow grasses and other forage plants growing in uncultivated soil, and the plant canopy is uncut except for trimming of the tall weeds once or twice a year.

A geological description by Mills and Zwarich (1970) mentions four basic units overlying bedrock; an upper lacustrine silt unit (2-5 m), a lower lacustrine clay unit, glacial till and a sandy deposit on Precambrian bedrock, for a total of 10 to 20 meters.

4.2 Measurements

Soil temperatures were measured with a stack of platinum resistance thermometers fastened to a wooden support at 1, 10, 50, 100 and 200 cm. The platinum sensor coils inside the end of thin metal support tubes extended horizontally about 40 cm into undisturbed soil, towards the south. Four lead wires were used for each thermometer to minimize telemetering errors (McLernon 1969). Sensor signals were taken through linearizing bridges to a 12 point strip chart recorder. Thermometers were sampled sequentially for six seconds each to give a total cycle time of 72 seconds. Measurement error was conservatively estimated to be less than 0.5°C.

Daily soil moisture measurements were available from electrical resistance sensors (Coleman et al 1949) at 10 and 100 cm for a few years (1969,1970,1971).

Air temperatures were taken with standard Meteorological Service of Canada (MSC) maximum and minimum thermometers in a Stevenson Screen. Hours of bright sunshine were obtained from a Campbell Stokes sunshine recorder and wind was measured at seven meters above ground with a Bendix Frieze aerovane. Rainfall and snowfall were obtained from standard MSC rain and snow gauges and snow depth was read from snow stakes.

4.3 Data Preparation

Soil temperature records were scaled hourly for 1, 10 and 50 cm levels and twice daily (noon and midnight) for 100 and 200 cm levels. Calibration corrections were applied on the basis of characteristic equations provided by the manufacturer of the thermometers. Five day means were calculated and ensemble averaged over the five year period. Diurnal cycles for each month were obtained by averaging hourly temperatures for the five year period.

Best fit Fourier series coefficients for the annual and diurnal waves were obtained with a calculator program (HP 9820A, III-9) and the curves were drawn with an X-Y plotter (HP9862A) (Figure 3,4 and 6). About 95% or more of the temperature variance was accounted for by these curves.

The vertical time section of soil temperatures (Figure 2) was prepared by plotting five day mean temperatures on a large rectangular grid with each horizontal line representing a particular soil depth. Smooth isotherms were then drawn by hand at one degree C intervals by a linear interpolation technique.

Thermal diffusivity of the soil between 1 and 10 cm was calculated for five day periods by the amplitude ratio and phase lag methods from the diurnal wave. Maximum and minimum temperatures and the phase at these points was estimated from best fit curves for five day means. Hand smoothed curves drawn through these points were presented in Figure 18. An average annual cycle of thermal diffusivity based on these three years (1969,1970,1971) was used for prediction purposes as outlined in the next section (Figure 18). Thermal diffusivities based on the annual wave were also calculated by these methods for 10,50,100 and 200 cm (Table 13).

4.4 Prediction

To predict actual soil temperature for a particular week or month, it is necessary to understand how soil temperature is related to meteorological variables such as air temperatures, sunshine, wind, precipitation, snow depth and so on. Linear correlation and regression coefficients were obtained for monthly mean values with a University of Manitoba program (STAT 18) run on a PDP/10 computer (Table 1-6,10,11).

Daily soil surface temperature can be predicted in a somewhat similar way. For convenience, daily prediction was grouped into monthly units. The trend line or shape of the monthly unit was obtained from the annual soil temperature curve. Then a regression equation based on daily meteorological observations was used to estimate the departure of the soil temperature from the 5 day mean. The mean of the resulting daily estimates was then adjusted to the predicted monthly mean to give the final daily temperature output. These calculations were executed on a programmable calculator (HP9820A) and concurrently plotted on a temperature versus time graph with an X-Y plotter (HP0962A).

The input data were subjected to symmetrical statistical weighted smoothing to filter out unwanted random fluctuations. Care was taken in all smoothing processes to ensure that useful frequencies were not distorted or filtered out.

The summer of 1976 was chosen to test the prediction capability of this technique. As it turned out, June 1976 was a very wet month in Manitoba (17.3 cm at WNRE), and soil moisture was above field capacity for a significant period of time (surface water). Since soil thermal properties are strongly dependent on soil moisture and since the surface radiation balance is strongly affected by moisture conditions

TABLE I

SIMPLE CORRELATION COEFFICIENTS

SOIL SURFACE (1 CM) TEMPERATURES VS METEOROLOGICAL OBSERVATIONS

MAY-<u>JUNE</u>-JULY

		SOILT	AIRT	PCPN	AMIN	WNDS	SUNS	AMAX
	VARIABLE	1	2	3	4	5	6	7
SOILT	1	1.00						
AIRT	2	0.97	1.00					
PCPN	3	-0.02	0.03	1.00				
AMIN	4	0.95	0.99	0.14	1.00			
WNDS	5	-0.73	-0.67	0.35	-0.57	1.00		
SUNS	6	0.47	0.46	-0.54	0.34	-0.58	1.00	
AMAX	7	0.96	0.98	-0.07	0.94	-0.75	0.57	1.00

Note: SOILT is soil temperature, AIRT is mean air temperature, PCPN is precipitation, AMIN is mean minimum air temperature, WNDS is winds speed, SUNS is mean hours of bright sunshine, AMAX is mean maximum air temperature.

TABLE 2

SIMPLE CORRELATION COEFFICIENTS
SOIL SURFACE (1 CM) TEMPERATURES VS METEOROLOGICAL OBSERVATIONS
JUNE-JULY-AUGUST

		SOILT	AIRT	PCPN	AMIN	WNDS	SUNS	AMAX
	VARIABLE	1	2	3	4	5	6	7
SOILT	1	1.00						
AIRT	2	0.88	1.00					
PCPN	3	-0.07	-0.05	1.00				
AMIN	4	0.85	0.95	0.10	1.00			
WNDS	5	-0.38	-0.39	0.38	-0.19	1.00		
SUNS	6	0.45	0.50	-0.56	0.27	-0.48	1.00	
AMAX	7	0.84	0.96	-0.18	0.82	-0.53	0.65	1.00

TABLE 3

SIMPLE CORRELATION COEFFICIENTS

SOIL SURFACE (1CM) TEMPERATURES VS METEOROLOGICAL OBSERVATIONS

JULY-AUGUST-SEPTEMBER

		SOILT	AIRT	PCPN	AMIN	WNDS	SUNS	AMAX
	VARIABLE	1	2	3	4	5	6	7
SOILT	1	1.00						
AIRT	2	0.96	1.00					
PCPN	3	0.09	0.01	1.00				
AMIN	4	0.95	0.98	0.09	1.00			
WNDS	5	-0.27	-0.35	0.50	-0.19	1.00		
SUNS	6	0.86	0.91	-0.21	0.85	-0.41	1.00	
AMAX	7	0.94	0.99	-0.05	0.93	-0.45	0.93	1.00

TABLE 4

LINEAR MULTIPLE CORRELATION AND REGRESSION COEFFICIENTS

MONTHLY MEAN SOIL TEMPERATURE (1CM) VS AIR TEMPERATURE, PRECIPITATION (P) SUNSHINE (S), PINAWA, MANITOBA, 1968 TO 1972 AND 1976.

JUNE

	R	C_0	C_1 (AIRT)	C_2 (AMAX)	C_3
1.	.971*	-0.9165	0.6412	0.2775	
2	.972	0.2212	0.7666	0.1486	-0.0914(P)
3.	.972	-1.6521	0.4768	0.4669	-0.1005(S)

* $R^2_{1.27} = (r_{12}^2 + r_{17}^2 - 2r_{12}r_{17}r_{27})/(1 - r_{27}^2)$ (Panofsky and Brier 1968)

TABLE 5

LINEAR MULTIPLE CORRELATION AND REGRESSION COEFFICIENTS

MONTHLY MEAN SOIL TEMPERATURE (1CM) VS AIR TEMPERATURE, PRECIPITATION (P) SUNSHINE (S), PINAWA, MANITOBA, 1968 TO 1972 AND 1976.

JULY

	R	C_0	C_1 (AIRT)	C_2 (AMAX)	C_3
1.	.885	4.3063	0.8495	-0.1060	
2.	.886	4.8285	0.9190	-0.1732	-0.0614(P)
3.	.887	4.7269	0.9837	-0.2691	0.1139(S)

For legend see Table 1.

TABLE 6

LINEAR MULTIPLE CORRELATION AND REGRESSION COEFFICIENTS

MONTHLY MEAN SOIL TEMPERATURE (1CM) VS AIR TEMPERATURE, PRECIPITATION (P) SUNSHINE (S), PINAWA, MANITOBA, 1968 TO 1972 AND 1976.

AUGUST

	R	C_0	C_1 (AIRT)	C_2 (AMAX)	C_3
1.	.959	2.6060	1.2002	-0.2998	
2.	.961	1.6853	1.0669	-0.1826	0.1683(P)
3.	.959	2.8941	1.2113	-0.3380	0.0496 (S)

(Idso et al 1975 a, 1975 b, Yu 1977) a soil moisture term was introduced for June. Rainfall was used as a soil moisture estimator approximately along these lines. The first 0.5 cm of each rainfall was ignored, as this amount would not normally reach the soil due to interception by vegetation and evaporation. Daily evapotranspiration of 0.5 cm was assumed for wet soil (Reimer et al 1973). A moist surface soil would require a nearly fixed fraction of available energy for daily evapotranspiration as long as the moisture lasts (Rosenberg 1974). The term

$$a(\log ACPN)$$

was used to express this, where a is a constant (-2.0) and ACPN is accumulated precipitation (5 APCN 200 mm). This term operated for most of the month of June with the results in Figure 7. This term was not used for July and August (Figure 8,9).

Daily subsoil temperature prediction at 10 cm was tested with an extension of the regression method outlined earlier. The forcing function based on equation (37) was combined with the regression equation to estimate daily departure from the mean, and applied to July 1976 WNRE data, with the remarkably good results seen in Figure 15 ($s = 0.4^{o}C$).

For subsoil temperature prediction equation (37) can also be used to produce an annual soil temperature wave at any depth provided the soil physical properties are known. If the mean annual soil temperature is assumed constant with depth (Table 14) and the surface amplitude (A_{on}) of each harmonic is known and if $K = k/\rho c$ is known so that D may be calculated, equation (37) is applicable. Predicted and observed annual soil temperature curves are presented for comparison in Figure 16 for all four subsoil levels. Although K is known from the amplitude ratio and phase shift methods it is also estimated by an iteration method with equation (37) (Table 13). This technique gives basically the same results as Lettau's (1954) temperature integral method.

WNRE annual soil surface (1 cm) Fourier-series coefficients are displayed in Figure 7.

5. RESULTS AND DISCUSSION

5.1 Average Observed Soil Temperatures

Five day average observed temperatures at five soil levels are displayed in Figure 1. Large random, temperature fluctuations are apparent at the soil surface, but these fluctuations decrease rapidly with depth. This is consistent with the exponential damping term (exp -z/D) in equation (35). For soil with a thermal diffusivity of 12 x $10^{-4} cm^2 s^{-1}$ this gives a damping depth of about 13 cm for a hypothetical periodicity of 5 days. Thus, the amplitude at 13, 26 and 39 cm should be 1/e, $1/e^2$ and $1/e^3$ of that at the surface for a 5 day wave.

In winter with snow on the ground somewhat less damping is observed. Snow with a density of 0.3 g/cm^3 and thermal diffusivity of 0.0045 cm^2/s would have a damping depth of 25 cm for a five day wave. Thus, 1/e of the snow surface temperature wave is damped out before it reaches the soil surface through a 25 cm layer of snow.

The vertical time section of soil temperatures (Figure 2) clearly shows the time of rapid soil warming in spring and cooling in fall, where the isotherms are closely packed. Delay of the maximum and minimum with depth is also evident.

Frost penetration does not seem to exceed 50 cm, on the average, for the years from 1968 to 1972 under forage conditions. The presentation of an overall average temperature for each measurement level (Figure 3) clearly demonstrates the nearly sinusoidal form of the annual temperature wave. The amplitude of the annual temperature wave decreases from 11°C at the surface to 3.5°C at 200 cm (Figure 3).

Diurnal surface temperature waves for June, July and August (Figure 4) also have a nearly sinusoidal wave form. About 95% of the variance is accounted for by the fundamental wave. The amplitude of the diurnal surface temperature wave has a maximum in summer of 5°C and a minimum in winter (Fig. 6) of near zero. The amplitude of the diurnal wave also decreases with depth to near zero values at 50 cm (not shown).

Prediction of the diurnal temperature wave is not attempted in this work. For an application that is not too demanding, an observed long-term average diurnal temperature wave imposed on a predicted daily mean temperature may suffice. In a more demanding application, a knowledge of cloud cover, wind and precipitation may be required. For a series of completely cloud-free days without much wind and with fairly dry soil, the diurnal soil surface temperature wave will be a fairly symmetrical sinusoidal wave form with greater amplitude (double) than the longterm mean. For a completely overcast day, the amplitude will be small or zero, provided there are no other meteorological events (i.e. a frontal passage, thunderstorm, etc.) to disturb this pattern.

Therefore, a typical soil temperature wave for a clear or overcast day is a somewhat idealistic concept, and a good example is difficult to find. For general information, however,

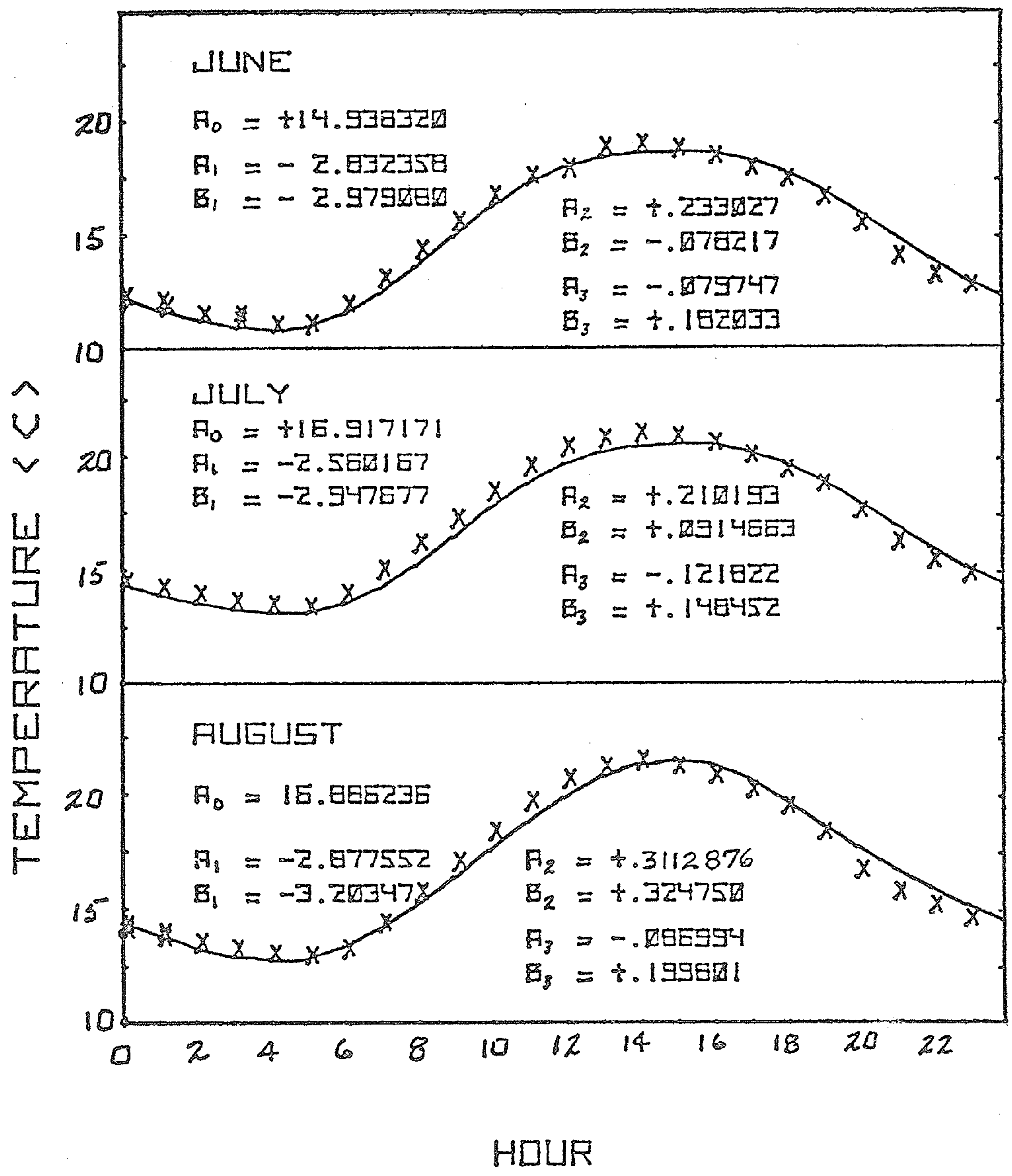


Figure 4. Observed (X) average diurnal soil surface (1 cm) temperature waves for June, July and August (1968-1972) and best-fit Fourier-series curves (solid line) at WNRE.

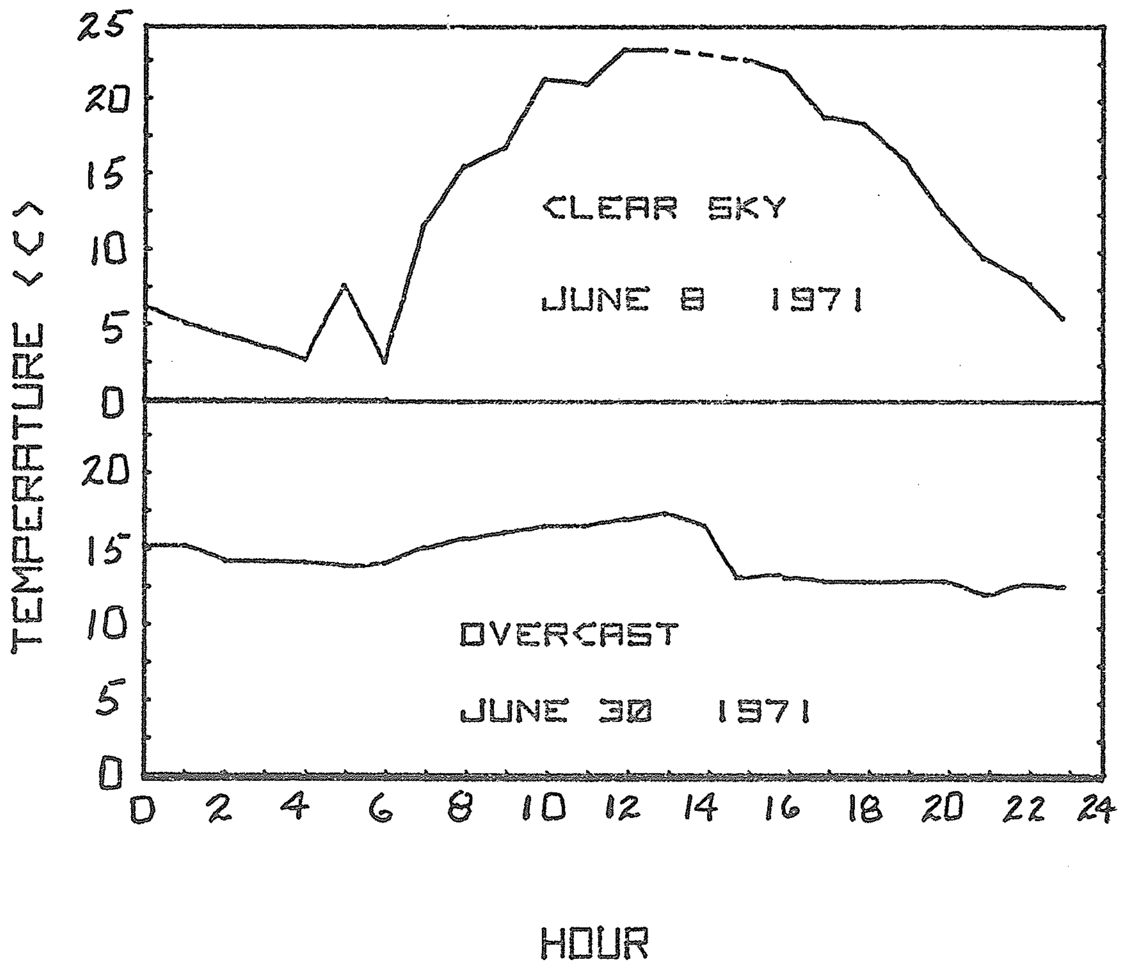


Figure 5. Observed soil surface (1 cm) temperature fluctuations on a clear day and on a cloudy day at WNRE (the sudden cooling at mid-afternoon on June 30 is not typically diurnal).

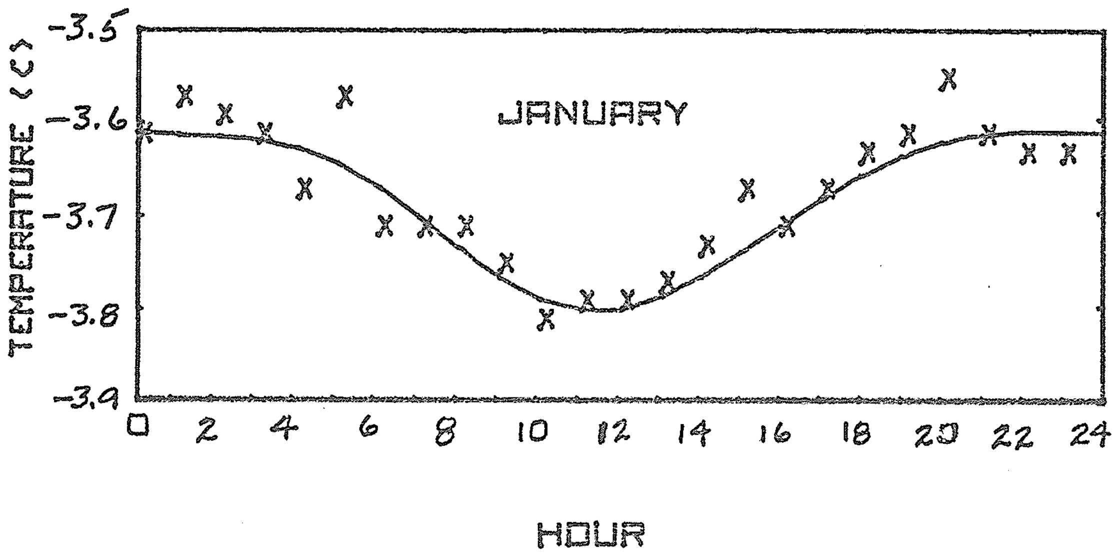


Figure 6. Observed average diurnal soil surface (1 cm) wave (x) at WNRE for January (1968-1973) and best-fit Fourier-series curve (solid line).

clear day and overcast day summer curves are given in Figure 5. Note, for example, that the amplitude of a clear day wave is about $10^{\circ}C$, twice that of the mean wave. With a certain amount of care and discretion, such knowledge can give a more useful estimate than a long-term average curve.

In January the diurnal fluctuation of the soil surface temperature is very small - a few tenths of a degree Celsius - and out of phase with the above snow diurnal cycle. This damping and phase shift is, of course, caused by the layer of snow on the ground (Figure 6). For most applications, the diurnal wave in winter can safely be ignored, especially with a normal amount of snow cover. For research purposes, however, the amplitude and phase shift of the diurnal wave may be of interest.

5.2 Prediction - Monthly Mean

By comparing Figures 1 and 3 it is obvious that a surface temperature for any particular time exhibits a lot of variation about the long-term mean. To predict such a highly variable quantity it is necessary to understand how the soil temperature (predictand) is related to the meteorological variables (predictor).

Five-day mean values were studied individually and grouped into pairs and tetrads, but the results were inconsistent, and predicted values showed an unacceptable amount of scatter. A longer time unit was indicated for this study and the month was chosen for convenience. This was done with the realization that five years is not a sufficiently large sample to give stable regression coefficients. The problem was largely overcome by pooling adjacent months with the month in question (see Tables 1,2,3,10).

These monthly mean studies showed that soil temperature is strongly associated with air temperature in June, July and August (see Tables 1-3). Association with wind speed and sunshine varies considerably over the summer months although the correlation coefficient (r) is consistently negative for wind speed and positive for sunshine (Table 1-3). The smallest $|r|$ values found were for precipitation.

Linear regression equations of soil temperature on air temperature, sunshine and precipitation were formed for June, July and August (Table 4-6,11). Wind was not used for two reasons; (1) the relationship with soil temperature is apparently not linear and (2) wind is affected by local surface roughness which presents problems for regional application. It is obvious from the multiple correlation coefficients (R) that air temperatures are the best predictors and precipitation and sunshine add very little to the capability of the equations.

Predicted monthly soil temperatures were compared with observed values in Tables 7-9. The standard deviation of the difference is less than one degree C in all cases and the smallest value (best prediction) is 0.84 (August).

Wintertime soil surface temperature estimation introduces another dimension - the effect of snow cover. Sample size was even more important here, so the observations for Winnipeg, Manitoba, with a longer record, were used. January

TABLE 7

MEAN MONTHLY OBSERVED AND CALCULATED
SOIL TEMPERATURES (C)
(1 CM)

JUNE

	OBSERVED	REGRESSION EQUATION NUMBER		
		1	2	3
1968	14.45	13.87	13.84	13.95
1969	11.00	10.94	11.05	10.91
1970	16.36	16.85	17.01	16.82
1971	16.94	16.02	16.04	16.05
1972	16.03	15.62	15.60	15.53
1976	15.46	16.94	17.10	16.98

TABLE 8

MEAN MONTHLY OBSERVED AND CALCULATED
SOIL TEMPERATURES (C)
(1 cm)

JULY

	OBSERVED	REGRESSION EQUATION NUMBER 1	2	3
1968	17.66	16.46	16.36	16.41
1969	15.94	16.91	16.95	16.88
1970	19.16	18.62	18.59	18.65
1971	15.83	15.84	15.92	15.97
1972	15.93	15.51	15.54	15.39
1976	16.45	17.63	17.59	17.67

TABLE 9

MEAN MONTHLY OBSERVED AND CALCULATED
SOIL TEMPERATURES (C)

(1 CM)

AUGUST

	OBSERVED	REGRESSION EQUATION NUMBER 1	2	3
1968	15.03	14.93	15.03	14.93
1969	18.08	18.56	18.73	18.63
1970	17.69	16.35	16.34	16.37
1971	16.89	16.57	16.29	16.62
1972	15.75	15.92	16.22	15.90
1976	16.29	17.29	17.11	17.30

TABLE 10

SIMPLE CORRELATION COEFFICIENTS

SOIL TEMPERATURES (5 CM) VS METEOROLOGICAL OBSERVATIONS

WINNIPEG INTL A

DECEMBER - JANUARY - FEBRUARY

		SOILT 1	AIRT 2	SNOWD 3	AMIN 4	WNDS 5	SUNS 6	AMAX 7
SOILT	1	1.00						
AIRT	2	0.30	1.00					
SNOWD	3	0.12	-0.47	1.00				
AMIN	4	0.28	0.98	-0.54	1.00			
WNDS	5	0.13	-0.12	0.63	-0.19	1.00		
SUNS	6	-0.28	-0.21	0.55	-0.31	0.37	1.00	
AMAX	7	0.30	0.98	-0.38	0.94	-0.04	-0.11	1.00

NOTE; SEE TABLE 1 FOR LEGEND.

TABLE II

LINEAR MULTIPLE CORRELATION AND REGRESSION COEFFICIENTS

MONTHLY MEAN SOIL TEMPERATURE VS AIR TEMPERATURE, SNOW DEPTH AND SUNSHINE, WINNIPEG INTL A, 1969 TO 1976.

DECEMBER - JANUARY - FEBRUARY

	R	C_0	C_1 (AMIN)	C_2 (AMAX)	C_3 (SNOW D)	C_4 (SUNS)
1.	.343	-0.7242	0.1996	0.0186		
2.	.429	-2.7337	-0.4355	0.6079	-0.9295	
3.	.607	2.2230	-0.0368	0.3353	0.2377	-1.3094

TABLE 12

MEAN MONTHLY OBSERVED AND CALCULATED
SOIL TEMPERATURES (C)
(5 CM)
WINNIPEG INTL A

JANUARY

	OBSERVED	CALCULATED		
		1	2	3
1969	-4.72	-4.71	-5.08	-3.55
1971	-9.72	-4.99	-5.23	-7.01
1973	-2.50	-3.26	-3.20	-4.29
1974	-1.97	-4.85	-3.57	-2.78
1975	-3.17	-3.64	-3.59	-3.05
1976	-2.33	-3.88	-4.68	-3.63

was pooled with December and February and soil temperature was correlated with air temperatures, snow depth, wind, and sunshine (Table 10). Association was weaker than summer, for most variables, but it was more uniformly distributed. Thus, a three or four variable equation is a much improved predictor over a two variable equation, in this case (Table 11). In spite of this, the best four variable equation still does not predict as well (R=0.6) as those for the summer months. The standard deviation of the difference between observed and predicted is 1.7 (Table 12).

5.3 Prediction - Daily Mean

The daily prediction sequence begins by smoothing the input data with a moving three-day function. The smoothed values for a particular day are then applied to the regression equation to obtain the departure from the long-term mean. The Fourier Series expression (33 or 34) was then solved for that day and combined with the departure value to give the final estimate.

Trials were run without any smoothing and the result was a widely fluctuating output that overshot the observed value by a wide margin. On the other hand, over-smoothed input gave a curve that smoothed out much or all of the actual observed fluctuations of a few days in length. The final choice of smoothing function was a compromise between an overall best fit curve for a month and rounding off of some of the sharpest peaks and valleys. This gave a standard deviation of less than $1^{\circ}C$ for the difference between predicted

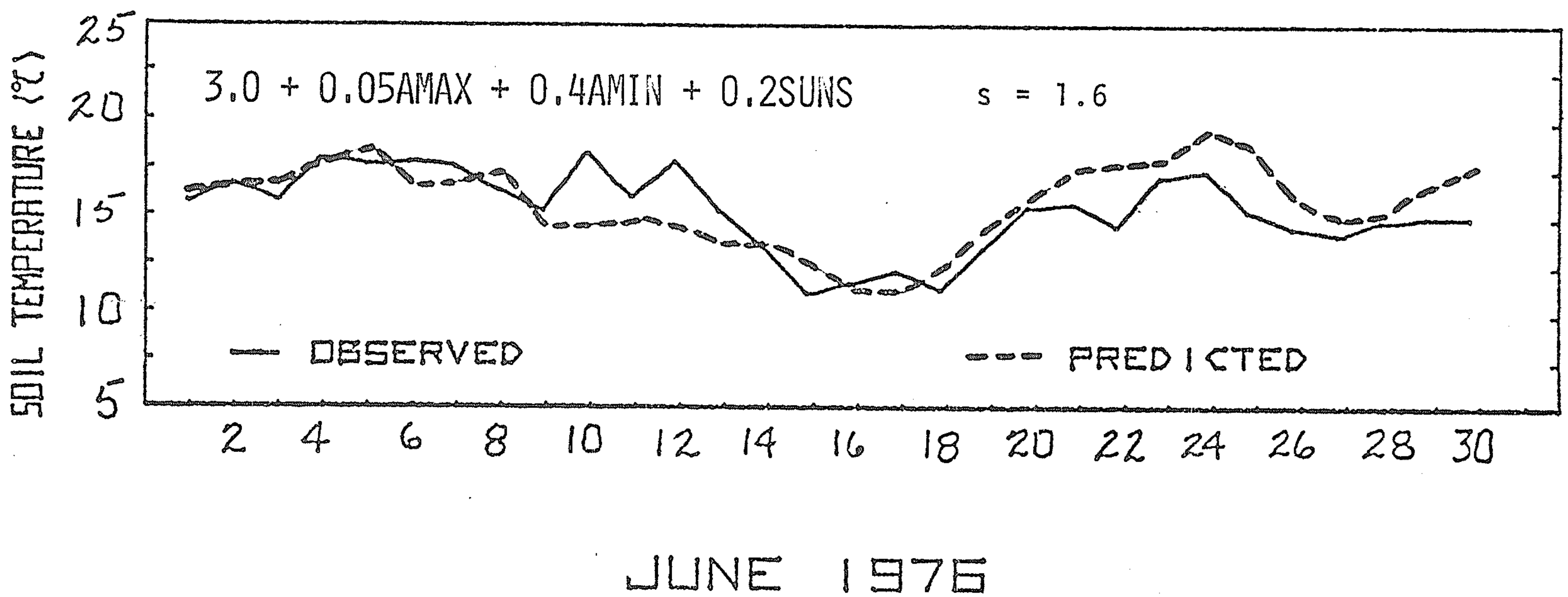


Figure 7. Daily observed and predicted soil surface (1 cm) temperatures for June, 1976 at WNRE (s is the standard deviation of predicted minus observed). WNRE annual soil surface (1 cm) Fourier coefficients are; $A_0 = 5.6$, $A_1 = 10.031$, $B_1 = -4.6911$, $A_2 = 0.8763$, $B_2 = 1.1149$, $A_3 = 0.2045$, $B_3 = 0.6605$.

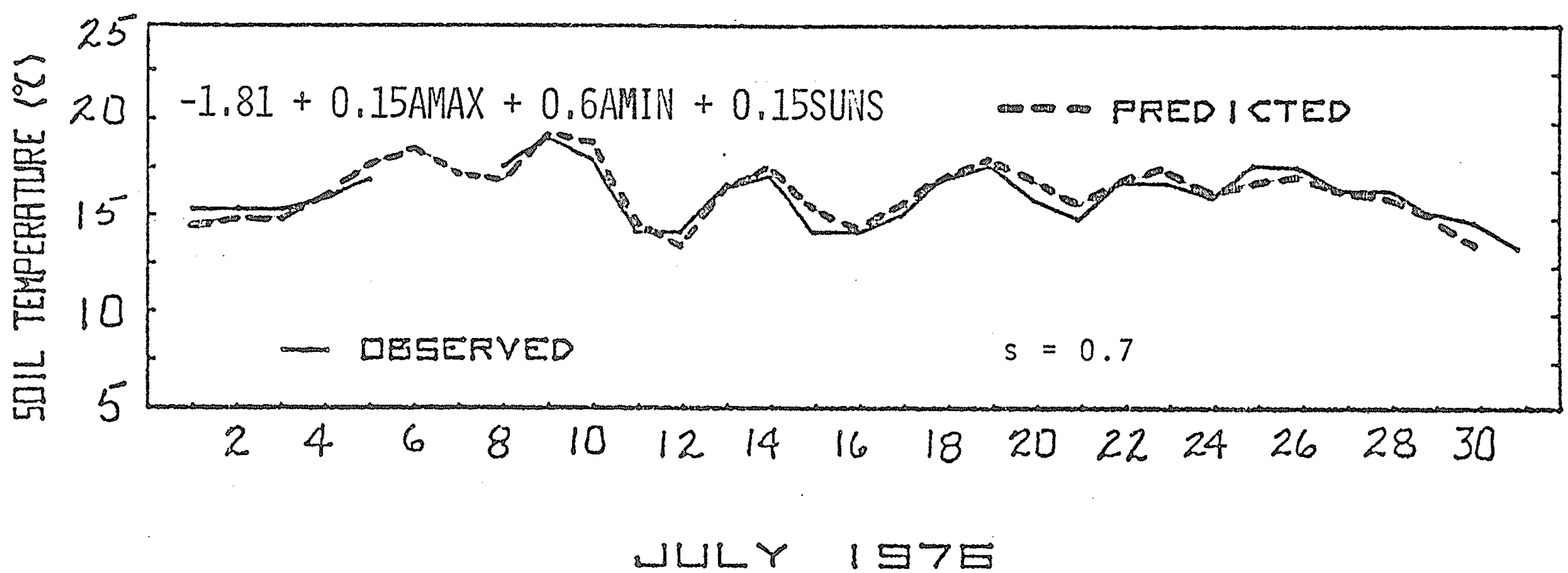


Figure 8. Daily observed and predicted soil surface (1 cm) temperatures for July, 1976 at WNRE. Note: Root-mean-square-error (RMSE) is nearly the same as s; i.e., the sum of squares is divided by N instead of N-1 (Panofsky and Brier 1968).

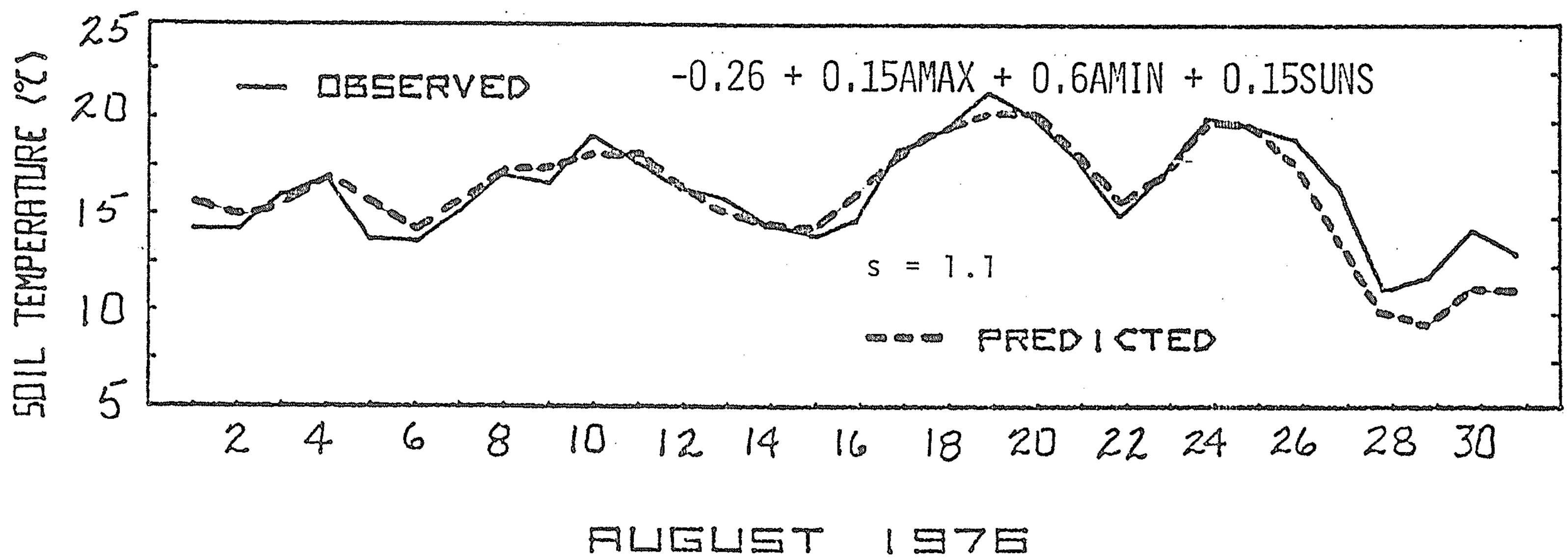


Figure 9. Daily observed and predicted soil surface (1 cm) temperatures for August, 1976 at WNRE.

and observed for all the summer months except June 1976 (Figure 7-9,12 and 15)(see Fig. 8 for a note on RMS-Error).

The phase of the input values must be adjusted to the observed temperatures to obtain a synchronized output curve. Minimum and maximum air temperatures occur about half a day apart, i.e., the maximum occurs about half a day after the minimum for any given day. Since the time step used here was a whole day, the maximum air temperature was delayed one day. In addition to this, the final output was delayed one day to give the best fit to the measured soil temperatures for the summer months. The cause of this may, at least in part, be attributed to the intervention of the plant canopy between the atmosphere and the soil and the effect that thermo-regulation of living tissue may have on soil temperature. As it turned out, the prediction scheme for January had somewhat the same features but perhaps for different reasons, as described below.

Unusually high soil moisture was a factor to be reckoned with because of its profound effect on the air-soil energy balance. The month of June in the test year (1976) had more than twice the normal rainfall, and surface water was observed on a number of days. Much of the incoming solar energy was thus consumed by direct evaporation and the soil temperature remained below the long-term mean. An attempt was made to simulate this with a precipitation term, but the results were not satisfactory, to say the least. The main difficulty seemed to be that precipitation was a poor estimator of soil

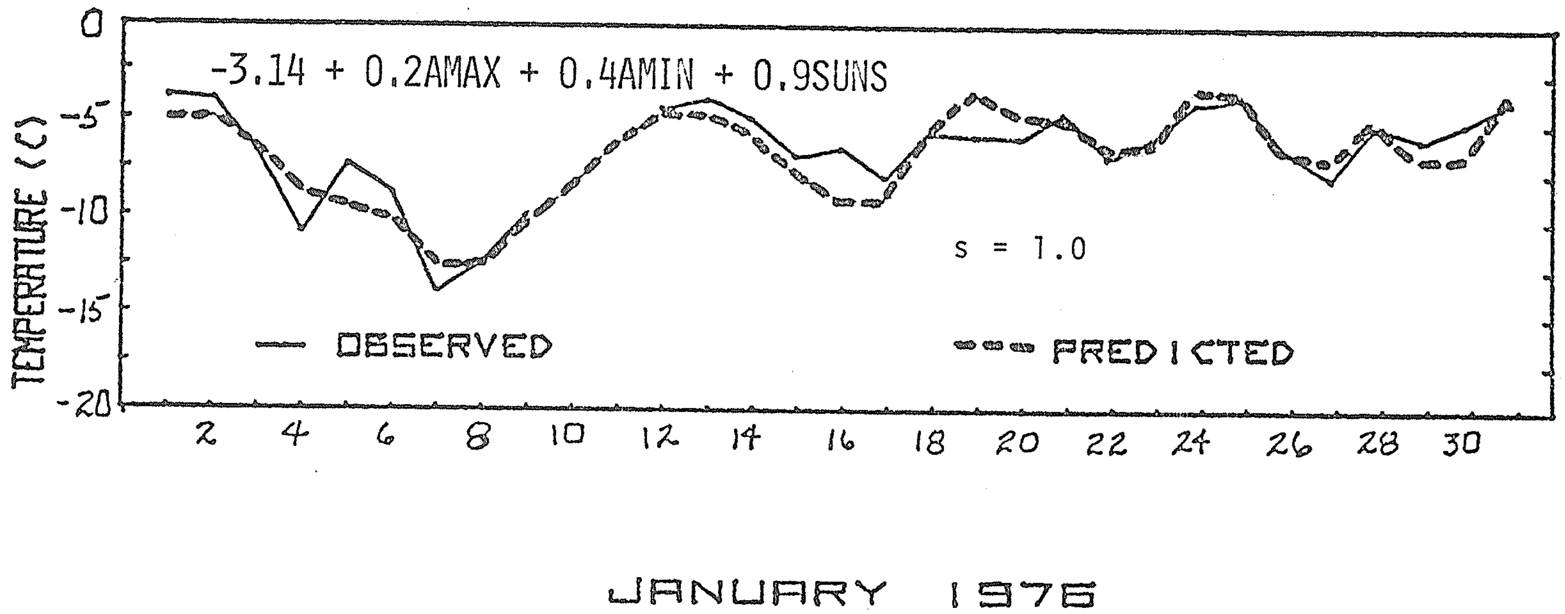


Figure 10. Daily observed and predicted soil surface (5 cm) temperatures for January 1976 at Winnipeg International Airport.

moisture because soil moisture is a function of several other factors such as plant canopy interception, runoff, evaporation, transpiration, infiltration, recharge and discharge, hydraulic conductivity, etc. Therefore actual soil moisture measurements should have been tested in such a prediction scheme.

For January soil temperature prediction, the same smoothing technique was used, and maximum air temperature was delayed for one day. However, a two-day delay for the final output gave the best results (Figure 10). The goodness of fit achieved for January was surprising (Fig. 10) in view of the smaller amount of association between the soil temperature and the meteorological variables. The standard deviation of the difference was 1.0°C.

5.4 Prediction - Regional

Regional application of this model seems quite feasible when certain conditions are met. Mean annual soil temperature is a very conservative quantity and has been mapped for most of Manitoba, south of 56 N latitude (Mills et al 1977). Similar information is available for the other Prairie Provinces (Ouellet 1973) and so a suitable mean can be obtained for equation (37) for most areas. The shape and amplitude of the annual cycle can be estimated for key locations in the Prairie Region, if different from that found for Pinawa, Manitoba and this model is then applicable.

Assuming a known monthly mean, daily soil temperature prediction for Winnipeg was attempted for July 1975. Agreement between observed and predicted is poor (Fig. 11), but it is noted that Winnipeg soil temperature is measured at 5 cm, not 1 cm. Applying a correction for damping, thus:

$$\exp(-z/D)$$

where z is depth and D is damping depth, the results in Figure 12 (for z=5 cm and D=10 cm) are obtained. The fit is remarkably good.

Daily January soil temperatures were predicted for Saskatoon and Thompson (Figures 13 and 14). Appropriate corrections were again made for the damping effect of snow and soil. The results were most encouraging. The standard deviation of the difference between predicted and observed temperatures was 1.1 and 0.5°C for Saskatoon and Thompson respectively.

5.5 Prediction - SubSoil

Prediction for subsoils was better than for surface soils, as might be expected, since random fluctuations are minimized or largely damped out at deeper levels. Soil thermal properties as expressed by thermal diffusivity (K) or damping depth (D) appear to be more stable and more representative at subsoil levels of 10 to 200 cm than in the surface few centimeters, as is amply demonstrated by the results in Figures 15 and 16.

A Chi-square test of all the prediction results in this work, assuming a normal or Gaussian distribution of the differences (Walpole 1968), indicated that all the variances were less than 1°C at the 99% confidence level for daily and monthly means, except June 1976.

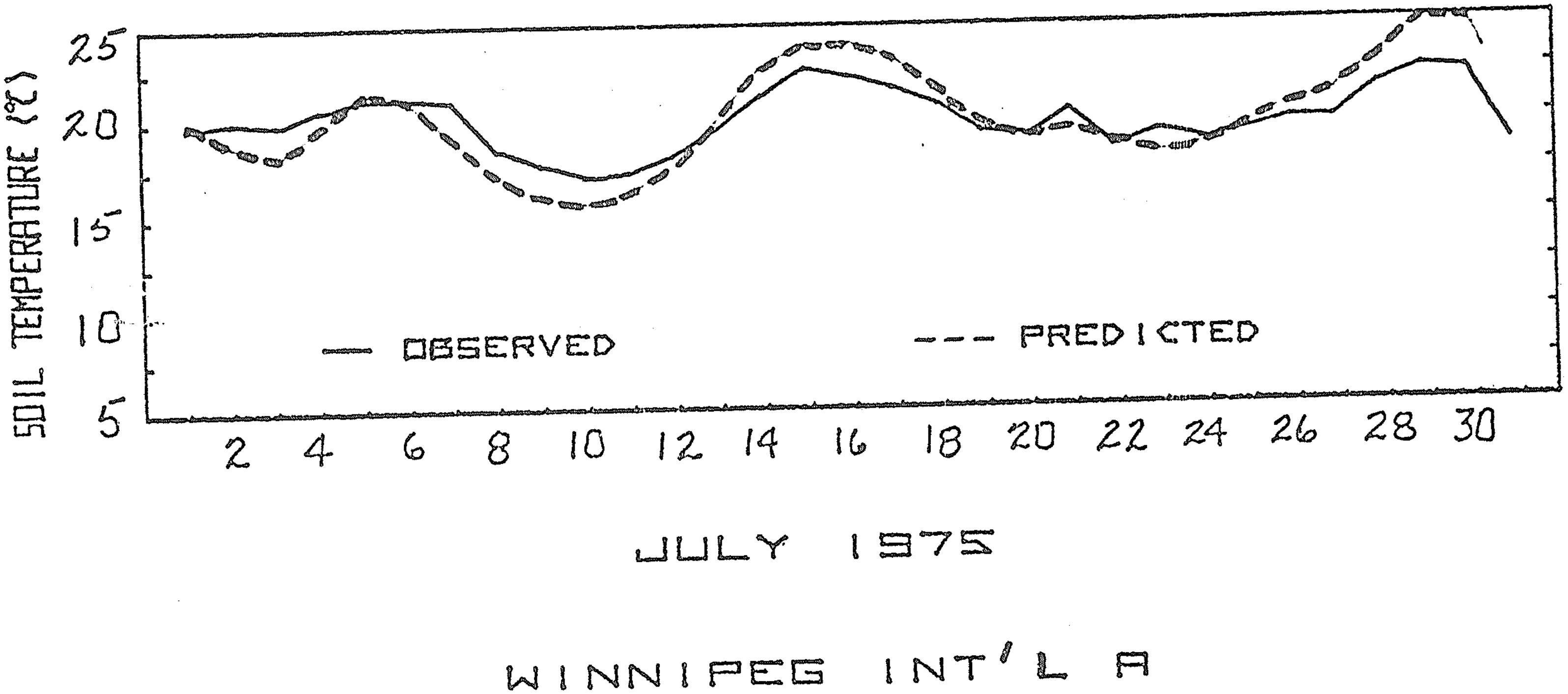


Figure 11. Daily observed and predicted (without damping correction) soil surface (5 cm) temperature for July 1975 at Winnipeg International Airport.

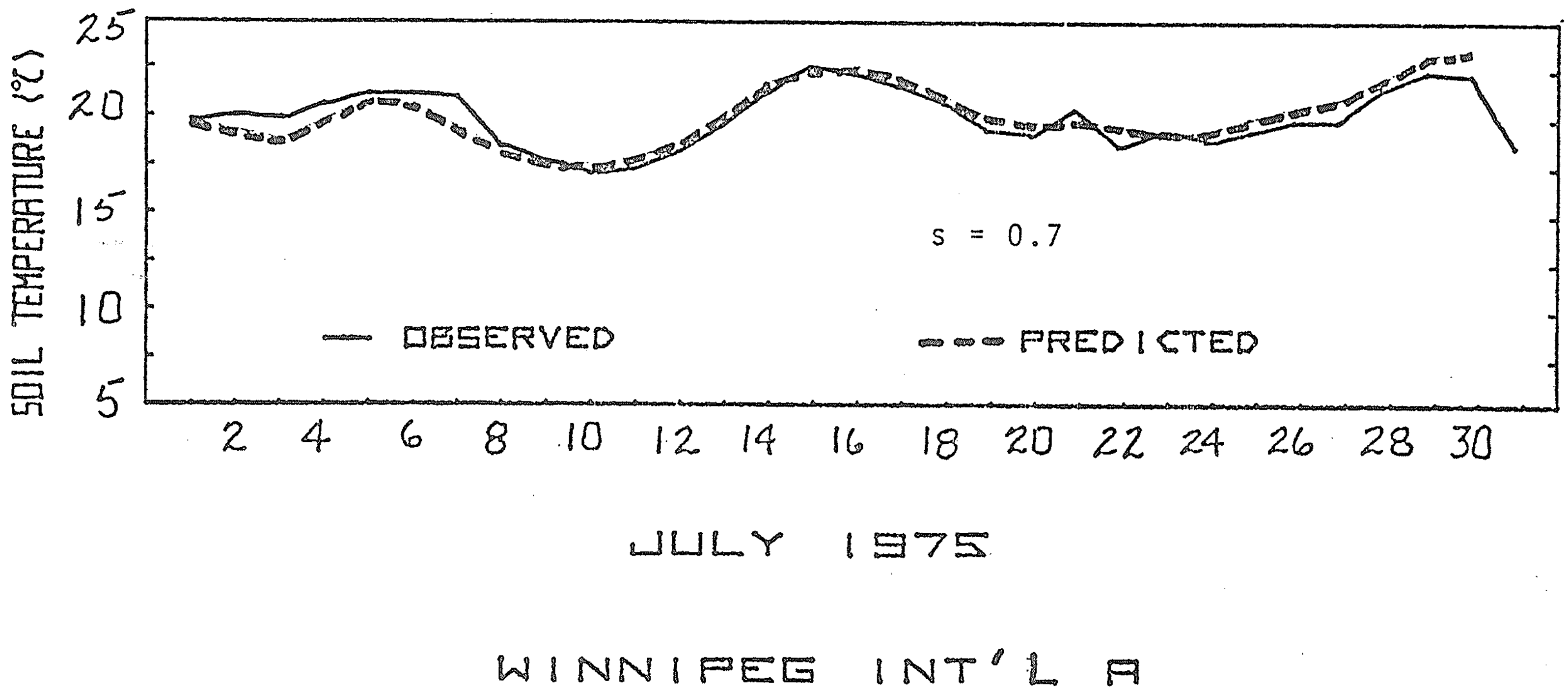


Figure 12. Daily observed and predicted (with damping correction) soil surface (5 cm) temperature for July 1975 at Winnipeg International Airport.

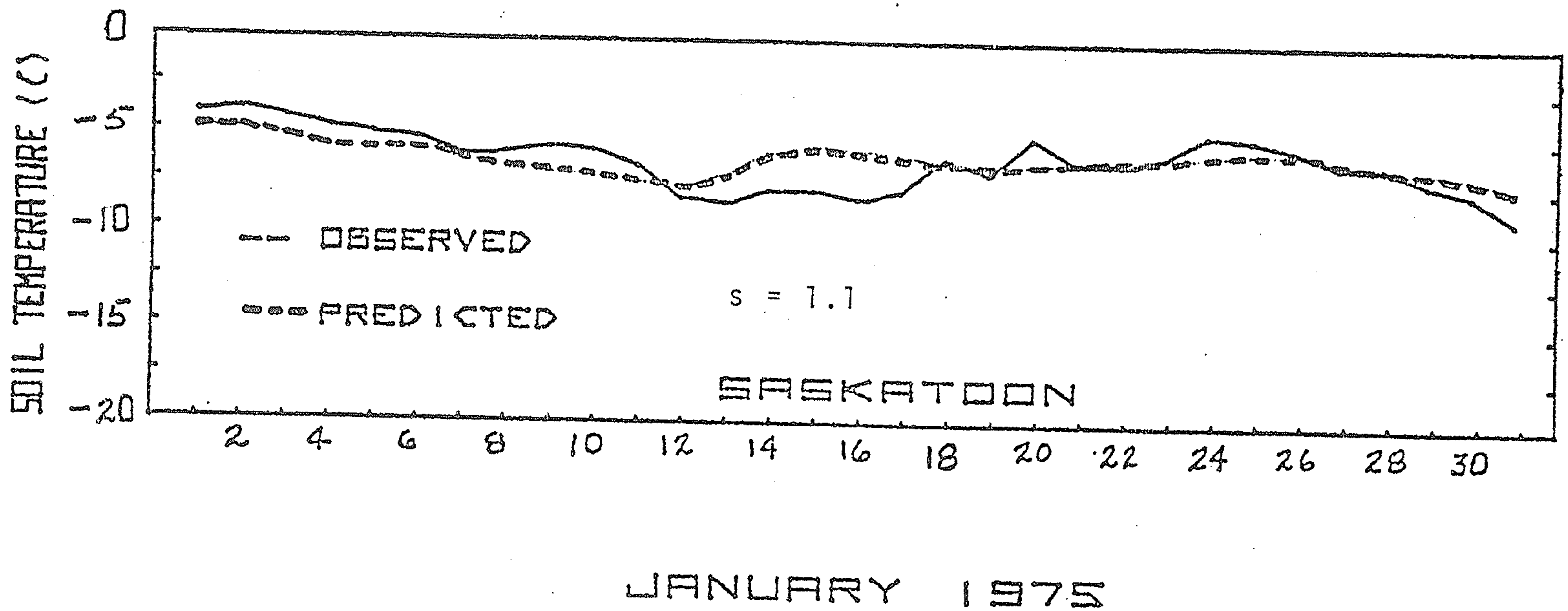


Figure 13. Daily observed and predicted soil surface (5 cm) temperature for January 1975 at Saskatoon.

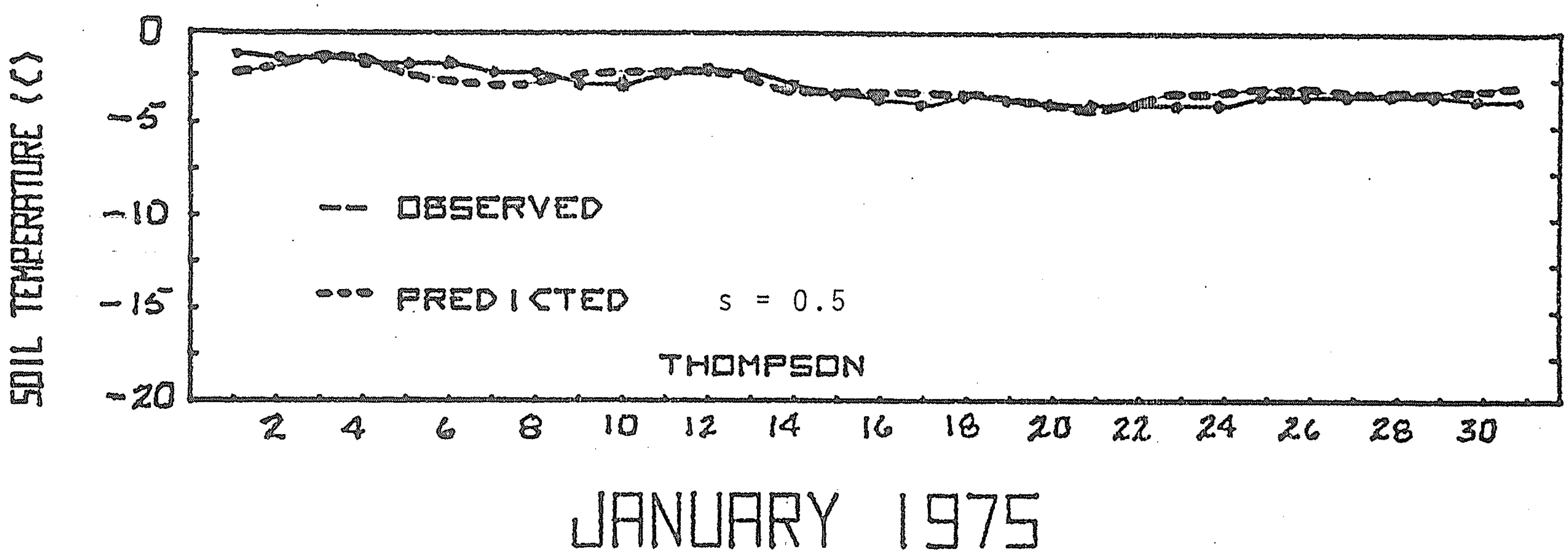
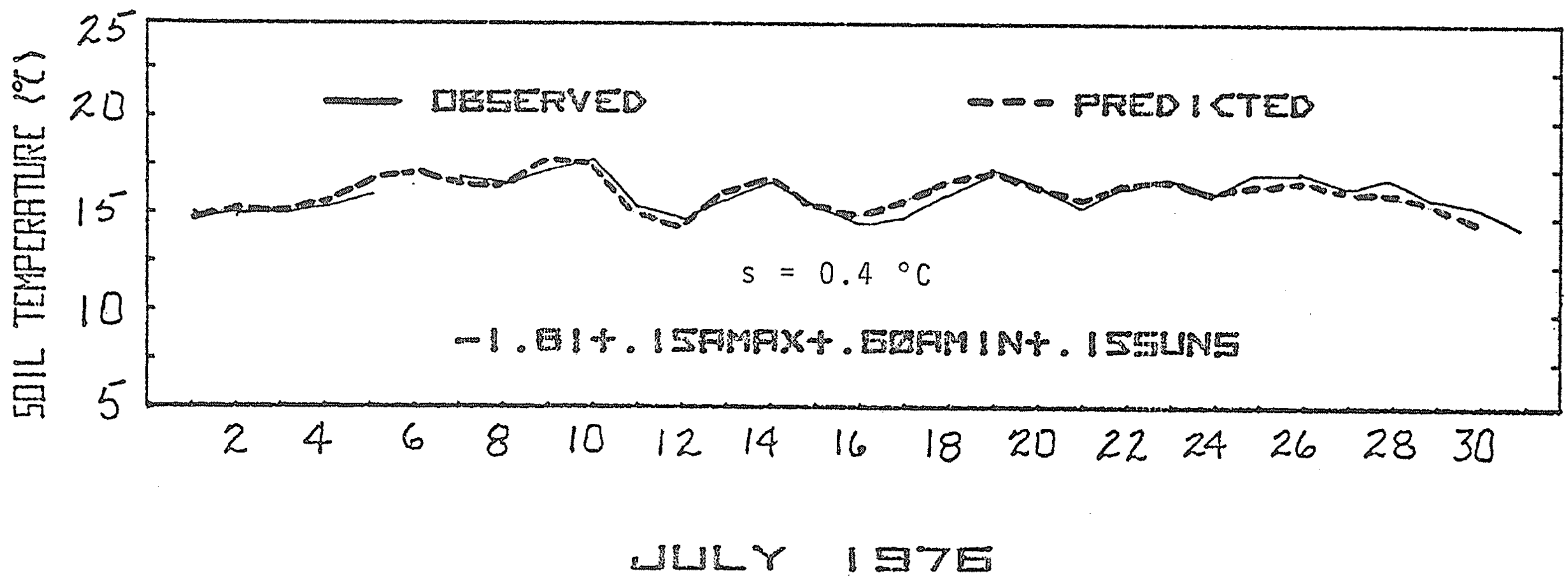


Figure 14. Daily observed and predicted soil surface (5 cm) temperature for January 1975 at Thompson.



Fibure 15. Daily observed and predicted subsoil (10 cm) temperatures for July 1976 at WNRE, and s is the standard deviation of predicted minus observed values.

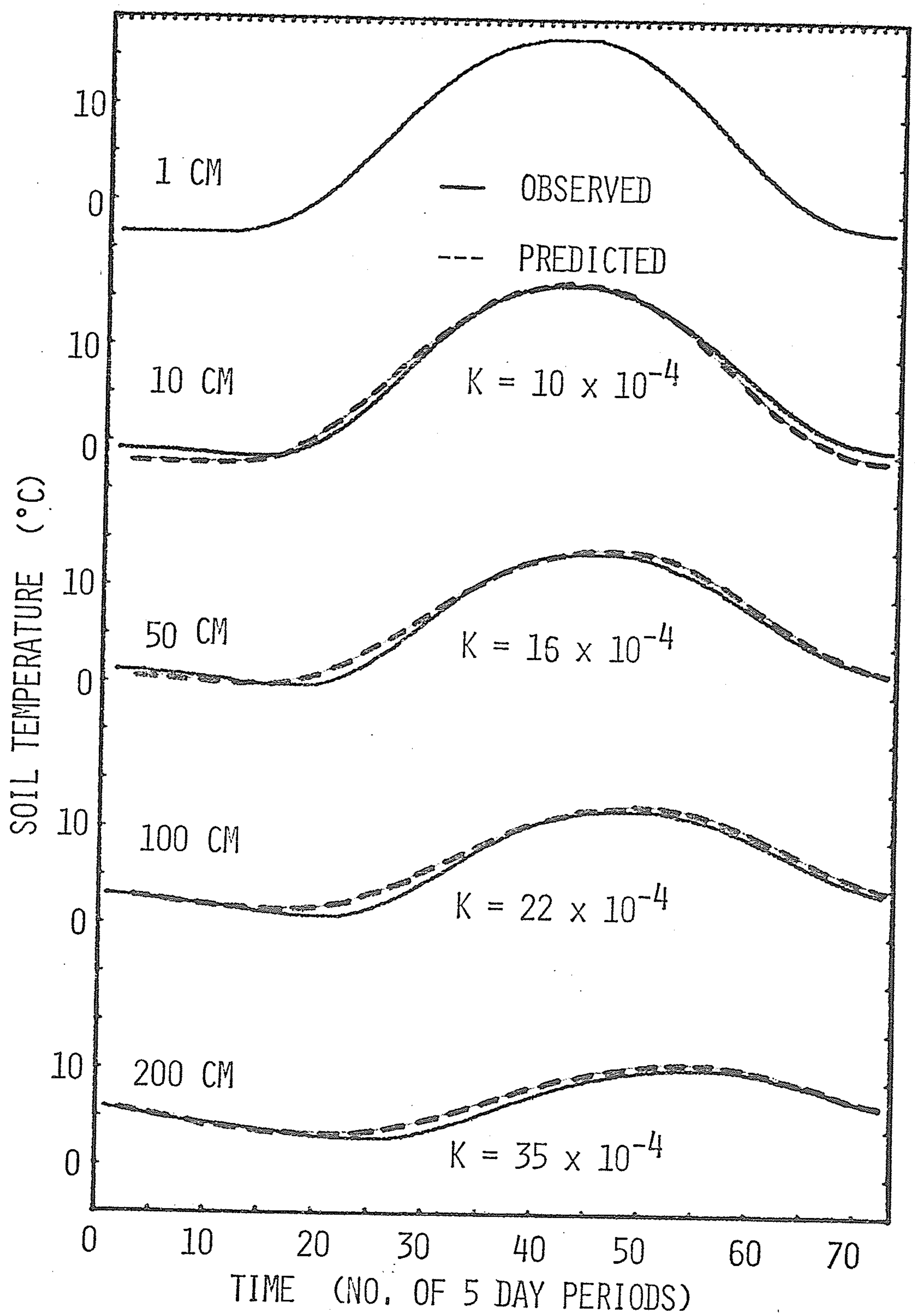


Figure 16. Observed and predicted subsoil annual temperature waves from Equation (37).

6. SUMMARY AND CONCLUSION

The temperature climate of a soil must be well understood before an effective prediction scheme can be brought into existence. A number of good descriptions of agricultural and grassland soils have been reviewed.

The long term (1969-1972) mean annual soil temperature at WNRE is 5.45^{o}C at 1 cm and 6.11^{o}C at 200 cm (Table 14). Most of this increase with depth occurs in the first 10 cm of soil. At the surface (1 cm) the mean seasonal temperature ranges from 18^{o}C in summer to -4^{o}C in winter. At 200 cm the mean temperature ranges from a maximum of about 10^{o}C in late summer to a minimum of 2.5^{o}C in late winter.

Mean hourly soil surface temperature ranges from a daytime maximum of about 20^{o}C to a nightime minimum of 12^{o}C in summer. In January the mean surface temperature is about -4^{o}C and the range is a few tenths of a degree, or neglegible for most purposes.

The mechanistic approach, to give precise results requires precise measurements of the heat sinks and sources and the flow of energy from one soil layer to another. If the necessary transfer coefficients are accurately defined and processes are precisely described by appropriate mathematical expressions then precise prediction is possible as long as these parameters are valid.

The theory of non-equilibrium thermodynamics has been applied by several authors (Taylor and Cary 1964, Bolt and Groenevelt 1967, Jury and Miller 1974, Groenevelt and Kay 1974I & II etc.), and if attention is given to all phases of water and heat transfer, that is, total possible coupling between water and heat transfer processes, good results can be obtained. Under controlled conditions for very specific applications these methods give excellent results and the potential of this appraoch for future development is very good.

The present model is however based on known principles and existing techniques to model actual soil temperatures and it is less sensitive to changing conditions. The prediction technique consists of the following basic steps for any location:

(1) Obtain monthly mean soil surface temperatures if available or estimate these from the regression equations as outlined.

(2) Obtain a Fourier-series equation for the annual surface temperature wave for the site in question or a nearby site.

(3) Estimate daily mean soil surface temperatures from the Fourier-series curve plus the regression equation for the month. Adjust these values to the monthly mean from (1).

(4) Calculate daily subsoil temperatures if required from equation (37) using the best available information

on soil thermal properties (Figure 15).

(5) Calculate subsoil annual temperature waves if required using soil thermal properties from (4)(Fig.16).

As is seen from figures 7 to 14 the standard deviation of the difference between predicted and observed temperatures for surface soils is less than $1^{o}C$ (p=99%), except for June 1976 (Figure 7). Prediction of subsoil temperatures is expected to be better (Figure 15) since most of the troublesome random fluctuations found at the surface have been damped out.

The ultimate accuracy of any prediction scheme, however, is dictated, first, by the accuracy of the information used to design the model and, second, by the accuracy of the input data. The overall error in the WNRE soil temperature measurements used here is conservatively estimated to be $0.5^{o}C$. Therefore this is the limit of accuracy of the present model.

As noted earlier, it is clear from Figure 4 that the AM and PM method of measuring soil temperatures is subject to considerable uncertainty. Therefore it is recommended here that an attempt should be made to measure the daily minimum and the daily maximum soil temperatures in place of the fixed time AM and PM measurements.

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Table 13. Average thermal diffusivity of three soil layers by the amplitude ratio, phase shift and iteration method ($cm^2 s^{-1} x\ 10^4$).

Layer (cm)	Amplitude Ratio(68-72)	Phase Lag (1968-72)	Iteration Meth.* (1968-1972)
10-50	22	22	16
50-100	43	30	22
100-200	55	38	35

* Note: Diffusivities in this column are averaged from the soil surface to the lower boundary in each case.

TABLE 14

AVERAGE ANNUAL SOIL TEMPERATURE (^{o}C)

AT FIVE LEVELS AT WNRE

DEPTH (CM)	1	10	50	100	200
YEAR					
1969	5.22	5.63	5.58	5.62	5.91
1970	5.83	6.20	6.10	5.99	6.16
1971	6.00	6.02	6.24	6.16	6.33
1972	4.76	4.97	5.39	5.44	6.02
MEAN	5.45	5.71	5.83	5.80	6.11
STD DEV	.57	.55	.41	.33	.18

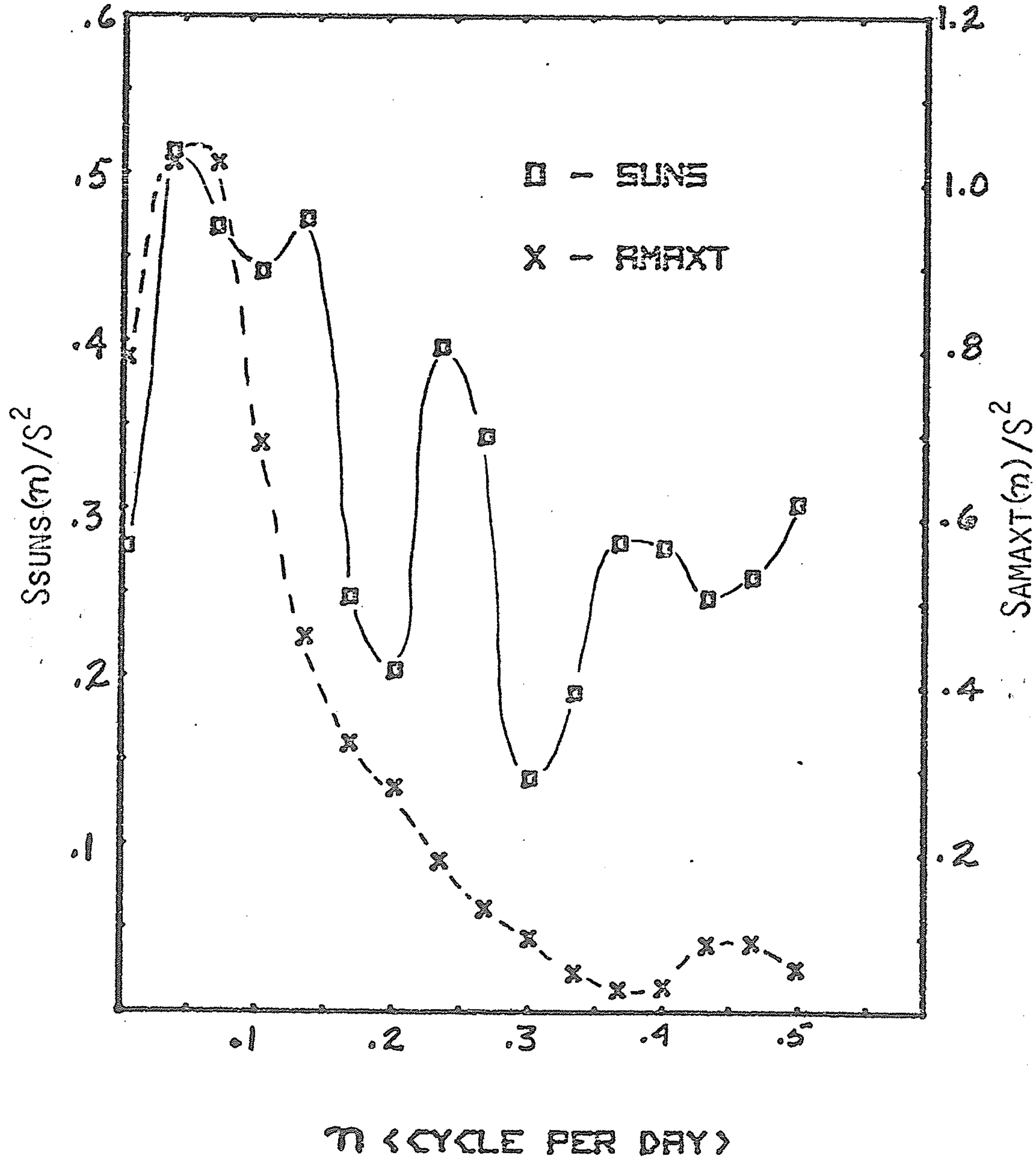

Figure 17. Normalized spectrum estimates of daily sunshine (SUNS) and daily maximum air temperature (AMAXT) for January by ensemble averaging over five years (1969-1973) at WNRE (see Appendix 1).

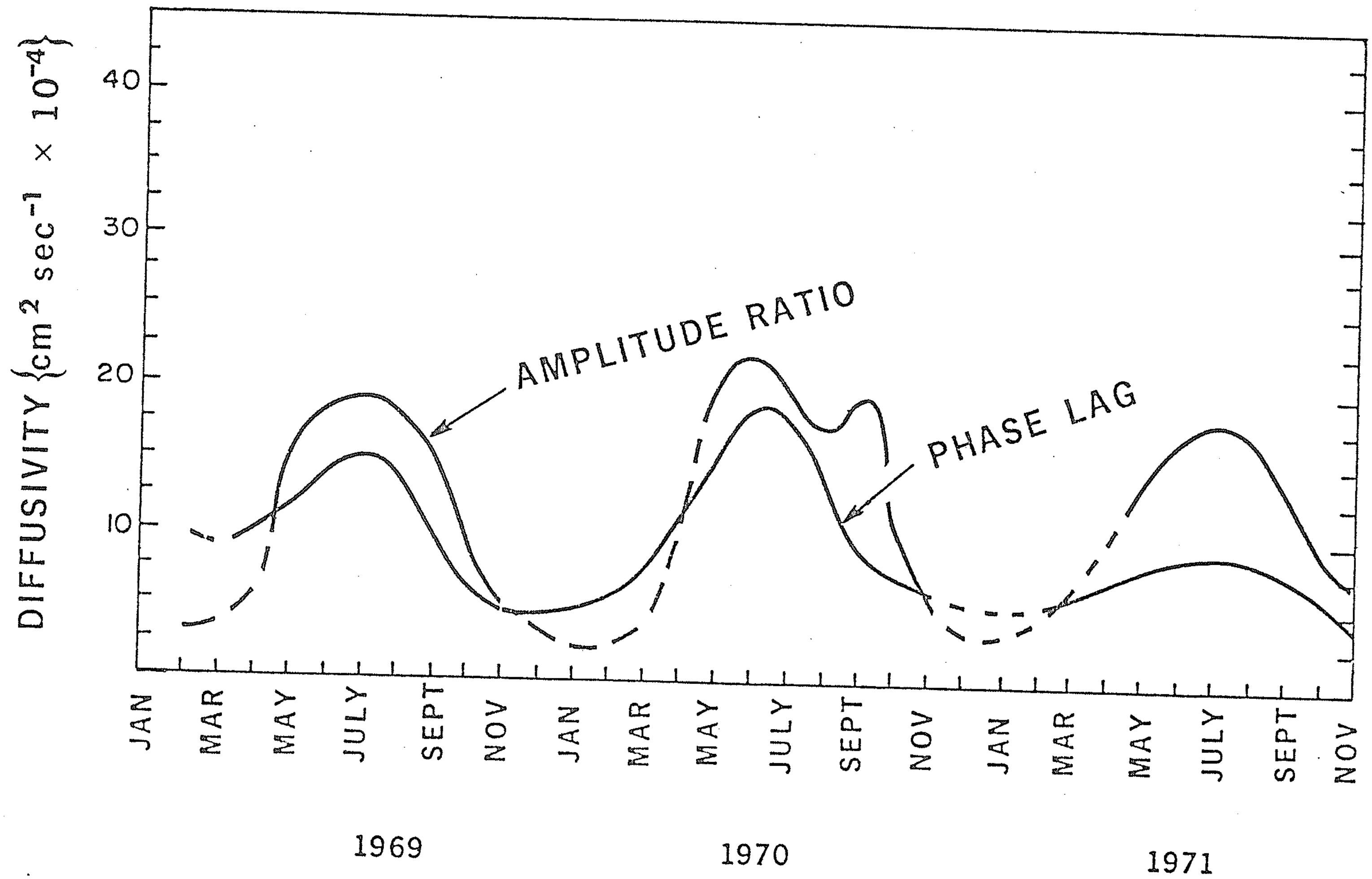

Figure 18. Seasonal variation of thermal diffusivity in the top 10 cm of soil at WNRE by amplitude ratio and phase shift methods (dotted line interpolated).

APPENDIX 1

Spectrum analysis was done by IMSL (1977) subroutine FTFREQ on a CDC 6600 computer. Given a time series such as air temperature or sunshine the program calculates the auto-covariance and the power spectrum of the series (Dixon 1965, Jenkins and Watts 1968).

The power spectrum of the daily maximum air temperature and daily hours of bright sunshine for five January months (1969-1973) was obtained for each variate with FTFREQ. These five power spectra were then ensemble-averaged to produce the final spectrum estimate (Figure 17).

White noise filtering was not applied since the whole frequency region is of interest and since ensemble-averaging has somewhat the same effect. Since the annual temperature wave has a maximum in winter with a rather flat peak in January no detrending is required.

The lowest frequency that can be resolved is usually assumed to have a period about 1/5 of the sample length (Munn 1965), which is onecycle in about six days in this case. The upper limit of resolution, the Nyquist frequency, is one cycle in two days.

Within this range there is one peak in the AMAX spectrum and what appear to be two peaks in the SUNS spectrum. The AMAX peak of 2 to 2.5 days exceeds the 5% confidence limit for a 1 point peak and is just equal to the 1% confidence limit for a 1 point peak, using the chi square and degrees of freedom criterion of Panofsky and Brier (1968). The variance accounted for by this peak is obviously small.

In the SUNS spectrum the peaks at 4 days and 2.5 days are just significant at the 5% level for one point peaks. Since these peaks are also two point or wider they likely are significant at the 5% level.

Significance tests for meteorological data are always open to question because successive data points may not be independent. However since the final estimates are the average of five independent data sets one can hardly reject the reality of these periodicities completely (Misra 1971, Tilley and McBean 1973, Reimer et al 1974).

Further work should be done to verify these findings and also to broaden the limits of resolution at both ends of the spectrum to attempt to identify a one day cycle, if one exists in winter, and to investigate low frequencies as well.

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