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The Estimation of Soil Moisture Content and Actual Evapotranspiration Using Thermal Infra-Red Remote Sensing

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Abstract

Two digital models for estimating daily evaporation and soil moisture have been developed as part of the Tellus Project, sponsored by the Joint Research Centre of the European Community. One, for grassland areas, relates surface temperatures to evaporation, and the other, for bare soils, relates day and night temperatures to both evaporation and thermal inertia, which may in turn be related to soil moisture content. Surface temperatures may be estimated using measurements of thermal emission, such as those recorded using an infra-red line scanner.

These models have been tested through a flight experiment at Grendon Underwood, in Buckinghamshire. Whilst primarily research tools, it is hoped that in the future similar models may be used operationally to give areal estimates of mean soil moisture and evapotranspiration. To this end, data from the Heat Capacity Mapping Mission are being used to test the models.

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Introduction

Estimates of soil moisture content and of evapotranspiration are useful for hydrological and agricultural purposes; accurate estimates can on occasions lead to considerable cost savings, by more effective planning of water supply, irrigation and drainage systems. Regional estimates of soil moisture content or of evapotranspiration may nevertheless be difficult to obtain because of measurement problems. One method of estimating both soil moisture content and evapotranspiration is to use a model of the energy balance of the soil surface and of the atmospheric boundary layer; such a model may be written in a form that most of the input data used are routinely observed, and may be adapted so that remotely-sensed measurements may be included.

It is possible to estimate the evaporation for one instant by solving the transport equation for evaporative heat transfer, and also to estimate variables related to instantaneous soil moisture content from the diffusion equation. However, the users of soil moisture content or evapotranspiration data generally require not merely instantaneous values, but cumulative values for particular time intervals, particularly daily values. In order to do this, it is necessary either to have a time series of measurements of soil moisture or of evapotranspiration, or to use a model which estimates these quantities from variables which are routinely measured continuously. While continuous measurements of soil moisture or of evapotranspiration may be accurate, it is difficult to deploy a large number of such sensors, and so the spatial variations of either variable would only be poorly estimated, although these variations may be considerable. The employment of a digital model is thus very desirable. In order to employ such models, it is necessary to estimate the components of the energy balance of the soil surface for each of the time steps of measurements input.

The energy balance of the soil or crop surface is governed by a continuity equation

$$R_N + G + H + L.E = 0$$

where R_N is the net radiation flux, G the soil heat flux, ~~if~~ H the sensible heat flux in the atmospheric boundary layer and L.E. the latent heat flux. L.E is the product of E, the evapotranspiration flux, and L, the latent heat of vaporisation of water per unit mass. Conventionally, downward fluxes are given a positive value.

Consider the components of the radiation balance in more detail. The net radiation flux is the sum of the incoming and outgoing short and long wave radiation fluxes:

$$R_N = (1 - \alpha_s)R_s + (1 - \alpha_1)R_1 - \varepsilon_c \sigma T_c^4 \qquad (2)$$

where R_s is the measuring short wave and R_1 the incoming long wave radiation, α_s and α_1 the short and long-wave soil/crop reflection coefficients, ε_c the soil/crop emission coefficient, σ the Stefan Botzmann constant, and T_c the soil/crop surface temperature. Radiation is considered short-wave up to a wavelength of 1.1 µm. For long-wave radiation, $\varepsilon_c = 1 - \alpha_1$, so equation (2) may be written

$$R_N = (1 - \alpha)_s R_s + \varepsilon_c (R_1 - \sigma T_c^4) \qquad (3)$$

The soil heat flux G through any horizon $z = z'$ is related to the temperature gradient in the soil $\partial T/\partial z$ and the thermal conductivity λ, both evaluated at z':

$$G = -\lambda \frac{\partial T}{\partial z} . \tag{4}$$

This relationship leads to a form of the diffusion equation

$$\frac{\partial}{\partial z}\left(\lambda \frac{\partial T}{\partial z}\right) = C_v \frac{\partial T}{\partial t}$$

where C_v is the volumetric heat capacity of the soil. This equation may be solved numerically for each level z' to give a temperature profile using a finite element or finite difference method, taking the soil/crop surface temperature as the upper boundary condition, and some known temperature at a given depth in the soil as a lower boundary condition.

The sensible heat flux may be estimated using a transport equation, depending on the air temperature T_a, the soil/crop surface, T_c, the aerodynamic resistance, r_a, and the air density ρ and specific heat c_p:

$$H = \rho c_p \frac{(T_a - T_c)}{r_a} \tag{5}$$

The latent heat flux may also be expressed as a transport equation

$$L.E = \frac{\rho c_p}{\gamma} \frac{(e_a - e_s)}{r_a + r_s} \tag{6}$$

where γ is the psychrometric constant, e_a the water vapour pressures in the atmosphere layer, e_s the saturated water vapour pressure at temperature T_c, and r_s the stomatal diffusion resistance to water vapour transport. The quantities e_a and e_s may be estimated from T_a and T_c.

By comparing (3), (4), (5) and (6), it may be seen that the components of the energy balance may be expressed as functions of the air temperature, the temperature at a given depth in the soil, and the soil surface temperature, as well as of on the incoming radiation and the aerodynamic and stomatal resistances. Most of these variables may be measured or estimated from routine observations, which are widely collected continuously in time. The exception is the crop/soil surface temperature, T_c, which is not routinely measured except at a few sites. However, if a time series of estimates of the crop surface temperatures may be made then the evapotranspiration flux E and the soil moisture may be estimated. Such estimates may be provided using digital models.

Digital models

Equations (3), (4), (5), and (6) may be re-written to give an expression for the soil/crop surface temperature T_c in terms of routine measurements. The initial estimate of the surface temperature will probably not satisfy the continuity requirement of equation (1), because of simplifications which are necessary to estimate parameters such as r_a, and so it is necessary to use an iterative numerical optimization technique to ensure that the condition in equation (1) is satisfied. This procedure may be adopted to give estimates of the surface temperature T_c corresponding to every time step of a set of routine measurements of air temperature, wet-bulb temperature, wind speed and incoming short and long wave radiation.

Although continuous measurements of surface temperature are not made routinely, it is nevertheless possible to make instantanteous measurements, by using a radiometer to measure the thermal emission from the soil /crop surface. The measured emission may be related to the surface temperature, as shown in the next section. It is possible to compare the measured surface temperatures with those estimated from a model, and then to adjust the parameters of the model until the estimated surface temperatures are the same as the measured surface temperatures at the same time. These parameters such as the aerodynamic resistance r_a and the thermal heat capacity of the soil C_v, may then be used to give more accurate estimates of daily evapotranspiration or soil moisture. Two models have been developed using these methods. One model, the "Tergra" model, developed by Soer (1977), is for use in grassland areas and, using one measurement of the soil surface temperature, gives estimates of the daily evaporation. The other, the "Tellus" model, developed by Rosema (Rosema *et al.*, 1978), is for use on bare soil, and estimates both daily evaporation and thermal inertia using measurements of both day-time and night-time surface temperatures. The thermal inertia p may be defined as

$$p = (\lambda C_v)^{\frac{1}{2}}$$

The relationship between thermal inertia and soil moisture may be determined experimentally. Both models are layered models, which require known boundary conditions at a given height in the atmospheric boundary layer and at a given depth in the soil in order to solve the equation (1) using equations (3), (4), (5) and (6). A time series of measurements of incoming radiation, temperature, vapour pressure, and wind speed at (say) 2 m is then input to both models, together with soil conditions at a sufficient depth in the soil that they may be assumed not to vary during the period over which the models are applied. For each forward time step, equation (1) is solved to estimate the components of the energy balance, which may then be used to estimate the quantities of interest in each model. Each model will now be considered briefly, to point to the differences in the way equation (1) is solved in each case.

The Tergra model for grassland areas uses a Businger-Dyer approach to estimate the turbulent diffusion resistance r_a, which is used in equations (5) and (6). This gives r_a as a function of wind velocity, the stability of the atmospheric boundary layer just above the surface, and the nature of the surface. Different conditions are applied under stable, neutral and unstable atmospheric conditions. The stomatal resistance, r_s, used in equation (6), is found from a set of empirical relationships which relate it to the crop height, the short-wave radiation and the leaf-water pressure (Soer, 1977).

The soil heat flux is estimated using an explicit finite difference method with the spacings between each node being equal. The iteration of the surface temperature is also simple, the surface temperature being altered by successive small amounts until an energy balance has been achieved. This approach is satisfactory where the components of the energy balance are not changing rapidly between successive time steps, so that the initial estimate of the surface temperature, (the temperature estimated from the previous time step) is close to the final estimate. As noted above, the Tergra model directly estimates the daily evaporation. Changes in the soil moisture may also be estimated from the evapotranspiration flux, E, estimated in equation (6).

Given the evaporation, soil moisture is obtained by a set of relationships as follows. For grassland, E may be expressed as

$$E = \frac{1}{g} \frac{\psi_l - \psi_s}{r_{plant} + r_{soil}} \qquad (7)$$

where ψ_1 is the leaf water pressure, ψ_s the soil water pressure, r_{plant} the plant resistance for water transport, and r_{soil} the soil hydraulic resistance. Given estimates of ψ_1, and the resistances r_{plant} and r_{soil}, ψ_s may be estimated. These may be found using empirical relationships with other variables (Soer, 1977) and then ψ_s may be related to the volumetric water content, θ, by

$$\psi_t^{-m} = \frac{S - S_r}{1 - S_r} \qquad (8)$$

where S is the saturation, defined as θ/θ_s (θ_s being the moisture content at saturation). S_r is the rest saturation, m is a pore size distribution factor, and ψ_t is a re-scaled soil water pressure equal to ψ_s/ψ_a, where ψ_a is the air entry value.

The Tellus model for bare soil sites uses two separate estimates of surface temperature, ideally at the times of maximum and minimum thermal emission (about 0200 and 1400 local time), and so allows two unknowns to be estimated directly, without the need to use empirical relationships such as (7) and (8) to derive one of these unknowns. In this model, thermal inertia and evaporation and transpiration are estimated; an experimentally-derived relationship may be found between thermal inertia and soil moisture. Whilst the general structure of the model is similar to that used in the Tergra model, there are some differences. The soil heat flux is again found using an explicit finite difference method, but in the Tellus model the spacing between nodes is allowed to increase downwards, using a Dufort and Frankel method. The iteration at each time step is also rather faster than for the Tergra model, as a Newton-Ralphson method is used; this converges rapidly.

The most critical input . , in the sense of being most difficult to estimate, is the surface temperature. Other inputs are either measured, or may be estimated. The roughness length, which may be difficult to estimate and is used in the estimation of the aerodynamic resistance, is estimated using empirical relationships, as suggested by Monteith (1973).

Estimation of surface temperature using radiometers

Radiometers measure the radiant energy received at the sensor in a set of wavelengths. When directed at an area of the ground surface, the energy received must first be related to the radiant emission from that area which in turn must be related to the temperature of that area. The radiometer itself is calibrated using two black body radiators, one hotter and one colder than the area of ground being studied.

The relationship between emission and temperature usually employed - Stefan's Law - is only valid when the emission is integrated over all wavelengths. However, it has been shown (eg Scarpace, 1974) that to use Stefan's Law for the 8 - 14 µm waveband often adopted for radiometers introduces only very small errors, usually less than 0.03K. The temperature of the soil surface may then be estimated from the measurements E_* by a relationship of the form

$$E_* = (1 - \varepsilon_c)\, \varepsilon_s \sigma T_s^4 + \varepsilon_c \sigma T_c^4 \qquad (9)$$

where ε_c is the emissivity of the crop or soil surface, ε_s is the sky emissivity, T_s is the apparent sky temperature (K) T_c is the crop surface temperature, and σ is Stefan-Boltzmann's constant. If the emission as measured at the radiometer is digitised by dividing it into equal-sized steps between the two calibration black-body emissions on the radiometer, the measured emission E_* may be estimated; if (for example) the data are digitized into 8-bit numbers, scanner step 255 will be set equal to the emission at the upper calibration temperature, scanner step 0 set equal to that at the lower calibration temperature, and E_* is estimated as

$$E_* = \frac{255 \times \varepsilon_* \sigma T_0^4 + \varepsilon_* \sigma T_{255}^4 \times SS - \varepsilon_* \sigma T_0^4 \times SS}{255} \qquad (10)$$

where ε_* is the emissivity of the "black body" calibration on the radiometer, T_0 and T_{255} the temperatures of the lower and upper calibration "black bodies", and SS the scanner step equivalent to the measured ground temperature. By substitution of (10) into (9), the soil surface temperature, T_c, may be found in terms of the scanner step, SS. It is necessary to know the emissivity of the soil surface, and the apparent sky temperature and its emissivity, as well as the emissivity of the calibration black bodies, their temperatures and the scanner step of the area being examined. These may be found from sample ground measurements: the temperature estimated is not very sensitive to these parameters for typical bare soil or short vegetation, where the emissivity is high. Figures 1 and 2 show the absolute errors which are produced for a typical set of values, if the sky temperature and crop emissivity are incorrectly estimated. For this example, the correct values are a sky temperature of 270 K, crop and sky emissivities of 0.95, a black body emissivity of 0.99 and calibration temperatures of 273 K and 291 K and a crop temperature of 281 K. The sky temperature was altered by $\pm$ 20K in 1K steps and the crop emissivity was allowed to vary from 0.76 to 1.0, the absolute errors in temperature being noted. It may be seen that the sky temperature need only be found within 15K of the correct value, and the emissivity need only be estimated within 0.05 for the crop surface temperature to be estimated to within 0.5K of the correct value.

Atmospheric effects are also very important, particularly when measurements from satellites are used. Although it is possible to construct equations to estimate the absorption, scattering and re-emission of radiation using Mie theory, it is not practically possible to collect enough data to allow these equations to be solved, and so approximations are necessary. One such approximation where the assumption is made that all radiation is absorbed and re-emitted by water vapour has been suggested by Rangaswamy *et al.* (1978). In all cases it is necessary to have information about ground temperatures and about the temperature and humidity profile of the atmosphere.

Application of the models

This project is part of the Tellus project, sponsored by the Joint Research Centre, Ispra. As part of this project a Joint Flight Experiment was performed at Grendon Underwood, Buckinghamshire in September 1977. Most of the components of the energy balance were measured directly, apart from actual evaporation and transpiration, which are difficult to measure. Two sites were monitored in detail, one field of bare soil and one field of grass of length 5 - 10 cm. Two flights were made, using a Daedelus DS-1250 multi-spectral scanner with a thermal infra-red sensor in the 8 - 14 μm waveband, one flight being at 5 am and the other at 2 pm BST. Simultaneous ground measurements with PRT-5 radiometers were made; these measure emission in the 8 - 14 μm waveband with a 2^{o} field of view. Temperatures estimated from the aircraft and ground-mounted radiometers were very similar for both day-time and night-time data.

Some preliminary results of modelling are presented in Table 1. Mean soil surface temperatures were estimated from radiometer data for the bare soil site and the graasland site. Both morning and afternoon temperature estimates were input to the Tellus model, and the morning temperature estimate only was input to the Tergra model. For comparison, the mean volumetric soil moistures are included, taken from 15 cm core samples for each site. The actual evaporation for 13th September 1977 is difficult to estimate. However, if the run-off from the River Ray catchment within which the test areas are sited is subtracted from the rainfall then some idea of the evaporation may be obtained if change in water storage within the catchment is neglected. This obviously involves many assumptions about the behaviour of the catchment, but if this exercise is performed for data for the month of September 1977, and then divided to give a daily figure, then this may give an estimate of the actual evaporation to an order of magnitude. Such a figure is included in Table 1. As 13th September, the day of the flight experiment, was clear and rainless, the actual evaporation and transpiration were probably above the mean figure given. The Tergra model rather underestimated the evaporation and transpiration, but as the results are only preliminary the reasons for this are not yet understood.

Further analysis of these data is continuing. It is hoped that data from the Heat Capacity Mapping Mission for the U.K. area will also be used as inputs to the models. Much further work is needed to test the models outlined here. Nevertheless, the preliminary results presented are very encouraging and so it may be hoped that regional estimates of soil moisture and of evaporation and transpiration will soon be available using this approach.

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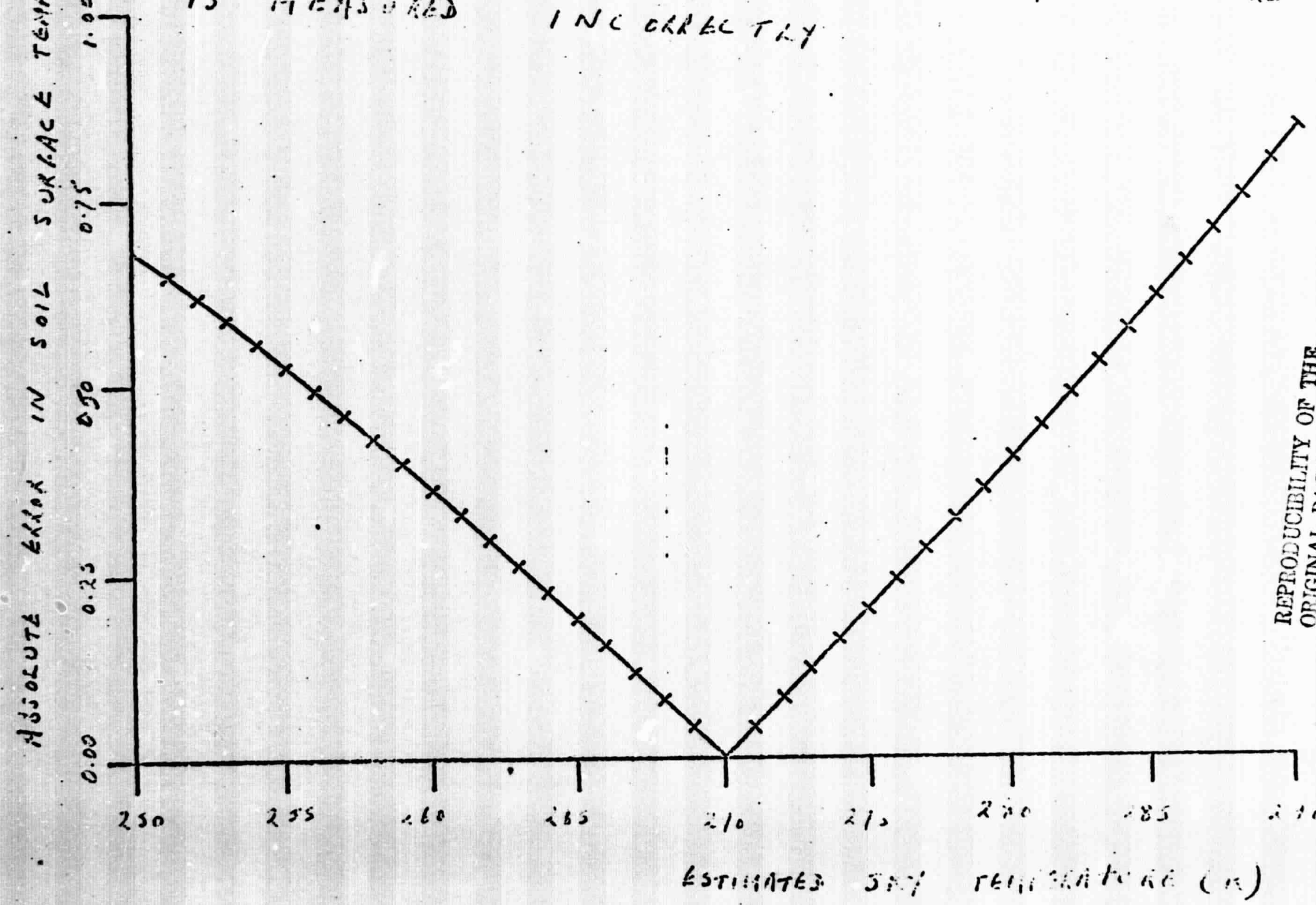
FIGURE 1: ERRORS OF ESTIMATION OF SURFACE TEMPERATURE FROM RADIOMETER DATA WHEN SKY TEMPERATURE IS MEASURED INCORRECTLY



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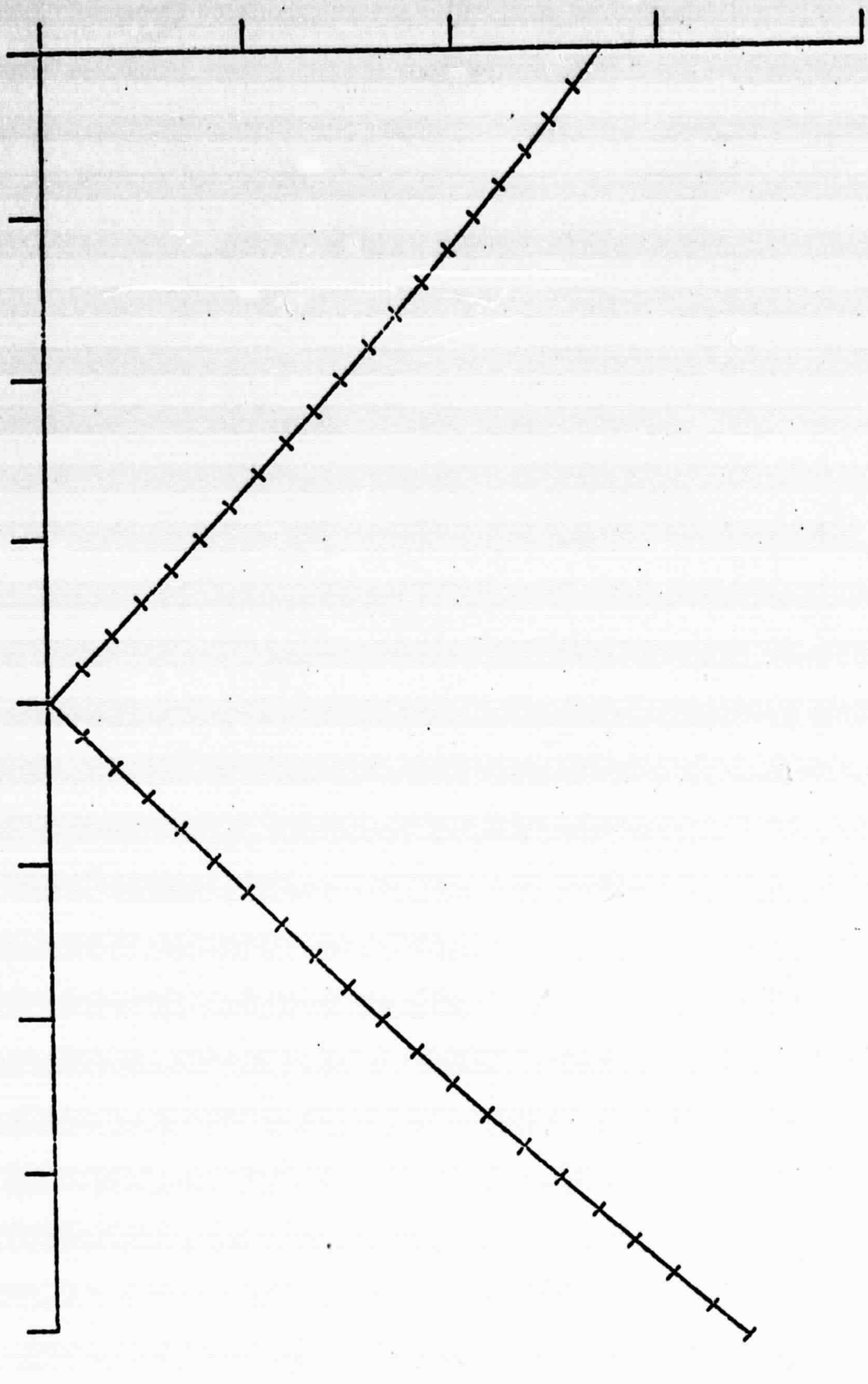

FIGURE 2: ERRORS OF ESTIMATION OF SURFACE TEMPERATURE FROM RADIOMETER DATA WHEN SURFACE EMISSIVITY IS ESTIMATED INCORRECTLY

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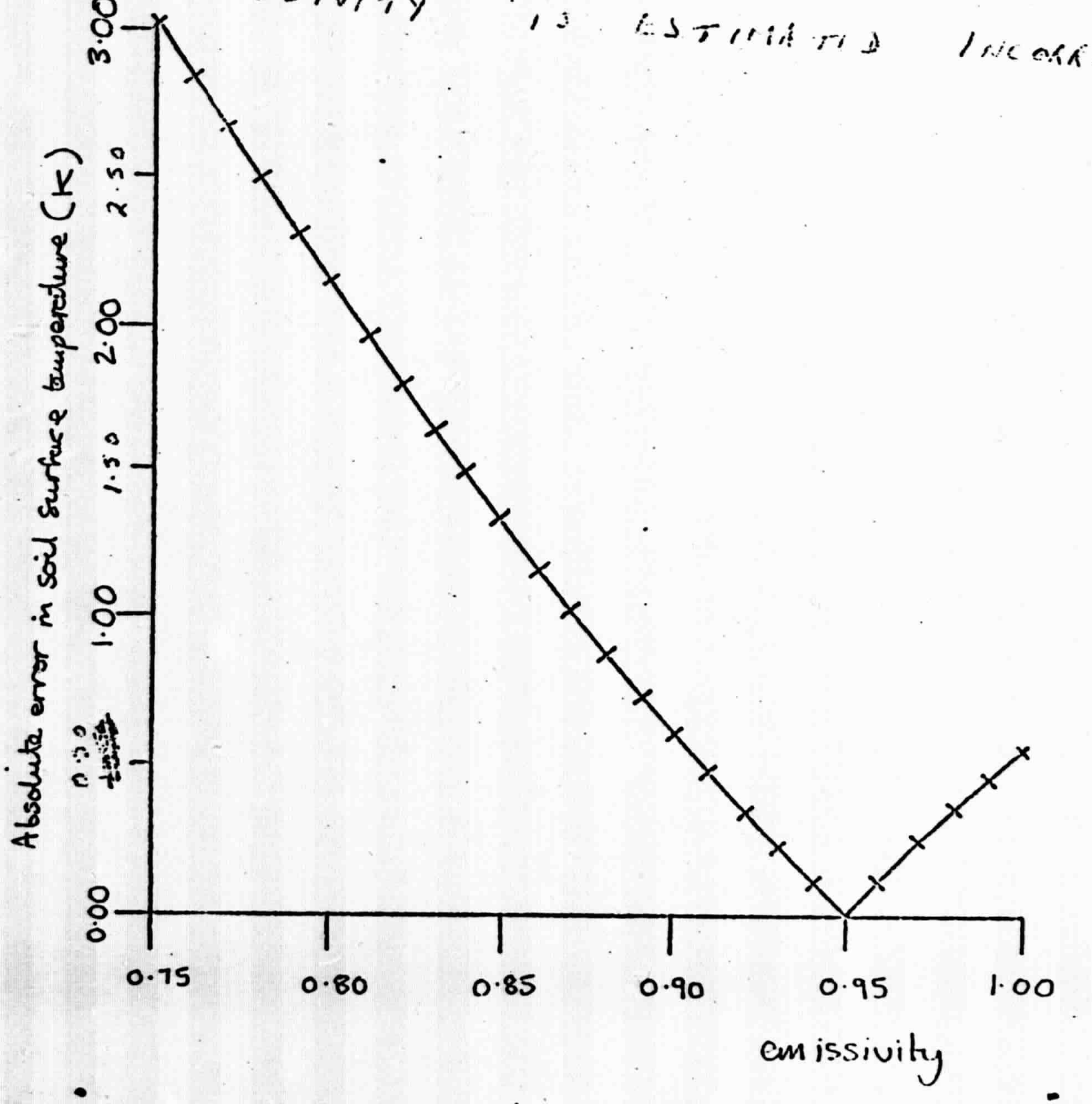

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