SIMULATING TROPICAL BARE SOIL SURFACE TEMPERATURE USING THE

FORCE-RESTORE METHOD

BY

OLANIRAN JONATHAN MATTHEW

B.Sc. Educ. (Physics), OAU, Ile-Ife.

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DEDICATION

This thesis is dedicated to the glory of God Almighty.

CERTIFICATION

This is to certify that this work was carried out by Mr. MATTHEW, Olaniran Jonathan under my supervision and that to the best of my knowledge it has not been submitted elsewhere for the award of a degree.

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Date

Supervisor

Prof. O. O. Jegede

Department of Physics

Obafemi Awolowo University

Ile-Ife, Nigeria

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OBAFEMI AWOLOWO UNIVERSITY

TABLE OF CONTENTS

LIST OF FIGURES

LIST OF TABLES

ABSTRACT

The force-restore (FR) method is a simple and efficient approximation procedure that can be used for predicting soil thermal characteristics such as ground heat flux and the surface temperature. It uses a prognostic energy-balance equation and the response to a periodic heating caused by diurnal variation of the solar radiation. Several variants of the FR method exist and have been developed and tested for the high latitude regions, but is rare for the tropical areas. In this particular study, a force-restore (FR) model is adapted to simulate bare soil surface temperature and other thermal properties at an experimental site a (tropical location) chosen inside the Teaching and Research Farm of Obafemi Awolowo University Ile-Ife, Nigeria (7.55^{o}N, 4.56^{o}E).

In the model, the surface energy balance equation used does not contain the latent heat flux term. Thus it was assumed that evaporation/condensation processes at the surface was neglected. The relative contributions of the imposed initial conditions; deep ground temperature, soil thermal characteristics, surface temperature, and cloud cover fractions, were used to investigate the performance of the FR model. The numerical simulations of the energy fluxes and surface temperature were then compared with the observed field data.

The FR method worked quite well for estimating the diurnal variation of bare (dry) soil surface temperature within a range of ± 2.0^{o}C from actual values. On clear and relatively dry days, the simulated net radiative flux attained a maximum value of (420± 10)Wm^{-2} at about 1400hrs local time.

It was found that the bare soil surface temperature depends essentially on the soil thermal properties, deep soil temperature and the cloud cover fractions. The result of this study is found useful

for agrometeorological purposes particularly for predicting the soil surface temperature and frost conditions.

CHAPTER ONE

INTRODUCTION

1.1 Background to the Study

The primary forcing of the atmospheric boundary layer over the land is through absorption of solar radiation at the ground surface. Such generally results in diurnal variation of both the soil thermal characteristics (e.g. surface temperature) and the turbulent heat fluxes. The surface temperature at a given location depends on the radiation balance, atmospheric exchange processes in the immediate vicinity of the surface, presence of vegetation or plant cover and thermal properties of the sub-surface medium.

In the absence of vegetation (i.e. bare soil) the atmosphere-surface interface is relatively well defined. Conservation of energy at the interface both instantaneously and averaged in time requires that:

$$R_n = H_s + H_l + H_g \qquad 1.1$$

which is a surface energy balance equation. Here R_n is the net radiation flux density at the surface, H_s is the sensible heat flux, H_l is the latent heat flux and H_g is the heat flux into the soil. The net flux (R_n-H_g) represents the "available energy". This is the amount of energy available for conversion into sensible heat, H_S, and latent heat, H_l, at the surface.

The sub-soil temperature can be measured by a variety of instrumentation such as resistance wires, thermocouples, thermistors, etc. The surface temperature is often determined by extrapolation of measured temperature profiles in soil and air with the knowledge of their expected theoretical behaviors. Another, perhaps better ,method of determining the surface temperature when emissivity of the surface is known is through remote sensors, such as a downward–looking radiometer which measures the flux of out-going longwave radiation from the surface and hence the surface temperature,T, using the modified Stefan–Boltzmann equation is given as:

$$R_l = -\varepsilon\sigma T^4 \qquad 1.2$$

where R_l is the outgoing longwave radiation ,ε is the emmisivity of the surface and σ the Stefan-Boltzmann constant given as 5.67 x 10^{-8} $Wm^{-2}K^{-4}$. The times of minimum and maximum in (soil) surface temperatures as well as the diurnal range are of considerable interest to micrometeorologists. For example, the knowledge of surface and soil temperatures is of immense importance in the studies of microenvironment of plant cover including the root zone and surface energy balance as well as in the prediction of frost condition. On clear days, the maximum surface temperature is attained typically an hour or two after the time of maximum insolation while the minimum temperature is reached in early morning hours. The maximum diurnal range is achieved for a relatively dry and bare surface under relatively calm air and clear skies. For example, maximum surface temperature occur at about 1400hrs local time in Ile-Ife, Nigeria, while minimum is attained in the early morning hours (Babatunde, 1980).

The presence of moisture at the surface and in the subsurface soil greatly moderates the diurnal range of surface temperatures. This is due to the increased evaporation from the surface and also due to increased heat capacity and thermal conductivity of the soil. Balogun (2000) in an independent investigation ascertained that the soil heat capacity, thermal conductivity and diffussivity vary linearly with the soil moisture content.

Thermal properties of soils also depend on the properties of the solid particles and their size distribution as well as the porosity of the soil. For example, fine textured soils (e.g. clay) have greater heat capacities as compared to coarse soils (e.g. sand). Over a wet, bare soil about 65% of the net radiation is utilized for evaporation in the beginning but the percentage reduces as the soil surface dries up (Balogun, 2000). Increased heat capacity of the soil further slows down the warming of the upper layer of the soil in response to radiative heating of the surface. The (soil) surface heat flux is also reduced by evaporation. The range of soil temperatures decreases exponentially with depth and becomes insignificant at a depth of the order of 1m (Babatunde, 1980). There is damping of the temperature wave in the soil such that the temperature fluctuates more at the surface than at 0.10m, more at 0.10m than at 0.20m, and so on (Hanks and Ashcroft, 1986). There is also an appreciable lag between the times of the occurrences of maximum of the temperature at the surface and in the soil as the heat wave penetrates downwards.

The presence of vegetation on the surface reduces the diurnal range of surface temperature. Part of the incoming solar radiation is intercepted by plant surfaces reducing the amount reaching the surface. Therefore surface temperatures during the day are uniformly lower under vegetation than over a bare soil surface. At night the outgoing longwave radiation is partly intercepted by vegetation but the latter radiates energy back to the surface. This slightly slows down the radiative cooling of the surface. Vegetation also enhances the latent heat exchange due to evapotranspiration and increases turbulence near the surface.

The question of the diffusion of heat in response to periodic forcing is a well known problem in classical physics (Carslaw and Jaeger, 1959). This approach is useful for understanding diurnal and seasonal variations of soil temperatures in response to the solar heating at the surface (Sellers, 1965). The "force- restore" (FR) method (Blackadar, 1976) uses a prognostic equation for soil temperature that reproduces exactly its response to periodic heating (diurnal variation). The method involves the use of a finite difference scheme to solve the heat diffusion equation by stepping forward in time from some known initial conditions (e.g., deep ground temperature, air temperature, soil thermal parameters, surface temperature, and cloud cover fractions).