INLAND WATERS BRANCH

EVAPORATION AND CLIMATE
A study in cause and effect

F. I. MORTON

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DEPARTMENT OF ENERGY,
MINES AND RESOURCES

Evaporation and climate

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Preface

Evaporation from the soil and vegetation surfaces of a region depends on the availability of both water and energy. Water supply variations, by changing the amount of energy used for evaporation and the amount of heat and vapour convected to the air, have substantial effects on the weather and hence on the energy available for evaporation. As the evaporation also has an effect on the water supply there is a problem of feedback and of distinguishing cause from effect. Consideration of energy and vapour transfers suggests that the regional evaporation is equal to the absorbed insolation less the evaporation from a small continuously moist surface, i.e., to the absorbed insolation less the potential evaporation, and that the latter quantity can be derived from the most appropriate of several well known climatological techniques. Such a model permits the regional evaporation, a product of climatic, soil moisture and vegetative processes, to be estimated by its effects on evaporimeter or weather observations.

The model is developed and its implications with regard to climate and the evaporation problem discussed. Differences between annual rainfall and annual runoff for river catchments in Ireland provide reasonable confirmation of the model when compared with predicted values. Analysis of records of evaporation pans and grass evaporimeters located in various parts of Ireland provides further verification and a deeper insight into the implications of the model.

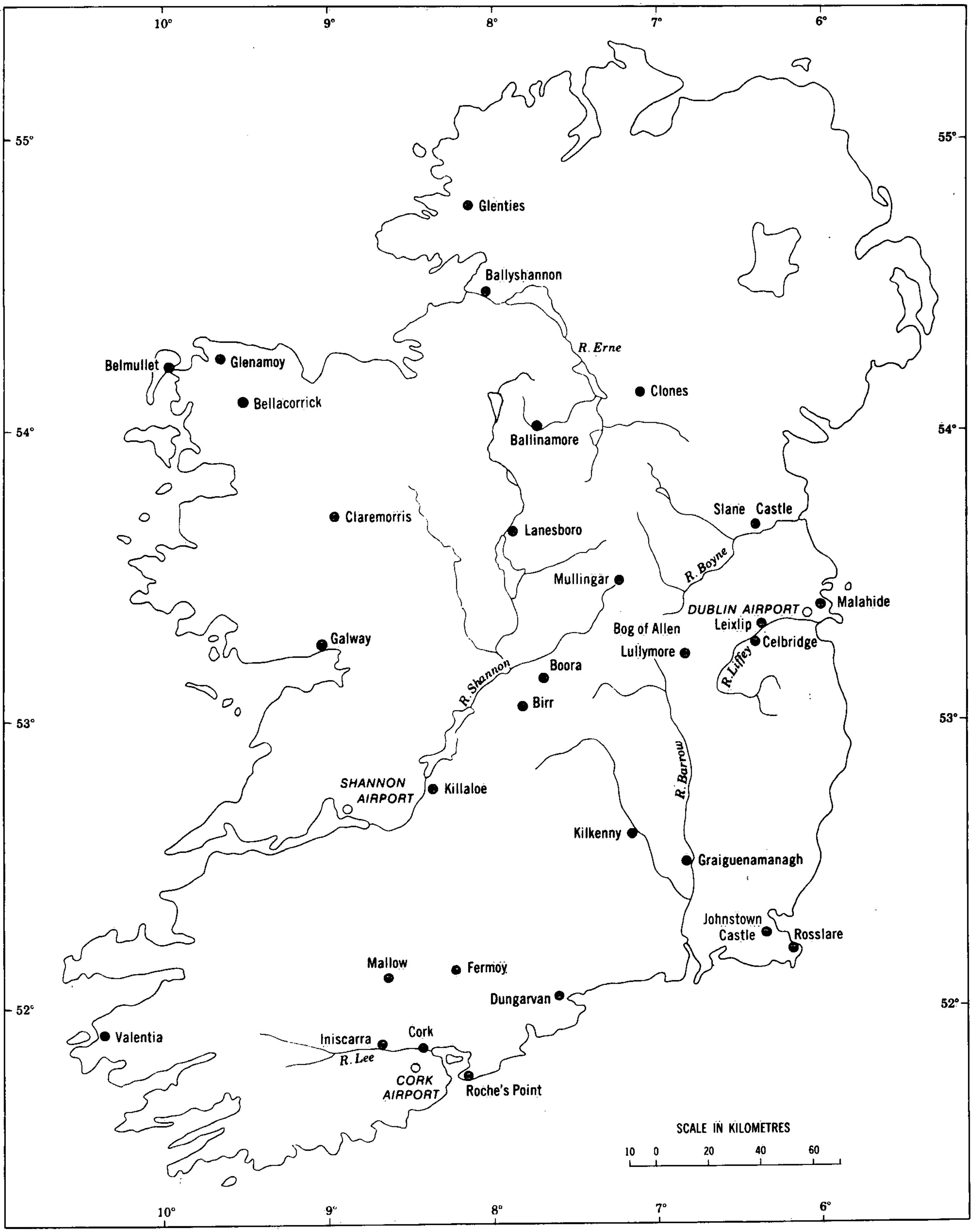

Figure 1 Location Map - Ireland

Introduction

Evaporation from the soil and vegetation surfaces of a region depends on the availability of both water and energy. Water from rain or snow becomes available for evaporation on a significant scale only if it is retained in storage, i.e., if it does not flow immediately into a river system. The stored water may be on the surface in snow cover, swamps or lakes, or below the surface, as soil moisture and ground water. Although water in sub-surface storage requires some soil or vegetative process such as capillarity or transpiration to bring it to the surface again, it is usually the most important component of water supply for evaporation from a land region. The amount of sub-surface water that can be made available for evaporation depends on the amount of water in storage and the withdrawal processes, and these are influenced by the evaporation itself in addition to the pluvial, vegetal, and geological characteristics of the region. The energy required to transform water into vapour in the evaporation process is derived from the solar radiation. However, the proportion of the energy from solar radiation that is available for evaporation depends on the vapour content and heat content of the lower atmosphere, and these in turn are appreciably influenced by the evaporation. Thus the process of evaporation is subject to feedback at all stages, as it has significant effects on its own causes, i.e., on the supply of energy and the supply of water. This complicates the problem of analyzing cause and effect in the evaporation phenomenon, a problem which is of great importance in climatological studies and in assessing the environmental modifications associated with such human activities as irrigation, land drainage, creation of lakes, afforestation, urbanization, atmospheric pollution and rainfall stimulation.

Two techniques have been developed for estimating evaporation from its effects on the lower atmosphere. The eddy correlation approach uses an instrument that automatically measures the covariance of specific humidity and vertical wind speed. Such an instrument, named the evapotron, has been described by Dyer and Maher (1965). Dyer (1967) states that the instrument may be regarded as having an accuracy within 10 per cent. The aerodynamic approach to estimating evaporation depends on carefully measured profiles of humidity, wind speed and temperature, and on the assumption of similarity between transfers of momentum, water vapour and heat. Both the eddy correlation and aerodynamic approaches have theoretical and instrumental difficulties which must be overcome before they can be widely used with any confidence. Furthermore, they are applicable only to surfaces that are homogeneous with regard to height and the availability of both energy and water.

The energy balance and vapour transfer approaches are used to provide estimates of evaporation from its causes. In the former approach, the evaporation is the unknown quantity in the energy conservation equation for the surface whereas, in the latter, the evaporation is a function of the wind speed and the difference between the vapour pressure of the surface and that of the air. Unfortunately both approaches require knowledge of the temperature and vapour pressure at the evaporating surface and this is exceedingly difficult, if not impossible to obtain. However, both Penman (1948) and Ferguson (1952) formulated combinations of the energy balance and vapour transfer equations which eliminate the need for surface temperature observations in estimating the evaporation from a continuously moist surface when sub-surface heat storage changes are insignificant. The two combinations appear to be different but are based on similar assumptions and provide similar results. As they permit the evaporation to be estimated from routine weather observations, the methods proposed by Penman and Ferguson are referred to as the climatological approach.

When the surface is not continuously moist the problem is more complex. Lysimeters, with vegetation and soil moisture similar to those of their surroundings, are used in the hope that the observed evaporation accurately reflects the causes of evaporation from the surrounding area. Their use is limited to micrometeorological research because they are expensive to install, difficult to operate and representative of only a small area. The generally accepted alternative is to assume that the evaporation from a continuously moist surface, as derived from weather observations using the climatologic approach, or from evaporimeter observations, is the energy available for evaporation. With this assumption the evaporation from the continuously moist surface accurately reflects one of the causes of evaporation and is referred to as the potential evaporation. The other cause of evaporation, the availability of water, is estimated from a trial-and-error soil moisture budget, with the ratio of actual evaporation to potential evaporation being some assumed function of the soil moisture content. Sellers (1965) has provided a detailed example of the application of one such technique. Although it is questionable whether soil moisture by itself is an adequate index of water availability, the most important objection to such an approach is the assumption that the potential evaporation is the energy available for evaporation. This assumption is true only for very small areas as exemplified by evaporimeters and lysimeters. It is not valid when the area is sufficiently large for the convected evaporation and sensible heat transfer to have a significant influence on the vapour and heat content of the overpassing air. A decrease in the availability of water to the surfaces of such an area and the resultant increase in the heat content and vapour pressure deficit of the lower atmosphere, would result in an increase in the potential evaporation. When regarded in this light the potential evaporation is an effect rather than a cause of the actual evaporation.

An approach more consistent with reality has been suggested by Bouchet (1963). It is based on a distinction between regional evaporation and potential evaporation; the former from an area so large that convection from the water, soil, and vegetation surfaces effaces advection from adjacent regions as a factor governing the heat and vapour content of the overpassing air, and the latter, from a continuously moist surface so small that advection effaces convection. Consideration of the changes in the heat and vapour content of the lower atmosphere associated with changes in regional water supply suggests that the sum of the potential evaporation and regional evaporation remains constant under conditions of changing water availability, and that the sum is equal to the evaporation equivalent of the absorbed insolation. Morton (1965, 1967b) developed this concept into a model in which the potential evaporation is estimated from routine weather observations using the climatological approach. This takes advantage of the large scale turbulence and movement of the lower atmosphere which makes the weather observations representative of the heat and vapour content of a large area. With such a model the end results of complex climatic, soil moisture, and vegetative processes on regional evaporation may be estimated by their effects on weather observations. The potential evaporation also may be assumed equal to evaporimeter observations provided that the evaporimeter has insolation absorption, vapour transfer, and heat transfer characteristics similar to those of the region.

The concept and model provide a novel and potentially valuable approach to various aspects of meteorology, hydrology, and engineering. Amongst other applications they provide a coherent frame of reference for design and evaluation of evaporation experiments; a basis for assessing environmental modifications associated with cultural changes, for estimating irrigation water requirements and for classifying climatic humidity, and a technique for estimating river and ground water yields from weather observations. However, they are based on assumptions which need to be tested under conditions of wide climatic diversity. Bouchet (1963) presented mean annual rainfall, runoff, and insolation data from regions of equatorial rain forest which indicated that the concept is reasonable, although the quality and quantity of the data were inadequate for a systematic test. Solomon (1967) presented annual and mean annual data from tropical and equatorial regions which support the concept in a general way. Morton (1965, 1967b) compared model predictions with monthly values of rainfall less runoff for Canadian river catchments. Although the selection techniques used to avoid the effects of soil moisture and ground water storage changes were somewhat arbitrary, the comparison did provide systematic evidence that the model reflects physical processes with reasonable fidelity.

In the research for this study the concept and model were reviewed and refined. This involved a neater formulation of the basic concept, a more comprehensive exploration of its implications, the use of an alternative technique for estimating potential evaporation, and the selection of more reliable empirical approximations. Differences between annual rainfall and annual runoff for river catchments in Ireland were compared with model predictions and the agreement between the two independent estimates of regional evaporation provides impressive evidence for the validity of the concept and model. Monthly evaporation observations for pans and grass evaporimeters in Ireland were examined in two different ways. In the first, they were compared with values computed from weather observations using the climatological approach. In the second, the pan observations were adjusted to provide estimates of potential evaporation and these were compared with the evaporation equivalent of one half the absorbed insolation, the conceptual value of both potential and regional evaporation under hypothetical humid conditions. These two comparisons provide further confirmation of the model and graphical insight into its implications.

Section 2

Units, Notation and Definitions

Components of the energy budget are expressed in evaporation equivalents based on a latent heat of vaporization of 590 calories per gram. Unless otherwise noted the evaporation and energy units are in millimeters per day, the temperatures are in degrees Celsius, the pressures are in millibars and the wind speeds are in knots. The symbols used are as follows:

a = psychrometric constant

B = radiant heat transfer to space at air temperature

B_E = radiant heat transfer to space from evaporimeter

B_P = radiant heat transfer to space from small moist surface evaporating at potential rate

B_P' = B_P in an arid region

B_R = radiant heat transfer to space from region

b = rate of change of radiant heat transfer to space with surface temperature

C = ratio of cloud cover to total sky

D = vapour pressure deficit of air at screen height = $e(1 - r)$

d_T = rate of change of saturation vapour pressure with temperature between surface and air temperatures

E_E = evaporimeter evaporation

E_P = potential evaporation

E_P' = E_P in an arid region

E_R = regional evaporation

e = saturation vapour pressure at screen height air temperature

e_S = saturation vapour pressure at surface temperature

f_E = vapour transfer function for evaporimeter

f_R = vapour transfer function for region

G = incident insolation

G_E = insolation absorbed by evaporimeter

G_O = extra-atmospheric insolation

G_R = insolation absorbed by region

h_E = heat transfer coefficient for evaporimeter = $\frac{b}{f_E}$ + nap

h_R = heat transfer coefficient for region = $\frac{b}{f_R}$ + ap

K_E = sensible heat transfer to air from evaporimeter

K_P = sensible heat transfer to air from small moist surface evaporating at potential rate

K_P' = K_P in an arid region

K_R = sensible heat transfer to air from region

M = seasonal coefficient in relationship between cloud cover and sunshine duration

m = rate of change of d_T with respect to e

n = ratio of sensible heat transfer area to vapour transfer area

P = rainfall plus water content of snowfall

p = atmospheric pressure

R = runoff in depth on catchment

r = relative humidity at screen height

S = ratio of observed to maximum possible sunshine duration

T = air temperature at screen height

T_S = surface temperature

w = wind speed at 10 meters above surface

Several of the terms used in the foregoing list need further definition. Insolation is the total short wave direct solar radiation and short wave diffuse sky radiation incident on a horizontal surface. Radiant heat transfer to space is the difference between the upward long wave radiation from the surface and the downward long wave radiation from the atmosphere and clouds. Sensible heat transfer to the air is the difference between convection of heat from the surface to the air and that from the air to the surface.

Section 3

Physical Considerations

(a) Moist surface evaporation

The energy balance for an evaporimeter differs from that of a natural evaporating surface because of the effects of the wall, bottom, and surface on the insolation absorption, vapour transfer and heat transfer characteristics. Therefore a generalized form of the climatological approach is necessary for application to evaporimeters. However, the more general form is applicable also to any continuously moist surface for which representative weather observations are available, i.e., to moist surfaces which are of evaporimeter size or of regional size but not those which are intermediate in size.

A simplified energy balance for an evaporimeter is shown in Equation 1. It does not contain terms for sub-surface heat storage changes or for snow melt and thus is not applicable to short time intervals or to snow cover conditions.

$$G_E = E_E + K_E + B_E \qquad (1)$$

The absorbed insolation, G_E, is the source of the energy. The evaporation, E_E, the sensible heat transfer to the air, K_E, and the radiant heat transfer to space, B_E, are all dependent on the surface temperature.

The vapour transfer equation for an evaporimeter is:

$$E_E = f_E(e_S - re) \qquad (2)$$

This equation has an adequate physical basis with the evaporation proportional to the difference between the surface vapour pressure, e_S, and the vapour pressure at the dew point, re, as first suggested by Dalton. However, the vapour transfer function, f_E, is an empirically derived function of the horizontal wind speed. There is reason to believe that f_E should be some function of the atmospheric pressure, but any dependence is so slight that it is practically impossible to obtain a significant empirical relationship.

The sensible heat transfer to the air from an evaporimeter may be estimated from:

$$K_E = napf_E(T_S - T) \qquad (3)$$

Equation 3 has reasonable physical justification with the sensible heat transfer proportional to the atmospheric pressure, p, and the difference between the surface temperature, T_S, and the air temperature, T. It may be derived from the well known Bowen (1926) ratio. The psychrometric constant, a, has limiting values of 0.00058 $°C^{-1}$ and 0.00066 $°C^{-1}$, depending on the state of the atmosphere, and Bowen (1926) concluded that under normal atmospheric conditions it was approximately 0.00061 $°C^{-1}$. For the populated parts of Ireland the atmospheric pressure, p, may be assumed constant at 1,000 millibars without significant error. The ratio of sensible heat transfer to vapour transfer area, n, is equal to one for a two dimensional surface such as a puddle of water, but for a pan of water exposed to the air it must be determined experimentally.

A relationship between the radiant heat transfer to space from an evaporimeter and the surface temperature is presented in the following equation:

$$B_E = b(T_S - T) + B \qquad (4)$$

The radiant heat transfer to space at air temperature, B, depends on the cloud amount and height and, in a marginal way, on atmospheric temperature and humidity. Empirical estimates of B are discussed in Section 4(a). The rate of change of radiant heat transfer with temperature, b, may be considered constant at approximately 0.18 mm day^{-1} $°C^{-1}$ within the temperature range experienced on naturally evaporating surfaces.

It is apparent from Equations 1 to 4 inclusive that the impact of any climatic factor on the energy balance is brought about by its effect on the surface temperature. In practice, the surface temperature is exceedingly difficult, if not impossible, to measure. However, this difficulty may be overcome by expressing Equation 2 as:

$$\begin{aligned} E_E &= f_E(e_S - e) + f_E e(1 - r) \\ &= f_E d_T(T_S - T) + f_E D \qquad (5) \end{aligned}$$

and rearranging it to give:

$$T_S - T = \frac{E_E - f_E D}{f_E d_T} \qquad (6)$$

The rate of change of saturation vapour pressure with temperature, d_T, is a function of T_S and T or of e_S and e. As the difference between the surface and air temperature is small it is sufficiently accurate to estimate d from T or e alone. The following simple relationship has been derived from saturation vapour pressure tables and is applicable at temperatures between 0°C and 20°C:

$$d_T = 0.10 + 0.0583e \qquad \text{mb } °C^{-1} \qquad (7)$$

If Equation 6 is inserted into Equations 3 and 4, the results are:

$$K_E = \frac{nap}{d_T}(E_E - f_E D) \qquad (8)$$

$$B_E = \frac{b}{f_E d_T}(E_E - f_E D) + B \qquad (9)$$

and when these are inserted in Equation 1 the energy balance becomes:

$$\begin{aligned} G_E = E_E &+ \frac{nap}{d_T}(E_E - f_E D) \\ &+ \frac{b}{f_E d_T}(E_E - f_E D) + B \qquad (10) \end{aligned}$$

so that:

$$E_E = \frac{d_T}{d_T + h_E}(G_E - B) + \frac{h_E}{d_T + h_E} f_E D \qquad (11)$$

Equation 11 is the well known formulation by Penman (1948), as generalized with the heat transfer coefficient $h_E = \frac{b}{f_E} + nap$. The slope b takes into account the variation of radiant heat transfer to space with surface temperature and the ratio n takes into account sensible heat transfer through the sides and bottom of the evaporimeter.

Using Equations 1, 3, 4 and 5 and similar algebraic procedures it may be demonstrated that:

$$K_E = \frac{nap}{d_T + h_E}(G_E - B - f_E D) \qquad (12)$$

and:

$$B_E = \frac{h_E - nap}{d_T + h_E}(G_E - B - f_E D) + B \qquad (13)$$

Equations 11, 12 and 13 permit the surface temperature dependent components of the energy balance for a continuously moist surface to be estimated by their causes as reflected in routine weather observations. When considered in conjunction with Equation 1, 2, 3 and 4, they provide an insight into the basic physical processes. Thus an increase in absorbed insolation, G_E, or a decrease in radiant heat transfer to space at air temperature, B, increases the surface temperature to some equilibrium value at which increases in evaporation, sensible heat transfer and radiant heat transfer maintain the energy balance. A reduction in atmospheric vapour pressure, re, resulting in an increase in vapour pressure deficit, D, causes the evaporation to rise and the surface temperature to fall until the additional evaporative energy is balanced by decreases in the sensible and radiant heat transfers. Similarly, an increase in evaporation due to an increase in wind speed and hence in the vapour transfer function, f_E, is balanced by decreases in the sensible and radiant heat transfers. This happens because the effects of the lower surface temperatures on sensible heat transfer are greater than those of the higher wind speed, whereas the opposite is true for evaporation. If the air temperature rises the radiant and sensible heat transfers fall, and this causes the surface temperature to increase until the energy balance is restored by an increase in evaporation.

The way in which the surface temperature distributes energy amongst the various components of the energy balance ensures good accuracy for the climatological approach to estimating evaporation. An error in one of the weather observations causes an equivalent error in the entire energy balance, but this is shared amongst the evaporation, sensible heat transfer and radiant heat transfer, instead of being concentrated in the evaporation alone as it would be in the energy balance approach of Equation 1 or the vapour transfer approach of Equation 2. The effects of such sharing on the evaporation may be estimated by differentiating Equation 11. In the results of the differentiation ΔE_E is the error or small change in evaporimeter evaporation due to ΔG_E, ΔB, ΔT, Δre, and Δf_E, which are the errors or small changes in absorbed insolation, radiant heat transfer to space at air temperature, air temperature, vapour pressure at dew point, and vapour transfer function (i.e., wind speed) respectively.

$$\Delta E_E = \frac{d_T}{d_T+h_E}(\Delta G_E - \Delta B) + \frac{d_T}{d_T+h_E}\left[h_E f_E + m(G_E-B-E_E)\right]\Delta T - \frac{h_E}{d_T+h_E} f_E \Delta re + \left[\frac{h_E}{d_T+h_E} E_E - \frac{d_T}{d_T+h_E}\,\frac{nap\,(G_E-B-E_E)}{h_E}\right]\frac{\Delta f_E}{f_E} \quad (14)$$

According to Equation 7 the value of m, the rate of change of d_T with respect to e, is 0.0583 $^\circ C^{-1}$.

There is another method of estimating evaporation using the climatological approach. It was developed by Ferguson (1952), who generalized it to take into account the change of radiant heat transfer with surface temperature, but did not generalize it to take into account sensible heat transfer through the walls and bottom of an evaporimeter. As indicated by Morton (1965, 1967b) this may be accomplished by combining Equations 1, 2, 3 and 4 and rearranging the terms as follows:

$$e_S + h_E T_S = re + h_E T + \frac{G_E - B}{f_E} \quad (15)$$

The right hand side of Equation 15 may be estimated from weather observations, but both e_S and T_S on the left hand side are unknown. However e_S and T_S are related to each other in a manner defined by saturation vapour pressure tables, so that a family of curves for different values of the heat transfer coefficient, h_E, may be prepared to relate e_S to $e_S + h_E T_S$. Values of the right hand side of Equation 15 may be substituted for $e_S + h_E T_S$ so that, by interpolation on the basis of known values of h_E, the family of curves yields estimates of e_S from records of routine weather observations. The evaporimeter evaporation may then be computed by inserting the estimated value of e_S into Equation 2.

There is practically no difference in assumptions or results between the Penman and the Ferguson formulations of the climatological approach. The Ferguson formulation is slightly superior, as it eliminates the need for the quantity d_T. However, because it is graphical and difficult to use analytically, it was not used in this investigation.

(b) Potential evaporation

As usually defined, potential evaporation is the evaporation which would occur if there was an adequate moisture supply at all times. This definition does not specify the area of adequate moisture supply and therefore fails to allow for the effect of the evaporation and sensible heat transfer on the evaporability of the overpassing air. As defined herein the potential evaporation is the evaporation from a continuously moist surface with regional insolation absorption, vapour transfer, and heat transfer characteristics and with an area so small that the convected heat and vapour have no significant effect on the overpassing air. With such a definition the potential evaporation may be estimated from weather observations by substituting G_R for G_E, f_R for f_E and h_R for h_E in Equation 11. The result is:

$$E_P = \frac{d_T}{d_T+h_R}(G_R - B) + \frac{h_R}{d_T+h_R} f_R D \quad (16)$$

The ratio of sensible heat transfer area to vapour transfer area, n, for a small plot of vegetation would vary with its dimensions and exposure. However for an area as large as a region the effective ratio would be equal to one, and this value is used in the definition of the regional heat transfer coefficient, $h_R = \frac{b}{f_R} + ap$.

The sensible heat transfer to the air and the radiant heat transfer to space associated with potential evaporation, K_P and B_P, may be estimated by making similar substitutions in Equations 12 and 13.

In order to derive potential evaporation from evaporimeter observations it is necessary to adjust for the differences in the absorbed insolation, in the sensible heat transfer, and in the vapour transfer function. This may be accomplished by solving for the vapour pressure deficit, D, in Equation 11 and inserting the result in Equation 16. The result is:

$$E_P = \frac{h_R f_R}{h_E f_E}\,\frac{(d_T+h_E)}{(d_T+h_R)} E_E + \frac{d_T}{d_T+h_R}\left[(G_R - B) - \frac{h_R f_R}{h_E f_E}(G_E - B)\right] \quad (17)$$

In Equation 17, $h_R f_R = b + apf_R$ and $h_E f_E = b + napf_E$, both of them functions of the wind speed only. These quantities, expressed in terms of their ratio, have considerable influence on the accuracy of the results but unfortunately they are not well defined. The reliability of the potential evaporation estimates from Equation 17 is influenced only slightly by the accuracy of the air temperature, absorbed insolation and radiant heat transfer values that are used, so that very rough estimates of these quantities are satisfactory.

In Equation 17 the potential evaporation is treated as the result of the same factors that cause evaporimeter evaporation. In Equation 16 it is treated as the effect of three kinds of causes. The most important of these are the insolation and atmospheric radiation which are the results of processes in the upper atmosphere or beyond. Of intermediate importance are heat, humidity, and turbulence in the lower atmosphere which are the results of heat and vapour transfers from adjacent regions. The least important of the three classifications are the surface colour and roughness which can cause small variations in the absorbed insolation and in the vapour transfer function. It should be pointed out that the lack of importance given to surface conditions in potential evaporation is due to the continuity of water supply to the surface. When surface moisture can change, it becomes one of the controlling factors in the evaporation process.

(c) Potential evaporation and regional evaporation

Comparisons of evaporimeter and weather observations in an arid region with those in a humid region provide an insight into the relationship between potential and regional evaporation. Table 1 provides such a comparison for the rather arid region near Swift Current, Saskatchewan and for the rather humid region near Truro, Nova Scotia for the month of July, 1964.

Table 1

	SWIFT CURRENT	TRURO AIRPORT
Latitude	50° 16′ N	45° 22′ N
Longitude	107° 44′ W	63° 16′ W
Elevation	825 m	40 m
Class A pan evaporation (E_E)	332 mm	110 mm
Incident insolation (G)	344 mm	238 mm
$\frac{E_E}{G}$	0.97	0.46
Pan water temperature (T_S)	20.0 °C	20.0 °C
Air temperature (T)	20.4 °C	17.4 °C
$T_S - T$	−0.4 °C	+2.6 °C
Vapour pressure at dew point	12.2 mb	16.6 mb
Relative humidity	51 %	83 %
Wind speed at pan rim	5.4 knots	5.5 knots

These data were abstracted from the *Monthly Records of Meteorological Observations in Canada.* They are typical of observations from both arid and humid regions throughout the world during the seasons when significant quantities of evaporative energy are available. Thus in arid regions, where there is little or no water for evaporation from the regional surfaces, the pan evaporation uses practically all the incident insolation (97 per cent at Swift Current), whereas in humid regions, with only occasional water supply limitations to regional evaporation, the pan evaporation uses a much lower proportion (46 per cent at Truro) of the incident insolation. Furthermore, the difference between the average temperatures of the pan water and the air, (T_S - T), is consistently negative in arid regions and consistently positive in humid regions. Such patterns provide a basis for an understanding of the relationship between regional and potential evaporation.

The relationship may be derived by considering simplified surface energy balances for a region and for a potential evaporimeter well exposed to the regional weather. A potential evaporimeter is defined as a continuously moist surface with regional insolation absorption, vapour transfer and heat transfer characteristics. The existence of such an evaporimeter is not essential to the analysis as the limited surface area does not yield or absorb sufficient heat and vapour to have a perceptible effect on the overpassing air. The energy balance for the potential evaporimeter is shown in Equation 18 and that for the region is shown in Equation 19.

$$E_P = G_R - B_P - K_P \tag{18}$$

$$E_R = G_R - B_R - K_R \tag{19}$$

Neither of these equations includes terms for sub-surface heat storage change or snow melt so that the analysis does not apply to evaporation for short periods of time, from deep lakes or from snow covered surfaces.

Under conditions of adequate moisture supply to the soil and vegetation surfaces of the region, each of the regional energy quantities in Equation 19 equals its potential evaporimeter counterpart in Equation 18, i.e., $E_R = E_P$, $K_R = K_P$, and $B_R = B_P$. However, when inadequate water supplies limit regional evaporation there are complicated changes in the energy terms of both Equations 18 and 19. With no change in the absorbed insolation, G_R, the energy not used for evaporation heats up the regional surfaces, thereby increasing the radiant heat transfer to space, B_R, and the sensible heat transfer to the air, K_R. Because of decreased evaporation and increased sensible heat transfer from the surface of the region, the air over the region becomes drier and hotter, and as it passes over the evaporimeter it increases the potential evaporation, E_P. This causes the evaporimeter surface temperature to decrease until reductions in the radiant and sensible heat transfers, B_P and K_P, balance the energy used for increased evaporation. Thus a decrease in regional evaporation due to lack of water results in an increase in the evaporation from a potential evaporimeter within the region, and energy conservation is maintained by compensating changes in the respective radiant and sensible heat transfers. Therefore, it seems reasonable to assume that the sum of the regional evaporation and potential evaporation, $E_R + E_P$, remains constant under conditions of constant insolation but varying water availability. As the effects of changes in the regional energy balance are transferred to the evaporimeter through changes in the heat and vapour content of the overpassing air, the assumption is not valid when net advections of evaporability in the lower atmosphere over the region are significant. Therefore, the results do not apply near a sharp climatic boundary such as the edge of an oasis or a coastline, or to short time intervals.

As the region becomes more arid, E_R continues to decrease and E_P continues to increase. At the same time the surface temperature of the potential evaporimeter continues to decrease relative to the air and eventually becomes less. When the region becomes completely arid the regional evaporation is zero and the potential evaporation is equal to the assumed constant value for $E_R + E_P$. With prime signs indicating the arid condition the constant may be evaluated by:

$$E_R + E_P = E_P' = G_R - B_P' - K_P' \tag{20}$$

As the surface temperature of a potential evaporimeter under arid conditions is less than the air temperature, the radiant heat transfer to space, B_P', is positive and small, whereas the sensible heat transfer to the air, K_P', is negative and small. Therefore it seems reasonable to assume that the sum $B_P' + K_P'$ is insignificant. With this assumption:

$$E_R + E_P = G_R \tag{21}$$

Equation 21 was derived in different ways by both Bouchet (1963) and Morton (1965).

It would be fortuitous indeed if the two assumptions basic to Equation 21 were correct in all respects. However, despite over-simplification, they do reflect reality as it is observed in climatically diversified evaporimeter and weather records, as exemplified by those at Swift Current and Truro. By the standards of hydrometeorology the physical basis for Equation 21 is good but, because of its hypothetical nature, it can be judged only by the results of systematic tests with data from a wide variety of environments.

An extreme example of the workings of Equation 21 is provided by a hypothetical desert gradually subjected to irrigation. For the sake of simplicity it is assumed that the absorbed insolation and atmospheric radiation remain constant. A potential evaporimeter is located in the centre of the proposed irrigated region. In its original state the desert has no evaporation, so that all the energy received as absorbed insolation is dissipated to the air as sensible heat or to space as radiant heat. With no evaporation and a high rate of sensible heat transfer, the air over the desert is dry and hot, and as it passes over the evaporimeter it causes a high rate of potential evaporation. The energy used in evaporation causes the surface temperature of the evaporimeter to be much lower than the surface temperature of the desert and somewhat lower than the temperature of the air. Under such conditions the loss of energy through radiant heat transfer to space is less than that from the desert and the sensible heat transfer is from the air to the surface of the evaporimeter, i.e., it is negative. Because the relatively small radiant and sensible heat transfer terms tend to cancel each other out, the potential evaporation is equal to G_R, the absorbed insolation, as indicated by Equation 21 when $E_R = 0$. The interaction between the energy balances for a completely arid region and for a representative potential evaporimeter is shown schematically in Figure 2.

As the irrigated area is extended outward from the evaporimeter, the evaporation from the water, soil and vegetation surfaces modifies the overpassing air. The modified air would be cooler and more humid than the desert air and could maintain the same potential evaporation rate only if it were possible to keep the surface of the evaporimeter at a higher temperature without increasing the radiant heat losses or decreasing the sensible heat gains. As this is impossible, the evaporimeter surface temperature increases to a new equilibrium value at which the energy made available by reduced evaporation is balanced by increased radiant heat losses and decreased sensible heat gains. The difference between the potential evaporation rate under the desert air, i.e., the absorbed insolation, and the potential evaporation rate under the modified air provides an indication of the amount of the modification. According to Equation 21, the difference is equal to the regional evaporation but in the case under consideration, where there is no evaporation from the part of the region still desert, it represents the evaporation from the irrigated area weighted by the proportion of irrigated area in the region.

As the irrigated area is extended, the potential evaporation decreases further and the evaporimeter surface temperature rises until it eventually becomes higher than the air temperature. The sensible heat gain then becomes a sensible heat loss. Progressive increases in regional evaporation through extensions of the irrigated area cause further decreases in potential evaporation and increases in sensible and radiant heat transfers from the evaporimeter. At some stage in the extension process the irrigated area becomes so large that the heat and vapour content of the lower atmosphere are not significantly influenced by advection to and from the surroundings but are controlled by convection from the irrigated surfaces. Under these conditions Equation 21 indicates that the evaporation from the irrigated region is equal to the absorbed insolation less the potential evaporation as measured by the evaporimeter. If the irrigation process is such that the land and vegetation surfaces of the region are continuously moist, the evaporation from the irrigated region is equal to the potential evaporation and, in accordance with Equation 21, both are equal to one half of the absorbed insolation. The energy balances for a completely humid region and for a representative potential evaporimeter are shown schematically in the lower part of Figure 2 in such a way that they may be readily compared with their counterparts for an arid region.

From the formulation of Equation 21 and the foregoing example, it is evident that the potential evaporation is not a cause of regional evaporation and is not even the energy available for regional evaporation under conditions of adequate moisture supply. As the ratio of regional evaporation to absorbed insolation increases from a minimum value of zero under arid conditions to a maximum value of one half under humid conditions, the ratio of potential evaporation to absorbed insolation decreases from a maximum value of one to a minimum value of one half. Therefore the potential evaporation changes when regional water availability and regional evaporation change. Although both the potential and regional evaporation share the same upper and extra-atmospheric causes, i.e., insolation and atmospheric radiation, the potential evaporation is significantly influenced by changes in the heat and vapour content of the lower atmosphere and is, therefore, more an effect than a cause of regional evaporation.

According to Equation 21 the potential evaporation is equal to the absorbed insolation that is not used for regional evaporation and according to Equation 19 this is equal to the sum of the sensible and radiant heat transfers from the surfaces of the region. Using this conclusion with the regional equivalents of Equations 3 and 4 results in:

$$K_R = \frac{ap}{h_R} (E_P - B) \qquad (22)$$

$$B_R = \frac{h_R - ap}{h_R} E_P + \frac{ap}{h_R} B \qquad (23)$$

Another perspective to the concept of Equation 21 is provided by the difference between the potential evaporation and one half of the absorbed insolation, $E_P - 1/2\ G_R$. It is equal to the difference between one half of the absorbed insolation and the regional evaporation, $1/2\ G_R - E_R$, and as such it is the energy equivalent of the water that is not available for evaporation because of soil moisture or vegetative conditions. Because of the equality of the two differences the potential evaporation may be visualized as the reflection of the regional evaporation on the opposite side of a line with a locus at one half the absorbed insolation.

As is evident from the foregoing example of the progressive irrigation of a hypothetical desert, the evaporation from a moist surface is a function of its size. This is because of the changes that take place in the heat and vapour content of the air as it passes over the surface. For this reason the usual definition of potential evaporation, which specifies a moist surface of indefinite size, is inadequate. Despite the possibility of some confusion, the term potential evaporation was retained herein for what is essentially a new and more useful definition because of the

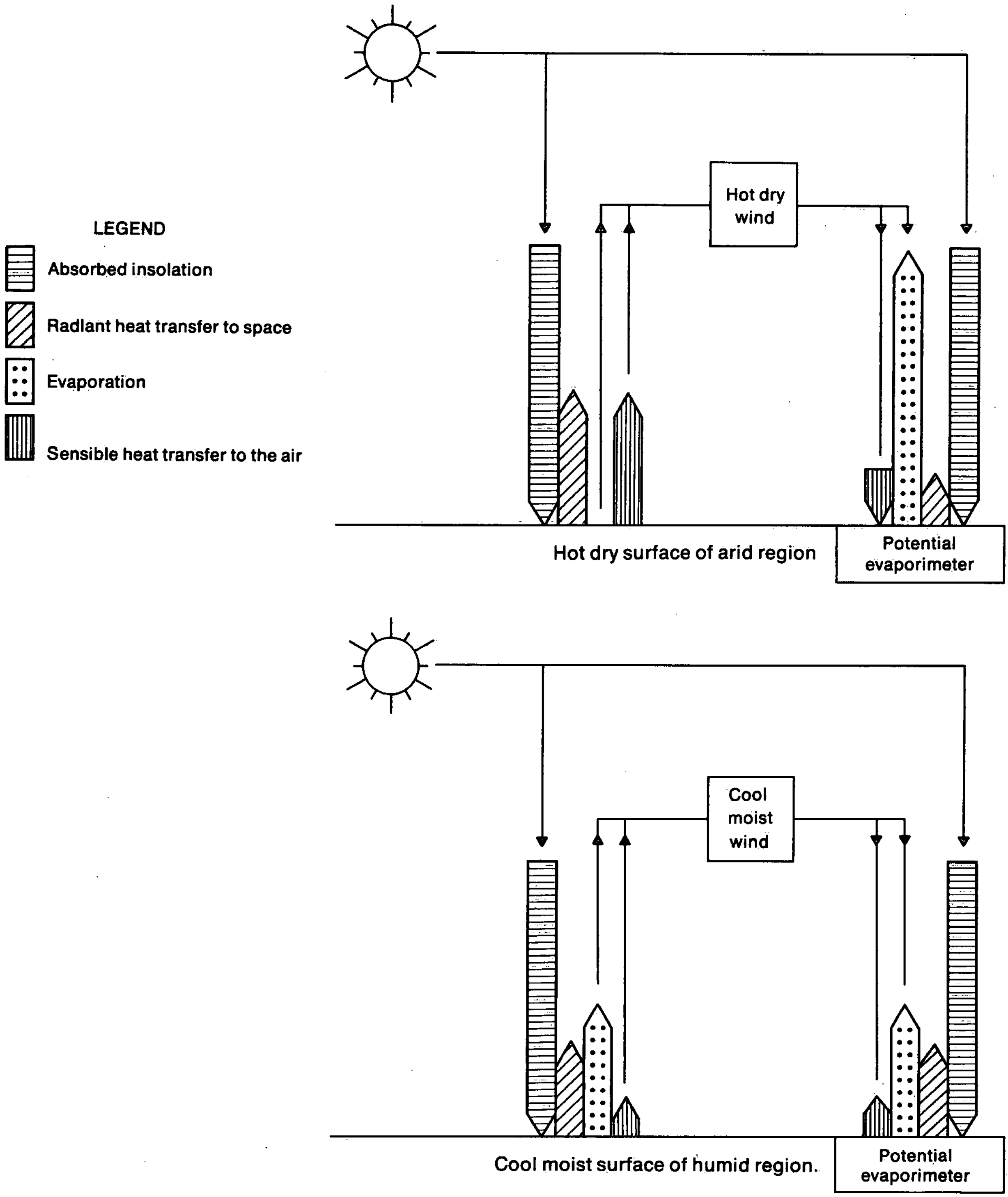

Figure 2   Interactions between potential and regional energy balance

analogy in Equation 21 between potential evaporation and potential energy, regional evaporation and kinetic energy, and absorbed insolation and total energy. Thus changes in regional evaporation (kinetic energy) are balanced by changes in potential evaporation (potential energy) under conditions of constant absorbed insolation (total energy). The condition of complete regional humidity even corresponds to the hydraulic condition of critical flow when kinetic energy is equal to the potential energy. However, the hydraulic analogy breaks down at supercritical conditions because the regional evaporation cannot exceed potential evaporation.

The maximum dimension of a moist surface for the measurement of potential evaporation is not known with any exactitude. Dyer and Crawford (1965) present data which indicate that the decrease in evaporation 150 meters downwind of a sharp transition from arid to humid surface conditions may be only 4 per cent under conditions of high insolation and may be up to 40 per cent under conditions of low insolation. Therefore it would seem that an evaporimeter with a fetch of more than five to ten meters could give values of potential evaporation that are too low. However, the actual maximum dimension is not critical as economic considerations tend to keep the size of evaporimeters much smaller.

The minimum dimension for a region depends on its location. Ideally a region is an area in which the transitional zone, i.e., the zone where advected vapour and heat interacts and reaches equilibrium with the vapour and heat convected from the regional surfaces, is so small that its effects may be ignored. When there is little contrast with the neighbouring environments a region may be quite small. However, when near a sharp climatic boundary, such as the edge of an oasis or a coastline, the transitional zone becomes much larger and the minimum size for a region increases. Modification of the lower atmosphere in transition from sea to land environments is discussed further in Section 6 in connection with its effects on evaporation from pans.

In accordance with the concept of Equation 21 a large lake or swamp may be classified as a region if the evaporation approximates one half of the absorbed insolation. Such a classification is rare for large lakes because of the sharp environmental contrasts associated with the climate over the adjacent land or with sub-surface heat storage changes. The first of these reasons for environmental contrast may be exemplified by the Salton Sea, with an area of 900 square kilometers and an average depth of eight meters, and the second by Lake Superior, with an area of 83,000 square kilometers and a maximum known depth of 410 meters. Hughes (1967) has presented data which indicate that the average ratio of water budget evaporation to absorbed insolation for the Salton Sea is approximately two thirds. This high value of the ratio may be attributed to excessive evaporation from a transitional zone large enough to bring the lower layers of hot dry air from the nearby desert into equilibrium with the lake environment. Thus the transitional zone acts as a sink for the excess evaporability resulting from deficient evaporation at the desert surface. Morton (1967a) presented data which show that the average ratio of water budget evaporation to absorbed insolation for Lake Superior is approximately one third. This low value may be attributed to the effects of sub-surface heat storage changes, which keep the climate over the lake continuously out of phase with the climate over the land, and result in the advection of large quantities of heat from the lake environment. During the spring and early summer when there is an inversion over the surface, the lake uses practically all the absorbed insolation for internal heating. During the remaining eight or nine months of the year the heat is released for evaporation, radiant heat transfer, and sensible heat transfer, with the largest releases occurring in December when the surface water is near maximum density and the land environment is frigid. Therefore the adjacent land environment acts as a sink for the excess heat transferred to the lower atmosphere from the depths of the lake during the greater part of the year.

A large deep lake may be treated as a region if sub-surface heat storage changes and advections of heat between the lake and land environments are taken into account. By using an elaboration of Equation 21 and linking the advections of heat with the convections of heat from the surface to the air, Morton (1967a) developed the hypothesis that the monthly evaporation from Lake Superior or Lake Ontario is equal to the monthly radiant heat transfer to space.

The minimum interval of time which must be used if sub-surface heat storage changes and net advections of evaporability are to be ignored as insignificant is not known with any certainty. For a region which includes little deep water and is remote from sharp climatic boundaries one day is too short and one month is more than ample. A period as short as one week is probably adequate.

(d) Regional evaporation

The evaporation from the water, soil and vegetation surfaces of a region may be estimated from representative weather observations by combining Equations 16 and 21. The resultant model is:

$$E_R = G_R - E_P = \frac{d_T}{d_T+h_R} B + \frac{h_R}{d_T+h_R} (G_R - f_R D) \quad (24)$$

Estimates of regional evaporation from Equation 24 increase with increasing radiant heat transfer to space at air temperature and with decreasing wind speeds and vapour pressure deficits. This seems contrary to all reason until it is remembered that regional evaporation is being estimated by its effects, not by its causes.

The regional sensible heat transfers to the air and radiant heat transfers to space may be estimated from representative weather observations in a similar way by combining Equation 16 with Equations 22 and 23.

The regional evaporation may be estimated from representative evaporimeter and weather observations using the following solution of Equations 17 and 21.

$$E_R = \frac{h_R}{d_T+h_R} G_R + \frac{d_T}{d_T+h_R}\left[\frac{h_R f_R}{h_E f_E}(G_E - B) + B\right] - \frac{h_R f_R (d_T+h_E)}{h_E f_E (d_T+h_R)} E_E \quad (25)$$

Estimates of regional evaporation from Equation 25 increase with decreasing evaporimeter evaporation. Although contrary to superficial expectations, this is in accord with the concept that the evaporation from an evaporimeter is a measure of the amount of energy not used for regional evaporation because of inadequate moisture supply.

Empirical Considerations

(a) General

In applying the equation presented and developed in the last section it is necessary to use empirical approximations for estimating some of the quantities that appear in them. Thus when radiation observations are not available it is necessary to estimate incident insolation and radiant heat transfer to space from other weather observations. Such estimates are discussed in this sub-section. The amount of incident insolation absorbed, the vapour transfer function, and the ratio of sensible heat transfer area to vapour transfer area depend on the nature of the surface and these are discussed in sub-sections 4(b), 4(c), and 4(d) for river catchments, Class A evaporation pans, and Thornthwaite type grass evaporimeters, respectively.

Incident insolation is observed by an instrument called the pyranometer which measures the sum of the short wave direct solar radiation and the short wave diffuse sky radiation incident on a horizontal surface. The only long term record of such observations in Ireland is on the southwest coast at Valentia Observatory. However, there are records of sunshine duration at many locations throughout the country. Under such circumstances it is usual to estimate the incident insolation, G, from an empirical relationship with the extra-atmospheric insolation, G_O, and the ratio of observed to maximum possible sunshine duration, S. Connaughton (1967) has shown that the relationship at Valentia Observatory is:

$$G = G_O\,(0.25 + 0.58S) \qquad (26)$$

This equation gives results that are anomalously high when compared with results of similar equations for stations in Great Britain. Monteith (1966) has shown that an equation giving similar results exists for Aberporth on the south shore of Cardigan Bay and has speculated that "at stations on western coasts such as Aberporth and Valentia, the air may be clean throughout the day when the wind is westerly or when a sea breeze is blowing inland toward the shore". However, he presented data from a station near Aberystwyth, fifty kilometers northeast of Aberporth and only three kilometers inland from Cardigan Bay, which indicate that any such effects are confined to a narrow coastal zone. The data from near Aberystwyth, along with data from many other inland British stations, support a more general equation proposed by Black, Bonython, and Prescott (1954).

$$G = G_O\,(0.23 + 0.48S) \qquad (27)$$

Equation 27 is based on data from 32 stations in many parts of the world including Great Britain and Western Europe, and it is suspected that any effects of urban or industrial smog included in it would be balanced by the effects of high relative humidity in Ireland. Therefore, in the light of the doubtful applicability of Equation 26 outside a narrow coastal zone, it seemed reasonable to use Equation 27 in this investigation. The use of Equation 27 in Ireland is not meant as an endorsement of its world-wide usefulness. For example, it gives lower estimates than Equation 28, which was used in this investigation to compute incident insolation for river catchments in the naturally forested regions of Canada east of the Rocky Mountains.

$$G = G_O\,(0.23 + 0.54S) \qquad (28)$$

The extra-atmospheric insolation, G_O, is a function of the solar constant, the season of the year, and the latitude. The monthly values used in this investigation were derived from the values presented by Shaw (1942) increased by 3 per cent to make them consistent with the presently accepted solar constant of 2.0 calories per square centimeter per minute. Table 2 shows the values thus derived expressed in millimeters per day of evaporation equivalent on the basis of latent heat of vaporization of 590 calories per gram.

Table 2

MONTH	Extra-atmospheric insolation in mm day^{-1}	
	Latitude 50°N	Latitude 60°N
January	3.8	1.5
February	6.1	3.6
March	9.4	7.2
April	13.1	11.5
May	15.9	15.1
June	17.1	16.9
July	16.5	16.0
August	14.1	12.9
September	10.8	8.8
October	7.2	4.8
November	4.4	2.0
December	3.2	1.0

Values of extra-atmospheric insolation for intermediate latitudes in Ireland may be estimated from the tabulation by linear interpolation.

Radiant heat transfer to space is the difference between the upward long wave radiation from the surface and the downward long wave radiation from the atmosphere and clouds. Observational estimates may be made by subtracting the difference between incident insolation and reflected insolation, as measured by pyranometers, from net radiometer observations. As there are no long term net radiometer observations for Ireland it is necessary to base estimates on Equation 4 using empirical estimates of B, the radiant heat transfer to space at air temperature.

There are many empirical methods for estimating B from routine weather observations, most of them limited to clear sky conditions. In these the upward long wave radiation depends on air temperature and surface emissivity and the estimate is based on the Stefan-Boltzman law. The downward atmospheric radiation under clear sky conditions is assumed to conform to the Stefan-Boltzman law applied to observed air temperatures, with an emissivity depending on observed atmospheric vapour pressures. Such methods are both complex and approximate and the effects of vapour pressure variations tend to nullify the effects of temperature variations. Moreover the most important variables, the cloud amount and cloud height, are taken into account in a very crude manner, if at all. Therefore it is evident that the complexity is not related to reliability of results. Because of this and because radiant heat transfer to space is small compared to the insolation during the months when evaporation is high, it was decided to use the much simpler equation suggested by Monteith (1961). In the appropriate units and with an assumed emissivity of 0.97 Monteith's equation is:

$$B = 3.29 - 2.93\ C \qquad \text{mm day}^{-1} \quad (29)$$

This is expressed in terms of sunshine duration by:

$$B = 3.29 - 2.93\ (M - 0.8S) \qquad \text{mm day}^{-1} \quad (30)$$

The quantity (M - 0.8S) is an estimate of C, the ratio of cloud cover to total sky, derived from over 600 monthly averages of six-hourly cloud cover and daily sunshine observations, which were obtained from the *Monthly Weather Reports* of the Irish Meteorological Service. The data used were from a total of eight years and the number of stations represented varied from four to fifteen. The quantity M is a seasonal coefficient with a value of 1.00 during the summer months of May, June, July and August, a value of 0.90 during the winter months of November, December, January and February, and a value of 0.95 during the transitional months of March, April, September and October. The seasonal effect is significant, despite the wide scatter of the data, and may be explained partially as a day length factor reflecting the tendency for cloud cover to be less during the hours when the sun does not shine. It is improbable that these values for M would be applicable in latitudes much different from those of Ireland.

(b) River catchments

Barry and Chambers (1966) have presented data indicating that the albedo, i.e., the proportion of reflected to incident insolation, during the summer in England and Wales varies from 0.12 for peat and moss to 0.24 for agricultural grassland, with intermediate values of 0.15 for heather moor and 0.18 for deciduous woodland. Most river catchments in Ireland would have combinations of these four types of cover and the proportions would vary with agricultural development. Because of difficulty in quantifying the effects of seasonal changes and variations in cover, it has been assumed that the albedo for Irish catchments is constant at a value of 0.18. When combined with Equation 27 this value of albedo yields the following estimate for regional absorbed insolation:

$$G_R = G_O\,(0.19 + 0.39\ S) \quad (31)$$

As discussed previously the effective ratio of sensible heat transfer area to vapour transfer area for a region such as a river catchment is equal to one and this value is used in the catchment heat transfer coefficient, $h_R = \frac{b}{f_R} + ap$.

Consideration of Equation 2 indicates that the vapour transfer function is equal to the evaporation divided by the difference between the surface and the atmospheric vapour pressures. This definition is of no use in the formulation of a relationship between the vapour transfer function and the wind speed for river catchments as it is not possible to observe or estimate the surface vapour pressure. However this difficulty may be overcome by re-arranging Equation 24 to give:

$$f_R = \frac{1}{D}\left[(G_R - E_R) - \frac{d_T}{h_R}\,(E_R - B)\right] \quad (32)$$

Equation 32 has been used to evaluate monthly values of the vapour transfer function from the data for Canadian river catchments presented by Morton (1965). In order to avoid the effects of albedo changes the computations were limited to the months of May, June, July and August, resulting in a total of 46 sets of data. Of these, the four with reference numbers 17, 27, 38 and 50 were discarded because they failed to satisfy a criterion designed to detect suspiciously low values of the vapour pressure deficit D. In the solution of Equation 32, E_R was the difference between rainfall and runoff during months when hydrologic criteria indicated that there were no significant changes in surface, soil moisture, and ground water storage; G_R was computed from Equation 28 assuming an albedo of 0.18; and B was derived from Equation 30. The quantity h_R depends on f_R to such a slight extent that it seemed reasonable to compute it from estimates of f_R based on the equation in the reference. Figure 3 shows the values of the vapour transfer function thus computed and plotted against the 10-meter wind speeds. The linear least squares solution is shown as a solid line. In the appropriate units and with the wind velocity in knots, the equation of the line is:

$$f_R = 0.086\ w \qquad \text{mm day}^{-1}\text{mb}^{-1} \quad (33)$$

There are three arguments against the use of Equation 33. These are that it assumes the validity of Equation 24; that it is based on data from the forested regions of Canada, and that it has a coefficient of correlation of only 0.56. The broken line in Figure 3 has been included to partially offset these arguments. It is the well known equation proposed by Penman (1948) as revised for the changes in units and wind measurement height. The two lines intersect near the average of the 42 points. Moreover, although the two lines appear to diverge, the slopes are not different at the 10 per cent level of significance. However, despite the general agreement between the two equations, the chief argument for the use of Equation 33 is the lack of an alternative based on catchment data.

(c) Class A evaporation pans

The United States Weather Bureau Class A evaporation pan is 1210 millimeters in diameter, 250 millimeters in depth and constructed of galvanized iron, normally unpainted. It is set a short distance above the ground on 50 by 100 millimeter wooden supports in such a way that approximately one half the bottom area is exposed to the air. The water level is kept between 40 and 70 millimeters below the rim. Normally observations are made once a day, early in the morning. The evaporation is obtained by adding the change in water level (positive downward and negative upward), as obtained by reading a hook gauge or changing the water content to conform to a point gauge, to the rainfall as measured in a standard rain gauge.

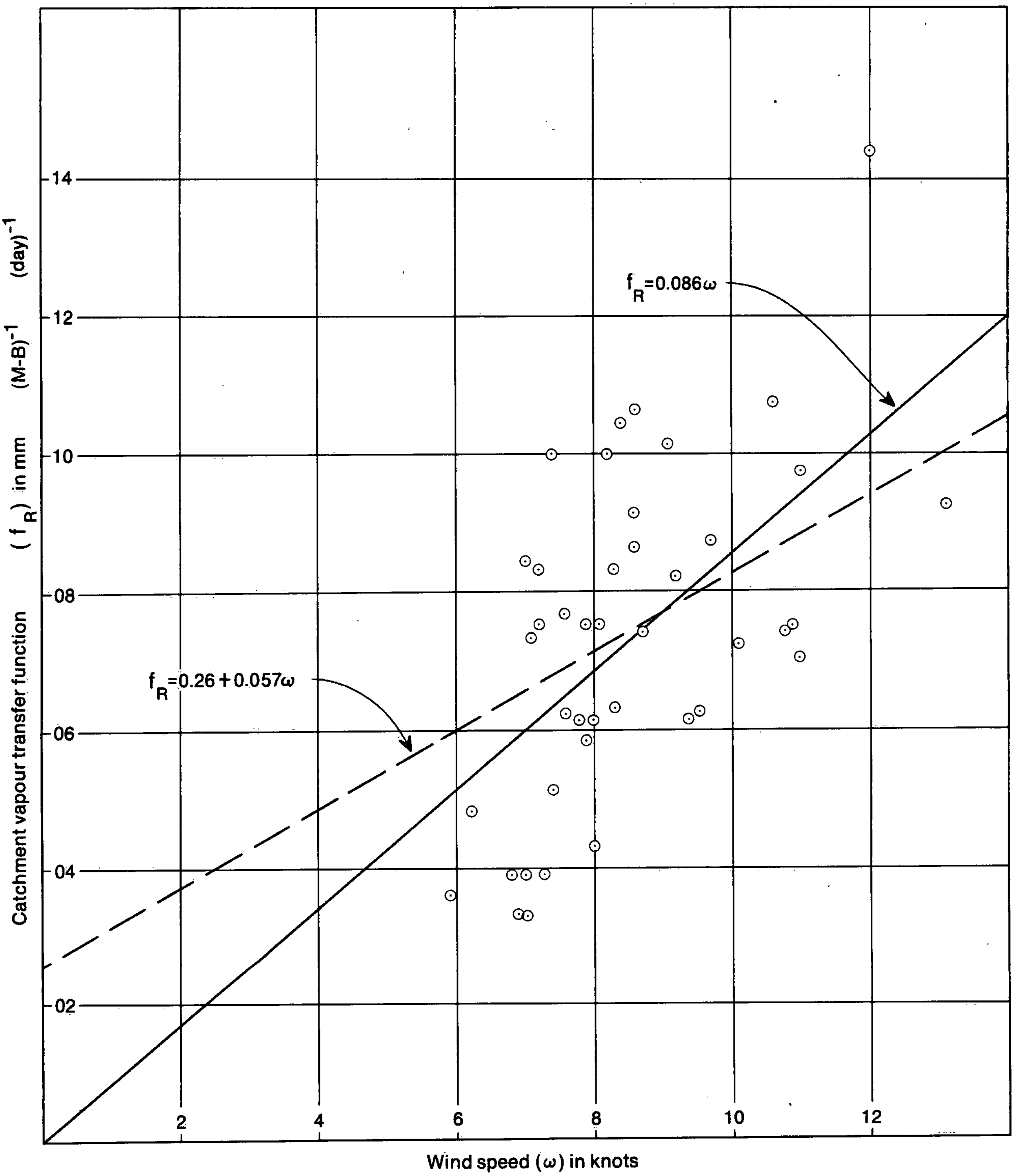

Figure 3 Vapour transfer function for river catchments

The vapour transfer function for Class A evaporation pans was developed in the course of the *Lake Hefner Studies* (1954). In the appropriate units and with the wind speed in knots at 10 meters the function is:

$$f_E = 0.35\ (0.9 + 0.1\ w) \qquad \text{mm day}^{-1}\text{mb}^{-1} \quad (34)$$

Conversion to the appropriate units utilized the relationship presented by Morton (1965), which showed that the wind speed per day at the pan rim is approximately ten times the wind speed per hour at 10 meters.

The absorbed insolation for a Class A evaporation pan differs from that of a flat surface with the same albedo because of the effects of the rim, wall and bottom. These effects were evaluated from the data presented with the *Lake Hefner Studies* (1954) for the two pans located at the South and Southwest stations. Average energy com-

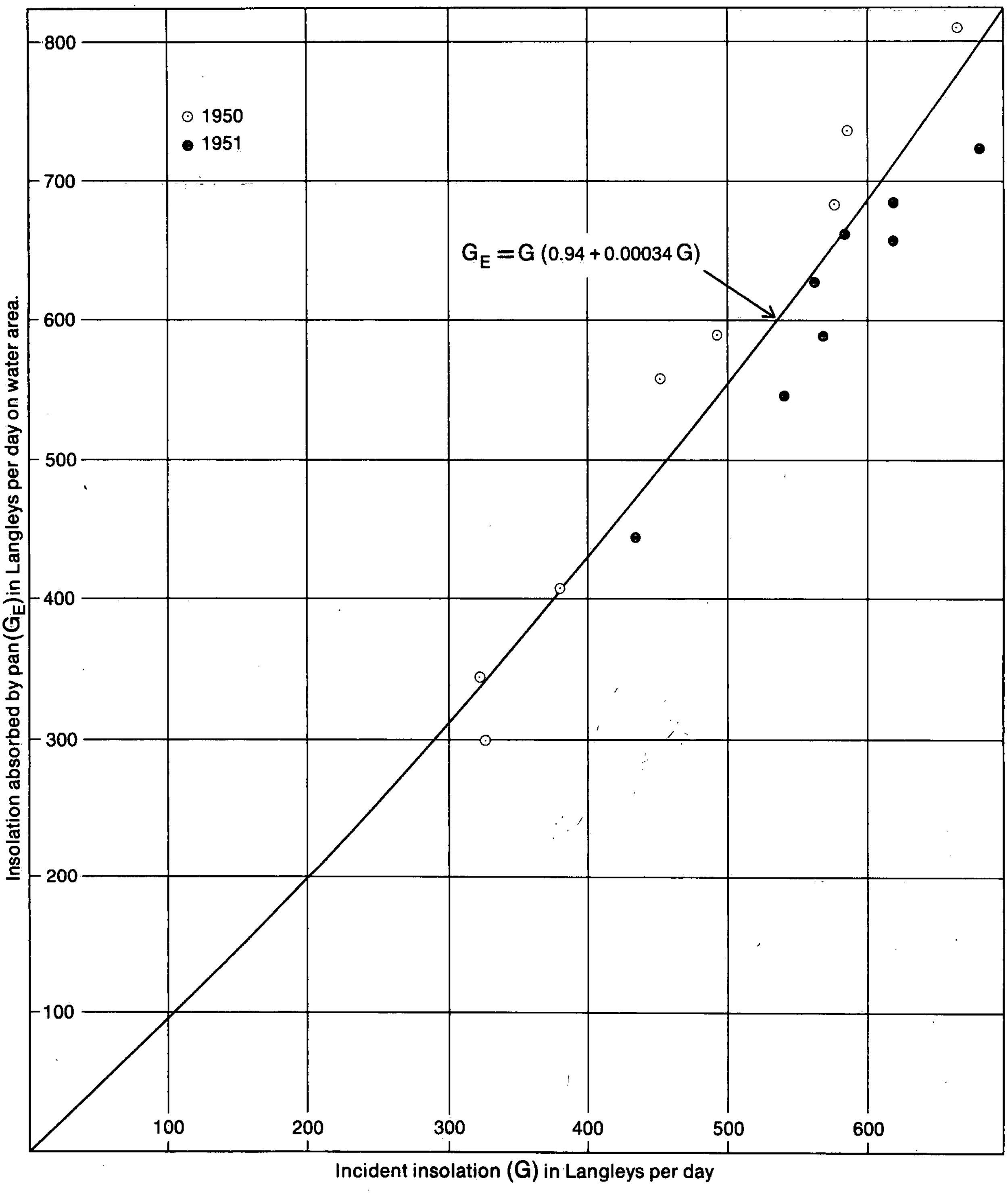

Figure 4 Absorbed insolation for Class A evaporation pans

ponents for the two pans were assembled for each day that evaporation, surface temperature, air temperature, wind, incident insolation and atmospheric radiation observations were available, a total of 167 days in all. The radiant heat transfers to space were estimated from surface temperature and atmospheric radiation observations and the changes in heat storage were estimated from changes in daily minimum surface temperatures. By grouping the daily data in such a way that the total sensible heat transfer, as computed by adding the daily values of

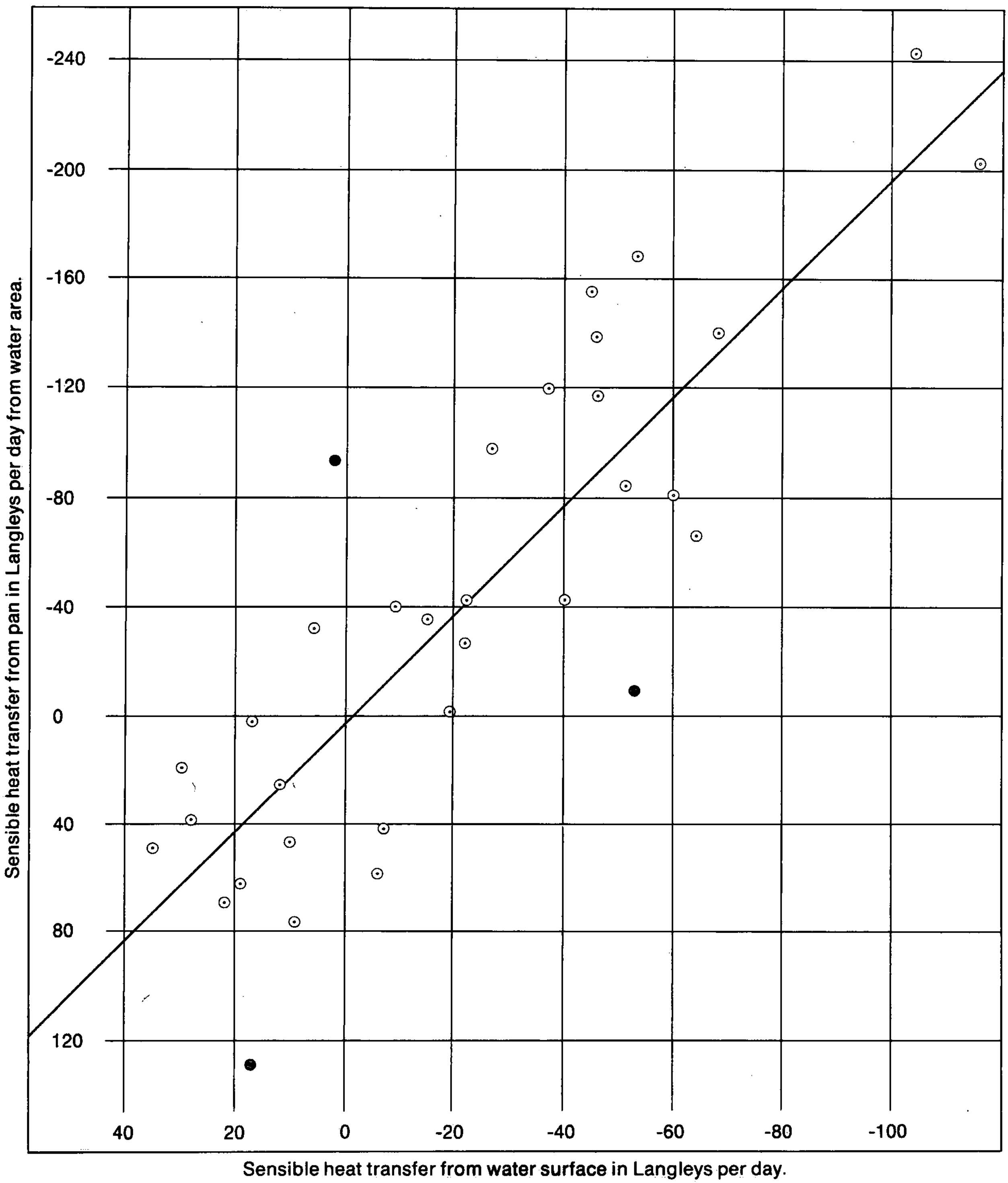

Figure 5 Sensible heat transfer from Class A evaporation pans

$2apf_E(T_S - T)$, was roughly equal to zero, the energy budget for each group provided an estimate of the absorbed insolation. The days used in each group were not necessarily consecutive but they were not allowed to overlap with those of another group. The 16 groups thus selected represented a total of 96 days and varied in size from a minimum of four days to a maximum of nine days. The absorbed insolation for these groups, as derived by the energy budget, is plotted against the observed incident insolation in Figure 4. The absorbed insolation averages 12 per cent higher than the incident insolation, but the difference seems to be some function of the latter quantity. This dependence on incident insolation seems reasonable, because during the winter the shading effect of the pan rim is accentuated and during cloudy periods the diffuse radiation provides little energy to the pan walls. The least squares curve shown of Figure 4 provides the simplest possible recognition of this tendency. In the appropriate units, with a regional albedo of 0.18, the least squares equation is:

$$G_E = G_R(1.14 + 0.03\ G_R) \qquad \text{mm day}^{-1} \quad (35)$$

There are three qualifications to the use of Equation 35. These are: that it does not differ from $G_E = 1.12\ G = 1.37\ G_R$ at the 10 per cent level of significance; that the incident insolation in Ireland is less than that used to define the equation for about seven months of the year; and that the data for 1951 in Figure 4 plot lower than the data for 1950 thereby indicating a possible time trend. However, the lack of a realistic alternative provides adequate justification for the use of Equation 35 to estimate the insolation absorbed by Class A evaporation pans, despite these qualifications.

The sensible heat transfer characteristics of the wall and bottom of a Class A pan cannot be evaluated from *a priori* reasoning. The metal areas exposed to the air are 150 per cent of the water surface area but, because they include the bottom area between the wooden supports and the inner and outer sides of the pan rim above the water level, their effectiveness is problematic. However, the effective ratio of sensible heat transfer area to vapour transfer area, n, may be estimated from the energy components for the Lake Hefner pans which were used to formulate Equation 35. The sensible heat transfers for 33 pentades of consecutive days were computed from the pan energy budgets using the curve shown in Figure 4 to estimate the absorbed insolation. These were plotted against the pentade values of $apf_E(T_S - T)$, the sensible heat transfer from the water surface, in Figure 5. The coefficient of correlation is 0.84, but the elimination of the three solidly marked outliers would increase the coefficient to 0.90. The least squares line shown in Figure 5 passes near the origin and has a slope of 2.00 with 10 per cent confidence limits of 1.62 and 2.38. The elimination of the three outliers would change the slope by 2 per cent. Therefore, despite the wide confidence limits, it seemed justifiable to use the slope value of 2.00 as the best estimate of n, the ratio of sensible heat transfer area to vapour transfer area for Class A pans.

(d) Thornthwaite type grass evaporimeters

The Thornthwaite type grass evaporimeters are metal containers, 560 millimeters in diameter and 760 millimeters deep, sunk in the soil with their rims slightly above the surrounding soil surface. They are filled with soil and grass is grown on the surface. The grass, together with the grass growing on the surrounding soil is kept as short as possible. A drain pipe leads from the bottom of the evaporimeter to a pit where the seepage is measured. Each morning water is sprinkled on the evaporimeter until seepage starts. The evaporation for the preceding day is equal to the amount of water applied less the measured seepage plus the rainfall measured at a nearby standard rain gauge. This calculation assumes that the amount of water stored in the soil is the same each time that seepage starts, an assumption that could cause error for daily observations although the errors would cancel out for periods of greater length.

In this investigation it was assumed that a Thornthwaite Type grass evaporimeter has an albedo equal to that of grassland, turbulence producing characteristics similar to those of a Class A pan, and no heat transfer through the walls and bottom. With these assumptions the absorbed insolation is 93 per cent of that for a region, the vapour transfer function may be estimated from Equation 34, and the ratio of sensible heat transfer area to vapour transfer area, n, is equal to one.

Section 5

Catchment Evaporation

The model, as formulated in Equation 24, was tested by a comparison of the model predictions with annual values of the difference between rainfall and runoff for six Irish river catchments over the decade 1952 – 1961. The rivers are the Shannon, the Erne, the Boyne, the Liffey, the Barrow and the Lee. They were selected because of their size, which justified ignoring percolation to or from adjacent catchments, and because of the availability of rainfall and river flow data. During the selected decade there were only eight years of homogeneous river flow records for the River Lee, and only one year of continuous river flow records coinciding with adequate contiguous meteorological observations for the River Barrow. The locations of the rivers are shown in Figure 1.

Estimates of calendar year rainfall on the Shannon, Boyne, Liffey, and Lee catchments were supplied by the Electricity Supply Board of Ireland, as were those on the Erne catchment for the first four years of the decade. Those for the four catchments first mentioned were derived from Theissen polygons while those for the Erne were derived from isohyetal maps. These data were supplemented by isohyetal estimates for the Erne catchment during the last six years of the decade and for the Barrow catchment during 1958.

Estimates of annual runoff from the catchments were based on monthly river flow records supplied by the Electricity Supply Board. The records for the Shannon and Erne were derived from machine and spillway rating curves while those from the Boyne, Barrow, and Lee were derived from stage-discharge relationships. The runoff for the Liffey was computed from the average of the recorded flows at the Celbridge gauging station and the Leixlip Power House, because of an apparent downstream decrease in flow in the short distance between the two measurement sites. It was assumed that runoff resulting from the calendar year rainfall and evaporation is equal to the average calendar year river flow plus an adjustment equal to 10 per cent of the change in the average January flow during the year, or to the equivalent of 50 millimeters on the catchment, whichever is less. This adjustment, which is roughly equivalent to a 37-day lag between the rainfall year and the runoff year, was used to make allowance for the change in surface and sub-surface storage during the rainfall year, i.e., during the calendar year. Although arbitrary, it is probably adequate for the Barrow and Boyne, with headwaters in the Bog of Allen and other large bogs and for the Shannon and Erne, with large lakes and bogs in their catchments. However, the adequacy of the adjustment for the Liffey and Lee catchments is more doubtful, because of the large proportion of mountainous area in their catchments. It should be noted that adjustment-induced errors for consecutive years are compensatory and thus have little significance for periods of more than two years.

The calendar year model predictions were based on monthly values for individual weather stations, as computed from Equation 24 and the empirical approximations of Sections 4(a) and 4(b). Appropriate records of air temperature, relative humidity, 10-meter wind speed, and ratio of observed to maximum possible sunshine duration were obtained from the *Monthly Weather Reports* of the Meteorological Service. If, as sometimes happened during the winter months, the use of Equation 24 resulted in model predictions greater than one half the absorbed insolation, it was assumed that the latter value was correct. The predictions for the Shannon catchment were obtained by applying the Theissen polygon weights of 0.49, 0.19, 0.23, and 0.09 to the annual values for the weather stations at Mullingar, Shannon Airport, Claremorris and Clones respectively. The predictions for the Liffey catchment were obtained from the annual values averaged for Mullingar and Dublin Airport, while those for the Erne, Boyne and Barrow catchments were obtained from annual values at individual weather stations, i.e., at Clones, Mullingar and Kilkenny respectively. For the Lee catchment, the predictions were based on the Mallow air temperature and sunshine records and on the Shannon Airport relative humidity and wind speed records. In the computations for the Lee catchment, the Shannon Airport wind speeds were increased by 1.5 knots, an adjustment indicated by a graphical comparison with monthly values at Cork Airport for the years 1962 to 1965 inclusive.

Table 3 summarizes the relevant water and energy budget components for the six catchments. It presents annual values of the components for the Shannon, Erne, Boyne and Barrow catchments and two year average values for the Liffey and Lee catchments. This latter arrangement is justified by the relatively poor accuracy of the rainfall, storage effect and energy component estimates for the Liffey and Lee catchments, and by the relatively poor accuracy of the Liffey runoff estimates as highlighted by the apparent downstream decrease in flow from Celbridge to Leixlip. The use of two year averages decreases erratic storage error due to fast runoff and snow accumulation, and reduces by one half the weight given to the data in any analysis.

The data in Table 3 is used to provide two estimates of the river basin evaporation, i.e., rainfall less runoff (P - R) and the model predictions (G_R - Ep). The rainfall has a large potential error which is caused by both the method of measurement and the use of isohyetal maps or Theissen polygons to determine areal values from point measurements. The potential error in the runoff is due to small but consistent error in the river flow measurements and to changes in surface and sub-surface storage not accounted for by the rather crude adjustment discussed previously. The use of isohyetal maps or Theissen polygons and the use of the adjustment to account for storage changes may be the cause of much error in mountainous areas such as those that make up such large proportions of the Liffey and Lee catchments. The potential error in the model predictions, i.e., the absorbed insolation less potential evaporation, is due to error in the weather observations, the inadequacy of the empirical approximations of Sections 4(a) and 4(b) and the inadequacy of the basic concept of Equation 21. The former two sources of error could be quite large but are diminished in importance by the use of Equation 24, which has error attenuation characteristics somewhat similar to those discussed in Section 3(a) as qualities of Equation 11. The last source of potential error, the inadequacy of the basic concept, may or may not exist, and this of course is the reason for the test.

A statistical analysis of the 40 sets of data in Table 3 indicates that there is no significant simple correlation between rainfall less runoff, P - R, and net radiation at air temperature, G_R - B. This may be partially explained by reference to Equation 24, in which regional evaporation depends more on the sum of absorbed insolation and net radiation to space at air temperature than on the difference between them. The same analysis shows that rainfall less runoff is related to both absorbed insolation and potential evaporation with coefficients of correlation equal to 0.34 and -0.57 respectively. Although significant at the 10 per cent level these simple correlations are not adequate for prediction. However, a multiple correlation provides much improved predictive power with a coefficient or multiple correlation of 0.86 and a probable error of 18 millimeters per year. The equation is:

$$P - R = 237 + 0.886\ G_R - 1.180\ Ep \qquad \text{mm year}^{-1} \qquad (36)$$

Equation 36 is the statistical equivalent of Equation 21. As such it provides good evidence for the validity of Equation 21, because

TABLE 3

Water and Energy Budget Component for Catchments

RIVER Gauging Station	Catch-ment Area Km^2	Year or Two Years Averaged	RAINFALL		RUNOFF		Rainfall less Runoff	Absorbed Inso-lation	Net Radiation at Air temperature	Potential Evapor-ation	Model Pre-diction
				P		R	P – R	G_R	G_R – B	E_p	$G_R - E_p$
			in yr^{-1}	mm yr^{-1}	$m^3 sec^{-1}$	mm yr^{-1}	mm yr^{-1}	mm yr^{-1}	mm yr^{-1}	mm yr^{-1}	mm yr^{-1}
SHANNON Killaloe	10410	1952	32.59	828	122.4	372	456	1088	650	606	482
		1953	32.83	834	122.8	372	462	1107	673	620	487
		1954	47.83	1215	246.5	747	470	1048	641	610	438
		1955	35.30	897	144.7	438	459	1144	680	634	510
		1956	40.25	1022	177.1	538	484	1083	661	620	463
		1957	38.5	978	171.4	519	459	1105	667	628	477
		1958	43.42	1103	204.7	620	483	1021	627	563	458
		1959	39.2	996	165.4	501	495	1134	677	634	500
		1960	47.0	1194	237.8	722	472	1099	659	609	490
		1961	40.8	1036	185.9	563	473	1044	632	595	449
ERNE Bally-shannon	4350	1952	35.40	899	68.6	499	400	1040	623	619	421
		1953	34.20	869	64.8	470	399	1050	641	629	421
		1954	51.22	1301	130.9	949	352	991	613	630	361
		1955	37.28	947	76.6	555	392	1081	646	643	438
		1956		1064	87.2	634	430	1029	633	628	401
		1957		1109	94.5	685	424	1064	642	615	449
		1958		1182	101.5	736	446	985	606	544	441
		1959		1048	78.3	568	480	1084	650	602	482
		1960		1206	106.4	774	432	1049	639	567	482
		1961		1129	96.3	698	431	990	600	590	400
BOYNE Slane Castle	2490	1952	31.55	801	20.97	267	534	1082	642	576	506
		1953	29.36	746	19.82	251	495	1114	675	606	508
		1954	44.02	1118	49.69	630	488	1042	635	583	459
		1955	31.82	808	22.05	280	528	1169	690	622	547
		1956	32.64	829	24.72	314	515	1101	668	607	494
		1957	36.61	930	30.70	389	541	1110	664	614	496
		1958	42.44	1078	43.15	547	531	1030	625	547	483
		1959	34.14	867	27.90	354	513	1158	682	629	529
		1960	44.87	1140	45.79	582	558	1122	666	688	514
		1961	33.35	847	25.51	324	523	1050	633	575	475
LIFFEY Celbridge	808	1952-3	33.56	852	9.75	381	471	1119	670	647	472
		1954-5	42.85	1088	15.62	609	479	1126	674	647	479
		1956-7	40.18	1021	13.48	527	494	1124	678	658	466
		1958-9	42.90	1090	15.87	619	471	1116	664	613	503
		1960-1	44.80	1138	17.49	684	454	1112	662	650	462
LEE Iniscarra	793	1952-3	47.58	1208	20.11	800	408	1102	683	703	399
		1954-5	57.14	1452	26.96	1073	379	1116	690	728	388
		1956-7	57.42	1458	27.11	1079	379	1100	682	712	388
		1957-8	64.85	1647	32.20	1281	366	1108	686	715	393
BARROW Craigue-Mamanagh	2800	1958		1101	47.63	537	564	1057	652	539	518

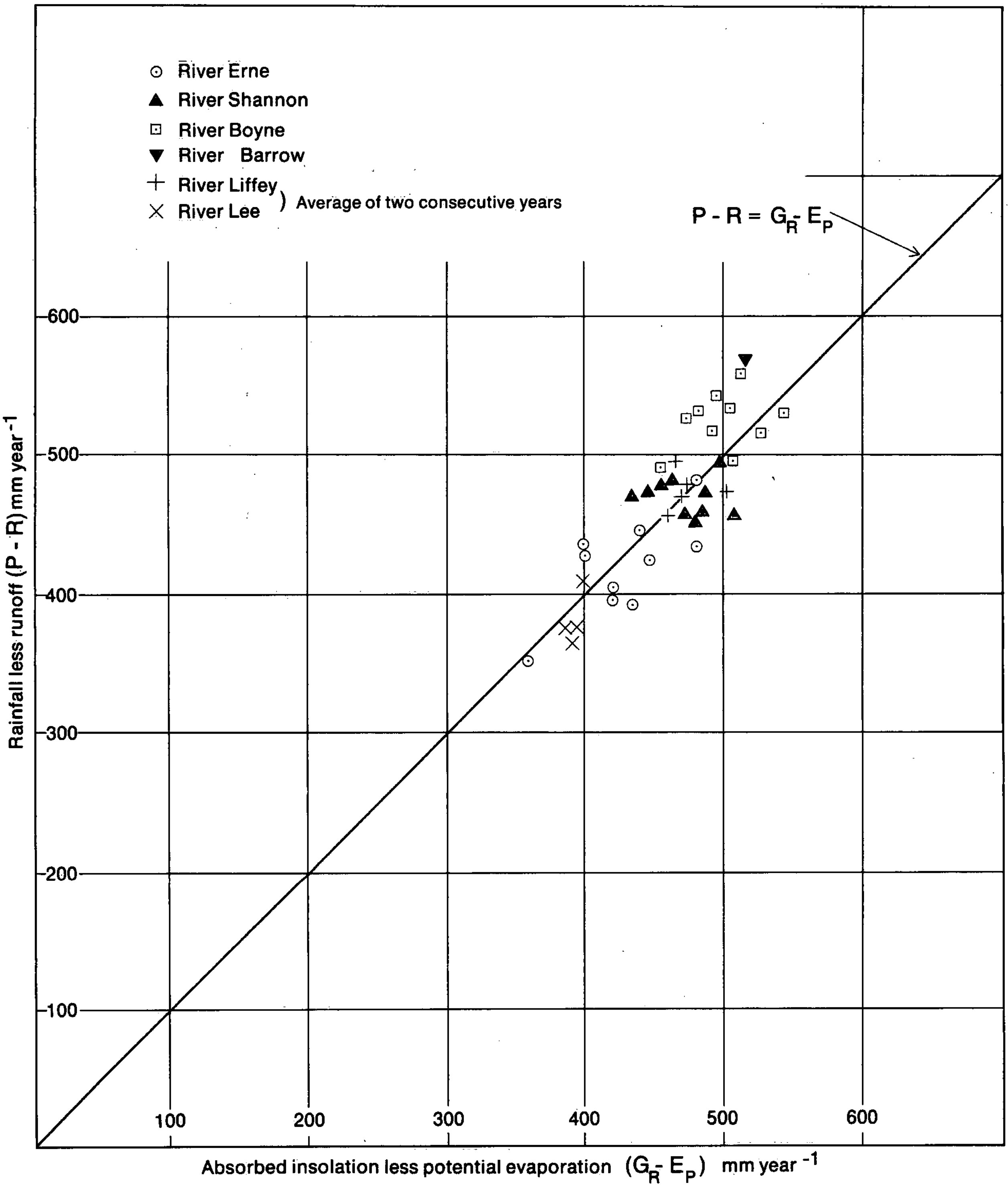

Figure 6 Evaporation from river catchments

the partial regression coefficients, 0.886 and -1.180, do not differ from the conceptual values of 1.00 and -1.00 at the 10 per cent level of significance.

Figure 6 shows the differences between rainfall and runoff plotted against the differences between absorbed insolation and potential evaporation, i.e., the model predictions. The coefficient of correlation is 0.84. The data plot evenly about the one-to-one line with a maximum error of 51 millimeters per year, or 10 per cent, and a probable error of 19 millimeters per year, or 4 per cent. As the probable error is of the same general magnitude as the error in estimating rainfall, it is of little use in evaluating error in the model predictions. It may be noted from Figure 6 that the deviations from the model predictions are not random, but are generally positive for the Boyne and negative for the Erne. Such non-random deviations could be due to consistent errors in the river flow records, to unrepresentative weather observations, or to catchment albedos which differ from each other and from the assumed value.

The comparison in Figure 6 is between two completely independent estimates of regional evaporation. Not one of the empirical approximations used in the model predictions is dependent in any way on the values of rainfall less runoff, and in fact only one, that between the cloud and sunshine ratios in Equation 30, was derived from Irish data. Thus, in evaluating the validity of the model, the favourable results of the statistical analysis are out-weighed by the excellent visual agreement between the rainfall less runoff values and the absorbed insolation less potential evaporation values shown in Figure 6. With such agreement there can be little doubt that the concept of Equation 21 and the model of Equation 24 reflect physical reality with a fidelity that is as good as, if not better than, the rainfall less runoff estimates.

Section 6

Pan Evaporation

The Class A evaporation pan network in Ireland has been expanded progressively since its inception in March 1962. The pans are located at stations of the Meteorological Service, An Foras Taluntais (the Agricultural Institute), Bord na Mona (the Peat Board), the Electricity Supply Board, the Forestry Division of the Department of Lands, the Office of Public Works, and at two colleges of the National University of Ireland. The records are processed and distributed to interested organizations or individuals by the National Committee for Geodesy and Geophysics. By September 1967, the last month included in this study, there were twenty-four pans in the network, although two had been in operation for less than five months. Records for the latter two paris and for the seven pans lacking nearby sunshine observation were not given any weight in this investigation. They include three in very sheltered locations and three in partially sheltered locations. Significant details of the fifteen pans whose records are analyzed herein are summarized in Table 4 and their locations are shown in Figure 1.

Data from four of the pans listed in Table 4 showed sharp discontinuities during 1964. For the pan at Boora the discontinuity could have been due to movement of the pan from a sheltered to a slightly more exposed location, and for the pans at Bellacorrick and Glenamoy the discontinuities could have been due to lowering of the pan water levels to prevent water from spilling out during the high winds experienced so frequently in northwest Mayo. Therefore the data for the three pans prior to the discontinuities are not presented. The discontinuity for the Rosslare pan in 1964 was not so extreme and no good reason for it can be suggested. Therefore, the data both before and after the discontinuity are presented, and the earlier period is distinguished by an asterisk, e.g., Rosslare*.

The earlier pan observations at Lanesboro are not presented because of the lack of any weather observations at the site. The analysis that is presented begins in June 1965, when temperature observations became available, although sunshine observations from a different station had to be used until a sunshine recorder was established at the site in February 1966.

In analyzing pan records, no computations were made for months with more than 40 per cent of their daily observations missing. Furthermore, the analyzed data for particular months were omitted when the observations were obviously in error. The number of outlier months thus omitted is shown in Table 4. The total number is 24 or ar average of 1.6 per pan.

The monthly pan evaporation observations were analyzed in two steps. In the first step they were compared with predictions derived from weather observations using Equation 11 and the applicable empirical approximations of Sections 4(a) and 4(c). In the second step the potential evaporation was derived from pan observations using Equation 17 and the applicable empirical approximations, and then compared with the absorbed insolation. The required weather obser-

TABLE 4

Evaporation Pan Location	Latitude North	Longitude West	Period begins	Months omitted		Stations used for Estimating			
				Insufficient Observations	Outliers	Relative Humidity		Wind Speed	
INLAND									
Lullymore	53°17′	6°58′	July '63	0	0	Dublin A	Birr	Lullymore	–
Lanesboro	53°40′	7°57′	June '65	0	0	Clones	Birr	Clones	–
Ballinamore	54°04′	7°47′	July '63	1	2	Clones	Mullinger	Clones	–
Fermoy	52°10′	8°16′	Oct. '64	1	0	Cork A	Kilkenny	Cork A	Kilkenny
Boora	53°14′	7°44′	March '64	0	0	Mullinger	Birr	Birr	–
COASTAL									
Valentia	51°56′	10°15′	March '62	2	2	Valentia	–	Valentia	–
Rosslare*	52°15′	6°20′	March '62	0	1	Rosslare	–	Rosslare	–
Rosslare	52°15′	6°20′	July '64	0	0	Rosslare	–	Rosslare	–
TRANSITIONAL									
Dungarvan	52°15′	7°39′	Feb. '64	0	4	Cork A	Roches Pt.	Cork A	–
Johnstown Castle	52°18′	6°31′	April '62	3	2	Cork A	Rosslare	Cork A	Dublin A
Malahide	53°25′	6°10′	May '62	2	5	Dublin A	Kilkenny	Dublin A	–
Ballyshannon	54°30′	8°12′	March '63	4	2	Belmullet	Clones	Clones	–
Glenamoy	54°14′	9°43′	June '64	3	0	Belmullet	Claremorris	Belmullet	–
Bellacorrick	54°07′	9°34′	June '64	0	4	Belmullet	Claremorris	Belmullet	–
University College Galway	53°17′	9°03′	Sept. '66	0	0	Shannon A	Claremorris	Claremorris	–
University College Cork	51°54′	8°29′	March '62	3	2	Cork A	Kilkenny	Cork A	Kilkenny

vations were abstracted from the *Monthly Weather Reports* or from the records of the Meteorological Service. Sunshine duration observations were made at or near all of the pans. Average temperatures were available for the pans at Valentia and Rosslare and averages of the maximum and minimum temperatures were available for twelve other pans. For the exception, Ballyshannon, records of maximum and minimum temperatures at Glenties were used. Average relative humidities and 10-meter wind speeds were available for the pans at Valentia and Rosslare and 2-meter wind speeds were available for the pan at Lullymore. For the other pans, estimates of average relative humidities were made by averaging values at two of the most suitably situated observing stations, and estimates of 10-meter wind speeds were made from values at either one or two of the most suitably situated observing stations. Details of the humidity and wind data used are summarized in Table 4.

The results of the analysis are shown in Figure 7 to Figure 12 inclusive. Each figure presents the results for two or three pans, all of them at either inland, coastal or transitional locations. In the top panels of these figures the observed monthly evaporation (in millimeters per day of observation) is plotted against the monthly average value computed from weather observations, with the one-to-one line included for purposes of comparison. In the bottom panels the potential evaporation derived from the monthly pan observations is plotted against the monthly average value of regional absorbed insolation and the one-to-two line is included to facilitate further examination of the data.

The points plotted on Figures 7 to 12 inclusive are subject to various types of error. The Class A evaporation pan is not a precision instrument. In addition to observer error there is potential error in the rainfall estimates as it is not known whether the amount of rain intercepted by the pan is the same as that intercepted by the rain gauge. Furthermore, there are other hazards such as spillage during high wind and absorption and splashing by gulls enjoying a fresh water bath. The computed values and the absorbed insolation are also affected by errors in the weather observations and in the empirical approximations of Sections 3(a) and 3(c). Although these two sources of error could be quite large, they are diminished in importance when used in computing pan evaporation by the energy distribution effect of the surface temperature discussed in Section 3(a).

Figure 7 presents the results for three inland evaporation pans. These are located at the Lullymore research centre of An Foras Taluntais, the Lanesboro peat harvesting works of Bord na Mona, and the Ballinamore research centre of An Foras Taluntais. The Lullymore pan is situated in an exposed location on a strip of pasture about 150 meters wide running through a large area of bare peat, part of a bog

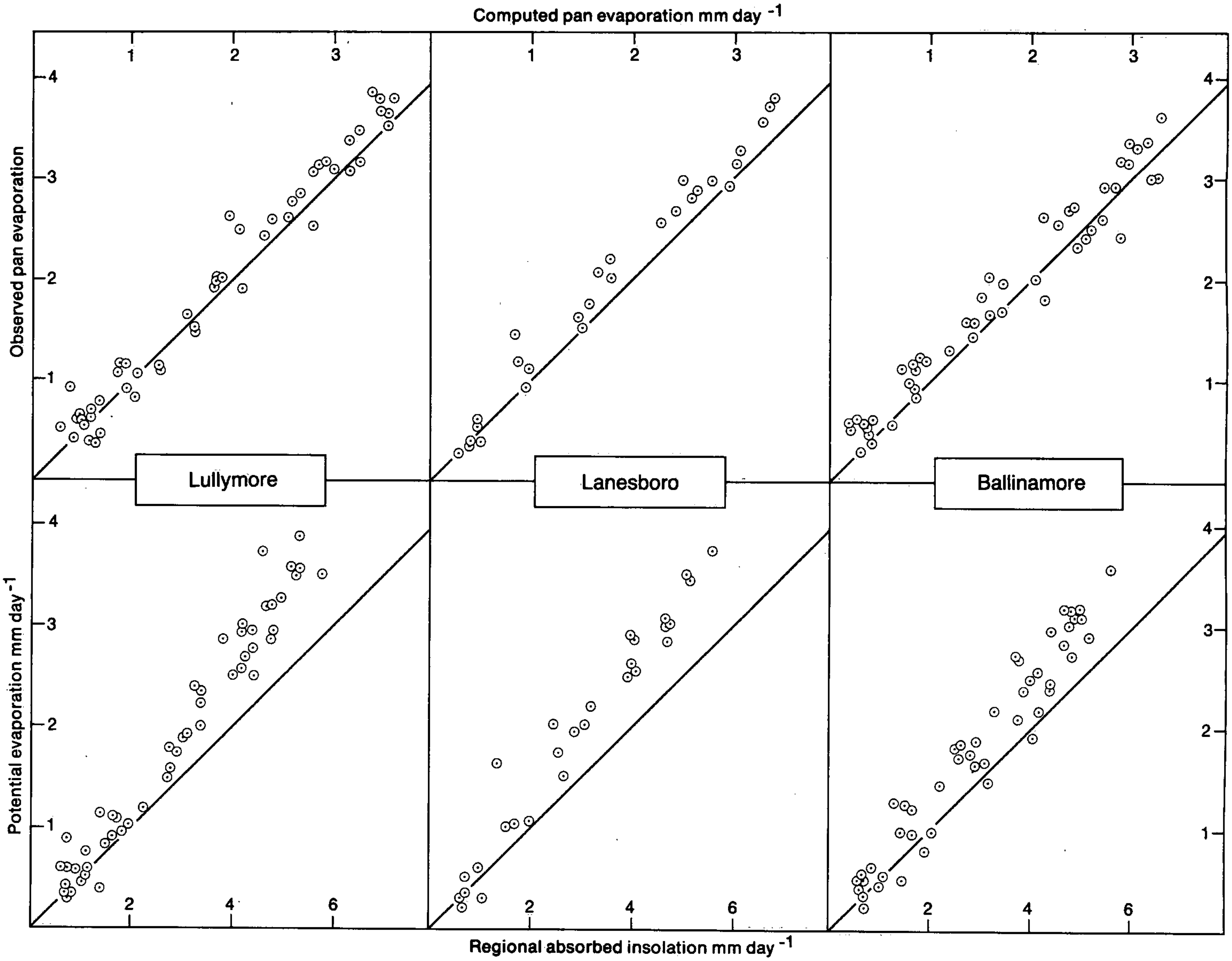

Figure 7 Evaporation from inland Class A pans - Lullymore, Lanesboro and Ballinamore

that has been drained and harvested by Bord na Mona. The Lanesboro pan is in a somewhat similar situation, although it is less exposed because of its proximity to a road, a railway cut and administration buildings. The Ballinamore pan is located near the top of a hillock in a grassed drumlin region. The evaporation observations for the three pans agree quite well with the computed values although there is a tendency for them to plot slightly higher than the one-to-one line. This might be explained by the effects of drainage around the Lullymore and Lanesboro pans and the hillslope location for the Ballinamore pan, which could result in lower relative humidities than those assumed. The significance of the plots of potential evaporation against absorbed insolation in the lower panels of Figure 7 may be evaluated in the light of the discussion on potential and regional evaporation in Section 3(c). Thus the potential evaporation may be thought of as the reflection, on the opposite side of the one-to-two line, of the regional evaporation. During the winter, when the regional evaporation is limited by the available energy to one half of the absorbed insolation, the potential evaporation should scatter about the one-to-two line; whereas during the summer, when the regional evaporation is limited by the available water to less than one half of the absorbed insolation, the potential evaporation should plot above the line. Furthermore, because bare drained peat absorbs more insolation and evaporates less water during dry weather than vegetation, the summer values of potential evaporation for Lullymore and Lanesboro should plot higher above the one-to-two line than those for Ballinamore. The agreement between these implications of Equation 21 and the inland pan data presented in Figure 7 provides additional evidence in support of the concept and model.

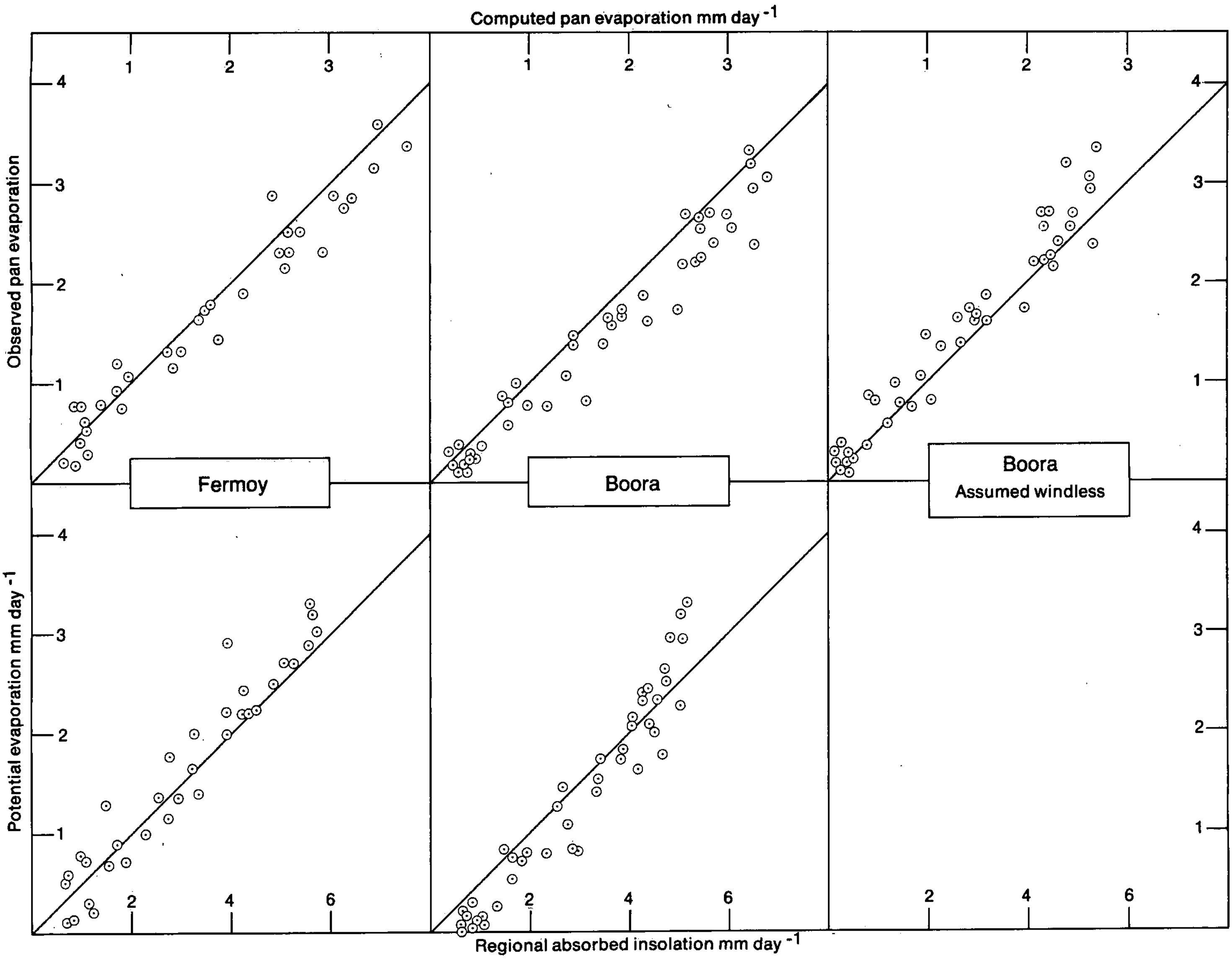

Figure 8 Evaporation from inland Class A pans - Fermoy and Boora

The results of the analysis for the Class A evaporation pans at the Fermoy research centre of An Foras Taluntais and the Boora peat harvesting works of Bord na Mona are presented in Figure 8. Both are inland stations. The pan at Fermoy has an exposed location in an agricultural region, whereas the pan at Boora has a sheltered location amongst trees, hedges and buildings in a region of partially harvested bogs. The observed pan evaporation at Fermoy agrees quite well with computed pan evaporation during the winter months but tends to be somewhat lower during the summer months. The potential evaporation scatters about the one-to-two line when plotted against the absorbed insolation and does not rise significantly above it even in summer. This could be due to continuously moist conditions in the region during the summer, but is more probably caused by the error in observed evaporation suggested by the negative deviations in the top panel. The observed evaporation at Boora plots lower than the computed pan evaporation and this is reflected in the plot of potential evaporation against absorbed insolation. The negative deviations may be attributed to the sheltered location of the Boora pan and the resultant low wind speeds. To illustrate this, the evaporation has been computed assuming windless conditions and the results compared graphically with the observed values in the upper right hand panel of Figure 8. The observed evaporation plots somewhat higher than the computed windless evaporation, but agrees with it more closely than with the evaporation computed assuming full wind effect. Therefore it seems that the pan is not fully sheltered from the wind, although the sheltering does decrease the effect of the regional wind by more than one half.

Figure 9 presents the results of the analysis for two coastal

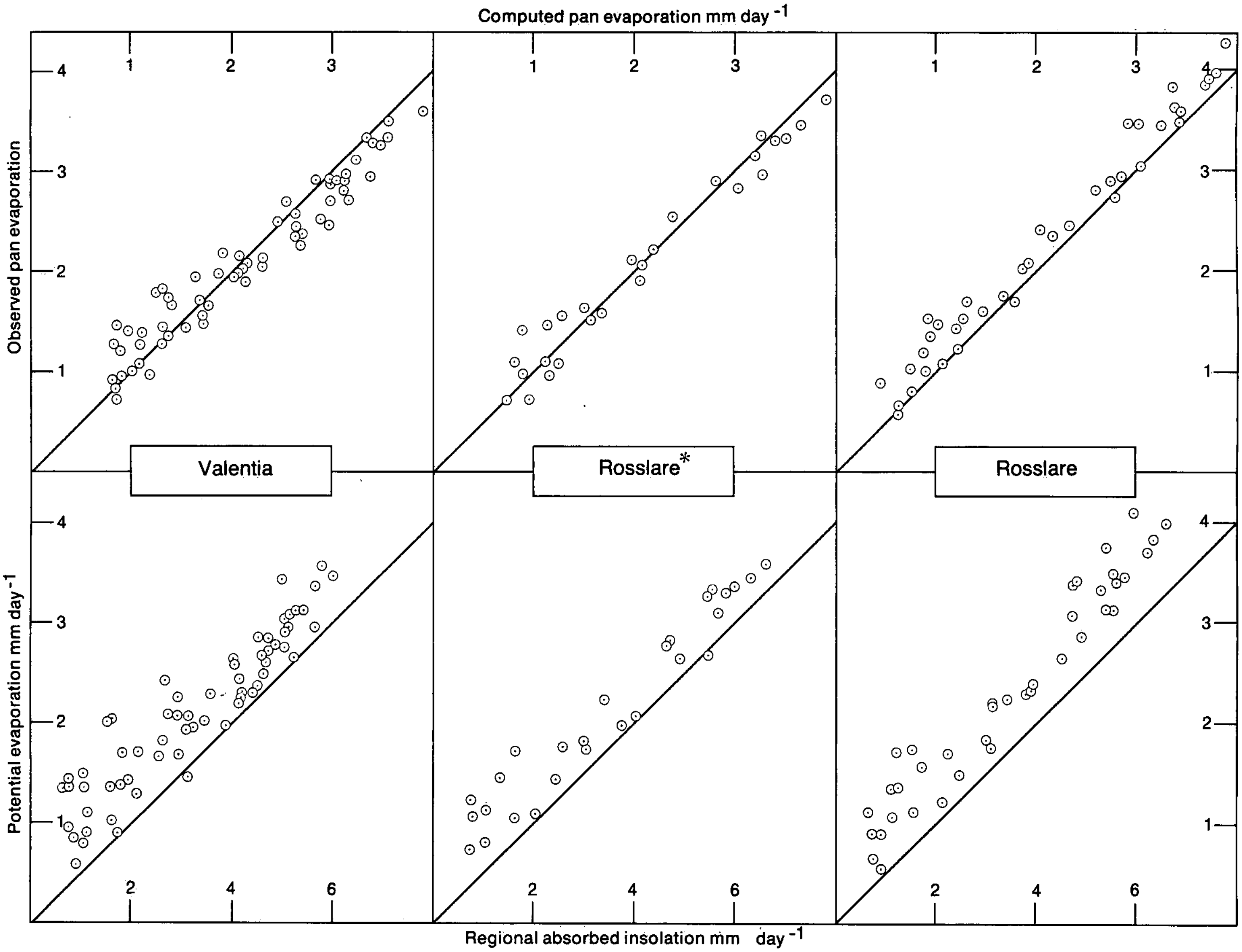

Figure 9 Evaporation from coastal Class A pans

evaporation pans. These have reasonably well exposed locations next to the coast at the Valentia and Rosslare meteorological observatories. The observations at Valentia agree quite well with the computed values although the deviations tend to be positive in the winter and negative in the summer. The negative summer deviations, if real, could lead to some doubt concerning the reliability of the insolation observations at Valentia, because the use of Equation 27, rather than the Valentia based Equation 26, for estimating incident insolation should produce contrary results. The evaporation observations at Rosslare for the period prior to July, 1964 (as distinguished by the asterisk on the station name) agree quite well with the computed values, whereas the observations from that month onward (as distinguished by the lack of an asterisk on the station name) are somewhat higher than the computed values. No reason is known for the change. The plots of potential evaporation against absorbed insolation for both Valentia and Rosslare have a pattern that is distinct from the pattern for inland stations. The winter potential evaporation plots much above the one-to-two line rather than about it and the deviations of the summer potential evaporation from the one-to-two line are less pronounced. The difference in pattern may be attributed to the evaporability, as manifested in heat and vapour pressure deficit, which is advected to the land from the sea during the winter and from the land to sea during the summer.

Figure 10 presents results for three transitional pans located in agricultural areas near the south and east coasts. All are in reasonably

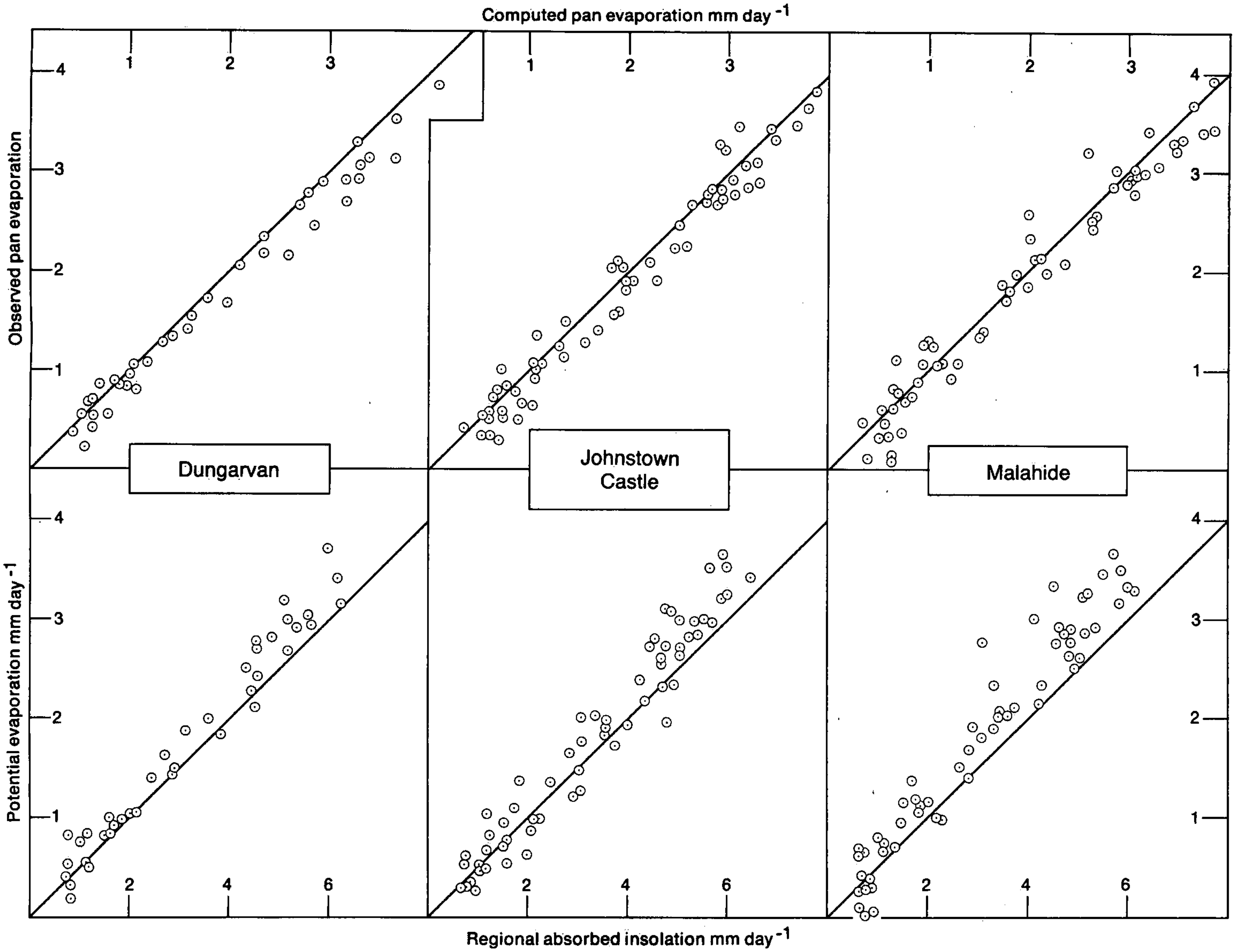

Figure 10 Evaporation from transitional Class A pans - Dungarvin, Johnstown Castle and Malahide

exposed locations at research centres of An Foras Taluntais. The pan at Dungarvan is two kilometers west of Dungarvan Harbour and about seven kilometers from the open sea, the pan at Johnstown Castle is four kilometers west of Wexford Harbour and about seven kilometers from the open sea, and the pan at Malahide is about four kilometers west of the open sea. There is a reasonable agreement between observed and computed evaporation for the three pans, although the observed values tend to be somewhat lower than the computed values. The plots of potential evaporation against absorbed insolation are similar to those for inland pans, i.e., the points scatter about the one-to-two line during the winter and plot higher than the one-to-two line during the summer. Therefore it seems that the effects of advections of evaporability between the sea and land environments are limited to a distance of less than four kilometers from a south or east coast.

The results for three transitional pans located near the west coast are presented in Figure 11. The Ballyshannon pan is at an Electricity Supply Board pumping station about seven kilometers east of Donegal Bay. It is partially sheltered from the north by a three-meter embankment which retains the head pond of the Cathleen Falls power development, but has agricultural land in all other directions. Its characteristics, as presented in Figure 11, are similar to those of the transitional pans near the south and east coasts, with the observed evaporation plotting slightly lower than the computed evaporation, and with the potential evaporation plotting against the absorbed insolation in the same way as for inland pans. The Glenamoy pan is located at an An Foras Taluntais research centre in a windswept area of natural bog. It is three kilometers east of the nearest arm of the sea and is about ten kilometers east of and ten kilometers south of the open sea. There is

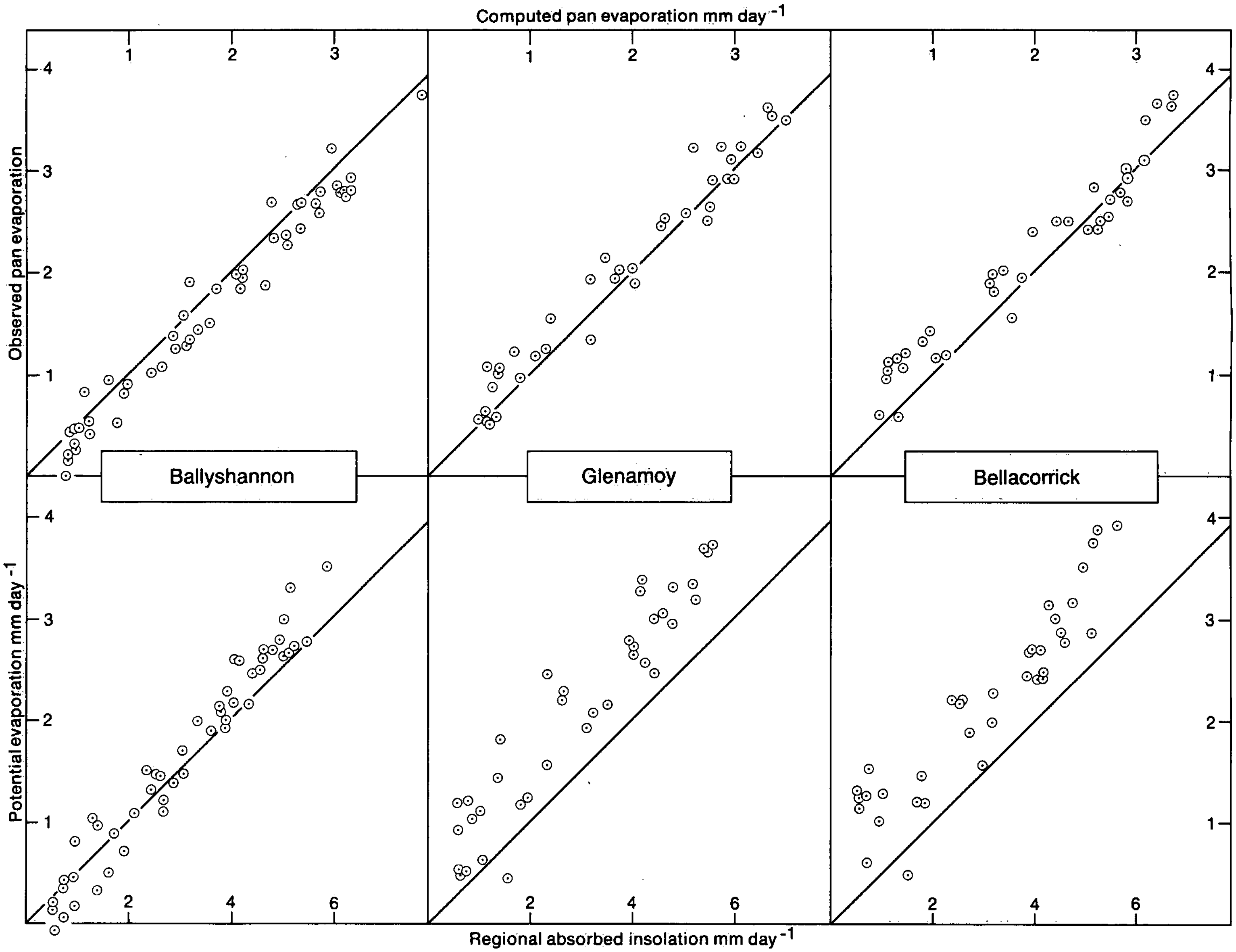

Figure 11 Evaporation from transitional Class A pans - Ballyshannon, Glenamoy and Bellacorrick

reasonable agreement between the observed and computed evaporation although the observed values tend to be somewhat higher. The potential evaporation appears to have coastal characteristics during the winter months, plotting above the one-to-two line, and inland characteristics during the summer months, plotting further above the line than normal for a coastal section. The Bellacorrick pan, located at a Bord na Mona station, in a similar area of windswept bog, has almost identical characteristics even though it is more than twice as far from the open sea in both directions. It is not known why the effects of advections of evaporability should persist as far inland as they appear to do at Glenamoy and Bellacorrick when they are not in evidence at Ballyshannon, Malahide, Johnstown Castle and Dungarvan. If they are real, the answer may be that the pans at Glenamoy and Bellacorrick are more exposed to oceanic influences that the other four transitional pans, with the Gulf Stream impinging directly on the nearby coasts and with only barren bogland to modify its effects.

Figure 12 shows the results for the evaporation pans located at University College, Galway and University College, Cork. Both have transitional locations between coastal and inland environments. The evaporation observations are much lower than the computed values for both pans and this is reflected in the plots of potential evaporation against absorbed insolation. As the Galway pan is reasonably well exposed, no reason for such results can be suggested. However, the results for the Cork pan may be explained by its sheltered location amongst trees and college buildings. This is illustrated by the comparison between the observed evaporation and evaporation computed for assumed windless conditions, which is presented in the upper

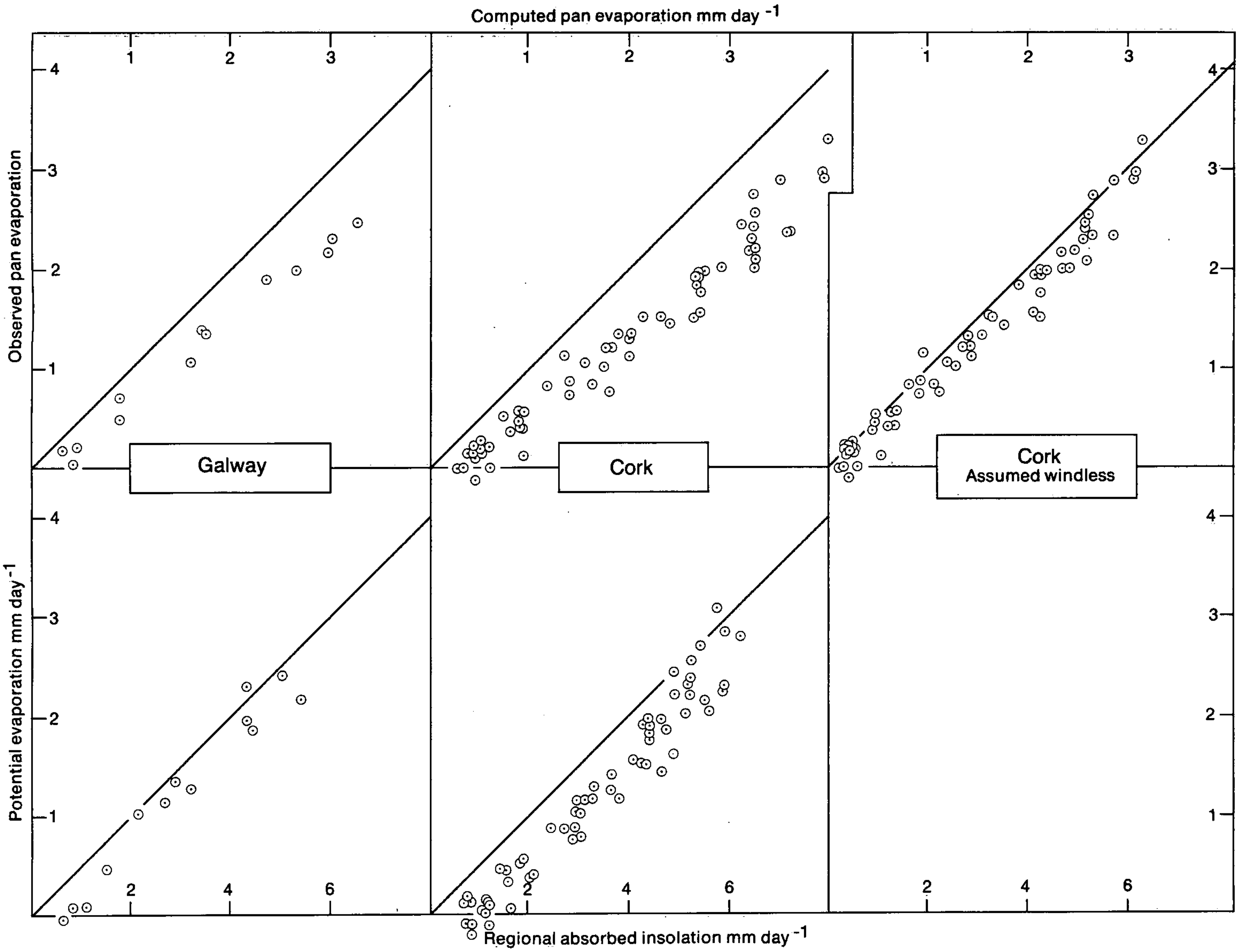

Figure 12 Evaporation from transitional Class A pans - Galway and Cork

right hand panel of Figure 12. With the assumed windless conditions the scatter is reduced and the agreement is reasonably good. However, the observed values still tend to plot below the computed values, and this suggests that some of the sheltering objects may partially shade the pan from the sun.

The conclusions reached from the foregoing analysis of Class A evaporation pan observations may be summarized as follows: -

(1) Reasonable estimates of Class A pan evaporation may be obtained from weather observations using Equation 11 and the empirical approximations of Sections 4(a) and 4(c). Therefore Equations 16 and 17, together with the empirical approximations of Sections 4(a) and 4(b), should provide reasonable estimates of potential evaporation.

(2) The potential evaporation data for the inland pans and the majority of the transitional pans provide supporting evidence for the implications of Equation 21.

(3) The effects of advections of evaporability between sea and land environments are evident in the potential evaporation data for coastal pans.

(4) Normally the effects of advections of evaporability between sea and land environments are not evident more than four kilometers inland. However, in coastal regions with characteristics similar to the regions around Glenamoy and Bellacorrick, the transitional zone may be several times wider.

Section 7

Grass Evaporimeter Evaporation

There are five batteries of Thornthwaite type grass evaporimeters in Ireland. These are located in the same instrument enclosures as the evaporation pans at Valentia, Glenamoy, Johnstown Castle, Malahide and Ballinamore. The records at Valentia, which go back to 1952, were supplied by the Meteorological Service, and the records for Glenamoy, Johnstown Castle, Malahide and Ballinamore, which began late in 1963, were supplied by An Foras Taluntais. No analysis was made of records for the periods before March, 1962 and after September, 1967.

Figure 13 compares the monthly observed evaporation with the monthly computed evaporation for each of the grass evaporimeter batteries at Valentia, Glenamoy, Johnstown Castle and Ballinamore. The observed evaporation is an average for either the four evaporimeters in a battery or as many of the evaporimeters as were in operation. The computed evaporation was derived from Equation 11, the empirical approximations of Section 4(a), and the assumptions of Section 4(d) using the weather observations that were assembled for the analysis of the neighbouring pan. The solid points shown on Figure 13 represent outliers which have been averaged with the observations of either the preceding or succeeding months in order to moderate the effects of end-of-month weather disturbances or observational gaps. The use of two month averages permits the use of all the data, as the outliers do not need to be excluded.

The main reason for analyzing grass evaporimeter records was to find out whether vegetative processes control evaporation from grass when soil moisture is unlimited. The observations from Valentia shown in Figure 13 would indicate that they do, but the observations from Glenamoy provide impressive evidence to the contrary. As the more scattered observations from Johnstown Castle and Ballinamore support the Glenamoy observations and, as the observations from Malahide are too scattered to be of any help in settling the issue, the weight of evidence favours the hypothesis that there are no vegetative controls on evaporation from grass during periods of ample moisture supply.

The foregoing tentative conclusion implies also that Equation 11, with the empirical approximations of Section 4(a) and the assumptions of Section 4(d), provides adequate estimates of grass evaporimeter evaporation from weather observations. Although the observations from Valentia provide evidence to the contrary, this is outweighed by the excellent agreement between the observed and computed evaporation of Glenamoy and the more scattered agreement at Johnstown Castle and Ballinamore.

Section 8

Concluding Discussion

(a) General

The foregoing analyses of catchment and evaporimeter data, together with the analysis of Canadian catchment data presented by Morton (1965), provide impressive support for the concept of Equation 21 and the model of Equation 24. Although observations from a wider climatic range are needed to provide a detailed evaluation of the basic assumptions, there is sufficient evidence available now to indicate that the concept provides a reasonable model of the underlying physical processes. Even with such qualifications, the concept has a number of important implications and applications and these are summarized and discussed in the following sub-sections.

(b) Evaporation and climate

With the perspective provided by Equation 21, climate at a constant altitude is governed by the availability of both solar radiation and water at the regional evaporating surfaces. When the availability of water is the limiting factor the evaporative energy can vary from zero under completely arid conditions to one half the absorbed insolation under humid conditions; whereas when the availability of water is not limiting, the evaporation is limited by the availability of energy to the equivalent of one half of the absorbed insolation. Such changes in evaporation, from zero to one half the absorbed insolation under conditions of increasing water availability, represent considerable amounts of energy and water vapour which cannot help but have a governing influence on the regional climate. This influence is usually reinforced by the tendency for the clouds that bring the water to obscure the sun.

Changes in evaporative energy due to changes in the availability of water are manifested in the climate through sensible heat transfer from the surface to the air. Although a relatively small component of the surface energy balance, the sensible heat transfer to

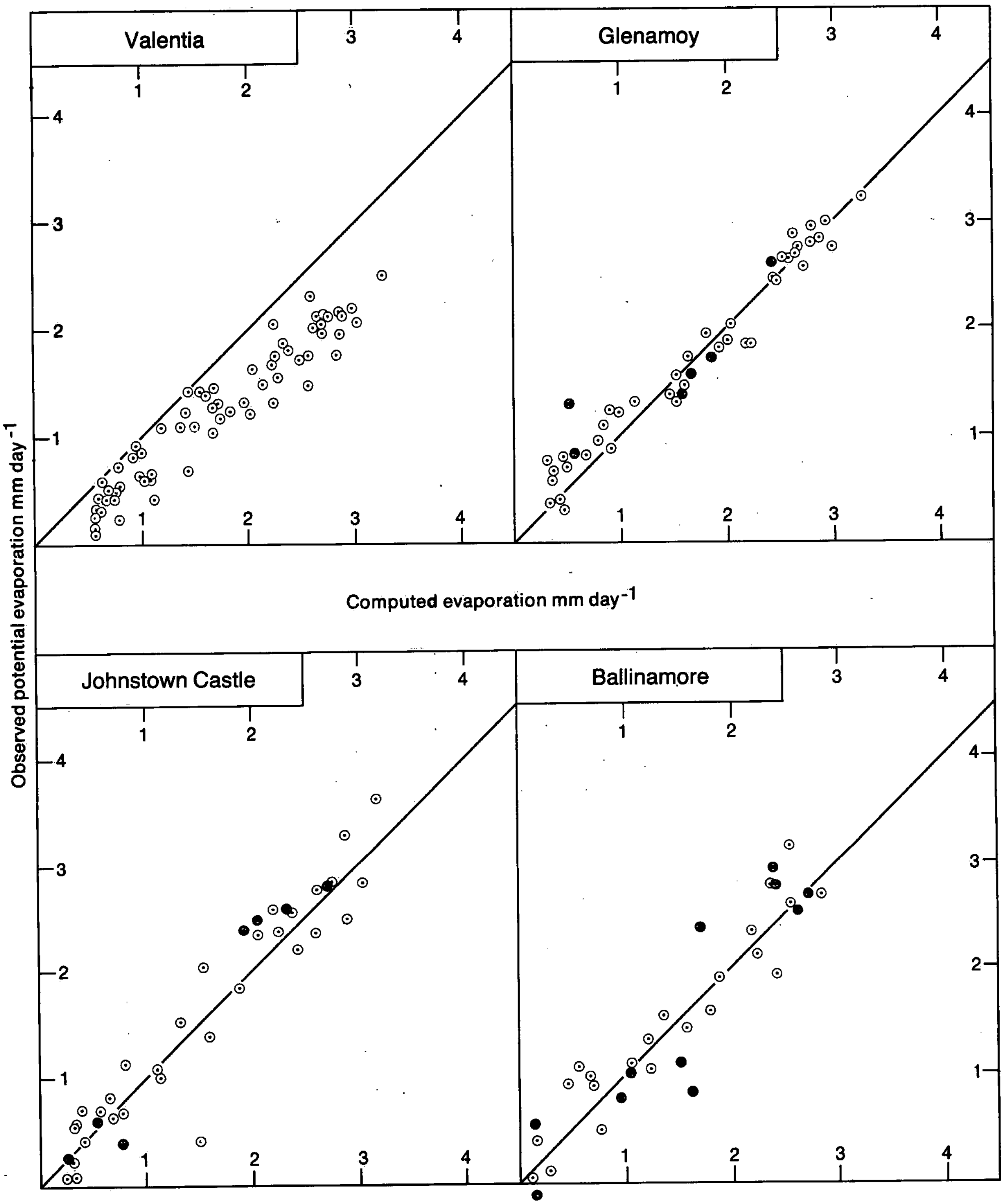

Figure 13 Evaporation from grass evaporimeters

the air appropriates $\frac{ap}{h_R}$, or approximately three quarters, of the evaporative energy made available by a reduction in the availability of water. Changes in the vapour content of the atmosphere are brought about directly by evaporation.

The concept of Equation 21 is based on a simple model of the atmosphere. In the lower atmosphere the vertical transfer of water vapour and sensible heat is largely the result of turbulence generated at or near the surface, whereas in the part of the upper atmosphere below the tropopause it is largely the result of cloud formation. The lower atmosphere is normally 600 to 1000 meters thick but the thickness can vary from less than 300 to more than 3000 meters depending on the intensity of turbulence, the humidity and the temperature. Because the intensity of turbulence varies with the contrast between the surface and the lowest layer of air, the heat and vapour content of the lower atmosphere can adjust quickly to bring the fluxes of heat and vapour from the surface into equilibrium with those to the upper atmosphere. Thus, the lower atmosphere may be regarded as a storage layer with turbulence, humidity and temperature partially controlling the influxes of vapour and heat from the surface and completely controlling the effluxes to the upper atmosphere. In this model the upper atmosphere is simply a sink for vapour and heat from the lower atmosphere, although the clouds formed in it do influence convection at the surface indirectly by reducing incoming insolation, increasing atmospheric radiation and supplying water for evaporation.

Convection of heat and vapour from the regional surfaces to the lower atmosphere takes place in a context of advection from other regions. Short term advections, such as those due to large scale air mass movements, are brought into equilibrium with the regional surface without significant effects on evaporation by the changes in the turbulence of the lower atmosphere mentioned in the preceding paragraph. However, sustained advections, such as those that occur at climatic boundaries, can have significant effects on evaporation in transitional zones where the advections and convections of heat and vapour interact and reach equilibrium with each other. The width of the transitional zone depends on the contrast between the two environments and the nature of the surface. In the change from sea to land environments in Ireland the transitional zone may vary from less than four to more than twenty kilometers as discussed in Section 6, whereas in the change from desert to lake environments discussed in Section 3(c) with reference to the Salton Sea, the transitional zone appears to be about fifteen kilometers.

In the light of the foregoing discussion, it is apparent that the climatic effects of ocean currents are achieved by processes in the upper atmosphere. Thus, heat and water vapour from the Gulf Stream influence the climate throughout Ireland, although its manifestations in the lower atmosphere are not evident outside of a coastal zone that varies from four to twenty kilometers in width. The climate in the interior is to a large extent the result of the clouds formed in the upper atmosphere from water vapour originating in the Gulf Stream. These clouds reduce the main seasonal components of the inland radiation budget, i.e., the incident insolation during the summer and the radiant heat transfer to space during the winter, thereby moderating the temperatures. Furthermore, the ample rainfall resulting from the clouds maintains high evaporation rates which tend to keep the atmospheric humidity high and the summer temperatures low.

The value of the potential evaporation as a climatic indicator may be assessed in the context of the foregoing discussion. The potential evaporimeter absorbs the same amount of insolation and atmospheric radiation as the surfaces of the region. However, the proportions of these energy sources that are used for evaporation, sensible heat transfer to the air and long wave radiation from the surface of the evaporimeter are determined by the heat and water vapour content of the overpassing air, and these are determined by the amount of water that is available for evaporation from the regional surfaces. According to Equation 21, any change in the regional evaporation due to soil moisture or vegetative restrictions on water availability is manifested by an equal and opposite change in the potential evaporation. In a region remote from climate boundaries, the restrictions are equal to the potential evaporation less one half the absorbed insolation, or to one half of the difference between potential and regional evaporation. According to Equation 21 the potential evaporation also can be regarded as the amount of energy remaining from the absorbed insolation after regional evaporation has occurred and, as such, it is equal to the sum of the sensible and radiant heat transfers from the surfaces of the region.

The potential evaporation also provides a measure of the heat and humidity of the lower atmosphere. Thus in a region near a climatic boundary the quantity $E_P + E_R - G_R$ may be used as an index of advected evaporability. Such an index can be used only when an independent estimate of regional evaporation exists, which, in practice, limits its application to arid environments where regional evaporation is zero or to humid environments where regional evaporation should be equal to one half of the sum of the absorbed insolation and any effects of advected evaporability. Class A pan observations are used to detect net advections of evaporability across the Irish coastline in Section 6, taking advantage of the humid conditions that prevail during the winter and the near-humid conditions that prevail during the remainder of the year.

(c) Climatological applications

One of the more important potential uses of the concept of Equation 21 is in estimating the climatic effects of such cultural activities as irrigation or drainage, which change the availability of water for evaporation, and of vegetation changes, which change the availability of both water and energy. For example, the drainage of the Lullymore bog and the removal of the turf has had a significant effect on the climate and this can be evaluated from Class A pan and weather observations, if it is assumed that the bog was a continuously moist region in its natural state and that the stripping process reduced the albedo from 0.15 to 0.12. Averages of observations at or near Lullymore for the month of July 1966 are $T = 14.0°C$, $r = 0.79$, $w = 8.9$ knots, $S = 0.35$, and $E_E = 3.7$ mm day^{-1}. With these observations $d_T = 1.03$ mb °C^{-1}, $f_R = 0.76$ mm mb^{-1} day^{-1}, $h_R = 0.85$ mb °C^{-1}, $B = 1.2$ mm day^{-1}, G_R (with albedo of 0.15) = 5.5 mm day^{-1} and the computed value of $E_E = 3.6$ mm day^{-1}. The potential evaporation, E_P, computed from the observed value of E_E, and Equation 17, assuming an albedo of 0.12, is 3.7 mm day^{-1}.

From previous discussion of Equation 21 it is evident that the decrease in vapour transfer to the air due to the change from continuously moist natural bog to drained bare peat is $-\Delta E_R = \Delta E_P = E_P - 1/2 G_R = 1.0$ mm day^{-1}. The resulting increase in sensible heat transfer to the air due to drying is equal to $\frac{ap}{h_R}\left(E_P - 1/2\ \frac{0.88}{0.85}\ G_R\right)$ and that due to change in albedo is $\frac{ap}{h_R}\ \frac{0.03}{0.85}\ G_R$. The total $\Delta K_R = \frac{ap}{h_R}$

$$(E_P - \frac{0.41}{0.85} G_R) = \frac{0.61}{0.85} \times 1.1 = 0.8 \text{ mm day}^{-1}.$$

The rate of dehumidification of the entire atmosphere is approximately $\frac{0.1\Delta E_P}{0.622}$ or 0.16 mb day^{-1}. The rate of heating of the entire atmosphere is approximately $\frac{980 \times 590 \times 0.1\Delta K_R}{0.24 \times 1{,}000{,}000}$ or 0.19°C day^{-1}. In this computation, 980 is the acceleration of gravity in cm sec^{-2}, 590 is the latent heat of vaporization in cal g^{-1}, 0.1 is in g mm^{-1}, 0.24 is the specific heat of air at constant pressure in cal g^{-1} °C^{-1}, and 1,000,000 is the atmospheric pressure in dynes cm^{-2}. As interest is focused on the equilibrium changes in the temperature and vapour pressure in the lower atmosphere, the rates of change appear to be of little use. However, it is reasonable to assume that the ratio of the equilibrium changes to each other is proportional to the ratio of the rates of change to each other and with this assumption

$$\Delta r_e = -\frac{0.16}{0.19}\Delta T = -0.84\Delta T$$

With the foregoing assumption the values of the equilibrium changes may be obtained by substituting E_P for E_E, h_R for h_E, f_R for f_E, $\frac{0.88}{0.85} G_R$ for G_E, and $\frac{0.03}{0.85} G_R$ for ΔG_E in Equation 14 and assuming the vapour transfer function is unchanged by the drainage and the albedo change. Then $\Delta E_P = 1.0 = 0.55 \times \frac{0.03}{0.85} G_R + 0.38\Delta T - 0.34\Delta r_e = 0.1 + 0.67\Delta T$, so that $\Delta T = 1.3$°C and $\Delta r_e = -1.2$ mb.

From the foregoing calculation it would appear that drainage and peat harvesting of the bog at Lullymore increased the air temperature and decreased the vapour pressure during June 1966 by approximately 1.3 degrees Celsius and 1.2 millibars respectively. Although based on a number of unverified assumptions, the values seem reasonable. The concept of Equation 21 could be used also to evaluate climatic modifications due to irrigation or to vegetation changes if the necessary data were available.

The concept of Equation 21 may be used to classify climatic humidity. The ratio of potential evaporation to absorbed insolation, with the potential evaporation estimated from weather or evaporimeter observations using Equation 16 or Equation 17, provides an excellent index since it should vary from 1.00 under completely arid conditions to 0.50 under completely humid conditions. As an alternative, the ratio of regional evaporation to absorbed insolation is also an excellent index, as it varies from 0.00 under completely arid conditions to 0.50 under completely humid conditions. In the computation of the latter ratio, the regional evaporation may be estimated from weather or evaporimeter observations using Equation 24 or Equation 25, or from the difference between rainfall and runoff.

(d) Engineering applications

River catchment evaporation may be computed from weather observations using Equation 24 and the appropriate empirical approximations and, when subtracted from rainfall, will provide an estimate of the water available for runoff from a catchment large enough for percolation to and from neighbouring catchments to be ignored. For estimates of long term average runoff the difference between average rainfall and average evaporation is adequate. However, to provide a continuous record of monthly river flows, the differences between monthly rainfall and monthly evaporation must be routed through catchment storage in a manner similar to that proposed by Dooge (1961).

The foregoing procedure would be used when there are no river flow records available but there are records of rainfall, temperature, relative humidity, wind and either sunshine duration or insolation. However, in continuously humid regions, such as equatorial rain forests, only rainfall and either sunshine or insolation records are necessary since the river basin evaporation is equal to one half the absorbed insolation.

Water in storage beneath the surface is either soil moisture or ground water, with the soil moisture defined as the water that can be brought back to the surface by capillarity or some vegetative process. Changes in soil moisture may be computed as the difference between rainfall (or snow melt) and the sum of the evaporation, surface runoff and ground water recharge. As the latter two quantities are not significant until the soil moisture nears the storage limit, the basic problem is that of estimating evaporation. Therefore, Equation 24, with the appropriate empirical approximations, provides a means for keeping a continuous soil moisture budget from records of weather observations and a knowledge of the storage capacity. Such a budget can be used for river flow forecasting, irrigation scheduling, trafficability studies or to provide estimates of ground water recharge.

The implications of Equation 21 should prove helpful in estimating the limiting evaporation from an irrigated area. Thus for a small isolated plot the limiting evaporation would approach the potential evaporation, whereas for larger areas it would decrease to some value between the potential value and the humid regional value of one half the absorbed insolation. Of course the actual amount of water supplied is a matter of balancing increasing crop yields with increasing costs, but the evaporation from a continuously moist surface plus desirable percolation does set an upper limit.

(e) Scientific applications

One of the more important potential applications of the concept of Equation 21 is in the field of research on evaporation. It provides a coherent frame of reference against which the effects of evaporating area, water availability, heat storage and proximity to climatic boundaries may be assessed. Such evaluations are exemplified by the discussion on the effects of changes in water availability in the Lullymore Bog in Section 8(c), by the discussion of the effects of heat storage in Lake Superior in Section 3(c), and by the discussion on the effects of a coastal climatic boundary on pan evaporation in Section 6.

As has been stressed repeatedly throughout the foregoing sections, the size of an evaporating area is one of its more important characteristics. Thus the evaporation from a moist surface can vary from the potential value for a very small area down to one half the absorbed insolation for a very large area. As the potential evaporation in a desert is equal to the absorbed insolation this can be a very wide spread. With such possible variations it is obvious that the results of evaporation experiments are not reproducible unless carried out on surfaces with the same dimensions. The limits set by the potential evaporation and one half the absorbed insolation are also very useful in evaluating the results of evaporation experiments on moist surfaces of intermediate size.

(f) Further research

Definitive proof for Equation 21 would require close agreement between potential evaporation and absorbed insolation in completely arid regions, and close agreement between potential evaporation and one half of the absorbed insolation in completely humid regions. The potential evaporation could be computed from weather observations using Equation 16 and the appropriate empirical approximations but the evidence would be more convincing in the form of potential evaporimeter observations, or in the form of ordinary evaporimeter observations adjusted using Equation 17 and the appropriate empirical approximations. Good agreement between one half of the absorbed insolation and the difference between rainfall and runoff in completely humid regions would also be very convincing but, because of the nature of such regions, adequate data may be difficult to obtain.

The concept, as developed thus far, deals only with the boundary conditions of an evaporation problem, i.e., with regional evaporation from an area so large that convection of heat and vapour from the surface thoroughly modifies the overpassing air, and the potential evaporation from an area so small that convection of heat and vapour from the surface has no significant effect on the overpassing air. Both "large" and "small" are loosely defined and many evaporation problems involve areas which are intermediate in size. Therefore, research on modification of the lower atmosphere at climatic boundaries is essential if the concept is to be exploited to its full potential. The work of Rider, Philip and Bradley (1963) and that of Dyer and Crawford (1965) provide good examples of the use of sophisticated instrumentation in such research. The humble evaporimeter also may prove to have some use in this field as a number of them, strategically located near a climatic boundary, would provide a continuous temporal and areal record of the modification of the lower atmosphere.

There are many other problems which require solutions in order to extend the scope of the concept. Some of the needs are:-

(1) For improvements to the design of evaporimeters so that they may be used to provide undistorted and accurate observations of potential evaporation.

(2) For research on sub-surface heat storage changes so that the concept may be applied with confidence to short time intervals and to deep lakes.

(3) For studies on the relationship between snow evaporation and snow melt so that the concept may be applied with confidence to snow covered surfaces and glaciers.

Solutions to such problems will not be simple, but should be facilitated if experiments are designed and data are analyzed in the light of the implications of Equation 21. Therefore it is apparent that, in addition to providing a probable solution to the areal boundary conditions of the evaporation problem, the concept of Equation 21 has opened up an interesting, complex and potentially profitable field of research.

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