EVAPORATION FROM NATURAL NONSATURATED SURFACES

A Thesis Submitted to the College of
Graduate Studies and Research
in Partial Fulfilment of the Requirements
For the Degree of Doctor of Philosophy
in the Department of Agricultural Engineering
University of Saskatchewan
Saskatoon

by

Raoul Joseph Granger

TABLE OF CONTENTS

Page

ABSTRACT

Evaporation from nonsaturated surfaces is investigated. The concept of potential evaporation is first examined; a series of definitions are developed and classified, and it is shown that the relationship between potential and actual evaporation rates depends on the controlling variables in the chosen definition for potential evaporation.

Extending the work of Penman (1948) to the unsaturated case, a general equation is derived to describe evaporation from nonsaturated surfaces. Applying Bouchet's (1963) hypothesis with a consistent set of definitions also leads to the same general equation. To account for departures from the saturated condition, the equation makes use of the concept of relative evaporation, the ratio of the actual evaporation rate to the rate which would occur under the prevailing atmospheric conditions if the surface was saturated at the actual surface temperature. A relationship is established between the relative evaporation, G, and a dimensionless parameter called the relative drying power, D, the ratio of the drying power (the evaporation rate which would occur if the surface was saturated at the actual air temperature) to the sum of the drying power and the net available energy. The relationship is non-dimensional and appears to be single-valued.

An experimental investigation of evaporation from bare soil and growing wheat is carried out; data from an energy balance installation show that the G-D estimates of evaporation are in close agreement with calculations obtained using the Bowen ratio approach. The data are also used to refine the relationship between relative evaporation and relative drying power. The G-D estimates of evaporation also agree closely with independent estimates obtained from the soil water balance at three sites during the growing season.

The combination of this G-D relationship with the derived general evaporation equation constitutes a simple model for obtaining estimates of evaporation from nonsaturated surfaces; no prior estimate of the potential evaporation is required, and the surface conditions of temperature and humidity need not be known.

A preliminary relationship is found for the vapour transfer coefficient (used in a Dalton type transfer equation) for daily time periods. Algorithms are presented for the estimation of daily net radiation and for soil heat flux.

A new approach is proposed for the application of remotely-sensed surface temperature data to the estimate of regional evaporation. A relationship is derived between the surface temperature and the vapour pressure deficit in the air. The relationship allows for the use of remotely-sensed data in evaporation models such as that presented above. The method is shown to provide superior results to the simplified energy balance approach currently being applied.

RÉSUMÉ

L'évaporation des surfaces non-saturées est étudiée. Dans un premier temps le concept d'évaporation potentielle est examiné, une série de définitions est présentée, et il est démontré que la relation entre les taux d'évaporation potentielle et réelle dépend des variantes caractérisant la définition d'évaporation potentielle en question.

L'extension de l'approche de Penman (1948) conduit à une équation générale décrivant l'évaporation d'une surface non-saturée. L'application de l'hypothèse de Bouchet (1963) avec les définitions appropriées conduit également à cette même équation. L'équation fait appel au concept d'évaporation relative, soit le rapport du taux d'évaporation réelle à celui correspondant à une surface saturée à la même température et sous les mêmes conditions atmosphériques. Une relation est établie entre l'évaporation relative, G, et un deuxième paramètre adimensionné, D, le pouvoir évaporatif relatif, soit le rapport du pouvoir évaporatif (le taux d'évaporation correspondant à une surface saturée à la température de l'air) à la somme du pouvoir évaporatif et de l'énergie disponible. Cette relation semble se prêter à une application générale.

Une étude expérimentale de l'évaporation d'un sol nu et d'une couverture végétale est effectuée utilisant une installation à bilan énergétique; les données obtenues démontrent que la méthode G-D produit des estimations de l'évaporation qui correspondent de près avec les calculs produits par la méthode du rapport Bowen. Ces données servent également à raffiner la relation entre l'évaporation relative et le pouvoir évaporatif relatif. Les estimations produits par la méthode G-D correspondent également de près avec les résultats du bilan d'eau du sol obtenu à trois emplacements durant la période d'étude.

L'utilisation de la relation G-D avec l'équation générale développée ici constitue une approche simple pour le calcul de l'évaporation des surfaces non-saturées; elle requiert ni un calcul préalable de l'évaporation potentielle, ni une mesure des conditions de température et d'humidité à la surface.

Une relation tentative est présentée pour le coefficient de transfert de vapeur pour un pas de temps journalier. Des algorithmes sont présentés pour le calcul du rayonnement net et du flux de chaleur dans le sol.

Une nouvelle approche est proposée pour l'application de la télédétection à l'estimation de l'évaporation régionale. Une relation est établie entre la température à la surface et le déficit de pression de vapeur d'eau dans l'air. Cette relation permet l'utilisation directe des données obtenues par télédétection dans un modèle d'évaporation tel que celui présenté ici. La méthode donne des résultats supérieurs à ceux produits par l'approche du bilan énergétique simplifié couramment utilisée.

ACKNOWLEDGEMENTS

The author wishes to express his sincere gratitude to Dr. D. M. Gray for the encouragement he provided at the outset and patience he has shown during the course of this study. His insightful advice, both technical and editorial, is respectfully acknowledged.

The author also wishes to express his appreciation to the members of the Division of Hydrology, College of Engineering, for their friendship and their untiring assistance in the collection of soil moisture data.

The financial support provided by the Atmospheric Environment Service, Environment Canada, Science Subvention Program, and by the National Sciences and Engineering Research Council of Canada and Atmospheric Environment Service Research Partnership Program is gratefully acknowledged. The author also wishes to express his gratitude to the National Hydrology Research Institute for the financial and technical support received during the course of this study.

L'auteur doit aussi une note de reconnaissance aux membres de sa famille qui ont su lui apporter un appui constant et faire preuve d'une patience examplaire tout au long de cette étude. Cette oeuvre leur est dédiée.

LIST OF FIGURES

LIST OF TABLES

LIST OF SYMBOLS

a, b, c	regression coefficients
A,B	regression coefficients
C_p	specific heat of air [kJ/kg·°C]
D	relative drying power
d_0	displacement height for rough surfaces [m]
e	vapour pressure [kPa]
e^*	saturation vapour pressure [kPa]
E	evaporation rate [mm/d]
E_a	drying power [mm/d]
E_p	potential evaporation [mm/d]
E_{pw}	wet-environment evaporation [mm/d]
f(u)	vapour transfer function [mm/d·kPa]
g	acceleration due to gravity [m/s^2]
G	relative evaporation
h	mean height of vegetation [m]
H	sensible heat (L_eQ_h) [mm/d]
$K_{e,h,m}$	turbulent diffusivities for vapour, heat, and momentum [m^2/s]
L	Monin-Obukhov length [m]
L_e	latent heat of vaporization [kJ/kg]
n	hours of bright sunshine
N	maximum possible hours of bright sunshine
P	atmospheric pressure [Pa]

P_n	precipitation [mm]
q	specific humidity [kg/kg]
Q	net available energy [W/m^2 or mm/d evaporation equivalent]
Q_e	latent energy flux [W/m^2]
Q_g	soil heat flux [W/m^2]
Q_h	sensible heat flux [W/m^2]
Q_n	net all-wave radiation [W/m^2]
Q_r	reflected short-wave radiation [W/m^2]
Q_{dfs}	diffuse short-wave radiation [W/m^2]
Q_{drs}	direct-beam short-wave radiation [W/m^2]
Q_{ln}	net long-wave radiation [W/m^2]
Q_{lno}	clear-sky net long-wave radiation [W/m^2]
Q_{sn}	net short-wave radiation [W/m^2]
Q_{so}	clear-sky short-wave radiation [W/m^2]
r_a, r_s	aerodynamic and stomatal resistances
R_o	runoff [mm]
T	temperature [°C]
T_*	characteristic temperature [°C]
u	wind speed [m/s]
u_*	friction velocity [m/s]
z	height [m]
z_0	roughness parameter [m]
z_{0e}	roughness length for water vapour [m]
α	ratio K_e/K_m
β	Bowen ratio

δS	change in soil water storage [mm]
δT	daily change in air temperature [°C]
ΔT	temperature gradient between surface and air
γ	psychrometric constant [kPa/K]
Δ	slope of the saturation vapour pressure vs temperature curve [kPa/K]
ρ	density of air [kg/m^3]
ζ	z/L, dimensionless stability parameter

1. INTRODUCTION

Evaporation from natural surfaces, open water and land surfaces with various vegetative covers, is one of the important phases of the hydrologic cycle. Much of the water that is received at the surface through precipitation is returned to the atmosphere through evaporation[1]. Brutsaert (1982), examining the water budget for the entire earth, indicates that annual evaporation from land surfaces is approximately 65% of annual precipitation; Gray et al. (1970) place this figure near 75%. Unfortunately, evaporation is also one of the least understood components of this water cycle. Whereas the other major components of the hydrologic cycle (precipitation and runoff) lend themselves to direct measurement, and monitoring on an operational basis, practical estimates of evaporation are still obtained indirectly from the measurement of other parameters and through the use of known, inferred or derived relationships. Whereas there is a need to obtain reliable estimates of evaporation on a regional scale, at the present time the determination of the evaporation at a point in operational practice is still problematic.

For evaporation to occur there must be a supply of water at the surface, a supply of energy to satisfy the requirement for the phase change, and a transport mechanism to carry the

---

[1] Throughout this dissertation the term "evaporation" is taken to include both direct evaporation from soil and water surfaces and transpiration from plants.

vapour away from the surface. A great number of procedures have been developed to provide estimates of evaporation. With the exception of the strictly empirical techniques, all these procedures stem from consideration of one or a combination of these three requirements. Brutsaert (1982) presented a comprehensive review and analysis of the various techniques available for estimating evaporation, and provided a complete listing of the pertinent literature to that date. Providing a detailed literature review on the subject would consist to a large degree of a repetition of the excellent work presented by Brutsaert; therefore, herein reference is made only to those works pertinent to the thesis, and the reader is referred to Brutsaert (1982) for a more detailed review.

Mass balance and energy balance techniques are available in which evaporation is determined as the residual in a balance equation for which all the other terms have been independently determined. These balance techniques have the inherent drawback that errors in the measurement or estimate of the various terms accumulate in the estimate of the residual. The mass balance for a watershed is expressed as

$$\delta S = P_n - E - R_o \, , \qquad 1.1$$

where δS is the change in soil water stored, P_n is the precipitation, R_o is the runoff, and E is the evaporation; all terms are expressed as equivalent depth of water over the area. The fluxes of precipitation and runoff can generally be monitored in "real time"; however, the short-term fluctuations in soil water storage are difficult to determine with acceptable accuracy, so that the mass balance techniques are generally not practical for short time periods. They do represent the most reliable means of verifying the seasonal evaporation.

The energy balance at a surface is often expressed in a form which describes the dissipation of the incoming radiant energy into the various other forms of energy flux, that is:

$$Q_n = L_e E + Q_h + Q_g \text{ ,} \tag{1.2}$$

where Q_n is the net all-wave radiation flux, Q_h is the turbulent transfer of sensible heat, Q_g is the heat flux by conduction into the soil, and L_e is the latent heat of vaporization. In Eq. 1.2 the radiation flux is taken positive toward the surface, while the other fluxes are positive away from the surface. In that the fluxes of radiant energy (Q_n), energy conducted in the soil (Q_g), and the turbulent transfer of sensible heat (Q_h) are relatively easy to determine, the energy balance approach lends itself to practical application for estimating evaporation over a wide range of time scales. However, the estimate of the sensible heat flux requires measurements at more than one height as well as knowledge of the relationship between the flux and the gradient, which is presumed to be similar to that for the flux of water vapour; thus the only real advantage of the energy balance approach stems from the fact that temperature measurements are generally more easily obtained and more reliable than are humidity measurements.

When measurements of the humidity gradient above a surface are available, aerodynamic formulations can be used to estimate the evaporation rate. These formulations are of the general form

$$E = -\rho K_e \frac{\partial q}{\partial z} , \qquad 1.3$$

where ρ is the air density, q is the specific humidity, z is the height above the surface. K_e is the turbulent diffusivity for vapour flux, which is generally expressed as a function of the wind speed and is affected by the thermal stratification, or stability, of the boundary layer. The relationship between the flux of water vapour and the mean profiles of wind and humidity has been the subject of much study. These flux-gradient relationships, and the effects of atmospheric stability, are reasonably well established for periods representing quasi-steady conditions, that is, of the order of one half to one hour; however, for daily periods or longer (those periods generally of interest in operational practice), the net effect of the diurnal cycling of the atmospheric stratification on the flux-gradient relationship has yet to be determined or quantified. These profile methods also have the inherent drawback of requiring measurements at more than one observation level, and these are rarely obtained on an operational basis.

Two modifications to the energy balance approach which have proven to be useful both in providing practical estimates of evaporation and as the basis for the development of other relationships warrant mentioning at this point. The first was the introduction by Bowen (1926) of the ratio of the latent and sensible heat fluxes, $\beta = Q_h/L_eE$, into the energy equation (Eq. 1.2) which then becomes

$$L_eE = \frac{Q_n - Q_g}{1 + \beta} \quad . \qquad 1.4$$

Use of the Bowen ratio, based on the assumption of the equality of turbulent transfer coefficients for heat and water vapour, eliminates the need for the measurement of wind speed and thus does not require knowledge of the flux-profile relationships. The method has proven reliable where the gradients are well defined and the soil heat flux term can be determined with confidence.

The second significant modification of the energy balance approach was provided by Penman (1948), who combined the energy and vapour transfer equations, for saturated surfaces, in such a way as to obtain the following general equation:

$$E = \frac{\Delta (Q_n - Q_g) + \gamma E_a}{\Delta + \gamma} \quad , \qquad 1.5$$

where Δ is the slope of the saturation vapour pressure curve at the given air temperature, and γ is the psychrometric constant. The term E_a is called the drying power of the air, and is defined as the evaporation rate that would occur if the surface was saturated at the temperature of the air; it is generally represented by a Dalton-type formulation

$$E_a = f(u)\,(e_a^* - e_a)\ , \qquad 1.6$$

where e_a is the actual vapour pressure of the air, e_a^* is the saturation vapour pressure at the air temperature, and $f(u)$ is a function of the wind speed, u.

The most attractive feature of the Penman equation is that, while it has a sound physical basis, it requires the measurement of temperature and humidity at only one height. Its major drawback is that it is applicable only to saturated surfaces. The Penman equation has been used extensively as an index value with which the evaporation from nonsaturated surfaces can be related, and has formed the framework for the development of other relationships based on soil moisture, vegetative characteristics and other parameters.

The work of Penman (1948) also marks the beginning of the use of "potential evaporation", although the concept was actually introduced the same year by Thornthwaite (1948). It has become a commonly used approach for estimating the "actual" evaporation in operational practice to moderate the "potential" rate according to moisture and energy parameters. Modelling evaporation in this way has the advantages that it is simple, direct and makes use of easily-measured atmospheric variables. A major problem with the approach however is that it requires a prior calculation of "potential" evaporation and despite its importance, potential evaporation has yet to be clearly defined in a universally-accepted manner. There is much ambiguity regarding the concept of potential evaporation; this ambiguity stems from a lack of consensus regarding (i) the approach to defining potential; many different definitions can be found in the literature; and (ii) the relationship between the actual and potential rates; the tendency has been to assume a direct relationship, however, Bouchet (1963) demonstrated that the relationship can be an inverse one. In this thesis the use of the potential evaporation approach will be examined, a classification of the various types of definition will be presented, and the relationship between actual and potential evaporation will be clarified.

Another common approach to the extension of the Penman equation (Eq. 1.5) to the nonsaturated case involves the introduction of a stomatal resistance term (or soil water resistance term in the case of bare soil) to account for the reduced vapour pressure at the surface (Penman and Schofield, 1951; Penman, 1956; Monteith, 1965).

$$E = \frac{\Delta(Q_n - Q_g) + \gamma E_a/r_a}{\Delta + \gamma(1 + r_s/r_a)} , \qquad 1.7$$

where r_a and r_s are the aerodynamic and stomatal resistance terms, respectively. Eq. 1.7 is referred to as the Penman-Monteith equation. Attempts have been made to relate these resistance terms to other parameters; for example, Monteith (1965) relates the bulk stomatal resistance to the Bowen ratio, van Bavel (1967) and Tan and Black (1976) relate it to soil water stress, and Federer (1979) to the plant water potential. Brutsaert (1982) points out that, thus far, causal relationships of this type have not proven to be sufficiently general to be transferable or practical.

In this thesis, a somewhat different approach will be presented for extending the Penman equation (Eq. 1.5) to the nonsaturated case; the approach makes use of the concept of the dimensionless relative evaporation rather than resistance to moisture flow. A general equation for evaporation from nonsaturated surfaces is thus derived.

The relationship introduced by Bouchet (1963), based on the feedback links between evaporation and the atmosphere, provides another framework for the development of techniques for estimating evaporation. Bouchet (1963) postulated that as a large surface dries from initially saturated conditions and the evaporation rate decreases, the humidity of the overlying air will decrease and its temperature will increase, thus increasing the "potential" evaporation rate. Thus Bouchet (1963) showed that the relationship between actual and "potential" evaporation is an inverse one; his relationship is often called the complementary relationship and is given as

$$E + E_p = 2E_{pw} , \qquad 1.8$$

where E_p is the potential evaporation rate, and E_{pw} is the wet-environment evaporation rate, or the evaporation rate from an extensive wet surface.

However, those models which have been based on this approach (Brutsaert and Stricker, 1979; Morton, 1983) have not found general acceptance, primarily because the inverse relationship suggested by Bouchet (1963) appears to be inconsistent with the direct relationship inferred by the proponents of the Penman approach. It will be demonstrated in this thesis that when a consistent set of definitions is used to describe the potential and wet-environment evaporation rates, the Bouchet approach results in the same general equation as is derived using an extension of the Penman approach.

Use of this new general equation requires that an expression be found for the relative evaporation. A relationship will be derived between the relative evaporation and a second dimensionless parameter called the relative drying power; the relationship draws on feedback rather than causal mechanisms, and the experimental verification suggests that it may be applicable over a variety of surfaces.

In the last decade, much attention has been given to the estimation of regional evaporation. The development of techniques for obtaining reliable estimates of the evaporation from large-scale areas requires the availability of data representative of the scale considered as well as algorithms capable of translating the "areal" data into the required regional flux. Seguin (1980) points out that in the case of a heterogeneous region the total evaporation should be obtained through the integration of the rates from the individual land units. Such an approach requires an information network of sufficient density to cover the various landuses present. This can best be achieved through the use of remote sensing techniques.

The only evaporation model designed specifically to provide "areal" estimates is the CRAE model (Morton, 1983). However, it requires reliable humidity data as input, and remote sensing techniques have thus far not provided such data. Review of the literature shows that only two indirect methods for obtaining evaporation estimates using remotely-sensed data are being tested. The first involves a soil moisture balance technique in which the vapour flux is inferred from measured changes in the surface soil moisture content which can be detected using thermal infrared (Price, 1977, 1980; Taconet et al., 1986) or microwave techniques (Wang, 1987; Wang et al., 1987; Autret et al., 1989). In the second method evaporation is calculated as the residual in a simplified energy balance equation in which the sensible heat flux is inferred from the remotely-sensed surface temperature (Soer, 1980; Seguin, 1980; Nieuwenhuis et al., 1985; Kustas et al., 1989; Vidal and Perrier, 1989). The approach involves a linear regression between the daily sensible heat flux and the observed mid-day surface - air temperature gradient (Seguin and Itier, 1983; Jackson et al., 1983; Brunel, 1989).

Both the water and energy balance approaches used are at best very simple approximations of the balance equations, and have as well the inherent limitations listed above. A procedure for the direct determination of evaporation using remotely-sensed data, or for the application of remotely-sensed data in areal evaporation models, is thus a desirable objective. Such a procedure will be derived and evaluated.

2. THE USE OF POTENTIAL EVAPORATION

2.1 Introduction

Potential evaporation has been commonly used as a hydrological parameter for nearly half a century. The term potential evaporation was introduced by Thornthwaite (1948) in the context of the classification of climate: *"There is a distinction, then, between the amount of water that actually transpires and evaporates and that which would transpire and evaporate if it were available. When water supply increases, as in a desert irrigation project, evapotranspiration rises to a maximum that depends only on the climate. This we may call "potential evapotranspiration", as distinct from actual evapotranspiration."* Thornthwaite then classed a climate as moist or dry depending on whether precipitation was greater or less than potential evapotranspiration.

Although the concept of potential evaporation was first developed as a climatic index, it has been most extensively used by agriculturalists, interested in the efficient use of water by crops, as an indication of the potential <u>for</u> evaporation, and by hydrologists as a more predictable, or accessible, index from which the actual evaporation could be estimated. Indeed, the latter is what motivated Penman (1948), who was *"seeking an absolute relation between weather elements and open water evaporation, and comparative relations between losses from the soil and losses from open water* *exposed to the same weather.*" His intent was not to develop or conceptualize a "potential" evaporation. However, perhaps because what Penman calculated (Eq. 1.5) appeared to be similar to what Thornthwaite had defined, his expression came to be known as the "Penman potential evaporation", and has since been extensively used by both agriculturalists and hydrologists.

Potential evaporation has become a standard ingredient in the development of schemes for estimating actual evaporation; many of the evaporation models in use estimate evaporation as some fraction of this "maximum potential evaporation". However, for all its years of existence, and for all the use that it has seen, the concept of potential evaporation has yet to be adequately and clearly defined in a generally accepted manner. There is ambiguity surrounding the definition of the concept (is it governed by available energy, weather or moisture conditions?), and there is uncertainty as to the relationship between the actual and potential evaporation rates (is the relationship direct or inverse?).

A survey of the appropriate literature reveals that there are many different "definitions" available for potential evaporation. These vary from a simple statement equating potential evaporation to evaporation from a free water surface, to verbal descriptions of the Penman equation, and to somewhat elaborate descriptions of conditions under which potential evaporation might be measured. Selecting the appropriate "potential" is not always a straightforward task.

The often used definition equating potential evaporation to evaporation from a saturated or free water surface (ie. Konstantinov, 1963; LeDrew, 1979) is an example of an "open-ended" statement; an infinite number of values are possible if the atmospheric conditions or the conditions of the evaporating surface (ie. temperature) are not specified. This then is an inadequate definition.

Bouchet (1963) recognized the need to specify the conditions for potential evaporation; he defined "*l'évapotranspiration potentielle, l'eau susceptible d'être perdue dans les mêmes conditions quand elle n'est plus facteur limitant*". However, Bouchet did not specify what was meant by the statement "under the same conditions".

Others as well have identified the need to specify conditions for potential evaporation; however, there appears to be no general consensus as to which conditions need be specified. Some (ie. WMO, 1974; Fortin and Seguin, 1975) indicate that potential evaporation represents the "demand" exerted by the climate, and is thus governed by atmospheric conditions. For example, the WMO (1974), in its International Glossary of Hydrology, defines potential evaporation (evaporation capacity) as the "*Quantity of water vapour which could be emitted by a surface of "pure" water, per unit surface area and unit time, under the existing atmospheric conditions.*"

Others (van Bavel, 1966; Priestley and Taylor, 1972) place more importance on the surface parameters. Priestley and Taylor state that the apportionment of energy between sensible and latent heat fluxes will be governed by the dryness of the surface and the surface temperature. van Bavel (1966) indicated that "*the fundamental condition that defines potential evaporation is that the surface vapour pressure can be found from the surface temperature.*" He thus provided the following definition: "*A direct approach is to state that the potential evaporation rate* E_o *shall be defined from a Dalton-type equation:*

$$E_o = f(u)(e_o' - e_a), \qquad 2.1$$

in which f(u) is a function of wind speed at a stated height, e_a *is the vapour pressure of the air at the same height, and* e_o' *the saturation vapour pressure of the surface, determined by its temperature,* T_o."

Some definitions encountered in the literature appear to be descriptions of the equations used to calculate the parameter; for example, the WMO (1974) definition and others like it appear to be attempts to describe the Penman equation. On the other hand, some definitions appear to be descriptions of an "ideal" evaporation pan, or of how one might attempt to measure the potential. For example, Morton (1969) stated "*Potential evapotranspiration is defined as the evapotranspiration that would occur from a continuously moist surface with regional radiation absorption, vapour transfer,* *and heat transfer characteristics, which has an area so small that the transfers of heat and water vapour from the surface have no significant effect on the evaporability of the overpassing air. Thus it may be visualized as the evaporation from a hypothetical potential evaporimeter with reflectivity and roughness similar to those of the surrounding region."* It is unlikely that such calculated and measured potential evaporation rates would concur; however, notwithstanding his verbal definition, Morton (1983) uses a variation of the Penman method to calculate potential evaporation in his Complementary Relationship Areal Evapotranspiration (CRAE) model.

This brief survey shows that not only are there several different definitions in use for identifying the parameter called potential evaporation, but also that these represent distinctly different terms. This in fact represents the major source of confusion surrounding the concept of potential evaporation. Since more than one value is possible, these different terms will be referred to generally as potential evaporation parameters in the following discussions.

Given that more than one potential evaporation parameter can be described, it should be possible to classify these. A first distinction can be made between measured and calculated parameters. Perrier (1977) provides a classification of the "measured" parameters, and suggests appropriate nomenclature for each. These definitions are reproduced in Table 2.1.

Table 2.1. Summary of the definitions describing "measured" potential evaporation parameters (after Perrier, 1977).

EP_*	Theoretical Potential Evaporation	The evaporation from a saturated vegetative cover with complete radiation absorption at the upper surface (a hypothetical situation).
EP	Potential Evaporation	The evaporation from a saturated cover (ie. vegetative cover after rain or irrigation, free water surface).
ETR_m	Maximum Actual Evapotranspiration	The evaporation from a vegetative cover well supplied with water.
ETP	Potential Evapotranspiration	The evapotranspiration from an actively growing, well-watered short grass surface.
ETP_c	Climatic Potential Evapotranspiration	Refers to the results of formulae or models used to estimate the above parameters.
ETP_o	Limiting Regional Potential Evaporation	An evaporation rate related only to available energy (generally equated to the first term of the Penman equation).
E_o	Point Evaporation	The evaporation rate as measured with evaporation pans, Piche evaporimeters, atmometers, etc.

Perrier (1977) suggested that the parameters in Table 2.1 could be related in the following general manner:

$$EP_{*} \geq EP \geq ETP_o \quad \text{and,} \quad EP > ETR_m \geq ETR, \qquad 2.2$$

where ETR is the actual evapotranspiration rate.

2.2 Development of potential evaporation parameters

Although Perrier (1977) has included a calculated parameter, ETP_c, in his classification, it is limited to those formulae used to estimate the "measured" parameters which are commonly used. Other calculated potential evaporation parameters can be developed and classified; these need not be confined to representing physically-attainable situations.

The type of parameter being sought will depend on the end use, and this can influence the approach used in its development. If the main purpose of the potential evaporation parameter is to serve as an index of the actual evaporation from a natural nonsaturated surface, the major criteria governing its usefulness are: (i) that it represent a unique value for each set of conditions; (ii) that it be easily calculated from readily measured input data; or (iii) that it be easily measured; (iv) that the relationship between it and the actual evaporation can be established; or (v) that its introduction into the analysis facilitate the establishment of a relationship between other parameters from which the actual evaporation can be calculated.

Brutsaert (1982) indicated that the fact that potential evaporation is often calculated by means of meteorological data observed under non-potential conditions represents a source of ambiguity since this is not the same rate as that which would be observed if the surface had been saturated. When using Penman's equation some hydrologists (Kohler and Parmele, 1967; Morton, 1983) provide a correction for the net radiation to account for the difference between actual and potential conditions. However, if the Penman equation is to be used only as an index (potential evaporation parameter) of the actual evaporation from a nonsaturated surface, then this type of adjustment becomes unnecessary. It is not necessary that the index represent a real or attainable situation to be useful; it is more important that it be unique, definable, predictable and related to the actual conditions. If, on the other hand, the Penman equation is to be used to provide an estimate of the evaporation from an open body of water or from an irrigated field (ETP_c from Table 2.1), then some adjustments become necessary for the radiant transfer and the turbulent transfer coefficients, since what is being sought in this case is a model describing evaporation from such surfaces.

It is unlikely that a parameter can be found such that criteria (ii) and (iii) are both satisfied. For example, although evaporation pan measurements (which are easily and readily obtained) have proven useful as an evaporation index in many situations, the calculation of pan evaporation is not a simple task given the complex advection and heat transfer regimes involved. It follows as well that these pan measurements do not accurately reflect those calculated terms such as the Penman evaporation parameter.

The approach used to develop potential evaporation parameters involves considering a natural evaporating surface as a system to or from which both energy and mass (water vapour) can be transferred. The assumption is made, for the sake of convenience, that the simplest situation exists: the net supply of radiant energy is the dominant source of energy; the transfer of heat by conduction in the soil is negligible even when relatively large changes in surface temperature are encountered; the surface is of a sufficient size such that there are no significant inputs of energy by horizontal advection. The system then is one where the net energy supplied to the surface, Q, will be dissipated by the turbulent transfer of sensible, Q_h, and latent heat, Q_e, into the overlying air layer. These energy fluxes are governed by a turbulent transfer coefficient and by the gradients defined by the difference between the surface parameters (temperature, T_s; vapour pressure, e_s) and those measured in the overlying air layer (T_a, e_a).

To define a "potential evaporation" for this system one would have to specify those changes to the system, or to the state variables or parameters of the system, that would result in evaporation taking place at a maximum or "potential" rate.

Those parameters which can be controlled in this exercise are the net available energy, Q, and the atmospheric and the surface parameters. It is convenient to group the atmospheric variables (wind, temperature and humidity) into a single parameter. The drying power of the air, which takes into account the wind speed and the vapour pressure deficit of the air, is an appropriate parameter. The minimum condition for the definition of potential evaporation would be the specification that the surface be brought to saturation; the surface temperature is then the parameter controlling the vapour pressure at the surface. By specifying different conditions or combinations of conditions for these parameters (ie., by providing different definitions) different values would be obtained for the "potential" evaporation rate.

(i) Consider first the minimum condition alone. Suppose that the surface, for which the initial conditions are known, was to be supplied with water so as to become saturated. All the other parameters (Q, Q_h, Q_e, T_a, e_a, T_s, and e_s) could change and a new equilibrium state could be reached. Let the equilibrium evaporation rate in this case be known as EP1. (EP1 represents the situation where potential evaporation is simply defined as evaporation from a saturated surface.) The solution of this new equilibrium state is indeterminate since there are too many unknowns for the number of available equations. Thus the specification of a saturated surface alone is not sufficient to define a useful parameter which might be called potential evaporation.

(ii) If the net energy supply, Q, is held constant while the surface becomes saturated, the surface temperature and the temperature and vapour pressure of the air will change and a new equilibrium state will be established (T_{s2}, T_{a2} and e_{a2}). Let the evaporation rate, governed by the gradient $e_{s2}^*-e_{a2}$, in this case be denoted by EP2. This equilibrium condition cannot be solved since there too many unknowns for the available equations. The evaporation rate resulting from the specification of the available energy with a saturated surface has been used as a hydrological parameter. It has been referred to as the "wet-environment", "equilibrium" or the "advection-free potential evaporation". Because it cannot be calculated directly, EP2 has been empirically related to the radiant energy supply (Priestley and Taylor, 1972; Davies and Allen, 1973; Brutsaert and Stricker, 1979). In this respect it is also related to ETP_0 described by Perrier (1977).

(iii) If the net energy supply, Q, and the atmospheric parameters, T_a and e_a, are held constant while the surface becomes saturated, the surface temperature will change and a new equilibrium state will be established (T_{s3}). This equilibrium condition can be solved by using the vapour transfer and energy balance equations, as was done by Penman (1948). This evaporation rate, EP3, defined by the gradient $e_{s3}^*-e_a$, has most often been referred to as the "Penman potential evaporation".

(iv) If the energy supply, Q, and the surface temperature, T_s, are held constant and the surface saturated, the air temperature and vapour pressure will change and a new equilibrium state will be established. In this case again there is insufficient information to solve for the equilibrium state. This evaporation rate, EP4, has not been used as a hydrological parameter, probably because it is not practical to hold constant a parameter which is not usually measured while allowing those readily measured parameters to vary.

(v) If the atmospheric parameters, T_a and e_a, and the surface temperature, T_s, are held constant while the surface becomes saturated, the evaporation rate established, EP5, defined by the gradient $e_s^*-e_a$, can be calculated using the vapour transfer equation if the surface temperature is known. This parameter has also been used to define potential evaporation (ie. van Bavel, 1966). Although EP5 is a very simple parameter, the general lack of availability of surface temperature measurements has inhibited its use.

There are other possible combinations of conditions which could be imposed on the system and which could result in the definition of other parameters, but these are sufficient to demonstrate the multiplicity of definitions. Table 2.2 provides a summary of the parameters presented above, along with the corresponding "definitions" and how they may be estimated. The table also indicates which of these have been referred to as "potential evaporation".

Table 2.2 Summary of definitions of the potential evaporation parameters.

Name	Definition	Calculation	Common usage
EP1	Evaporation rate which would occur if the surface was brought to saturation.	Indeterminate	"definition" of potential evaporation
EP2	Evaporation rate which would occur if the surface was brought to saturation and the energy supply to the surface was held constant.	Indeterminate, Regression with solar radiation (Priestley and Taylor, 1972)	wet-environment or equilibrium evaporation. advection-free evaporation.
EP3	Evaporation rate which would occur if the surface was brought to saturation and the atmospheric parameters and the energy supply to the surface were held constant.	Energy balance, vapour transfer equations	Penman potential. wet-surface evaporation.
EP4	Evaporation rate which would occur if the surface was brought to saturation and the energy supply and the surface temperature were held constant.	Indeterminate	not used
EP5	Evaporation rate which would occur if the surface was brought to saturation and the atmospheric parameters and the surface temperature were held constant.	Vapour transfer equation (if surface temperature is known)	Potential (van Bavel, 1966)

With the exception of EP4, all of the above evaporation rates have from time to time been referred to as the "potential evaporation" rate. Of these, EP3, the Penman evaporation, has found the widest use, probably because of those which can be calculated it is the only one which can be obtained from readily available meteorological variables. Although the verbal definition associated with EP1 is frequently encountered in the literature, the parameter itself is not a practical one; whereas a wet surface will evaporate at the potential rate, the specification of a saturated surface alone is not sufficient to define the potential condition. The parameter EP5 is the only one which accounts directly for the effect of the surface conditions; however, since it requires knowledge of the surface temperature which is rarely measured, it has found less use.

Although the term wet-environment evaporation is most often associated with the parameter EP2, some authors (Fortin and Seguin, 1975; Granger, 1989) utilize the Penman equation, EP3, to describe it. Seguin (1975) analyzed Morton's verbal definition of potential evaporation, for which the effects of advection are significant, and he showed that as the area of the hypothetical evaporimeter was reduced and approached zero, the calculated value for this potential evaporation rate increased and approached that of EP5.

Of the parameters presented in Table 2.2, EP2, EP3 and EP5 appear to be the most useful, either because they are easily calculated (EP3) or because they represent useful limits for evaporation from a saturated surface (EP2, EP5). These different potential evaporation parameters are of course not equal; they are related in the following general manner:

$$EP5 \geq EP3 \geq EP2 \geq E \qquad 2.3$$

where E is the actual evaporation rate. Because they are different, it is important that they be recognized as such, and appropriate and consistent designations should be adopted for these.

The net energy available, Q, is the only variable governing the parameter EP2. The equilibrium condition described in this case cannot be solved; there are four unknowns (EP2, T_o, T_a, e_a) and only the energy balance and vapour transfer equations available. Priestley (1959) provides the basis for an approximation; under these ideal conditions the overlying air will tend toward saturation ($e_a \approx e_a^*$) and the Bowen ratio can be approximated by $\beta \approx \gamma/\Delta$. The resultant expression for EP2 then is identical to the first term of the Penman equation, that is: $EP2 \approx \Delta Q_n/(\Delta + \gamma)$. Slatyer and McIlroy (1961) and Brutsaert and Stricker (1979) discuss this term in the context of evaporation from large, homogeneous moist surfaces under advection-free conditions and suggest that it represents a lower limit to evaporation from wet surfaces; they refer to this term as "equilibrium evaporation".

EP3, given by the Penman equation, is the most frequently used potential evaporation parameter. It is governed by the energy supply and atmospheric considerations. EP3 represents the evaporation rate from a moist surface exposed to the existing available energy and atmospheric conditions. It might thus be more appropriate to refer to this parameter as the "Penman wet-surface evaporation", rather than as the "Penman potential evaporation" rate as is frequently done.

The parameter EP5 is the largest of the defined parameters; it is the only parameter which is governed by both atmospheric and surface conditions. These considerations are probably sufficient to justify calling this term the "potential evaporation", or the maximum potential evaporation. The fact that the surface temperature, which is not routinely measured, is required for its calculation does appear to detract from its usefulness as an index from which relationships can be developed for estimating the actual evaporation. However, the inclusion of surface conditions can be most useful in the development of relationships for estimating actual evaporation.

2.3 The relationship between actual and potential evaporation

It has been the normal practice, when a potential evaporation parameter was adopted or defined, to establish an empirical relationship between it and the actual evaporation rate for the given conditions. Such regressions are generally not single-valued, and require calibration for each different situation. The implied assumption in this approach is that the actual and potential rates are directly related, that is, an increase in the potential rate will result in an increase in the actual evaporation rate.

Bouchet (1963) showed that as a surface dried from initially saturated conditions (where the actual and potential evaporation rates are equal), the actual evaporation decreases while the potential increases. He postulated that the decrease in the actual evaporation rate is equal to the increase in the potential evaporation rate. This relationship between actual and potential evaporation which Bouchet (1963) introduced is an inverse one. It has become known as the complementary relationship.

The difference between the inverse relationship introduced by Bouchet and the "traditional" direct relationship approach, and the validity of each, has been the subject of much debate; the natural tendency is, of course, to believe only one side or the other. As will be shown, both approaches are to a certain degree correct, but neither defines the relationship completely.

It can be shown that the parameters EP2, EP3 and EP5 behave in a complementary-like manner; the difference between the respective evaporation rates is greatest for dry conditions and they are of course equal for initially wet conditions. Fig. 2.1 is a schematic representation of an evaporating surface depicting the vapour pressure gradients

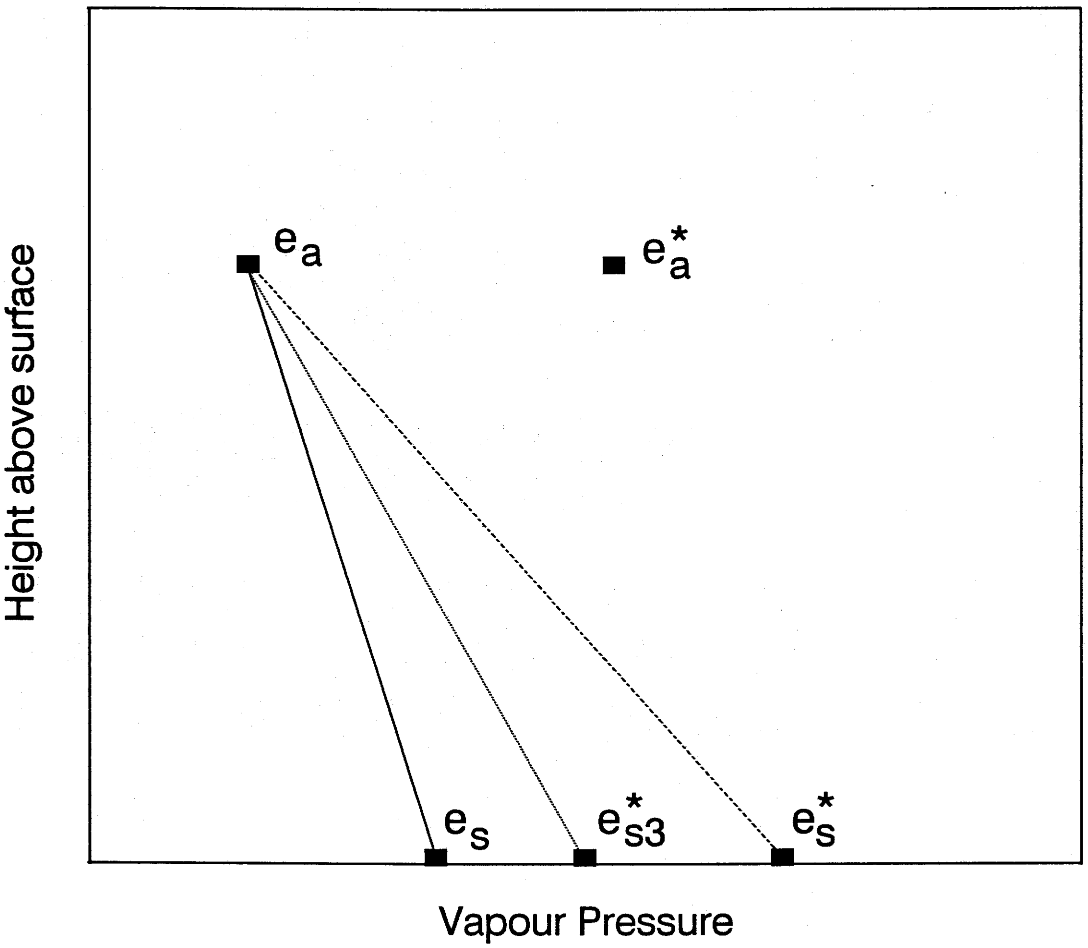

Figure 2.1 Schematic representation of the vapour pressure gradient above a nonsaturated surface and the gradients defining the parameters EP3 and EP5.

governing the evaporation rates E, EP3 and EP5. For a given set of conditions (net available energy, Q; drying power, E_a, defined by the vapour pressure deficit of the air; surface temperature, T_s; and surface vapour pressure, e_s), the actual evaporation, E, is governed by the vapour pressure gradient (e_s - e_a), EP5 is defined by the gradient (e_s^* - e_a), where e_s^* is the saturation vapour pressure at the temperature T_s. If the conditions of Q and E_a are maintained while the surface is supplied with water (the conditions defining the parameter EP3), the evaporation rate would increase and the surface temperature would decrease to some value T_{s3}, with a saturation vapour pressure e_{s3}^*. The wet-surface evaporation, EP3, would then be defined by the gradient (e_{s3}^* - e_a), and would be somewhat less than EP5. The evaporation rates E, EP3 and EP5 thus behave in a complementary-like manner; however, a truly complementary relationship as described by Bouchet (1963), where the differences between respective evaporation rates are equal, does not necessarily follow.

The situation presented in Fig. 2.1 is very much a hypothetical one, as was that described by Bouchet (1963). With the available energy held constant, the relationship between actual and "potential" evaporation has been demonstrated to be an inverse one. However, the available energy, consisting for the most part of the net radiation, is not constant; it follows pronounced diurnal and seasonal cycles. The behaviour of the actual and "potential" evaporation rates in this case can be demonstrated by using the Penman evaporation, EP3, as an example. Penman's expression is governed by the available energy, Q, and the drying power, E_a; an increase in either of these terms results in an increase in the Penman evaporation. The responses of the actual (E) and Penman (EP3) evaporation rates to changes in Q and E_a are demonstrated schematically in Fig. 2.2. The actual evaporation increases with increases in available energy. It is well known that the diurnal cycle of evaporation follows closely that of the net radiation (cf. Figs. 4.2 and 4.3). Thus both the actual and the "potential" evaporation rates respond directly to the changes in available energy. However, for non-advective conditions, the vapour pressure deficit of the air is a reflection of the partitioning of energy at the surface. Decreased evaporation results in a decreased vapour pressure in the air, which in turn increases the drying power; thus the actual and "potential" rates respond in opposite directions to changes in the drying power. This response will depend on the partitioning of energy at the surface and the available moisture.

Fig. 2.2 and the above discussion demonstrate why it has not been possible to develop universally-applicable relationships which provide a reliable estimate of the actual evaporation rate from the calculation of a single "potential" parameter; the two terms of the Penman wet-surface evaporation

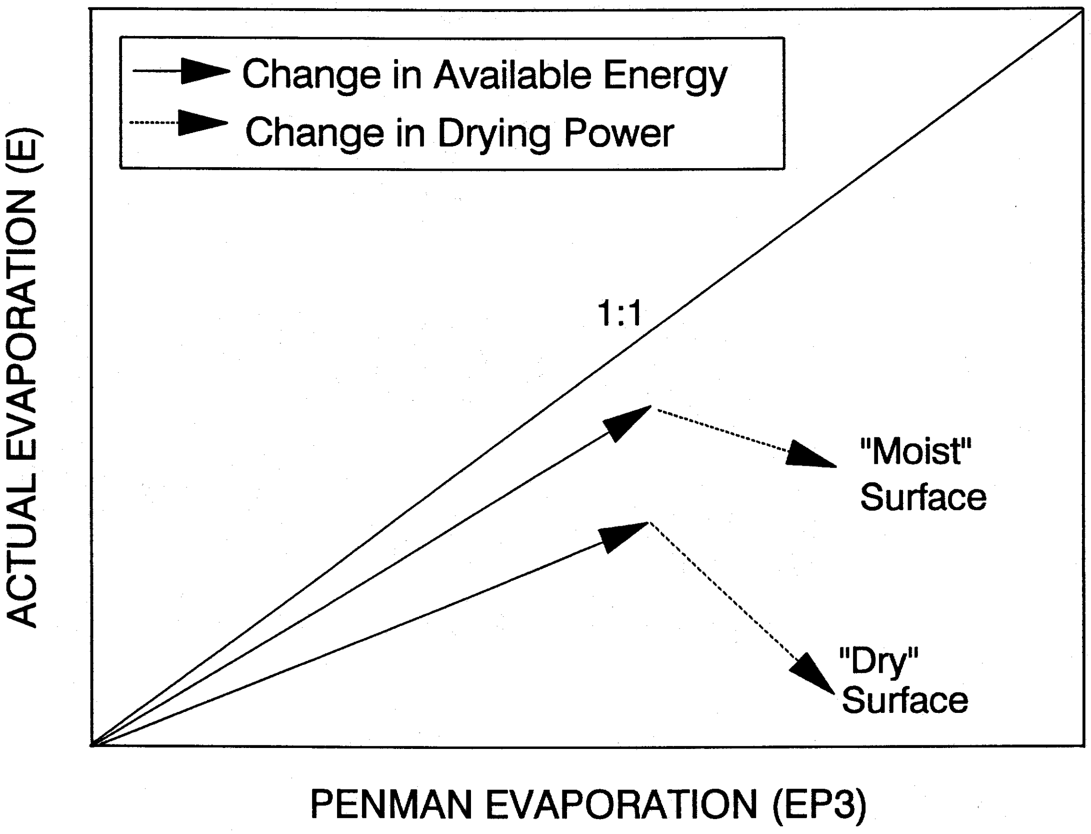

Figure 2.2 Schematic representation of the responses of the actual (E) and Penman (EP3) evaporation rates to changes in the available energy (Q_n) and the drying power (E_a).

rate respond differently to changes in the actual evaporation rate. The use of a second potential evaporation parameter, as in the complementary approach, makes it possible to account for both the direct and inverse responses. Those complementary relationship evaporation models developed to date (Brutsaert and Stricker, 1979; Morton, 1983) have made use of the parameters EP2 and EP3. This combination works because EP2, which is a function only of the net radiation, defines the contribution of the available energy, and the relationship then focuses essentially on the inverse relationship between the drying power component of EP3 and the actual evaporation rate. However, because EP2 is generally estimated through regression, the exact form of the relationship between E, EP2 and EP3 has not been derived; the above modellers have assumed the relationship to be truly complementary.

The inequality presented in Eq. 2.3 can form the basis for an evaporation model if the relationship between the actual evaporation, E, and one or more of the other evaporation parameters can be established. EP3 and EP5 can both be expressed in terms of gradients defined by the actual conditions of temperature and humidity; it should thus be possible to derive the form of the relationship between these and the actual evaporation rate. As will be shown in the following chapter, EP3 and EP5 are also suited to form the basis of a complementary-type of evaporation model.

3. DEVELOPMENT OF APPROPRIATE RELATIONSHIPS

A nonsaturated surface is characterized by the fact that the vapour pressure at the surface, e_s, is less than the saturation vapour pressure at the surface temperature, e_s^*. This has traditionally been a major source of difficulty in obtaining useful relationships for evaporation from nonsaturated surfaces. For example, Penman (1948) limited his development to the saturated case for the simple reason that it represents a situation where the relationship between the vapour pressure and the temperature at the surface is known. He recognized that his was only a first step, and that further steps would be required to extend this to the nonsaturated case; his attempt at this second step was the introduction of resistance terms to moderate the evaporation (Penman and Schofield, 1951; Penman, 1956). The development of relationships to describe the resistance terms continues to be an active area of research.

The existence of feedback relationships between evaporation and the temperature and humidity of the overlying air has also been used as a basis for establishing relationships between evaporation from nonsaturated surfaces and one or more potential evaporation parameters. This approach was first used by Bouchet (1963) who demonstrated that the relationship between actual and potential evaporation was an inverse one. Brutsaert and Stricker (1979) and Morton

(1983), assuming that the actual and potential evaporation rates respond in a symmetrical, or truly complementary fashion, adapted this approach to the development of evaporation models.

Both the Penman approach and the Bouchet approach will be examined as possible bases for the development of a general form of equation to describe evaporation from nonsaturated surfaces.

3.1 Extension of the Penman approach

The energy balance in the vertical direction at a horizontal surface, assuming advection is negligible, can be written (with the terms expressed as an equivalent depth of evaporation) as:

$$E + H = Q \;, \qquad 3.1$$

where E is the evaporation, H is the sensible heat, and Q is the total energy available from net radiation and soil heat.

Introducing the Bowen ratio,

$$\beta = \frac{H}{E} = \frac{\gamma(T_s - T_a)}{(e_s - e_a)} \;, \qquad 3.2$$

and the slope of the saturation vapour pressure curve at the mean temperature, $\Delta = de^*/dT$, expressed in the form:

$$\Delta = \frac{(e_s^* - e_a^*)}{(T_s - T_a)} , \qquad 3.3$$

into Eq. 3.1 gives:

$$Q = E\left[1 + \frac{\gamma(e_s^* - e_a^*)}{\Delta(e_s - e_a)}\right] , \qquad 3.4$$

where: γ is the psychrometric constant; e_s^* and e_s are the saturation and actual vapour pressures at the evaporating surface at a temperature, T_s; and e_a^* and e_a are the saturation and actual vapour pressures of the air at a temperature, T_a.

Evaporation can also be expressed by the Dalton - type bulk transfer equation:

$$E = f(u)(e_s - e_a) \qquad 3.5$$

or:

$$e_s - e_a = \frac{E}{f(u)} , \qquad 3.6$$

where $f(u)$ is a wind speed function. Substituting equation (3.6) into (3.4) and rearranging terms yields:

$$E = Q + \frac{\gamma}{\Delta} f(u)(e_a^* - e_s^*) . \qquad 3.7$$

If the vapour pressure of the air, e_a, is added and subtracted to the difference in vapour pressure Eq. 3.7 becomes:

$$E = Q + \frac{\gamma}{\Delta}[f(u)(e_a^* - e_a) - f(u)(e_s^* - e_a)] . \qquad 3.8$$

The first term in the brackets in Eq. 3.8, $f(u)(e_a^* - e_a)$, the product of the wind function and the vapour pressure deficit of the air, is generally referred to as the "drying power" of the air, and is denoted by E_a. The second term in the brackets, $f(u)(e_s^* - e_a)$, is the evaporation rate which would occur under the same atmospheric conditions for wind and humidity, if the surface was saturated at the temperature of the surface. This term therefore represents a form of the potential evaporation, E_p (van Bavel, 1966), and is equivalent to the parameter EP5 defined in Table 2.2. Eq. 3.8 is a general equation relating evaporation to the net available energy, the drying power of the air and this potential evaporation. It can be used to calculate E when an adequate expression for E_p is available. The above formulation for E_p although quite simple, has the major drawback of requiring the surface temperature, a parameter which is rarely measured.

For a nonsaturated surface, for which evaporation occurs at some rate less than the potential, ($0 < E < E_p$), a relative evaporation can be defined as the ratio of actual to potential evaporation, $G = E/E_p$. The relative evaporation, G, should be an unique parameter for each set of atmospheric and surface conditions. Using Eq. 3.5 and the equivalent expression for E_p, G can be expressed as :

$$G = \frac{E}{E_p} = \frac{f(u)\,(e_s - e_a)}{f(u)\,(e_s^* - e_a)} \quad . \qquad 3.9$$

Eq. 3.9 shows that the relative evaporation is governed by the availability of water at the surface as expressed by the vapour pressure deficit, $e_s^* - e_s$. In other words surface conditions play a predominant role in the partitioning of the available energy, and thus also control the potential evaporation, as defined in this case. For a wet surface, where $e_s = e_s^*$, G will be equal to unity; for a very dry surface, e_s will tend toward e_a, and G will approach zero. Thus, excluding the possibility of condensation, the relative evaporation is confined to the limits $0 \leq G \leq 1$. Substituting $E/G = E_p = f(u)(e_s^* - e_a)$ into Eq. 3.8 yields:

$$E = \frac{\Delta G Q}{(\Delta G + \gamma)} + \frac{\gamma G E_a}{(\Delta G + \gamma)} \quad . \qquad 3.10$$

Eq. 3.10 is a general equation describing evaporation from nonsaturated surfaces. It can be used to estimate the evaporation if a suitable expression for the relative evaporation, G, is available. Eq. 3.10 is similar in form to the Penman equation, but differs through the inclusion of the relative evaporation, G, which accounts for departures from saturated conditions. It is interesting to note that Barton (1979), in a parallel development, introduced the relative humidity at the soil surface (rather than the relative evaporation) and produced an equation somewhat different from Eq. 3.10 to account for nonsaturated conditions. He used his expression to attempt to extend the Priestley and Taylor (1972) hypothesis to include nonsaturated surfaces.

3.2 The relationship of Bouchet

Bouchet (1963), utilizing an analysis based on the energy balance, demonstrated that as a surface dried from initially moist conditions the potential evaporation increased while the actual evaporation was decreasing; he thus corrected the misconception that a larger potential evaporation necessarily signified a larger actual evaporation. The relation which he derived (Eq. 1.8), which has come to be known as the complementary relationship between actual and potential evaporation, states that as the surface dries the decrease in actual evaporation is accompanied by an equal, but opposite, change in the potential evaporation; the potential evaporation thus ranges from its value at saturation to twice this value. Bouchet cautioned that this relationship was an approximate one. Nonetheless, the complementary relationship has formed the basis for the development of some evaporation models (Brutsaert and Stricker, 1979 ; Morton, 1983). Its behaviour for different scales of space and time has been analyzed (Seguin, 1975; Fortin and Seguin, 1975).

The complementary relationship (Eq. 1.8) makes use of two potential evaporation parameters, a "potential evaporation", E_p, and a second parameter, E_{pw}, which Bouchet defined as the value of the potential evaporation when the actual regional evaporation rate is equal to the potential rate. The major benefit accruing from the use of two potential evaporation parameters, rather than a single one, is that the resulting relationship accounts for both the direct and inverse responses between the "potential" and actual rates, and appears to be universally applicable. On the other hand, it is now dependant on the proper selection, interpretation and application of two parameters, rather than one. This should not normally represent a drawback; however, as was shown in the previous chapter, the concept of potential evaporation has been the source of some ambiguity. The choice of definitions adopted by those who have developed complementary relationship evaporation models (Morton, 1983; Brutsaert and Stricker, 1979) has not permitted them to examine or derive the exact form of the relationship, and has thus forced them to assume that it is truly complementary, or that $dE/dE_p = -1$. The following is an analysis of the Bouchet relationship, using a set of definitions for the potential and wet-surface evaporation rates which allow for the derivation of the general form of the relationship.

3.2.1 The relationship between E, E_p, and E_{pw}

Bouchet (1963) postulated that as a wet surface dried the decrease in actual evaporation was matched by an equivalent increase in potential evaporation, ie. $\delta E = -\delta E_p$. This resulted in the relationship:

$$E + E_p = 2E_{pw} , \qquad 3.11$$

in which Bouchet defines E_{pw} as the evaporation rate which would occur under conditions where $E = E_p$. In deriving his relationship Bouchet did not use known or derived expressions for E_p or E_{pw}; he simply based his analysis on the assumption that as E was reduced the energy liberated must necessarily go to increasing the "potential", and that the changes in the actual and potential rates would be similar in magnitude. The inequality (Eq. 2.2) presented above provides a framework which supports the concept introduced by Bouchet (1963); however, it does not necessarily support the assumption that $\delta E = -\delta E_p$.

Those complementary relationship models developed to date have utilized the parameters EP3 and EP2 to represent potential and wet-environment evaporation, respectively. Because EP2 cannot be derived directly from energy balance and mass transfer equations, the form of the relationship between E, EP3 and EP2 was not derived, but was assumed to be truly complementary. The parameter EP2 has been obtained through regression. Brutsaert and Stricker (1979) utilize a regression similar to that presented by Priestley and Taylor (1972). Morton (1983) establishes a regression in such a way as to make it equal to one half of the potential which he adopts for a "dry" environment; by doing so he is in fact redefining the parameter and ensuring that the relationship will truly be a complementary one.

However, since all three potential evaporation parameters (EP2, EP3, and EP5) behave in a complementary-like manner, it should be possible to develop a "complementary relationship model" using any two of these. In addition, if the two parameters selected can be derived from energy balance or mass transfer equations and expressed in terms of appropriate vapour pressure gradients, it should also be possible to derive the general form of the relationship between the actual evaporation and the two potential evaporation parameters. EP3 and EP5 can both be calculated directly; EP3 is represented by the Penman combination equation, and EP5 is derived directly from the mass transfer equation. There is some justification and precedence for using the parameters EP5 and EP3 to represent the potential evaporation (E_p) and the wet-surface evaporation (E_{pw}), respectively; EP5 is the largest of the potential evaporation parameters, and van Bavel (1966) adopts this form of expression for potential evaporation, as well, Fortin and Seguin (1975) suggest that the Penman equation, which is independent of surface values of temperature and humidity, represents the wet-surface evaporation rather than the potential evaporation.

Fig. 2.1, a schematic representation of an evaporating surface, illustrates that for a given set of conditions [net available energy, Q; drying power, E_a, defined by the vapour pressure deficit of the air; surface temperature, T_s; and the vapour pressure at the surface, e_s], actual evaporation (E) is governed by the vapour pressure gradient ($e_s - e_a$), and the potential (E_p) is defined by the gradient ($e_s^* - e_a$), where e_s^*

is the saturation vapour pressure at the surface temperature T_s. If the conditions of Q and E_a were maintained while the surface was supplied with water, the evaporation rate would increase and the surface temperature would decrease to some value T_{s3}, with a saturation vapour pressure $e_s^*{}_3$. The wet-surface evaporation, E_{pw}, would be defined by the gradient ($e_s^*{}_3$ - e_a), and would be somewhat less than the potential evaporation, E_p. As the actual (initial) surface conditions approached saturation both E and E_p would approach the wet-surface value, E_{pw}. Using a Dalton-type formulation, these rates can be written as follows:

Actual Evaporation $$E = f(u)\,(e_s - e_a) \qquad 3.12a$$

Drying Power $$E_a = f(u)\,(e_a^* - e_a) \qquad 3.12b$$

Wet-surface Evaporation $$E_{pw} = f(u)\,(e_{s3}^* - e_a) \qquad 3.12c$$

Potential Evaporation $$E_p = f(u)\,(e_s^* - e_a) \qquad 3.12d$$

For a given set of conditions (Q and E_a) the wet-surface evaporation rate (E_{pw}) will be fixed; the potential evaporation rate (E_p) is then governed by the surface temperature, and the actual evaporation (E) by the actual vapour pressure at the surface.

Referring to Fig. 2.1, and using Eqs. 3.12a, c and d to describe E, E_{pw} and E_p, respectively, we can parallel Bouchet's development and write:

$$\begin{aligned}\delta E = E_{pw} - E &= f(u)\,(e^*_{s3} - e_a) - f(u)\,(e_s - e_a) \\ &= f(u)\,(e^*_{s3} - e_s)\ ,\end{aligned} \qquad 3.13$$

and

$$\begin{aligned}\delta E_p = E_p - E_{pw} &= f(u)\,(e^*_s - e_a) - f(u)\,(e^*_{s3} - e_a) \\ &= f(u)\,(e^*_s - e^*_{s3})\ .\end{aligned} \qquad 3.14$$

Eqs. 3.13 and 3.14 can be combined in the following manner:

$$E_{pw} - E = (E_p - E_{pw})\,\frac{(e^*_{s3} - e_s)}{(e^*_s - e^*_{s3})}\ . \qquad 3.15$$

The slope of the saturation vapour pressure curve at the surface temperature, T_s, can be approximated by

$$\Delta = \frac{(e^*_s - e^*_{s3})}{(T_s - T_{s3})}\ .$$

For the situation presented in Fig. 2.1, where the available energy and the atmospheric conditions are held constant, only the latent and sensible heat fluxes are allowed to change as the evaporation rate changes from E to E_{pw}. The integrity of the energy balance then requires that the change in latent heat flux result in an equal but opposite change in the sensible heat flux. It is assumed that a change in surface temperature will result in a change in soil heat flux which is small compared to the resultant change in turbulent sensible heat flux. The Bowen ratio relating these changes in latent and sensible heat fluxes can be expressed as:

$$\beta' = -1 = \gamma \frac{(T_{s3} - T_s)}{(e^*_{s3} - e_s)} ,$$

where γ is the psychrometric constant. Eq. 3.15 can then be simplified to:

$$E_{pw} - E = (E_p - E_{pw}) \frac{\gamma}{\Delta} ,$$

or

$$E + E_p \frac{\gamma}{\Delta} = E_{pw}\left(1 + \frac{\gamma}{\Delta}\right) , \qquad 3.16$$

which replaces Bouchet's relationship (Eq. 3.11). Eq. 3.16 holds for a wet surface where $E = E_p = E_{pw}$, and reduces to Bouchet's relationship only when $\Delta = \gamma$, a condition which, for standard atmospheric pressure, occurs when the temperature is near 6°C. Eq. 3.16 shows that for constant available energy and atmospheric conditions (ie. constant E_{pw}), $\partial E/\partial E_p = -\gamma/\Delta$, and that this "complementary" relationship is not normally symmetric as Bouchet had assumed.

This result does not necessarily void all the other developments which have been advanced relative to Bouchet's relationship. The exact form of the complementary relationship will depend on the selection of the potential evaporation parameters. Morton (1983), for example, uses a variation of the Penman method to calculate E_p, and actually redefines E_{pw} in such a way as to ensure that Bouchet's relationship (Eq. 3.11) holds.

Eq. 3.16 can be used to provide estimates of actual evaporation if appropriate expressions are available for estimating E_{pw} and E_p. E_{pw} is given by the Penman combination equation; however, E_p requires that the surface temperature be known. Substituting the Penman equation into Eq. 3.16, and introducing the concept of relative evaporation ($G = E/E_p$), the ratio of actual to potential evaporation, results in the following general equation for evaporation:

$$E = \frac{\Delta G Q}{(\Delta G + \gamma)} + \frac{\gamma G E_a}{(\Delta G + \gamma)} \ .$$

This is identical to the general equation (Eq. 3.10) describing evaporation from nonsaturated surfaces which was derived using an approach similar to that of Penman (1948). Thus both the combination approach (paralleling Penman's development) and the complementary approach (Bouchet's development) yield the same general evaporation equation. This, however, should be expected since the definitions used for potential and wet-surface evaporation were consistent.

3.3 Relative evaporation

To determine evaporation using the general equation (Eq. 3.10) requires an estimate of the relative evaporation, G. An appropriate relationship for G would be one which relies on readily available data and which maintains the inherent advantage of the Penman approach, that is, one which does not require measurements of the surface conditions of temperature and humidity.

Several investigators, for example Priestly and Taylor (1972), Black (1979) and Federer (1979), have related parameters similar to G to soil water deficits or to available soil water. The major disadvantages of such relationships are they are empirical, depend on the specific conditions encountered, and are generally not transferable. Thus, the practical application of the method would be greatly enhanced if G could be related in a universal manner to readily-measured or calculated parameters. These would exclude the surface parameters, such as temperature and vapour pressure, which are rarely measured but which govern both actual and potential evaporation. Morton (1983) points out that the effects of changes in the availability of soil water on potential evaporation can be assessed by their effects on the temperature and humidity gradients. Since an increase in actual evaporation causes the vapour pressure of the overlying air to increase, the drying power, E_a, reflects to some extent the "dryness" of the surface. Further, the sum of the drying power, E_a, and the available energy, Q, should index the "potential" <u>for</u> evaporation. Following this logic, it is reasonable to assume that a relationship can be established between G and the ratio $D = E_a/E_a+Q$, which can be referred to as the relative drying power. The relationship, which is also dimensionless, is an inverse one; for a dry surface $G = 0$, E_a is large and D approaches unity. For a wet surface $G = 1$, E_a

and D approach zero.

If actual evaporation data are available it is possible to obtain estimates of the relative evaporation. Rearranging the general equation (Eq. 3.10) and simplifying gives an expression for G equal to:

$$G = \frac{\gamma E_a}{\Delta Q + \gamma E_a - \Delta E} \; . \qquad 3.17$$

An investigation of the relationship between the dimensionless parameters, G and D, was undertaken using field data monitored at two stations located in the semi-arid climatic zone of western Canada. Both sites, Bad Lake and Saskatoon, include silty clay and heavy clay soils which form part of the Brown soil zone in the Province of Saskatchewan. Relative evaporation, G, was evaluated by Eq. 3.17 from estimates of E, E_a, Q, γ and Δ calculated using hydrological and meteorological observations made at the stations. A brief review of the procedures used for the calculations is given below.

The data were separated and grouped into two periods, postmelt - the period extending from the end of snowcover ablation to the day of seeding of an annual crop and the crop growing season. As well the data were stratified according to landuse and vegetative cover, that is as fallow, stubble, growing crop (wheat) and grass. Stratification was necessary for several reasons: (a) the ground heat flux, surface albedo, and soil moisture and temperature regimes and other factors affecting the net energy available for evaporation (Q) differ widely between postmelt and growing-season periods and (b) the calculation of the drying power (E_a) includes a wind function (f(u)) that is affected by surface roughness, which in turn depends on the type, density and height of vegetation.

Evaporation was calculated as the residual term in a soil water balance applied at field sites where precipitation and soil moisture were measured. Precipitation estimates used in the budget calculations were daily values collected by a standard Canadian rain gauge. Profiles of changes in soil moisture were obtained with a two-probe gamma, density meter. Complete details on the determination of soil moisture by this method are well-documented in the literature (for example see Troxler, undated; Smith et al., 1967; Reginato and Jackson, 1971; and Jame and Norum, 1980). In principle the method measures soil wet density and a change in density in a measurement interval is assumed to be due to a soil moisture change. Profiles of soil moisture changes were obtained at 20 mm increments of depth to 1 m and 40 mm increments between 1 m and 1.6 m. Details of the measurement procedures used are described by Granger et al. (1984) and Gray et al. (1985). Repeatability tests conducted with their equipment in the field gave a standard error of estimate in soil moisture content of 2.5 mm in a 1-m profile.

Although radiation measurements were available at both sites, these were plagued by frequent intervals of missing data. Hence, they were not used directly for calculating the energy flux Q, except as a check of values of net radiation estimated by the model developed by Gray and Landine (1988). Soil heat flux was assumed to be negligible and omitted from the calculation of Q except for those periods during postmelt when a soil was thawing. It was observed that during the "thaw" period a significant fraction of the net radiation is apportioned to soil heat. Soil temperature measurements taken daily at the sites themselves allowed for estimation of this term assuming a simple, steady-state conduction model for which the thermal conductivity was estimated using the method described by de Vries (1963).

Since a variety of surface conditions were encountered, use of a single wind function such as that proposed by Penman (1948) was inappropriate. The function was estimated from boundary layer and surface roughness considerations based on the well-known aerodynamic formula for evaporation:

$$E = \frac{0.622 k^2 \rho u_1 (e_s - e_a)}{P \ln\left[\frac{z_1 - d_0}{z_0}\right] \ln\left[\frac{z_2 - d_0}{z_0}\right]}, \qquad 3.18$$

in which E is the evaporation rate (mm/s); k = 0.4 is the von Karman constant; ρ is the air density (kg/m^3); P is the atmospheric pressure (Pa); u is wind speed (m/s) and e is vapour pressure (Pa). The subscripts 1 and 2 refer to the measurement heights z_1 and z_2. d_0 is the zero-plane displacement of the vegetative surface (m), and z_0 is the roughness height (m). According to Eq. 3.18, the wind function of the Dalton-type equation can be taken equal to:

$$f(u) = \frac{0.622k^2\rho u_1}{P\ln\left[\frac{z_1 - d_0}{z_0}\right]\ln\left[\frac{z_2 - d_0}{z_0}\right]} \quad . \qquad 3.19$$

In addition to atmospheric measurements, Eq. 3.19 requires information on the roughness elements of the ground surface. The displacement and roughness heights were estimated from measurements of crop height in the following expressions (Brutsaert, 1982):

$$d_0 = 0.67h \qquad 3.20$$

and:

$$z_0 = h/7.35 \; , \qquad 3.21$$

in which h is the mean height of the vegetation. Values of d_0 and z_0 assigned to fallow were 0 and 0.006 m, respectively, and to stubble (postmelt) and grass surfaces 0 and 0.016 m.

Values for E, Q and E_a estimated by these procedures were used in Eq 3.17 to calculate G for 158 periods, varying from 2 to 30 days, for which soil moisture changes had been monitored. The mean values of atmospheric and soil parameters observed during the measurement interval were used. Relative evaporation is plotted against relative drying power for each individual period in Fig. 3.1. The data for the different landuses have been pooled because the differences between the

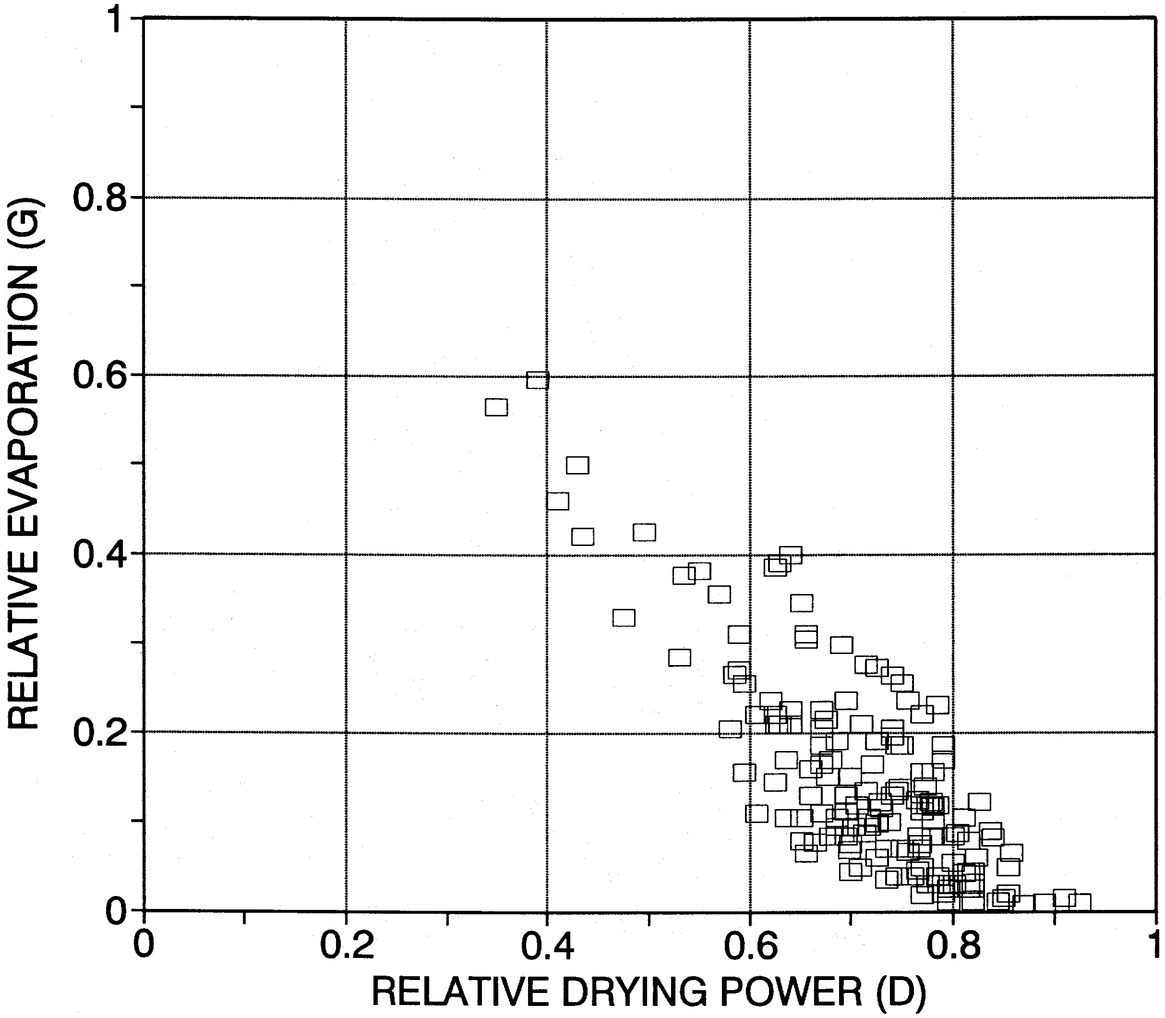

Figure 3.1 Plot of relative evaporation against relative drying power showing all data points.

means of G for the different landuse groups are small and the scatter in G-values within groups is of similar magnitude. In fact the relationship between G and D, which are both dimensionless, should be independent of landuse provided the respective calculations properly account for the effects of surface condition on the energy and mass exchange processes.

Fig. 3.1 shows a trend for G to decrease nonlinearly with increasing D, however the shape of the decrease is masked by the scatter of the data. A clearer picture of this trend is shown in Fig. 3.2 in which the mean values of G for different class intervals of D are plotted. It appears that the relationship is S-shaped, a shape similar to the curve of the ratio of actual to potential evaporation versus soil moisture content. An exponential function fitted to the data gave the expression:

$$G = \frac{1}{1 + 0.028e^{8.045D}} , \qquad 3.22$$

with a standard error of the estimate of G of 0.051. Eq. 3.22 can be used with Eq. 3.10 to calculate evaporation from nonsaturated surfaces. The equations are independent of surface parameters (temperature and vapour pressure) and they do not require an estimate of potential evaporation. Unfortunately, the lack of measurements from wet environments (G > 0.7) does not allow the development of a functional relation between G and D that can be treated with confidence over the entire range in G. In fact Fig. 3.2 suggests that

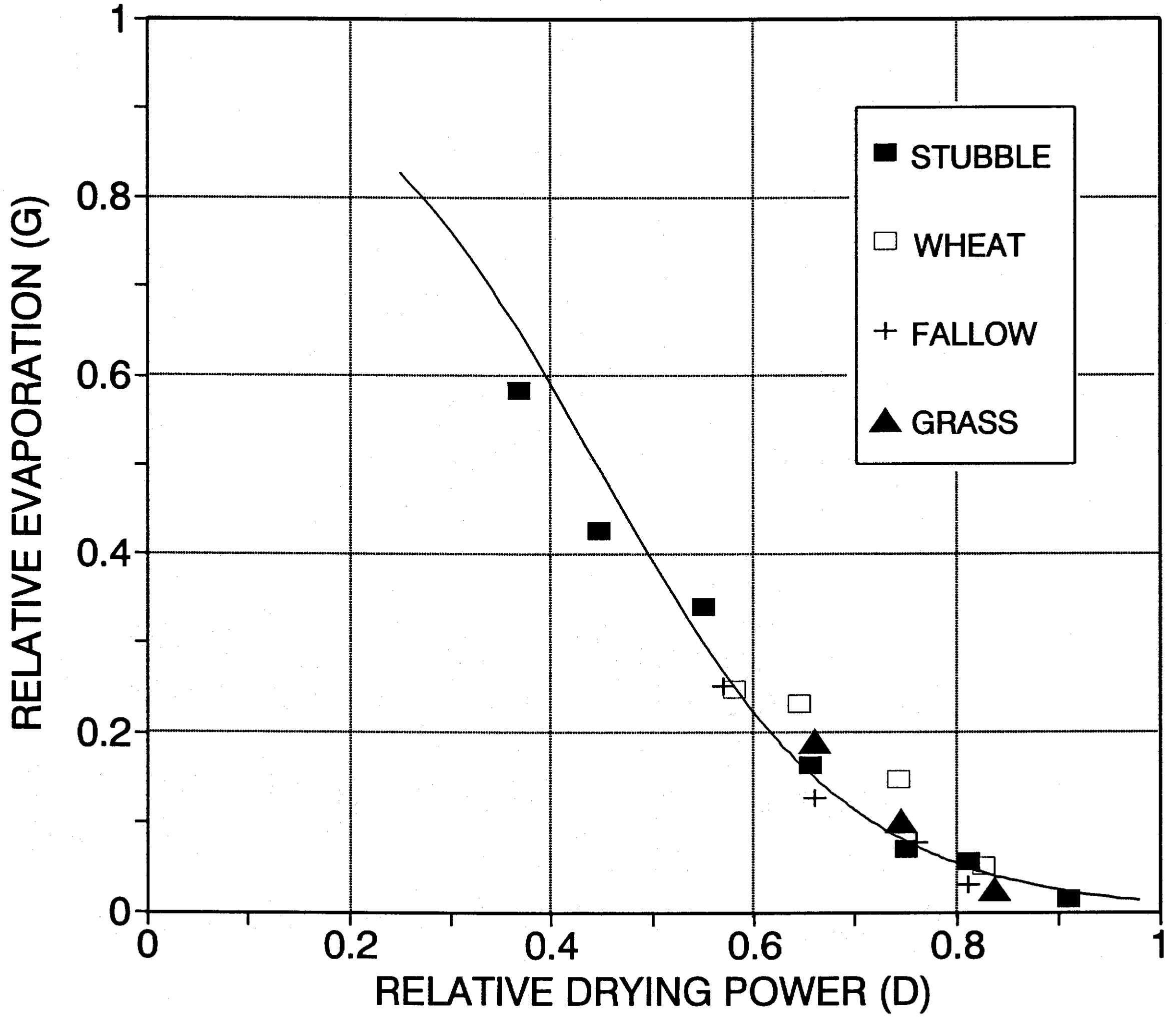

Figure 3.2 Mean values of relative evaporation plotted against the mean values of different class intervals of relative drying power for various landuses.

the derived relationship may also slightly overestimate the values of G for values of D near 0.4. Additional data, in particular in the wetter region, are required to confirm or refine the relationship.

It is also possible to relate the relative evaporation, G, to the aerodynamic and bulk stomatal resistance terms (or soil water resistance in the case of bare soil) used in the Penman-Monteith approach to extending the Penman equation to the nonsaturated case. These resistances are defined by the expressions:

$$E = \frac{\rho C_p}{\gamma L_e}\frac{(e_s - e_a)}{r_a} = \frac{\rho C_p}{\gamma L_e}\frac{(e_s^* - e_s)}{r_s} \ , \qquad 3.23$$

where r_a and r_s are the aerodynamic and stomatal resistances, respectively, and C_p is the specific heat of air at constant pressure. From these expressions and the definition of the relative evaporation (Eq. 3.9) it can be shown that

$$G = \frac{e_s - e_a}{e_s^* - e_a} = \frac{r_a}{r_a + r_s} \ . \qquad 3.24$$

In other words, the relative evaporation is equivalent to the ratio of the aerodynamic resistance (the only resistance term involved when the surface is saturated) to the sum of the aerodynamic and stomatal resistances encountered under nonsaturated conditions. The aerodynamic resistance can be determined from formulations such as Eq. 3.19. However, the interaction of the factors which govern the stomatal resistance (the distribution of radiation absorption, biological activity, moisture stress in the soil, and the physiological characteristics of each vegetation species) are very complex and difficult to parameterize. Attempts have been made to relate this resistance term to other parameters; for example, Monteith (1965) relates r_s to the Bowen ratio, van Bavel (1967) and Tan and Black (1976) relate it to the soil water stress, and Federer (1979) to the plant water potential. However, although one can rationalize causal relationships between the stomatal resistance and those factors mentioned, relationships of this type are generally not sufficiently transferable to be practical.

On the other hand, there are generally feedback mechanisms involved which tend to define the equilibrium state. The examination of the feedback effects, rather than the causal relationships, may in fact provide a more reliable approach to the development of a "general" relationship. That this is probable has been demonstrated through the development of the relationship between relative evaporation and relative drying power. However, further experimental verification would be required to confirm the validity and general applicability of the relationship.

4. EXPERIMENTAL VERIFICATION

A field experiment was established at the University of Saskatchewan Crop Development Farm to:

i) collect reference evaporation data against which to compare the results of calculations obtained using the relationship between relative evaporation and relative drying power in conjunction with the general equation for evaporation from nonsaturated surfaces;

ii) derive an appropriate relationship for the vapour transfer function for periods of one day or longer;

iii) refine the relationship between relative evaporation and relative drying power;

iv) derive a relationship for the soil heat flux;

v) evaluate the net radiation model; and

vi) assess the possibility of applying remotely-sensed surface temperature data for providing estimates of regional evaporation.

4.1 Experimental procedures

The experimental installation was designed to provide estimates of evaporation using the energy balance and Bowen ratio approaches. Table 4.1 provides a description of the essential instrumentation involved, and Fig. 4.1 is a schematic representation of the deployment of the sensors.

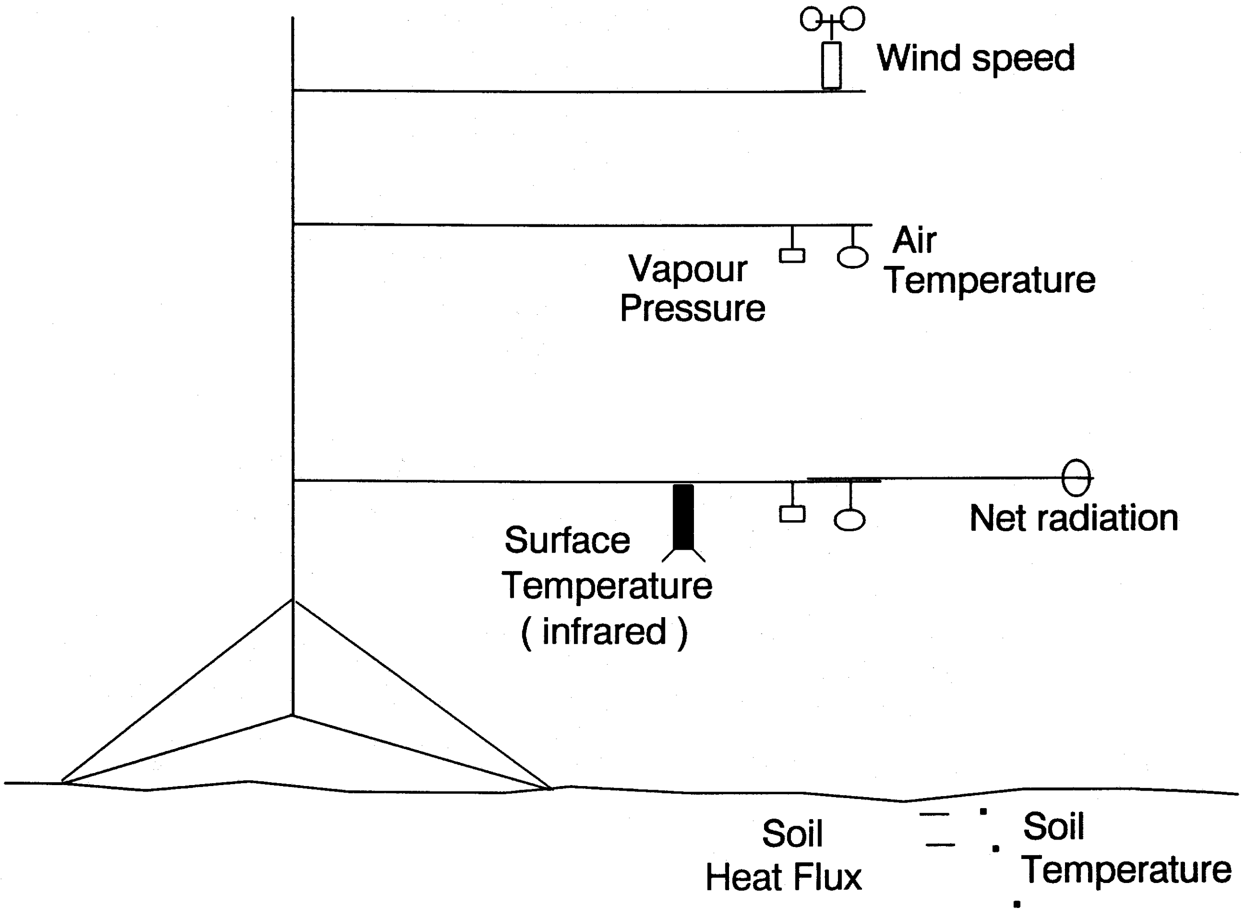

Figure 4.1 Schematic representation of the instrument mast showing the deployment of the principal sensors.

Table 4.1 Description of the instruments deployed for the measurement of evaporation/evapotranspiration.

ELEMENT	SENSORS	MANUFACTURER	HT or DEPTH	COMMENTS
Net Radiation	1 Net Pyrradiometer	Middleton Co.	1 m	calibrated March 1990
Soil Heat Flux	2 Heat Flux Transducers	Radiation and Energy Balance Systems, inc.	0.02 and 0.08 m	calibration provided by manufacturer
Soil Temperature	Averaging Soil Thermocouple probe	Campbell Scientific Co.	0.02 to 0.06 m	
Air Temperature	Type-E fine-wire Thermocouple	Campbell Scientific Co.	0.8 and 1.5 m	wired for differential measurement
Humidity	Chilled Mirror Hygrometer	General Eastern	0.8 and 1.5 m	single sensor for both heights
Wind Speed	Micro-response Cup Anemometer Model 2031	WeatherMeasure Weathertronics	2 m	NBS calibration
Surface Temperature	Infrared Thermometer	Everest Interscience, inc.	1 m	60° field of view

Data were logged using a Campbell Scientific 21X data logger. Air temperature and humidity channels were sampled every second, while all other parameters were sampled at 10-second intervals. All data were averaged and stored as the means for 30-minute periods. Other measurements obtained at the site included precipitation, soil moisture, canopy temperature and crop height and development. As well, measurements of sunshine duration and insolation were available from the Saskatchewan Research Council meteorological observation station located approximately 2 km from the site.

Three observation periods were used during 1989 and 1990, providing a total of 141 days of observation. In the summer of 1989 measurements were taken over a crop of growing wheat for a period of 23 days (July 9 to July 31). The period includes the rapid growth stages and ripening of the crop. During this observation period infrared surface temperature measurements were not available. In 1990 experiments were conducted during two periods. The first was over fallow in the period following the disappearance of the seasonal snow cover (April 27 to May 22). The soil had already thawed to depth, and the field remained bare throughout the observation period. The equipment was removed to allow for the seeding of the annual wheat crop; it was then replaced and measurements were taken throughout the growing season (May 29 to August 29).

4.1.1 Calculation of the latent heat flux

The instrumentation used allowed for the calculation of the latent heat flux with three different methods: (i) the Bowen ratio, (ii) an aerodynamic formula using the gradient of humidity and the wind speed, and (iii) the energy balance.

The Bowen ratio

The Bowen ratio approach (Bowen, 1926) represents a simple manipulation of the energy balance equation:

$$Q_n = Q_g + L_e E + H \, , \qquad 4.1$$

where Q_n is the net radiation, Q_g is the soil heat flux (taken positive into the soil), H is the turbulent transfer of sensible heat, L_e is the latent heat of vaporization, and E is the vapour flux.

The Bowen ratio is defined as the ratio of the sensible and latent heat fluxes: $\beta = H/L_e E$. The energy equation can then be rearranged in terms of the evaporation rate:

$$E = \frac{Q_n - Q_g}{L_e(1 + \beta)} \, . \qquad 4.2$$

Assuming that the transfer coefficients for heat and vapour fluxes are equal, the Bowen ratio can be written as:

$$\beta = \frac{c_p \Delta T}{L_e \Delta q} \, , \qquad 4.3$$

where ΔT and Δq are the gradients of temperature and specific humidity, respectively.

The net radiation, Q_n, was measured directly, and the soil heat flux at the surface, Q_g, was estimated from the measured heat flux at a depth of 8 cm and the change in energy stored in the soil layer 0 to 8 cm.

The Bowen ratio method has the advantage of being independent of the wind speed; however, as can be seen from Eq. 4.2, when the Bowen ratio values approach -1, the method becomes unreliable. Ohmura (1982) developed some objective criteria for identifying those periods when the Bowen ratio results are not usable; the integrity of the energy balance (Eq. 4.1) is used to ensure that the method produces the proper sign for the vapour flux, and the limits of resolution of the temperature and humidity measurements are used to determine when the ratio approaches -1 and when the gradients are too small to be used with confidence in the calculation. The criteria of Ohmura (1982) were applied to the data, and evaporation calculations for periods in which the Bowen ratio was rejected were replaced by estimates provided by the other calculation techniques.

Aerodynamic formula

Whereas the Bowen ratio approach does not require the measurement of wind speed, its availability allows for the calculation of the vapour flux using an aerodynamic formula such as that introduced by Thornthwaite and Holzman (1939):

$$L_eE = \frac{0.622\alpha k^2 \rho L_e}{P} \frac{u_2(e_1-e_2)}{\ln\left[\frac{z_2 - d_0}{z_0}\right] \ln\left[\frac{z_2 - d_0}{z_1 - d_0}\right]} , \qquad 4.4$$

where k=0.4 is the von Karman constant; ρ is the density of air; P is the atmospheric pressure; u is the wind speed; z is the measurement height; α is the ratio of the turbulent diffusivities for vapour (K_e) and momentum (K_m); and d_0 and z_0 are the displacement and roughness heights, respectively. The subscripts 1 and 2 are used to denote measurements taken at different heights. The expressions relating the displacement and roughness heights to the crop height were given in Eqs. 3.20 and 3.21. The ratio of diffusivities, $\alpha = K_e/K_m$, is a function of the atmospheric stability. The generally accepted approach is to assume that the diffusivities for vapour (K_e) and heat (K_h) are identical; thus, for the unstable range, the ratio α can be estimated using the expression derived by Granger (1977):

$$\alpha = \frac{K_h}{K_m} = \frac{K_e}{K_m} = (1 - 58\zeta)^{0.25} , \qquad 4.5$$

where ζ = z/L is a dimensionless stability parameter; L is a characteristic scaling length introduced by Obukhov (1946) and is defined as:

$$L = -\frac{u_*^3 C_p \rho T}{kgH} , \qquad 4.6$$

where u_* is the friction velocity; C_p is the specific heat of air; g is the acceleration due to gravity; and T is the mean absolute temperature of the boundary layer. The calculation of L requires that the sensible heat flux, H, be known. Granger (1977) provides an approximate solution of L. The heat flux can be expressed in terms of scaling parameters as

$$H = \rho C_p u_* T_* \ , \tag{4.7}$$

where T_* is a characteristic temperature, which can be estimated from the temperature gradient near the surface using the expression:

$$T_* = \frac{k(T_2 - T_1)}{\ln\left[\frac{(z_2 - d_0)}{(z_1 - d_0)}\right]} \ . \tag{4.8}$$

The Obukhov scaling length (Eq. 4.6) then becomes:

$$L = -\frac{u_*^2 T}{kgT_*} \ . \tag{4.9}$$

Energy balance

The energy balance approach consists of calculating L_eE as the residual in Eq. 4.1, in which the sensible heat flux, H, has been estimated using an aerodynamic formula including a correction for stability:

$$H = \alpha \rho C_p k^2 \frac{u_2(T_1 - T_2)}{\ln\left[\frac{z_2 - d_0}{z_0}\right] \ln\left[\frac{z_2 - d_0}{z_1 - d_0}\right]} \ . \tag{4.10}$$

It is interesting to note that whereas the Bowen ratio approach has the relative advantage of being independent of the wind speed, or of the transfer coefficients for heat and vapour, the aerodynamic formula and the energy balance methods are sensitive to errors in the estimate of these transfer coefficients. However, the latter two methods can be combined in such a way as to partially eliminate the effect of such errors. An overestimate of the transfer coefficient results in an overestimate of evaporation when the aerodynamic formula is used, but it results in an underestimate of evaporation when the energy balance is applied. When the evaporation rate is estimated as the average of the aerodynamic formula and energy balance calculations, the errors resulting from uncertainties in the transfer coefficients tend to be compensating; they are in fact completely eliminated when the Bowen ratio is equal to unity. This combination approach gives results which are similar to the Bowen ratio method, without the obvious instabilities resulting from Bowen ratio values near -1.

4.2 Results of the latent heat flux calculations

Figs. 4.2 and 4.3, showing the diurnal cycle of net radiation and the L_eE calculations, compare the three methods of estimating the latent heat flux for two days over growing wheat. The figures show that the methods for estimating L_eE are in reasonable agreement.

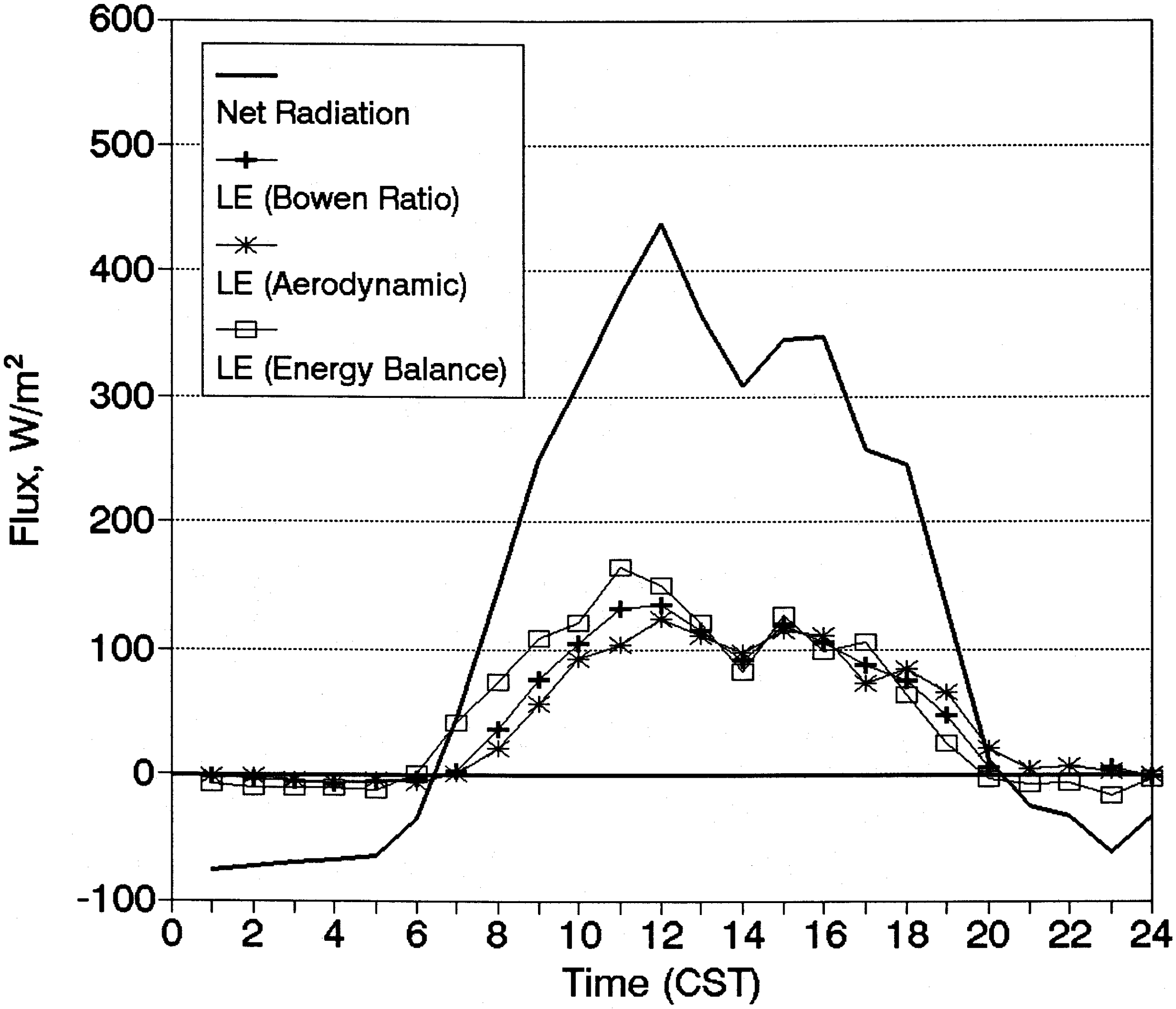

Figure 4.2 Hourly values of the observed net radiation and the calculated latent heat flux over wheat at Saskatoon (June 16, 1990).

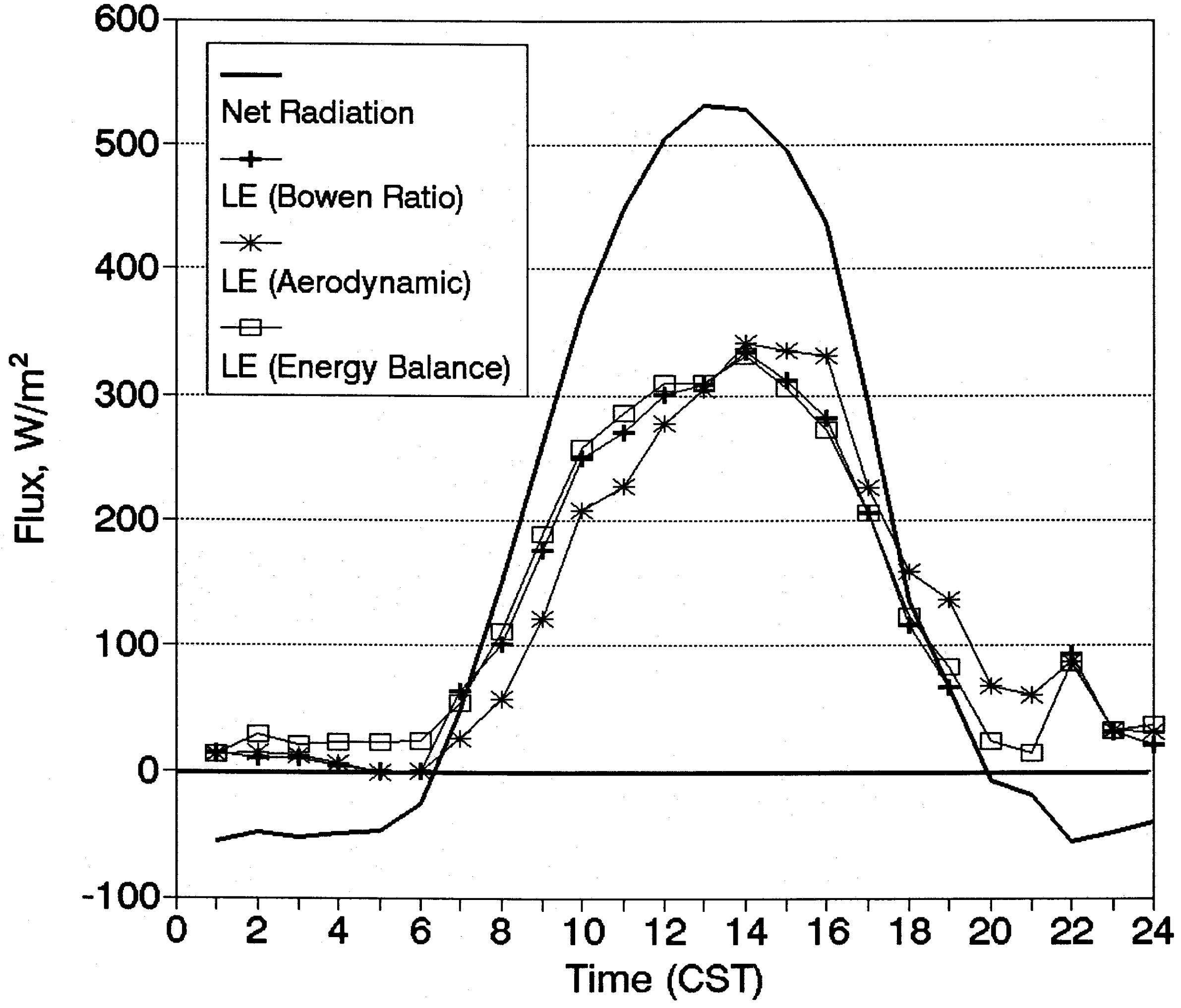

Figure 4.3 Hourly values of the observed net radiation and the calculated latent heat flux over wheat at Saskatoon (June 24, 1990).

The major sources of error are presumed to be the estimate of the soil heat flux, Q_g, and the wind function which relies on estimated values for the displacement and roughness heights. The effect of these respective errors is demonstrated in Figs. 4.2 and 4.3. An error in the wind function results in a divergence between the aerodynamic formula and energy balance estimates; an example of this can be seen in Fig. 4.2 where, in the evening, these two estimates differ by approximately the same amount, but on opposite sides, from the Bowen ratio estimate. An error in the estimate of the soil heat flux may be the cause of the apparent phase lag between the estimates produced by the Bowen ratio and the aerodynamic formula; this is likely the case in Fig. 4.3 where the Bowen ratio and energy balance methods correspond, but the diurnal cycle produced by the aerodynamic formula estimates lags by approximately one hour.

For the daily totals, however, the differences are not large. For example, for June 24 (Fig. 4.3), the Bowen ratio and aerodynamic formula estimates of evaporation were 4.33 and 4.45 mm/d, respectively. Fig. 4.4 compares the daily totals of the estimates produced by the aerodynamic formula and the Bowen ratio approach. The agreement between the two methods is good; the Bowen ratio method produces estimates which are slightly larger than those from the aerodynamic formula. The mean difference between the estimates is 0.186 mm/d, with a standard deviation of 0.318 mm/d.

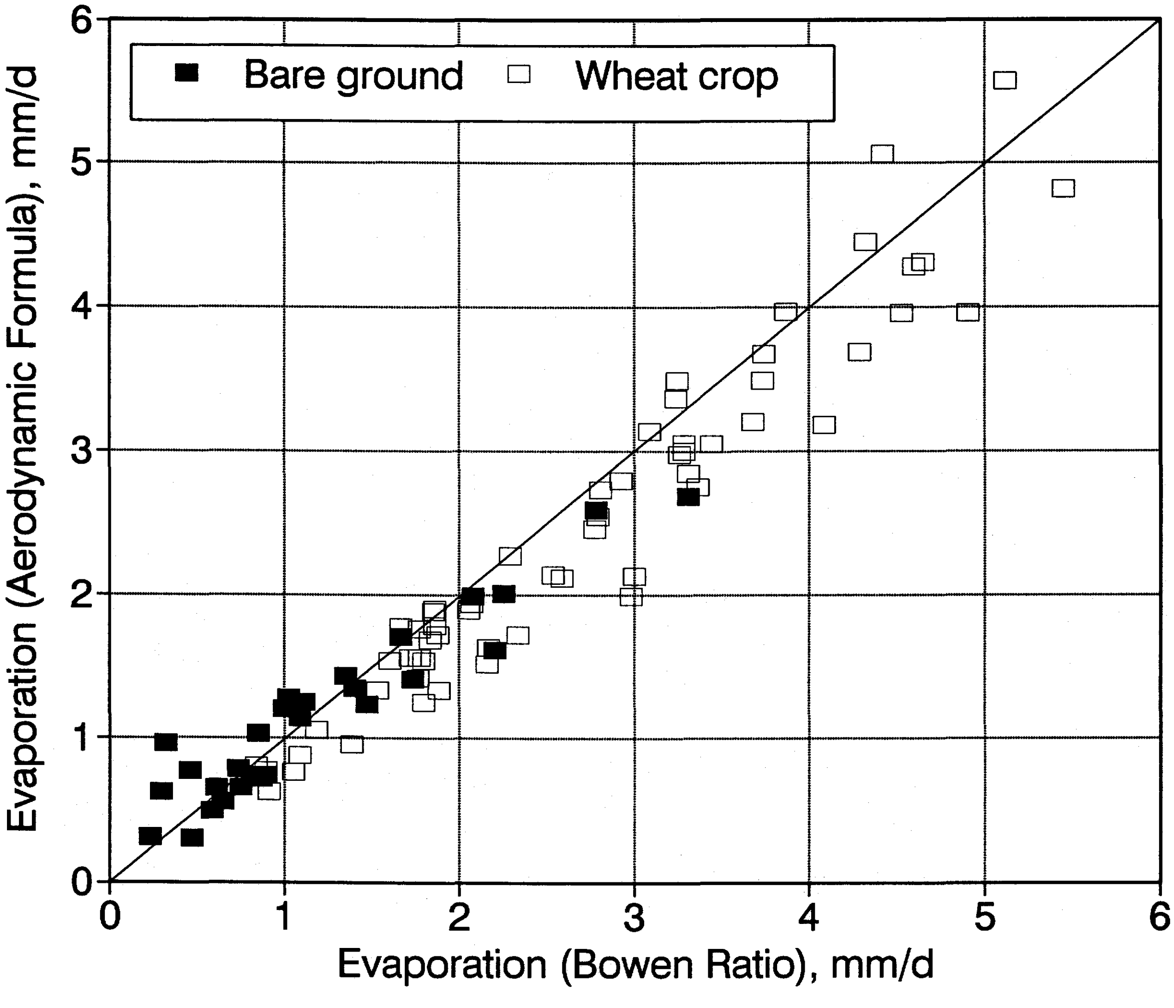

Figure 4.4 Comparison of the daily totals of evaporation as calculated by the aerodynamic formula and the Bowen ratio approaches.

For the purpose of the analyses which follow, the Bowen ratio estimates were used. For periods which were rejected by the criteria of Ohmura (1982) the mean of the aerodynamic formula and energy balance methods was used. The hourly values of the latent heat flux were converted to their equivalent vapour flux and summed to provide daily totals for the evaporation. These daily evaporation estimates obtained from the experimental installations provided reference values against which to compare estimates obtained by the G-D relationship derived in chapter 3, and with which the relationship could be refined.

4.3 The vapour transfer function

Use of the G-D relationship to estimate evaporation requires a prior calculation of the drying power, E_a, which is the product of a wind function and the vapour pressure deficit of the air. In developing the G-D relationship (chapter 3), a wind function derived from the aerodynamic formula, without correction for stability, was used. The aerodynamic formulae of that type, and the transfer functions that are derived from them, are designed for short time periods where quasi-steady conditions can be assumed; for time periods covering one or more diurnal cycles this approach can be questioned. It was used in the initial development because no other appropriate wind function was available. The relatively long time periods used ensured, however, that the mean wind speeds encountered were limited to a small range centred about the long-term mean wind speed for the region; this limited the range of values obtained for the transfer function. As well, it was observed that an error in the drying power causes a point to be displaced along the curve in Fig. 3.2 rather than away from it. In other words, the G-D relationship is much less sensitive to errors in the estimate of drying power than to errors in the estimate of the available energy.

The vapour flux estimates obtained at the measurement site allowed for an examination of the effect of wind speed on the effective vapour transfer function for daily periods. With the daily vapour flux and the mean vapour pressure gradient known, the effective vapour transfer function can be determined as:

$$f(u) = \frac{E}{e_s - e_a} \ . \qquad 4.11$$

The vapour pressure at the surface, e_s, was estimated by extrapolating the logarithmic vapour pressure gradient down to the effective surface at $d_0 + z_{0e}$. z_{0e} is the roughness length for water vapour. Brutsaert (1982) points out that the roughness lengths for scalar admixtures, such as heat and water vapour, are much smaller that the momentum roughness parameter, z_0. It was assumed here that the roughness lengths for heat and water vapour are identical. By assuming a logarithmic temperature profile and extrapolating from the mean temperature gradient in the air, the value of z_{0e} was determined as the height at which the temperature was equal to the mean surface temperature observed with the infrared temperature detector. These observations confirmed that the roughness parameter for scalar admixtures is much smaller, generally by at least one order of magnitude, than the z_0 values obtained from Eq. 3.21. For example, for bare ground, a value of 0.0002 m was found to be typical for z_{0e}, while for z_0 a value of 0.006 m was used. This value of the scalar roughness length, z_{0e}, was then used to obtain an estimate of the vapour pressure at the effective surface, e_s. Eq. 4.11 was then used to determine the effective vapour transfer function for daily periods.

Fig. 4.5 plots the calculated values of the daily vapour transfer coefficient for bare ground and growing wheat against the mean wind speed at 2 m. The scatter in the points is to be expected since the estimates do not account for the effect of stability or for the change in surface roughness with crop height. Nevertheless, as a first approximation, the best-fit lines were drawn through the data sets for bare ground and growing wheat. The equations describing these lines are, for bare ground:

$$f(u) = 7.50 + 1.36u \quad (\mathrm{mm \cdot d^{-1} \cdot kPa^{-1}}), \qquad 4.12$$

with $r^2 = 0.648$, and for growing wheat:

$$f(u) = 11.75 + 1.69u \quad (\mathrm{mm \cdot d^{-1} \cdot kPa^{-1}}), \qquad 4.13$$

with $r^2 = 0.553$.

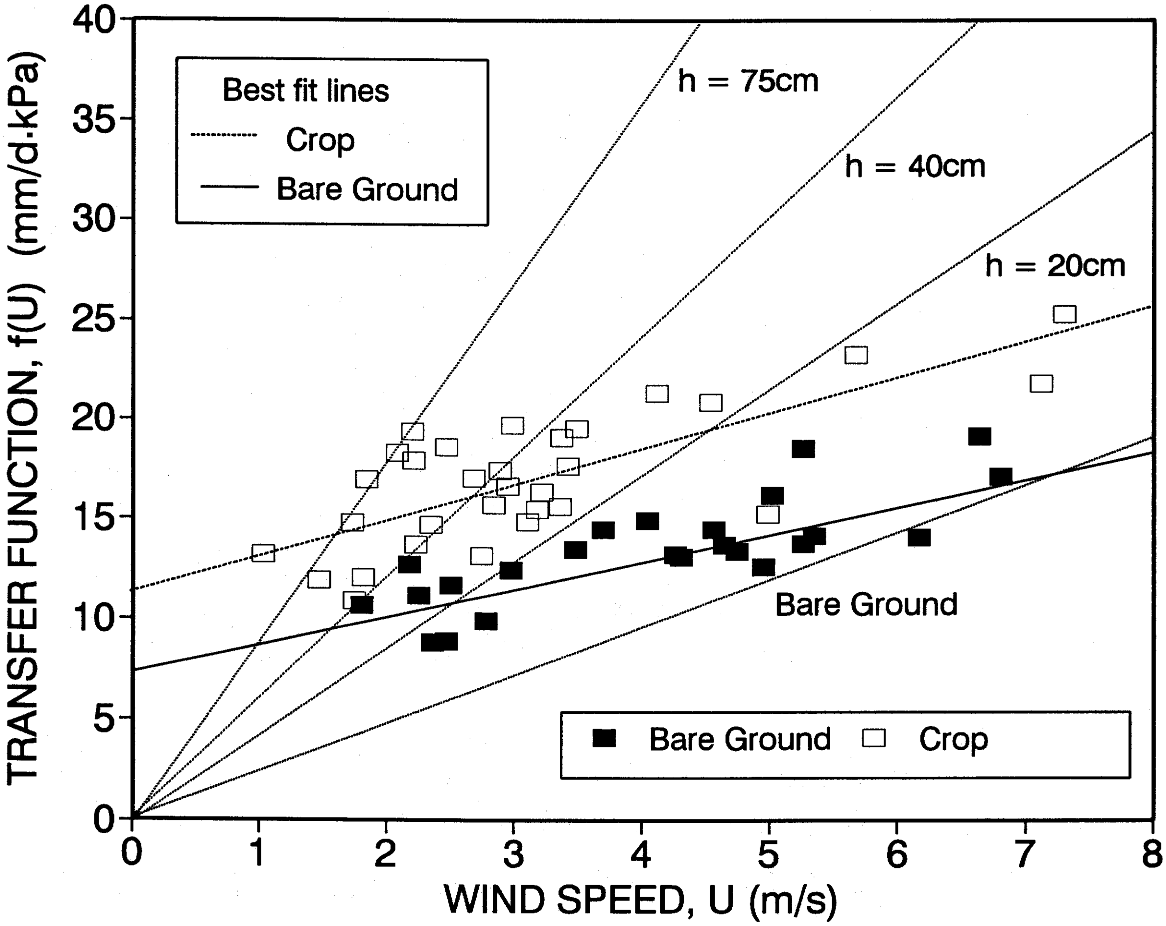

Figure 4.5 Effective vapour transfer function for daily periods plotted against mean wind speed at 2 m for bare ground and wheat.

Also shown in Fig. 4.5 are the vapour transfer coefficients calculated using the aerodynamic formulation for bare ground and for surfaces with vegetation 20 cm, 40 cm and 75 cm in height. The figure shows that for daily periods, the effect of wind speed on the vapour transfer coefficient is reduced. Morton (1983) presented intuitive arguments to suggest that in fact the effect of wind speed would be eliminated and that the transfer function could be taken as a constant. Fig. 4.5 shows that although the effect of the wind speed is significantly reduced for daily time periods, it is by no means eliminated.

It is interesting to note that for wind speeds near the long term mean value for the region (approximately 4 m/s), and for bare surfaces and surfaces covered with vegetation shorter than 0.5 m (which covers the bulk of the cases used in the analysis in chapter 3), the aerodynamic estimates for the transfer coefficient are in the same range as those obtained from Eqs. 4.12 and 4.13.

4.4 Daily estimates of evaporation

The observations obtained in the field experiment provided a set of reference evaporation data (daily totals of the Bowen ratio estimates) against which the estimates produced by the G-D relationship could be compared. Because the Bowen ratio values are themselves only estimates of evaporation, the comparison does not constitute an absolute verification; but it does provide an appreciation of the relative performance of the procedure.

The expressions (Eqs. 4.12 and 4.13) derived for the daily vapour transfer coefficient were used to calculate the drying power, E_a; and the relative drying power, D, was obtained from the calculated drying power and the daily totals of the observed net radiation and soil heat flux. The relative evaporation, G, was obtained from its derived relationship with D (Eq. 3.22), and the general equation (Eq. 3.10) was used to estimate the daily totals of evaporation. The calculations were carried out with the daily values for the three observation periods.

The calculated evaporation was also compared with the estimates produced by the Complementary Relationship Areal Evapotranspiration (CRAE) model. [The model, developed by Morton (1983) is based on the relationship presented by Bouchet (1963), and has been subjected to extensive verification and refinement by its author (Morton, 1983; Morton et al., 1985)]. It requires as input mean period values for the temperature and humidity of the air, and an index of solar radiation (observed duration of bright sunshine or observed global radiation). The temperature and humidity data were those obtained at the observation site. The CRAE model contains a series of algorithms to calculate the net radiation based on observations of the sunshine duration; however, since Granger and Gray (1989, 1990a) have assessed that the radiation estimates represent a weak link in the CRAE model, the observed insolation (global radiation) from the Saskatchewan Research Council meteorological observation station (approximately 4 km from the site) was used as input. All calculations were carried out for daily periods.

The daily evaporation estimates produced by the G-D relationship are compared to the Bowen ratio and the CRAE calculations in Figs. 4.6, 4.7 and 4.8 for the three observation periods, respectively. The figures present accumulation curves of evaporation obtained from the summation of the daily totals.

Figs. 4.6 to 4.8 show a good agreement between the G-D and the Bowen ratio estimates, both as to the seasonal patterns and the period totals. In all three cases the G-D curve corresponds more closely to the Bowen ratio curve than does the line representing the CRAE results. The differences between the G-D and Bowen ratio estimates of the period totals are -1.3 mm in 23 days, 1.5 mm in 26 days and 6.6 mm in 93 days for the three observation periods shown in Figs. 4.6, 4.7 and 4.8, respectively. The CRAE estimates are consistently higher, probably due in large part to the fact that the CRAE model ignores the soil heat flux term, while the other two methods take this flux into account. (It will be shown in the following chapter that the daily soil heat flux term is important, representing near 10% of the net incoming radiation in the case of a growing crop and 20 % for bare ground.)

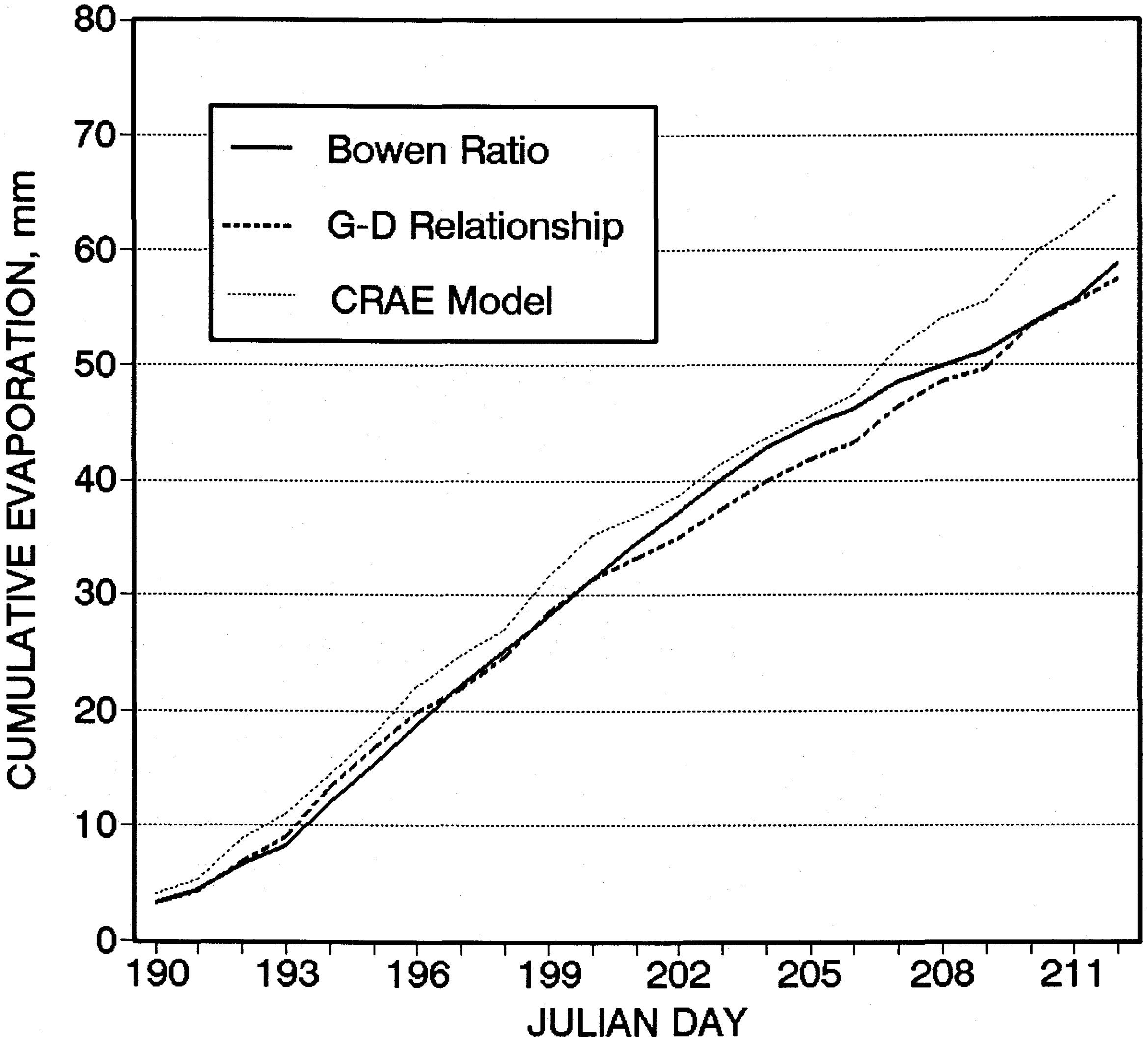

Figure 4.6 Mass curves of daily evaporation over a growing wheat crop for the period July 9 to 31, 1989, at Saskatoon, as estimated by the Bowen ratio approach, the G-D relationship and the CRAE model.

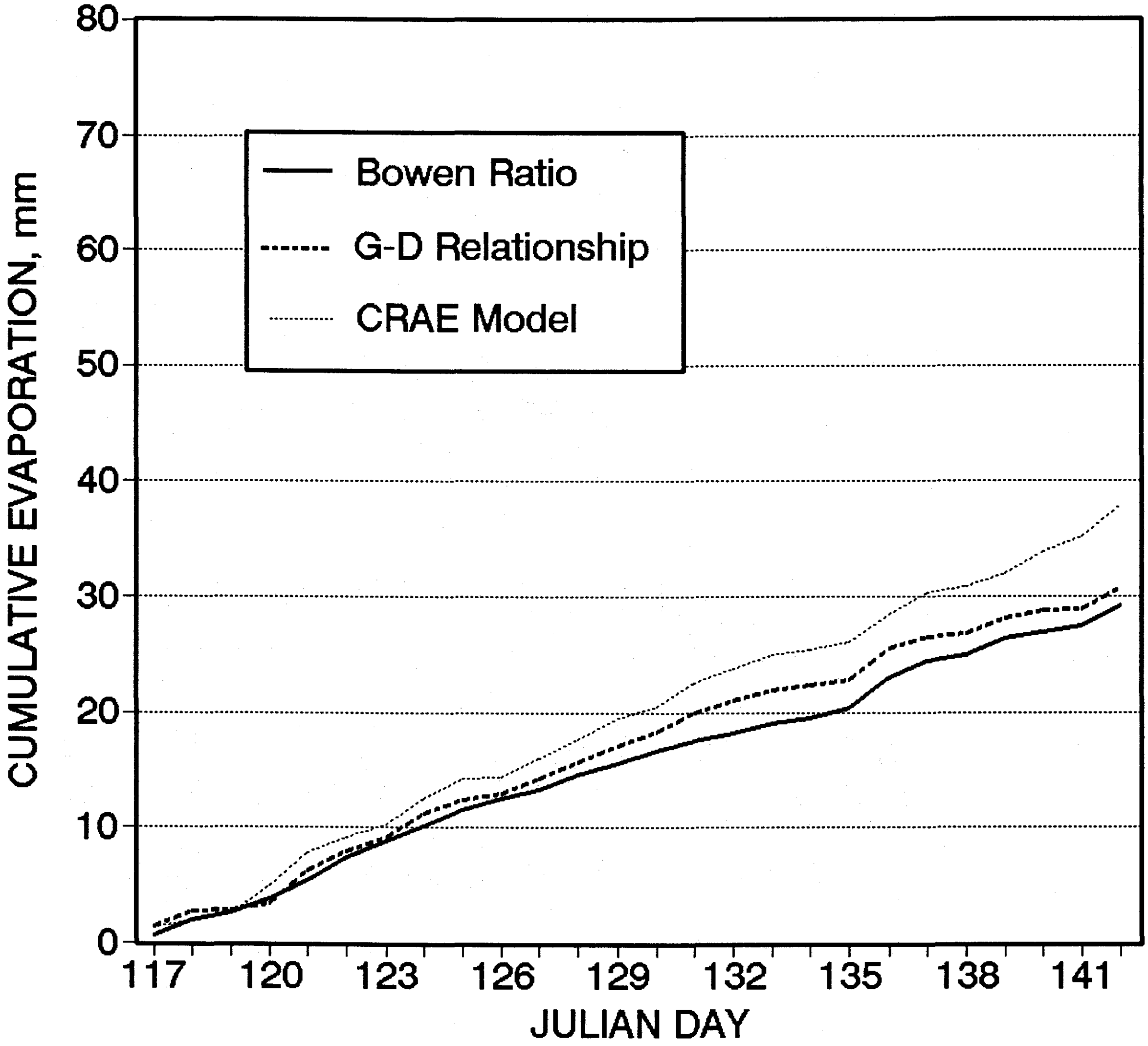

Figure 4.7 Mass curves of daily evaporation over bare ground for the period April 27 to May 22, 1990, at Saskatoon, as estimated by the Bowen ratio approach, the G-D relationship and the CRAE model.

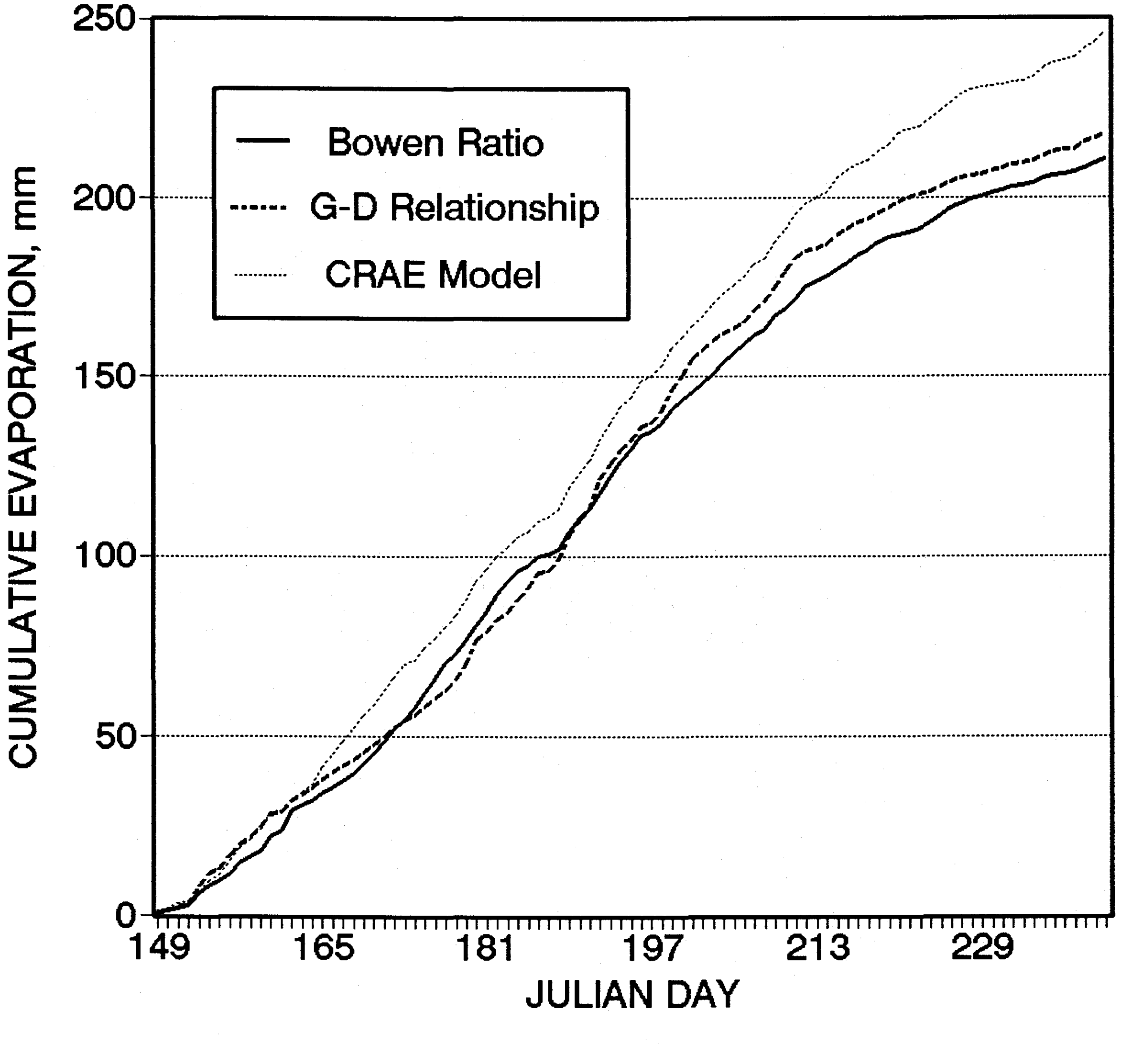

Figure 4.8 Mass curves of daily evaporation over a growing wheat crop for the period May 29 to August 29, 1990, at Saskatoon, as estimated by the Bowen ratio approach, the G-D relationship and the CRAE model.

While accumulation curves such as those presented in Figs. 4.6 to 4.8 can provide a reasonable visual comparison of the trends and the accumulations, they do not constitute the best test of concurrence of two methods of obtaining estimates of the evaporation rates. Fig. 4.9 presents a comparison, in the form of a scatter diagram, of the G-D calculations of the daily evaporation rates against those obtained using the Bowen ratio method. The data points from the three periods are presented. The observed daily evaporation rates varied from 0 to approximately 4.5 mm/d, with the largest rates occurring over the actively growing crop. The largest discrepancies between the estimates by the two methods occurred during the 1989 growing season; however, no obvious differences in observation technique or procedure could be found to explain this trend. Considering the three data sets, the mean difference between the daily estimates of evaporation by the G-D relationships and Bowen ratio method shown in Fig 4.9 is 0.11 mm/d, with a standard deviation of the difference of the estimates of 0.48 mm/d. For the 1990 data alone, the standard deviation of the difference between the two methods is 0.41 mm/d, or 9% of the observed maximum daily flux. This compares favourably, for example, with Morton (1983) who showed that his model agrees within 9.9% of monthly basin water balance estimates of evaporation. If the Bowen ratio estimates can be considered reasonable, then the G-D relationship approach appears to provide a good estimate of daily evaporation.

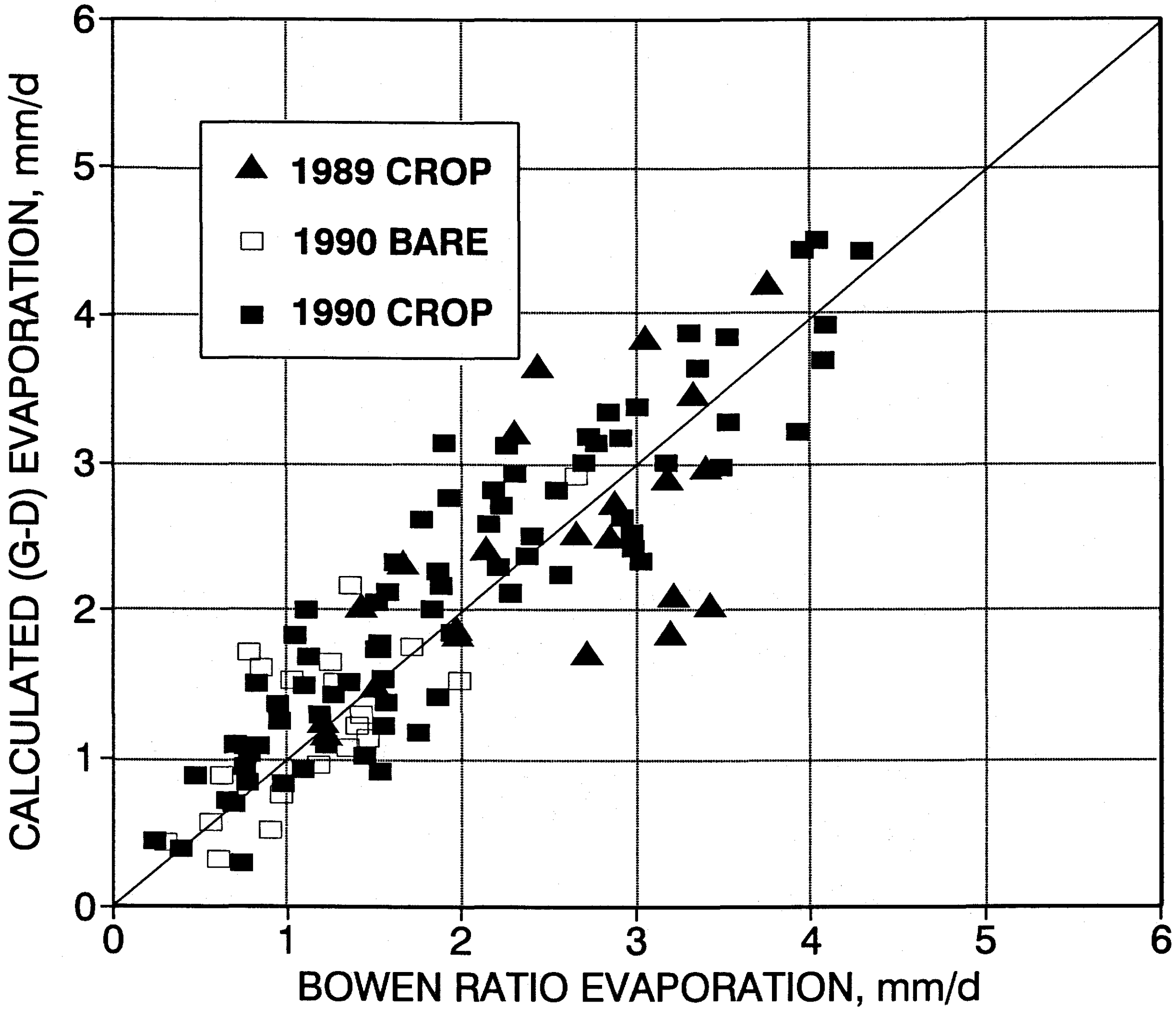

Figure 4.9 The daily values of evaporation estimated using the G-D relationship (Eq. 3.22) plotted against the daily values of evaporation obtained with the Bowen ratio approach, for the three observation periods.

4.5 Refinement of the G-D relationship

The relationship (Eq. 3.22) between the relative evaporation, G, and the relative drying power, D, presented in chapter 3 was derived using soil moisture data for periods of variable length ranging from 2 to 30 days. That data set allowed only the dryer portion ($D > 0.35$) of the range of available moisture to be sampled (cf. Fig. 3.2). The data collected during the three observation periods in this study provide a data set with which to re-evaluate the G-D relationship using consistent time periods of one day, and to extend the range of the relationship toward the wetter region.

The derived vapour transfer functions for bare ground (Eq. 4.12) and growing wheat (Eq. 4.13) were used to calculate the drying power, E_a. Using the daily totals of the Bowen ratio evaporation estimates and the observed daily totals of the net radiation and soil heat flux, the relative evaporation, G, for the daily periods was recalculated from Eq. 3.17. These daily values of G are shown plotted against the corresponding values of the relative drying power, D, in Fig. 4.10. The data in the figure show somewhat less scatter than those used in the original development (Fig. 3.1). The reduced scatter may be attributable in part to the use of a transfer function appropriate to the time scale. Although the use of a shorter time period did result in more points in the mid-range of wetness, no data from the wet range ($D < .2$) were obtained.

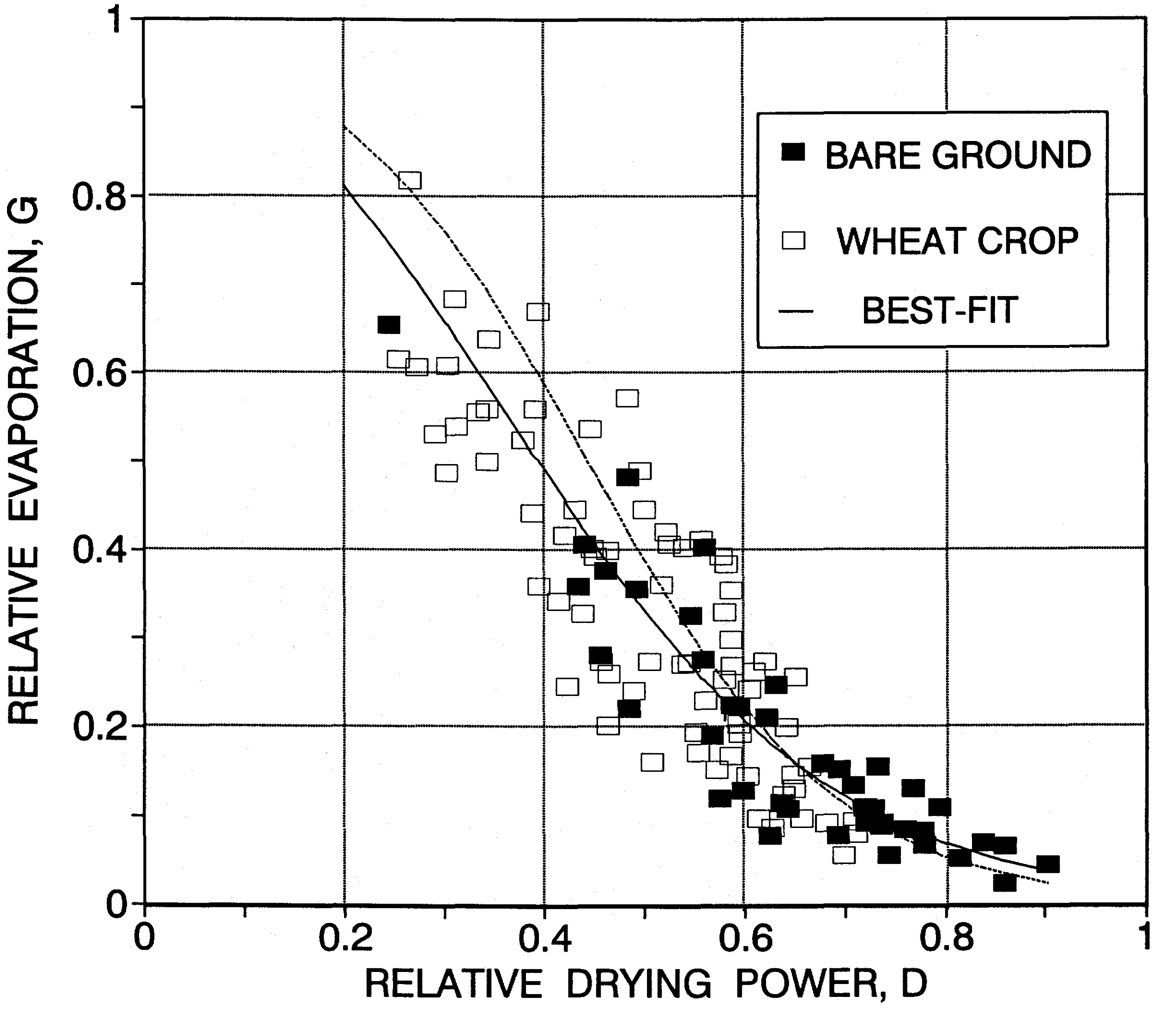

Figure 4.10 Daily values of the relative evaporation, G, plotted against the relative drying power, D, for bare ground and growing wheat. The best-fit line (solid) and the original relationship (Eq. 3.22, dashed) are shown.

The data in Fig. 4.10 do not show any systematic difference between bare ground and growing wheat; this lends support to the assertion that the relationship between relative evaporation and relative drying power is single-valued, that is, that it is the same regardless of the landuse or surface conditions encountered provided that the different vapour transfer functions are correctly taken into account.

The solid line shown in Fig. 4.10 is the "best-fit" line (fitted by trial and error). The dashed line represents the relationship (Eq. 3.22) derived using soil moisture balance data. The best-fit (solid) line is represented by the equation:

$$G = \frac{1}{0.905 + 0.095\, e^{6.20\, D}} \quad . \qquad 4.14$$

The line confirms the "ogee" shape, or S-shape, of the relationship, although it is less pronounced than the original curve (Eq. 3.22). Eqs. 4.14 and 3.22 provide similar results for G in the dryer range (D > 0.6), however Eq. 4.14 gives lower values for G in the wetter region than does Eq. 3.22. Inspection of Fig. 3.2 suggests that the original Eq. 3.22 does give estimates for G which are too high in this region. The mean difference between values of G calculated with Eq. 4.14 and those obtained the daily Bowen ratio data in Eq. 3.17 is 0.007, with a standard deviation of 0.083.

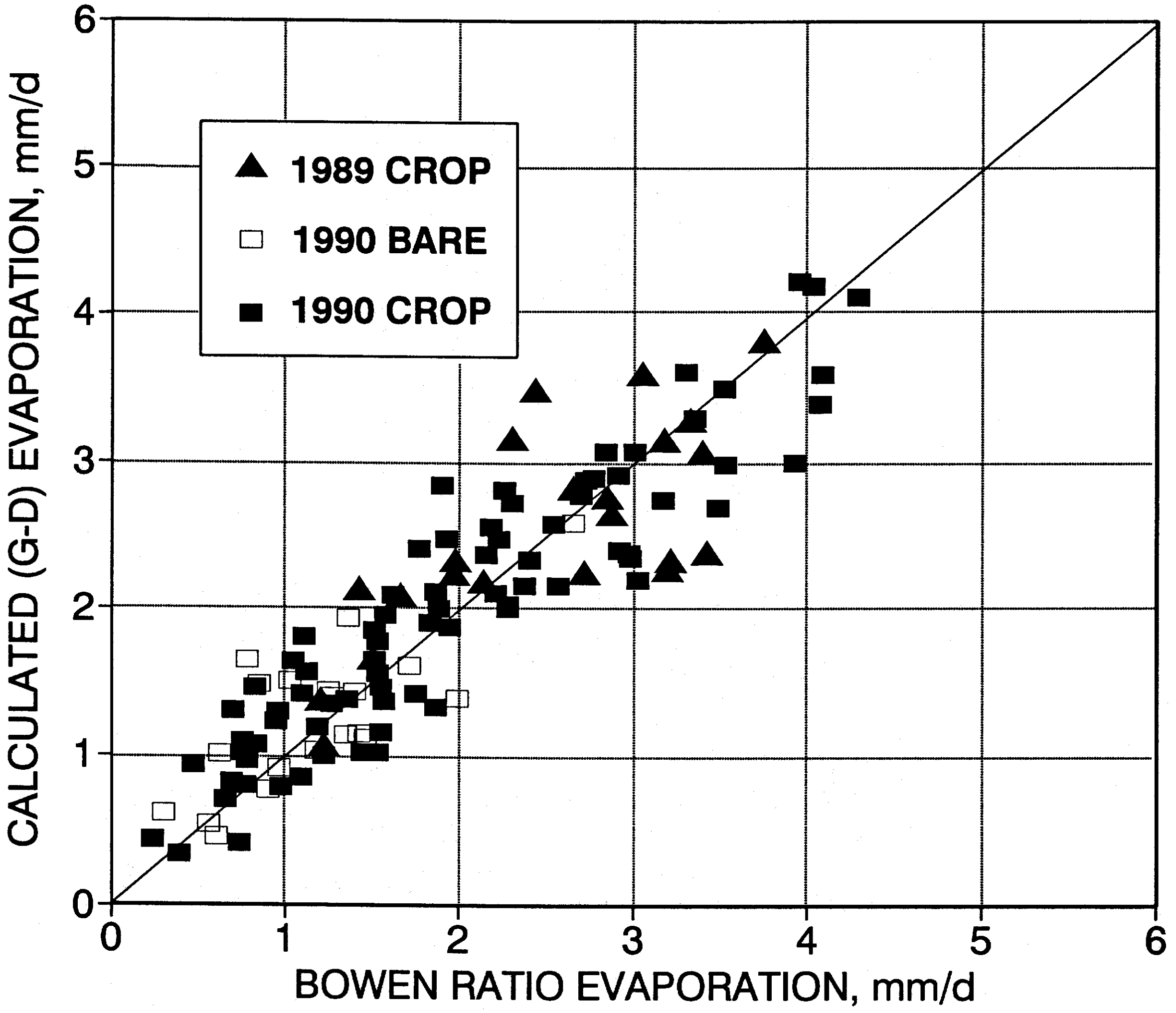

Figure 4.11 The daily values of evaporation estimated using the revised G-D relationship (Eq. 4.14) plotted against the daily values of evaporation obtained with the Bowen ratio approach, for the three observation periods.

The daily evaporation was recalculated using the G-D relationship (Eq. 4.14) in conjunction with the general equation (Eq.3.10), and the results plotted against the reference Bowen ratio evaporation values in Fig. 4.11. The mean difference between the two estimates is 0.03 mm/d, and the standard deviation of the difference is 0.41 mm/d. As expected this comparison is somewhat better than that resulting from the original relationship (cf. Fig. 4.9).

4.6 Comparison with soil water balance results

Soil moisture and precipitation measurements were also obtained during the growing season in the 1990 observation period; the data allowed for the estimation of evaporation losses from the growing wheat using a soil water balance. Because these data are different from those used in the model calculation, this approach provides an independent estimate of the evaporation from the field against which the results of the calculations from the G-D approach can be compared.

Twin-probe gamma soil moisture profiling sites were located approximately 20 m north and southwest of the experimental installation in the wheat field (these sites were designated as #14 and #2, respectively); a third site (#20) was located approximately 180 m south of the installation. The procedure used for determining soil moisture changes with this method has been described in Chapter 3. Soil moisture measurements were obtained on the first day of the observation period (May 29) and approximately bi-weekly during the growing season; precipitation was monitored continuously. The total accumulated evaporation during measurement intervals is obtained as the difference between the total precipitation and the accumulated soil moisture change, ΔS. Shrinkage cracks developed in the soil between the access tubes after July 17 at sites #2 and #14, and the moisture estimates were unreliable after this date; however, site #20 provided useful data throughout the growing season.

The soil moisture changes and evaporation estimates obtained at these three sites are presented in Table 4.2, and the cumulative evaporation is compared with the G-D model and Bowen ratio estimates in Fig. 4.12. There is excellent agreement between the estimates obtained at the three sites; the maximum difference between accumulation estimates on July 17 is 11.4 mm, which is equivalent to a rate of 0.2 mm/d. There is also excellent agreement between the G-D model and the moisture balance estimates as shown in Fig. 4.12; at July 17 the G-D estimate of the total evaporation is identical to the mean of the three moisture balance estimates, and the difference between these estimates of accumulated evaporation to August 20 is 5.2 mm, which represents a mean rate of 0.06 mm/d. This compares favourably with the results obtained for the comparison of the G-D and Bowen ratio results (Fig. 4.11).

Table 4.2 Summary of the evaporation estimates obtained at the soil moisture monitoring sites on a wheat field at Saskatoon, 1990. (All accumulated changes are referenced to May 29, 1990)

		SITE #2		SITE #14		SITE #20		G-D
Date	Precip. mm	ΔS mm	E mm	ΔS mm	E mm	ΔS mm	E mm	E mm
Jun 13	43.6	+15.9	27.7	+10.3	33.3	-0.3	43.9	37.5
Jun 21	49.2	+1.3	47.9	+3.0	46.2	-1.5	50.7	56.6
Jul 5	75.9	-23.5	99.4			-31.9	107.8	102.0
Jul 17	99.8	-40.0	139.8	-51.4	151.2	-48.2	148.0	146.5
Aug 8	120.1					-81.0	201.1	207.0
Aug 20	122.6					-94.1	216.7	221.9

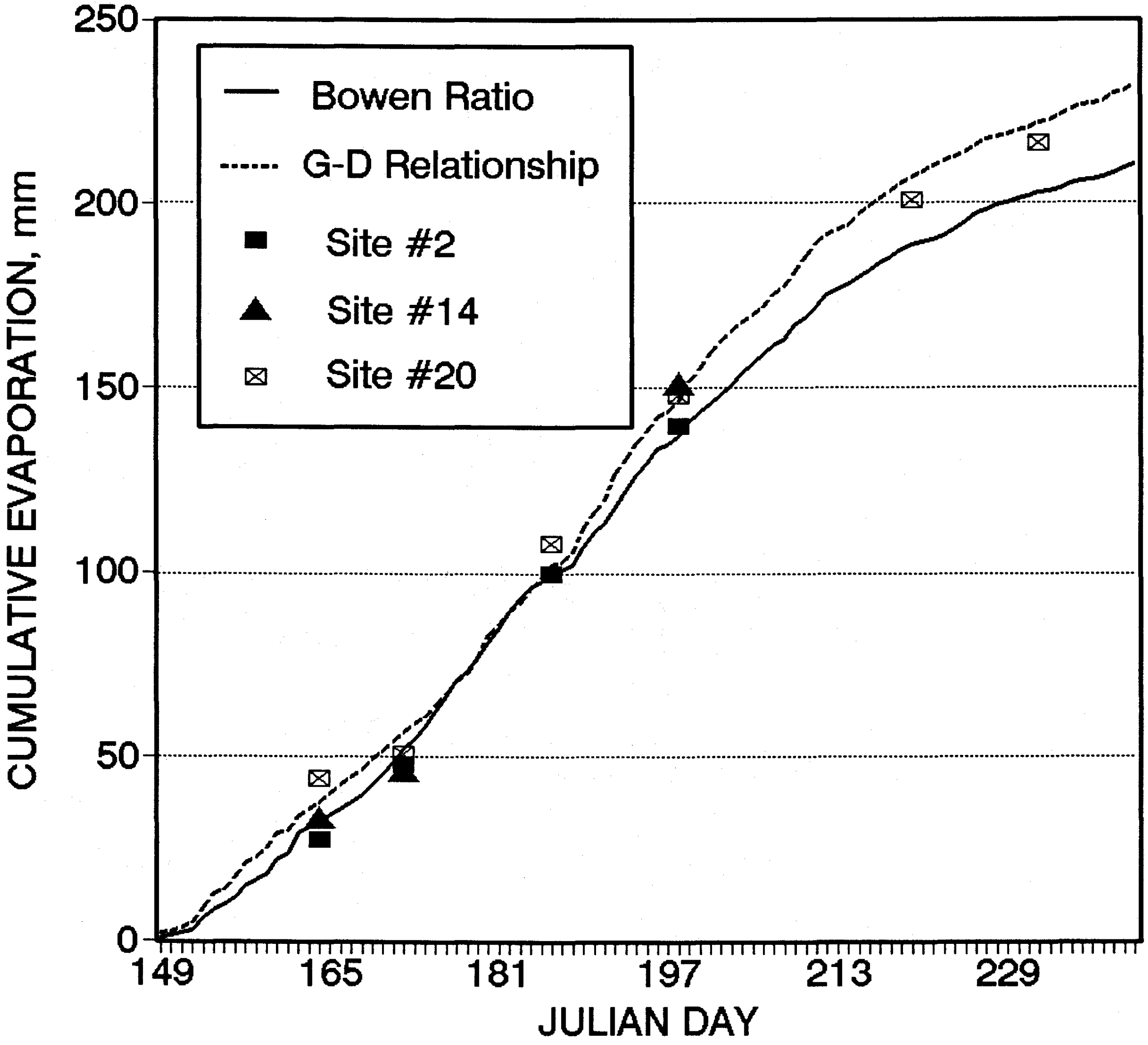

Figure 4.12. Comparison of the G-D estimates of evaporation with those obtained from the soil moisture balance sites on a wheat field at Saskatoon, 1990.

4.7 Hourly estimates of evaporation

The field data also allowed for the assessment of the G-D relationship for time periods less than one day. Hourly periods were used for this test. The hourly values for the drying power, E_a, were calculated using the wind function derived from the aerodynamic formula (Eq. 3.19) with the correction for stability (Eq. 4.5) included. The observed hourly values of the net radiation and soil heat flux terms and the calculated drying power were used to derive hourly values for D. The hourly Bowen ratio estimates of evaporation were used, with the hourly values of the available energy and drying power, in Eq. 3.17 to calculate hourly values for the relative evaporation, G. Because of the nature of the relationship, those periods when the available energy was negative, as well as when condensation was occurring, were necessarily omitted.

Figs. 4.13 and 4.14 show the relationship between hourly estimates of relative evaporation, G, and relative drying power, D, for bare ground and growing wheat, respectively. The curve representing the relationship derived using daily values (Eq. 4.14) is also shown in the figures. These data confirm that (i) the relationship between relative evaporation and relative drying power holds for the shorter time periods; (ii) Eq. 4.14 is a good predictor of the relationship for hourly periods as well; (iii) the relationship also applies in the wet range ($D < 0.2$).

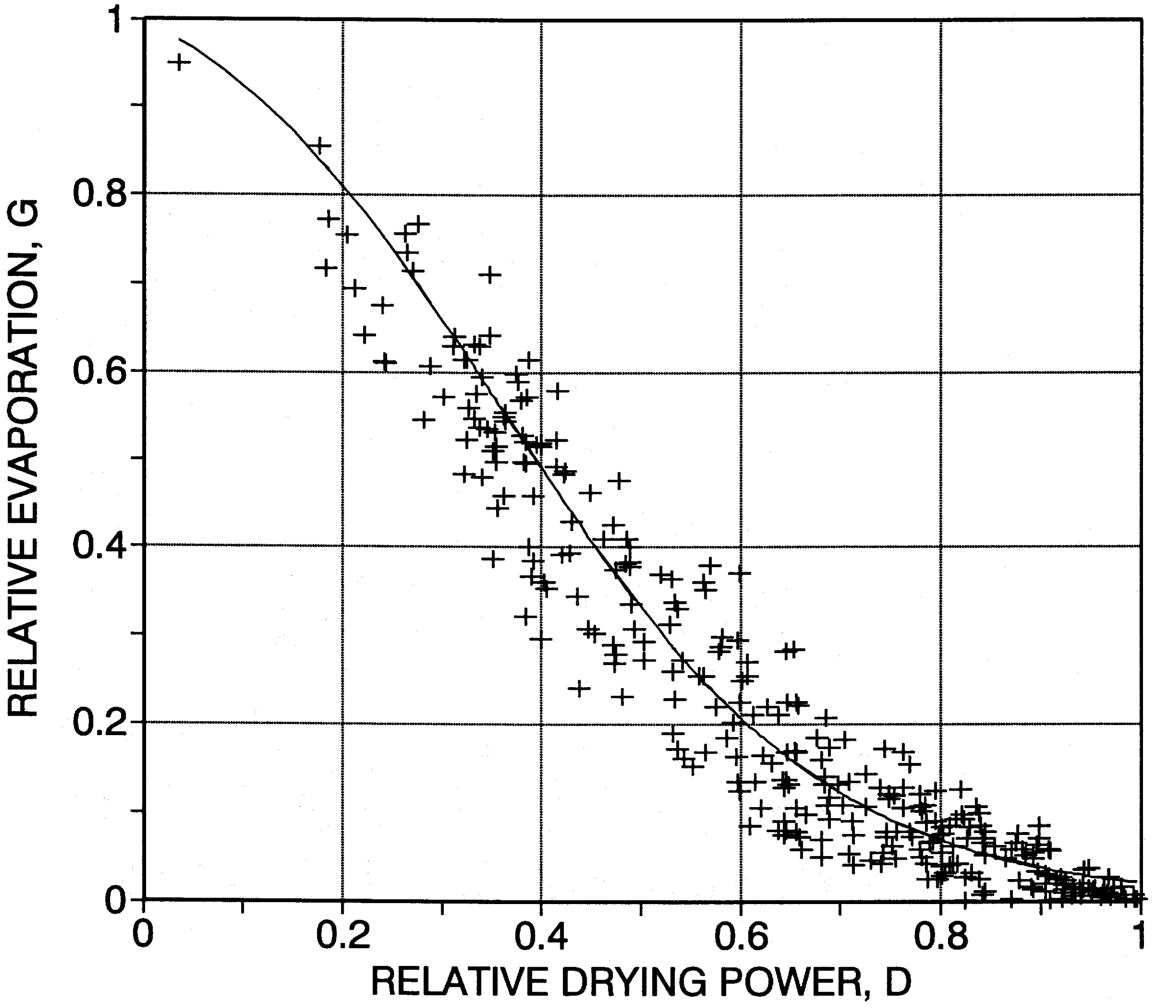

Figure 4.13 Hourly values of the relative evaporation, G, plotted against the corresponding values of the relative Drying power, D, for bare ground. The curve represented by Eq. 4.14 is also shown.

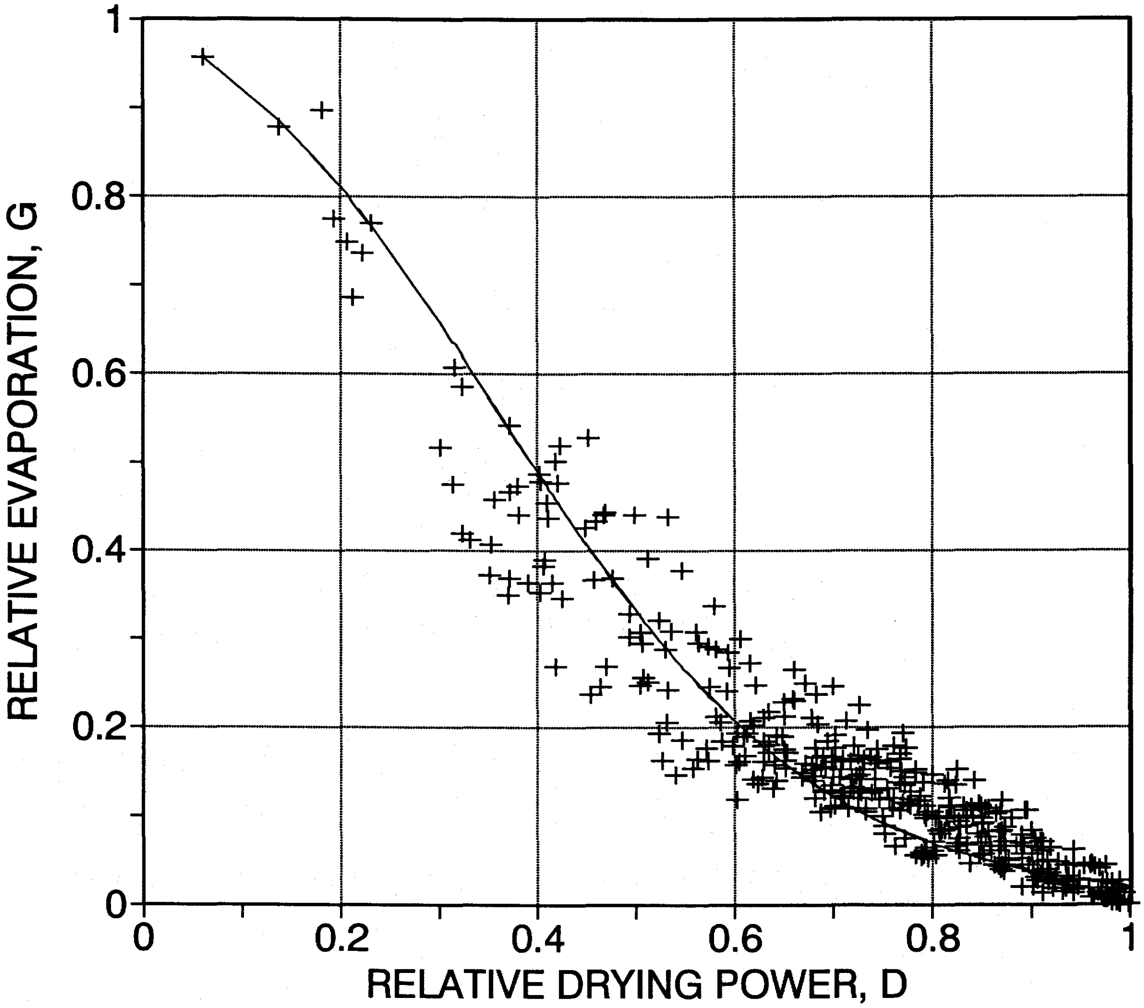

Figure 4.14 Hourly values of the relative evaporation, G, plotted against the corresponding values of the relative Drying power, D, for a growing wheat crop. The curve represented by Eq. 4.14 is also shown.

For the case of bare ground (Fig. 4.13), where 301 points were used, the mean difference between the values of G estimated using Eq. 4.14 and those represented by the data points is 0.012, and the standard deviation of the difference is 0.059. For the growing crop (Fig. 4.14) 403 data points were used; the corresponding mean difference is -0.009, and the standard deviation of the difference is 0.054. These statistics suggest that the relationship (Eq. 4.14) applies equally well to both landuses. However, Fig. 4.14 shows that for a growing crop and D in the range from 0.6 to 0.9, the data points are for the most part concentrated above the line. Such a deviation could be caused by an underestimate of either the actual evaporation rate (error in the wind function for taller vegetation or in the stability correction for very unstable cases) or the available energy.

4.8 Summary

The experimental data obtained during the three observation periods, over bare ground and growing wheat, served to develop a preliminary relationship for the vapour transfer function for daily time periods. The evaporation calculations resulting from the experimental data were used to confirm the proposition that there exists an apparently single-valued relationship between the relative evaporation, G, and the relative dying power, D. This relationship is described by Eq. 4.14. With the use of an appropriate wind function, the G-D relationship is shown to be applicable for hourly periods and for periods of one day or longer.

The equation representing this relationship (Eq. 4.14), when used in conjunction with the derived general equation (Eq. 3.10) forms the basis of a simple, reliable method for estimating evaporation from nonsaturated surfaces. Independent estimates of evaporation obtained using soil moisture balance data were also used to confirm the validity of this approach.

5. MODELLING AVAILABLE ENERGY AND EVAPORATION

The relationships developed in the preceding chapters (Eq. 3.10, the general equation for evaporation from nonsaturated surfaces, and Eq. 4.14, the relationship between the dimensionless parameters relative evaporation and relative drying power) form the basis of a relatively simple, yet effective, routine for estimating the evaporation from natural surfaces. The method has been shown to be applicable for daily time periods or longer; it may also be used for shorter periods of the order of one hour; however, at the present stage of development, its applicability is limited to periods when the available energy is positive and evaporation (not condensation) is occurring.

The method requires as input data the same parameters as used in the Penman equation, that is, available energy (net radiation and soil heat flux), wind speed, and the vapour pressure deficit of the air (temperature and vapour pressure). Whereas air temperature and wind speed, and in some cases humidity, are routinely observed and can be attributed some measure of areal representativeness, the net radiation and soil heat flux terms are much more site dependant and are not so frequently observed. It is often more convenient to estimate the net radiation term from other routinely measured parameters, such as the observed duration of bright sunshine.

This chapter will examine techniques for estimating the daily net radiation and the soil heat flux for use in modelling evaporation. As well, the use of remotely-sensed surface temperature data in the parameterization of areal evaporation will be explored, and a new approach will be presented.

5.1 Modelling net radiation

Evaporation models, such as the CRAE model (Morton, 1983), usually incorporate algorithms for estimating the net radiation input to the surface being considered. An examination of the CRAE estimates of the radiation fluxes (Granger and Gray, 1990a) showed that the monthly structure of the radiation model and the inherent errors in the estimation of the extra-terrestrial flux, the transmittancy of clouds to short-wave radiation, the surface albedo and the net long-wave flux can result in significant errors in the modelled net all-wave radiation.

An attempt was thus made to establish a series of relationships which could provide the framework for a more reliable estimate of the daily net radiation. Climatological observations from three stations located in the Prairie provinces (Edmonton, Bad Lake, Winnipeg) were used to establish empirical relationships for the daily values of the components of the net radiation. Table 5.1 presents the coordinates of the stations, the parameters measured and the period of observation of the data used in the analysis.

Table 5.1 Information on the meteorological stations and observations used in the analysis of the daily radiation fluxes.

Station	Latitude (deg)	Longitude (deg)	Altitude (masl)	Parameters[a]	Observation Period
Edmonton	53°33'N	114°06'W	766	n, RF1,2,4	1977-83
Bad Lake	51°20'N	108°24'W	637	n, RF1,2,3,4	1980-85
Winnipeg	49°54'N	97°14'W	239	n, RF1,2	1977-83

[a] n = hours of bright sunshine; RF1 = global short-wave; RF2 = diffuse short-wave; RF3 = reflected short-wave; RF4 = net all-wave radiation.

In the analysis of the radiation fluxes, the net radiation, Q_n, was divided into its net short-wave, Q_{sn}, and net long-wave, Q_{ln}, components; the net short-wave is taken as the difference between the incident short-wave, Q_s, and the reflected short-wave, Q_r; and the incident short-wave flux is further divided into its direct beam, Q_{drs}, and diffuse, Q_{dfs}, components. Equations describing each of these components under clear-sky conditions were derived, and relationships to account for the moderating effects of cloud cover were subsequently established. The details of the development of the radiation model are described by Granger and Gray (1990b); the relationships along with the relevant statistics are presented in Appendix A.

The use of regression relationships such as those presented in Appendix A is not entirely new. However, the approach used to parameterize the net long-wave flux is innovative and warrants being highlighted here. Whereas the short-wave fluxes are readily derived from geometric considerations and other parameters such as the effective atmospheric transmissivity and duration of bright sunshine, the long-wave fluxes are more complex. The atmospheric flux of long-wave radiation is emitted by ozone, carbon dioxide, dirt particles, water vapour and other contaminants from all atmospheric levels; under clear skies the major portion originates from the lowest 100 m layer. Since the largest contributor is presumed to be the water vapour, many investigators have related the atmospheric clear-sky long-wave flux to the temperature and vapour pressure measured near the surface. Brunt (1932), Brutsaert (1975) and Satterlund (1979) present expressions of this type. Using the data presented above, Granger and Gray (1990b) showed that this traditional approach using air temperature and vapour pressure is not generally reliable; in nature, the atmospheric conditions of temperature and humidity lapse rates vary considerably and are often quite different from the ideal atmosphere upon which the empirical relationships are based. For a natural surface the fluctuations in surface and atmospheric temperatures (and to a large extent the variations in humidity) are driven by the energy supplied to the surface; thus an alternative approach for estimating the long-wave flux is to correlate it with the incident short-wave flux. Fig. 5.1 shows daily values of the clear-sky net long-wave radiation, Q_{lno}, plotted against the incident short-wave flux, Q_{so}, for Bad Lake.

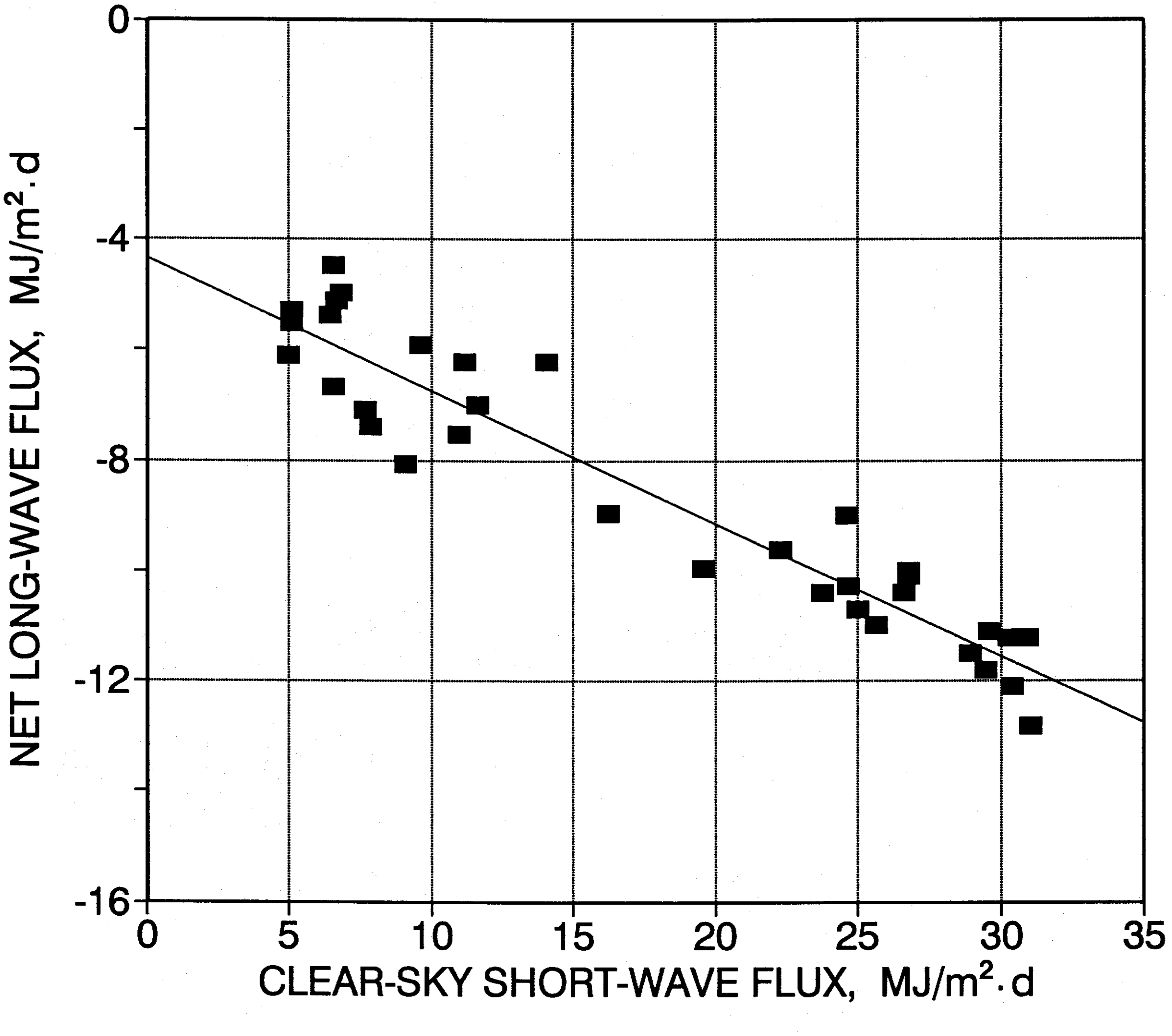

Figure 5.1 Relationship between daily net long-wave and incident short-wave fluxes on clear days at Bad Lake.

Fig. 5.1 shows that a well-defined relationship exists between the net long-wave and incident short-wave fluxes. The best-fit line describing the data in Fig. 5.1 is given by:

$$Q_{lno} = -4.25 - 0.24\, Q_{so} \,, \qquad 5.1$$

where Q_{lno} and Q_{so} are in $MJ.m^{-2}.d^{-1}$. The regression gives a correlation coefficient of 0.904, with a standard error of the estimate of 0.79 $MJ\cdot m^{-2}\cdot d^{-1}$. The regression against incident short-wave radiation provides more reliable estimates of the net long-wave flux than does the traditional approach for summer conditions. It does not require the air temperature and vapour pressure as input parameters, and the clear-sky short-wave flux can be reliably modelled (see Appendix A).

The radiation algorithms presented in Appendix A were used to estimate the net radiation at the Saskatoon site. Fig. 5.2 shows the calculated daily values of net radiation plotted against the observed values for the two observation periods in 1990. The mean difference between calculated and observed values is 0.09 $MJ\cdot m^{-2}\cdot d^{-1}$, and the standard deviation of the difference is 1.44 $MJ\cdot m^{-2}\cdot d^{-1}$. These translate to equivalent evaporation rates of 0.02 mm/d and 0.5 mm/d, respectively. This represents an improvement over the approach used by Morton (1983), for example; Granger and Gray (1990a) showed that for the CRAE model the mean difference between calculated and measured daily net radiation values was 0.39 $MJ\cdot m^{-2}\cdot d^{-1}$, with a standard deviation of 2.58 $MJ\cdot m^{-2}\cdot d^{-1}$.

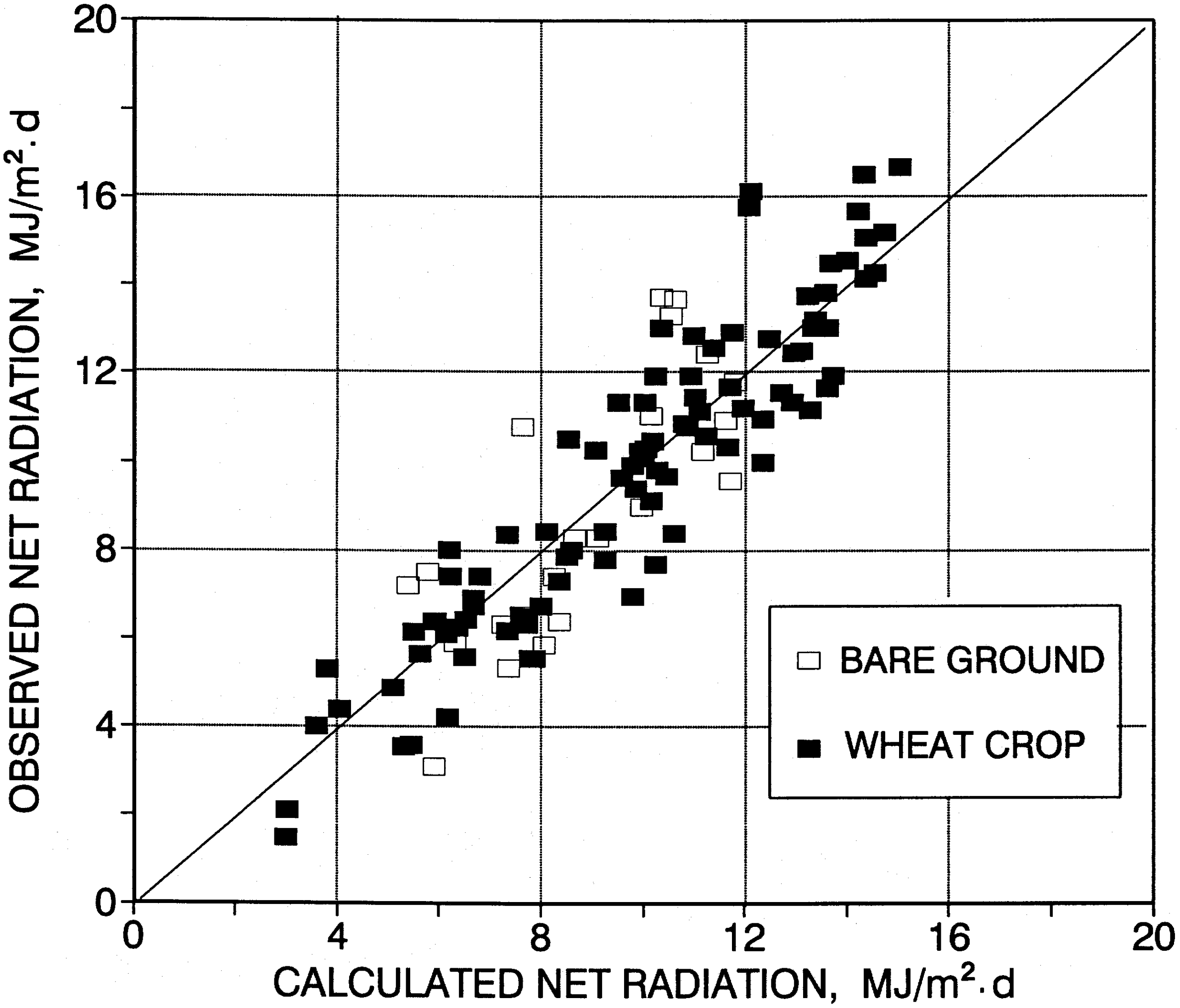

Figure 5.2 Comparison of calculated and observed daily values of the net radiation over bare soil and growing crop at Saskatoon, 1990.

5.2 Soil heat flux

From the literature it appears to be common practice to ignore the effect of the soil heat flux in schemes to model evaporation from natural surfaces; the rationale for this approach is usually based on the assumption that the soil heat flux, for daily periods, is negligible or very small compared to the other terms of the energy balance. Whereas there are situations (and time scales) where the soil heat flux is in effect very small, the assumption that it is negligible should not be made lightly. The soil heat flux term will usually be significant, even for daily periods or longer, in regions subjected to appreciable seasonal variations in temperature; most mid- to northern latitudes would fall in this category. If seasonal freezing and thawing of the soil occurs, the importance of the soil heat term is further increased. The data collected during the three observation periods also show that the soil heat term is significant. For the case of a growing crop, on average the daily soil heat flux represents approximately 9% of the net radiation input, with an observed maximum near 30%; for the case of bare ground, the corresponding average is near 20%, with a maximum near 50%.

Brutsaert (1982) presents a brief review of empirical relationships developed to estimate the soil heat flux. The general approach has been to assume that the soil heat flux is proportional to one of the other terms of the energy balance, generally the sensible heat flux or the net radiation. His review shows that the empirical constants of proportionality can vary widely. Relating the soil heat flux to only the radiation or only sensible heat term is an oversimplification since it is in fact affected by all the other terms of the energy balance.

An attempt was thus made to include the effects of both the radiation and sensible heat components on the daily total soil heat flux. In the analysis the net radiation was used as measured, and the sensible heat flux was parameterized by the change in daily mean air temperature. The following expression was tested:

$$Q_g = a + bQ_n + c\delta T \,, \qquad 5.2$$

where Q_n is the daily net radiation and δT is the change in average air temperature from the previous day; a, b and c are the regression coefficients. The coefficients b and c were found to vary with the crop height (and density). Using the data from the three observation periods, the following expression was derived for the soil heat flux:

$$Q_g = -0.065 + (0.171 + 0.18h)Q_n + (0.236 - 0.15h)\delta T \,, \qquad 5.3$$

where h is the crop height in metres; and the fluxes, Q_g and Q_n, are in $MJ/m^2 \cdot d$.

The daily soil heat flux values calculated from Eq. 5.3 are plotted against the corresponding observed values in Fig. 5.3. The mean difference between calculated and observed

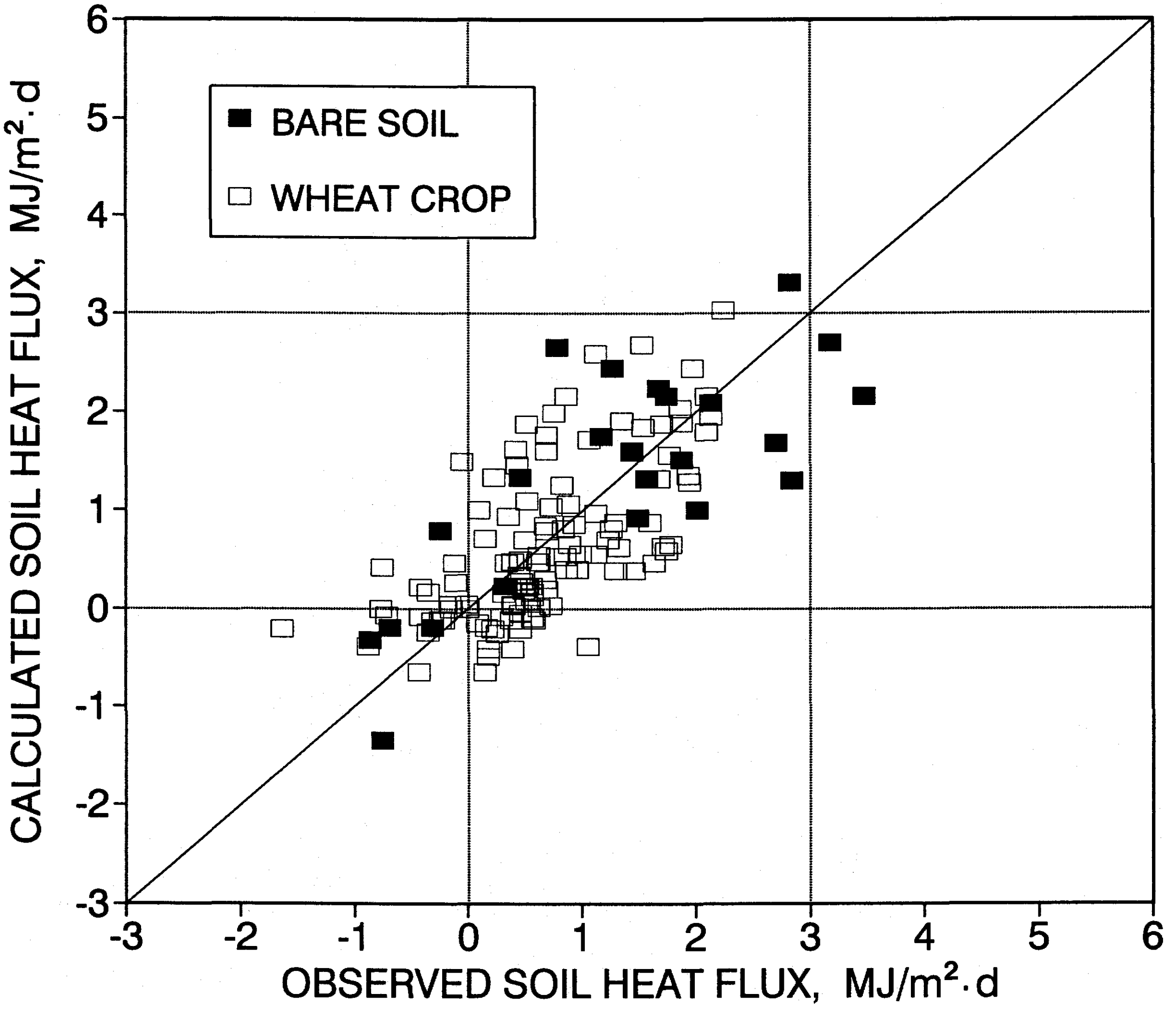

Figure 5.3 Comparison of the calculated and observed daily values of the soil heat flux term, Q_g, for bare ground and conditions of growing crop at Saskatoon.

values shown in Fig. 5.3 is -0.01 $MJ/m^2 \cdot d$, and the standard deviation of the difference is 0.67 $MJ/m^2 \cdot d$. The figure shows that the scatter is relatively large; the standard deviation of the difference between the calculated and observed values indicates that the error of estimate for this expression is still approximately 20% of the maximum flux. Although relating the soil heat term to both the net radiation and sensible heat, as with an expression such as Eq. 5.3, may represent a slight improvement over the simple regressions described by Brutsaert (1982), the results are still far from satisfactory. As well, in view of the large variability in soil and vegetative conditions, it is unlikely that the results presented here could be transposed to other areas with any reasonable degree of confidence. A great deal more work will be required before an adequate and convenient scheme for parameterizing the soil heat flux term can be established.

5.3 Use of remotely-sensed surface temperature data

The development of methods for obtaining reliable estimates of the evaporation from large-scale areas requires (i) the availability of information or data representative of the area being considered (and consequently the development of techniques for collecting such data), and (ii) the availability of algorithms or models capable of translating these "areal" data into the required regional vapour flux.

The transition from the measurement of "local" evaporation to that of the "region" is not necessarily achieved in a simple step. Brutsaert (1982) reviewed a number of studies of boundary layers over natural terrain and showed that wind, temperature and humidity measurements taken a few meters above the ground can be representative of fetches of the order of hundreds of meters, that is the field scale. Measurements at higher levels in the atmospheric boundary layer are likely to reflect surface conditions extending downwind over greater distances. Such higher elevation measurements, generally obtained from the periodic deployment of atmospheric sounding devices or with airborne sensors, are still relatively scarce, awkward and expensive; and when the region is heterogeneous, as in the case of an agricultural region made up of a patchwork of various land uses, the representativeness of these upper level measurements is subjected to an additional level of uncertainty. In the case of a heterogeneous region, Seguin (1980) points out that the total evaporation is best obtained through integration of the evaporation rates from the individual land units. This approach then requires an information network of sufficient density to cover all the individual land units; this can best be achieved through the use of remote sensing techniques.

Review of the literature indicates that essentially only two approaches are being tested as regards the application of remotely-sensed data for estimating areal evaporation. One involves a soil moisture balance technique in which evaporation is inferred from measured changes in surface soil moisture content which can be detected using thermal infrared (Price, 1977, 1980; Taconet et al., 1986) or microwave techniques (Wang, 1987; Wang et al., 1987; Autret et al., 1989). In the other method evaporation is calculated as the residual in a simple energy balance equation in which the convective heat flux is inferred from remotely-sensed surface temperature (Soer, 1980; Seguin, 1980; Brunel, 1989).

Seguin and Itier (1983) and Brunel (1989) developed linear correlations between the turbulent heat flux and the surface - air temperature gradient, and thus further simplified the energy balance to the form:

$$E = Q_n + A + B\Delta T \, , \tag{5.4}$$

where ΔT is the difference between the surface and air temperatures, and A and B are regression coefficients.

Seguin and Itier (1983) and Jackson et al. (1983) also demonstrated the practicality, at least for clear days, of using single-measurement values of the temperature gradient in the air near mid-day to represent the sensible heat flux for the entire day.

The reliability of the surface temperature measurements obtained from satellites has improved considerably. Since the publication of the Manual of Remote Sensing (American Society of Photogrammetry, 1983) some progress has been made in the area of data interpretation. Wan and Dozier (1989) show that it is possible to simultaneously make atmospheric corrections and emissivity corrections by using multiple wavelength thermal infrared data in obtaining estimates of surface temperatures of different land covers, and that accuracies of around 0.2 K can be achieved. Measurements during cloudy periods will require further developments in the field of microwave radiometry. In the mean time, for humid regions where cloud cover is frequent, Nieuwenhuis and Menenti (1986) suggest that remote sensing applications such as this will be mainly based on scanning techniques using aircraft-mounted sensors.

Both the water and energy balance approaches have their inherent limitations. The major drawback is that evaporation is calculated as the residual in the balance equation; and errors associated with the estimate of the other terms accumulate in the estimate of the residual. A procedure for the direct determination of evaporation using remotely-sensed data, or for the application of remotely-sensed data in current evaporation models, is thus a desirable objective. However, the current evaporation models (including the procedures presented in the previous chapters), invariably require as input the humidity in the air overlying the surface. Remote sensing techniques have not yet progressed to the point of providing a measure of this parameter.

However, the surface temperature may prove to be a sufficiently reliable indicator of the humidity of the air. The existence of feedback links between the surface and the overlying air has been demonstrated. This feedback is what Bouchet (1963) described when he showed that the relationship between the actual and "potential" evaporation rates (for a constant energy supply) is an inverse one. Morton (1983) states that the temperature and humidity observed in the air are a reflection of what is occurring (partitioning of energy) at the surface. It is plausible that the converse also holds; that is, that what is observed at the surface (temperature) is indicative of the conditions in the overlying air.

The surface temperature data collected during the 1990 observation periods were used to test the hypothesis that the surface measurements might be used to parameterize the conditions in the air. Fig. 5.4 shows the mean daily vapour pressure deficit of the air plotted against the saturation vapour pressure for the mean daily surface temperature as measured by the infrared sensor. The figure shows data for bare ground and for the wheat crop; the two groups of data show somewhat different relationships. The difference in the relationships for bare ground and wheat might be explained by the nature of the respective "surfaces". In the case of bare ground, the infrared sensors effectively "sees" a surface; while in the case of a vegetated cover, the sensor "sees" an apparent surface comprised of the upper portion of the vegetation, the depth of which will depend on the canopy density and the view angle of the sensor.

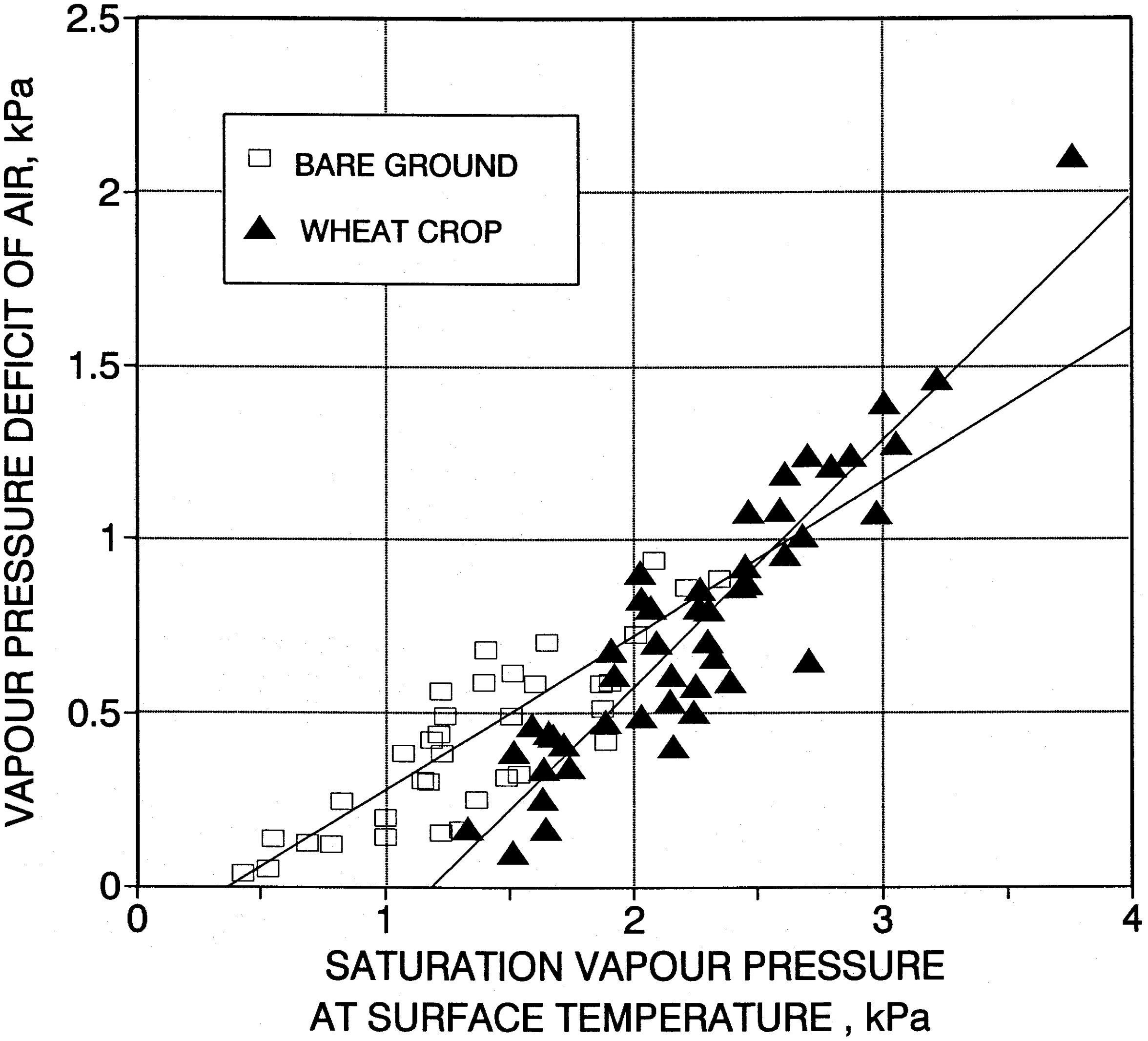

Figure 5.4 Plot of the daily vapour pressure deficit in the air against the saturation vapour pressure at the observed surface temperature, for bare ground and wheat at Saskatoon.

Straight lines were fitted through the two groups of data in Fig. 5.4. For bare ground the derived relationship between the vapour pressure deficit in the air and the saturation vapour pressure at the surface temperature is given by:

$$e_a^* - e_a = -0.170 + 0.436 e_s^* \, . \qquad 5.5$$

The regression gives a coefficient of determination, r^2 of 0.734, and a standard error of the estimate of 0.128 kPa. The corresponding regression equation for wheat is:

$$e_a^* - e_a = -0.840 + 0.706 e_s^* \, , \qquad 5.6$$

with $r^2 = 0.822$, and a standard error of the estimate equal to 0.171 kPa.

Eqs. 5.5 and 5.6 provide a means of using remotely-sensed data, which have a certain areal representativeness, in evaporation models, such as the G-D relationships. If the estimates of the vapour pressure deficit provided are reliable, this approach represents a potential improvement over the simplified energy balance approach currently being used in conjunction with remotely-sensed surface temperature data, since the evaporation would no longer be estimated as the residual in a balance equation.

5.3.1 Use of mid-day values of surface temperature.

Remote sensing, whether from polar-orbiting satellites or from aircraft, generally provides data only once daily. Routines must be established to estimate the daily mean from one-time measurements. Seguin and Itier (1983) and Jackson et al. (1983) provide regressions, for clear days, of the daily sensible heat flux against the mid-day values of the temperature gradient between the surface and the air; these were used in a simple energy balance to obtain estimates of the daily evaporation.

Using the data collected during the 1990 observation periods, the daily mean surface temperature, T_s, was correlated against the mid-day value, T_{smid} (taken here as the mean hourly value nearest solar noon). The following regression equations were obtained:

$$\textit{Bare Ground}: \quad T_s = 0.66 + 0.406\,T_{smid}\,, \qquad 5.7$$

with $r^2 = 0.714$, and a standard error of the estimate of 3.22°C; and,

$$\textit{Wheat}: \quad T_s = 5.52 + 0.457\,T_{smid}\,, \qquad 5.8$$

with $r^2 = 0.834$, and a standard error of the estimate of 1.51°C. The regression for the conditions of growing crop provides significantly better results than that for bare ground.

Because the measurements were obtained with a sensor mounted on an instrument mast, the data include days with cloud cover as well as clear days. A second regression was carried out to assess the effect of cloud cover on the relationship between the mid-day and daily mean surface temperature values. The following equations represent the best fit for the temperature data obtained with bare ground and wheat as surface conditions:

$$\text{Bare Ground}: \quad T_s = 4.36 + 0.428T_{smid} - 5.29\frac{n}{N} \qquad 5.9$$

and,

$$\text{Wheat}: \quad T_s = 5.51 + 0.465T_{smid} - 0.359\frac{n}{N}\,, \qquad 5.10$$

where n/N is the sunshine ratio, the ratio of the observed duration of bright sunshine to its maximum possible value for the day.

It is interesting to note that for wheat the effect of cloud cover on the estimate of T_s appears to be small, and the coefficients in Eq. 5.10 differ only slightly from those in Eq. 5.8. However, for the bare ground the effect of cloud cover on the relationship is much greater, as evidenced by the coefficient which is one order of magnitude larger (Eq. 5.9). The data sets used are limited and insufficient to establish the reason for the difference in importance of the cloud cover to the regression under these two conditions. One can speculate that differences in the partitioning of energy (larger soil heat flux and smaller evaporation rates for bare ground) as well as seasonal differences (the bare ground data were obtained in the Spring, following the disappearance of the annual snow cover) may account in part for this effect.

The values of the daily mean surface temperature obtained from Eqs. 5.9 and 5.10 are plotted against the observed values in Fig. 5.5.

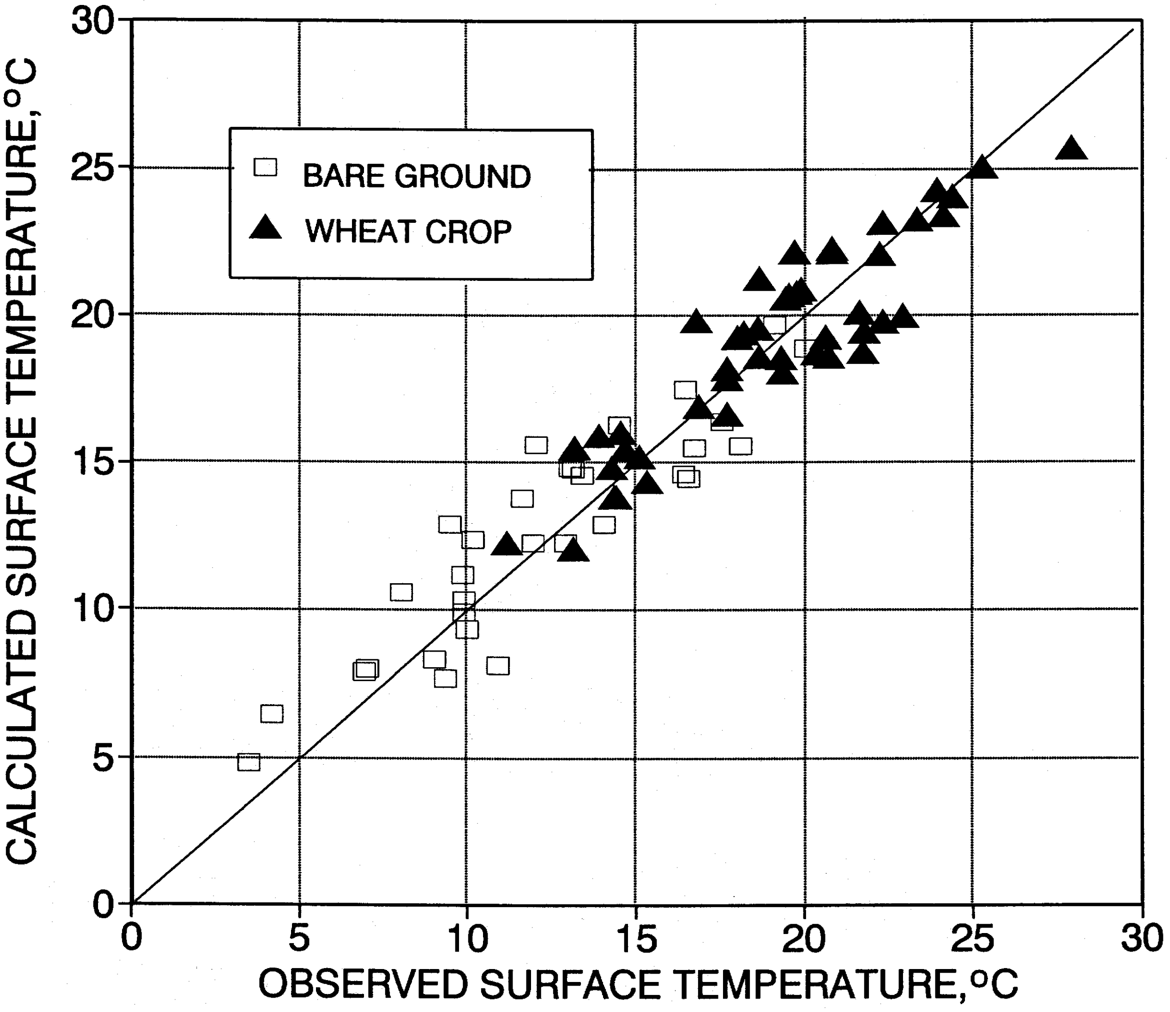

Figure 5.5 The calculated values of the mean daily surface temperature plotted against the corresponding observed values for bare ground and wheat, at Saskatoon.

Fig. 5.5 shows remarkable agreement between the calculated and observed values considering that the daily mean is estimated from a single mid-day value. The mean difference between the calculated and observed values is 0.14 °C, and the standard deviation of the difference is 1.50 °C. Thus for bare ground, comparison with the standard error of the regression for Eq. 5.7 indicates that including the effects of cloud cover results in a significant improvement in the correlation; whereas, for the growing crop conditions, the improvement is small.

The values of the daily mean surface temperature obtained from Eqs. 5.9 and 5.10 were used in Eqs. 5.5 and 5.6 to calculate the vapour pressure deficit of the air over bare ground and wheat, respectively. The calculated values of the vapour pressure deficit obtained are compared to the observed values in Fig. 5.6. The figure shows that the method can provide a good estimate of the vapour pressure deficit under the conditions encountered; the mean difference between the calculated and observed values is 0.0023 kPa, with a standard deviation of the difference of 0.147 kPa. When the observed values of the surface temperature were used to estimate the vapour pressure deficit, the agreement between the calculated and observed values was slightly better; the mean difference between calculated and observed values was 0.003 kPa, with a standard deviation of 0.131 kPa.

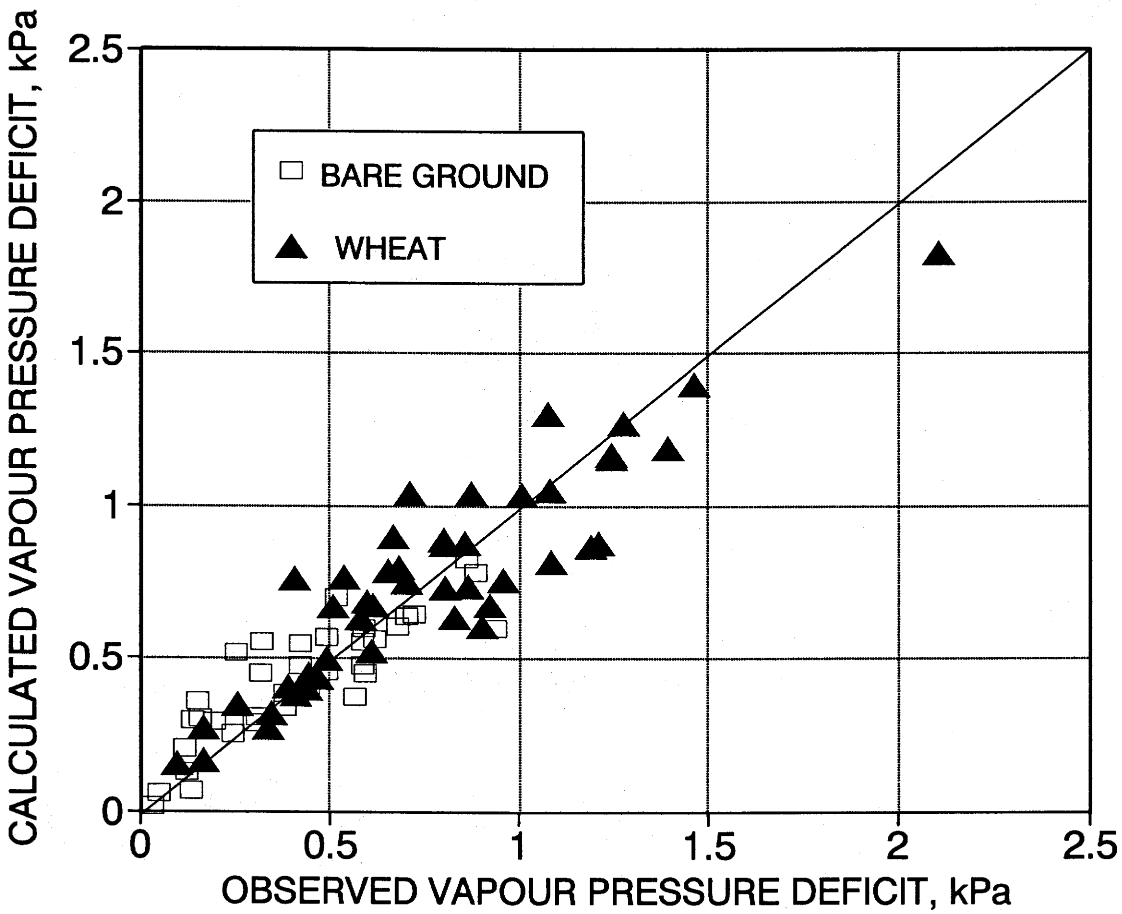

Figure 5.6 Comparison of calculated and observed values of the vapour pressure deficit in the air over bare ground and wheat at Saskatoon.

5.3.2 Application to the estimate of evaporation.

This method of obtaining the vapour pressure deficit of the air using remotely-sensed surface temperature data can be applied in a model to provide estimates of the daily evaporation rates. The following procedure was used to assess the applicability of the approach:

i) the mid-day values of surface temperature were used to estimate its daily mean value (Eqs. 5.9 and 5.10);

ii) the calculated daily mean surface temperature values were used to estimate the vapour pressure deficit in the air (Eqs. 5.5 and 5.6);

iii) the calculated values of the vapour pressure deficit were used, along with the appropriate vapour transfer function (Eqs. 4.12 or 4.13), to provide values of the drying power, E_a, and the relative drying power, D; and

iv) The relative evaporation was obtained from Eq. 4.14, and the evaporation was then calculated using the general equation (Eq. 3.10).

The calculated values of the daily evaporation thus obtained are compared to the reference values, obtained using the Bowen ratio approach, in Fig. 5.7. Although there appears to be a trend towards underestimation at the higher evaporation rates over the growing crop, the overall agreement is good; the mean difference between the calculated and observed evaporation values is 0.01 mm/d, and the standard deviation of the difference is 0.42 mm/d.

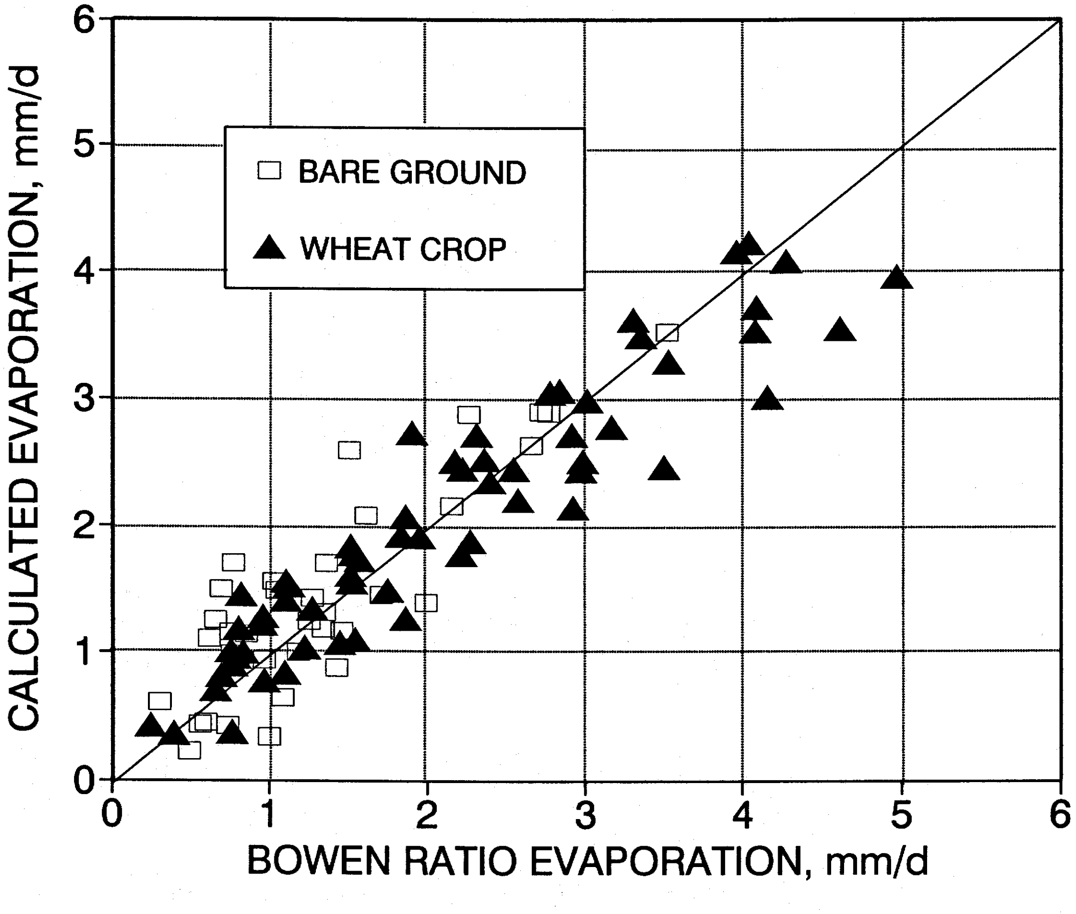

Figure 5.7 Calculated values of the daily evaporation obtained using remotely-sensed surface temperature data plotted against the reference values of evaporation obtained with the Bowen ratio approach, for bare ground and wheat at Saskatoon.

Fig. 5.7 compares favourably with Fig. 4.9, which shows the results of the verification of the G-D relationship, and where the mean difference was 0.11 mm/d and the standard deviation was 0.48 mm/d. There are, however, fewer data points shown in Fig. 5.7 because surface temperature measurements were not available during the 1989 observation period and the sensor did not function for the whole of the 1990 period.

Seguin et al. (1990) state that the simplified energy balance approach for the estimation of evaporation from remotely-sensed surface temperature measurements works satisfactorily as a first order approximation, providing an estimated error of ±1 mm/d (or about 20%). The standard error for the method described above, and demonstrated in Fig. 5.7, is less than half this value. This supports the suggestion that the approach which consists of using the relationship between surface temperature and the vapour pressure deficit in conjunction with an evaporation model, such as the G-D relationship, does represent a potentially significant improvement over the currently used methods of applying remotely-sensed data to the estimates of areal evaporation. The results warrant further testing with surface temperature data derived from satellite or aircraft-mounted sensors to confirm the applicability of the method to the large scale.

6. SUMMARY AND CONCLUSIONS

The concept of potential evaporation and its use in the development of schemes for estimating evaporation from nonsaturated surfaces ware examined.

i) It was demonstrated that the concept of potential evaporation is not limited to a single, unique potential, but that several different values are possible depending on the conditions specified in the chosen definition. A series of potential evaporation parameters are developed and classified.

ii) The nature of the relationship between actual and potential evaporation was shown to depend on the variables controlling the definition of the potential evaporation in use. (For example, a potential evaporation defined as a function of the available energy tends to show a direct relationship with actual evaporation since evaporation responds directly to the net radiation; whereas, if the atmospheric conditions are specified in the definition of the potential, the relationship between actual and potential evaporation is an inverse one because the humidity of the air is a reflection of the evaporation occurring at the surface. Thus the Penman equation, which contains both a radiation and a drying power term, shows both effects; this is why it has not been possible to establish generally useful correlations between the Penman expression and actual evaporation.)

A new approach was used to extend the Penman (1948) equation to the nonsaturated case. Rather than using stomatal or soil water resistance terms to describe the departure from saturated conditions, the development makes use of the relative evaporation, G, the ratio of the actual evaporation to the evaporation which would occur if the surface was saturated at the actual surface temperature.

iii) The Penman combination approach, with the introduction of the relative evaporation, G, results in a general equation describing evaporation from nonsaturated surfaces:

$$E = \frac{\Delta GQ}{(\Delta G + \gamma)} + \frac{\Delta GE_a}{(\Delta G + \gamma)} \qquad 3.10$$

iv) It is also shown that the development of Bouchet (1963), who first suggested an inverse relationship between actual and potential evaporation, also leads to the same general equation when consistent definitions are chosen for the potential and wet-environment evaporation rates. Thus the apparently contradictory approaches of Penman and Bouchet are reconciled.

v) An inverse relationship was developed between the relative evaporation, G, and the relative drying power, D (the ratio of the drying power to the sum of the available energy and the drying power):

$$G = \frac{1}{0.905 + 0.095e^{6.20D}} \qquad 4.14$$

vi) The G-D relationship (Eq. 4.14) appears to be single-valued. Estimates of evaporation obtained from soil moisture balances and from an experimental investigation were used to show that the relationship holds for a variety of land uses, as well as for daily and hourly time periods.

vii) The use of the G-D relationship (Eq. 4.14), along with the general evaporation equation (Eq. 3.10), represents a useful approach for estimating evaporation from natural nonsaturated surfaces. The method retains the principal advantage of the Penman equation, that is, it requires only measurements from one observation height. It presents the added advantage of not requiring a prior calculation of the potential to obtain the actual evaporation. Evaporation estimates obtained using this new G-D approach were shown to be in close agreement with independent evaporation estimates obtained using a soil moisture balance.

The use of approaches based on the Penman equation, or modifications of this equation, require a calculation of the drying power. The data obtained during the experimental investigation also allowed for the examination of the vapour transfer function used in the calculation of the drying power.

viii) It was demonstrated that, in the vapour transfer equation, and in the expression for drying power, the transfer function used must be appropriate for the time period being considered. For periods of the order of one hour, expressions such as those provided by the Thornthwaite and Holzman (1939) equation may be appropriate. Expressions for the vapour transfer function, developed using the experimental data, show that for daily periods or longer the effect of the wind speed on the vapour transfer is significantly reduced (but not eliminated as was suggested by Morton, 1963).

Most evaporation models also require that the available energy (net radiation and soil heat flux) to the surface be estimated. The experimental data allowed for the development and evaluation of appropriate algorithms.

ix) The net radiation model, developed by Granger and Gray (1990b), was shown to be a reliable predictor of the daily net radiation for the surface conditions encountered.

x) The clear-sky value of the net long-wave flux can be related directly to the incoming short-wave energy, thus eliminating the (often invalid) necessary assumption that an "ideal" atmosphere is present when the term is related to the observed temperature and humidity near the surface.

xi) In any approach based on the energy balance, the soil heat flux, or heat storage, term should not be neglected.

xii) A method for estimating the daily soil heat flux term was presented. The method accounts for the energy supply (net radiation), the partitioning of the energy at the surface (as indexed by the daily change in air temperature), and the insulating effect of the vegetative cover.

Finally, the application of remotely-sensed surface temperature data to the estimate of regional evaporation was investigated.

xiii) Using data from a mast-mounted infrared temperature sensor, a direct linear relationship was established between the vapour pressure deficit in the air and the saturation vapour pressure at the observed surface temperature; expressions were developed for bare soil and growing wheat surfaces (Eqs. 5.5 and 5.6, respectively).

xiv) The relationship between surface temperature and the vapour pressure deficit in the air allows for the application of remotely-sensed temperature data directly in evaporation models, such as CRAE or the G-D model.

xv) It was demonstrated that this new approach to the application of remote sensing may provide more reliable estimates of evaporation than does the current approach which uses remotely-sensed data to index the sensible heat flux and estimates evaporation using a simplified energy balance.

2.1 Recommendations

The following recommendations for further study are presented:

i) The G-D relationship developed here should be tested over a wider range of surfaces, such as forested land surfaces and, and in regions characterized by seasonal freezing of the soil.

ii) The effective vapour transfer function for daily time periods requires further study; the effects of surface roughness and stability must be parameterized.

iii) Expressions for estimating the daily soil heat flux term must be developed for a wide range of surface and climatic conditions.

iv) The relationship developed here between surface temperature and the vapour pressure deficit in the air must be tested for a variety of surfaces using data from aircraft- and satellite-mounted sensors.

7. REFERENCES

American Society of Photogrammetry. 1983. Manual of Remote sensing. (R.N. Colwell, Ed. in chief), 2440 pp (in two volumes).

Autret, M., R. Bernard and D. Vidal-Madjar. 1989. Theoretical study of the sensitivity of the microwave backscattering coefficient to the soil surface parameters. Int. J. Remote Sensing. 10(1): 171-179.

Barton, I.J. 1979. A parameterization of the evaporation from nonsaturated surfaces. J. Appl. Meteorol. 18: 43-47.

Black, T. A. 1979. Evapotranspiration from douglas fir stands exposed to soil water deficits. Water Resour. Res. 15(1): 164-170.

Bouchet, R.J. 1963. Evapotranspiration réelle et potentielle, signification climatique. Int. Assoc. Sci. Hydrol. Proc. Berkeley, CA Symp. Publ. 62: 134-142.

Bowen, I.S. 1926. The ratio of heat losses by conduction and by evaporation from any water surface. Phys. Rev. 27: 779-787.

Brunel, J.P. 1989. Estimation of sensible heat flux from measurements of surface radiative temperature and air temperature at two meters: application to determine actual evaporation rate. Agric. Forest Meteorol. 46: 179-191.

Brunt, D. 1932. Notes on the radiation in the atmosphere. Quart. J. Royal Meteorol. Soc. 58: 389-420.

Brutsaert, W. 1975. On a derivable formula for long-wave radiation from clear skies. Water Resour. Res. 11: 742-744.

Brutsaert, W. 1982. Evaporation into the atmosphere: theory, history, and application. D. Reidel Publishing Co., Dordrecht, Holland, 299 pp.

Brutsaert, W. and H. Stricker. 1979. An advection aridity approach to estimate actual regional evapotranspiration. Water Resour. Res. 15(2): 443-450.

Davies, J.A. and C.D. Allen. 1973. Equilibrium, potential and actual evaporation from cropped surfaces in southern Ontario. J. Appl. Meteorol. 12: 649-657.

de Vries, D.A. 1963. Thermal properties of soils. In Physics of Plant Environment (W.R. van Wijk, ed.), North-Holland, Amsterdam, 382 pp.

Federer, C. A. 1979. A soil-plant-atmosphere model for transpiration and availability of soil water. Water Resour. Res. 15(3): 555-562.

Fortin, J.P. and B. Seguin. 1975. Estimation de l'ETR régionale à partir de l'ETP locale: utilisation de la relation de Bouchet à différentes échelles de temps. Ann. Agron. 26(5): 537-554.

Garnier, B.J. and A. Ohmura. 1970. The evaluation of surface variations in solar radiation income. Solar Energy. 13: 21-34.

Garnier, B.J. and A. Ohmura. 1968. A method of calculating the direct shortwave radiation income of slopes. J. Appl. Meteorol. 7: 796-800.

Granger, R.J. 1977. Energy exchange during melt of a prairie snowcover. Unpublished M.Sc. dissertation, Univ. Saskatchewan, Saskatoon, SK.

Granger, R.J. 1989. A complementary relationship approach for evaporation from nonsaturated surfaces. J. Hydrol. 111: 31-38.

Granger, R.J. and D.M. Gray. 1989. An assessment of Morton's model for estimating areal evapotranspiration. Report No. 89003, National Hydrology Research Institute, Environment Canada, Saskatoon, SK.

Granger, R.J. and D.M. Gray. 1990a. Examination of Morton's CRAE model for estimating daily evaporation from field-sized areas. J. Hydrol. 120: 309-325.

Granger, R.J. and D.M. Gray. 1990b. A net radiation model for calculating daily snowmelt in open environments. Nordic Hydrol. 21: 217-234.

Granger, R.J., D.M. Gray and G.E. Dyck. 1984. Snowmelt infiltration to frozen Prairie soils. Can. J. Earth Sci. 21(6): 669-677.

Gray, D.M. and P.G. Landine. 1988. An energy budget snowmelt model for the Canadian Prairies. Can. J. Earth Sci.

Gray, D.M., G.A. McKay and J.M. Wigham. 1970. Energy, evaporation and evapotranspiration. In Handbook on the Principles of Hydrology (D.M. Gray, ed.) National Research Council of Canada, Ottawa, ON.

Gray, D.M., R.J. Granger and G.E. Dyck. 1985. Overwinter soil moisture changes. Trans. ASAE. 28(2): 442-447.

Jackson, R.D., R.L. Hatfield, R.J. Reginato, S.B. Idso and P.J. Pinter. 1983. Estimation of daily evapotranspiration from one time-of-day measurements. Agric. Water Manage. 7: 351-362.

Jame, Y.W. and D.I. Norum. 1980. Heat and mass transfer in freezing unsaturated porous medium. Water Resour. Res. 16: 811-817.

Kohler, M.A. and L.H. Parmele. 1967. Generalized estimates of free-water evaporation. Water Resour. Res. 3(4): 997-1005.

Konstantinov, A.R. 1963. Evaporation in Nature. Israel program for Sci. Translations, 1966, 523 pp.

Kustas, W.P., B.J. Choudhury, M.S. Moran, R.J. Reginato, R.G. Jackson, L.W. Gay and H.L. Weaver. 1989. Determination of sensible heat flux over sparse canopy using thermal infrared data. Agric. Forest Meteorol. 44: 197-216.

LeDrew, E.F. 1979. A diagnostic examination of the complementary relationship between actual and potential evapotranspiration. J. Appl. Meteorol. 18: 495-501.

List, R.J. 1968. Smithsonian Meteorological Tables, 6th edition. The Smithsonian Institute, Washington, DC.

Monteith, J.L. 1965. Evaporation and environment. In The State and Movement of Water in Living Organisms. Proc. 19th Symp., Soc. Exp. Biol., Cambridge University Press, London, England, 205-233.

Morton, F.I. 1969. Potential evaporation as a manifestation of regional evaporation. Water Resour. Res. 5(6): 1244-1255.

Morton, F.I. 1983. Operational estimates of areal evapotranspiration and their significance to the science and practice of hydrology. J. Hydrol. 66: 1-76.

Morton, F.I., F. Ricard and S Fogarasi. 1985. Operational estimates of areal evapotranspiration and lake evaporation - program WREVAP. NHRI paper No. 24, National Hydrology Research Institute, Inland Waters Directorate, Ottawa, ON.

Nieuwenhuis, G.J.A. and M. Menenti. 1986. Application of thermal infrared remote sensing in water management of humid and arid areas. Geocarto International. 1: 35-46.

Nieuwenhuis, G.J.A., E.H. Smidt and H.A. Thunnissen. 1985. Estimation of regional evapotranspiration of arable crops from thermal infrared images. Int. J. Remote Sensing. 6(8): 1319-1334.

Obukhov, A.M. 1946. Turbulence in an atmosphere with a non-uniform temperature. Trudy Instituta Teoreticheskio Geofiziki, AN-SSSR No. 1 (Boundary-Layer Meteorology, 1971, 2: 7-29)

Ohmura, A. 1982. Objective criteria for rejecting data for Bowen ratio flux calculations. J. Appl. Meteorol. 21: 595-598.

Penman, H.L. 1948. Natural evaporation from open water, bare soil and grass. Proc. R. Soc. London, Ser.A, 193: 120-145.

Penman, H.L. 1956. Evaporation: an introductory survey. Neth. J. Agric. Sci. 4: 9-29.

Penman, H.L. and R.K. Schofield. 1951. Some physical aspects of assimilation and transpiration. In Carbon Dioxide Fixation and Photosynthesis, Symposia of the Society for Experimental Biology, (5): 115-129.

Perrier. A. 1977. Projet de définitions concernant l'évapotranspiration en fonction de considérations théoriques et pratiques. Météorologie 11: 7-16.

Price, J.C. 1977. Thermal mapping: A new view of the earth. J. Geophys. Res. 82: 2582-2590.

Price, J.C. 1980. The potential of remotely sensed thermal infrared data to infer surface soil moisture and evaporation. Water Resour. Res. 16(4): 787-795.

Priestley, C.H.B. 1959. Turbulent transfer in the lower atmosphere. University of Chicago Press, Chicago, IL, 130 pp.

Priestley, C.H.B. and R.J. Taylor. 1972. On the assessment of surface heat flux and evaporation using large-scale parameters. Monthly Weather Rev. 100(2): 81-92.

Reginato, R.J. and R.D. Jackson. 1971. Field measurement of soil-water content by gamma-transmission compensated for temperature fluctuations. Soil Sci. Soc. Am. Proc. 35: 529-533.

Satterlund, D.R. 1979. An improved equation for estimating long-wave radiation from the atmosphere. Water Resour. Res. 15: 1643-1650.

Seguin, B. 1975. Influence de l'évapotranspiration régionale sur la mesure locale d'évapotranspiration potentielle. Agric. Meteorol. 15: 355-370.

Seguin, B. 1980. Détermination de l'évaporation réelle dans les bilans hydrologiques par la télédétection en thermographie infra-rouge. Bull. Sci. Hydrol. 25(2): 143-153.

Seguin, B. and B. Itier. 1983. Using midday surface temperature to estimate daily evaporation from satellite thermal IR data. Int. J. Remote Sensing. 4(2): 371-383.

Seguin, B., M. Savane and B. Guillot. 1990. Estimation of evaporation from thermal infrared meteorological satellite data: a case study with Meteosat and NOAA for France. In Proc. Int. Symp. on Remote Sensing and Water Resources, Enschede, The Netherlands, August 20-24, 1990: 215-227.

Slatyer, R.O. and I.C. McIlroy. 1961. Practical Microclimatology. Commonwealth Scientific IRO, Melbourne, Australia, 310 pp.

Smith, E.M., T.H. Taylor and S.W. Smith. 1967. Soil moisture measurement using gamma transmission techniques. Transactions Am. Soc. Agric. Eng. 10(2): 205-208.

Soer, G.J.R. 1980. Estimation of regional evapotranspiration and soil moisture conditions using remotely sensed crop surface temperatures. Remote Sensing of Environment. 9: 27-45.

Taconet, O., R. Bernard and D. Vidal-Madjar. 1986. Evapotranspiration over an agricultural region using a surface flux/temperature model based on NOAA-AVHRR data. J. Climate Appl. Meteor. 25: 284-307.

Tan, C.S. and T.A. Black. 1976. Factors affecting the canopy resistance of a Douglas-fir forest. Boundary-Layer Meteorol. 10:475-488.

Thornthwaite, C.W. 1948. An approach toward a rational classification of climate. Geograph. Rev. 38: 55-94.

Thornthwaite, C.W. and B. Holzman. 1939. The determination of evaporation from land and water surfaces. Mo. Wea. Rev. 67: 4-11.

Troxler. undated. Operation and maintenance manual, two probe density gauge. Troxler Electronics Laboratories, Raleigh, NC.

van Bavel, C.H.M. 1966. Potential evaporation: The combination concept and its experimental verification. Water Resour. Res. 2(3): 455-467.

van Bavel, C.H.M. 1967. Changes in canopy resistance to water loss from alfalfa induced by soil water depletion. Agric. Meteorol. 4:165-176.

Vidal, A. and A. Perrier. 1989. Analysis of a simplified relation for estimating daily evapotranspiration from satellite thermal IR data. Int. J. Remote Sensing. 10(8): 1327-1337.

Wan, Z. and J. Dozier. 1989. Land-surface temperature measurement from space: Physical principles and inverse modeling. IEEE Trans. Geosci. Remote Sensing. 27(3): 268-277.

Wang, J.R. 1987. Microwave emission from smooth bare fields and soil moisture sampling depth. IEEE Trans. Geosci. Remote Sensing. GE-25(5): 616-622.

Wang, J.R., E.T. Engman, T. Mo, T.J. Schmugge and J.C. Shiue. 1987. The effects of soil moisture, surface roughness, and vegetation on L-band emission and backscatter. IEEE Trans. Geosci. Remote Sensing. GE-25(6): 825-832.

World Meteorological Organization. 1974. International glossary of hydrology. WMO report No.385, Geneva, Switzerland, 393 pp.

APPENDIX A

ALGORITHMS FOR ESTIMATING NET RADIATION

Net radiation, Q_n, is the sum of the net short-wave, Q_{sn}, and the net long-wave, Q_{ln}, fluxes:

$$Q_n = Q_{sn} + Q_{ln} \qquad \text{A1}$$

Net short-wave radiation is equal to the incident short-wave flux, Q_s, less the amount reflected by the surface, Q_r:

$$Q_{sn} = Q_s - Q_r \qquad \text{A2}$$

The incident flux, Q_s, is expressed as the sum of its direct beam, Q_{drs}, and diffuse, Q_{dfs}, components:

$$Q_s = Q_{drs} + Q_{dfs} \qquad \text{A3}$$

The direct beam component of solar radiation incident to a horizontal plane on the earth's surface under clear skies, Q_{dro}, is estimated by the expression proposed by Garnier and Ohmura (1970):

$$Q_{dro} = \frac{I_o}{r^2} \int p^m \cos(X \wedge S)\, dH \qquad \text{A4}$$

where I_o is the solar constant; r is the radius vector of the earth's orbit; p is the mean transmissivity of the atmosphere along the zenith path; m is the optical air mass, the ratio of the distance the sun's rays travel through the atmosphere to the depth of the atmosphere along the zenith path; cos(X^S) is the cosine of the angle of incidence of the sun's rays on a slope; and H is the hour angle measured from solar noon. The integral is taken over the duration of sunlight on the surface.

Values of I_o, r and m are given by List (1968). Based on the data presented by List (1968) the following expression is used to calculate the radius vector of the earth's orbit:

$$r = 0.01676 \cos[\pi - .017262\,(DAY - 3)\,] + 1 \qquad \text{A5}$$

Garnier and Ohmura (1968) provide the following secant approximation for the optical air mass:

$$m = (P/P_o)\ \sec(X \wedge S) = (P/P_o)/\cos(X \wedge S) \qquad \text{A6}$$

where P is the atmospheric pressure, and P_o is the standard sea-level pressure.

The seasonal variation in the transmissivity is described by the expression:

$$p = 0.818 - 0.064 \sin[(DAY - 90)\ 2\pi/365] \qquad \text{A7}$$

in which DAY is the Julian day number.

In the presence of cloud cover, the direct-beam radiation, Q_{drs}, is attenuated; the following expression was derived by regression:

$$Q_{drs} = Q_{dro}[0.024 + 0.974\ (\frac{n}{N})^{1.35}] \qquad \text{A8}$$

in which n/N, the sunshine ratio, is the ratio of the observed hours of bright sunshine, n, over the maximum possible duration of bright sunshine. The mean difference between calculated and observed Q_{drs} values and the standard deviation of the difference were respectively, 0.12 $MJ/m^2 \cdot d$ and 1.84 $MJ/m^2 \cdot d$ for Bad Lake; -0.08 $MJ/m^2 \cdot d$ and 1.76 $MJ/m^2 \cdot d$ for Edmonton and 0.32 $MJ/m^2 \cdot d$ and 1.64 $MJ/m^2 \cdot d$ for Winnipeg.

The daily clear-sky diffuse radiation was found to exhibit a seasonal variation described by the following expression:

$$Q_{dro} = 3.5\left(\frac{P}{P_o}\right)\cos(X \wedge S)/p + 0.45\sin[(172 - DAY)2\pi/365] \qquad \text{A9}$$

In the presence of cloud cover the diffuse radiation, Q_{dfs}, is described by the quadratic:

$$\frac{Q_{dfs}}{Q_{dfo}} = 2.68 + 2.2\left(\frac{n}{N}\right) - 3.85\left(\frac{n}{N}\right)^2 \qquad \text{A10}$$

Comparisons between the calculated and observed values of Q_{dfs} showed a mean and standard deviation of the difference of -0.15 $MJ/m^2 \cdot d$ and 1.30 $MJ/m^2 \cdot d$ for Bad Lake; -0.08 $MJ/m^2 \cdot d$ and 1.24 $MJ/m^2 \cdot d$ for Edmonton; and 0.2 $MJ/m^2 \cdot d$ and 1.32 $MJ/m^2 \cdot d$ for Winnipeg, respectively.

The albedo, a, of the surface at Bad Lake was observed to increase gradually, during the snow-free season; the seasonal trend was approximated linearly by:

$$a = 0.16 + 0.0001224DAY \qquad \text{A11}$$

The clear-sky net long-wave radiation was regressed against the clear-sky incident short-wave flux:

$$Q_{lno} = -4.25 - 0.24\ Q_{so} \qquad \text{A12}$$

In the presence of clouds, the net long-wave radiation, Q_{ln}, is defined by the equation:

$$\frac{Q_{ln}}{Q_{lno}} = 0.25 + 0.75\left(\frac{n}{N}\right) \qquad \text{A13}$$

A comparison of the calculated and observed Q_{ln} values for the Bad Lake data gave a mean difference of 0.16 $MJ/m^2 \cdot d$, and a standard deviation of the difference of 1.25 $MJ/m^2 \cdot d$.

APPENDIX B

TABLES OF EXPERIMENTAL DATA

Table B-1 Data from the 1989 observation period at Saskatoon (growing wheat).

Julian Day	Air Temp. °C	Vapour Pres. kPa	Net Rad. MJ/m^2	Soil Heat MJ/m^2	Wind Speed m/s	Evaporation (Bowen ratio) mm/d
190	17.60	1.319	11.29	-0.40	2.86	3.200
191	15.25	1.549	4.80	0.15	2.58	1.221
192	16.79	1.695	8.73	-0.10	2.18	2.143
193	14.48	1.407	8.65	0.01	1.90	1.666
194	17.11	1.511	15.86	-0.52	1.87	3.758
195	19.96	1.665	15.85	-0.73	2.35	3.334
196	21.72	1.605	16.85	-0.69	2.46	3.403
197	22.42	1.393	15.19	-0.47	3.19	3.428
198	20.72	1.774	13.17	-0.40	1.97	2.885
199	20.84	1.745	16.22	-0.43	1.22	3.061
200	23.62	1.479	18.14	-0.62	1.74	3.181
201	25.35	1.497	14.42	-0.49	1.77	3.201
202	27.55	1.631	14.70	-0.68	2.73	2.723
203	22.10	1.430	16.13	-0.18	2.75	2.855
204	20.51	0.951	17.28	-0.30	1.92	2.658
205	24.45	1.314	15.43	-0.62	2.94	1.980
206	17.88	1.241	10.50	-0.16	2.77	1.500
207	18.71	1.289	16.71	-0.57	1.53	2.306
208	17.76	1.317	12.78	-0.46	3.06	1.424
209	17.88	1.422	8.78	-0.41	2.10	1.201
210	18.92	1.409	17.26	-0.70	1.25	2.437
211	23.42	1.331	14.61	-0.79	1.37	1.969
212	22.90	1.746	14.05	-0.62	2.68	3.215

Table B-2 Data from the 1990 observation period at Saskatoon (bare ground).

Julian Day	Air Temp. °C	Vapour Pres. kPa	Net Rad. MJ/m^2	Soil Heat MJ/m^2	Wind Speed m/s	Evaporation (Bowen ratio) mm/d
117	-2.40	0.378	7.48	0.88	4.75	0.655
118	-4.06	0.386	7.17	0.23	6.17	1.274
119	-4.85	0.371	0.22	-0.30	6.81	0.701
120	-1.68	0.395	3.70	0.47	4.31	1.136
121	1.91	0.454	13.28	1.00	2.26	1.504
122	5.72	0.531	10.74	3.18	1.81	1.992
123	9.02	0.765	8.25	1.75	2.97	1.459
124	10.10	0.918	11.01	2.15	2.18	1.362
125	15.33	0.746	12.39	2.84	3.48	1.401
126	17.69	0.767	6.30	1.45	5.02	0.901
127	6.78	0.734	7.39	0.75	4.65	0.849
128	1.38	0.554	6.51	0.69	5.27	1.248
129	5.36	0.450	10.21	1.78	3.07	1.031
130	6.98	0.575	8.24	1.17	3.70	1.179
131	4.21	0.331	11.76	2.32	2.37	0.777
132	7.36	0.434	8.96	1.69	2.31	0.618
133	7.76	0.352	9.53	2.71	1.81	0.967
134	9.71	0.523	6.35	1.89	3.61	0.303
135	7.35	0.827	1.64	-0.86	4.13	0.999
136	7.96	0.766	13.64	2.51	3.46	2.659
137	9.36	0.559	10.88	3.48	2.47	1.343
138	9.92	0.656	5.33	2.02	4.96	0.600
139	6.77	0.844	5.88	0.34	5.27	1.425
140	8.25	0.929	5.83	2.85	2.36	0.557
141	10.12	1.084	3.10	1.50	2.78	0.493
142	12.25	1.002	13.70	2.34	1.45	1.713

Table B-3 Data from the 1990 observation period at Saskatoon (growing wheat).

Julian Day	Air Temp. °C	Vapour Pres. kPa	Net Rad. MJ/m^2	Soil Heat MJ/m^2	Wind Speed m/s	Evaporation (Bowen ratio) mm/d
149	20.70	0.938	10.27	1.94	7.13	0.748
150	18.52	1.182	3.59	0.73	6.64	0.738
151	19.24	1.328	7.29	1.96	3.17	0.469
152	16.35	1.128	6.90	0.87	4.57	0.737
153	8.26	0.782	10.46	-1.63	5.35	3.150
154	10.61	0.692	12.85	1.13	2.51	2.266
155	12.77	0.979	6.49	-0.05	4.26	1.470
156	11.99	1.077	9.66	0.24	4.05	2.327
157	14.68	1.081	13.03	1.53	3.36	2.868
158	14.18	1.094	7.78	0.89	1.47	1.416
159	15.12	1.124	9.10	0.43	2.34	1.878
160	17.57	1.139	15.79	2.25	1.74	4.215
161	20.42	1.332	6.17	1.78	5.00	1.665
162	17.21	1.008	16.13	0.70	7.31	5.119
163	13.85	0.873	9.81	0.15	5.68	1.863
164	10.25	0.995	5.58	-0.76	4.16	1.088
165	12.96	0.786	12.49	1.98	2.20	1.930
166	15.31	0.768	13.18	1.97	3.49	1.775
167	17.03	1.118	10.97	2.15	2.68	1.678
168	19.53	1.274	11.36	2.10	2.84	2.290
169	17.37	1.157	10.32	1.13	1.82	2.716
170	14.95	1.293	6.95	-0.75	2.95	2.704
171	16.49	1.042	11.67	0.52	2.23	3.523
172	17.12	1.036	11.16	0.70	2.99	3.201
173	19.28	1.124	11.97	1.88	2.35	3.061
174	22.18	1.235	9.64	1.55	4.54	2.901
175	22.16	1.435	13.76	2.10	2.74	3.927
176	22.07	1.314	14.50	1.71	3.11	4.417
177	23.29	1.346	13.02	1.89	3.18	3.870
178	19.56	1.593	9.99	0.34	3.39	3.036
179	17.59	1.466	11.21	0.35	2.05	3.142
180	18.44	1.530	15.67	0.77	2.89	4.268
181	22.05	1.433	13.01	0.88	4.05	4.156
182	22.78	1.526	14.55	1.07	3.43	4.613
183	21.26	1.544	10.85	0.68	4.12	3.506
184	15.43	1.403	9.90	-0.71	7.40	3.040
185	10.14	1.070	6.11	-0.87	3.22	1.575
186	13.57	1.212	9.39	0.44	1.73	2.428
187	13.24	1.422	1.64	-0.16	3.37	0.761
188	14.15	1.443	6.73	0.51	2.21	1.515
189	17.28	1.354	16.70	1.36	2.25	4.387
190	20.03	1.459	15.10	1.46	2.32	4.311
191	17.78	1.618	8.39	0.53	2.08	2.517

Table B-3 (cont.)

192	19.03	1.679	16.53	1.68	1.03	3.961
193	21.13	1.559	15.18	0.84	3.69	4.975
194	21.35	1.431	14.27	0.93	3.14	4.075
195	17.53	1.092	12.76	0.32	3.08	3.517
196	17.04	1.106	14.16	0.90	3.10	3.562
197	14.95	1.285	4.19	-0.30	1.84	1.430
198	13.93	1.332	7.69	-0.35	2.44	2.412
199	13.29	1.058	13.82	-0.11	2.95	3.562
200	12.24	1.033	10.56	-0.10	1.68	2.627
201	13.16	1.073	11.94	0.63	1.59	2.155
202	16.54	1.171	11.34	1.26	1.25	2.593
203	18.62	1.277	11.55	1.33	1.94	2.778
204	20.35	1.128	12.47	1.75	1.76	2.919
205	21.38	1.278	11.70	1.65	2.24	2.693
206	22.12	1.556	10.78	1.46	2.71	2.432
207	20.53	1.743	10.25	0.52	1.91	2.501
208	17.69	1.866	4.89	-0.35	2.34	1.850
209	17.17	1.458	11.15	-0.41	3.86	3.518
210	13.74	1.124	11.44	-0.21	1.16	2.212
211	17.81	1.231	12.91	0.85	2.10	3.088
212	22.71	1.334	12.58	1.60	2.42	2.999
213	19.67	1.679	5.59	0.27	2.25	1.544
214	17.65	1.205	5.56	0.09	2.31	1.487
215	14.48	0.964	11.96	0.57	2.00	1.904
216	18.78	1.073	10.24	1.24	1.92	1.956
217	22.25	1.190	11.33	1.72	2.52	2.034
218	25.99	1.173	10.48	1.79	4.31	1.781
219	20.90	1.176	10.09	1.05	2.65	1.732
220	16.07	1.016	7.86	0.40	2.09	1.284
221	15.63	1.061	8.42	0.57	1.39	1.063
222	19.02	1.180	8.34	0.99	2.05	0.841
223	17.10	1.070	6.22	0.59	1.37	0.889
224	19.31	1.171	7.99	0.97	1.77	1.562
225	16.64	1.227	6.30	0.16	2.52	2.053
226	13.76	1.095	6.71	0.22	1.21	2.064
227	20.00	1.264	8.39	1.30	3.45	0.958
228	14.03	1.190	3.52	-0.42	2.99	1.319
229	13.59	1.196	3.99	0.01	2.58	1.094
230	14.48	1.146	5.31	0.50	2.94	0.782
231	15.86	1.188	4.40	0.52	1.59	0.660
232	19.72	1.195	6.40	1.14	1.69	0.769
233	22.18	1.325	6.40	1.31	2.37	0.692
234	19.40	1.219	6.43	0.47	3.53	0.754
235	17.17	1.163	7.39	0.39	2.88	1.102
236	12.76	1.035	5.63	0.18	1.78	0.950
237	11.80	1.073	2.11	-0.23	1.84	0.244
238	11.26	1.122	1.49	-0.41	2.07	0.395
239	12.83	1.050	8.01	0.45	2.08	1.516
240	13.99	1.100	6.13	0.56	2.80	1.189
241	17.35	1.226	7.41	0.85	2.47	1.268