OPERATIONAL ESTIMATES OF AREAL EVAPOTRANSPIRATION AND THEIR SIGNIFICANCE TO THE SCIENCE AND PRACTICE OF HYDROLOGY

I THE PROBLEM

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ABSTRACT

Morton, F.I., 1982a. Operational estimates of areal evapotranspiration and their significance to the science and practice of hydrology: I The problem

Reliable estimates of areal evapotranspiration are essential to significant improvements in the science and practice of hydrology. Direct measurements; such as those provided by lysimeters, eddy flux instrumentation or Bowen-ratio instrumentation; give point values, require constant attendance by skilled personnel and are based on unverified assumptions. A critical review of the methods used for estimating areal evapotranspiration is presented. It indicates that the conventional conceptual techniques, such as those used in current watershed models, are based on assumptions that are completely divorced from reality. Furthermore, it provides evidence to the effect that causal techniques based on processes and interactions in the soil-plant-atmosphere continuum are not likely to prove useful for at least another generation. However, the review demonstrates that models based on the complementary relationship between areal and potential evapotranspiration could do much to fill the gap until such time as the causal techniques become practicable. The complementary relationship is introduced briefly but its conceptual and empirical foundations, its use in providing the basis for operational estimates of areal evapotranspiration and lake evaporation, the testing of such estimates against comparable water budget values and their potential use in providing the impetus for transforming hydrology from a descriptive to a predictive science are presented in a series of companion papers.

INTRODUCTION

Water planning and management require the ability to predict the hydrologic regime of river basins over the lifetime of proposed projects and the ability to forecast the hydrologic responses of river basins to specific precipitation events. The prediction of hydrologic regimes is based on the assumption that the streamflow records of the past are representative of those of the future, and thus ignores the effects of the land use or climatic changes that have taken place during the period of record and that will take place during the lifetime of the project. Forecasts of hydrologic response tend to be based on some crude modelling device, such as the antecedent precipitation index, with the implicit assumption that the water stored in the soil, swamps and lakes of a basin has little effect on subsequent runoff events. Although water planning and management have changed significantly during the last two decades, primarily through the use of the computer, there has been little progress in evaluating or improving the basic assumptions. Thus the current state of the art is that of a car spinning its wheels while stuck in a swamp, with frenetic computer applications making the problem worse by devaluing human judgement and experience.

Evaluation and improvement of the water planning and management assumptions have been impeded by the apparent determination of hydrologists and engineers to ignore the significance of areal evapotranspiration in the modelling of the hydrologic cycle. This determination persists despite widespread recognition that evapotranspiration is a much larger proportion of the precipitation than runoff in most populated parts of the world; that the accumulated effects of evapotranspiration on the water stored in soil, swamps and lakes control the basin response to precipitation events; and that evapotranspiration is the component of the hydrologic cycle most directly influenced by land use and climate changes. This attitude has resulted in the diversion of research effort from the real problems of hydrology to unreal mathematical elaborations of the more tractable peripheral problems.

A somewhat similar problem exists with regard to meteorology. Although air mass properties are the integrated results of hemispheric or even global phenomena, the way they are modified in passing over the continents is influenced to a large extent by the proportion of net radiant energy that is used for areal evapotranspiration and transformed into latent heat. The proportion can vary from zero to almost one. This significant transformation of energy tends to be ignored because of the difficulties associated with estimating areal evapotranspiration and with coupling the microscale phenomena near the ground to the macroscale phenomena in the free atmosphere.

The main reason for ignoring the hydrological and meteorological effects of areal evapotranspiration is that it is extremely difficult to measure or estimate. There is as yet no way to make routine direct measurements. Those methods in current use, e.g. the covariance or eddy flux technique, the Bowen-ratio energy balance technique and the lysimeter technique, give only point values and are not amenable to routine observations because of the high cost of instrumentation and the need for constant attendance by scientific or technical personnel. Furthermore they are all based on unverified assumptions. Thus the evapotranspiration from a weighing lysimeter is of significance only if it is assumed that the soil, soil moisture and vegetation are identical to those in the immediate environment. The water balance technique is an exception in that it can be applied to an area that drains into a stream where the flow is measured. However this requires assumptions concerning changes in the amount of water stored in the snow packs, lakes, swamps, soil and rock of the drainage basin. In practice it is usually necessary to ignore the changes in storage so that the resultant estimates of areal evapotranspiration are valid only when averaged over a period of years long enough for the changes to become a negligible part of the total water balance. Therefore the only practicable method for measuring areal evapotranspiration has little practical use except in calibrating and testing models.

The development of models for estimating areal evapotranspiration is hindered by the number of processes and feedback mechanisms involved as well as by the lack of good data. Evapotranspiration depends on the availability of energy, a function of the net radiant energy supply and the temperature and humidity of the overpassing air, and on the availability of water, a function of the precipitation and a complex system of soil moisture and plant processes. Furthermore, the soil-plant-atmosphere continuum has a significant number of interactions and feedback mechanisms. Those that are known tend to be ignored and there is reason to suspect that others remain unidentified. One interaction that is particularly relevant to the distinction between point and areal evapotranspiration concerns changes in the availability of water from place to place in a heterogeneous area and the effects of such changes on the overpassing air. Thus the decrease in evapotranspiration from a dry patch will make the overpassing air hotter and drier, thereby increasing the availability of energy and producing compensatory increases in evapotranspiration from any downwind wet patches.

Model development is also impeded by lack of communication between scientific disciplines. For example, the assumptions used for the evapotranspiration components of current watershed models are inconsistent with findings published in the meteorological, botanical and soil science literature. A critical review of the techniques for estimating evapotranspiration is presented herein. It indicates that the conventional conceptual techniques, such as those used in the current watershed models, are based on assumptions that are completely divorced from reality. Furthermore it provides some evidence to the effect that causal techniques based on the processes and interactions in the soil-plant-atmosphere continuum are not likely to prove useful for at least another generation. However the review demonstrates that techniques based on a complementary relationship between areal and potential evapotranspiration could do much to fill the gap until such time as the causal techniques become practicable.

The concept of a complementary relationship between areal and potential evapotranspiration is based on the previously mentioned interaction between evaporating surfaces and overpassing air. By incorporating this crucial feedback mechanism it avoids the complexities of the soil-plant system so that areal evapotranspiration can be estimated from its effects on the routinely observed temperatures and humidities used in computing potential evapotranspiration with no need for locally optimized coefficients. This means that the results are falsifiable so that errors in the associated assumptions and empirical relationships can be detected and corrected by progressive testing against long-term water balance estimates of river basin evapotranspiration from an ever-widening range of environments. During the past decade the test range has been expanded from Canada and Ireland to the United States of America, to Africa and to Australia and New Zealand.

The complementary relationship is introduced briefly herein but its conceptual and empirical foundations, its use in providing the basis for operational estimates of areal evapotranspiration and lake evaporation, the testing of such estimates against comparable water budget values and the potential use of such estimates to provide the impetus for transforming hydrology from a descriptive to a predictive science are presented in a series of companion papers (Morton, 1982b, 1982c, 1982d, and 1982e).

A tabular summary of the literature on the complementary relationship since it was introduced by Bouchet (1963) is presented in the Appendix.

EVALUATION OF CONVENTIONAL CONCEPTUAL TECHNIQUES

The currently accepted conceptual techniques for estimating evapotranspiration have many applications, all of them slightly different. Examples are in Budyko (1974), in the new versatile soil moisture budget presented by Baier and Robertson (1966) and in a technique used by the United Kingdom Meteorological Office to map soil moisture deficits and actual evapotranspiration (Grindley, 1970). Despite superficial differences the basic assumptions are the same and can be evaluated by examining one example, in this case the technique used in one of the better known watershed models.

The SACRAMENTO WATERSHED MODELLING SYSTEM (Burnash, Ferrel and McGuire, 1973) employs five conceptual storage reservoirs in the soil-plant continuum, two of which store tension water and three of which store free water. The tension storages are emptied by evapotranspiration and filled by interception or percolation whereas the free-water storages are emptied by interflow, baseflow or subsurface outflow and are filled by infiltration or percolation. There are also transfers of water from free-water storage to tension storage. All of these processes depend on the proportional loading (i.e. the ratio of current storage to storage capacity) for one or more of the reservoirs. The storage capacities and the parameters that control the transfers of water to and from the reservoirs are calibrated to provide the best possible fit between model output and recorded streamflows. The evapotranspiration is assumed equal to the evaporation demand (normally represented by pan evaporation or potential evapotranspiration) multiplied by the difference between the sum and product of the proportional loadings for the total tension storage and the upper zone tension storage. Because it abstracts a large proportion of the total precipitation directly from the tension storages and indirectly from the free-water storages, the need to adjust the evapotranspiration to make the water-balance balance over the calibration period has a dominant influence on the selection of storage capacities and transfer parameters. Therefore the reality of the calibrated capacities and transfer parameters, i.e. their ability to simulate flows before and after the calibration period and their applicability to the development of relationships for transposition to other basins, depends to a large extent on the validity of the assumptions used in estimating evapotranspiration.

The technique used for estimating evapotranspiration in the SACRAMENTO WATERSHED MODELLING SYSTEM is based on a number of implicit assumptions about soil, plant and atmospheric processes. The discussion that follows furnishes published experimental evidence that these are either wrong or questionable.

Soil Assumptions: The tension storage capacities are supposed to reflect the amount of moisture that can be held in the root zone between an upper threshold value of moisture content (e.g. the field capacity) and a lower threshold value of moisture content (e.g. the wilting point). The specific assumptions that are made in modelling the tension storages are:

(1) That the tension storages are constant and do not depend on seasonal or annual changes in vegetation cover and root development. The problems inherent in this assumption are self-evident and are particularly important in cultivated areas.

(2) That there is no drainage from tension storage when the water content is below the upper threshold value, the field capacity. This assumption is contrary to the theoretical and experimental findings of Rubin and Steinhardt (1963), Rubin, Steinhardt and Reiniger (1964) and Rubin (1966) that soil water content during rain infiltration rises only to the point where the hydraulic conductivity becomes equal to the rainfall or snowmelt rate and then water is transferred downward. Thus it is possible to have

significant groundwater recharge from long periods of rainfall or snowmelt with the water content well below field capacity. Moreover, evidence provided by Robins, Pruitt and Gardner (1954), Ogata and Richards (1957), Neilsen, Kirkham and Van Wijk (1959), Nixon and Lawless (1960) and Hewlett (1961) indicates that slow vertical percolation can provide an important supply of groundwater recharge during long dry periods after soil-moisture contents have declined below field capacity.

(3) That moisture fluxes into or out of the tension storages are independent of temperature, temperature gradients and vapour pressure gradients. Carey (1966) has shown that soil-moisture transport induced by thermal gradients and their associated vapour pressure gradients can be significant when compared with transport induced by hydraulic gradients, particularly when the temperatures or soil-moisture contents are low. Peck (1974) has provided a brief review of the theoretical and empirical evidence for the concept that the large observed increases in soil-moisture at below-freezing temperature or under a snow cover are caused by upward migration of water. Abramova (1968) has presented evidence of extensive upward and downward transports of water as vapour in the relatively dry soils of the Precaspian Lowland in the U.S.S.R. The experiments demonstrating how Chinese elm could flourish with the soil in the root zone at wilting point were particularly interesting because they indicated that the required water was supplied by an upward

transport of water vapour and that this was induced by a relative humidity gradient in the soil air rather than by a temperature gradient.

Plant Assumptions: The conventional conceptual techniques for estimating evapotranspiration, including that used in the SACRAMENTO WATERSHED MODELLING SYSTEM, are based on the assumption that the vegetation acts as a passive wick that transfers water from the tension storages to the atmosphere. This requires that leaf stomatal resistances be responsive only to internal leaf water controls and not to external environmental controls such as those imposed by irradiance, carbon dioxide in the ambient air, leaf temperature and humidity in the ambient air. Documentation for the existence of such controls and hence for the lack of reality in the assumption is presented in a detailed review by Cowan (1977) and in a shorter review by Hall, Schulze and Lange (1976).

The notion that the ambient humidity or the difference between the vapour pressure in the leaf and the vapour pressure in the air can have a significant direct effect on stomatal resistance may seem novel and strange to hydrologists and meteorologists, although the indirect effects, whereby an increase in vapour pressure difference would increase transpiration which in turn would reduce internal leaf water content thereby causing the stomata to contract, have long been recognized. Thus the indirect effects represent a feedback mechanism which reduces transpiration after leaf water contents have been reduced and the direct effects represent a feedforward mechanism which reduces transpiration before leaf water contents have been reduced. Schultze et al. (1972) have shown that the direct effects of an increase in water vapour concentration difference (or vapour pressure deficit) on apricots growing in the Negev Desert is to reduce transpiration and increase leaf water content. Kaufmann (1979) provided evidence that the direct effects of an increase in the absolute humidity difference (or vapour pressure deficit) on four year old potted Engelmann spruce seedlings was to increase the stomatal resistance sufficiently to make the transpiration independent of the difference. The significance of such findings is that an increase in vapour pressure difference which produces an increase in potential evapotranspiration may cause an increase in stomatal resistance large enough to decrease the actual transpiration.

Much of the evidence for the direct effects of vapour pressure differences on stomatal resistance is based on instrumentation and techniques that are unfamiliar to hydrologists and meteorologists. However, Tan and Black (1976) have used Bowen-ratio energy balance measurements of evapotranspiration to estimate the daytime mean canopy resistance of a 10-m high Douglas-fir forest. Their results are shown in Fig. 1 and indicate that the canopy resistance increases more in response to increases in the daytime mean vapour pressure deficit (range approximately 8 to 20 mbars) than to decreases in daily soil water potential (range 0 to -10 bars) or in daily soil water content (range 23 to 8 percent by volume).

The effects of ambient carbon dioxide concentrations on stomatal resistance have significance for current concerns about the hydrologic and climatic consequences of the combustion of fossil fuels. Based on their experiments with potted carnation plants, Enoch and Hurd (1979) have estimated that increases in carbon dioxide concentrations have decreased transpiration by 1.6 percent this century and could decrease transpiration by about 10 percent during the next 50 years. Such decreases could cause significant increases in runoff and a significant warming of the atmosphere which would augment the better known greenhouse effect.

Atmospheric Assumpions: The currently accepted conceptual techniques for estimating evapotranspiration, including that used in the SACRAMENTO WATERSHED MODELLING SYSTEM, are based on the assumption that pan evaporation or potential evapotranspiration are independent of the actual evapotranspiration and may therefore be used to reflect the evaporation demand. This assumption ignores the well known fact that a decrease in areal evapotranspiration caused by a reduction in the availability of water will increase the temperature and decrease the humidity of the overpassing air and that these in turn will increase pan evaporation or potential evapotranspiration.

Potential evapotranspiration is normally estimated from a solution of the energy balance equation and the vapour transfer equation. These equations can be solved with a graphical technique, an iterative technique or an approximate analytical technique. The approximate solution as developed by Penman (1948) is

$$E_{TP} = R_{TP}\ \Delta/(\Delta+\gamma p) \ + f_T(v-v_D)\ \ \gamma p/(\Delta+\gamma p) \qquad (1)$$

in which E_{TP} = potential evapotranspiration, R_{TP} = net radiation at the temperature of the evaporating surface, v = saturation vapour pressure at air temperature, v_D = saturation vapour pressure at dew point temperature, f_T = vapour transfer coefficient, γ = psychrometric constant, p = atmospheric pressure and Δ = rate of change of saturation vapour pressure with respect to temperature.

Kohler, Nordenson and Fox (1955) have shown that eq. (1) can provide good estimates of evaporation from the Class A evaporation pan if modified to take into account both the difference between vegetated and water surfaces and the sensible heat flux through the wall and bottom of the pan. Therefore there are grounds for belief that the evaporation from any small water-filled container will provide a reasonable reflection of potential evapotranspiration, although the reflection will be somewhat distorted by differences in albedo, surface roughness and in the ratio of the latent heat transfer area to sensible heat transfer area.

Eq. (1) shows that potential evapotranspiration increases when the vapour pressure deficit $(v-v_D)$ increases. This results from an increase in air temperature (as reflected in v) and/or a decrease in humidity (as reflected in v_D). Following the cause-and-effect chain beyond the customary first link, this is the kind of response that can be expected from the increase in heat flux and the decrease in vapour flux associated with a reduction in the availability of water for evapotranspiration from the surrounding area. The resultant feedback mechanism invalidates the currently accepted hydrometerological assumption that in any relationship between areal evapotranspiration and potential evapotranspiration the latter is the independent variable. Thus the use of potential evapotranspiration as a causal factor or forcing function is limited to two specific cases: (1) to existing moist areas so large that the effects of the evapotranspiration on the temperature and humidity of the overpassing air are fully developed and the areal evapotranspiration equals the potential evapotranspiration; and (2) to existing or hypothetical moist areas so small that the effects of the evapotranspiration on the temperature and humidity of the overpassing air are negligible. Pans and other evaporimeters fit into the latter category insofar as size is concerned.

In Bernouilli's equation for open-channel flow, the potential energy responds negatively to changes in the kinetic energy. Thus there appear to be analogies between potential evapotranspiration and potential energy and between areal evapotranspiration and kinetic energy. To be completely analogous the relationship would have to be complementary, i.e. the responses would have to be both opposite in sign and equal, and both observational data and theory show that this is a distinct possibility. However it should be noted that the analogy can not be extended into the critical flow regime because areal evapotranspiration cannot exceed potential evapotranspiration.

The oasis effect is the best known phenomenon associated with the foregoing interactions. It can be demonstrated by comparing monthly evaporation from two U.S. Weather Bureau Class A Pans in the southeast desert basin of California during the years 1962, 1963 and 1964. Figures 2 and 3 show the pan evaporation in a large well-irrigated area at Indio U.S. Date Garden plotted against the pan evaporation in the desert at Death Valley for the first six months and the last six months of the calendar year respectively. The plotted regression lines have coefficients of correlation of 99 percent. The small zero-intercepts can be attributed to the effect of the 2.7 degree difference in latitude in increasing the winter radiation inputs at Indio so that the slopes represent the effects on pan evaporation of increasing the areal evapotranspiration from a near-zero value in the desert to a near-maximum value in an irrigated oasis. Thus Fig. 2 shows that the oasis pan evaporation is only 57 percent of the desert pan evaporation during the first six months of the calendar year and Fig. 3 shows that the oasis pan evaporation is only 52 percent of the desert pan evaporation during the last six months of the calendar year.

The oasis effect and the negative relationship between actual and potential evapotranspiration have been demonstrated by Davenport and

Hudson (1967). They measured the variation in evaporation across a series of irrigated cotton and unirrigated fallow fields in the Sudan Gezira, using fiberglass dishes with black painted wells 113 mm in diameter and 36 mm in depth. The dish evaporation observations provided a somewhat distorted reflection of the potential evapotranspiration. The passage of air from the upwind desert (and/or the upwind unirrigated fallow fields) over the irrigated cotton caused the dish evaporation to decrease rapidly and approach a low constant value within 300 m - the width of the fields. Furthermore, as the air passed from irrigated cotton across unirrigated fallow the dish evaporation increased rapidly and approached a high constant value within 300 m. Fig. 4 shows the variation of dish evaporation across three irrigated fields on December 27, 1963. The description of the soil condition is presumably based on visual inspection and is related to the number of days since the last irrigation. Thus it is reasonable to assume that the actual evapotranspiration is greater in the "moist" field than in the "dry" field and that it is somewhere near its maximum possible value in the "wet" field.

At the upwind edge of the irrigated fields, where the dish evaporation decreases rapidly, the hot dry air from the desert (or the unirrigated fallow) loses heat and gains vapour from contact with the transpiring cotton leaves. Downwind from the transitional zone, where the dish evaporation approaches a low constant value, the effects of the evapotranspiration on the temperature and humidity of the overpassing air are well developed and approaching equilibrium. (Decreases in temperature and increases in humidity as the air moved across the irrigated cotton were observed over one field. The vapour pressure appeared to attain equilibrium values within 300 m but the temperatures were still decreasing, possibly because the observations were made above the level of the crop and dishes.) As it is only in an equilibrium zone that the term areal evapotranspiration has any meaning, the minimum size of an area for which the term is applicable is one in which the edge effects in the upwind transitional zone become insignificant.

The ratio of daily dish evaporation at the downwind edges of the individual irrigated cotton fields to that at the upwind edge of the irrigated area was 0.69 for the field with "dry" soil, 0.60 for the field with "moist" soil and 0.53 for the field with "wet" soil. This provides good evidence that the dish evaporation, and presumably the potential evapotranspiration, respond negatively to changes in areal evapotranspiration induced by changes in the availability of water.

Army and Ostle (1957) have inadvertently provided more direct evidence for the existence of a negative relationship between areal and potential evapotranspiration. They calculated actual areal evapotranspiration (E_T) from 0.1 acre plots of spring wheat by adding the total precipitation from seeding to harvest to the amount of soil moisture lost from seeding to harvest for 29 years at Huntley, Montana and for 26 years at Havre, Montana. The results were then related to the corresponding values of evaporation from nearby sunken B.P.I. tanks (E_{PW}), which provide a somewhat distorted reflection of potential evapotranspiration. The resultant regression equations for plots that were cropped every year were:

$$E_T \text{ (at Huntley)} = 30.8 - 1.025\, E_{PW} \qquad (2)$$

$$E_T \text{ (at Havre)} = 24.5 - 0.644\, E_{PW} \qquad (3)$$

and the regression equations for two plots that were cropped and left in fallow on alternate years were:

$$E_T \text{ (at Huntley)} = 34.6 - 1.142\, E_{PW} \qquad (4)$$

$$E_T \text{ (at Havre)} = 28.5 - 0.725\, E_{PW} \qquad (5)$$

The zero-intercepts for the four equations are in inches of water and the coefficients of correlation vary from 0.790 to 0.826 with a mean value of 0.814. In evaluating the results it should be noted that the tank evaporation would be higher than the potential evapotranspiration because of the low albedo of water and that this would cause the negative slopes to be underestimated; that unreported differences in vegetation and soil immediately adjacent to the various plots and tanks could account for some of the differences between the equations; and that unreported year-to-year variations of absorbed incident radiation would add to the scatter of the points and the uncertainty concerning the regression coefficients. With these considerations in mind the equations show that the relationship between actual evapotranspiration and potential evaporation is negative and that there is reasonable probability that it is complementary, i.e., that the slope is -1.

Summation: The foregoing evaluation shows that the assumptions used to estimate areal evapotranspiration in the conventional conceptual models are completely divorced from reality. They continue to be used because of their tractability and because of the ease with which unreal coefficients can be optimized when the relevant variables are correlated, sometimes spuriously, through their pronouced seasonal patterns. This is evident in a discussion of Penman (1967) in which it is stated that "in terms of hydrology, one could be very wrong in the estimates of potential evaporation and still get the right answer if one had the right value of D" in which D is a poorly defined tension storage capacity which is usually obtained by optimization. The reality of "garbage in - garbage out" tends to be ignored because the scientific and engineering literatures provide few if any examples of the consequences of practical applications or of rigorous tests similar to those carried out by Garrick, Cunnane and Nash (1978).

The most disturbing implication of the foregoing evaluation is that it is unrealistic to keep a simple real-time water budget for the unsaturated root zone. Part of the problem may be that the field capacity has some basis in reality but is a function of temperature (as suggested by Peck, 1974) and is applicable only to the problem of drainage after infiltration has ceased. Thus the concept of a minimum drainage water content that changes with the seasons and a maximum infiltration water content that changes with the rainfall or snowmelt rate may provide a realistic basis for a simple isothermal water budget. The alternative appears to be a complex, physically-based and largely untested mathematical model of the type developed by Freeze (1971). However no matter what kind of model is selected it must be made more realistic by taking into account the effects of temperature and/or vapour pressure gradients on vertical moisture fluxes.

The other information used in the evaluation is more welcome because it can be used constructively to formulate more realistic models. Thus the information on plant processes that control transpiration provides the basis for what are referred to herein as causal models and the information on the negative or complementary response of potential evapotranspiration to changes in the availability of water for areal evapotranspiration provides the basis for a technique that avoids the complexities of the soil-plant system. The causal hydrobotanical models are evaluated in the next section and the complementary relationship is introduced in the concluding discussion.

EVALUATION OF CAUSAL HYDROBOTANICAL TECHNIQUES

The hydrobotanical models are based on an equation formulated by Monteith (1965). If the potential evapotranspiration (E_{TP}) is derived from eq. (1), the actual evapotranspiration from a small area (E_{TS}) is estimated from

$$E_{TS} = E_{TP}\ (\Delta+\gamma p)/[\Delta+\gamma p(1+r_c/r_a)] \qquad (6)$$

in which r_c is the canopy resistance, i.e. an areal average of stomatal resistances, r_a is the aerodynamic resistance and all other symbols are as defined previously. The aerodynamic resistance is related to the vapour transfer coefficient, f_T , in eq. (1) by

$$r_a = \rho c/(\gamma p f_T) \qquad (7)$$

in which ρ and c are the density and specific heat respectively of the air.

One of the simpler applications of a causal model is that described by Calder (1977). The required coefficients were obtained by an optimization process using data from a large natural lysimeter with twenty-six mature spruce trees in Plynlimon, Central Wales. The aerodynamic resistance was obtained from data observed during rain storms (when the canopy resistance would be zero) together with four other parameters in an interception model. The optimization process indicated that it was a constant, i.e. that it was independent of wind speed. With the aerodynamic resistance established, the four parameters needed to relate canopy resistance to a cosine function of the Julian day number (reflecting the seasonal pattern of irradiance) and the vapour pressure deficit were derived from data observed when the canopy was dry. The optimization process indicated that there was no relationship between canopy resistance and soil moisture even though soil moisture deficits in excess of 100 mm were recorded. Unfortunately there is no way to tell whether this lack of relationship is real or whether it is due to a causal relationship between soil moisture and vapour pressure deficit or to a relationship between soil moisture and the time of the year.

A more sophisticated model has been developed by Federer (1979). The aerodynamic resistance is estimated from wind speed and from estimated values of a roughness parameter and a zero-plane displacement; the canopy resistance is assumed proportional to stomatal resistance; the stomatal resistance is derived from a regression equation in which porometer measurements of stomatal resistance have been related to solar radiation, air temperature, vapour pressure deficit and the plant water potential; the plant water potential is related to soil moisture by an intricate and assumption-filled linkage through the plant system; and the soil moisture is related to precipitation by dividing the soil into a number of layers and applying some approximate equations. The model includes an iterative procedure.

Federer (1979) chose to simulate the behaviour of mature yellow birch and sugar maple in the Hubbard Brook Experimental Forest in New Hampshire. Although some of the predetermined parameters were selected objectively, there was so much judgement used in the selection of others that they could only be described as educated guesses. Three parameters; i.e. the ratio of canopy to stomatal resistance, the resistance to flow into plant storage and the internal plant resistance per unit of length, were determined by fitting the model to two days of data. Half-hourly atmospheric input data were derived from daily values using a number of assumptions. The model simulated field data from a hardwood forest in terms of both plant water potential and stomatal behaviour, within a day and over a summer. The results were used to develop simple relationships between the ratio of actual to unstressed transpiration and some function of the soil moisture where the unstressed transpiration is defined as the value computed from the model when the stomatal opening is not affected by plant water potential (or soil moisture).

Causal models of the type formulated by Federer (1979) represent the wave of the future since it is obviously impossible to predict the effects of natural or man-made changes without a detailed knowledge of all the causal factors and their interactions. Because of the complexity of the processes and interactions it would be more correct to state that they represent the wave of the far-distant future. Some of the problems are:

(1) Porometer estimates of stomatal resistance are based on the length of time required to increase the relative humidity in a small chamber from a selected artifically low value to a selected higher value (Kanemasu, Thurtell and Tanner, 1969). This either ignores the effects of humidity on stomatal resistance or assumes that the effects are negligible over short measurement periods. Hall, Schultze and Lange (1976) have pointed out that this assumption may not always be valid, in which case the measurements would provide a good example of the classical phenomenon whereby the act of measurement disturbs the process being measured.

(2) Correlations between stomatal resistance and environmental factors can lead to unrealistic results when the so-called independent variables, i.e. solar radiation, temperature, vapour pressure deficit and possibly plant water potential, have regular diurnal and seasonal patterns.

(3) Much more knowledge is needed on the plant processes that limit the amount of water reaching the leaves, particularly those at the soil-root interface.

(4) Such models must be based on the assumption that it is possible to compute a real-time soil moisture budget. This problem has been discussed at length in the preceding section. Although the method used by Federer (1979) is much more realistic than those used in the

current conceptual models, there is some indication that it represents an unhappy medium in that it is too complex for easy computation and too simple for realism.

(5) The application of eq. (6) to an area is based on the assumption that the potential evapotranspiration derived from eq. (1) is independent of the actual areal evapotranspiration and may therefore be used to reflect the evapotranspiration demand. The information presented in the preceding section shows that this assumption is invalid. Federer (1979) has tried to take this into account in using the results of his model by using the unstressed transpiration as a causal factor or forcing function but does not take it into account in the model itself.

CONCLUDING DISCUSSION

The foregoing critical review has indicated that the conventional conceptual techniques used in estimating areal evapotranspiration are based on assumptions that are completely divorced from reality and that the causal techniques currently under development are not likely to prove useful for at least another generation. However, this does not mean that the science of hydrology must continue to stagnate for another generation awaiting the operational estimates of areal evapotranspiration that are needed to analyze river basin water balances and the hydrological cycle because such estimates can be based on the the negative or complementary responses of potential evapotranspiration to changes in the water available for areal evapotranspiration discussed in a preceding section. If the relationship is complementary it can be represented by:

$$E_T + E_{TP} = 2\,E_{TW} \tag{8}$$

or by:

$$E_T = 2E_{TW} - E_{TP} \tag{9}$$

in which E_T is the areal evapotranspiration, the actual evapotranspiration from an area so large that the effects of upwind boundary transitions, such as those shown in Fig. 4 are negligible; E_{TP} is the potential evapotranspiration, as estimated from a solution of the vapour transfer and energy balance equations, representing the evapotranspiration that would occur from a hypothetical moist surface with radiation absorption and vapour transfer characteristics similar to those of the area and so small that the effects of the evapotranspiration on the overpassing air would be negligible; and E_{TW} is the wet environment areal evapotranspiration, the evapotranspiration that would occur if the soil-plant surfaces of the area were saturated and there were no limitations on the availability of water.

Fig. 5 provides a schematic representation of eq. (8) under conditions of constant radiant energy supply. The ordinate represents evapotranspiration and the abscissa represents the water supply to the soil-plant surfaces of the area, a quantity that is usually unknown. When there is no water available for areal evapotranspiration (extreme left of Fig. 5) it follows that $E_T = 0$, that the air is very hot and dry and that E_{TP} is at its maximum rate of $2E_{TW}$ (the dry environment potential evapotranspiration). As the water supply to the soil-plant surfaces of the area increases (moving to the right in Fig. 5) the resultant equivalent increase in E_T causes the overpassing air to become cooler and more humid which in turn results in an equivalent decrease in E_{TP}. Finally, when the supply of water to the soil-plant surfaces of the area has increased sufficiently, the values of E_T and E_{TP} converge to that of E_{TW}.

The conventional definition for potential evapotranspiration is the same as the definition for the wet environment areal evapotranspiration. However the potential evapotranspiration that is estimated from a solution of the vapour transfer and energy balance equations by analytical (Penman, 1948), graphical (Ferguson, 1952) or iterative (Morton, 1982b) techniques has reactions to changes in the water supply to the soil-plant surfaces that are similar to those shown for E_{TP} in Fig. 5, so what is being estimated can exceed what is being defined by as much as 100 percent. By taking into account such reactions the complementary relationship is analogous to the Bernouilli equation for open channel flow in which the potential energy responds in a complementary way to changes in kinetic energy.

A tabular summary of the literature on the complementary relationship since it was introduced by Bouchet (1963) is presented in the Appendix. Although there is little doubt that the relationship is negative there is no "proof" that it is complementary. It is explicable only in terms of the meteorological theory of advection which is, unfortunately, more qualitative than quantitative. Therefore attempts to provide the complementary relationship with a mathematical basis, such as those presented by Bouchet (1963), Morton (1971), and Seguin (1975), depend on too many unverified and unverifiable assumptions to be classified as anything other than rationalizations. Despite this shortcoming the complementary relationship is superior to most hydrological concepts (including the assumption of linearity that is basic to the unit hydrograph) because it is compatible with all current theoretical knowledge and reliable empirical evidence. This is discussed in detail in a companion paper (Morton, 1982b). The supporting evidence that follows is limited to data presented in preceding sections.

(1) In Fig. 3, the slope of the line that relates the evaporation from the pan in the irrigated area at U.S. Indio Date Garden, California, to the evaporation from the pan in the desert at Death Valley, California, has a slope of 52 percent. In Fig. 4, the ratio of the dish evaporation at the downwind edge of the cotton field with "wet" soil to the dish evaporation in the desert at the upwind edge of the irrigation project is 53 percent. These percentages agree closely with the prediction of the complementary relationship (see Fig. 5)

that the potential evapotranspiration in a completely moist environment should be 50 percent of the potential evapotranspiration in a completely dry environment.

(2) Eq. (9) bears a striking resemblance to eqs. (2), (3), (4) and (5), the regression equations relating water budget estimates of growing season evapotranspiration from spring wheat near Havre and Huntley, Montana, to corresponding values of sunken tank evapotration (Army and Ostle, 1957). The reasons for the variability in the regression coefficients have been adduced in the discussion of the equations which also indicated that the difference between the average slope (-0.884) and that required by the complementary relationship (-1.000) could be due to the low albedo of water which would make the tank evaporation higher than the potential evapotranspiration.

The chief use of the complementary relationship is to provide realistic independent estimates of areal evapotranspiration. Thus eq. (9) provides the basis for a model in which the left hand side, a product of complex processes and interactions in the soil-plant-atmosphere continuum, can be estimated from the routine observations of temperature, humidity and sunshine duration needed to compute the right hand side. As in all modelling exercises the basic concept must be associated with other assumptions and poorly defined empiricisms. However, the complementary relationship avoids the complexities of the soil-plant system and the need to represent such poorly-understood phenomena as infiltration, soil moisture storage and movement, groundwater recharge, uptake of water by roots and stomatal controls by locally optimized fudge factors. Therefore the model is falsifiable (i.e. can be tested rigorously) so that errors in the associated assumptions and relationships can be detected and corrected by progressive testing over an ever-widening range of envionments. Such a methodology uses the entire world as a laboratory and requires that a correction made to obtain agreement between model and river basin water budget estimates in one environment must be applicable without modification in all other environments.

The techniques that use the complementary relationship to estimate areal evapotranspiration are referred to as the complementary relationship areal evapotranspiration (CRAE) models. With a few minor changes they can be used to provide realistic estimates of lake evaporation from routine observations of temperaure, humidity and sunshine duration in the land environment and they are then referred to as the complementary relationship lake evaporation (CRLE) models.

The development of the complementary relationship to serve as a basis for the latest versions of the CRAE and CRLE models and a detailed documentation for these versions are presented in three companion papers (Morton, 1982b, 1982c and 1982e). Documentation of the combined program in FORTRAN and in RPN notation for the Hewlett Packard HP-67 hand-held calculator has been published by Morton, Goard and Piwowar (1980).

The procedure used to test the latest versions of the CRAE and CRLE models is presented in detail in two companion papers (Morton, 1982d and 1982e respectively). Fig. 6 shows the CRAE model estimates of areal evapotranspiration plotted against the corresponding water budget estimates for 143 river basins in North America, Africa, Ireland, Australia and New Zealand. Fig. 7 shows the CRLE model estimates lake evaporation plotted against the corresponding water budget estimates for nine lakes in North America and one lake in Africa. The areal evapotranspiration and lake evaporation were estimated for periods of a month or shorter but were compared with water budget estimates for a year or many years. In the case of evapotranspiration this minimizes the effects of unmeasured changes of water stored in the basins and in the case of lake evapotration this minimizes the effects of unmeasured seasonal changes in subsurface heat storage.

Although it is generally recognized that evapotranspiration is an important aspect of hydrology, the recognition is theoretical. On a practical basis, most hydrologists and engineers fail to understand how research and education in hydrology have been confined to a few narrow problems by the lack of reliable operational estimates of evapotranspiration. Therefore it is extremely difficult for them to visualize the opportunities that arise when such estimates become available. For this reason a large part of a companion paper (Morton, 1982d) is devoted to examples which show the potential impact of operational estimates of evapotranspiration on the science and practice of hydrology.

APPENDIX

LITERATURE ON COMPLEMENTARY RELATIONSHIP

Bouchet (1963)	Introduction and rationalization: Wet environment areal evapotranspiration assumed to be half the absorbed global radiation: Mean annual water budget data from equatorial regions used to demonstrate plausibility of concept.
Morton (1965)	Rationalization and model development: Potential evapotranspiration estimated from graphical technique of Ferguson (1952): Vapour transfer coefficient a function of wind speed derived from pan data: Model tested with monthly water budget estimates for 58 Canadian basins.
Solomon (1966)	Discussion of Morton (1965)
Christiansen (1966)	Discussion of Morton (1965)
Gillespie and King (1966)	Discussion of Morton (1965)
Morton (1967)	Reply to Solomon (1966), Christiansen (1966) and Gillespie and King (1966).
Solomon (1967)	See discussion of Figs. 6 and 7 in companion paper (Morton, 1982b).
Morton (1968)	Potential evapotranspiration estimated from Kohler and Parmele (1967) version of Penman (1948) equation: Concept tested with pan data from humid areas in Ireland and model tested with annual water budget data from six river basins in Ireland.
Morton (1969)	Test of concept extended to pan data from arid areas of the southwestern United States.
Morton (1970)	Model adapted for use during, snowy, high latitude winters. Tested with water budget data from 20 rivers in Ireland and Canada.
Morton (1971)	Model adapted for use without wind speed inputs, i.e. vapour transfer coefficient assumed to be constant.

APPENDIX (Cont'd)

LITERATURE ON COMPLEMENTARY RELATIONSHIP

Bouchet, Fortin, and Seguin (1974)	Application of basic concepts to the effects of the size of an irrigated area on evapotranspiration.
Fortin and Seguin (1975)	Rationalization and test with data from experimental plot for time scales of less than a month.
Seguin (1975)	Application of basic concepts to take into account the effect of the relative size of the potential evapotranspiration area to the areal evapotranspiration area.
Morton (1975)	Potential evapotranspiration estimated from Penman (1948) equation and wet environment areal evapotranspiration derived from Priestley and Taylor (1972) equation: Vapour transfer coefficient and the required advection energy calibrated with data from arid regions. Model tested with 5-year water budget data for 118 river basins in Canada, the southern United States and Ireland.
Morton (1976)	Potential evapotranspiration estimated from Kohler and Parmele (1967) version of Penman equation and wet environment areal evapotranspiration made compatible with it: Test range extended to include 3 more river basins in Kenya and in the United States.
Giusti (1978)	See discussion of Fig. 8 in companion paper (Morton, 1982b).
Morton (1978)	Wet environment areal evapotranspiration corrected for effects of changes in the availability of water on temperature: Vapour transfer coefficient made function of atmospheric stability and radiation estimates made more realistic. Test range extended to one more basin in Canada.
Wallace (1978)	Application of model (Morton 1978) to data from arid area near Hanford, Washington.
Morton (1979)	Modification of model (Morton, 1968) to provide estimates of lake evaporation. Tested with annual water budget data for seven North American lakes.

APPENDIX

LITERATURE ON COMPLEMENTARY RELATIONSHIP

Brutsaert and Stricker (1979)	Wet environment areal evapotranspiration estimated from Priestley and Taylor (1972) equation; potential evapotranspiration estimated from Penman (1948) equation; vapour transfer coefficient estimated from wind speed: Tested with 3-day energy budget data for a basin in the Netherlands during the dry summer of 1976.
LeDrew (1979)	A critical evaluation of the complementary relationship based primarily on simulated values of areal and potential evapotranspiration.
Morton (1980)	Discussion of LeDrew (1979) stressing the lack of reality in the criterion used to select which pairs out of the 540 simulated values of areal evapotranspiration and 100 simulated values of potential evapotranspiration were compatible with the requirement that profiles of temperature and humidity be completely adjusted to surface fluxes of heat and water vapour.
LeDrew (1980)	Reply to Morton (1980) stating that the selection criterion is unimportant thereby implying incorrectly that there is no requirements for the profiles of temperature and humidity to be completely adjusted to the surface fluxes of heat and water vapour.
Morton, Goard and Piwowar (1980)	Documentation of Fortran programs and hand-held programmable calculator programs for the latest versions of the areal evapotranspiration and lake evaporation models (Morton, 1982a,b,c,d,e).

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TITLES FOR FIGURES

Fig. 1 Relationship between mean daytime canopy resistance, soil water potential and vapour pressure deficit for Douglass Fir (Tan and Black, 1976).

Fig. 2. Relationship between Class A pan evaporation at Indio U.S. Date Garden, California, and at Death Valley, California for the months of January to June inclusive.

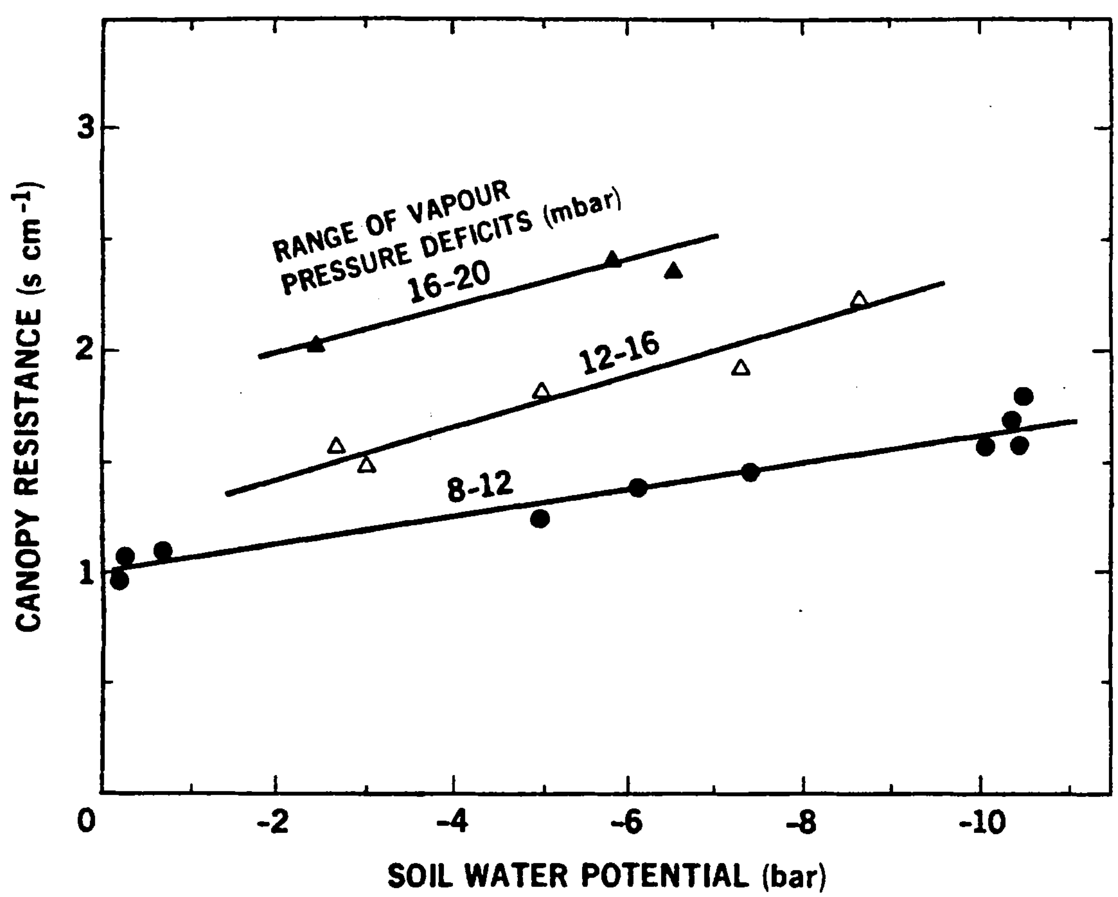
Fig. 3. Relationship between Class A pan evaporation at Indio U.S. Date Garden, California, and at Death Valley, California, for the months of July to December inclusive.

Fig. 4 Comparison of evaporation rates across irrigated cotton fields on December 27, 1963 (Davenport and Hudson, 1967).

Fig. 5 Schematic representation of complementary relationship between areal and potential evapotranspiration with constant radiant energy supply.

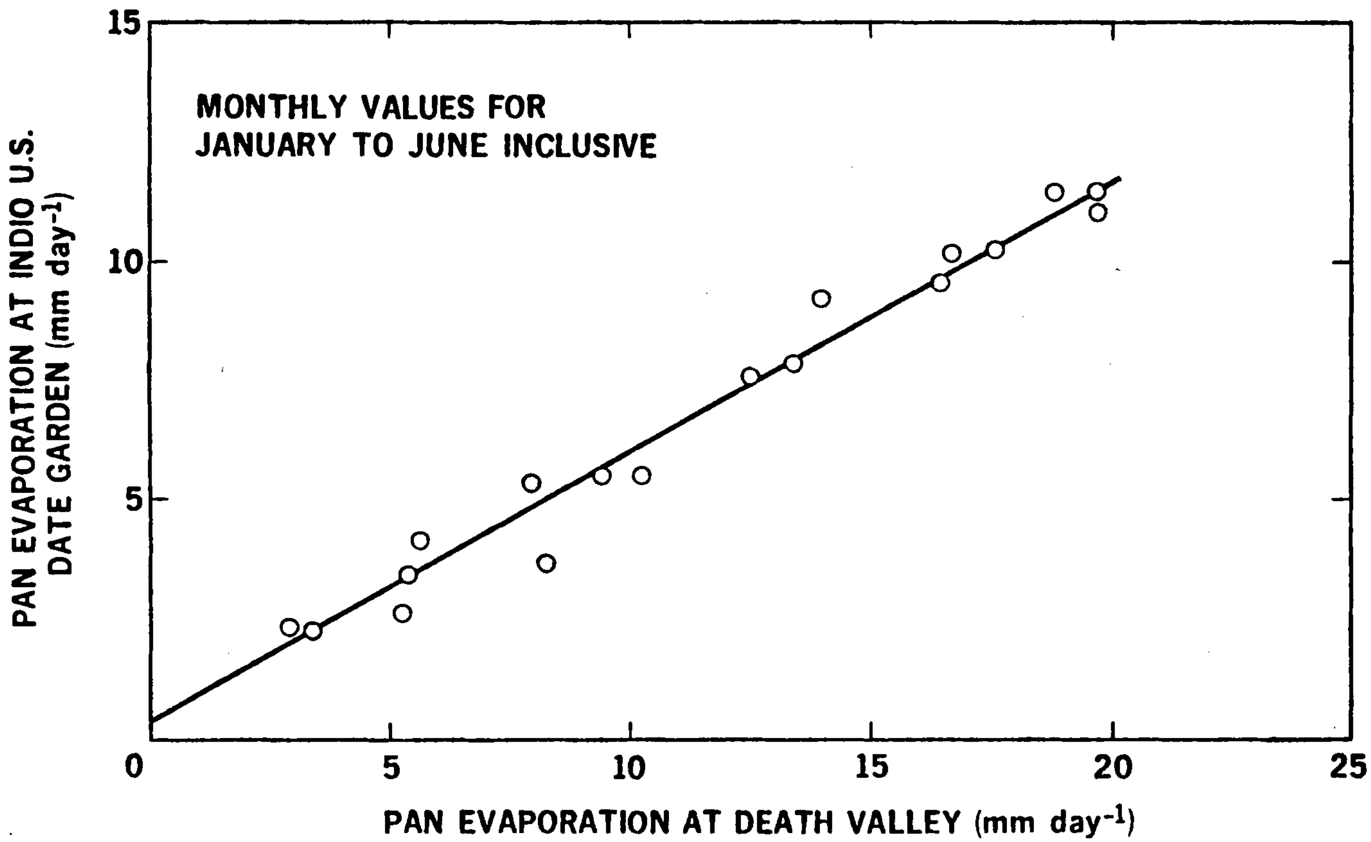
Fig. 6 Comparison of model estimates with water budget estimates of areal evapotranspiration for 143 river basins in North America, Africa, Ireland, Australia and New Zealand.

Fig. 7 Comparison of model estimates with water budget estimates of evaporation fromm Lake Victoria [V], Salton Sea [ss], Silver Lake [S], Lake Hefner [H], Pyramid Lake [P], Winnemucca Lake [W], Lake Ontario [O], Last Mountain Lake [L] and Dauphin Lake [D].

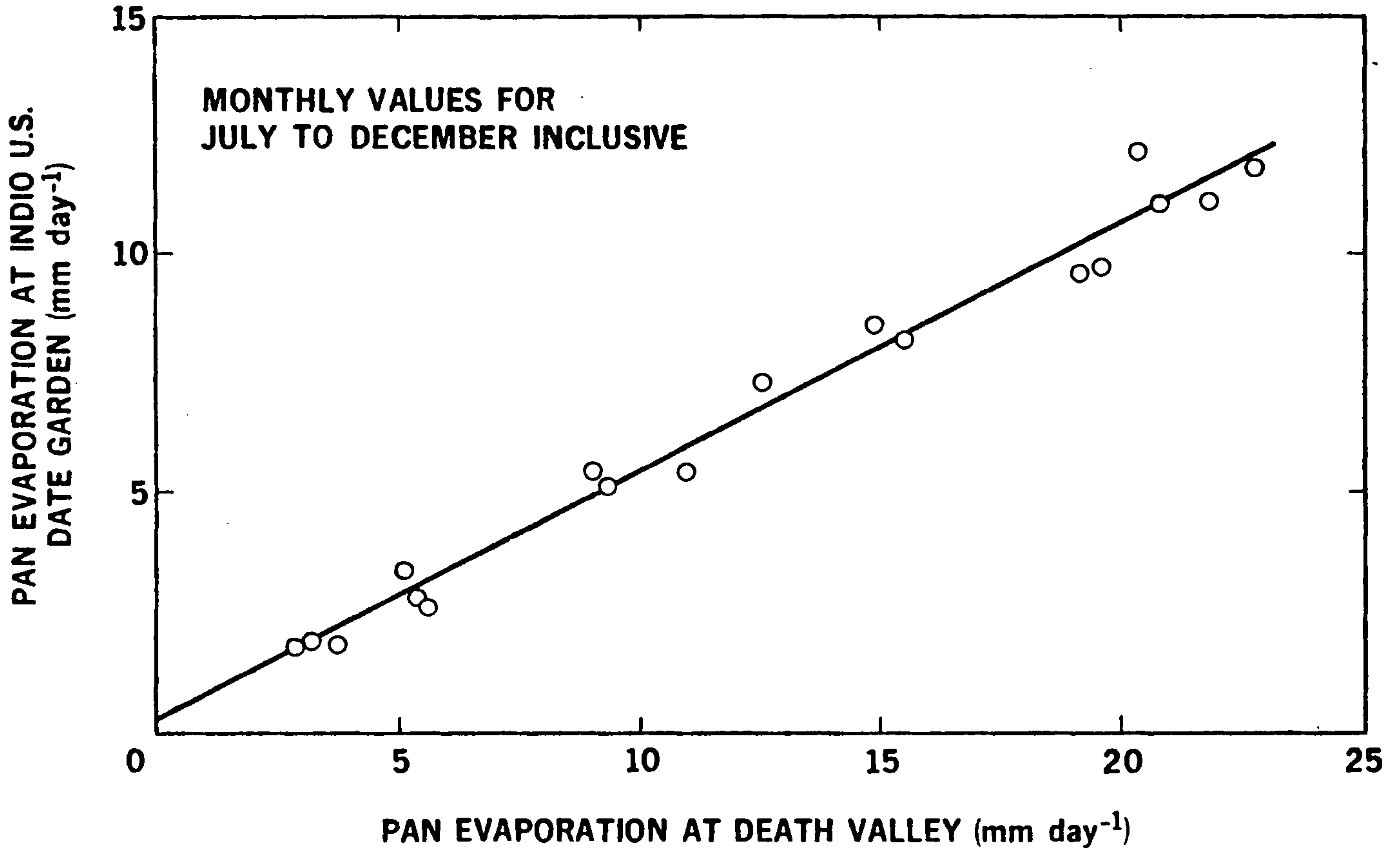

RANGE OF VAPOUR
PRESSURE DEFICITS (mbar)
16-20
12-16
8-12
CANOPY RESISTANCE (s cm^{-1})
SOIL WATER POTENTIAL (bar)
3
2
1
0
-2
-4
-6
-8
-10

1

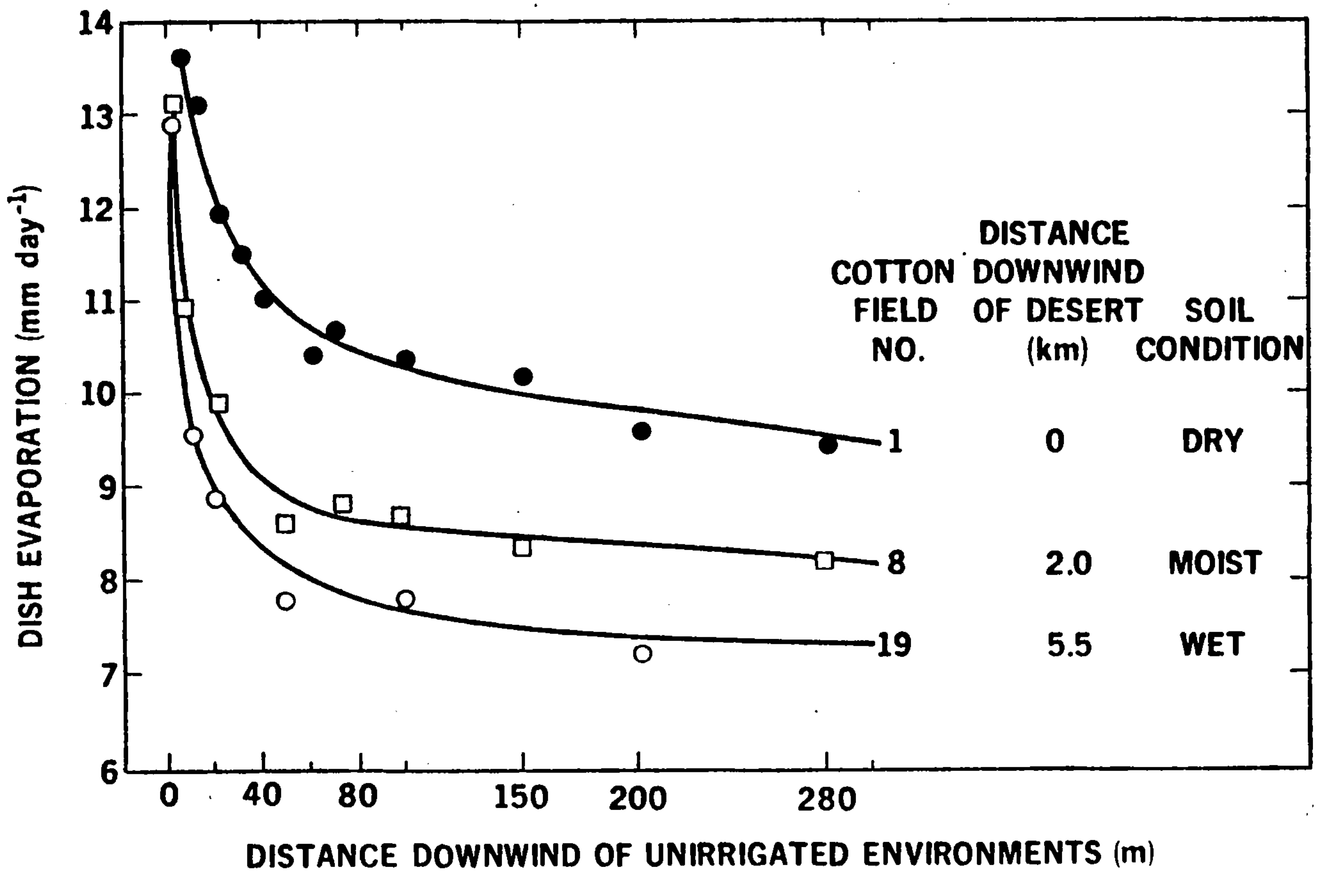
(2)


MONTHLY VALUES FOR
JANUARY TO JUNE INCLUSIVE
PAN EVAPORATION AT INDIO U.S. DATE GARDEN (mm day-1)
PAN EVAPORATION AT DEATH VALLEY (mm day-1)
0
5
10
15
20
25

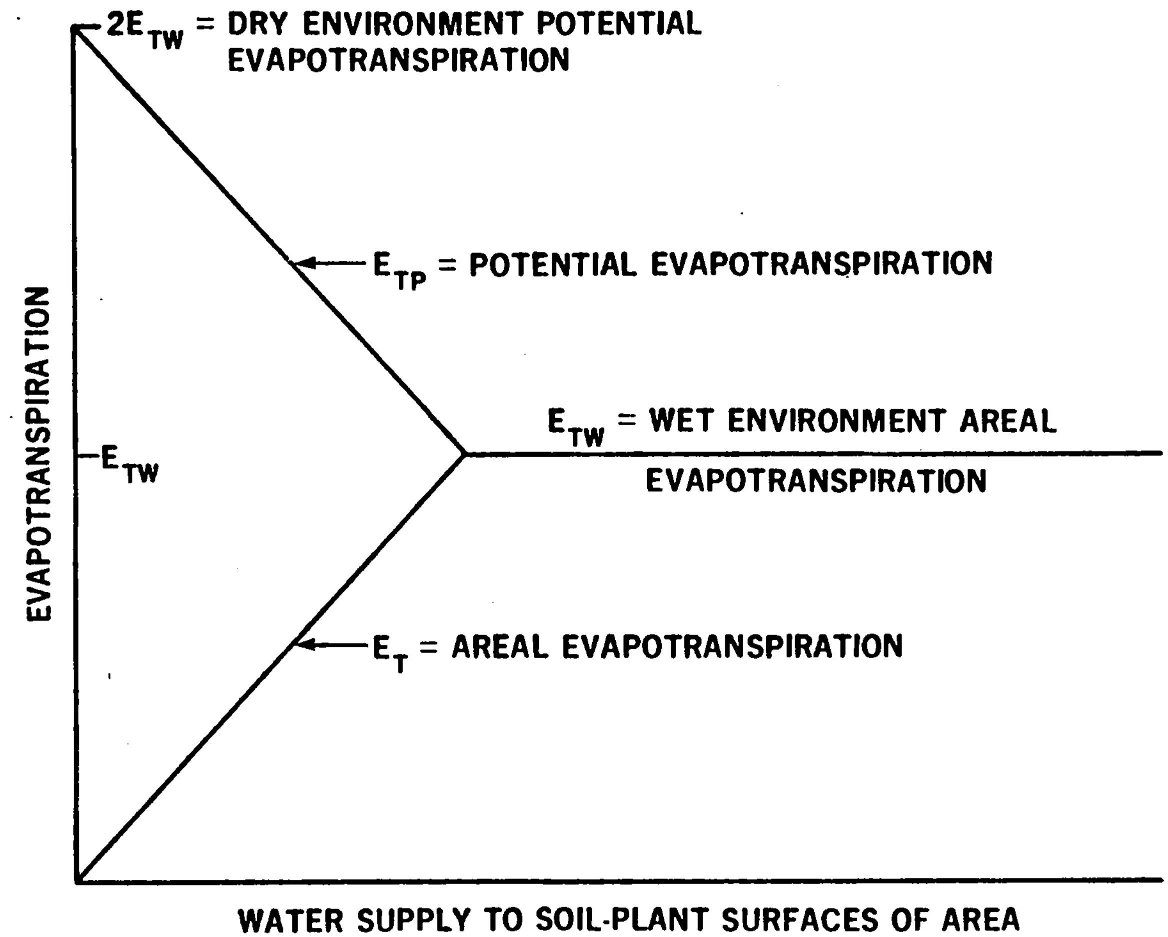
(3)


MONTHLY VALUES FOR
JULY TO DECEMBER INCLUSIVE
PAN EVAPORATION AT INDIO U.S. DATE GARDEN (mm day^{-1})
PAN EVAPORATION AT DEATH VALLEY (mm day^{-1})
0
5
10
15
20
25

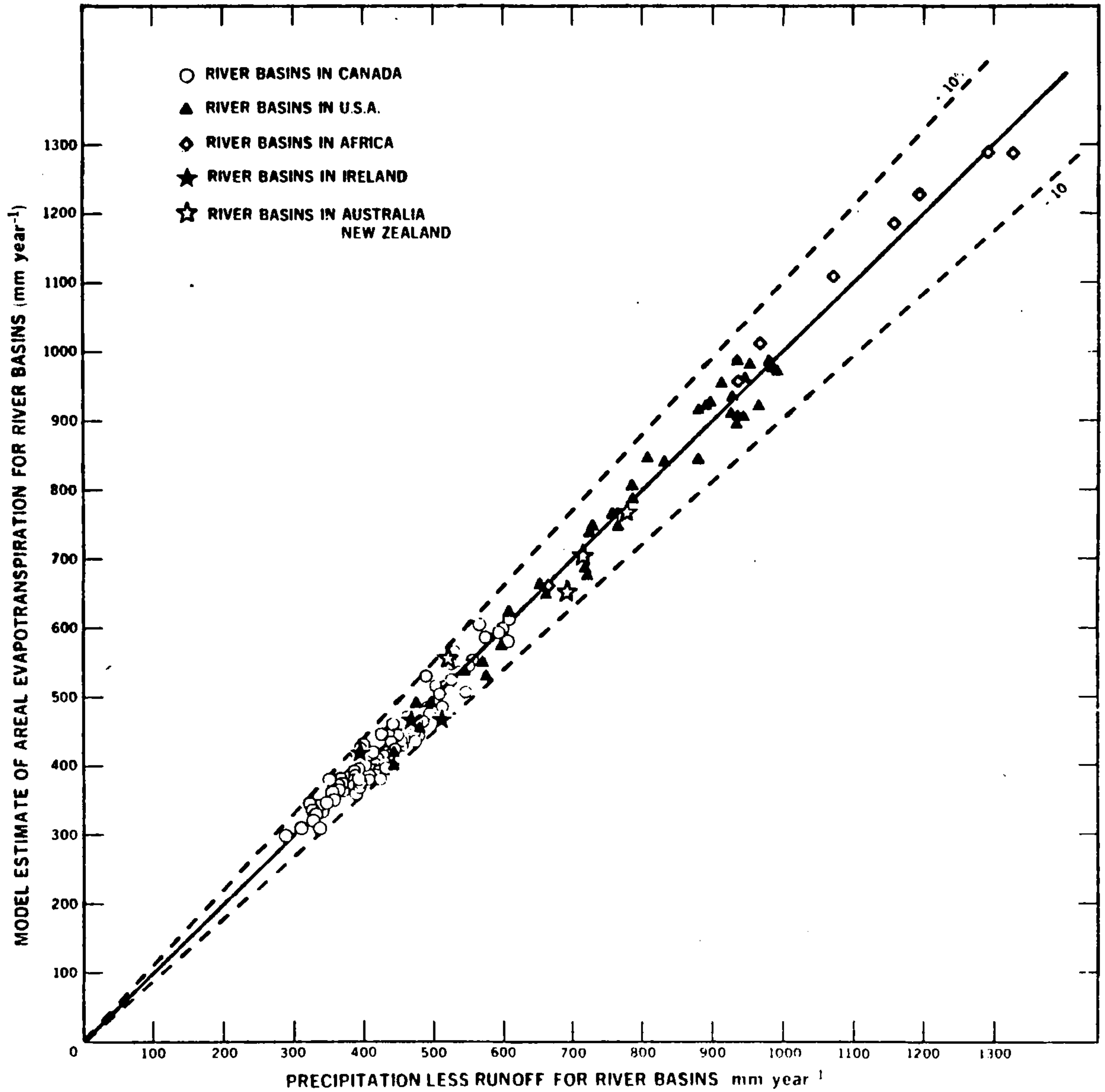
7

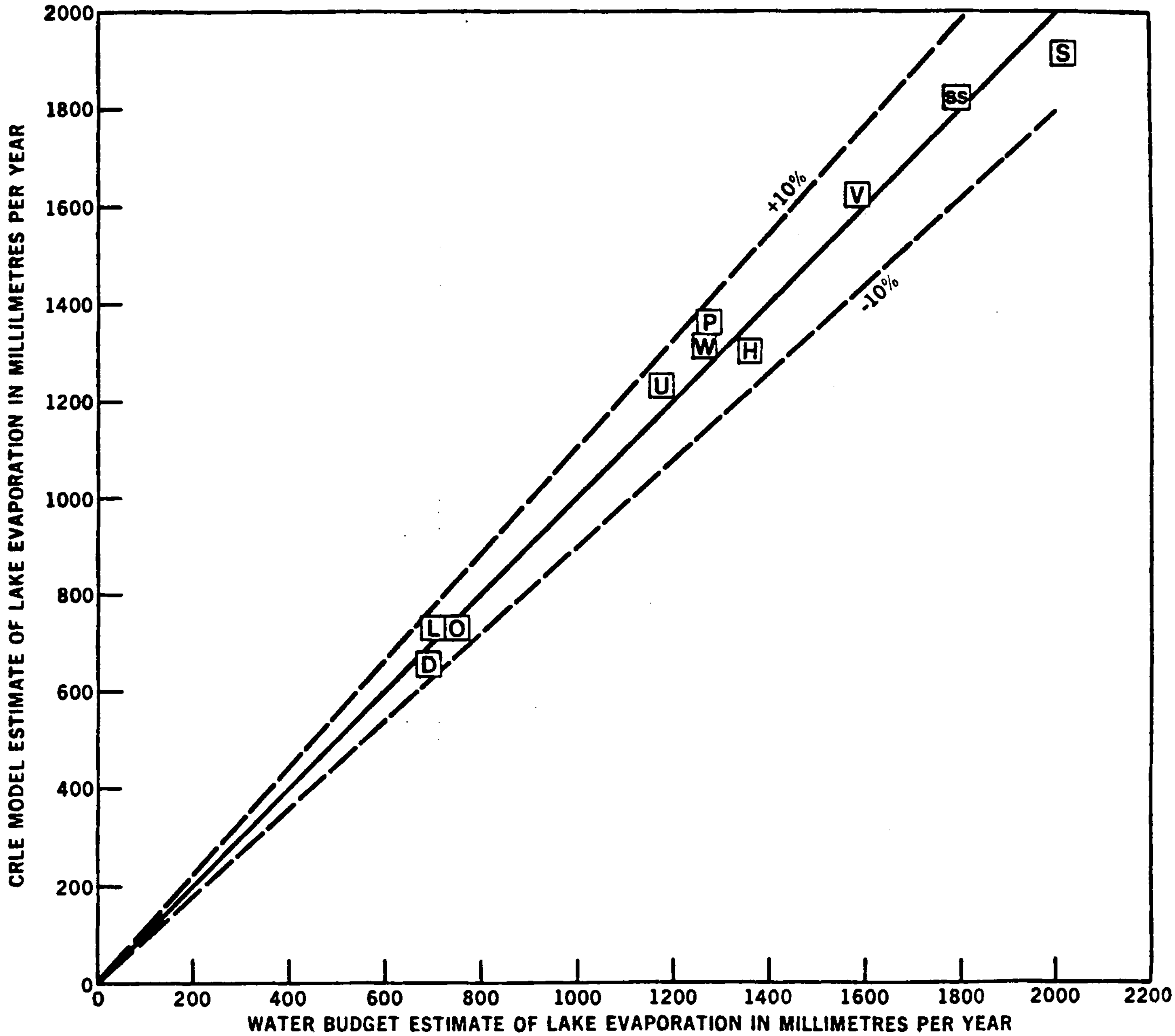


DISH EVAPORATION (mm day^{-1})
14
13
12
11
10
9
8
7
6
0
40
80
150
200
280
DISTANCE DOWNWIND OF UNIRRIGATED ENVIRONMENTS (m)
COTTON FIELD NO.
DISTANCE DOWNWIND OF DESERT (km)
SOIL CONDITION
1
0
DRY
8
2.0
MOIST
19
5.5
WET


2E_TW = DRY ENVIRONMENT POTENTIAL EVAPOTRANSPIRATION
E_TP = POTENTIAL EVAPOTRANSPIRATION
E_TW = WET ENVIRONMENT AREAL EVAPOTRANSPIRATION
E_TW
E_T = AREAL EVAPOTRANSPIRATION
EVAPOTRANSPIRATION
WATER SUPPLY TO SOIL-PLANT SURFACES OF AREA

5


RIVER BASINS IN CANADA
RIVER BASINS IN U.S.A.
RIVER BASINS IN AFRICA
RIVER BASINS IN IRELAND
RIVER BASINS IN AUSTRALIA
NEW ZEALAND
.10
.10
MODEL ESTIMATE OF AREAL EVAPOTRANSPIRATION FOR RIVER BASINS (mm year^{-1})
PRECIPITATION LESS RUNOFF FOR RIVER BASINS mm year^{-1}
0
100
200
300
400
500
600
700
800
900
1000
1100
1200
1300


CRLE MODEL ESTIMATE OF LAKE EVAPORATION IN MILLILMETRES PER YEAR
WATER BUDGET ESTIMATE OF LAKE EVAPORATION IN MILLIMETRES PER YEAR
+10%
-10%
S
BS
V
P
W
H
U
L
O
D
0
200
400
600
800
1000
1200
1400
1600
1800
2000
2200

OPERATIONAL ESTIMATES OF AREAL EVAPOTRANSPIRATION AND THEIR SIGNIFICANCE TO THE SCIENCE AND PRACTICE OF HYDROLOGY

II THE COMPLEMENTARY RELATIONSHIP

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ABSTRACT

Morton, F.I., 1982b. Operational estimates of areal evapotranspiration and their significance to the science and practice of hydrology: II The complementary relationship.

The complementary relationship predicts that the sum of the areal and potential evapotranspiration is equal to twice the wet environment areal evapotranspiration. It permits the areal evapotranspiration, a product of complex processes and interactions in the soil-plant-atmosphere continuum, to be estimated from its effects on the routine climatological observations that are required to estimate the other two quantities. An appraisal of the conceptual and empirical foundations for the complementary relationship demonstrates that it is realistic in the hydrometeorological context. A quickly converging iterative solution of the vapour transfer and energy balance equations for estimating potential evapotranspiration and the calibration of an equation for estimating wet environment areal evapotranspiration, using climatological data from desert regions where the monthly areal evapotranspiration approximates the monthly precipitation, are outlined. These formulations permit the complementary relationship to be used anywhere in the world with no further calibration. The development of complementary relationship models to provide operational estimates of areal evapotranspiration and lake evaporation from routine climatological observations, the testing of such estimates against comparable water budget values and their potential use in revitalizing the science and practice of hydrology are presented in subsequent companion papers.

INTRODUCTION

Advances in the science and practice of hydrology have been impeded by a dearth of reliable areal evapotranspiration estimates. Direct measurements, such as those provided by lysimeters, the eddy flux technique or the Bowen-ratio technique, give point values, require constant attendance by skilled personnel and are based on unverified assumptions. The water budget method provides good estimates for several years or more but requires excessive instrumentation and manpower for shorter term estimates. A companion paper (Morton, 1982a) presents a critical review indicating that the current conceptual modelling techniques, as exemplified in the SACREMENTO WATERSHED MODELLING SYSTEM, are based on assumptions about the soil, the vegetation and the atmosphere that are incompatible with published evidence and that causal modelling techniques that take into account the complex processes and interactions in the soil-plant-atmosphere continuum are not expected to have practical applications in the next generation. The review also indicates that the models based on the complementary relationship between potential and actual areal evapotranspiration can do much to fill the gap until such time as the causal models become practicable.

The chief advantage of the complementary relationship is that it permits the areal evapotranspiration, a product of complex processes and interactions in the soil-plant-atmosphere continuum, to be estimated by its effects on the routine climatological observations that are used to compute potential evapotranspiration. Because the model avoids the complexities of the soil-plant system it requires no local optimization of coefficients and is, therefore, falsifiable. This means that it can be tested rigorously so that errors in the associated assumptions and relationships can be detected and corrected by progressive testing over an ever-widening range of environments. Such a methodology uses the entire world as a laboratory and requires that a correction made to obtain agreement between model and river basin water budget estimates in one environment must be applicable without modification in all other environments. The estimates resulting from the most recent application of this methodology agree closely with comparable long-term water budget estimates for 143 river basins in North America, Africa, Ireland, Australia and New Zealand.

The complementary relationship also takes into account the modification of the air in passing from land across a lake. This provides further advantages in that minor changes can transform the complementary relationship areal evapotranspiration (CRAE) model to a complementary relationship lake evaporation (CRLE) model that permits lake evaporation to be estimated from routine climatological observations in the land environment. Results of the latest CRLE model have been tested against comparable water budget estimates for nine lakes in North America and two lakes in Africa.

The conceptual and empirical foundations for the complementary relationship and the developments needed to make it useable are presented herein. The formulation of the latest CRAE and CRLE models, the way they are used to provide operational estimates of areal evapotranspiration and lake evaporation, the test of such estimates against comparable water budget values and the potential use of such estimates to provide the impetus for transforming hydrology from a descriptive to a predictive science are presented in companion papers (Morton 1982c, 1982d, 1982e).

The conceptual basis for the complementary relationship, which includes a schematic model of the lower atmosphere and a mathematical rationalization, is presented in the Appendix.

COMPLEMENTARY RELATIONSHIP

Potential evapotranspiration is normally estimated from a solution of the energy balance equation and the vapour transfer equation for a moist surface. These equations can be solved by an approximate analytical technique (Penman, 1948) a graphical technique (Ferguson, 1952) or an iterative technique like the one presented herein. Kohler, Nordenson and Fox (1955) have shown that the Penman (1948) equation can provide good estimates of Class A pan evaporation if modified to take into account the difference between vegetated and water surfaces and the effects of sensible heat flux through the wall and bottom of the pan. Therefore it seems likely that the evaporation from any small water-filled container will provide a reasonable reflection of potential evapotranspiration, although the reflection will be somewhat distorted by differences in albedo, in surface roughness and in the ratio of latent heat transfer area to sensible heat transfer area.

The complementary relationship is based on a feedback mechanism whereby a decrease in areal evaporation caused by a reduction in the availability of water will increase the temperature and decrease the humidity of the overpassing air and this in turn will increase the potential evapotranspiration. Interactions such as these invalidate the conventional assumption that in any relationship between areal evapotranspiration and potential evapotranspiration the latter is the independent variable. Thus the use of potential evapotranspiration as a causal factor or forcing function is limited to two specific cases: (1) to existing moist areas so large that the effects of the evapotranspiration on the temperature and humidity of the overpassing air are fully developed and the areal evapotranspiration equals the potential evapotranspiration; and (2) to existing or hypothetical moist areas so small that the effects of the evapotranspiration on the temperature and humidity of the overpassing air are negligible. Pans and other evaporimeters fit into the latter category insofar as size is concerned.

In Bernouilli's equation for open-channel flow, the potential energy responds negatively to changes in the kinetic energy. Thus there appear to be analogies between potential evapotranspiration and potential energy and between areal evapotranspiration and kinetic energy. To be completely analogous the relationship would have to be complementary, i.e. the responses would have to be opposite in sign and equal, and both observational data and theory show that this is a distinct possibility. However it should be noted that the analogy cannot be extended into the critical flow regime because areal evapotranspiration cannot exceed potential evapotranspiration.

The oasis effect is the best known phenomenon associated with the foregoing interactions. It can be demonstrated by comparing monthly evaporation from two U.S. Weather Bureau Class A pans in the southeast desert basin of California during the years 1962, 1963 and 1964. Figs 1 and 2 show the pan evaporation in a large well-irrigated area at Indio U.S. Date Garden plotted against the pan evaporation in the desert at Death Valley for the first six months and the last six months of the calendar year respectively. The plotted regression lines have coefficients of correlation of 99 percent. The small zero-intercepts can be attributed to the effect of the 2.7 degree difference in latitude in increasing the winter radiation inputs at Indio so that the slopes represent the effects on pan evaporation of increasing the areal evapotranspiration from a near-zero value in the desert to a near-maximum value in an irrigated oasis. Thus Fig. 1 shows that the oasis pan evaporation is only 57 percent of the desert pan evaporation during the first six months of the calendar year and Fig. 2 shows that the oasis pan evaporation is only 52 percent of the desert pan evaporation during the last six months of the calendar year.

The oasis effect and the negative relationship between actual and potential evapotranspiration have been demonstrated by Davenport and Hudson (1967). They measured the variation in evaporation across a series of irrigated cotton and unirrigated fallow fields in the Sudan Gezira, using fiberglass dishes with black painted wells 113 mm in diameter and 36 mm in depth. The dish evaporation observations provided a somewhat distorted reflection of the potential evapotranspiration. The passage of air from the upwind desert (and/or the upwind unirrigated fallow fields) over the irrigated cotton caused the dish evaporation to decrease rapidly and approach a low constant value within 300 m - the width of the fields. Furthermore, as the air passed from irrigated cotton across unirrigated fallow the dish evaporation increased rapidly and approached a high constant value within 300 m. Fig. 3 shows the variation of dish evaporation across three irrigated fields on December 27, 1963. The description of the soil condition is presumably based on visual inspection and is related to the number of days since the last irrigation. Thus it is reasonable to assume that the actual evapotranspiration is greater in the "moist" field than in the "dry" field and that it is somewhere near its maximum possible value in the "wet" field.

At the upwind edge of the irrigated fields, where the dish evaporation decreases rapidly, the hot dry air from the desert (or the unirrigated fallow) loses heat and gains vapour from contact with the transpiring cotton leaves. Downwind from the transitional zone, where the dish evaporation approaches a low constant value, the effects of the evapotranspiration on the temperature and humidity of the overpassing air are well developed and approaching equilibrium. (Decreases in temperature and increases in humidity as the air moved across the irrigated cotton were observed over one field. The vapour pressure appeared to attain equilibrium values within 300 m but the temperatures were still decreasing, possibly because the observations were made above the level of the crop and dishes.) As it is only in an equilibrium zone that the term areal evapotranspiration has any meaning, the minimum size of an area for which the term is applicable is one in which the edge effects in the upwind transitional zone become insignificant.

The ratio of daily dish evaporation at the downwind edges of the individual irrigated cotton fields to that at the upwind edge of the irrigated area was 0.69 for the field with "dry" soil, 0.60 for the field with "moist" soil and 0.53 for the field with "wet" soil. This provides good evidence that the dish evaporation, and presumably the potential evapotranspiration, respond negatively to changes in areal evapotranspiration induced by changes in the availability of water.

In Fig. 3 the dish evaporation for the "wet" field provides an indication of what happens over a lake in an arid climate. Thus the upwind dish evaporation reflects the potential evaporation in the desert and the low, relatively constant dish evaporation near the downwind edge reflects the potential evaporation over most of the lake. Furthermore, the dish evaporation from the "moist" and "dry" fields provides an analogy for what happens over lakes in progressively more humid climates where the contracts between lake and land environments are less extreme. Because the transition zone is so narrow, the lake evaporation would approximate the low constant downwind value of potential evaporation.

If the negative reactions of potential evapotranspiration to changes in the availability of water for areal evapotranspiration are complementary, as suggested by the analogy to the Bernouilli theorem for open channel flow, the relationship is expressed in mathematical form by:

$$E_T + E_{TP} = 2E_{TW} \tag{1}$$

in which E_T is the areal evapotranspiration, the actual evapotranspiration from an area so large that the effects of upwind boundary transitions, such as those shown in Fig. 3, are negligible; E_{TP} is the potential evapotranspiration, as estimated from a solution of the vapour transfer and energy balance equations, representing the evapotranspiration that would occur from a hypothetical moist surface with radiation absorption and vapour transfer characteristics similar to those of the area and so small that the effects of the evapotranspiration on the overpassing air would be negligible; and E_{TW} is the wet environment areal evapotranspiration, the evapotranspiration that would occur if the soil-plant surfaces of the area were saturated and there were no limitations on the availability of water.

Fig. 4 provides a schematic representation of eq. (1) under conditions of constant radiant energy supply. The ordinate represents evapotranspiration and the abscissa represents the water supply to the soil-plant surfaces of the area, a quantity that is usually unknown. When there is no water available for areal evapotranspiration (extreme left of Fig. 4) it follows that $E_T \approx 0$, that the air is very hot and dry and that E_{TP} is at its maximum rate of $2E_{TW}$ (the dry environment potential evapotranspiration). As the water supply to the soil-plant surfaces of the area increases (moving to the right in Fig. 4) the resultant equivalent increase in E_T causes the overpassing air to become cooler and more humid which in turn produces an equivalent decrease in E_{TP}. Finally, when the supply of water to the soil-plant surfaces of the area has increased sufficiently, the values of E_T and E_{TP} converge to that of E_{TW}.

The conventional definition for potential evapotranspiration is the same as the definition for the wet environment areal evapotranspiration. However, the potential evapotranspiration that is estimated from a solution of the vapour transfer and energy balance equations by analytical (Penman, 1948), graphical (Ferguson, 1952) or iterative (see next section) techniques has reactions to changes in the water supply to the soil-plant surfaces similar to those shown for E_{TP} in Fig. 4 so that what is being estimated can exceed what is being defined by as much as 100 percent. By taking into account such reactions the complementary relationship is analogous to the Bernouilli equation for open channel flow in which the potential energy responds in a complementary way to changes in kinetic energy.

The evaporation from a shallow lake, E_W, differs from the wet environment areal evapotranspiration, E_{TW}, only because the radiation absorption and vapour transfer characteristics of water differ from those of vegetated land surfaces. The potential evaporation, E_P, differs from the potential evapotranspiration, E_{TP}, for the same reasons. Although the lake evaporation is equal to the potential evaporation in the lake environment it can differ significantly from the potential evaporation in the land environment.

Fig. 5 provides a schematic representation of the relationship between shallow lake evaporation and the potential evaporation in the land environment under conditions of constant radiant energy supply. The ordinate represents evaporation and the abscissa represents the water supply to the soil-plant surfaces of the land environment. Since a lake is defined to be so wide that the effects of the kind of upwind transition shown in Fig. 3 are negligible, the lake evaporation is independent of variations in the water supply to the soil-plant surfaces of the land environment. However, the complementary relationship predicts that the potential evaporation in a completely dry land environment would be twice the lake evaporation and that it would decrease in response to increases in the water supply to the soil-plant surfaces until it reached a minimum equal to the lake evaporation as shown in Fig. 5.

A tabular summary of the literature on the complementary relationship since it was introduced by Bouchet (1963) is appended to the preceding companion paper (Morton, 1982a). Although there is little doubt that the relationship is negative there is no "proof" that it is complementary. However the most important assumption underlying the relationship is compatible with the schematic model of the lower atmosphere that is documented herein as an Appendix. The schematic model includes an equilibrium sublayer, in which the gradients of temperature and humidity adjust to the surface fluxes of heat and water vapour; and a mixed sublayer, in which the heat and vapour fluxes are mixed by free buoyant convection, with no significant potential temperature or specific humidity gradients, under an overlying regional potential temperature inversion. The heights of the equilibrium and mixed sublayers have diurnal and seasonal patterns with typical summer afternoon values of 30 m and 800 m respectively. According to this model the large-scale weather systems influence potential evapotranspiration indirectly through their effects on the radiant energy and precipitated water received at the surface. However the direct effects are negligible because they are manifested by changes in the stability of the regional inversion and because the changes cause increases or decreases in the equilibrium sublayer temperature and in the equilibrium sublayer humidity that tend to be compensatory. Therefore the effects of changes in the availability of water for areal evapotranspiration on the potential evapotranspiration can be assessed by their effects on the temperature and humidity gradients in the equilibrium sublayer.

The complementary relationship is explicable only in terms of the meteorological theory of advection which is, unfortunately, more qualitative than quantitative. Therefore, attempts to provide it with a mathematical formulation, such as those presented by Bouchet (1963), Morton (1971) and Seguin (1975), depend on too many unverified and unverifiable assumptions to be classified as anything other than rationalizations. However, the appended schematic model of the lower atmosphere and the appended improved version of the previously published rationalization (Morton, 1971) can be combined to show that the complementary relationship has some basis in reality and to demonstrate how it works.

The complementary relationship cannot be verified with currently available theoretical knowledge and empirical evidence. However, in addition to the rationalizations mentioned previously, there is direct evidence for its plausibility. Figs. 6 and 7 were prepared in the same format as Fig. 4 by Solomon (1967) to show annual data from four river basins in Malawi. The availability of water is represented by the precipitation, the evapotranspiration (areal) is represented by the difference between precipitation and runoff and the potential evaporation is estimated from climatological data using the Penman (1948) equation. The sum of the potential evaporation and evapotranspiration remains relatively constant from basin to basin and from year to year. The lines of best fit are curvilinear rather than linear as shown in Fig. 4, because annual precipitation is greater than the water available for evapotranspiration and the difference increases with increases in the precipitation. Fig. 8 is a similar plot prepared by Giusti (1978) using average annual data for a number of closed river basins in various parts of Puerto Rico. In Fig. 8 the potential evapotranspiration is represented by pan evapotranspiration. All three figures show that the relationship between annual values of potential and areal evapotranspiration is negative in tropical countries like Malawi and Puerto Rico and there is a good probability that it is complementary.

Army and Ostle (1957) have inadvertently provided more direct evidence for the existence of a complementary relationship between areal and potential evapotranspiration. They calculated actual areal evapotranspiration (E_T) from 0.1 acre plots of spring wheat by adding the total precipitation from seeding to harvest to the amount of soil moisture lost from seeding to harvest for 29 years at Huntley, Montana and for 26 years at Havre, Montana. The results were then related to the corresponding values of evaporation from nearby sunken B.P.I. tanks (E_{PW}), which provide a somewhat distorted reflection of potential evapotranspiration. The resultant regression equations for plots that were cropped every year were:

$$E_T \text{ (at Huntley)} = 30.8 - 1.025\, E_{PW} \qquad (2)$$

$$E_T \text{ (at Havre)} = 24.5 - 0.644\, E_{PW} \qquad (3)$$

and the regression equations for two plots that were cropped and left in fallow on alternate years were:

$$E_T \text{ (at Huntley)} = 34.6 - 1.142\ E_{PW} \tag{4}$$

$$E_T \text{ (at Havre)} = 28.5 - 0.725\ E_{PW} \tag{5}$$

The zero-intercepts for the four equations are in inches of water and the coefficients of correlation vary from 0.790 to 0.826 with a mean value of 0.814. In evaluating the results it should be noted that unreported differences in vegetation and soil immediately adjacent to the various plots and tanks could account for some of the differences between the equations; that unreported year-to-year variations of absorbed radiation would add to the scatter of the points and the uncertainty concerning the regression coefficients; and that the tank evaporation provides an overestimate of potential evapotranspiration (because of the low albedo of water) and would thereby produce an underestimate of the negative slopes. In the light of these considerations the negative slopes do not differ significantly from each other and the average value (-0.884) does not differ significantly from that required by the complementary relationship (-1.000).

In Fig. 2, the slope of the line that relates the evaporation from the pan in the irrigated area at U.S. Indio Date Garden, California, to the evaporation from the pan in the desert at Death Valley,

POTENTIAL EVAPOTRANSPIRATION

The potential evapotranspiration is estimated by solving for the potential evapotranspiration equilibrium temperature (hereinafter referred to as the equilibrium temperature or T_P), the temperature at which the energy balance equation and the vapour transfer equation for a moist surface give the same result. The analytical solution of Penman (1948) and the modification suggested by Kohler and Parmele (1967) are accurate only under relatively humid conditions where the equilibrium temperature is near the air temperature; the graphical solution of Ferguson (1952) and the modification suggested by Morton (1965) are more precise but are inappropriate to computer use; and trial-and-error or iterative solutions can require excessive computer time to achieve the required precision. The technique used herein produces numerical accuracy within four iterations by incorporating certain aspects of the analytical solutions.

Eqs. (7) and (8) represent the energy balance and vapour transfer equations respectively

$$E_{TP} = R_T - [\gamma p f_T + 4\varepsilon\sigma(T_p+273)^3]\,(T_p-T) = R_T - \lambda f_T(T_P - T) \qquad (7)$$

$$E_{TP} = f_T\,(v_P - v_D) \qquad (8)$$

$$f_T = (p_s/p)^{0.5}\ f_Z/\zeta \tag{12}$$

in which f_Z is a constant, ζ is a dimensionless stability factor and p and p_s are the atmospheric pressure and the sea level atmospheric pressure respectively.

The square root term in eq. (12) is included to represent the effect of altitude or atmospheric pressure on evaporation and the vapour transfer coefficient. It provides a better fit to Rowher's (1931) data for sunken pans at Pikes Peak, Victor, Fort Collins, (all in Colorado), Lake Tahoe, Imperial, (both in California); Logan (in Utah) and Fort Calhoun (in Nebraska) than Rowher's (1931) linear representation.

The stability factor is included to take into account the decrease in the vapour transfer coefficient that occurs when the temperature of the evaporating surface is much below the air temperature. This is usual during the winter in extra-tropical latitudes and during other seasons in very dry environments. The relationship used herein is:

$$1/\zeta = 0.28(1 + v_D/v) + \Delta R_{TC}/[\gamma p\ (p_s/p)^{0.5}\ b_0\ f_Z(v - v_D)] \tag{13}$$

$$1/\zeta \leq 1 \tag{13a}$$

in which b_0 is 1.00 for areal evapotranspiration, R_{TC} is R_T with $R_{TC} \geq 0$, and all other symbols are as defined previously.

Eq. (13) is based on the reasonable assumption that the ratio of the radiation term to the vapour transfer term in the Penman (1948) equation for potential evaporation provides a good index of the effects of atmospheric stability on the vapour transfer coefficient. The quantity $\Delta R_{TC}/[\gamma p\ (p_S/p)^{0.5}\ b_o\ f_Z\ (v - v_D)]$ provides a preliminary estimate of this ratio. The other more arbitrary term was selected to reduce non-linearity and scatter in the 154 sets of data used to estimate the constants $2b_1$ and $2b_2$.

The vapour transfer coefficient was assumed to be independent of wind speed because: (1) it increases with increases in both surface roughness and wind speed and wind speeds tend to be lower in rough areas than in smooth areas; (2) it increases with increases in the instability of the atmosphere and this effect is more pronounced at low wind speeds than at high wind speeds; and (3) the use of climatological observations of wind speed can lead to significant error because of local variations in exposure and instrument height. These considerations indicate that the use of routinely observed wind speeds in estimating evapotranspiration does not significantly reduce error and may in fact increase it. Therefore there is ample justification for assuming that the quantity f_Z is a constant.

The constant f_Z was derived by trial and error in conjunction with the calibration of constants $2b_0$ and $2b_1$ to meet the following criterion:

$$b_1 \; \overline{(1 + \gamma p/\Delta_p)/R_{TP}} + b_2 = 1.32 \qquad (14)$$

in which the bar denotes an average for the 94 station-months that fell within the northern hemisphere high radiation season of March to September inclusive (i.e. 116.3 Wm^{-2}) and 1.32 is the constant in the equation proposed by Priestley and Taylor (1972) that was revised upward in the discussion of eq. (10) to make it compatible with their land surface data and incompatible with their water surface data.

The application of the criterion in eq. (14) produced values of f_Z, b_1 and b_2 of 28 Wm^{-2} $mbar^{-1}$, 14 Wm^{-2} and 1.20 respectively. Fig. 9 shows the 154 values of dry environment potential evapotranspiration plotted against corresponding values of $(1 + \gamma p/\Delta_p)^{-1} R_{TP}$ and the regression line representing the solution of eq. (11) doubled.

It may be noted that the values of f_Z are expressed in energy units. As they are meant to represent a vapour transfer coefficient they should be increased at below-freezing temperatures by a factor of 1.15, the ratio of the latent heat of sublimation to the latent heat of vaporization.

The same data and the same procedure were used in calibrating eq. (11) to provide estimates of lake evaporation. The only differences involved the use of water surface albedo and emissivity in estimating potential and lake evaporation (Morton 1982e) and the selection criterion.

Thus the constant f_Z was derived by trial-and-error in conjunction with the calibration of the constants $2b_1$ and $2b_2$ in order to minimize the absolute average percentage error between the annual lake evaporation estimated from the CRLE model (Morton, 1982e) and the corresponding water budget estimates for Dauphin Lake in Manitoba, Lake Ontario in the North American Great Lakes System, Pyramid and Winnemucca Lakes in Nevada, Lake Hefner in Oklahoma, Silver Lake and Salton Sea in California and Lake Victoria in East Africa [descriptions and data sources for these and other lakes are appended to a companion paper (Morton, 1982e)]. The application of this criterion produced values of f_Z, b_1 and b_2 of 25 Wm^{-2} $mbar^{-1}$, 13 Wm^{-2} and 1.12 respectively. A plot comparable to that shown in Fig. 9, with dry environment potential evaporation replacing dry environment potential evapotranspiration, shows even less scatter to the points. It should be noted that the value of the constant b_0 in eq. (13) is 1.12 for lake evaporation and that this is the ratio of the vapour transfer coefficient for areal evapotranspiration to the vapour transfer coefficient for lake evaporation.

CONCLUDING DISCUSSION

The only concepts used in hydrology that have any well-founded physical basis are the law of conservation of mass, the law of conservation of energy and approximations to the law of conservation of momentum such as those represented by the Darcy formula and modifications of the Chezy formula. Practically all others are based on assumptions that contravene published and verified field evidence. Even that cornerstone of modern river hydrology, the unit hydrograph, is based on the assumption of linearity - an assumption that has been known as a questionable approximation, at best, for about two decades. In comparison, the complementary relationship between areal and potential evapotranspiration has a realistic physical basis. It is known that a reduction in the water available for areal evapotranspiration makes the overpassing air hotter and drier and it is known that hotter and drier air increases the potential evapotranspiration so there is no doubt that the relationship is negative. While there is no "proof" that the negative reactions are complementary, the analysis of dish, pan and plot data presented in a preceding section provides good evidence that the complementary relationship is a reasonable working hypothesis. Furthermore the relationship remains compatible with all current theoretical knowledge and reliable empirical evidence, despite strong scepticism in the hydrometeorological community since it was first published by Bouchet (1963). LeDrew (1979) presented evidence that purported to cast doubt on the complementary relationship but on closer examination (Morton, 1980) this turned out to be simulated data based on inadequate assumptions and selected fragmentary field data interpreted in a questionable way. Therefore, the complementary relationship remains one of the better substantiated concepts in hydrometeorology and its demonstrated ability to provide reliable independent estimates of areal evapotranspiration from readily available data over an extremely wide range of environments makes it possible at last to analyze river basin water balances and thereby obtain much of the knowledge needed to transform hydrology from a descriptive to a predictive science.

According to the complementary relationship the areal evapotranspiration, a product of complex processes and interactions in the soil-plant-atmosphere continuum, is equal to twice the wet environment areal evapotranspiration less the potential evapotranspiration and both of these quantities may be estimated from routine climatological observations. As described in the preceding sections, the potential evapotranspiration is estimated from a quickly converging solution of the vapour transfer and energy balance equations and the wet environment areal evapotranspiration is estimated from an equation that was calibrated, in conjunction with the vapour transfer coefficient used in estimating potential evapotranspiration, using monthly climatological data from arid areas under conditions where the monthly areal evapotranspiration could be assumed equal to monthly precipitation. To be useful these basic concepts must be associated with a number of assumptions and poorly defined empiricisms, most of which are required to provide radiation estimates from routinely observed data. The latest version of the resultant complementary relationship areal evapotranspiration (CRAE) model is presented in detail in the next companion paper (Morton, 1982c). The minor changes needed to change this to a complementary relationship lake evaporation (CRLE) model are described in a subsequent companion paper (Morton, 1982e).

Although most of the associated empiricisms were selected from the literature or were developed independently from published data it was necessary to use a certain amount of jugement in the selection process and in making minor adjustments to obtain greater generality (e.g. in using atmospheric pressure to estimate atmospheric radiation). As these empiricisms were used in the once-only calibration of the vapour pressure coefficient and the wet environment areal evapotranspiration, the effects of the selection process and the minor adjustments are implicit in the models. Fortunately, the models are falsifiable, with no need for locally optimized coefficients, so that many of the errors in the associated assumptions and empiricisms have been detected and corrected during the years by progressive testing over an ever-widening range of environments. This methodology has required that a correction made to obtain agreement between model and river basin water budget estimates in one environment must be applicable without modification in all other environments. Later companion papers (Morton, 1982d, 1982e) test the latest version of the CRAE models with comparable water budget estimates of areal evapotranspiration for 143 river basins in North America, Africa, Ireland, Australia and New Zealand and test the latest version of the CRLE models with comparable water budget estimates of lake evaporation for 11 lakes in North America and Africa.

APPENDIX

A CONCEPTUAL BASIS FOR THE COMPLEMENTARY RELATIONSHIP

Schematic Model of the Lower Atmosphere

The structure of the atmosphere surface layer is shown in Fig. 10 (Geiger, 1966). The figure is based on the work of K. Brock who summarized all available observations made to 1938 in Central Europe and Egypt in the lowest 100 m of the atmosphere. The logarithms of the hourly mean temperature lapse rates at midday for the months of June and December are plotted against the logarithms of the height. Fig. 10 shows that the lower atmosphere can be divided into a lower part, where the temperature lapse rate decreases with height due to the changeover from forced frictional convection near the ground to free buoyant convection higher up, and a higher part, where the predominance of free buoyant convection keeps the lapse rate constant at a value somewhat lower than the dry adiabatic lapse rate (represented by the broken line). The former is referred to as the equilibrium sublayer, while the latter is referred to as the mixed sublayer. The equilibrium sublayer disappears at night with the formation of a temperature inversion. According to Geiger (1966) it appears again in the morning when the sun's elevation reaches about 10°. From then on it increases in thickness, and so do the temperature gradients observed in it. During this period a substantial amount of the heat transferred from the ground is used to increase the thickness of the equilibrium sublayer. When the sun reaches an elevation of about 30°, the temperature gradient continues to increase, but the height of the equilibrium sublayer no longer increases, because by that time the free convective motion has become so vigorous that further supplies of heat from the ground are transferred mainly into layers higher up. The midday thickness of the equilibrium sublayer can vary from a value of about 4 m in winter to values between 30 and 40 m by the time the sun changes direction in the summer.

The thickness of the mixed sublayer can reach values of more than 2000 m although the average midday value in summer is nearer to 800 m. The diurnal and seasonal patterns of changes in thickness are similar to those of the equilibrium sublayer.

Fig. 10 shows only two of many similar plots. According to Geiger (1966), a frequency distribution of the slopes for the equilibrium sublayer shows a well marked maximum for a value of -1. This implies that the air temperature decreases logarithmically with the height above the ground. The logarithmic relationship does not permit extrapolation to the ground surface but it can be used to show how changes in temperature decrease with height as the equilibrium sublayer is expanding upward against an essentially isothermal mixed sublayer. Thus an increase in temperature of 2.00°C at 0.014 m above the ground would induce increases of 0.99°C at 1.22 m (the height of a meteorological screen) and 0.31°C at 25 m. Similar results could be expected if the flux of heat from the ground heats the mixed sublayer with no expansion in the height of the equilibrium sublayer.

The distribution of humidity in the lower atmosphere has not been defined so rigorously. However there is a great deal of evidence in Geiger (1966) and elsewhere that the daytime vertical profiles of vapour pressure or specific humidity are similar to those of temperature. This is because the same processes of forced and free convection that are responsible for upward transport of heat are also responsible for the upward transport of water vapour.

In the rationalization of the complementary relationship there is only one assumption that is of sufficient import to justify an intensive evaluation. It requires that the potential evapotranspiration be unaffected by advections (horizontal transports) of heat and water vapour from elsewhere except insofar as they effect the radiant energy and precipitation at the surface. According to the McIlroy (1971) version of the Penman (1948) equation this can be achieved if the advections do not change the wet bulb temperature depression at the instrument level; i.e. if the change in humidity changes the wet bulb temperature by the same amount as the change in dry bulb temperature. In the case of small-scale advection, such as that shown by Davenport and Hudson (1967), the effects must be avoided by staying well down-wind of sharp environmental discontinuities. In the case of advection associated with large-scale weather systems, the effects can be assessed only in terms of the growth and evolution of the mixed sublayer.

Fig. 11 is a compositive of three plots published by Milford, Abdulla and Mansfield (1979) showing the results of temperature and humidity observations made over southern England on September 12, 1973. It shows the potential temperature (i.e. the temperature adjusted to the 1000 mbar level using the dry adiabatic lapse rate) plotted against height for 5 daylight soundings and the humidity mixing ratio plotted against height for 4 of the same 5 soundings. Fig. 12 is a composite of two plots presented Deardorff (1974) showing similar profiles for 4 daylight soundings during Day 33 of the Australian Wangara Experiment (Clarke et al. 1971). This is comparable to Fig. 11 because the specific humidity is practically identical to the humidity mixing ratio. Coulman (1978) has published similar profiles for other Australian data and those for May 21, 1974 are shown in Fig. 13. The virtual potential temperatures for 0937 hours exceeded the potential temperatures by 1.10° at a height of 100 m and 0.80° at a height of 1400 m. The corresponding figures for 1316 hours are 0.93° and 0.87° respectively.

During the night, when the air in contact with the ground is cooled, a nocturnal inversion is formed. This is evident in the early morning potential temperature profiles in Figs. 11, 12 and 13. When the sun rises in the morning the surface heats up, thereby inducing forced convection near the ground and free convection further up in the mixed sublayer. The mixed sublayer, where both the potential temperature and specific humidity are nearly constant, expands upward in response to increasing heat fluxes from the surface, its height at any time being the height at which the overlying inversion has the same potential temperature as the mixed sublayer. All three figures have a sharp height discontinuity between the nocturnal inversion and the regional inversion, and this causes a sudden expansion in the height of the mixed sublayer during the morning. The regional inversion, which is primarily the result of large-scale weather systems, is weaker and less stable than the nocturnal inversion in Fig. 12 and stronger and more stable than the nocturnal inversion in Figs. 11 and 13. The increase in potential temperature that results from a given afternoon surface heat flux is higher for a strong regional inversion than for a weak regional inversion. Heat loss from the mixed sublayer is primarily by radiation and has a night-time maximum when the surface is cooler than the air. Fig. 12 shows the first step in the formation of the nocturnal inversion as the surface cools off at 1800 hours.

Figs. 11, 12 and 13 show that strong mixing tends to keep the specific humidity constant between the top of the equilibrium sublayer and the potential temperature inversion. The specific humidity decreases abruptly across the potential temperature inversion thereby creating the high lapse rate needed for the inefficient vertical transport of water vapour from the mixed sublayer to the free atmosphere. All three figures show a somewhat different humidity pattern in the morning with an abrupt discontinuity between high values at low levels and low values at high levels. Fig. 13 shows that the increase at low levels continues until 1140 hours. This is thought to be due to the continuation of upward vapour flux during periods when there is no upward heat flux so that water vapour is trapped under the nocturnal inversion. Fig. 13 shows that this concentration of water vapour at lower levels disperses rapidly with the sudden expansion in the height of the mixed layer from that of the nocturnal inversion to that of the regional inversion.

From the viewpoint of the hydrologist, the mixed sublayer appears to act as a large reservoir that stores the widely varying heat and vapour fluxes from the surface and releases them through delayed radiation and/or inefficient transport across a potential temperature inversion. Thus the effects of the surface heat and vapour fluxes on the temperature and humidity at the bottom of the mixed sublayer, i.e. the top of the equilibrium sublayer, are very much less than they are near the surface where routine observations are made. Furthermore the mixed sublayer buffers the potential evapotranspiration from the effects of large-scale weather systems that are manifested primarily through the strength of the regional inversion. Thus an increase in the strength of the regional inversion traps both the heat and vapour fluxes at lower levels so that the effect of the increased temperature on the potential evapotranspiration is offset to a large extent by the effect of the increased humidity. Of course it takes some time for the mixed sublayer to come back into equilibrium with the surface after the passage of a weather system and during this time the potential evapotranspiration would be in a perturbed state.

There are three complications to the foregoing schematic model of the lower atmosphere:

(1) During periods of negative net radiation, potential temperature and specific humidity inversions persist down to the surface so that the potential evapotranspiration may be influenced directly by the advection of heat and water vapour associated with large-scale weather systems. However vertical fluxes under inversion conditions are notoriously inefficient so that the effects can be neglected unless they persist for long periods of time, as in the fall and winter. Even then they are quite small, so that they can be taken into account empirically.

(2) When the regional inversion is weak or non-existent the mixed sublayer may expand up to the condensation level with the resultant

formation of cumulus clouds. However, because the vapour used to form a particular cloud converges from a much larger area, only the top of the mixed layer is affected (Roll, 1965) so that the direct effects of cumulus convection on the potential evapotranspiration are probably negligible.

(3) Convection in the mixed sublayer is sustained by ascending thermals that penetrate into the nocturnal or regional inversions. The air that is displaced downward makes the mixed sublayer hotter and drier. The significance and extent of such effects are the subject of much controversy, but the prevalence of temperature lapse rates slightly less than adiabatic (Deardorff, 1972) and specific humidity lapse rates somewhat greater than zero provide some evidence that they are confined to the top part of the mixed sublayer. Therefore even if they are significant in the overall meteorological context they would have little effect on the potential evapotranspiration.

The foregoing complications detract little from the validity of the simple schematic model of the lower atmosphere that is basic to the conceptual rationalization of the complementary relationship. In this model the large-scale weather systems influence potential evapotranspiration indirectly through their effects on the radiant energy and the precipitated water received at the surface. However, the direct effects are negligible because they are manifested by changes in the stability of the regional inversion and because the changes cause increases or decreases in the equilibrium sublayer temperature and in the equilibrium sublayer humidity that tend to be compensatory. Therefore, the effects of changes in the availability of water for areal evapotranspiration on the potential evapotranspiration can be assessed by their effects on the temperature and humidity gradients in the equilibrium sublayer.

Mathematical Rationalization

With adequate moisture supplies to the soil and vegetation surfaces of an area, the areal evapotranspiration is equal to the potential evapotranspiration. However when availability of moisture limits areal evapotranspiration the resultant changes in the fluxes of heat and water vapour alter the lower atmosphere. Some of the results are so small that they are assumed to be negligible. These include changes in atmospheric radiation and in the vapour transfer coefficient. For the relationship between areal and potential evapotranspiration, the governing factors are the changes in the temperature and humidity of the equilibrium sublayer. If it is assumed that equality of eddy diffusivities for heat and water vapour prevails throughout the equilibrium sublayer, the changes are reflected in the following heat and vapour transfer equations.

$$[\gamma p f_{TL}/(\gamma p f_{TL}+4\epsilon\sigma T_L^3)]\ \delta E_T = -\delta[\gamma p f_{TU}(T_L-T_U)] = -\gamma p f_{TU}(\delta T_L-\delta T_U) \qquad (15)$$

$$\delta E_T = \delta[f_{TU}(v_{DL}-v_{DU})] = f_{TU}(\delta v_{DL}-\delta v_{DU}) \qquad (16)$$

in which E_T is the areal evapotranspiration, T_L and T_U are absolute air temperatures at a lower and an upper reference level respectively, v_{DL} and v_{DU} are dew point saturation vapour pressures at the lower and the upper reference levels respectively, f_{TL} and f_{TU} are vapour transfer coefficients between the surface and the lower reference level and between the lower reference level and the upper reference level respectively, γ is the psychrometric constant, p is the atmospheric pressure, ϵ is the surface emissivity, σ is the Stefan-Boltzman constant and δ is the partial differential with respect to changes in the supply of moisture to the evaporating surfaces of the area.

The negative sign in eq. (15) shows that energy is made available when δE_T is negative and that energy is used up when δE_T is positive. The ratio on the left side of eq. (15) represents the proportion of this energy that becomes sensible heat thereby changing the heat content of the lower atmosphere. The remainder of the energy causes changes in the net long-wave radiation. Eq. (16) describes changes in vapour transfer and vapour content of the lower atmosphere despite the use of energy units.

The lower reference level is defined as the height above the surface at which the average effect of surface temperature changes would be observed, i.e. the height at which the change in air temperatures is half the change in surface temperatures. The discussion of the logarithmic temperature profiles in the equilibrium sublayer has indicated that this occurs at a height that approximates the height of a meteorological screen. Although the height of the lower reference level might be subject to minor changes these have little effect since there would be compensatory changes in the vapour transfer coefficients f_{TL} and f_{TU}.

The upper reference level should be near enough to the top of the equilibrium sublayer for the ratios $\delta T_U/\delta T_L$ and $\delta v_{DU}/\delta v_{DL}$ to be much lower than one. As they are both of the same sign they can be assumed equal so that the solution of eqs. (15) and (16) is;

$$\delta v_{DL} = -\lambda_{TL}\,\delta T_L \qquad (17)$$

in which $\lambda_{TL} = \gamma p + 4\varepsilon\sigma T_L^3/f_{TL}$

The effects of δv_{DL} and δT_L on the potential evapotranspiration, E_{TP}, are estimated from the heat and vapour transfer equations for a hypothetical potential evapotranspirimeter at a representative location in the area, i.e.

$$[\gamma p f_{TL}/(\gamma p f_{TL}+4\varepsilon\sigma T_L^3)]\ \delta E_{TP} = -\delta[\gamma p f_{TL}(T_P-T_L)] = -\gamma p f_{TL}(\delta T_P-\delta T_L) \qquad (18)$$

$$\delta E_{TP} = \delta[f_{TL}(v_P - v_{DL})] = f_{TL}(\delta v_P - \delta v_{DL}) \qquad (19)$$

in which T_P = potential evapotranspiration equilibrium temperature, i.e. the temperature at the surface of a hypothetical potential evapotranspirimeter, v_P = saturation vapour pressure at T_P and all other symbols are as defined previously. Eq. (18) can be abbreviated to

$$\delta E_{TP} = \lambda_{TL} f_{TL} (\delta T_L - \delta T_P) \tag{20}$$

The solution of eqs. (17), (19) and (20) shows that T_P does not change in response to changes in the availability of water to the evaporating surfaces of the area. With $\delta v_P = \delta T_P = 0$

$$\delta E_{TP} = \lambda_{TL} f_{TL} \delta T_L = -f_{TL} \delta v_{DL} \tag{21}$$

The change in air temperature due to a change in the availability of water at the evaporating surfaces of the area is caused by a change in the temperature of the evaporating surfaces. the interactions are obtained from the sensible heat transfer equation for the areal surfaces

$$[\gamma p f_{TL} / (\gamma p f_{TL} + 4\varepsilon\sigma T_L^3)] \; \delta E_T = -\delta[\gamma p f_{TL} (T_T - T_L)] = -\gamma p f_{TL} (\delta T_T - \delta T_L) \tag{22}$$

in which T_T is the temperature of the areal surfaces and all other symbols are as previously defined.

The equilibrium heat content of the lower atmosphere is changed by the increment of sensible heat transfer from the areal surfaces. As the air passes over the potential evapotransprimeter the change in atmospheric heat content and the unchanged surface temperature induce an increment of sensible heat transfer that is opposite in sign and probably equal in magnitude to the areal increment. That is:

$$\gamma p f_{TL}(\delta T_T - \delta T_L) = -\gamma p f_{TL}(\delta T_p - \delta T_L) = \gamma p f_{TL}\delta T_L \qquad (23)$$

Eq. (23) requires that $\delta T_T = 2\delta T_L$. This is compatible with the definition of the lower reference level. As noted previously it is also consistent with a logarithmic temperature profile and a lower reference level near the height of a meteorological screen.

The solution of eqs. (22) and (23) is:

$$\delta E_T = -\lambda_{TL} f_{TL}\delta T_L \qquad (24)$$

The complementary relationship is obtained by summing eqs. (21) and (24) to obtain

$$\delta E_T + \delta E_{TP} = \lambda_{TL}\ f_{TL}(-\delta T_L + \delta T_L) = 0 \qquad (25)$$

Eq. (25), in its final form, does not take into account any data or properties pertaining to the lower reference level. Therefore the exact height of the lower reference level or its variations in time are not essential to the rationalization. It is possible that some of the other equations, e.g. eqs. (21) and (24), may eventually prove useful in remote sensing applications, in which case the data and properties pertaining to the meteorological screen level should approximate those of the lower reference level.

The solution for eq. (25) is the complementary relationship as presented in eq. (1).

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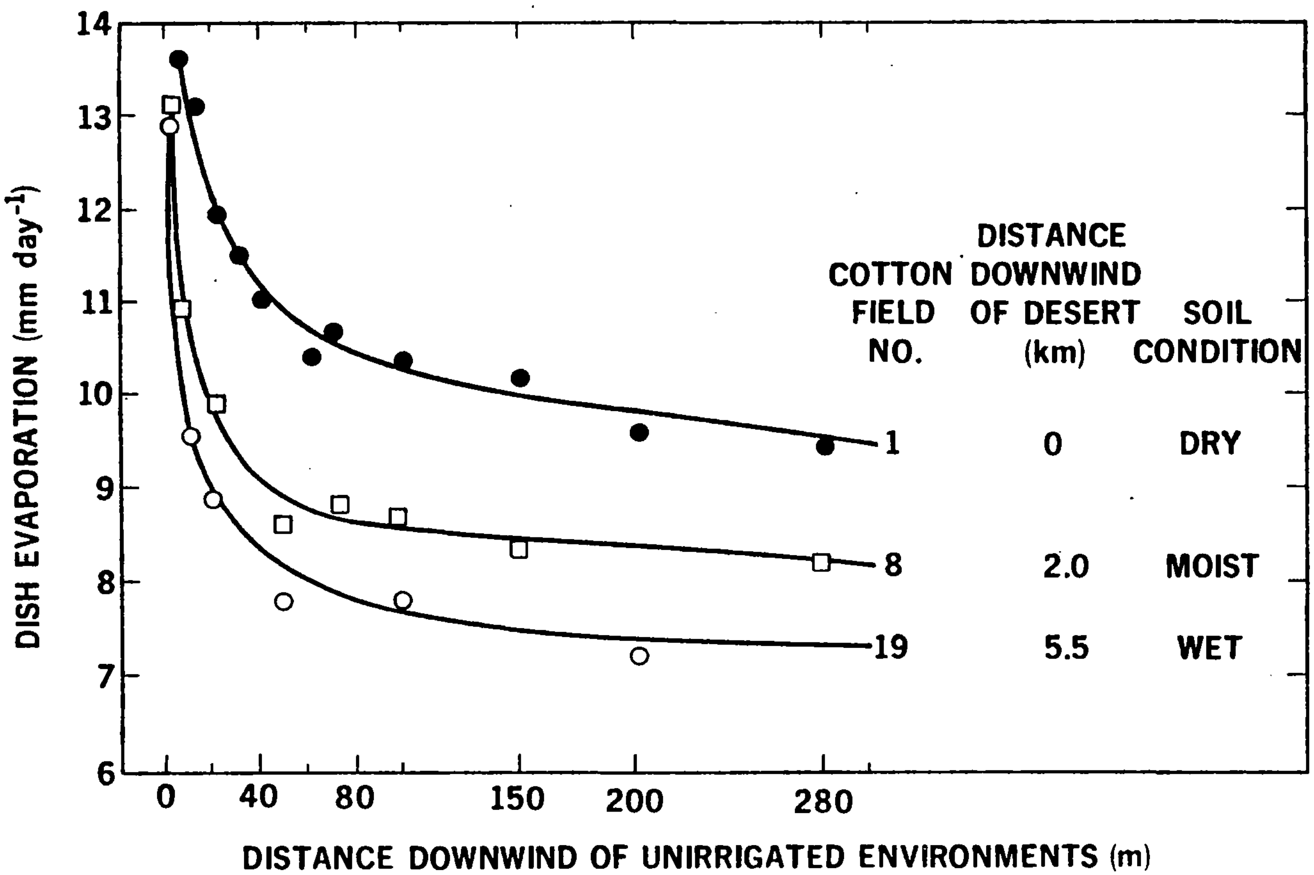
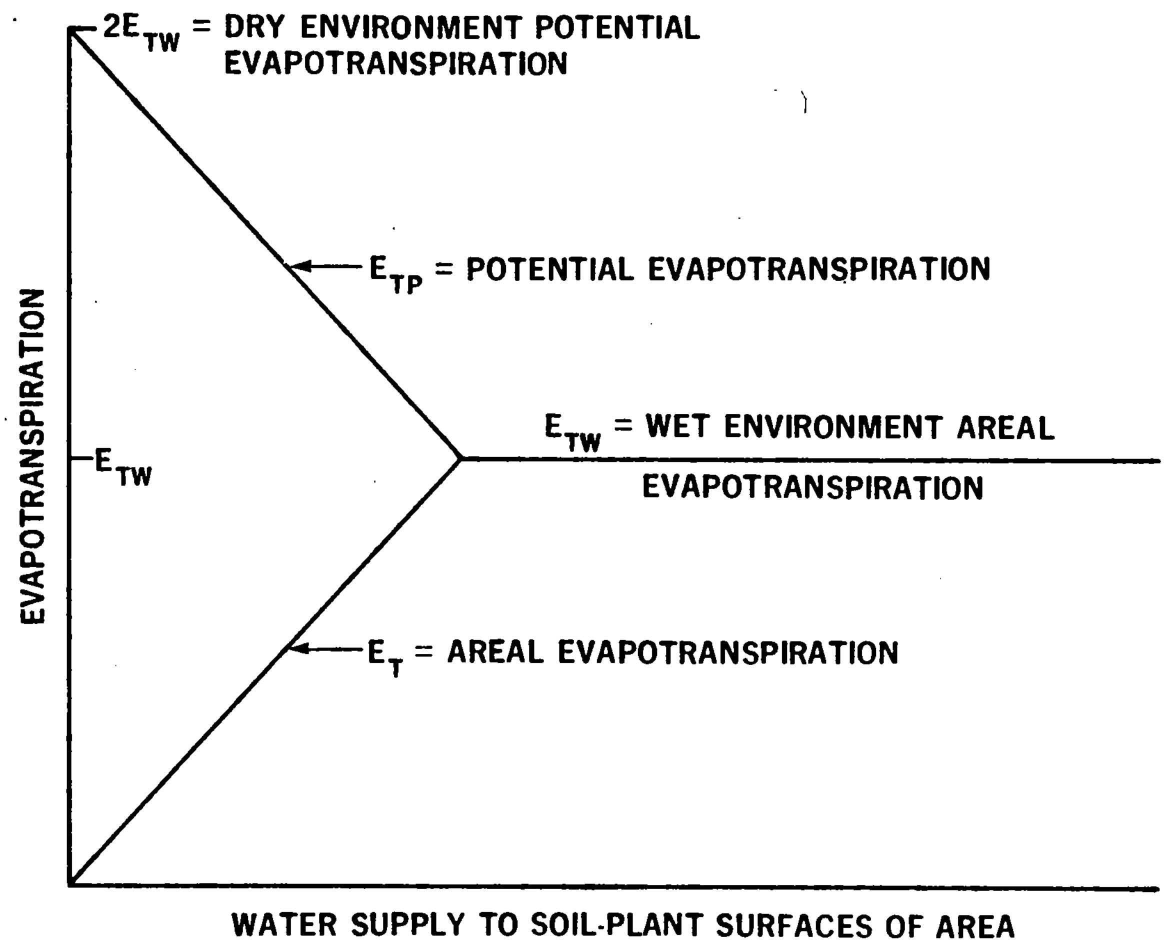
(3)

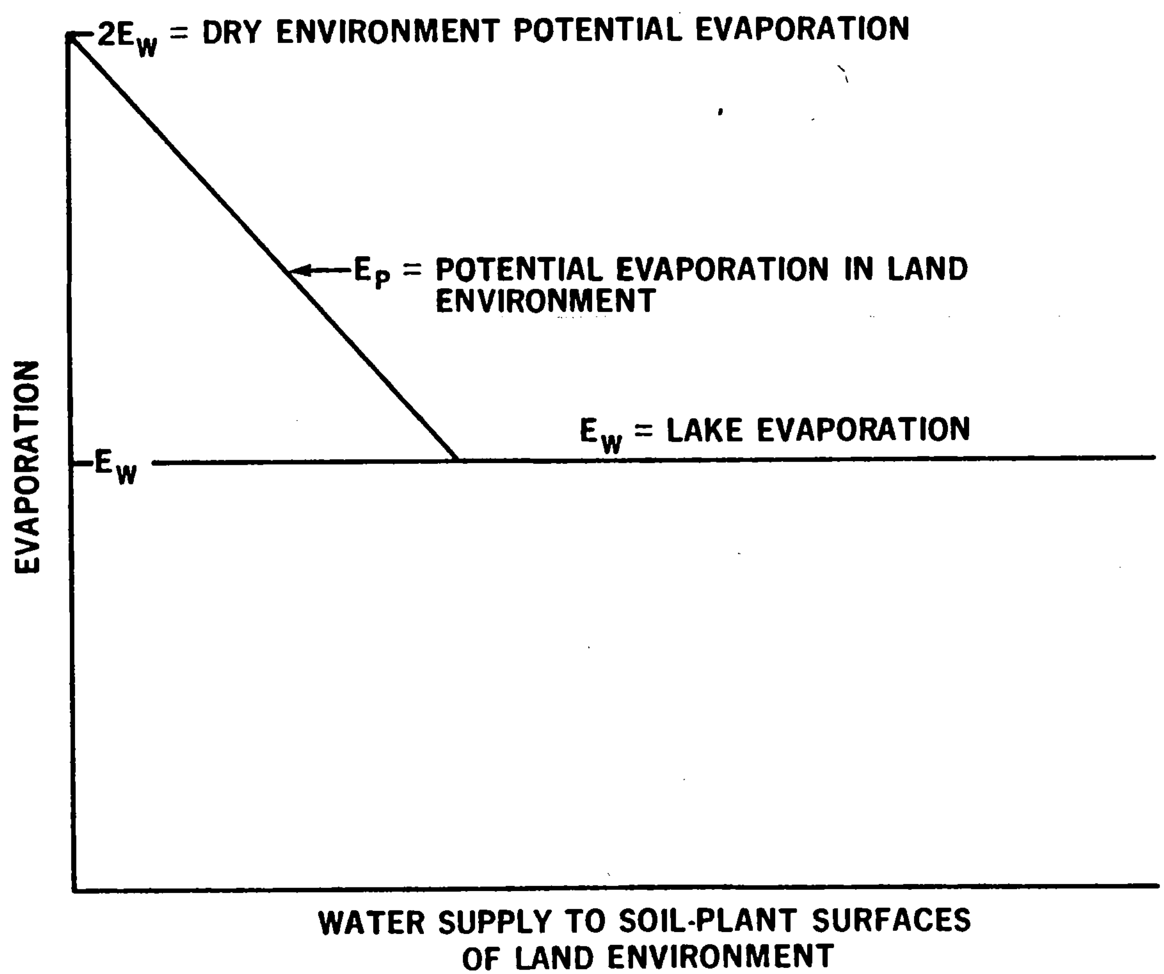


DISH EVAPORATION (mm day^{-1})
14
13
12
11
10
9
8
7
6
0
40
80
150
200
280
DISTANCE DOWNWIND OF UNIRRIGATED ENVIRONMENTS (m)
COTTON FIELD NO.
DISTANCE DOWNWIND OF DESERT (km)
SOIL CONDITION
1
0
DRY
8
2.0
MOIST
19
5.5
WET

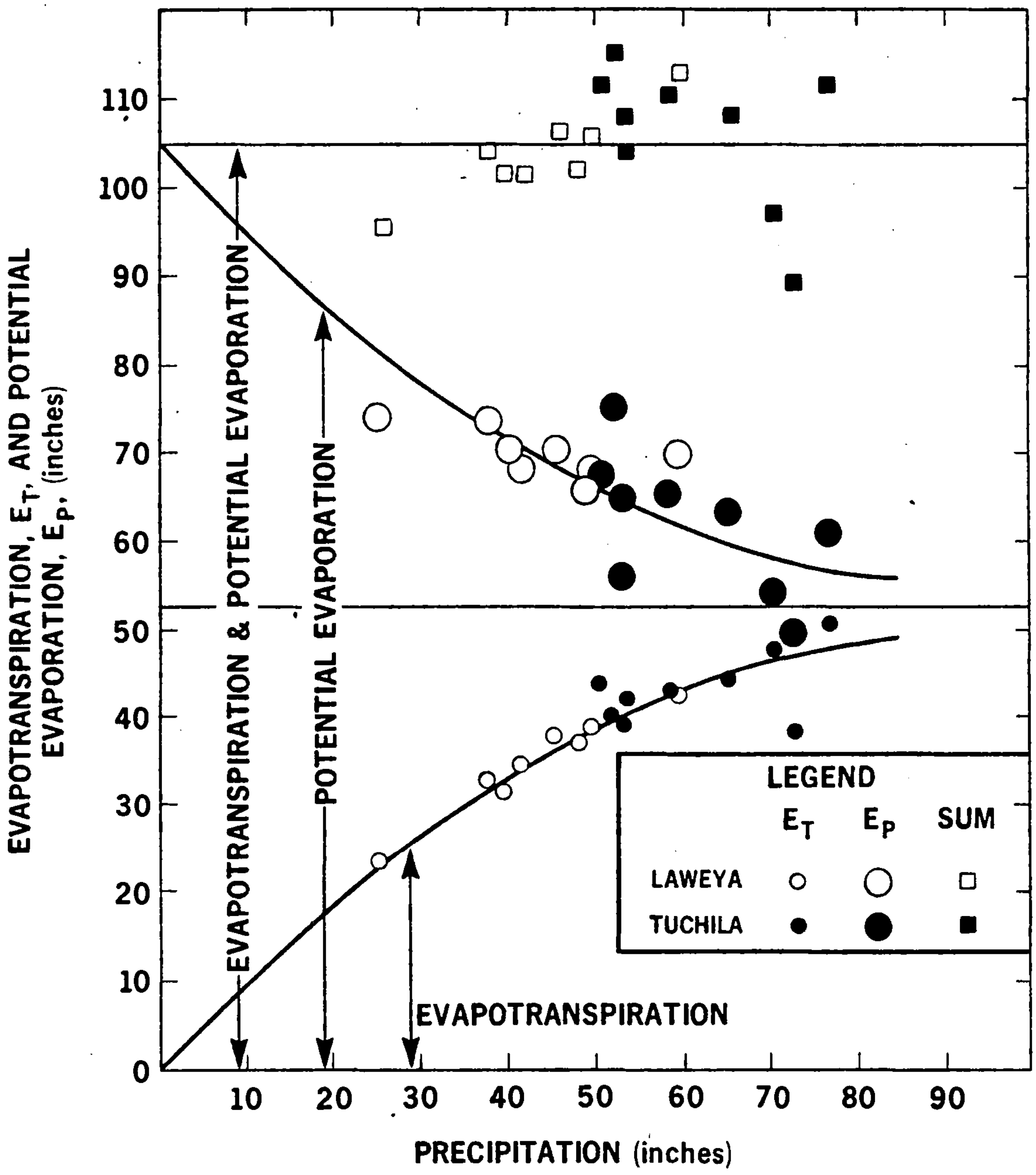
4



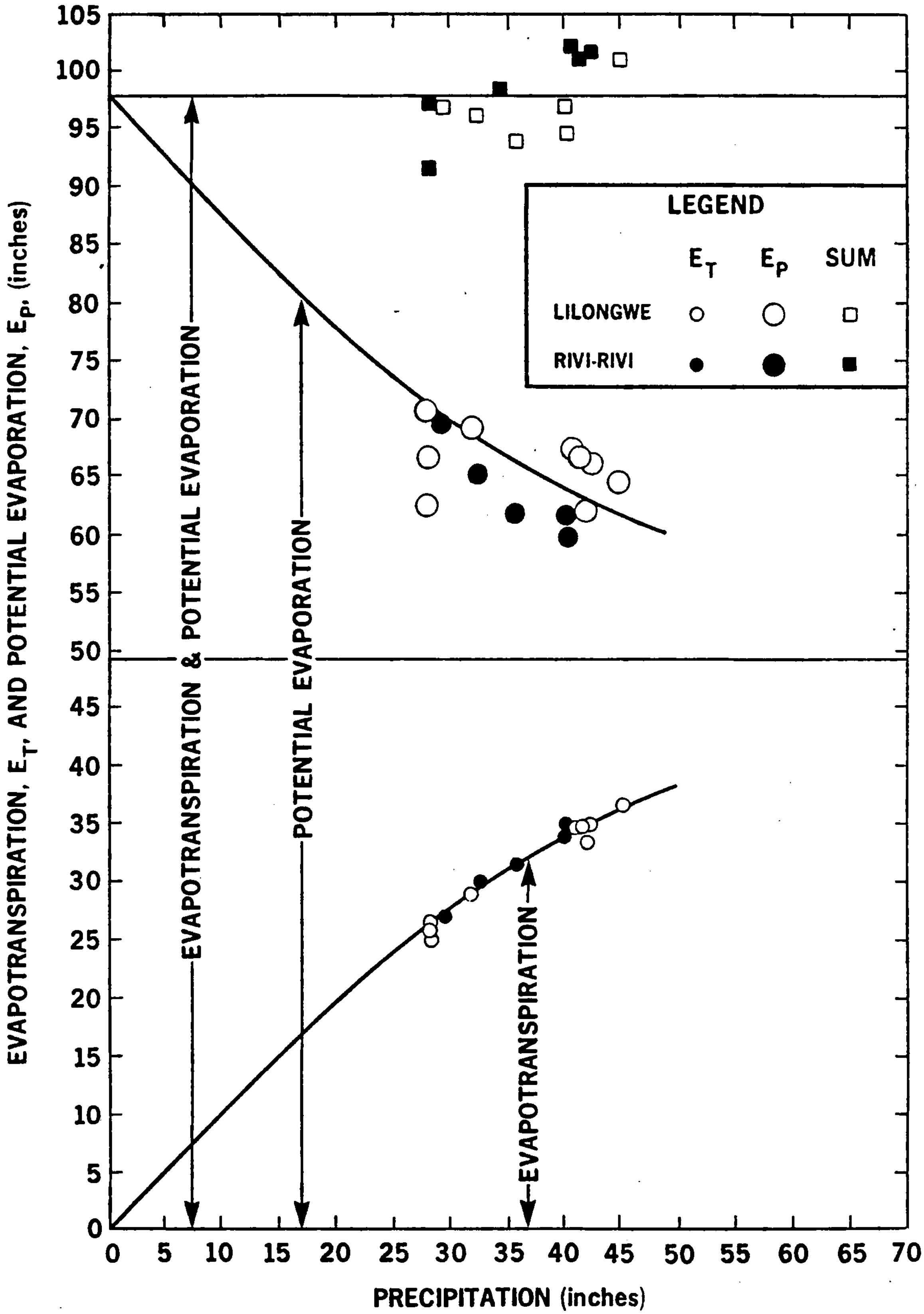
2E_TW = DRY ENVIRONMENT POTENTIAL EVAPOTRANSPIRATION
E_TP = POTENTIAL EVAPOTRANSPIRATION
E_TW = WET ENVIRONMENT AREAL EVAPOTRANSPIRATION
E_T = AREAL EVAPOTRANSPIRATION
E_TW
EVAPOTRANSPIRATION
WATER SUPPLY TO SOIL-PLANT SURFACES OF AREA


2E$_W$ = DRY ENVIRONMENT POTENTIAL EVAPORATION
E$_P$ = POTENTIAL EVAPORATION IN LAND ENVIRONMENT
E$_W$ = LAKE EVAPORATION
E$_W$
EVAPORATION
WATER SUPPLY TO SOIL-PLANT SURFACES OF LAND ENVIRONMENT

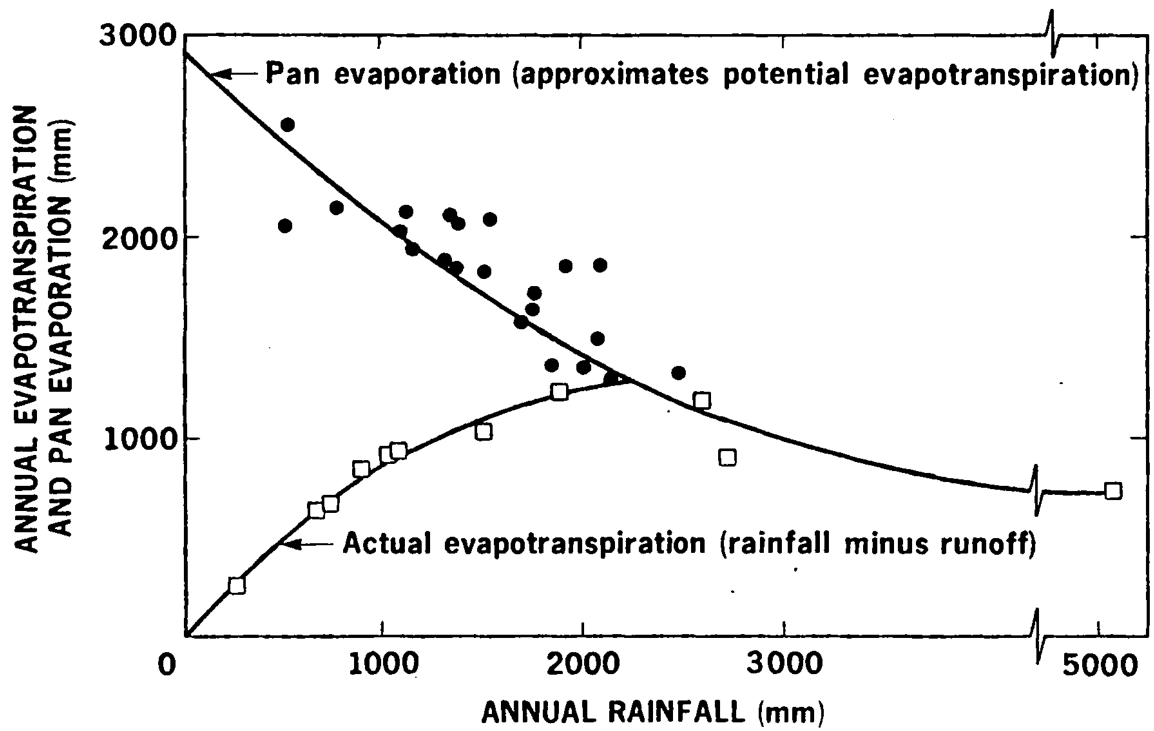
5

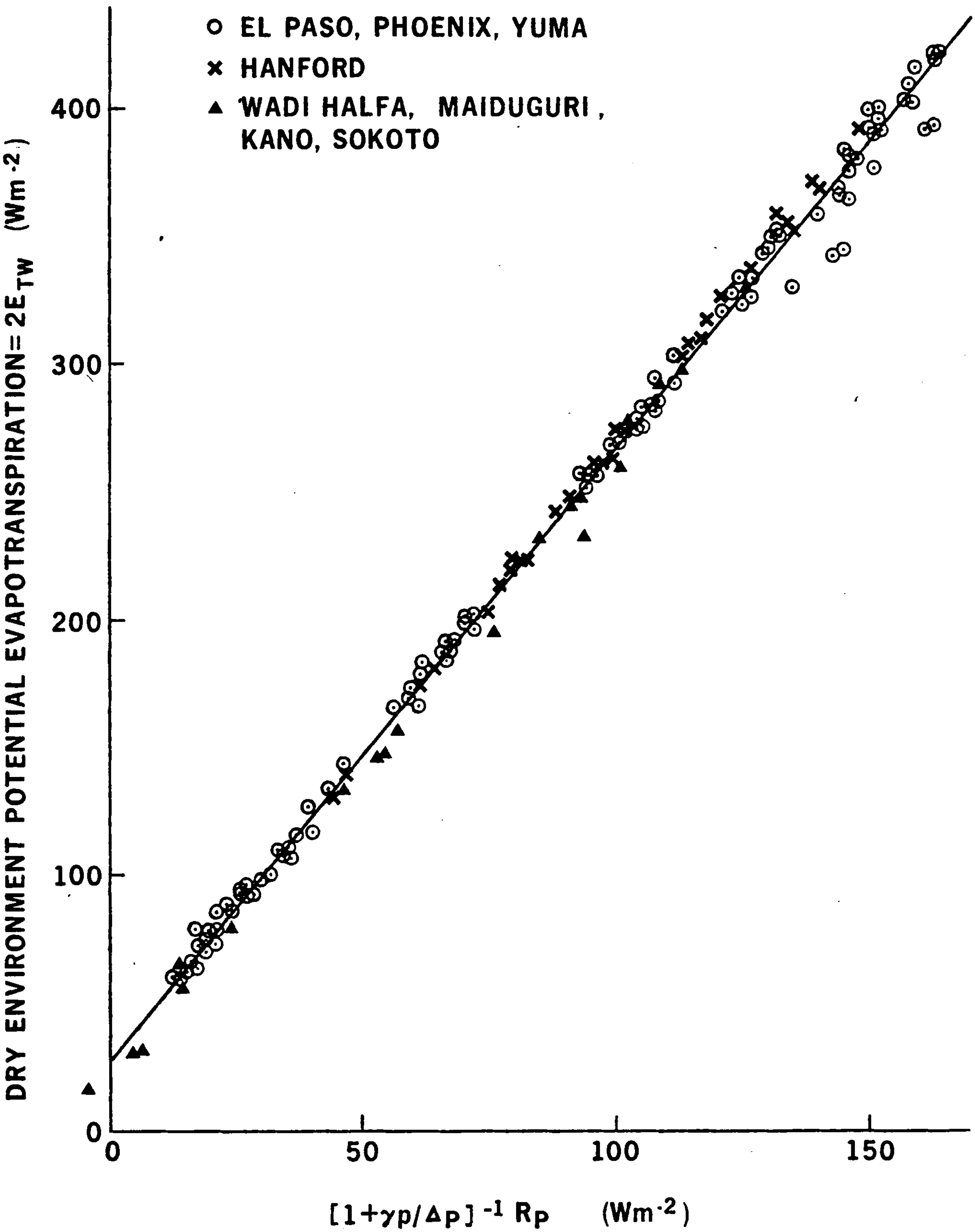

EVAPOTRANSPIRATION, E_T, AND POTENTIAL EVAPORATION, E_P, (inches)
PRECIPITATION (inches)
EVAPOTRANSPIRATION & POTENTIAL EVAPORATION
POTENTIAL EVAPORATION
EVAPOTRANSPIRATION
LEGEND
E_T
E_P
SUM
LAWEYA
TUCHILA
0
10
20
30
40
50
60
70
80
90
100
110

b

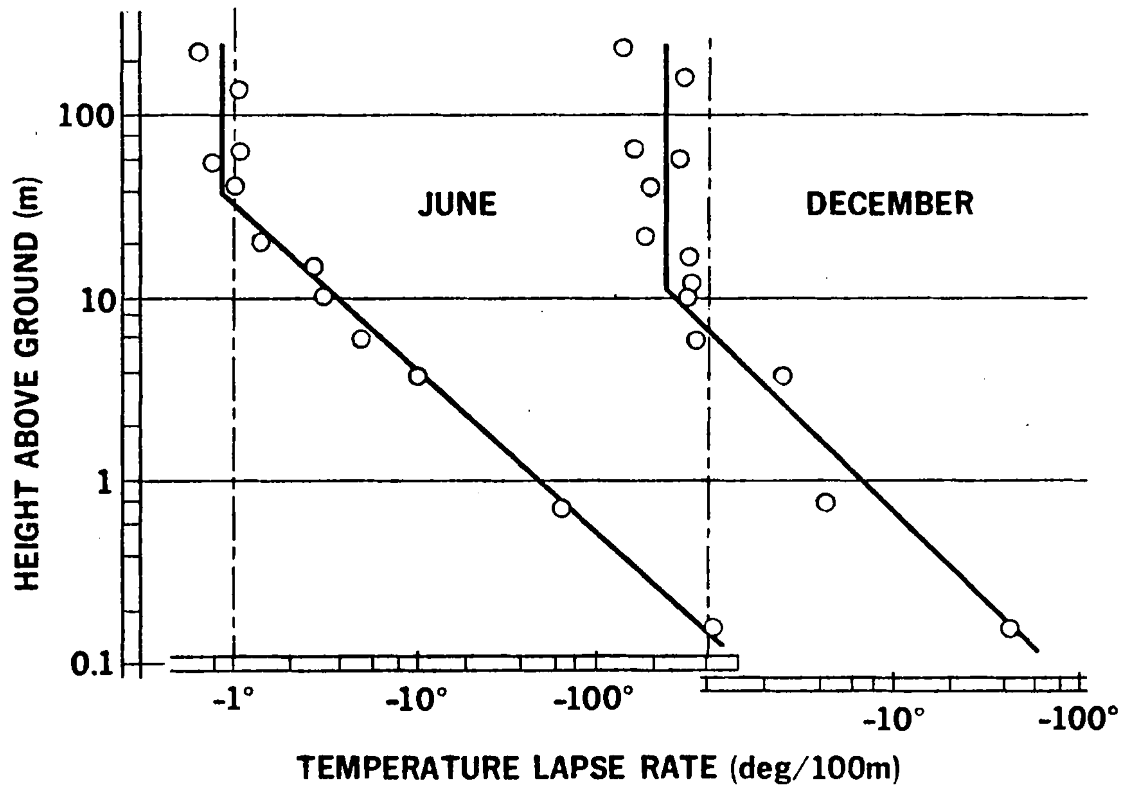


LEGEND
E_T
E_P
SUM
LILONGWE
RIVI-RIVI
EVAPOTRANSPIRATION & POTENTIAL EVAPORATION
POTENTIAL EVAPORATION
EVAPOTRANSPIRATION
EVAPOTRANSPIRATION, E_T, AND POTENTIAL EVAPORATION, E_P, (inches)
PRECIPITATION (inches)


Pan evaporation (approximates potential evapotranspiration)
Actual evapotranspiration (rainfall minus runoff)
ANNUAL EVAPOTRANSPIRATION AND PAN EVAPORATION (mm)
ANNUAL RAINFALL (mm)
3000
2000
1000
0
1000
2000
3000
5000

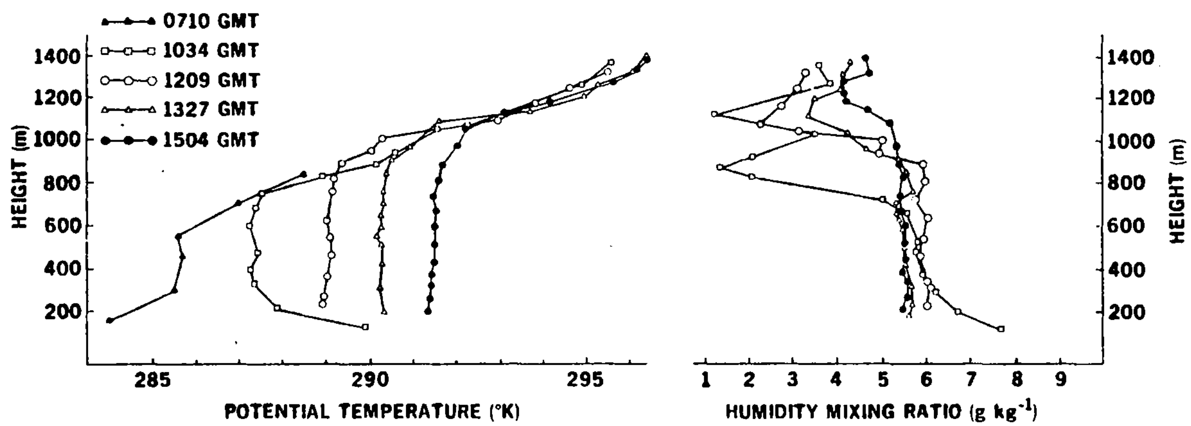

EL PASO, PHOENIX, YUMA
HANFORD
WADI HALFA, MAIDUGURI, KANO, SOKOTO
DRY ENVIRONMENT POTENTIAL EVAPOTRANSPIRATION= $2E_{TW}$ (Wm^{-2})
0
100
200
300
400
0
50
100
150
$[1+\gamma p/\Delta p]^{-1} R_P$ (Wm^{-2})

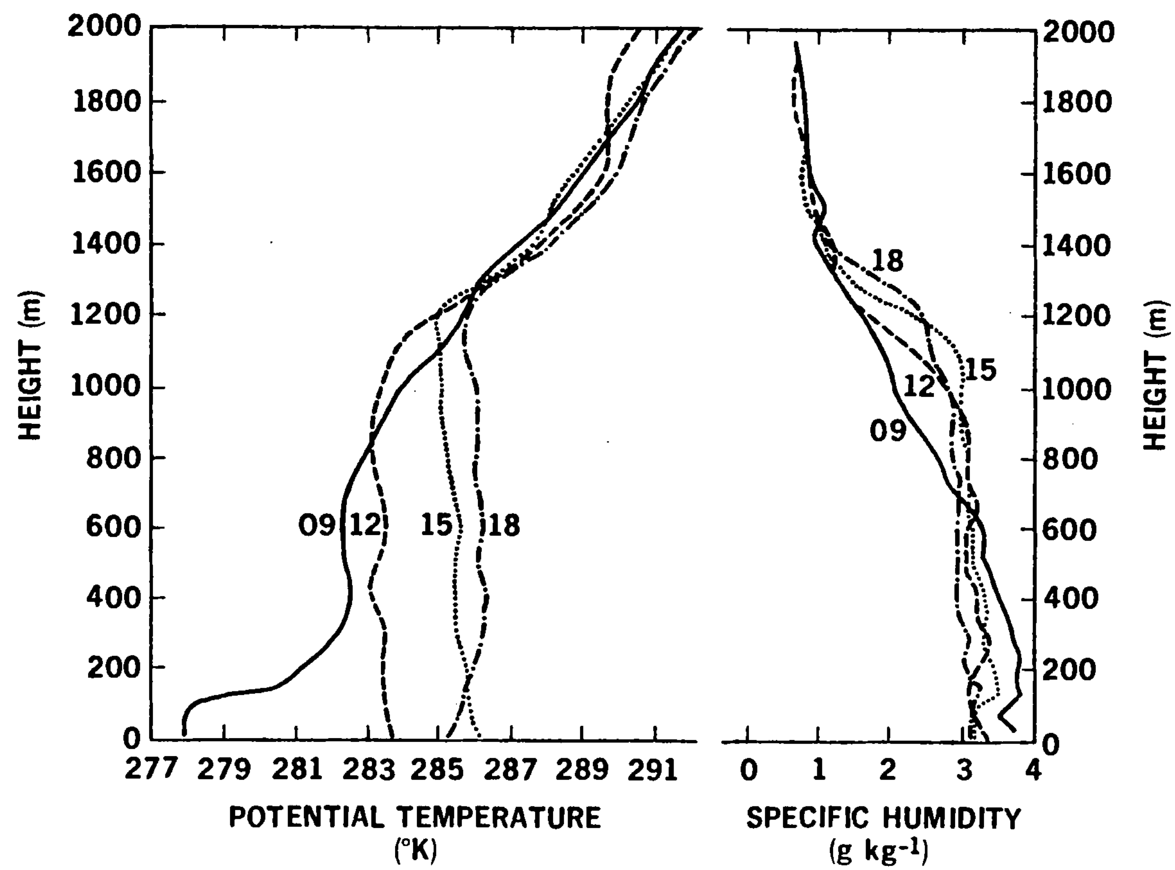
9


JUNE
DECEMBER
HEIGHT ABOVE GROUND (m)
100
10
1
0.1
-1°
-10°
-100°
-10°
-100°
TEMPERATURE LAPSE RATE (deg/100m)

10

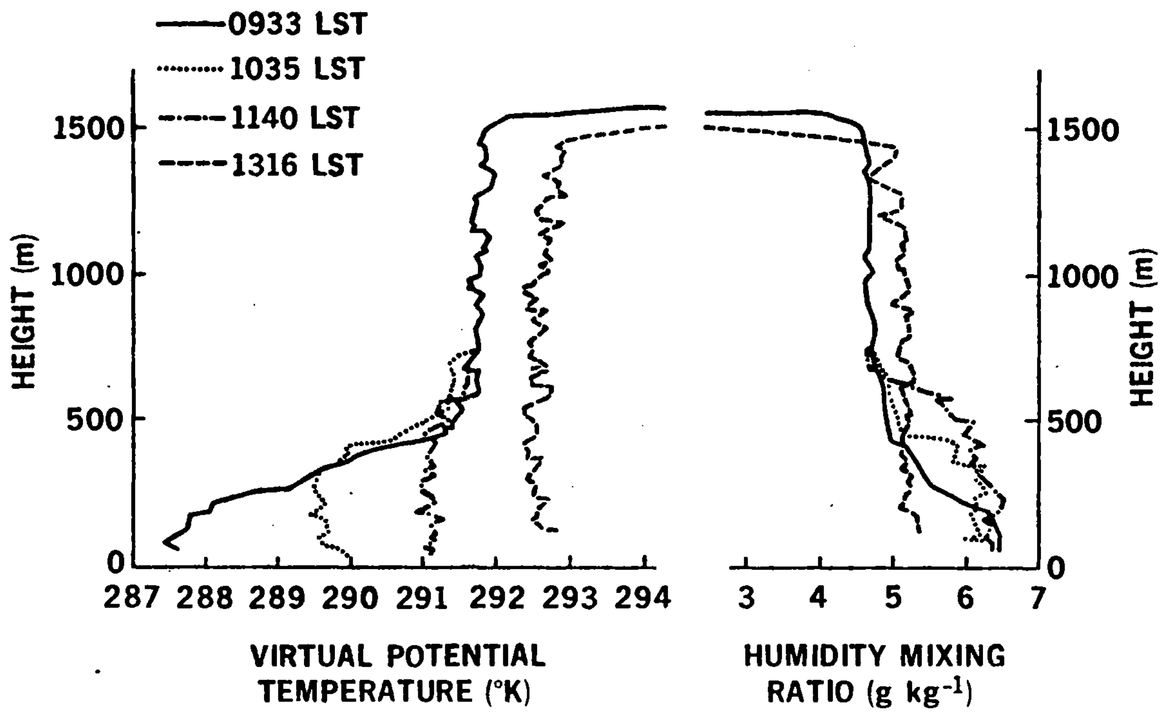
11


0710 GMT
1034 GMT
1209 GMT
1327 GMT
1504 GMT
HEIGHT (m)
1400
1200
1000
800
600
400
200
285
290
295
POTENTIAL TEMPERATURE (°K)
1 2 3 4 5 6 7 8 9
HUMIDITY MIXING RATIO (g kg-1)
HEIGHT (m)


HEIGHT (m)
2000
1800
1600
1400
1200
1000
800
600
400
200
0
277 279 281 283 285 287 289 291
POTENTIAL TEMPERATURE
(°K)
09
12
15
18
0 1 2 3 4
SPECIFIC HUMIDITY
(g kg-1)
HEIGHT (m)

12

13



0933 LST
1035 LST
1140 LST
1316 LST
HEIGHT (m)
0
500
1000
1500
287 288 289 290 291 292 293 294
VIRTUAL POTENTIAL TEMPERATURE (°K)
3 4 5 6 7
HUMIDITY MIXING RATIO (g kg-1)

OPERATIONAL ESTIMATES OF AREAL EVAPOTRANSPIRATION AND THEIR SIGNIFICANCE TO THE SCIENCE AND PRACTICE OF HYDROLOGY

III MODUS OPERANDI

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ABSTRACT

Morton, F.I., 1982c. Operational estimates of areal evapotranspiration and their significance to the science and practice of hydrology: III Modus operandi.

The complementary relationship permits the areal evapotranspiration, a product of complex processes and interactions in the soil-plant-atmosphere contiuum, to be estimated from its effects on the routine climatological observations needed to compute the potential evapotranspiration. The latest version of the complementary relationship areal evapotranspiration (CRAE) models is formulated in detail. The required station characteristics are the latitude, the altitude and a rough estimate of the long-term average annual precipitation. The climatological inputs are monthly values of dew point temperature, air temperature and sunshine duration. The minor changes needed for humidity, temperature and insolation input options and for shorter time period options are outlined. The resultant modus operandi appears quite complicated but most of the complexity is required to estimate the radiation components and to provide a generality that permits application during any season of the year in any part of the world. The minor changes required to convert the latest version of the CRAE models to a complementary relationship lake evaporation (CRLE) model, the testing of the CRAE and CRLE model estimates against comparable water budget values and the potential use of such estimates to expand the horizons of hydrology and the range of problems that can be addressed, studied and solved are presented in subsequent companion papers.

INTRODUCTION

Reliable estimates of areal evapotranspiration are essential to significant improvements in the science and practice of hydrology. Direct measurements, such as those provided by lysimeters, the eddy flux technique or the Bowen-ratio technique, give point values, require constant attendance by skilled personnel and are based on unverified assumptions. The water budget method provides good estimates for several years or more but require excessive instrumentation and manpower for shorter term estimates. A companion paper (Morton, 1982a) presents a critical review indicating that the current conceptual modeling techniques, such as that used in the SACREMENTO WATERSHED MODELING SYSTEM, are based on assumptions about the soil, the vegetation and the atmosphere that are incompatible with published evidence; and that causal modelling techniques that take into account the complex processes and interactions in the soil-plant-atmosphere continuum are not expected to have practical applications in the next three decades. The review also indicates that models based on the complementary relationship between potential and actual areal evapotranspiration can do much to fill the gap until such time as the causal models become practicable.

The complementary relationship is represented by:

$$E_T + E_{TP} = 2\ E_{TW} \tag{1}$$

or by:

$$E_T = 2E_{TW} - E_{TP} \tag{2}$$

in which E_T is the areal evapotranspiration, the actual evapotranspiration from an area so large that the effects of upwind boundary transitions are negligible; E_{TP} is the potential evapotranspiration, as estimated from a solution of the vapour transfer and energy balance equations, representing the evapotranspiration that would occur from a hypothetical moist surface with radiation absorption and vapour transfer characteristics similar to those of the area and so small that the effects of the evapotranspiration on the overpassing air would be negligible; and E_{TW} is the wet environment areal evapotranspiration, the evapotranspiration that would occur if the soil-plant surfaces of the area were saturated and there were no limitations on the availability of water.

Fig. 1 provides a schematic representation of eq. (1) under conditions of constant radiant energy supply. The ordinate represents evapotranspiration and the abscissa represents the water supply to the soil-plant surfaces of the area, a quantity that is usually unknown. When there is no water available for areal evapotranspiration (extreme left of Fig. 1) it follows that $E_T = 0$, that the air is very hot and dry and that E_{TP} is at its maximum rate of $2E_{TW}$ (the dry environment potential evapotranspiration). As the water supply to the soil-plant surfaces of the area increases (moving to the right in Fig. 1) the resultant equivalent increase in E_T causes the overpassing air to become cooler and more humid which in turn results in an equivalent decrease in E_{TP}. Finally, when the supply of water to the soil-plant surfaces of the area has increased sufficiently, the values of E_T and E_{TP} converge to that of E_{TW}.

The conventional definition for potential evapotranspiration is the same as the definition for the wet environment areal evapotranspiration. However the potential evapotranspiration that is estimated from a solution of the vapour transfer and energy balance equations by analytical (Penman, 1948), graphical (Ferguson, 1952) or iterative (Morton, 1982b) techniques, has reactions to changes in the water supply to the soil-plant surfaces that are similar to those shown for E_{TP} in Fig. 1 so that what is being estimated can exceed what is being defined by as much as 100 percent. By taking into account such reactions the complementary relationship is analogous to the Bernouilli equation for open channel flow in which the potential energy responds in a complementary way to changes in kinetic energy.

The preceding companion paper (Morton, 1982b) demonstrates that the complementary relationship is a working hypothesis with good conceptual and empirical foundations when compared with those of other hydrological concepts; presents a quickly converging iterative solution of the vapour transfer and energy balance equations for estimating potential evapotranspiration; and describes the calibration of equations for estimating the vapour transfer coefficient and the wet environment areal evapotranspiration using monthly climatological observations from stations in arid environments. With these developments, eq. (2) becomes the basis for a model in which the the left-hand side, a product of complex processes and interactions in the soil-plant-atmosphere continuum, is estimated from its effect on the routine climatological observations needed to estimate the right-hand side.

The modus operandi for the most recent version of the complementary relationship areal evapotranspiration (CRAE) models is presented in detail in the next section. It appears quite complicated but most of the complexity is required to estimate radiation components and to provide a generality that permits application during any season of the year in any part of the world. As in all modelling exercises the basic concept is associated with other assumptions and poorly defined empiricisms. However the model is falsifiable, with no need for locally optimized coefficients, so that errors in the associated assumptions and empiricisms have been detected and corrected by progressive testing over an ever-widening range of environments as demonstrated in the Appendix to the first companion paper (Morton, 1982a). This methodology requires that a correction made to obtain agreement between model and river basin water budget estimates in one environment must be applicable without modification in all other environments. Although most of the associated empiricisms were selected from the literature or were developed independently from published data it was necessary to use a certain amount of judgement in the selection process and in making minor adjustments to obtain greater generality (e.g. in using atmospheric pressure to estimate atmospheric radiation). As these empiricisms were used in the once-only calibration of the vapour transfer coefficient and the wet environment areal evapotranspiration, the effects of the selection process and the minor adjustments are implicit in the model.

The procedures set out in the modus operandi are for the most commonly used inputs and outputs. The required station characteristics are the latitude in degrees, the altitude in metres and a rough estimate of average precipitation in millimetres per year. The required climatological inputs are monthly values of dewpoint temperature, air temperature and the ratio of observed to maximum possible sunshine duration with the temperatures in degrees Celsius. The outputs are monthly values of the net radiation that would occur if the surface were at air temperature, the potential evapotranspiration and areal evapotranspiration, all in millimetres of evaporation or, in the case of net radiation, of evaporation equivalent. The minor changes that are needed for humidity, temperature and isolation input options and for shorter time period options are discussed in detail thereafter. The procedure used to test the model and details of its potential applications are presented in the subsequent companion paper (Morton, 1982d).

It may be noted that the information presented herein is of interest only to those who wish to apply the complementary relationship models so that those whose main interest is in the theoretical and practical significance of the complementary relationship can skip to the next companion paper (Morton, 1982d) with no loss of continuity.

The mathematical symbols are defined where first used and are summarized in the Appendix.

MODUS OPERANDI

For each station:

(1) Assemble input: ϕ = latitude in degrees (negative in southern hemisphere); H = altitude above sea level in metres; P_A = average annual precipitation in millimetres.

(2) Compute the ratio of atmospheric pressure at the station to that at sea level (p/p_s) with the pressure correction equation for the standard atmosphere:

$$p/p_s = [(288 - 0.0065\ H)/288]^{5.256} \tag{3}$$

(3) Estimate the zenith value of the dry-season, snow-free, clear-sky albedo (a_{zd}) from

$$a_{zd} = 0.26 - 0.00012 P_A (p/p_s)^{0.5}\ [1 + |\phi/42| + (\phi/42)^2] \tag{4}$$

$$0.11 \leq a_{zd} \leq 0.17 \tag{4a}$$

in which the absolute value $|\phi/42|$ is needed for application in the southern hemisphere. Eq. (4) attempts to take into account the effects of changes in vegetative cover with aridity. The term in brackets roughly reflects the reciprocal of latitudinal variations in evaporating power. Constraints (4a) and all other constraints used herein function as "IF" statements so that if the value of a_{zd} produced by eq. (4) is greater than 0.17 it is set equal to 0.17 and if the value of a_{zd} produced by eq. (4) is less than 0.11 it is set equal to 0.11. Over the greater part of the earth surface a_{zd} will be 0.11, whereas in deserts it will be 0.17. The nature of the transition can be seen most clearly in western Texas where the estimated values of a_{zd} at El Paso (P_A = 168 mm year^{-1}), Midland (P_A = 350 mm year^{-1}), San Angelo (P_A = 437 mm year^{-1}) and Abilene (P_A = 637 mm year^{-1}) are 0.170, 0.166, 0.143 and 0.110 respectively. In the transitions, the estimates of a_{zd} are sufficiently sensitive to changes in average annual precipitation (P_A) to justify an attempt to obtain a regional value by averaging the station value with any other long-term records in a 20 km radius. Note that the minimum constraint of 0.11 is redundant at this stage of the computations because it is applied later.

For each month:

(1) Assemble input: T = average of maximum and minimum air temperatures in degrees Celsius; T_D = average-dew point temperature in degrees Celsius; S = ratio of observed to maximum possible sunshine duration; i = month number beginning with 1 in January and ending with 12 in December; and n = number of days in the month.

(2) Compute v_D, the saturation vapour pressure at T_D in mbar; v, the saturation vapour pressure at T in mbar; and Δ, the slope of the saturation vapour pressure curve at T in mbar $^\circ C^{-1}$:

$$v_D = 6.11 \exp\,[17.27\,T_D/(T_D + 237.3)] \tag{5}$$

$$v = 6.11 \exp\,[\alpha T/(T + \beta)] \tag{6}$$

$$\Delta = dv/dT = \alpha\beta v/(T + \beta)^2 \tag{7}$$

in which α and β are 17.27 and 237.3°C, respectively, when $T \geq 0°C$, or 21.88 and 265.5°C, respectively, when $T < 0°C$.

(3) Compute various angles and functions leading up to an estimate of the extra-atmospheric global radiation (G_E):

$$\theta = 23.2 \sin(29.5i - 94) \tag{8}$$

$$\cos Z = \cos(\phi - \theta) \tag{9}$$

$$\cos Z \geq 0.001 \tag{9a}$$

$$\cos\omega = 1 - \cos Z/(\cos\phi \cos\theta) \tag{10}$$

$$\cos\omega \geq -1 \tag{10a}$$

$$\cos z = \cos Z + [(180/\pi)\sin\omega/\mathrm{acos}(\cos\omega) - 1]\cos\phi \cos\theta \tag{11}$$

$$\eta = 1 + (1/60)\sin(29.5i - 106) \tag{12}$$

$$G_E = (1354/\eta^2)(\omega/180)\cos z \tag{13}$$

in which θ is the declination of the sun in degrees, ω is the number of degrees the earth rotates between sunrise and noon, Z and z are the noon and average angular zenith distances of the sun, respectively, and η is the radius vector of the sun. Constraint (9a) applies during the arctic and antarctic winters when the sun does not rise, and constraint (10a) applies during the arctic and antarctic summers when the sun does not set. Eqs. (8) and (12) provide monthly average values that must be considered approximate because the solstices and equinoxes do not occur at midmonth and because the month lengths are unequal. The largest absolute error for eq. (8) is 0.48° in February and the average for all months is 0.13°. Alternatives to eqs. (8) and (12) for time periods of a fraction of a month are presented in the next section.

(4) Estimate the zenith value of snow-free, clear-sky albedo (a_{zz}), the zenith value of clear-sky albedo (a_z) and the clear-sky albedo (a_0):

$$a_{zz} = a_{zd} \tag{14}$$

$$0.11 \leq a_{zz} \leq 0.5\ (0.91 - v_D/v) \tag{14a}$$

$$c_0 = v - v_D \tag{15}$$

$$0 \leq c_0 \leq 1 \tag{15a}$$

$$a_z = a_{zz} + (1 - c_0^2)(0.34 - a_{zz}) \tag{16}$$

$$a_0 = \frac{a_z[\exp(1.08)-(2.16\ \cos Z/\pi + \sin Z)\exp(0.012Z)}{1.473\ (1-\sin Z)} \tag{17}$$

Constraint (14a) represents an attempt to decrease the albedo during the wet season of a subarid region where the estimate of a_{zd} may exceed the minimum value of 0.11. Eq. (16) represents an attempt to increase the albedo during the season of snow cover when the vapour pressure deficit is usually less than a millibar. Eq. (17) was suggested by Arnfeld (1975) and the numerical constants are equal to those derived from 110 observations in a cornfield when there was less than 33 percent cloud cover.

(5) Estimate precipitable water vapour (W) in mm and a turbidity coefficient (j):

$$W = v_D/(0.49 + T/129) \tag{18}$$

$$c_1 = 21 - T \tag{19}$$

$$0 \leq c_1 \leq 5 \tag{19a}$$

$$j = (0.5 + 2.5 \cos^2 z) \exp [c_1(p/p_s - 1)] \tag{20}$$

The derivation of eq. (18) has been presented elsewhere (Morton, 1978). Eq. (20) is based on the Angstrom formula for the turbidity coefficient as reported by Robinson (1966). However, the units are different, the average angular zenith distance (z) has been added to provide some sort of seasonal variation and the factor c_1 has been added to decrease the effects of altitude when the average temperature exceeds 16°C and to eliminate them completely when the average temperature exceeds 21°C.

(6) Compute the transmittancy of clear skies to direct beam solar radiation (τ) from an equation formulated by Brooks (1960):

$$\tau = \exp [-0.089(p/p_s/\cos z)^{0.75} - 0.083 (j/\cos z)^{0.90} - 0.029(W/\cos z)^{0.60}] \tag{21}$$

(7) Estimate the part of τ that is the result of absorption (τ_a):

$$\tau_a = \exp\left[- 0.0415\ (j/\cos z)^{0.90} - (0.0029)^{0.5}\ (W/\cos z)^{0.3}\right] \quad (22)$$

$$\tau_a \geq \exp\left[- 0.0415\ (j/\cos z)^{0.90} - 0.029\ (W/\cos z)^{0.6}\right] \quad (22a)$$

Eq. (22) is believed to be an improvement to the suggestion of Brooks (1960) that τ_a should be the exponent of the right-hand term in the argument of eq. (21). The estimate for absorption by precipitable water vapour that would result from eq. (22) is almost identical to that resulting from an equation presented by Lettau and Lettau (1973). The constraint prevents the effects of absorption by water vapour from exceeding the effects of absorption and scattering.

(8) Compute the clear sky global radiation (G_o) using the equation formulated by Brooks (1960); and then estimate the incident global radiation (G):

$$G_o = G_E\tau[1 + (1 - \tau/\tau_a)(1 + a_o\tau)] \quad (23)$$

$$G = SG_o + (0.08 + 0.30S)(1-S)G_E \quad (24)$$

The right-hand term in eq. (24), which represents the global radiation that is transmitted through clouds, was derived from zero-intercepts in the linear equation between the ratio of observed to extra-atmospheric global radiation and the ratio of observed to maximum possible sunshine duration that were tabulated by Rietveld (1978). The 27 values from Yugoslavia, Sweden, Portugal and the Netherlands were identified with either the summer or the winter seasons whereas the remaining 25 were for the entire year. Fig. 2 shows the 27 seasonally identified zero-intercepts plotted against the appropriate average sunshine duration ratios. The resultant regression line, which is shown on Fig. 2 and used in eq. (24), differs slightly from the one proposed by Rietveld (1978) because he used the values for the entire year as well.

(9) Estimate the average albedo (a) from

$$a = a_0 \; [S + (1-S)(1 - Z/330)] \tag{25}$$

The quantity (1 - Z/330) in eq. (25) was derived by computing the clear sky albedo from eq. (17) for the full range of noon angular zenith distances; estimating the cloudy sky albedo for the same zenith distances from a similar equation with constants that are somewhat more extreme than those derived by Arnfeld (1975) from 91 observations in a cornfield when the cloud cover exceeded 67 percent; dividing the cloudy sky values by the clear sky values and relating the ratios to the angular zenith distance. The resultant relationship fits the computed data very closely. Note that the clear sky and cloudy sky values are the same when the noon sun is at the zenith i.e. at certain times of year within the tropics.

(10) Estimate the proportional increase in atmospheric radiation due to clouds (ρ):

$$c_2 = 10\ (v_D/v - S - 0.42) \tag{26}$$

$$0 \leq c_2 \leq 1.0 \tag{26a}$$

$$\rho = 0.18[(1-c_2)(1-S)^2 + c_2(1-S)^{0.5}]\ p_s/p \tag{27}$$

Boltz (Geiger, 1966) suggested that ρ varies as the square of the cloud cover or the square of (1-S). This suggestion is realistic under most circumstances but becomes inadequate in very cloudy and humid conditions. These considerations are taken into account in relationships (26), (26a) and (27). For example, if the relative humidity (v_D/v) is 0.80, the weighting factor, c_2, varies from zero for sunshine duration ratios greater than or equal to 0.38 to one for sunshine duration ratios less than or equal to 0.28. This means that ρ would be proportional to the square of (1-S) for values of 1-S less than 0.62; would be a weighted average of the square and square root of (1-S) for values of (1-S) between 0.62 and 0.72; and would be proportional to the square root of (1-S) for values of (1-S) exceeding 0.72. The ratio p_s/p in eq. (27) reflects the tendency for clouds to be closer to the ground at high altitudes.

(11) Calculate the net long-wave radiation loss with the surface at air temperature (B):

$$B = \varepsilon\sigma\ (T + 273)^4\ [1 - (0.71 + 0.007\ v_D\ p/p_s)(1 + \rho)] \tag{28}$$

$$B \leq 0.05\ \varepsilon\sigma\ (T + 273)^4 \tag{28a}$$

in which ε is the emissivity and σ is the Stefan-Boltzmann constant. With a land surface emissivity of 0.92, $\varepsilon\sigma$ is $5.22 \times 10^{-8}\ Wm^{-2} \cdot K^{-4}$. The ratio of clear-sky atmospheric radiation to black-body radiation, i.e. $(0.71 + 0.007\ v_D\ p/p_s)$ has been derived from 15 out of the 18 average values for Tucson, Arizona tabulated by Sellers (1965). A plot of the data and the regression equation (without the atmospheric pressure adjustment) has been presented elsewhere (Morton, 1978). Constraint (28a) comes into effect only under very hot, humid and cloudy conditions such as those that might prevail in low altitude equatorial regions during certain seasons of the year. Note that the effects of atmospheric pressure or altitude in eqs. (27) and (28) will tend to cancel each other out under cloudy conditions.

(12) Estimate the net radiation if the surface were at air temperature (R_T), the stability factor (ζ), the vapour transfer coefficient (f_T) and the heat transfer coefficient (λ):

$$R_T = (1-a)G - B \tag{29}$$

$$R_{TC} = R_T \tag{30}$$

$$R_{TC} \geq 0 \tag{30a}$$

$$1/\zeta = 0.18\ (1 + \frac{v_D}{v}) + \Delta R_{TC}/[\gamma p\ (p_s/p)^{0.5}\ b_o\ f_Z(v - v_D)] \quad (31)$$

$$1/\zeta \leq 1 \quad (31a)$$

PG 71 it is 0.28.....
This could be an error here
All of the previous calculations use 0.28

$$f_T = (p_s/p)^{0.5}\ f_Z/\zeta \quad (32)$$

$$\lambda = \gamma p + 4\varepsilon\sigma\ (T + 273)^3/f_T \quad (33)$$

This should be:
γp = (γPs)(P/Ps)
From Actual paper in journal of Hydrology

in which b_0 = 1.00 for the CRAE model, $p=(\gamma p_s)(p/p_s)$ and f_Z and γp_s are 28.0 Wm^{-2} $mbar^{-1}$ and 0.66 mbar $°C^{-1}$ respectively when $T \geq 0°C$ or 28.0 x 1.15 Wm^{-2} $mbar^{-1}$ and 0.66/1.15 mbar $°C^{-1}$ when $T <$ 0°C. The development of eqs. (31) and (32) has been discussed in the preceding companion paper (Morton, 1982b).

(13) Choose initial values of T_P, v_P and Δ_P equal to T, v and Δ and estimate the final values from the following quickly converging iterative solution of the vapour transfer and energy balance equations:

$$\delta T_P = [R_T/f_T + v_D - v_P' + \lambda(T - T_P')]/(\Delta_P' + \lambda) \quad (34)$$

$$T_P = T_P' + \delta T_P \quad (35)$$

$$v_P = 6.11\ \exp\ [\alpha T_P/(T_P + \beta)] \quad (36)$$

$$\Delta_P = \alpha\beta v_P/(T_P + \beta)^2 \quad (37)$$

Eqs. (34), (35), (36) and (37) are repeated setting T_P', v_P' and Δ_P' equal to the values of T_P, v_P and Δ_P derived from the preceding iteration until $\delta T_P \leq 0.01^\circ C$. The purpose is to estimate the potential evapotranspiration equilibrium temperature (T_P) from a solution of the vapour transfer and energy balance equations for a small moist surface. The development of the iterative procedure is described in the preceding companion paper (Morton, 1982b).

(14) Estimate the potential evapotranspiration (E_{TP}), the net radiation if the surface were at the equilibrium temperature (R_{TP}) and the wet environment areal evapotranspiration (E_{TW}):

$$E_{TP} = R_T - \lambda f_T(T_P - T) \tag{38}$$

$$R_{TP} = E_{TP} + \gamma p f_T(T_P - T) \tag{39}$$

$$E_{TW} = b_1 + b_2 \, (1 + \gamma p/\Delta_P)^{-1} \, R_{TP} \tag{40}$$

$$1/2 \; E_{TP} \leq E_{TW} \leq E_{TP} \tag{40a}$$

in which the constants b_1 and b_2 are 14 Wm^{-2} and 1.20 respectively for the CRAE model. They have been determined from the calibration procedure described in the preceeding companion paper (Morton, 1982b). Constraints (40a) prevent the areal evapotranspiration from being negative in arid environments and from exceeding the potential evapotranspiration in humid environments. Eqs. (38) and (39) are just two different forms of the energy balance equation.

(15) Estimate the areal evapotranspiration, E_T, from the complementary relationship:

$$E_T = 2E_{TW} - E_{TP} \tag{41}$$

(16) Convert the net radiation with the surface at air temperature (R_T), the potential evapotranspiration (E_{TP}) and the areal evapotranspiration (E_T) from the power units of Wm^{-2} to the evaporation units of mm of depth by dividing by the latent heat of vaporization or sublimation and multiplying by the number of days. The latent heat of vaporization (for $T \geq 0°C$) is 28.5 W-days per kilogram and the latent heat of sublimation (for $T < 0°C$) is 28.5 x 1.15 W-days per kilogram.

INPUT OPTIONS

The modus operandi set out in the preceeding section uses as input the monthly values of dew point temperature, air temperature and ratio of observed to maximum possible sunshine duration. However there are a number of options available that require only minor changes. One such option, which converts the CRAE model to a complementary relationshp lake evaporation (CRLE) model is presented in a subsequent companion paper (Morton 1982e). The other options, which are concerned primarily with input data and shorter time periods, are presented below.

Altitude Input Option

If the average atmospheric pressure (p) is known it can be divided by the average sea level value (p_s) of 1013 mb thereby rendering eq. (3) superfluous.

Humidity Input Options

The humidity input is normally the average dew point temperature and this is used to estimate the average vapour pressure from eq. (5). The relationship is nonlinear so the result is somewhat less than the average of the vapour pressures that were used to estimate the individual values of dew point. The difference can be significant when the averages include the effects of frequent weather changes. However the models have been calibrated and tested with dew point temperature inputs despite this inconsistency because; (1) dew points are published more frequently than vapour pressures; and (2) the resultant vapour pressures are compatible with the saturation vapour pressures estimated from eq. (6) using average air temperatures. Therefore the use of average vapour pressure inputs requires a relatively small correction factor that may be estimated from:

$$\delta v_D = 0.71\ v_0^{0.25}\ [(\delta v_1)\ v_2/v_1]^{0.25}\ (\delta v_2)^{0.50} \qquad (42)$$

in which $(\delta v_1)\ v_2/v_1 \geq 0.5\ (\delta v_2)$ and $(\delta v_1)\ v_2/v_1 \leq 1.5\ (\delta v_2)$.

In eq. (42) δv_D is the correction to be subtracted from the average vapour pressure input, v_0 is the saturation vapour pressure at 0°C (6.11 mbar), v_1 is the saturation vapour pressure at the average maximum air temperature, v_2 is the saturation vapour pressure at the average minimum air temperature, δv_1 is the difference between the average of the saturation vapour pressures at the maximum air temperatures and v_1, and δv_2 is the difference between the average of the saturation vapour pressures at the minimum air temperatures and v_2. The equation is dimensionally consistent.

The data used in the formulation of eq. (42) were derived from the Monthly Records of Meteorological Observations in Canada for the months of April, May, July, August, November and December at 17 stations spread over nine of the ten provinces. They included monthly averages of vapour pressure, dew point temperature, maximum air temperature and minimum air temperature together with daily values of the latter two quantities. Fig. 3 shows the monthly values of δv_D that were derived from eq. (42) plotted against the differences between the recorded monthly vapour pressure and the vapour pressure computed from the recorded monthly dew point temperature. The standard error of estimate for the 92 plotted points is 0.108 mbar.

The physical basis for eq. (42) is the dependence of the dew point temperature on the minimum temperature and the dependence of the minimum temperature on the dew point temperature in humid and subhumid climates. However the dew point temperatures in arid and subarid climates are much lower than the minimum temperatures and because they are relatively independent they have much less variability. Eight station-months of data were omitted from the development of eq. (42) and from Fig. 3 because of this and these were the only station-months for which the observed average vapour pressures were less than 50 percent of the saturation vapour pressures at the average of the maximum and minimum temperatures. The average δv_D for the eight station-months was 0.18 mbar. Although there is not enough data to provide a firm basis for a further constraint the foregoing considerations suggest that the following would provide reasonable results:

> "When the observed average vapour pressure is less than 50 percent of the saturation vapour pressure at the average of the maximum and minimum temperatures, the vapour pressure correction factor (δv_D) is the value estimated from eq. (42) or 0.2 mbar whichever is the lesser."

With this further constraint the observed average vapour pressure less the vapour pressure correction can be used as input and this would render eq. (5) superfluous.

The modus operandi can easily accommodate average relative humidity inputs. All that is needed is to replace eq. (5) by the product of the saturation vapour pressure at air temperature that is computed from eq. (6) and the relative humidity expressed as a ratio. However nonlinearity in the relationship between saturation vapour pressure and temperature ensures that the resultant value of atmospheric vapour pressure (v_D) will be significantly overestimated. The greater part of the error is due to the inverse relationship between hourly values of saturation vapour pressure and hourly values of relative humidity that is the result of large hour-to-hour variations in temperature and small hour-to-hour variations in atmospheric vapour pressures over the period of a day and this is augmented by the effects of the significant day-to-day variations in atmospheric vapour pressures that have been discussed in the preceding paragraphs. If relative humidities must be used they should be applied to the saturation vapour pressures for the temperatures that occur at the same time and the resultant atmospheric vapour pressures should be converted to dew point temperatures. Even if this were done only once a day the average for the period of a week or more would provide more satisfactory results than the use of the average values of relative humidity and temperature. However it is recommended that the observations and computations be more frequent - once every six hours seems to be a good compromise value.

Temperature Input Options

Equations for converting air and dew point temperatures from Fahrenheit (or other units) to Celsius units can be included in the model before implementing eq. (5).

Insolation Input Options

If global radiation observations are available they can be used as input instead of the ratio of observed to maximum possible sunshine duration. This is done by using the observed global radiation (in Wm^{-2}) to replace the results of eq. (24) and to provide the estimates of the sunshine duration ratio that are required in further computations. The conversion to Wm^{-2} requires that cal cm^{-2} day^{-1} (langleys per day) be divided by 2.064 and that MJ m^{-2} day^{-1} be divided by 0.0864. The estimates of the sunshine duration ratio are estimated from

$$S = 0.53\ G/(G_o - 0.47\ G) \tag{43}$$

$$0 \leq S \leq 1.0 \tag{43a}$$

The formulation of eq. (43) is based on zero-intercepts and slopes for the 27 linear equations between the ratio of incident to extra-atmospheric insolation and the ratio of observed to maximum possible sunshine duration that were tabulated by Reitveld (1978) and identified with either the summer or winter seasons. The same data were used to formulate eq. (24) and prepare Fig. 2. Fig. 4 shows the zero intercepts plotted against the average value of the ratio of incident to extra-atmospheric global radiation as computed from the sum of the zero intercept and the product of the slope and the average sunshine duration ratio. The slope of the line shown in Fig. 4 is 0.47 and eq. (43) has been derived from a solution of

$$G = G_E \ (1-S) \ 0.47 \ G/G_E + SG_o \tag{44}$$

If net radiation observations are available they can simplify the model significantly. However they may introduce error because they depend on surface temperature whereas the net radiation used in the model (R_T) is the value that would occur if the surface were at air temperature. The error could be significant in arid or subarid climates. The difference can be taken into account with a rough estimate of the surface temperature by using the slope of the Stefan-Boltzmann equation, $4\epsilon\sigma \ (T + 273)^3$. The slope has values of 6.0, 5.0 and 4.0 $Wm^{-2} \ °C^{-1}$ at temperatures of 33°C, 15°C and -5°C respectively. When the net radiation observations have been converted to units of Wm^{-2} and adjusted for the difference between surface and air temperatures they can be used to estimate R_T and thereby eliminate the need for other insolation estimates, the station latitude input, the station long-term average annual precipitation input; and expressions (4), (4a) and (8) to (29) inclusive.

Shorter Time Period Options

The CRAE models cannot be used to provide daily estimates because of subsurface heat storage changes and because of the lag times associated with the change in storage of heat and water vapour in the atmospheric boundary layer after changes in surface conditions or the passage of frontal systems. There is every probability that the time periods could be shortened to five days without problems but for intervals of three days or less the results would always be suspect.

It is convenient to retain the monthly structure when estimating areal evapotranspiration from past records and this can be done quite simply. Let m equal the number of time periods in each month such that the first (m-1) periods have the same number of days and the last period has the number of days required to complete the month. To avoid large absolute and percentage variations in the lengths of the last periods, m should be restricted to 2, 3, 5 and 6. If I is the period number, with I = 1 for the first period in January and I = 12 m for the last period in December, the fractional month number, i, to be used in eqs. (8) and (12) is

$$i = [I + 0.5\,(m-1)]/m \tag{45}$$

and the constant 23.2° in eq. (8) is changed to 23.4°.

For real-time estimates of areal evapotranspiration it may be convenient to use weekly periods or some other time period that does not fit into the monthly structure. All that is needed is the procedure set out above with I equal to the number of days from the beginning of the calendar year to the middle day of the period (using a February of 28.5 days) and with m equal to 29.5 + I/270 or 30.4, whichever is smaller. As before, the constant 23.2° in eq. (8) is replaced by 23.4°.

Daily values of areal evapotranspiration may be needed in some potential applications, e.g. in real time forecasting. The errors resulting from the use of The CRAE models with daily inputs should not be hydologically significant unless accumulated for many days. Therefore the daily model estimates can be used on a provisional basis for a week but they should then be corrected proportionally to make the total agree with the model estimates derived from the average weekly inputs.

CONCLUDING DISCUSSION

Presented herein is a detailed documentation on how to use the latest version of the CRAE models to estimate areal evapotranspiration from monthly values of air temperature, dewpoint temperature and the ratio of observed to maximum possible sunshine duration and on how to make the adaptations needed for other temperature, humidity, insolation or shorter time period options. Documentation on how to make the conversion to a CRLE model that can be used to estimate lake evaporation from the same climatological inputs observed in the land environment is presented in a companion paper (Morton, 1982e). The models have also been documented in FORTRAN and in RPN for the Hewlett-Packard HP-67 hand-held calculator (Morton, Goard and Piwowar, 1980). A later RPN program for the Hewlett-Packard HP-41C hand-held calculator is documented and can be made available on request. The greater storage capacity of the HP-41C eliminates the need for using extra program cards during individual computations and permits the full range of options.

The model presented herein has the following limitations:

(1) It requires accurate humidity data and these have depended on frequent observations by skilled personnel. This is one of the more serious limitations to the use of the CRAE models at the present time. However the Humicell, a simple device developed by the Saskatchewan Research Council, now provides a convenient and reliable alternative. According to Langham (1969) the instrument provides integrated vapour pressures within $\pm$ 2 percent for periods exceeding three days. Other more convenient instrumentation is under development.

(2) It cannot be used for short time intervals because of subsurface heat-storage changes and because of the lag times associated with the change in storage of heat and water vapour in the atmospheric boundary layer after changes in surface conditions or the passage of frontal systems. There is every probability that the time periods could be shortened to five days but for intervals of three days or less the results would always be suspect. This limitation has little significance in hydrological applications because it does not

matter much whether the daily evapotranspiration is 3 mm or 6 mm as long as the accumulated values for the week or for a longer period are reliable. Short time period options are discussed more fully in the preceding section.

(3) It cannot be used near sharp environmental discontinuities, such as a high-latitude coastline or the edge of an oasis, because of advection of heat and water vapour in the lower layers of the atmosphere. The discussion of the data of Davenport and Hudson (1967) in some of the companion papers (Morton 1982a, 1982b and 1982d) indicates that the effects of such advections can decrease to near zero within 300 m but this finding may not be generally applicable.

(4) It requires input from a climatological station whose surroundings are representative of the area of interest. Some advantages that are associated with this and the preceding limitation are presented in a companion paper (Morton, 1982d).

(5) It cannot be used to predict the effects of natural or man-made changes since it neither uses nor requires knowledge of the soil-vegetation system. This is the most serious long-term limitation to the application of the CRAE models.

It is emphasized that the model presented herein is completely calibrated. Thus it is able to provide estimates of areal evapotranspiration anywhere in the world with no changes. The only local characteristics that are needed are the latitude, the altitude and a rough estimate of long-term average annual precipitation. These capabilities are tested in the next companion paper (Morton, 1982d) using water budget estimates of areal evapotranspiration for 143 river basins in North America, Africa, Australia, Ireland and New Zealand. The results are shown in Fig. 5.

As noted previously the CRAE model can be converted to a CRLE model with a few minor changes. In estimating lake evaporation from routine climatological observations in the land environment, it has the same capabilities, and these are tested in a companion paper using water budget estimates for eleven lakes in North America and Africa. The results for ten of these lakes are shown in Fig. 6.

Probably the greatest constraint to the use of the CRAE models is the failure of hydrologists and engineers to understand how the science and practice of hydrology have been stunted by the lack of operational estimates of areal evapotranspiration and to visualize how the availability of such estimates can suddenly expand the horizons of hydrology and the range of problems that can be addressed, studied and solved. For this reason a large part of the next companion paper (Morton, 1972d) is devoted to examples which show the potential impact of operational estimates of areal evapotranspiration on hydrological research and on water planning and management.

APPENDIX

DEFINITION OF SYMBOLS

a = albedo

a_o = clear-sky albedo

a_z = zenith value of clear-sky albedo

a_{zz} = zenith value of clear-sky, snow-free albedo

a_{zd} = zenith value of dry-season, snow-free, clear-sky albedo

b_0, b_1 and b_2 are constants

B = net long-wave radiation loss with the surface at air temperature (Wm^{-2})

c_0, c_1 and c_2 are constrained variables used in longer equations

E_T = areal evapotranspiration (Wm^{-2})

E_{TP} = potential evapotranspiration (Wm^{-2})

E_{TW} = wet environment areal evapotranspiration (WM^{-2})

f_T = vapour transfer coefficient (Wm^{-2} $mbar^{-1}$)

f_Z = constant used in estimating f_T - changes at below-freezing temperatures (Wm^{-2} $mbar^{-1}$)

G = incident global radiation (Wm^{-2})

G_E = extra-atmospheric global radiation (Wm^{-2})

G_o = clear-sky global radiation (Wm^{-2})

H = altitude (m)

i = month number

I = period number

j	=	turbidity coefficient
m	=	number of periods in month
n	=	number of days in period
p	=	atmospheric pressure (mbar)
p_s	=	atmospheric pressure at sea level (mbar)
P_A	=	long-term average annual precipitation (mm year^{-1})
R_T	=	net radiation with the soil-plant surfaces at air temperature (Wm^{-2})
R_{TC}	=	R_T with $R_{TC} \geq 0$ (Wm^{-2})
R_{TP}	=	net radiation with the surface at the potential evapotranspiration equilibrium temperature (Wm^{-2})
S	=	ratio of observed to maximum possible sunshine duration
T	=	average air temperature or average of maximum and minimum values (°C)
T_D	=	average dew point temperature (°C)
T_P	=	potential evapotranspiration equilibrium temperature (°C)
T_P'	=	trial value of T_P in iteration process (°C)
δT_P	=	correction to T_P' in iteration process (°C)
v	=	saturation vapour pressure at T (mbar)
v_D	=	saturation vapour pressure at T_D (mbar)
δv_D	=	difference between average of the saturation vapour pressures at the dew point temperatures and v_D (mbar)
v_P	=	saturation vapour pressure at T_P (mbar)
v_P'	=	trial value of v_P in iteration process (mbar)
v_0	=	saturation vapour pressure at 0°C = 6.11 mbar

v_1 = saturation vapour pressure at the average maximum air temperature (mbar)

δv_1 = difference between the average of the saturation vapour pressures at the maximum air temperatures and v_1 (mbar)

v_2 = saturation vapour pressure at the average minimum air temperature (mbar)

δv_2 = difference between the average of the saturation vapour pressures at the minimum air temperatures and v_2 (mbar)

W = precipitable water vapour (mm)

z = average angular zenith distance of sun (degrees)

Z = noon angular zenith distance of sun (degrees)

α = constant used in estimating vapour pressures - changes at below-freezing temperatures

β = constant used in estimating vapour pressures - changes at below-freezing temperatures (°C)

γ = psychrometric constant - changes at below-freezing temperatures (°C^{-1})

Δ = slope of saturation vapour pressure curve at T (mbar °C^{-1})

Δ_P = slope of saturation vapour pressure curve at T_P (mbar °C^{-1})

Δ_P' = slope of saturation vapour pressure curve at T_P' (mbar °C^{-1})

ϵ = surface emissivity

ζ = stability factor

η = radius vector of sun

θ = declination of sun (degrees)

λ = $\gamma p + 4\epsilon\sigma(T+273)^3/f_T$ (mbar °C^{-1})

π = 3.141592654

ρ = proportional increase in atmospheric radiation due to clouds

σ = Stefan–Boltzmann constant

τ = transmittancy of clear skies to direct beam solar radiation

τ_a = the part of τ that is the result of absorption

ϕ = latitude (degrees)

ω = angle the earth rotates between sunrise and noon (degrees)

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Morton, F.I., 1982b. Operational estimates of evaporation and their significance to the cience and practice of hydrology: II The complementary relationship, This issue.

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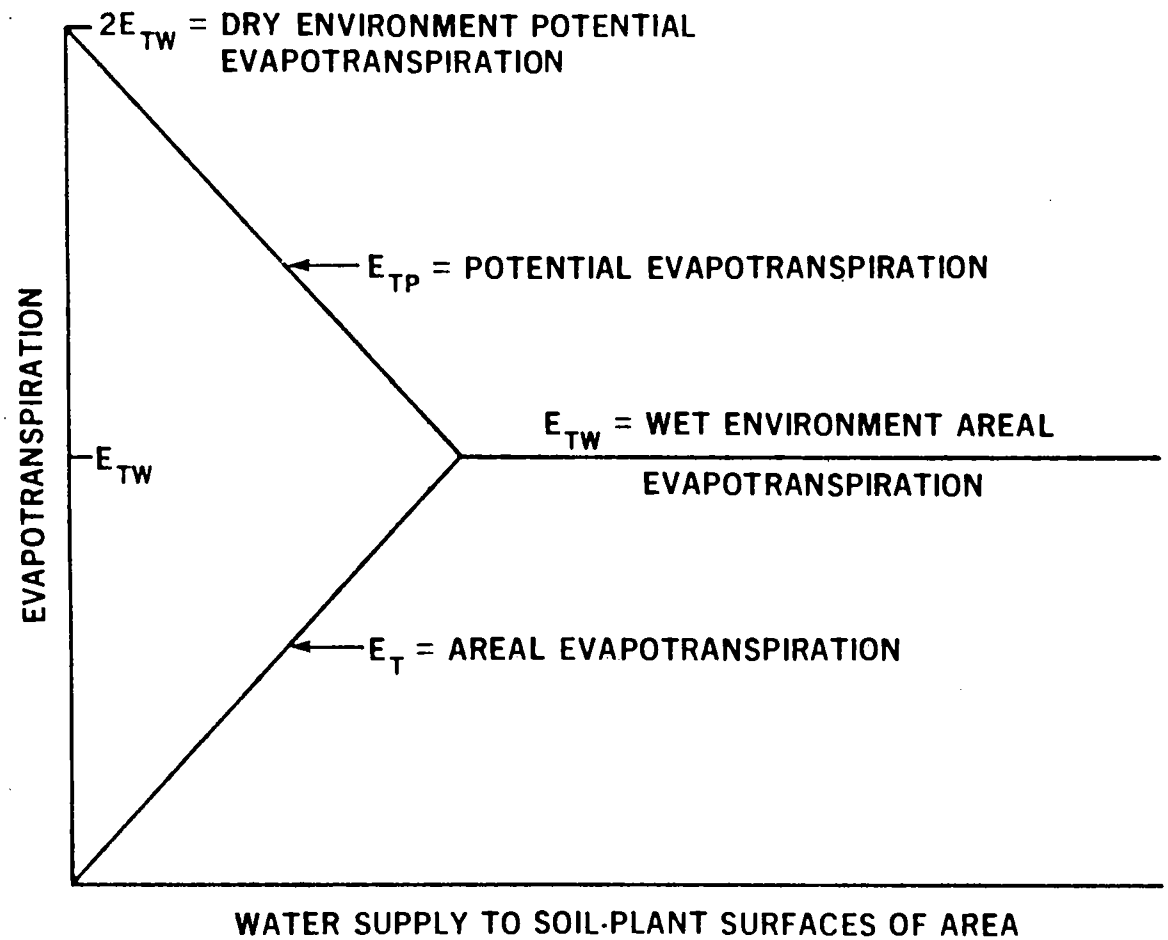
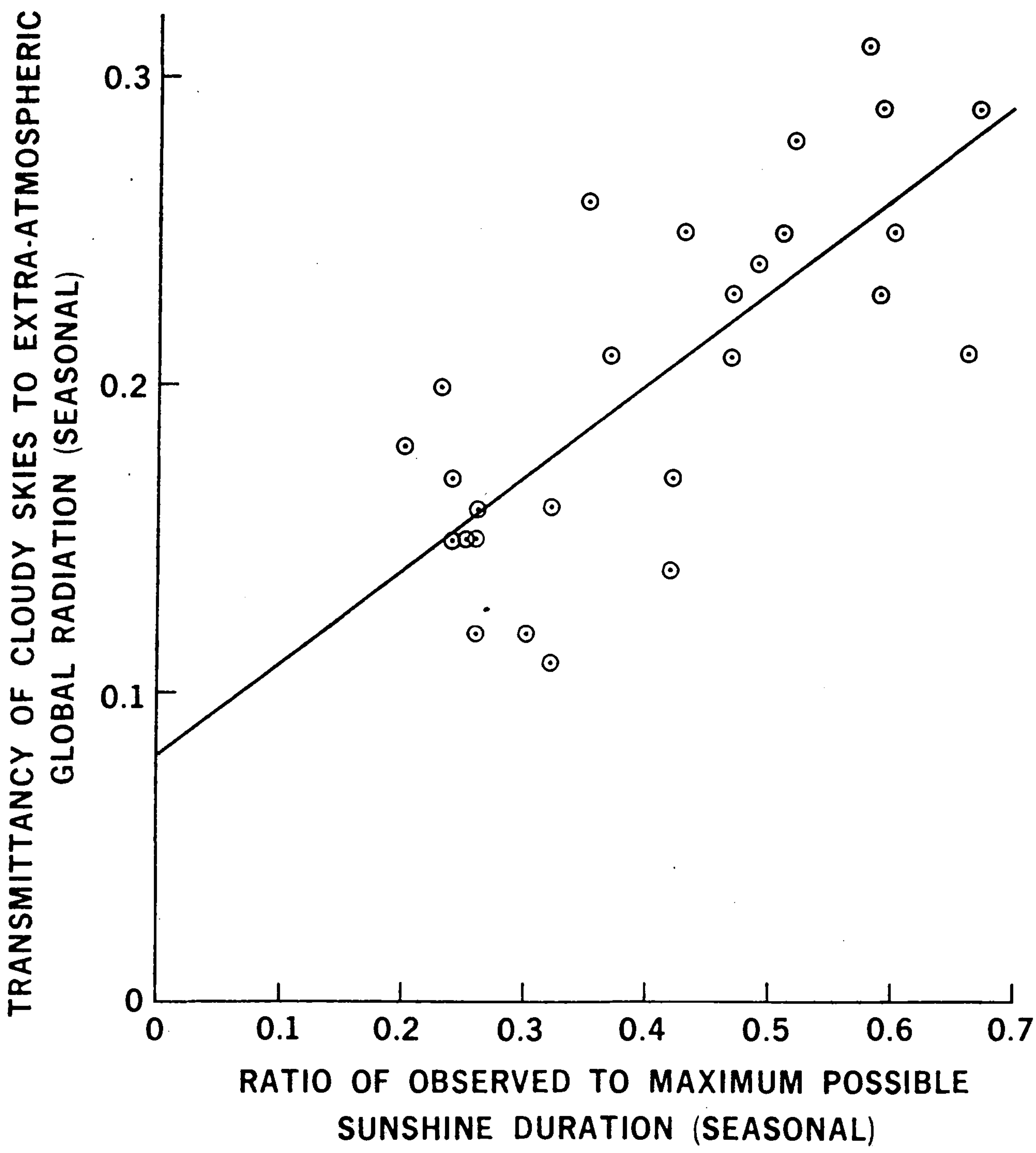
TITLES FOR FIGURES



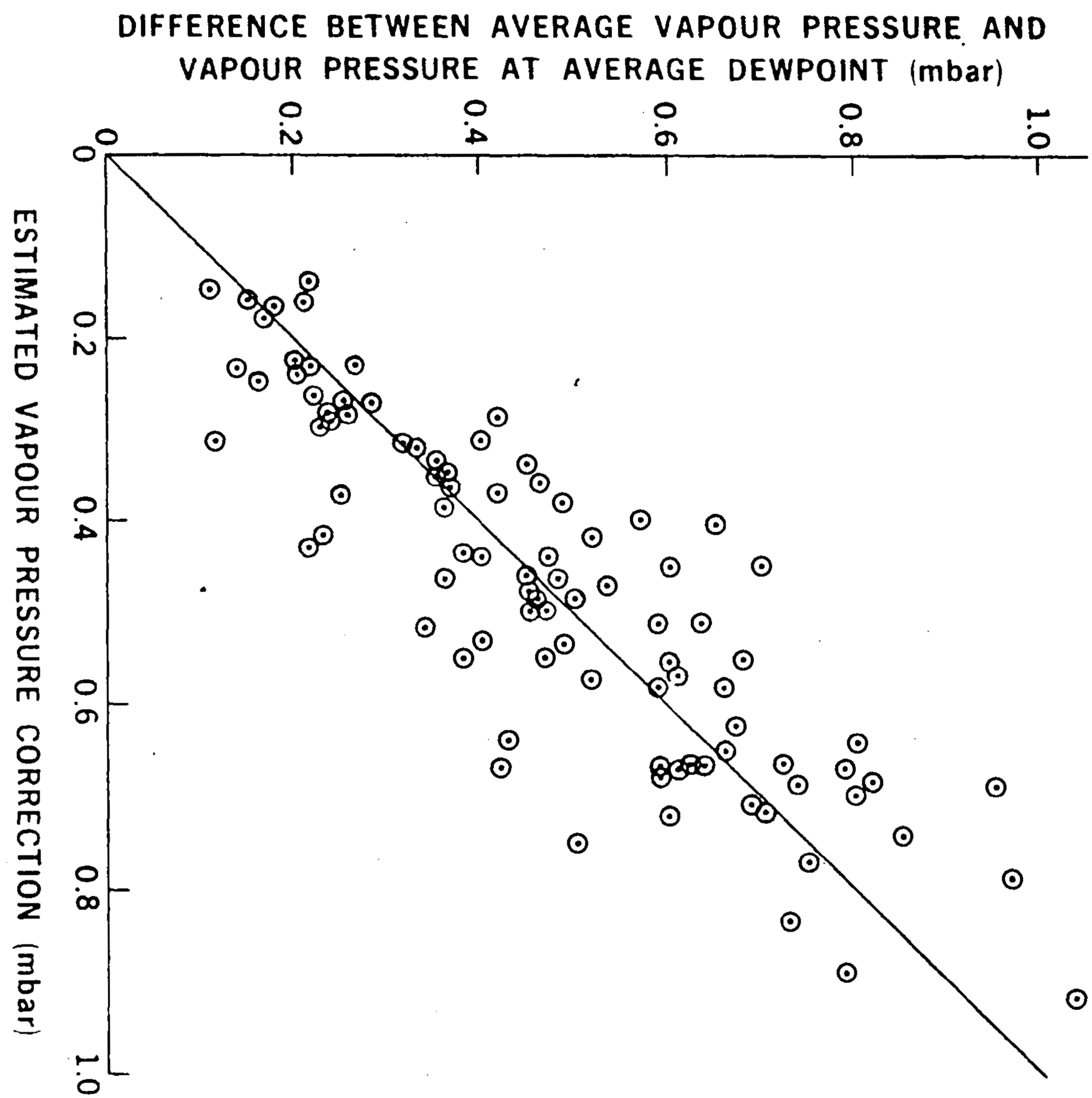
$2E_{TW}$ = DRY ENVIRONMENT POTENTIAL EVAPOTRANSPIRATION
E_{TP} = POTENTIAL EVAPOTRANSPIRATION
E_{TW} = WET ENVIRONMENT AREAL EVAPOTRANSPIRATION
E_{T} = AREAL EVAPOTRANSPIRATION
E_{TW}
EVAPOTRANSPIRATION
WATER SUPPLY TO SOIL-PLANT SURFACES OF AREA

1

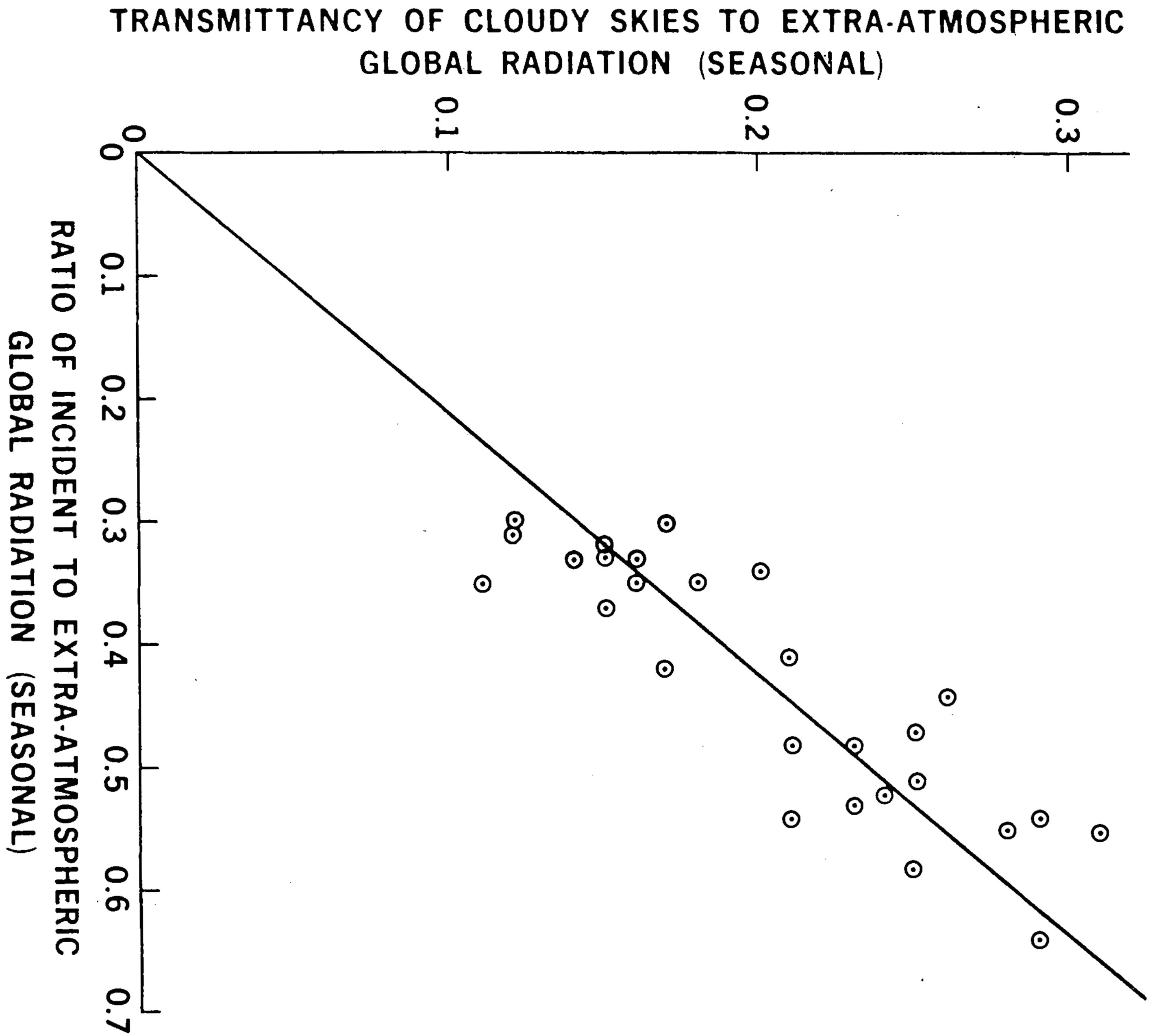

TRANSMITTANCY OF CLOUDY SKIES TO EXTRA-ATMOSPHERIC GLOBAL RADIATION (SEASONAL)
0
0.1
0.2
0.3
0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
RATIO OF OBSERVED TO MAXIMUM POSSIBLE SUNSHINE DURATION (SEASONAL)

2

3

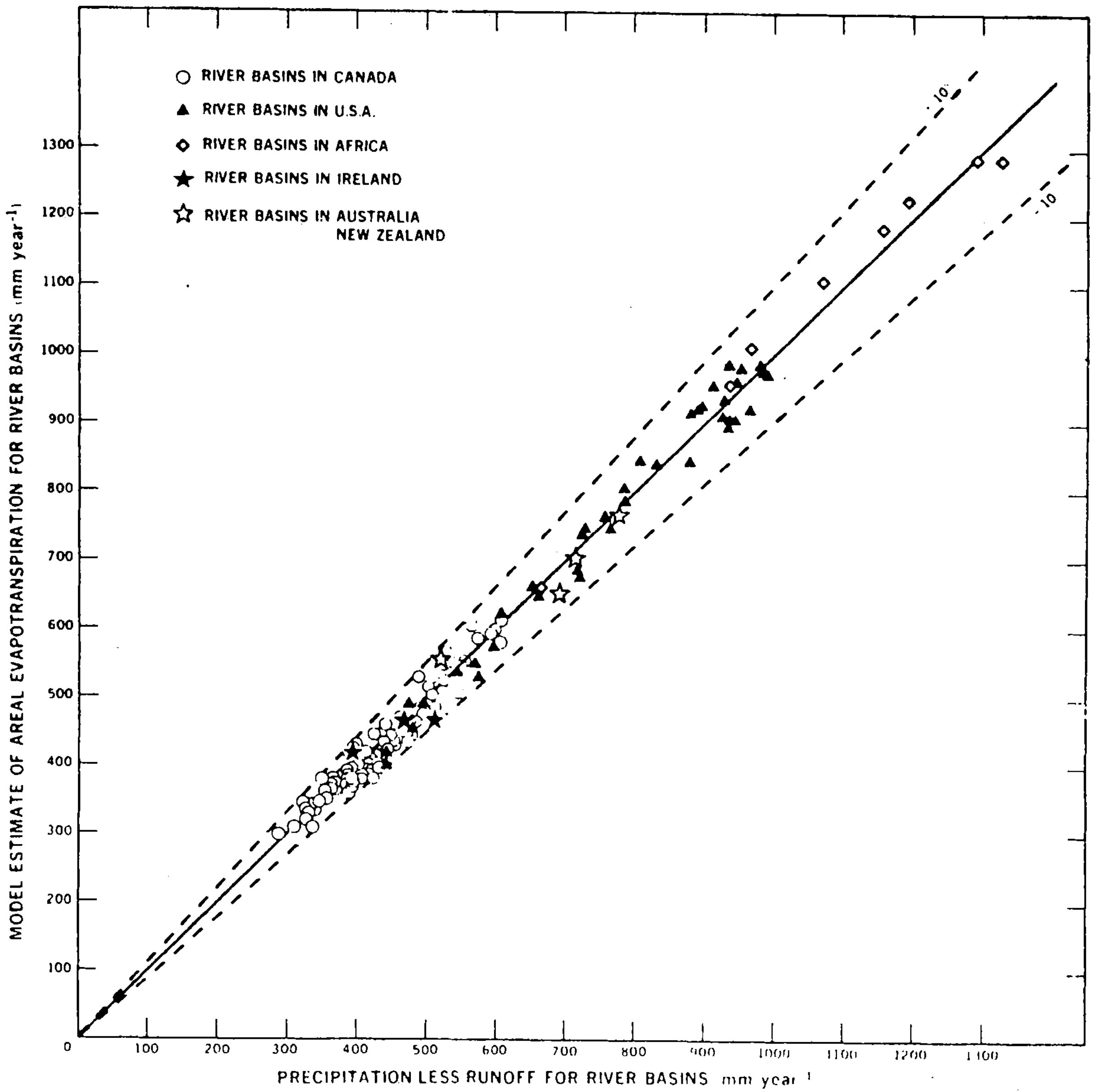


DIFFERENCE BETWEEN AVERAGE VAPOUR PRESSURE AND
VAPOUR PRESSURE AT AVERAGE DEWPOINT (mbar)
0
0.2
0.4
0.6
0.8
1.0
ESTIMATED VAPOUR PRESSURE CORRECTION (mbar)
0.2
0.4
0.6
0.8
1.0

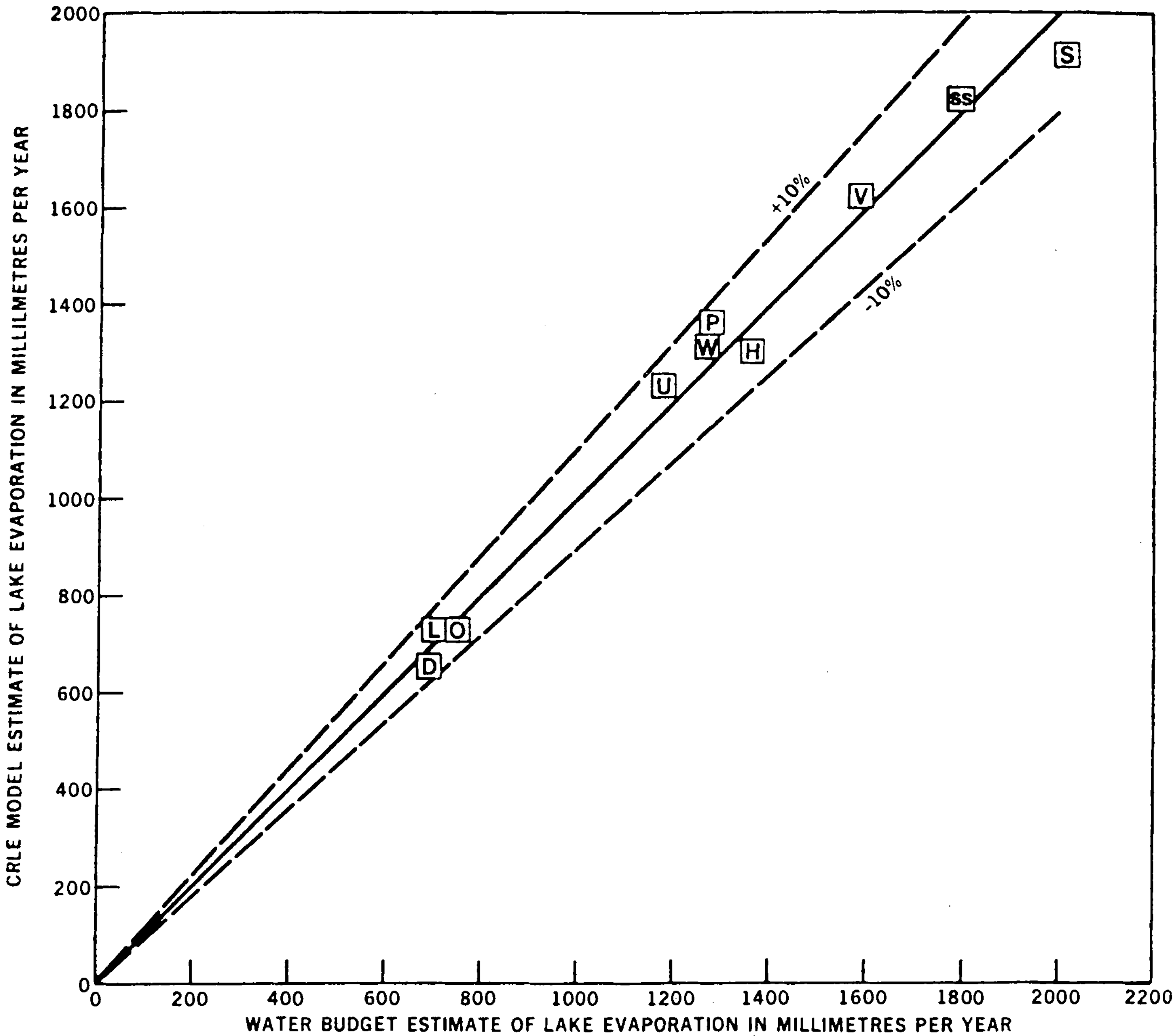


TRANSMITTANCY OF CLOUDY SKIES TO EXTRA-ATMOSPHERIC GLOBAL RADIATION (SEASONAL)
0
0.1
0.2
0.3
RATIO OF INCIDENT TO EXTRA-ATMOSPHERIC GLOBAL RADIATION (SEASONAL)
0
0.1
0.2
0.3
0.4
0.5
0.6
0.7

4



RIVER BASINS IN CANADA
RIVER BASINS IN U.S.A.
RIVER BASINS IN AFRICA
RIVER BASINS IN IRELAND
RIVER BASINS IN AUSTRALIA NEW ZEALAND
MODEL ESTIMATE OF AREAL EVAPOTRANSPIRATION FOR RIVER BASINS (mm year-1)
PRECIPITATION LESS RUNOFF FOR RIVER BASINS mm year-1


CRLE MODEL ESTIMATE OF LAKE EVAPORATION IN MILLILMETRES PER YEAR
WATER BUDGET ESTIMATE OF LAKE EVAPORATION IN MILLIMETRES PER YEAR
+10%
-10%
0
200
400
600
800
1000
1200
1400
1600
1800
2000
2200
S
SS
V
P
W
H
U
L
O
D

6

OPERATIONAL ESTIMATES OF LAKE EVAPORATION

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ABSTRACT

Morton, F.I., 1982e. Operational estimates of lake evaporation.

The complementary relationship between areal and potential evapotranspiration takes into account the changes in the temperature and humidity of the air as it passes from a land environment to a lake environment. Minor changes convert the latest version of the complementary relationship areal evapotranspiration (CRAE) models to a complementary relationship lake evaporation (CRLE) model. The ability of the CRLE model to produce reliable estimates of annual lake evaporation from monthly values of temperature, humidity and sunshine duration (or global radiation) observed in the land environment with no locally optimized coefficients is tested against comparable water budget estimates for 11 lakes in North America and Africa. Maps of annual lake evaporation and annual net reservoir evaporation (i.e. the difference between lake evaporation and areal evapotranspiration) for the part of Canada to the east of the Pacific Divide and for the southern United States are presented. An approximate routing technique, which takes into account the effects of depth and salinity on the seasonal pattern of monthly lake evaporation, is formulated and tested against comparable water budget estimates for 10 lakes in North America and Africa. The results indicate that the CRLE model, with its associated routing technique, is much superior to the other techniques in current use that rely on climatological or pan observations in the land environment.

INTRODUCTION

Operational estimates of lake evaporation must rely on readily available information and in practice this means climatological data or pan evaporation data that have been observed in the land environment. There are several problems associated with the use of such observations. The first is that seasonal changes in subsurface heat storage are not reflected directly in in pan evaporation or climatological data and that such changes are significant in determining seasonal variations in the evaporation from deep lakes. This problem is not too important because annual estimates are adequate for most water planning and management or environmental impact studies. Furthermore the effects of subsurface heat storage changes can be taken into account in an approximate way with routing techniques. A much more serious problem is that pan and climatological observations are influenced significantly by changes in the availability of water for evapotranspiration from the adjacent land and are, therefore, not representative of the environment over the lake.

Reliable information on the transition that takes place when the air passes from the land over a lake is rare if not non-existent. However, Davenport and Hudson (1967) have presented data on the transition from desert to irrigated cotton that can be used to show what happens at the upwind edge of a lake. They measured the variation in evaporation across a series of irrigated and fallow fields in the Sudan Gezira using fiberglass dishes with black painted wells 113 mm in diameter and 36 mm in depth. The passage of air from the desert (or from the unirrigated follow fields) over the irrigated cotton caused the dish evaporation above the cotton to decrease rapidly in the downwind direction and to approach a low constant value within 300 m, the width of the fields. Furthermore, as the air passed from irrigated cotton across unirrigated fallow, the dish evaporation above the fallow increased rapidly in the downwind direction and approached but did not reach the value observed at the upwind edge of the irrigation area. Fig. 1 shows the variation of dish evaporation across three irrigated fields on December 27, 1963. The ratio of daily dish evaporation at the downwind edges of the irrigated cotton to that at the upwind edge of the irrigated area was 0.69 for the field with "dry" soil, 0.60 for the field with "moist" soil and 0.53 for the field with "wet" soil.

The decreases in dish evaporation across the cotton fields were associated with decreases in temperature and increases in humidity. The vapour pressures appeared to attain equilibrium values within the 300-m width of the fields, but the temperatures were still decreasing, possibly because the observations were made above the level of the crop and the dishes.

Fig. 1 shows how the dish evaporation and potential evaporation increase when the water available for evapotranspiration from the area upwind decreases and how they decrease when the water available for evapotranspiration from the area upwind increases. Moreover, the dish evaporation for the "wet" field provides an indication of what happens over a lake in an arid climate. Thus the upwind dish evaporation reflects the potential evaporation in the desert and the low, relatively constant dish evaporation near the downwind edge reflects the potential evaporation over most of the lake. Furthermore, the dish evaporation from the "moist" and "dry" fields provides an analogy for what happens over lakes in progressively more humid climates where the contrasts between lake and land environments are less extreme. Because the transition zone is so narrow, the lake evaporation would approximate the low constant downwind value of potential evaporation.

The effects of changes in the availability of water to upwind areas can be taken into account by the complementary relationship between potential and areal evapotranspiration. A conceptual rationalization and a review of available theoretical knowledge and reliable empirical evidence in a companion paper (Morton, 1982b) have indicated that the complementary relationship is a much more plausible working hypothesis than most other hydrometeorological concepts. It is expressed in the following equation:

$$E_T + E_{TP} = 2\, E_{TW} \tag{1}$$

in which E_T is the areal evapotranspiration, the evapotranspiration from an area so large that the effects of upwind boundary transitions, such as those shown in Fig. 1 are negligible; E_{TP} is the potential evapotranspiration, as estimated from a solution of the vapour transfer and energy balance equations, representing the evapotranspiration that would occur from a hypothetical moist surface with radiation absorption and vapour transfer characteristics similar to those of the area and so small that the effects of the evapotranspiration on the overpassing air would be negligible; and E_{TW} is the wet environment areal evapotranspiration, the evapotranspiration that would occur if the soil-plant surfaces of the area were saturated and there were no limitations on the availability of water.

Fig. 2 provides a schematic representation of eq. (1) under conditions of constant radiant energy supply. The ordinate represents evapotranspiration and the abscissa represents water supply to the soil-plant surfaces of the area, a quantity that is usually unknown. When there is no water available for areal evapotranspiration (extreme left of Fig. 2) it follows that $E_T = 0$, that the air is very hot and dry and that E_{TP} is at its maximum rate of $2E_{TW}$ (the dry environment potential evapotranspiration). As the water supply to the soil-plant surfaces of the area increases (moving to the right in Fig. 2) the resultant equivalent increase in E_T causes the overpassing air to become cooler and more humid which in turn produces an equivalent decrease in E_{TP}. Finally, when the supply of water to the soil-plant surfaces of the area has increased sufficiently, the values of E_T and E_{TP} converge to that of E_{TW}.

The conventional definition for potential evapotranspiration is the same as the definition for the wet environment areal evapotranspiration. However the potential evapotranspiration that is estimated from a solution of the vapour transfer and energy balance equations by analytical (Penman, 1948), graphical (Ferguson, 1952) or iterative (Morton, 1982b) techniques has reactions to changes in the water supply to the soil-plant surfaces similar to those shown for E_{TP} in Fig. 2, so that what is being estimated can exceed what is being defined by as much as 100 percent. By taking into account such reactions the complementary relationship is analogous to the Bernouilli equation for open channel flow in which the potential energy responds in a complementary way to changes in kinetic energy.

The chief advantage of the complementary relationship is that it permits the areal evapotranspiration, a product of complex processes and interactions in the soil-plant-atmosphere continuum, to be estimated by its effects on the routine climatological observations that are used to compute potential evapotranspiration. Because the model avoids the complexities of the soil-plant system it requires no local optimization of coefficients and is, therefore, falsifiable. This means that it can be tested rigorously so that errors in the associated assumptions and relationships can be detected and corrected by progressive testing over an ever-widening range of environments. Such a methodology uses the entire world as a laboratory and requires that a correction made to obtain agreement between model and river basin water budget estimates in one environment must be applicable without modification in all other environments. The estimates resulting from the most recent application of this methodology (Morton, 1982d) agree closely with comparable long-term water budget estimates for 143 river basins in North America, Africa, Ireland, Australia and New Zealand.

The evaporation from a shallow lake, E_W, differs from the wet environment areal evapotranspiration, E_{TW}, only because the radiation absorption and vapour transfer characteristics of water differ from those of vegetated land surfaces. The potential evaporation, E_P, differs from the potential evapotranspiration, E_{TP}, for the same reasons. Although the lake evaporation is equal to the potential evaporation in the lake environment it can differ significantly from the potential evaporation in the land environment.

Fig. 3 provides a schematic representation of the relationship between shallow lake evaporation and the potential evaporation in the land environment under conditions of constant radiant energy supply. The ordinate represents evaporation and the abscissa represents the water supply to the soil-plant surfaces of the land environment. Since a lake is defined to be so wide that the effects of the kind of upwind transition shown in Fig. 1 are negligible, the lake evaporation is independent of variations in the water supply to the soil-plant surfaces of the land environment. However, the complementary relationship predicts that the potential evaporation in a completely dry land environment would be twice the lake evaporation and that it would decrease in response to increases in the water supply to the soil-plant surfaces until it reached a minimum equal to the lake evaporation as shown in Fig. 3.

The conventional techniques for estimating lake evaporation are based on the assumption that the evaporation estimated from pan or climatological observations in the land environment can be transposed to a nearby lake by applying a simple coefficient. The apparent lack of alternatives has led to the practice of ignoring the increasing body of evidence that these techniques are unrealistic. Such evidence is implicit in a tabulation published by Hounam (1973) which shows that the annual Class A pan coefficient is 0.81 for Lake Okeechobee in Florida, where the average annual precipitation is about 1400 mm; 0.70 for Lake Hefner in Oklahoma, where the average annual precipitation is about 800 mm; and 0.52 for the Salton Sea in California, where the average annual precipitation is about 60 mm. These kind of variations undermine the foundations of the conventional techniques because they indicate that lakes create their own environments which differ more and more from the land environments as the land environments become more arid. However they are compatible with the complementary relationship and the kind of interactions shown in Figs. 1 and 3, which predict that the lake evaporation would be equal to the potential evaporation in a wet land environment and would be equal to half the potential evaporation in a dry land environment.

The transposability from one lake to another of the coefficients required in the application of the conventional techniques will always be open to doubt because of the infinite variety of land environments and the lack of knowledge on now they influence the pan evaporation and the climatological observations. In this context the complementary relationship seems an attractive alternative. Its use as the basis for the latest version of what are now referred to as the CRAE (i.e. complementary relationship areal evapotranspiration) models; the application of this latest version to provide operational estimates of areal evapotranspiration from routine observations of temperature, humidity and sunshine duration (or global radiation); the test of the results against comparable water budget estimates for 143 river basins in North America, Africa, Ireland, Australia and New Zealand; and examples of how reliable operational estimates of areal evapotranspiration can be a major factor in the transformation of hydrology from a descriptive to a predictive science are described in companion papers (Morton, 1982a, 1982b, 1982c, 1982d). The transformation of the latest version of the CRAE models to the latest version of the CRLE (i.e. complementary relationship lake evaporation) models may be regarded simply as spinoff because the required changes are so minor. Presented herein are descriptions of the required changes; of the use of the resultant CRLE model to provide operational estimates of shallow lake evaporation from the routine observations of temperature, humidity and sunshine duration (or global radiation) in the land environment; of the development of a routing procedure to take into account the effects of seasonal subsurface heat storage changes in converting estimates of shallow lake evaporation to estimates of deep lake evaporation; of a test of the results against comparable water budget estimates for eleven lakes in Norther America and Africa; and of ways in which operational estimates of lake evaporation, in combination with operational estimates of areal evapotranspiration, can be applied to problems of water planning and management.

The Appendix presents descriptions and data sources for the ten lakes with water budget estimates of evaporation that are used in the development and testing of the CRLE model.

SHALLOW LAKE EVAPORATION

The modus operandi for the latest version of the CRAE (complementary relationship areal evapotranspiration) models is presented in detail in a companion paper (Morton, 1982c). The required station characteristics are the latitude in degrees, the altitude in metres and a rough estimate of average annual precipitation in millimetres per year. The required climatological inputs are monthly values of dewpoint temperature in degrees Celsius, air temperature in degress Celsius and the ratio of observed to maximum possible sunshine duration. The outputs are monthly values ot the net radiation that would occur if the surface were at air temperature, the potential evapotranspiration and the areal evapotranspiration, all in millimetres of evaporation or, in the case of net radiation, of evaporation equivalent. The minor changes that are needed for humidity, temperature and insolation input options and for shorter time period options are presented in detail thereafter. The modifications to the equations and constraints [of the reference (Morton, 1982c)] that are required to convert the CRAE model to a CRLE model are summarized hereunder.

(1) The zenith value of snow-free, clear-sky albedo (a_{zz}) in eq. (16) is assumed constant at 0.05. This change permits the deletion of eqs. (4) and (14) and their constraints.

(2) The emissivity (ϵ) in eqs. (28) and (33) and in constraint (28a) is assumed to be 0.97 so that the value of $\epsilon\sigma$ is 5.5 x 10^{-8} Wm^{-2} $(°K)^{-4}$.

(3) The value of the constant f_Z in eqs. (31) and (32) is 25.0 Wm^{-2} $mbar^{-1}$ with $T \geq 0°C$ and 25.0 x 1.15 Wm^{-2} $mbar^{-1}$ with $T < 0°C$.

(4) The value of b_0 in eq. (31) is 28/25 or 1.12.

(5) The values of b_1 and b_2 in eq. (40) are 13 Wm^{-2} and 1.12 respectively.

(6) Eq. (41) is deleted.

(7) The output symbols R_T, E_{TP} and E_{TW} in eqs. (29), (38), (40), and constraint (40a) should be changed to R_W (net radiation with water surface at air temperature), E_P (potential evaporation) and E_W (shallow lake evaporation) respectively.

Modifications 1) and 2) are self-explanatory since they make the radiation estimates compatible with water surfaces rather than soil-plant surfaces. Modifications 3), 4) and 5) take into account the differences in roughness between water and soil-plant surfaces. The procedure used to derive f_Z by trial-and-error in conjunction with the calibration of the constants b_1 and b_2 using monthly climatological data from arid regions is described in a companion paper (Morton, 1982b).

The latest versions of the CRAE and CRLE models have been documented in FORTRAN and in RPN notation for the Hewlett Packard HP-67 hand-held calculator (Morton, Goard and Piwowar, 1980). They have also been documented in RPN notation for the Hewlett Packard HP-41C hand-held calculator and this can be made available on request. The greater storage capacity of HP-41C eliminates the need for inserting extra program cards during individual computations and permits the full range of options.

A lake has been defined as a body of water so wide that the effects of the kind of upwind transitions shown in Fig. 1 are negligible. A shallow lake is one for which seasonal subsurface heat storage changes are insignificant. Estimates of subsurface heat storage changes can be incorporated into the CRLE models by adding them to the net radiation estimates in eq. (29) of the relevant companion paper (Morton, 1982c). In practical terms this is not possible because of the huge amounts of time, money and effort that are required for the measurement of temperature profiles with adequate spatial and temporal resolution. Seasonal subsurface heat storage changes can be taken into account in an approximate way with a routing technique that is described in a subsequent section. Furthermore they can be ignored if the period of interest is one or more integral years. In other words any lake is a shallow lake if one is interested only in the annual or mean annual evaporation.

The latest version of the CRLE model should provide reliable estimates of shallow lake evaporation anywhere in the world from records of temperature, humidity and sunshine duration (or global radiation) observed in the land environment with no need for locally optimized coefficients. This capability has been tested with water budget estimates of evaporation from Lake Victoria in East Africa, Salton Sea and Silver Lake in California, Lake Hefner in Oklahoma, Pyramid and Winnemucca Lakes in Nevada, Utah Lake in Utah, Lake Ontario in the North American Great Lakes System, Last Mountain Lake in Saskatchewan and Dauphin Lake in Manitoba. Descriptions and data sources for these lakes are presented in the Appendix. The CRLE model was used to provide monthly estimates of shallow lake evaporation for these ten lakes. To avoid the problems with seasonal changes in subsurface heat storage, the monthly model estimates were accumulated to provide annual or mean annual values and plotted against the comparable water budget estimates in Fig. 4. The maximum and average absolute deviations of the model estimates from the line of equality are 100 and 49 mm year^{-1} respectively while the maximum and average absolute percentage deviations are 5.6 and 3.9 percent respectively.

Pouyaud (1979) has published water budget estimates and the required climatological inputs for Lac de Bam, a long narrow lake located at 13°20' North and 1°30' West on an intermittent tributary of the White Volta River in the Republic of Upper Volta. The water budget estimates of evaporation are questionable and incomplete because of difficulties in estimating seepage outflows and rainy season inflows. In fact there was only one year, the year ending on September 30, 1976, with continuous water budget estimates and the required climatological data. Based on monthly temperature, vapour pressure and sunshine duration data observed a few kilometres from the lake, the CRLE model estimates of shallow lake evaporation for that year total 2153 mm, approximately 8 percent less than the comparable water budget estimate of 2346 mm. Furthermore, global radiation observations for eight of the months and global radiation estimates for the remaining four months (obtained by adjusting observations from the same months of other years in accordance with differences in sunshine duration) have been used to replace the sunshine duration observations as input. The resultant CRLE model estimates of shallow lake evaporation total 2304 mm, approximately 2 percent less than the comparable water budget estimate. The estimates for Lac de Bam are not included in Fig. 4 because the difference between the estimates based on sunshine duration inputs and those based on global radiation inputs is too large to be ignored.

The comparison between model and water budget estimates for ten lakes in Fig. 4 and for Lac de Bam demonstrates that the most recent version of the CRLE models can use routine climatological observations to provide reliable estimates of annual lake evaporation over a wide range of environments with no locally derived coefficients. However the model results do not agree nearly so well with published energy budget estimates for Lake Mead on the Colorado River and Lake Nasser on the Nile River. Thus the energy budget estimate for Lake Mead during the year ending September 30, 1953 (U.S. Geological Survey, 1958) was 2163 mm whereas the comparable model estimate, based on monthly dew point and air temperature observations at Boulder City and monthly global radiation observations at Boulder Island, was only 1680 mm. The two quantities are not strictly comparable because the energy budget estimates include the effects of net waterborne heat inputs, an energy term that is significant only for deep reservoirs on large rivers in hot climates, and releases of subsurface heat storage, an energy term that is usually insignificant on an annual basis. However, when these two terms are used in the CRLE model by adding them to the net radiation estimates in eq. (29) of the relevant companion paper (Morton, 1982c), the estimate is increased to

1855 mm year^{-1}, which is still almost 15 percent less than energy budget estimate. This type of discrepancy is even more evident in a comparison for Lake Nasser where Omar and El-Bakry (1981) have used monthly average estimates of different meteorological elements over the lake to produce an energy budget estimate of 2689 mm year^{-1}. This is much higher than the value of 2087 mm year^{-1} that is derived by using monthly averages of dew point temperature, air temperature and global radiation recorded near the south end of the lake at Wadi Halfa (Griffiths, 1972), as input to the CRLE model. By incorporating the monthly values of net waterborne heat inputs and releases of subsurface heat storage that have been estimated by Omar and El-Bakry (1981) the model estimate is increased to 2160 mm year^{-1}, which is still almost 20 percent less than the energy budget value.

In evaluating the foregoing results it should be noted that although the energy budget technique is based on the law of conservation of energy, it has a number of drawbacks. These are that it relies on the questionable assumption that the Bowen-ratio provides an adequate estimate of the ratio of sensible to latent heat fluxes under all the peculiar conditions of atmospheric stability that can occur over a lake; that the extrapolation of the inputs required for the energy budget technique from a few measurement points to an entire lake can lead to significant error; and that the energy budget had never been rigorously tested by applying an <u>identical</u> technique to a number of lakes in different environments and comparing the results with the applicable water budget estimates. The nearest approach to such a test has been performed in Australia (Hoy and Stephens, 1979) but this was not satisfactory because the lakes were unsuitable for water budget studies and it was assumed that the energy budget estimates were superior. The word "identical" has been stressed because it is easy to obtain significantly different results by seemingly arbitrary selection of techniques. For example, the Lake Nasser estimates produced by Omar and El-Bakry (1981) depend on a technique for estimating net long-wave radiation that does not take into account the effects of atmospheric water vapour and they differ significantly from those that would have resulted from the use of any of the alternate techniques that do take into account such effects.

It is easy to recognize that CRLE model estimates of annual lake evaporation would be orders of magnitude easier, faster and cheaper than energy budget estimates. What is not so easy to accept is that such estimates may also be more realistic. However, the considerations outlined above indicate that such a conclusion is not only possible but also probable.

The test of the areal evapotranspiration model [Morton, 1982d] required the preparation of evapotranspiration maps for the part of Canada to the east of the Pacific Divide and for the southeastern United States. This involved punching computer cards with 5 years of monthly air temperatures, dew point temperatures, and sunshine duration ratios for 37 climatological stations in the United States and 153 climatological stations in Canada. The availability of the cards provided the incentive to prepare similar maps for lake evaporation and for the difference between lake evaporation and areal evapotranspiration.

Fig. 5 is a map of the southeastern United States showing the average annual lake evaporation for the 5 years ending September 30, 1965, and Fig. 6 is a map of the part of Canada to the east of the Pacific Divide showing the average annual lake evaporation for the 5 yr ending December 31, 1969. They are based on the accumulation of monthly CRLE model estimates for all climatological stations in the two areas that report both air and dew point temperatures. The isopleths were plotted by linear interpolation between the stations. It should be noted that the estimates shown in the two maps do not apply to bodies of water so small that the type of edge effect shown in Fig. 1 would be significant or to lakes where there are large net waterborne heat inputs.

A comparison of Fig. 5 with the map on Plate 18 of the Water Atlas of the United States (Miller et al., 1963) and a comparison of Fig. 6 with the map on Plate 17 of the Hydrological Atlas of Canada (Canadian National Committee for the IHD,1978) indicate that the model estimates are generally higher in areas with high relative humidities and lower in areas with low relative humidities. The reasons for these discrepancies, which can be highly significant, are that all four maps are based on data from the land environment and that the CRLE model takes into account the differences between the land and the lake environments.

Fig. 7 is a map of the southeastern United States showing the difference between the average annual lake evaporation and the average annual areal evapotranspiration for the 5 years ending September 30, 1965, and Fig. 8 is a map of the part of Canada to the east of the Pacific Divide showing the difference between the average annual lake evaporation and the average annual areal evapotranspiration for the 5 years ending December 31, 1969. The lake evaporation estimates are those used to prepare Figs. 5 and 6, and the areal evapotranspiration estimates are those used in testing the areal evapotranspiration model (Morton, 1982d). The isopleths were plotted by interpolation between the 190 climatological stations in the two areas that report both air and dew point temperatures.

The purpose of Figs. 7 and 8 is to provide an estimate of net reservoir evaporation, or the effect on the long-term water balance of constructing a reservoir when edge effects and net waterborne heat inputs are insignificant. Such estimates are often needed for water resource or environmental impact investigations. The maps probably give values that are somewhat too high because the vegetation flooded by the reservoir would tend to transpire at a higher rate than the vegetation around the airports, where climatological stations are normally located.

DEEP LAKE EVAPORATION

The CRLE models do not take into account the effects of seasonal changes in subsurface heat storage so that the monthly estimates of the evaporation are realistic only for shallow lakes or when accumulated to provide annual totals. It is expected that the data required to provide physically based short term estimates will seldom be available on a routine basis so it is fortunate that annual estimates are adequate for most engineering and hydrologic applications. However it is possible to take subsurface heat storage into account in an approximate way by applying storage routing techniques similar to those used in routing water through natural reservoirs in hydrology. The storage (V) is related to the deep lake evaporation (E_L) in:

$$V = kE_L\ [1 + 7\exp(-\ E_L/12)] \tag{2}$$

in which k is the storage constant in units of months and the constant 12 in the argument of the exponential term is in units of mm month^{-1}. The nonlinearity represented by the exponential term is of significance only during high latitude winters.

The routing process also requires a time delay (t) which is taken into account by using input data for the preceding months in the following way:

$$E_W^t = E_W^{[t]} + (t - [t])(E_W^{[t+1]} - E_W^{[t]}) \quad . \tag{3}$$

in which the E_W are the CRLE model estimates of shallow lake evaporation, the superscripts refer to the delay time or the number of months backward it is necessary to go to obtain a value of E_W for the computations of the current month. The equation is used to estimate the value of E_W with a delay time that has both integral and fractional components, i.e. [t] and (t-[t]) respectively, from the values of E_W for two of the preceding integral months.

The solution of the storage and water balance equation must be iterative because of the exponential term in eq. (2). A rapidly converging solution may be obtained using the following relationships

$$\delta E_{LE} = \frac{E_W^t - \frac{1}{2}(E_{LB}+E_{LE}^*) + kE_{LB}\ [1+7\exp(-E_{LB}/12)] - kE_{LE}^*[(1+7\exp(-E_{LE}^*/12)]}{\frac{1}{2} + k + 7k\ (1 - E_{LE}^*/12)\ \exp\ (-\ E_{LE}^*/12)} \tag{4}$$

$$E_{LE} = \delta E_{LE} + E_{LE}^* \tag{5}$$

in which E_{LB} is the deep lake evaporation at the end of the preceding month, E_{LE} is the deep lake evaporation at the end of the current month, E_{LE}^* is a trial value of E_{LE} and δE_{LE} is the estimated correction. Eqs. (4) and (5) are repeated until $|\delta E_{LE}| \leq 0.01$ mm month^{-1} , a goal that is reached within four iterations. The initial value of E_{LE}^* is E_{LB} and subsequent values are those estimated from eq. (5) during the preceding iteration. In eq. (4) the numerator is the error in the water balance and the denominator is the derivative of the numerator with respect to E^*_{LE} with the sign changed.

The monthly value of deep lake evaporation is estimated from:

$$E_L = \frac{1}{2} (E_{LB} + E_{LE}) \tag{6}$$

The routing constant k and the delay time t (both in months) are estimated from:

$$k = d \left[0.04 + \frac{0.11}{1 + (d/16)^2}\right] \tag{7}$$

$$t = 0.50\ k \tag{8}$$

in which d is the effective depth of the lake in m. It is estimated from the average depth (d_A) in m and the concentration of total dissolved solids (s) in ppm using:

$$d = \frac{d_A}{1 + 0.00003s} \tag{9}$$

The derivation of eqs. (7), (8) and (9) is described subsequently.

The results of any routing procedure depend on a reasonable estimate of the starting point. Although errors resulting from an arbitrary starting point wear off quite quickly it is usually worthwhile to repeat the computations for the first year once and use the end-of-year results for the first trial as the starting point for the second trial. When using monthly values for a single year or average monthly values for an average year it is worthwhile to repeat the computations until E_{LE} for the last month approaches quite closely the value for the preceding trial. This makes the annual deep lake evaporation equal to the annual shallow lake evaporation. The nature of the routing equation produces fast convergence so that this criterion, which has been used to produce the results presented herein, can be met with few repetitions.

Eqs. (7), (8) and (9) were derived by using the monthly or average monthly water budget estimates of evaporation and model estimates of shallow lake evaporation for Lake Ontario (d_A = 86 m and s = 100 ppm), Pyramid Lake (d_A = 61 m and s = 3500 ppm), Lake Hefner (d_A = 8.2 m and s = 800 ppm), Salton Sea (d_A = 8.0 m and s = 37000 ppm) and Last Mountain Lake (d_A = 7.6 mm and s = 1700 ppm) in accordance with the procedure set out hereunder.

(1) The iterative procedure of eqs. (3), (4), (5) and (6) was used to estimate monthly or average monthly values of deep lake evaporation for Lake Ontario and Pyramid Lake with many different combinations of storage constant (k) and delay time (t). For the two combinations that gave the best fit to the comparable water budget estimates of each of these deep lakes, the delay time was very close to one half the storage constant as shown in eq. (8).

(2) The iterative procedure was then used again to estimate monthly or average monthly values of deep lake evaporation for Lake Ontario, Pyramid Lake, Lake Hefner, Salton Sea and Last Mountain Lake using many different trial estimates of the storage constant and the delay time estimated from eq. (8). The five trial estimates of the storage constant that gave the best fit to the monthly water budget estimates of each lake were used in the formulation of eqs. (7) and (9). Note that eq. (9) depends almost entirely on data for the Salton Sea where the concentration of total dissolved solids is more than ten times higher than that for Pyramid Lake, the lake with the next highest concentration.

Fig. 9 has been prepared to provide comparisons between monthly or average monthly values of shallow lake evaporation as estimated with the latest version of the CRLE models; of deep lake evaporation as estimated from E_W and the previously described routing procedure; and of the lake evaporation as estimated from the water budget. The comparisons are for the ten lakes described in the Appendix and used in the preparation of Fig. 4. These include the five lakes used to derive eqs. (7), (8) and (9). In evaluating the results, the uncertainties in the measurement of end-of-month lake levels and their contribution to error in the monthly water budget estimates should be kept in mind. Thus the Lake Hefner water budget estimates for the months of February and March provide grounds for suspicion that the estimated water level for the end of February was four to five cm too high. From this point of view Fig. 9 shows that the application of the previously described routing procedure to the CRLE model estimates of shallow lake evaporation can provide reasonably reliable estimates of monthly deep lake evaporation. In this regard it may be of interest to note that the average absolute deviation from the water budget estimates is 16.7 mm/month for the deep lake estimates as compared to 32.3 mm/month for the shallow lake estimates.

POND EVAPORATION

The CRLE model does not take into account the effects of increased evaporation at the upwind transition. These effects can be ignored for lakes but they could be significant for ponds or other small bodies of water. The nature of the transition is shown in Fig. 1 where the dish evaporation at the upwind edge of the cotton fields is analogous to the potential evaporation in the land environment and the dish evaporation at the downwind edge of the cotton fields is approaching a low constant value that is analogous to the lake evaporation. The transition can be approximated by

$$E_{PX} = E_L + (E_P - E_L)/(1 + X/C) \tag{10}$$

in which E_{PX} is the potential evaporation at distance X downwind of the upwind shoreline, E_P is the potential evaporation in the land environment, E_L is the deep lake evaporation (i.e. the potential evaporation downwind of the transition) and C is a constant. In Fig. 1, the value of C is 8 m for the "wet" field, 10 m for the "moist" field and 30 m for the "dry" field. Superficially it would seem that the value for the "wet" field is more appropriate for a lake. However the transition may be wider when the environmental contrasts are less extreme and when there is an increase in wind speed over water rather than a decrease over cotton. Therefore the value of C is conservatively estimated to be 13 m, the geometric mean of the value for the 3 fields.

The average evaporation for a lake that is X m wide in the crosswind direction (E_{LX}) can be estimated from the following integration of eq. (10):

$$E_{LX} = E_L + (E_P - E_L)\,\frac{\ln\,(1 + X/C)}{X/C} \tag{11}$$

and the percent error involved in using E_L as the lake evaporation is

$$100\,\left(\frac{E_{LX}}{E_L} - 1\right) = 100\,\left(\frac{E_P}{E_L} - 1\right)\,\frac{\ln\,(1 + X/C)}{X/C} \tag{12}$$

Lac de Bam, the lake in Upper Volta discussed in a preceeding section, has a length of about 30 km roughly perpendicular to the prevailing winds and a maximum width of 800 m. The latter figure provides a reasonable estimate of the average crosswind width X. The ratio E_P/E_W for the twelve months averaged 1.56 and this provides a good estimate of the annual value for E_P/E_L. When used in Eq. (13)

these figures indicate the annual CRLE evaporation estimates for Lac de Bam are approximately 3.8 percent too low.

CONCLUDING DISCUSSION

The CRLE model estimates are most sensitive to errors in the required sunshine duration or radiation inputs. They are relatively insensitive to errors in the dew point temperature or other optional humidity inputs and in the air temperature inputs. Furthermore it doesn't matter much where in the vicinity of the lake the temperature and humidity inputs are observed because the complementary relationship automatically takes into account the effects of differing surroundings. Thus the difference between estimates derived from observations in the land environment and estimates derived from observations over the lake would be due primarily to the relatively minor effect of the difference in humidity on the estimates of net radiation.

One of the objections to the CRLE models is that they do not take into account the effects of wind speed on lake evaporation. Such an objection ignores the following considerations:

(1) Vapour is transported away from a lake in a vertical direction by wind-induced turbulence and heat-induced turbulence. Heat-induced turbulence preponderates over wind-induced turbulence during periods of high evaporation and it becomes more pronounced at low wind

speeds so that attempts to plot evaporation per unit vapour pressure difference against wind speed always resemble buckshot on a barn door.

(2) The effects on evaporation of an increase in wind speed are partially offset by the effects on evaporation of the resultant decrease in surface temperature.

(3) Wind speed observations in the land environment are extremely sensitive to instrument height, upwind surface roughness and nearby obstructions, so they do not necessarily provide a good estimate of wind speed over a lake. Therefore the use of routinely observed wind speeds can lead to significant error.

The foregoing observations indicate that the use of routinely observed wind speeds in estimating lake evaporation does not significantly reduce error and may quite possibly increase it.

The CRLE models do not implicitly take into account seasonal change in subsurface heat storage. In this they are in no way inferior to the more conventional potential evaporation, pan evaporation and bulk aerodynamic techniques that also depend on data observed in the land environment. In fact, they are superior because they can readily accommodate explicit estimates of subsurface heat storage changes that have been determined from temperature soundings and when, as is usual, the results of such soundings are not available they can be coupled with the routing technique presented herein to produce estimates of deep lake evaporation that reflect the storage changes. Comparisons between monthly routed values of deep lake evaporation and the corresponding water budget values for ten lakes indicates that the routing technique provides reasonably realistic estimates.

The CRLE models and the routing technique do no take into account the kind of upwind shoreline transition shown in Fig. 1. Therefore they are applicable only to lakes. However the results can be applied to ponds or other small bodies of water when modified using eq. (11).

The limitations and disadvantages of the CRLE models seem minor in comparison with the advantages that are summarized below:

(1) They require as input only land environment observations of temperature, humidity and sunshine duration and the results are relatively insensitive to errors in temperature and humidity.

(2) They require no local optimization or fudging of coefficients. This means that the results are falsifiable and can be tested rigorously against comparable water budget estimates anywhere in the world. The results presented herein show good agreement with the corresponding water budget estimates for nine lakes in North America and two lakes (including Lac de Bam) in Africa.

(3) They have a sound physical basis and are therefore easily adaptable to unusual applications. Thus it is easy to make the adaptations needed to estimate the effects of heat rejection from thermal power plants and to estimate the effects of net waterborne heat inputs to deep reservoirs on large rivers in hot, arid climates.

(4) The same input data and an almost identical model can be used to provide an estimate of the evapotranspiration that has taken place in the area where a reservoir is planned or the evapotranspiration that would have taken place if a reservoir did not exist. The difference between the estimated lake evaporation and the estimated evapotranspiration, the net reservoir evaporation, is an important quantity because it represents the effect of an existing or planned reservoir on the water balance of a basin.

The foregoing advantages make the CRLE models much superior to the conventional potential evaporation, pan evaporation or bulk aerodynamic techniques that also rely on data observed in the land environment. With regard to the second advantage it should be noted that no other technique (including the energy budget technique) has been tested so rigorously and therefore no other technique can be used with such confidence to provide estimates of lake evaporation anywhere in the world without the need for locally calibrated coefficients.

APPENDIX

DOCUMENTATION OF LAKES

The lakes used to develop and test the CRLE model in this and a companion paper (Morton, 1982b) are documented herein in order of increasing latitude.

Lake Victoria is located in the headwaters of the White Nile River in East Africa with its centre near 1°S and 33°E. It has an attitude of about 1130 metres above sea level, an average depth of 40 metres, an area of 68000 square kilometres and a concentration of total dissolved solids that is believed to be near 400 ppm. The average annual rainfall is about 1670 mm. Monthly water budget estimates of evaporation and the required climatological data for the five years ending December 31, 1974 have been supplied by the Chief Technical Advisor of the WMO Hydrometeorological Survey of Lakes Victoria, Kyoga and Mobuto Sese Seko. Kite (1981) has tabulated the average monthly values of water budget evaporation. The CRLE model estimates are based on the averages of the monthly temperatures and dew point temperatures recorded at Entebbe and Mwanza and the monthly values of sunshine duration averaged over the lake using shoreline and island stations. A second set of CRLE model estimates that use the lake average values of global radiation as an input option provided results only 2 percent less on an average annual basis and are therefore not used separately in the test of the model.

The Salton Sea is located at 33°15'N and 115°50'W in the state of California. During 1961 and 1962 it was approximately 71 m below sea level with an average depth of 8.0 metres and an area of 920 km^2. The load of total dissolved solids has been observed at approximately 37000 ppm. The climate is arid with an average annual precipitation between 20 and 60 mm. Monthly water budget estimates of evaporation and the required climatological inputs for the years 1961 and 1962 have been presented by Hughes (1967). Two different sets of CRLE model estimates have been made, both based on dewpoint and air temperatures observed at Sandy Beach. They differ in that one uses the sunshine duration observed at Yuma as input whereas the other uses the global and hemispheric radiation observed at Sandy Beach as input. Since the former set of model estimates exceeds the latter set by less than 2 percent on an average annual basis it has been selected for use in the preparation of Figs. 4 and 9.

Silver Lake is located at 35°25'N and 116°05'W in the state of California at an altitude of approximately 280 m above sea level. In March, 1938 the lake had an area exceeding 50 km^2 and a maximum depth of about 2 m but by September, 1939 most of the water in the lake had evaporated. The average depth and total dissolved solids for this period are assumed to have been 0.5 m and 500 ppm respectively. The climate is arid with an annual rainfall of about 100 mm. Water budget estimates of evaporation and observations of temperature and relative humidity for the twelve months ending April 30, 1939 have been presented by Blaney

(1957). There are no sunshine duration or global radiation records available for use as input. However the lake is almost equidistant from Las Vegas, where later global radiation observations were high but within the expected range, and Inyokern (China Lake), where later global radiation observations were abnormally high. These later records indicated that there is little year-to-year variation in global radiation in the vicinity of the lake. Therefore the CRLE model estimates of the evaporation from Silver Lake are based on the temperature and relative humidities observed at the lake and on the monthly values of global radiation observed at Las Vegas averaged over the 5 years ending December 31, 1964.

Lake Hefner is located in the state of Oklahoma at 35°35'N and 97°40'W. It is approximately 365 m above sea level, has an average depth of 8.2 m, an average area of 10.5 km^2 and the concentration of total dissolved solid is estimated to be about 800 ppm. The climate is subhumid with an average annual precipitation near 790 mm. Water budget estimates of evaporation from the lake for the 16 months ending August 31, 1951 were an important part of the well-known Lake Hefner studies (U.S. Geological Survey, 1954). The CRLE model estimates are based on dewpoint temperature, air temperature and sunshine duration records for Oklahoma City. The values of water budget and model estimates of lake evaporation for the months of May, June, July and August are averages for two years while those for the remaining eight months are for one year only.

Pyramid Lake and Winnemucca Lake are located between 119°W and 120°W at a latitude of 40°N and an altitude of about 1160 m above sea level in the state of Nevada. The climate is arid with an average annual precipitation of approximately 165 mm. The depth and area of Winnemucca Lake in early 1935 were roughly 5 m and 60 km^2 respectively but by late 1938 the lake was practically dry. The average depth and the concentration of total dissolved solids are unknown but can be assumed to be 1.2 m and 13000 ppm respectively without contributing to significant error. The average depth and average area of Pyramid Lake during 1935 and 1936 were estimated to be approximately 61 m and 460 km^2 respectively taking into account an apparent error of 9 feet in the gauge reading prior to 1971 (Geological Survey Water Supply Paper, 2127, 1974). The concentration of total dissolved solids for Pyramid Lake has been measured and the reported value was approximately 3500 ppm. Harding (1962) has presented water budget estimates of evaporation from Pyramid Lake for the two complete years ending on December 31, 1936 and from Winnemucca Lake for 37 of the 41 months ending on September 30, 1938. The comparable CRLE model estimates are based on dewpoint temperatures, air temperatures, and sunshine duration recorded at Reno, Nevada. The monthly average values of water budget and model estimates used in Figs. 4 and 9 are for the calendar years 1935 and 1936 in the case of Pyramid Lake and for from one to four years, depending on the availability of water budget data, in the case of Winnemucca Lake.

Utah Lake is located at 40°10'N and 111°50'W at an altitude of approximately 1370 m in the state of Utah. It has an area of 380 km^2, an average depth of 2.7 m and concentration of total dissolved solids of about 1000 ppm. The climate is semiarid with an average annual precipitation of approximately 340 mm. The water budget estimates of monthly evaporation, averaged over the three years ending June 30, 1973, have been presented by Miller and Merritt (1980). The comparable CRLE model estimates have been derived from dewpoint temperature, air temperature and sunshine duration data recorded at Salt Lake City.

Lake Ontario, the furthest downstream of the North American Great Lakes, is centred at 43°39'N and 77°47'W with an altitude of about 75 m above sea level. It has an area of 19100 km^2, an average depth of 86 m and a concentration of total dissolved solids of roughly 100 ppm. Water budget estimates of the evaporation for the year ending March 31, 1973 were prepared for the IFYGL, the International Field Year on the Great Lakes (Witherspoon, 1978). During this year the precipitation on the lake, as estimated from shoreline stations and radar was 1025 mm. The comparable CRLE model estimates are based on dewpoint and air temperatures that are averages for Toronto International, Trenton, and Kingston airports and on sunshine duration ratios that are averages for Kingston, Morven, Smithfield, Toronto, Hamilton and Vineland.

Last Mountain Lake is a long, narrow north-south lake centred at 51°06'N and 105°12'W in the province of Saskatchewan. It is located at an altitude of 490 m above sea level, has an area of 185 km^2, an average width of 2.5 km, an average depth of 7.6 m and a concentration of total dissolved solids of approximately 1700 ppm. Details of the water budget estimates of evaporation for the two relatively dry years of 1973 and 1977 are presented in Table I. The comparable CRLE model estimates are based on dewpoint temperatures, air temperatures and sunshine duration ratios that are averages for Wynyard and the Regina, Moose Jaw and Saskatoon airports.

Dauphin Lake is located at 51°15'N and 99°46'W in the Province of Manitoba at an altitude of 260 m above sea level. It has an area of 521 km^2, an average depth of 2.0 m and the concentration of total dissolved solids has been observed at approximately 300 ppm. The average annual precipitation for the years 1967 and 1968 was between 390 and 400 mm. Details of the waer budget estimates of evapotranspiration for 1967 and 1968 have been presented elsewhere (Morton, 1979). The comparable CRLE model estimates are based on the records of dewpoint temperature, air temperature and sunshine duration at Dauphin Airport.

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TITLES FOR FIGURES

Fig. 1 Comparison of evaporation rates across irrigated cotton fields on December 27, 1963 (Davenport and Hudson, 1967).

Fig. 2 Schematic representation of complementary relationship between areal and potential evapotranspiration with constant radiant energy supply.

Fig. 3 Schematic representation of relationship between shallow lake evaporation and potential evaporation in the land environment with constant radiant energy supply.

Fig. 4 Comparison of model estimates with water budget estimates of evaporation from Lake Victoria [V], Salton Sea [ss], Silver Lake [S], Lake Hefner [H], Pyramid Lake [P], Winnemucca Lake [W], Lake Ontario [O], Last Mountain Lake [L] and Dauphin Lake [D].

Fig. 5 Lake evaporation for the southern United States during the five years ending September 30, 1965.

Fig. 6 Lake evaporation for the part of Canada to the east of the Pacific Divide during the five years ending December 31, 1969.

Fig. 7 Net reservoir evaporation for the southern United States during the five years ending September 30, 1965.

Fig. 8 Net resevoir evaporation for the part of Canada to the east of the Pacific Divide during the five years ending December 31, 1969.

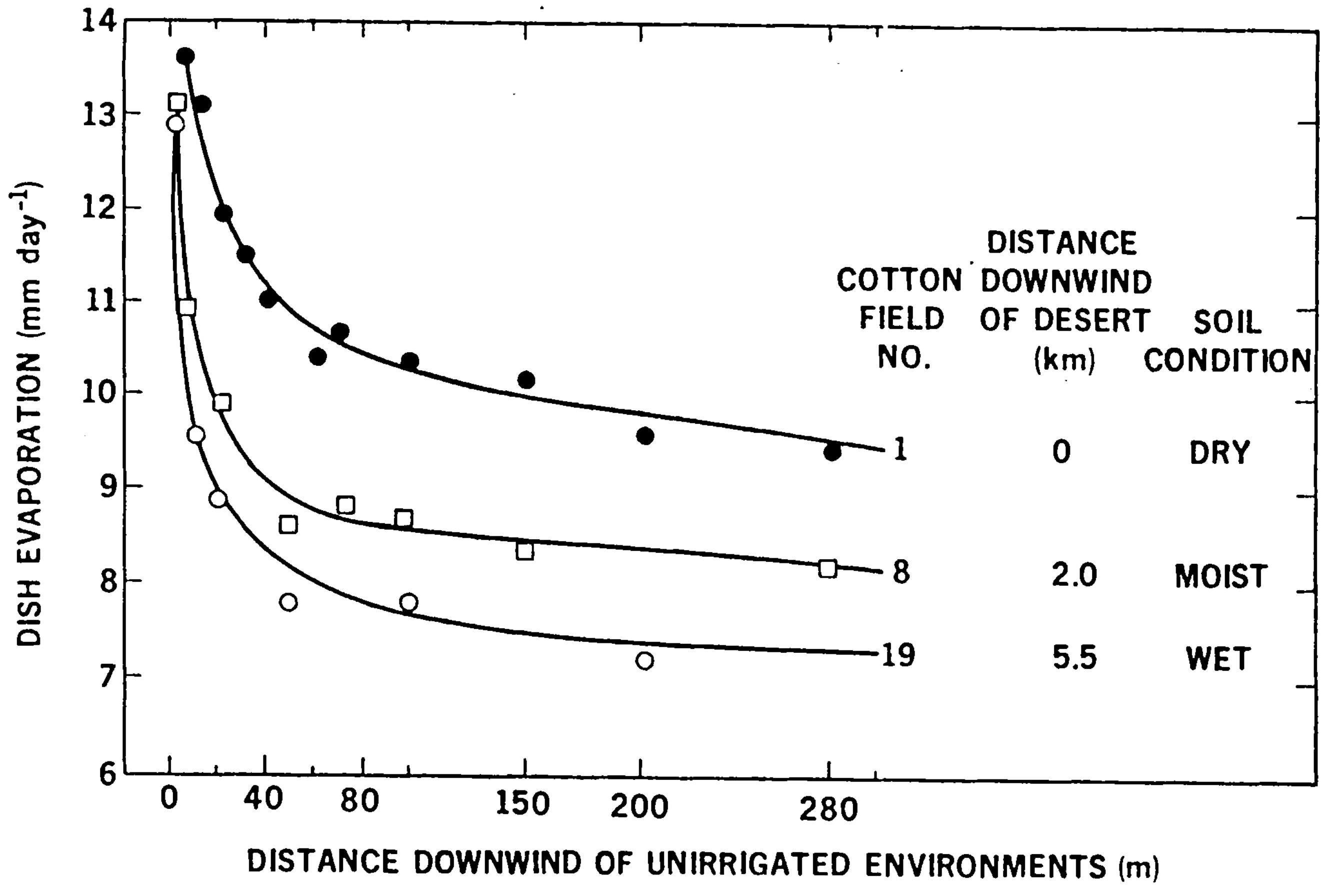
Fig. 9 Comparison of monthly model estimates of shallow lake and deep lake evaporation with comparable monthly water budget estimate.

TABLE I

COMPONENTS OF THE LAST MOUNTAIN LAKE WATER BUDGET

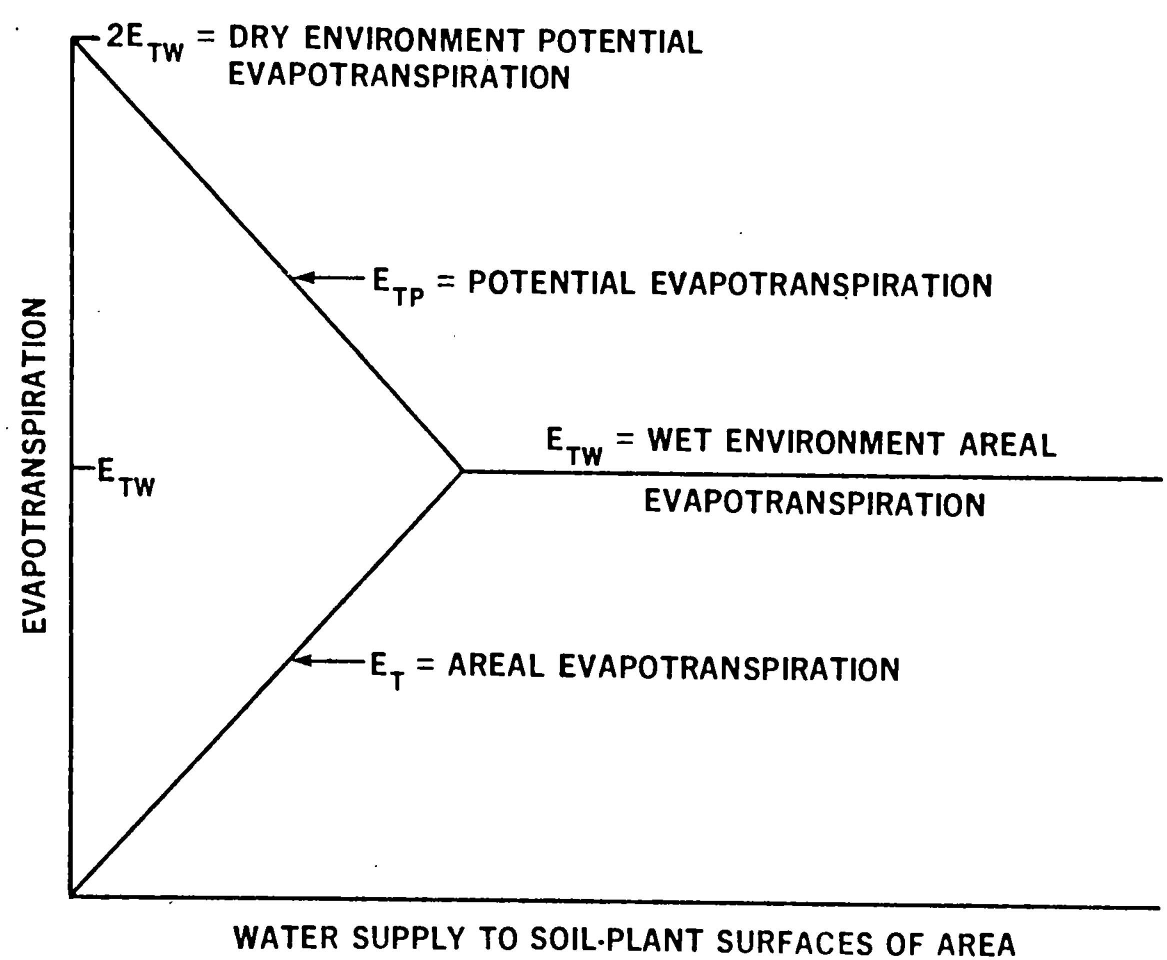
Year	Month	Precip- itation[1]	Gauged Inflow[2]	Ungauged Inflow[3]	Outflow[4]	Decrease in Level	Evapor- tion
		mm	mm	mm	mm	mm	mm
1973	Jan.	2	0	0	6	-1	-5
	Feb.	10	0	0	11	-1	-2
	Mar.	23	6	5	10	-32	-8
	Apr.	62	8	7	-19	-85	11
	May	38	10	8	-44	-93	7
	June	114	17	15	-9	-14	141
	July	38	11	9	-8	93	159
	Aug.	40	3	2	-5	97	147
	Sept.	41	5	4	-8	67	125
	Oct.	12	3	3	27	70	61
	Nov.	32	0	0	23	33	42
	Dec.	37	0	0	13	-11	13
	Total	449	63	53	-3	123	691
1977	Jan.	8	0	0	21[E]	13	0
	Feb.	1	0	0	10[E]	24	15
	Mar.	10	0	0	9	3	4
	Apr.	7	1	0	-9	-1	16
	May	145	3	3	-28	-69	110
	June	28	2	1	-18	57	106
	July	58	0	0	-11	52	121
	Aug.	26	0	0	1	111	136
	Sept.	48	0	0	0	53	101
	Oct.	5	1	0	-2[E]	59	67
	Nov.	21	0	0	-2[E]	28	51
	Dec.	30	0	0	5[E]	-40	-15
	Total	387	7	4	-24	290	712

[1] Average for Dilke, Duval, Last Mountain Wildlife., Lumsden, Nokomis and Strasbourg
[2] Total for Lanigan, Arm, Lewis and Saline Creeks (5793 km^2)
[3] Gauged value multipled by (5207 km^2/5793 km^2)
[4] Based on Rowans Ravine averages for 10 days before and after end of month
E Estimated from differences in discharges along Qu'Appelle River



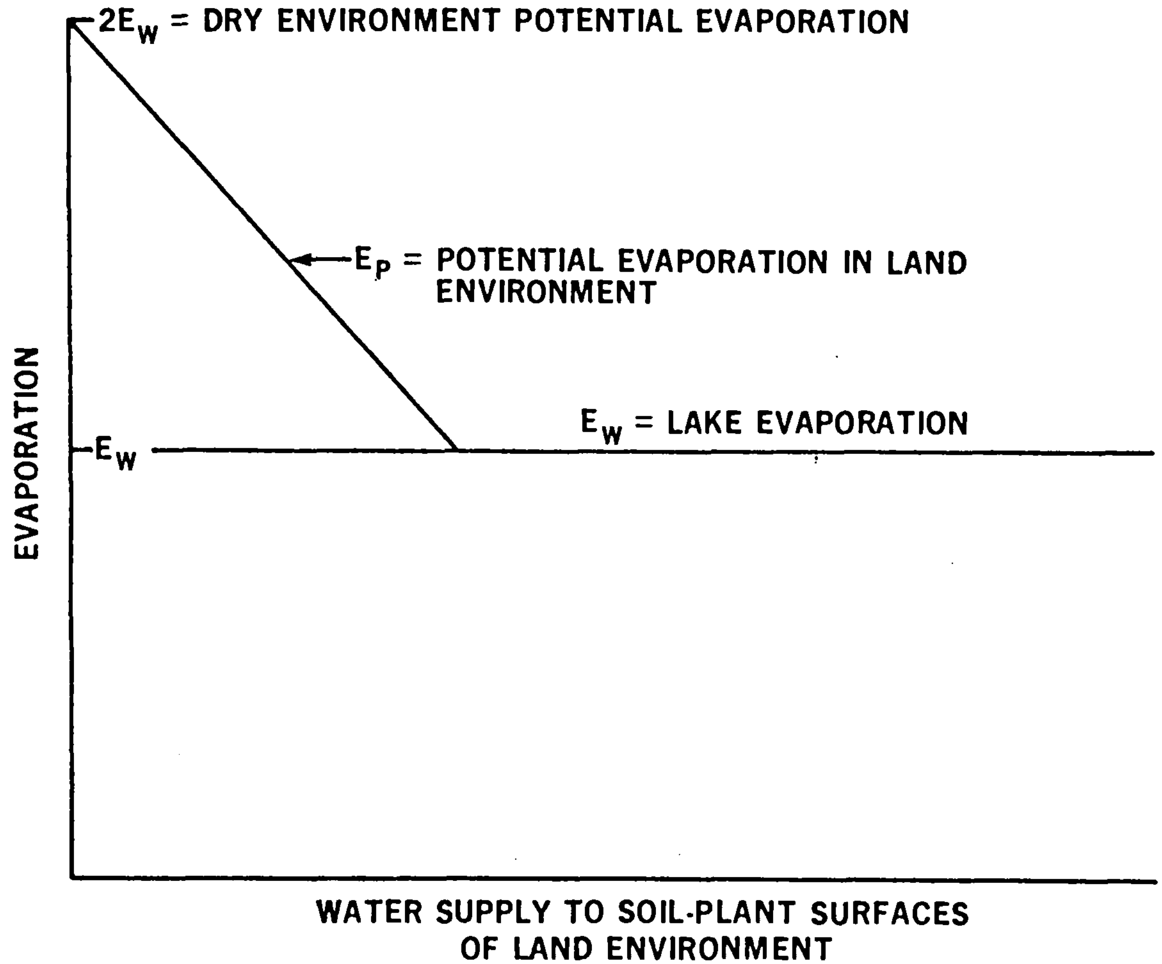
14
13
12
11
10
9
8
7
6
DISH EVAPORATION (mm day^{-1})
0
40
80
150
200
280
DISTANCE DOWNWIND OF UNIRRIGATED ENVIRONMENTS (m)
COTTON FIELD NO.
DISTANCE DOWNWIND OF DESERT (km)
SOIL CONDITION
1
0
DRY
8
2.0
MOIST
19
5.5
WET

(1)



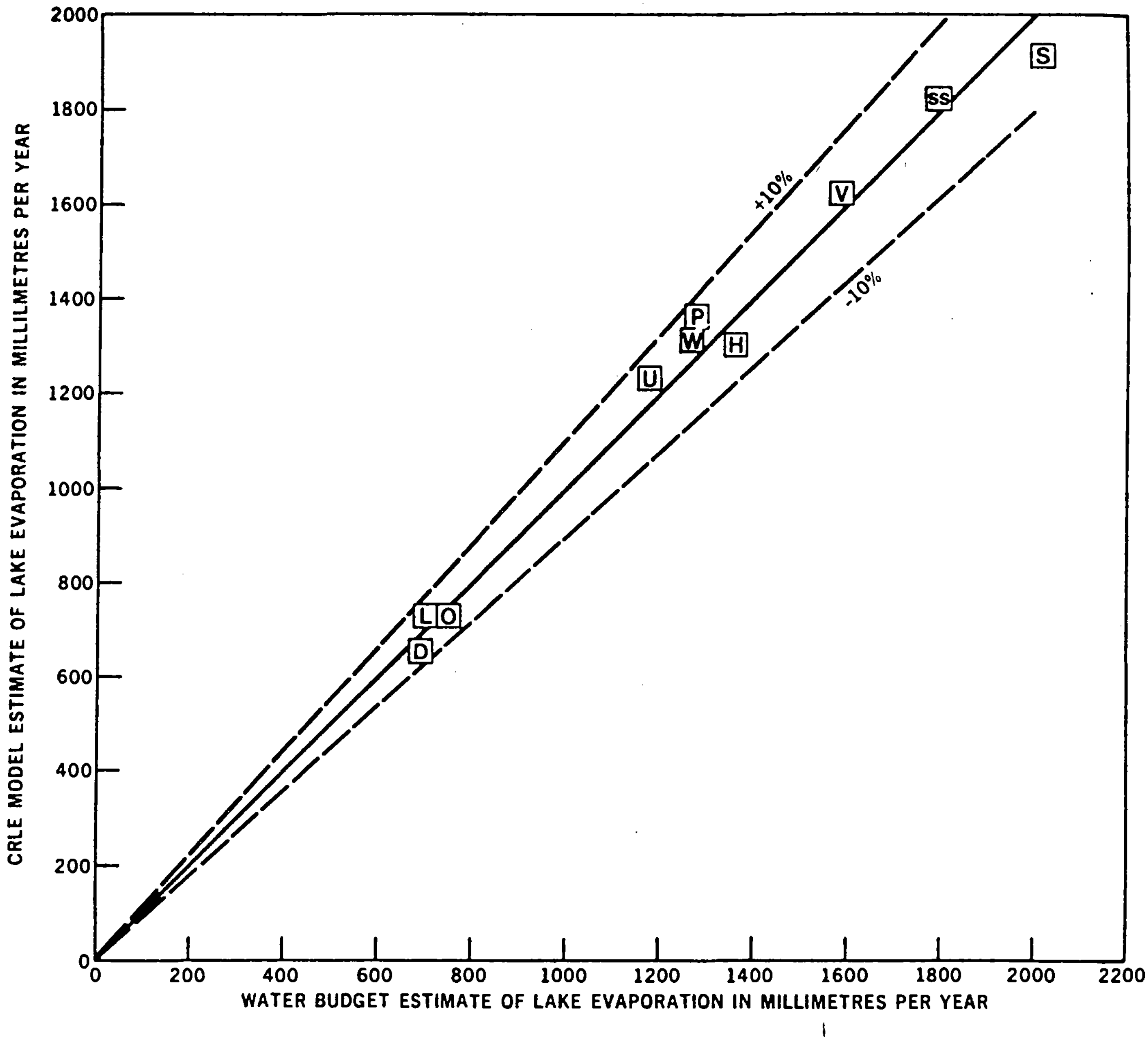
$2E_{TW}$ = DRY ENVIRONMENT POTENTIAL EVAPOTRANSPIRATION
E_{TP} = POTENTIAL EVAPOTRANSPIRATION
E_{TW} = WET ENVIRONMENT AREAL EVAPOTRANSPIRATION
E_T = AREAL EVAPOTRANSPIRATION
E_{TW}
EVAPOTRANSPIRATION
WATER SUPPLY TO SOIL-PLANT SURFACES OF AREA

(2)



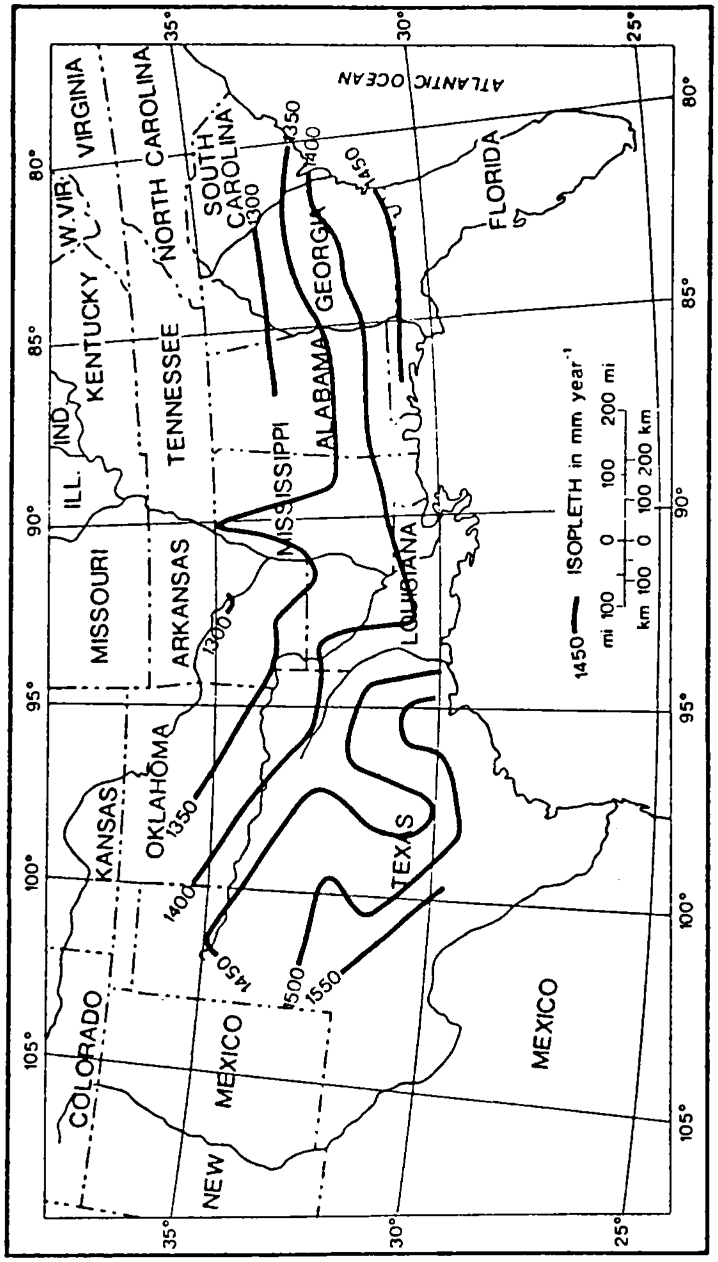
2E_W = DRY ENVIRONMENT POTENTIAL EVAPORATION
E_P = POTENTIAL EVAPORATION IN LAND ENVIRONMENT
E_W = LAKE EVAPORATION
EVAPORATION
E_W
WATER SUPPLY TO SOIL-PLANT SURFACES OF LAND ENVIRONMENT

3

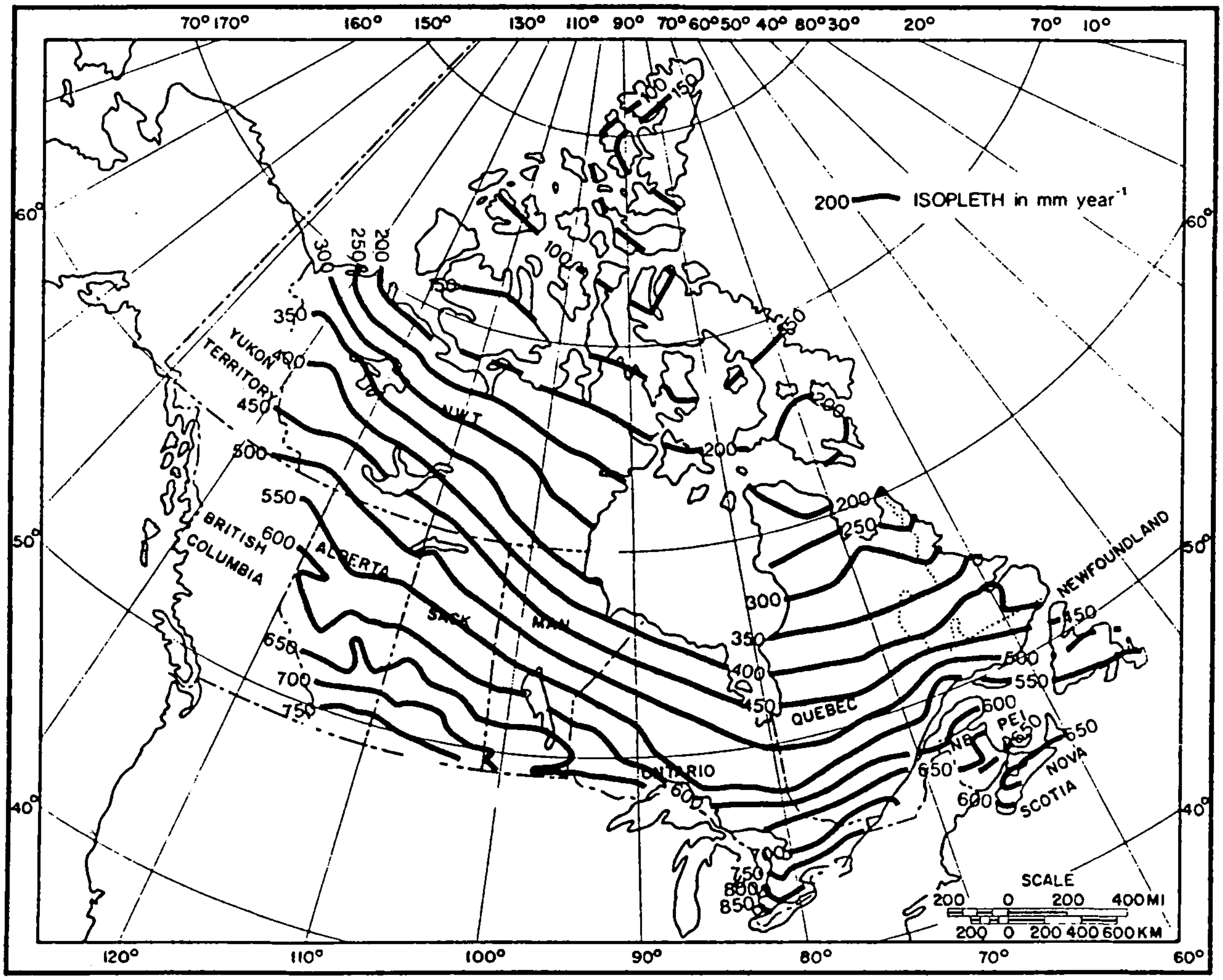


2000
1800
1600
1400
1200
1000
800
600
400
200
0
0
200
400
600
800
1000
1200
1400
1600
1800
2000
2200
CRLE MODEL ESTIMATE OF LAKE EVAPORATION IN MILLILMETRES PER YEAR
WATER BUDGET ESTIMATE OF LAKE EVAPORATION IN MILLIMETRES PER YEAR
+10%
-10%
S
SS
V
P
W
H
U
L
O
D

4

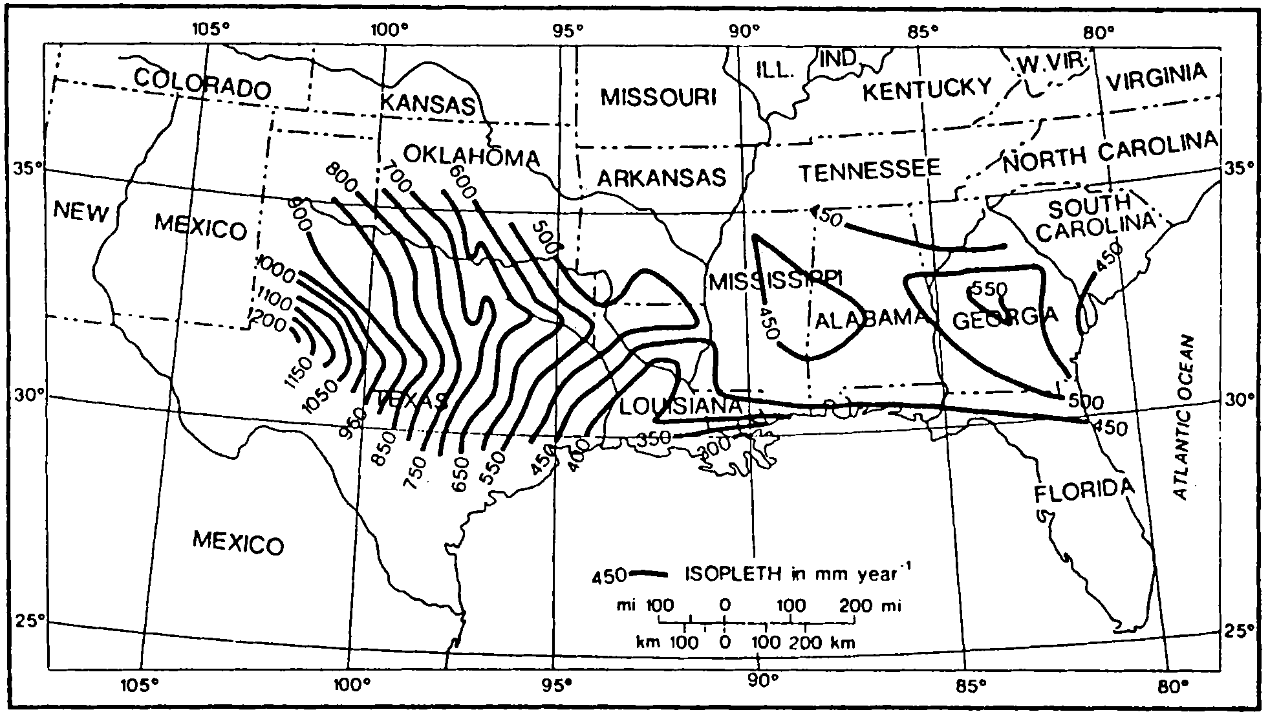


COLORADO
KANSAS
OKLAHOMA
NEW MEXICO
TEXAS
MISSOURI
ARKANSAS
LOUISIANA
ILL.
IND.
KENTUCKY
TENNESSEE
MISSISSIPPI
ALABAMA
GEORGIA
FLORIDA
SOUTH CAROLINA
NORTH CAROLINA
W.VIR.
VIRGINIA
MEXICO
ATLANTIC OCEAN
1300
1350
1400
1450
1500
1550
1450 — ISOPLETH in mm year⁻¹
mi 100 0 100 200 mi
km 100 0 100 200 km
105°
100°
95°
90°
85°
80°
35°
30°
25°

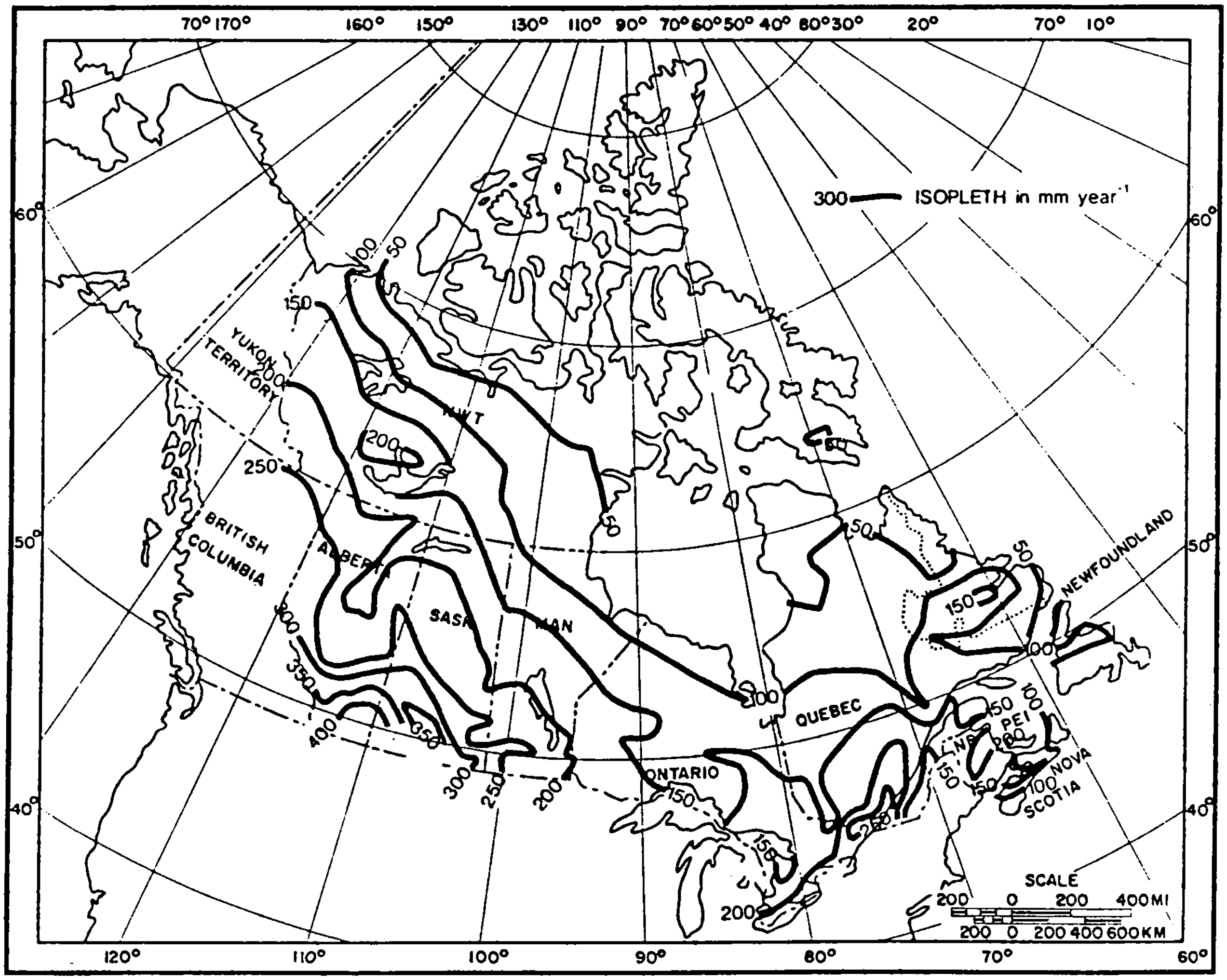


200 ISOPLETH in mm year⁻¹
YUKON TERRITORY
N.W.T.
BRITISH COLUMBIA
ALBERTA
SASK
MAN
ONTARIO
QUEBEC
NEWFOUNDLAND
NB
PEI
NOVA SCOTIA
SCALE
200 0 200 400 MI
200 0 200 400 600 KM

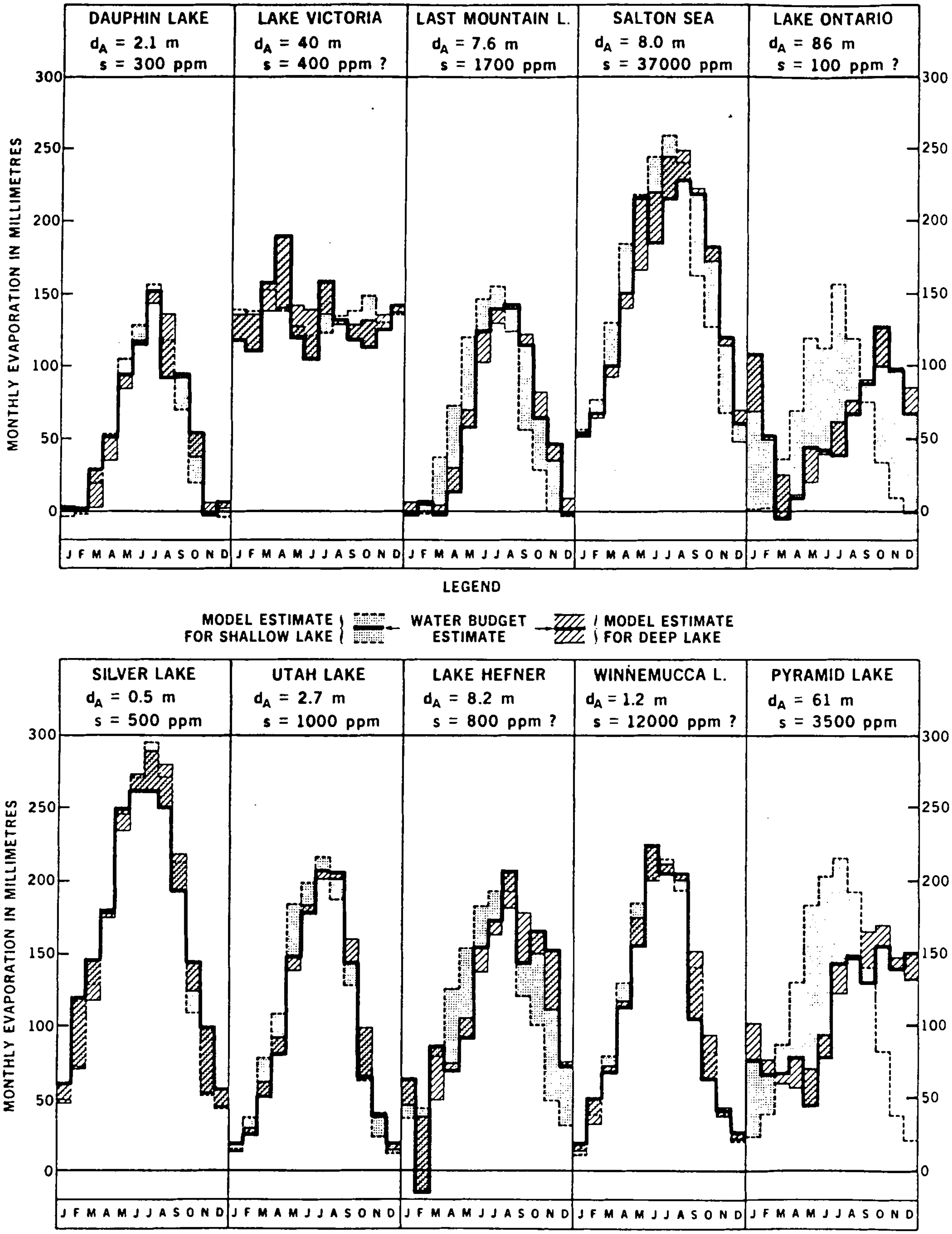
6



COLORADO
KANSAS
MISSOURI
ILL.
IND.
KENTUCKY
W.VIR.
VIRGINIA
NEW MEXICO
OKLAHOMA
ARKANSAS
TENNESSEE
NORTH CAROLINA
SOUTH CAROLINA
MISSISSIPPI
ALABAMA
GEORGIA
TEXAS
LOUISIANA
FLORIDA
MEXICO
ATLANTIC OCEAN
450 ISOPLETH in mm year⁻¹
mi 100 0 100 200 mi
km 100 0 100 200 km


300 ISOPLETH in mm year⁻¹
YUKON TERRITORY
NWT
BRITISH COLUMBIA
ALBERTA
SASK
MAN
ONTARIO
QUEBEC
NEWFOUNDLAND
NB
PEI
NOVA SCOTIA
SCALE
200 0 200 400 MI
200 0 200 400 600 KM

8


DAUPHIN LAKE
d_A = 2.1 m
s = 300 ppm
LAKE VICTORIA
d_A = 40 m
s = 400 ppm ?
LAST MOUNTAIN L.
d_A = 7.6 m
s = 1700 ppm
SALTON SEA
d_A = 8.0 m
s = 37000 ppm
LAKE ONTARIO
d_A = 86 m
s = 100 ppm ?
MONTHLY EVAPORATION IN MILLIMETRES
0
50
100
150
200
250
300
J F M A M J J A S O N D
LEGEND
MODEL ESTIMATE FOR SHALLOW LAKE
WATER BUDGET ESTIMATE
MODEL ESTIMATE FOR DEEP LAKE
SILVER LAKE
d_A = 0.5 m
s = 500 ppm
UTAH LAKE
d_A = 2.7 m
s = 1000 ppm
LAKE HEFNER
d_A = 8.2 m
s = 800 ppm ?
WINNEMUCCA L.
d_A = 1.2 m
s = 12000 ppm ?
PYRAMID LAKE
d_A = 61 m
s = 3500 ppm
MONTHLY EVAPORATION IN MILLIMETRES
J F M A M J J A S O N D

OPERATIONAL ESTIMATES OF AREAL EVAPOTRANSPIRATION AND THEIR SIGNIFICANCE TO THE SCIENCE AND PRACTICE OF HYDROLOGY

IV RELIABILITY, PRACTICALITY AND POTENTIAL IMPACT

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Ottawa, Canada K1A 0E7

ABSTRACT

Morton, F.I., 1982d. Operational estimates of areal evapotranspiration and their significance to the science and practice of hydrology: IV Reliability, practicality and potential impact.

The reliability of the independent operational estimates of areal evapotranspiration produced by the latest version of the complementary relationship models is tested with comparable long-term water budget estimates for 143 river basins in North America, Africa, Ireland, Australia and New Zealand. Their practicality and potential impact are demonstrated by examples which show how the availability of such estimates can revitalize the science and practice of hydrology by providing a reliable basis for detailed water balance studies; for further research on the development of causal models; for hydrological, agricultural and fire hazard forecasts; for detecting the development of errors in hydrometeorological records; for detecting and monitoring the effects of land use changes; for explaining hydrological anomalies; and for other better known applications. The use of the complementary relationship to provide reliable estimates of lake evaporation from routine climatological observations in the land environment is described in a subsequent companion paper.

INTRODUCTION

Evapotranspiration remains one of the most neglected and intractable problems in hydrology despite widespread recognition that it is a much larger proportion of precipitation than runoff in most populated parts of the world; that its accumulated effects on water stored in the soil, swamps and lakes control the basin response to precipitation events; and that it is the component of the hydrologic cycle most directly influenced by land use or other basin changes. Direct measurements, such as those provided by lysimeters, the eddy flux technique or the Bowen-ratio technique, give point values, require constant attendance by skilled personnel and are based on unverified assumptions. The water budget method provides good estimates for several years or more but requires excessive instrumentation and manpower for shorter term estimates. A companion paper (Morton, 1982a) presents a critical review indicating that the current conceptual modeling techniques, as exemplified in the SACREMENTO WATERSHED MODELING SYSTEM, are based on assumptions about the soil, the vegetation and the atmosphere that are incompatible with published evidence and that causal modeling techniques that take into account the complex processes and interactions in the soil-plant-atmosphere continuum are not expected to have practical applications in the next three decades. The review also indicates that the models based on the complementary relationship between potential and actual areal evapotranspiration can do much to fill the gap until such time as the causal models become practicable.

The complementary relationship is:

$$E_T + E_{TP} = 2\ E_{TW} \tag{1}$$

in which E_T is the areal evapotranspiration, the actual evapotranspiration from an area so large that the effects of upwind boundary transitions, such as those shown later in Fig. 1, are negligible; E_{TP} is the potential evapotranspiration, as estimated from a solution of the vapour transfer and energy balance equations, representing the evapotranspiration that would occur from a hypothetical moist surface with radiation absorption and vapour transfer characteristics similar to those of the area and so small that the effects of the evapotranspiration on the overpassing air would be negligible; and E_{TW} is the wet environment areal evapotranspiration, the evapotranspiration that would occur if the soil-plant surfaces of the area were saturated and there were no limitations on the availability of water.

The complementary relationship is produced by small-scale advection of heat and water vapour in the lower atmosphere as exemplified in the well-known oasis effect. These phenomena have been demonstrated graphically by Davenport and Hudson (1967). They measured the variation in evaporation across a series of irrigated cotton and unirrigated fallow fields in the Sudan Gezira, using fiberglass dishes with black painted wells 113 mm in diameter and 36 mm in depth. The dish evaporation observations provide a reflection of the potential evapotranspiration that is distorted somewhat by the difference between the radiation absorption and roughness characteristics of wet vegetation and water surfaces . The passage of air from the upwind desert (and/or the upwind unirrigated fallow fields) over the irrigated cotton caused the dish evaporation to decrease rapidly and approach a low constant value within 300 m - the width of the fields. Furthermore, as the air passed from irrigated cotton across unirrigated fallow the dish evaporation increased rapidly and approached a high constant value within 300 m. Fig. 1 shows the variation of dish evaporation across three irrigated fields on December 27, 1963. The description of the soil condition is presumably based on visual inspection and is related to the number of days since the last irrigation. Thus it is reasonable to assume that the actual evapotranspiration is greater in the "moist" field than in the "dry" field and that it is somewhere near its maximum possible value in the "wet" field.

At the upwind edge of the irrigated fields, where the dish evaporation decreases rapidly, the hot dry air from the desert or the unirrigated fallow loses heat and gains vapour from contact with evaporating and transpiring surfaces. Downwind from the transitional zone, where the dish evaporation approaches a low constant value, the effects of the evapotranspiration on the temperature and humidity of the overpassing air are well developed and approaching equilibrium. (Decreases in temperature and increases in humidity as the air moved across the irrigated cotton were observed at one site. The vapour pressure appeared to attain equilibrium values within 300 m but the temperatures were still decreasing, possibly because the observations were made above the level of the crop and dishes.) As it is only in an equilibrium zone that the term areal evapotranspiration has any meaning, the minimum size of an area for which the term is applicable is one in which the edge effects in the upwind transitional zone become insignificant.

The ratio of daily dish evaporation at the downwind edges of the individual irrigated cotton fields to that at the upwind edge of the irrigated area is 0.69 for the field with "dry" soil, 0.60 for the field with "moist" soil and 0.53 for the field with "wet" soil. This provides good evidence that the dish evaporation, and presumably potential evapotranspiration, respond negatively to the changes in areal evapotranspiration induced by changes in the availability of water. The ratio for the "wet" field provides some evidence that the negative response is also complementary since eq. (1) predicts that the ratio of potential evapotranspiration in a wet environment to potential evapotranspiration in a dry environment is 0.50.

The conceptual and empirical basis for the complementary relationship is presented in a companion paper (Morton, 1982b). Although there is no "proof" for its validity, it is compatible with all current theoretical knowledge and reliable empirical evidence. Such a claim cannot be matched by any concept used in hydrology with the exception of the law of conservation of mass, the law of conservation of energy and the approximations to the law of conservation of momentum represented by the Darcy formula and modifications to the Chezy formula.

The chief advantage of the complementary relationship is that it permits the areal evapotranspiration, a product of complex processes and interactions in the soil-plant-atmosphere continuum, to be estimated by its effects on the routine climatological observations that are used to compute potential evapotranspiration. The detailed formulation of the latest complementary relationship areal evapotranspiration (CRAE) model is presented in a companion paper (Morton, 1982c). The required climatological inputs are screen level values of the average dew point temperature (or vapour pressure), the average of the maximum and minimum temperatures, and the average sunshine duration (or global radiation) for periods of from five days to a month. The required station characteristics are the latitude, the altitude and a rough estimate of long-term average annual precipitation.

The CRAE models are beyond the capacity of slide rules or limited-memory calculators. However the latest version has been documented in FORTRAN and in RPN for the Hewlett-Packard HP-67 hand-held calculator (Morton, Goard and Piwowar, 1980). A later RPN program for the Hewlett-Packard HP-41C hand-held calculator is documented and can be made available on request. The greater storage capacity of the HP 41-C eliminates the need for using extra program cards during individual computations and permits the full range of options.

Because the models avoid the complexities of the soil-plant system they require no local optimization of coefficients and are, therefore, falsifiable. This means that they can be tested rigorously so that errors in the associated assumptions and relationships can be detected and corrected by progressive testing over an ever-widening range of environments. Such a methodology uses the entire world as a laboratory and requires that a correction made to obtain agreement between model and river basin water budget estimates in one environment must be applicable without modification in all other environments. A comparison of the estimates resulting from the most recent application of this methodology with comparable long-term water budget estimates for 143 river basins in North America, Africa, Ireland, Australia and New Zealand is presented herein.

Although it is generally recognized that evapotranspiration is an important aspect of hydrology, the recognition is theoretical. On a practical basis, most hydrologists and engineers fail to understand how research and education in hydrology have been confined to a few narrow problems by the lack of reliable operational estimates of evapotranspiration. Therefore it is extremely difficult for them to visualize the opportunities that arise when such estimates become available. For this reason, much of this paper is devoted to examples which demonstrate the potential impact of operational estimates of evapotranspiration on the science and practice of hydrology.

TEST PROCEDURE AND RESULTS

The CRAE models are based on the complementary relationship between potential and areal evapotranspiration. A conceptual rationalization, a schematic model of the lower atmosphere and a review of available theoretical knowledge and reliable empirical evidence (Morton, 1982b) have indicated that the concept reflects reality. Furthermore there is no documented evidence that the complementary relationship contravenes current knowledge of the underlying processes. "owever there is bound to be doubt that such a simple relationship can account for the complex interactions between the surface and the overpassing air. For this reason it is essential that the model be subjected to a rigorous testing procedure.

Ideally the model should be tested by comparing monthly estimates of areal evapotranspiration with monthly estimates of evapotranspiration derived from river basin water budgets. This requires knowledge of changes in the amount of water stored in the soil, rock, lakes, and snow cover of the basin. Unfortunately, such changes can be greater than the evapotranspiration and there is no satisfactory way to measure them. The only way that the problem can be overcome is to increase the comparison period to a number of years so that the changes in storage become a very small part of the total water budget. With the reasonable assumption that groundwater gains or losses are small, the water budget evapotranspiration is then approximately equal to precipitation less runoff.

The use of long comparison periods does not permit direct evaluation of the monthly estimates of areal evapotranspiration. However, the model was calibrated with monthly data under particularly stringent conditions, i.e., in arid areas where the dry environment potential evapotranspiration and the potential evapotranspiration are very large and the areal evapotranspiration is a small residual. Furthermore, the model can be applied, without the need for local optimization of coefficients, to river basins with different climates, vegetation covers, soils and landforms. Good agreement over a wide range of environments between long-term areal evapotranspiration and the corresponding values of precipitation less runoff should provide reasonable assurance that the seasonal distributions are realistic.

Fig. 2 is a map of the southeastern U.S.A. showing average annual areal evapotranspiration for the five years ending September 30, 1965. The map is based on the accumulation of monthly model estimates for the 37 climatological stations in the area that report both air and dew-point temperatures. For many of the stations it was necessary to estimate sunshine duration from averages for the two nearest recording stations. The isopleths were plotted by linear interpolation between the stations.

The isopleths in western Texas do not take into account the estimate of average annual areal evapotranspiration at Lubbock which was 540 mm year^{-1} or 100 mm year^{-1} higher than the average annual precipitation. Since this apparent error can be explained by large-scale irrigation from groundwater, it demonstrates how an apparent limitation to the widespread applicability of the CRAE models can be useful in detecting and monitoring the hydrometeorological effects of human activities.

Fig. 3 shows the average annual areal evapotranspiration over much of Canada for the five years ending December 31, 1969. It was prepared in the same way as Fig. 2, but with much more frequent use of two- or three-station averages for the sunshine duration. The isopleths were interpolated between values for the 153 climatological stations in the area that report both air and dew-point temperatures. They were not extended west of the Pacific Divide because large altitude differences would make the results unrepresentative.

Some features became evident in the preparation of Fig. 3 that bear on the limitations and potential applications of the CRAE models. Examples are:

(1) The anomalously high values of areal evapotranspiration near the east coast. These are particularly noticeable to the north of the Gulf of St. Lawrence where the isopleths are controlled by the areal

evapotranspiration estimates at Baie Comeau Airport (approximately 1500 m from the coast) and Sept Iles Airport (approximately 700 m from the coast). They provide some indication that advection of heat and water vapour in the lower layers of the atmosphere can affect the areal evapotranspiration estimates at a distance of more than 300 m from a high-latitude coastline.

(2) The difference between the areal evapotranspiration estimates at Edmonton International Airport (420 mm year^{-1}) and at Edmonton Industrial Airport (358 mm year^{-1}). This clearly reflects the effects of urbanization since the former is surrounded by agricultural land and the latter is surrounded by the city.

The model was tested by comparing the areal evapotranspiration for 115 river basins, as interpolated from Figs. 2 and 3, with the precipitation less runoff values for the corresponding 5-year periods. The locations of the hydrometric stations are shown in Figs. 2 and 3. The precipitation for the river basins was estimated from isohyetal maps by planimetry or by inspection. The runoff was estimated by converting the recorded flows (adjusted for reported diversions) to depth over the drainage area.

Similar comparisons have been made for 28 river basins in the vicinity of specific climatological stations in Canada, U.S.A., Ireland, Africa, Australia and New Zealand. Details of these comparisons are summarized in Table I.

Fig. 4 shows the CRAE model estimates of areal evapotranspiration for 143 river basins plotted against the corresponding precipitation less runoff values. The mean absolute deviation of the CRAE model estimates from the line of equality is 19.6 mm year^{-1} or 3.4 percent. The maximum deviations are 52 mm year^{-1} and 9.9 percent. As the deviations include error in the precipitation and runoff estimates in addition to the effects of basin storage changes and groundwater gains or losses, they are of little use in evaluating error in the model except to conclude that it is small.

The river basins used to test the model represent a extremely wide range of climate, vegetation, soil and topography. The environmental diversity, the calibration technique, and the good fit of average annual data in Fig. 4 combine to provide assurance that the monthly estimates of areal evaporation are realistic.

LIMITATIONS

As demonstrated in the preceding section, the CRAE model presented in a companion paper (Morton, 1982c) can provide reliable operational estimates of areal evapotranspiration. However it does have the following limitations:

(1) It requires accurate humidity data and these have depended on frequent observations by skilled personnel. This is one of the more

serious limitations to the use of the CRAE models at the present time. However the Humicell, a simple device developed by the Saskatchewan Research Council, now provides a convenient and reliable alternative. According to Langham (1969) the instrument provides integrated vapour pressures within ± 2 percent for periods exceeding three days. Other more convenient instrumentation is under development.

(2) It cannot be used for short time intervals because of subsurface heat-storage changes and because of the lag times associated with the change in storage of heat and water vapour in the atmospheric boundary layer after changes in surface conditions or the passage of frontal systems. There is every probability that the time periods could be shortened to five days but for intervals of three days or less the results would always be suspect. This limitation has little significance in hydrological applications because it does not matter much whether the daily evapotranspiration is 3 mm or 6 mm as long as the accumulated values for the week or for a longer period are reliable. The use of short time periods is discussed in the preceding section.

(3) It cannot be used near sharp environmental discontinuities, such as a high-latitude coastline or the edge of an oasis, because of advection of heat and water vapour in the lower layers of the atmosphere. The data of Davenport and Hudson (1967) indicate that

the effects of such advections can decrease to near zero within 300 m but this finding may not be generally applicable.

(4) It requires input from a climatological station whose surroundings are representative of the area of interest. Some advantages that are associated with this and the preceding limitation are presented subsequently.

(5) It cannot be used to predict the effects of natural or man-made changes since it neither uses nor requires knowledge of the soil-vegetation system. This is the most serious long-term limitation to the application of the CRAE models.

IMPACT ON SCIENCE AND PRACTICE OF HYDROLOGY

It is generally recognized that evapotranspiration is a much larger proportion of the precipitation than runoff in most populated parts of the world; that the accumulated effects of evapotranspiration on the water stored in soil, swamps, lakes and snow pack control the catchment response to precipitation events; and that evapotranspiration is the component of the hydrological cycle most directly influenced by land use and climatic changes. What is not recognized is the implication that without realistic independent estimates of evapotranspiration there is no chance for significant improvements in hydrological knowledge or for the transformation of hydrology from a descriptive to a predictive science. This failure to recognize the significance of evapotranspiration in the hydrological cycle is, to a large extent, responsible for the stagnation of the science during the past three decades and the concentration of research on marginal improvements to existing techniques which in themselves are open to doubt. It explains why so much time and effort have been expended on superficial improvements to time distributions of runoff during the half-century since the development of the unit hydrograph and why so little time and effort have been expended on those processes and interactions that control the volume of runoff. Furthermore it provides a partial explanation for the contrast between the amount of research that is directed toward statistically-based predictions of hydrologic regimes for periods of several decades or more and the amount of research that is directed toward detection of the effects of land use or climatic changes which, if significant, could destroy the statistical basis for the predictions. The substitution of hydraulics and mathematics for hydrology is apparently based on the perception that reliable operational estimates of evapotranspiration are impracticable. If so, the demonstrated ability of the current CRAE model to provide realistic estimates of areal evapotranspiration from routine climatological observations anywhere in the world should have a major impact on the science and practice of hydrology. Some examples, for the most part taken from the river basins listed in Table I, follow.

Water Balance Analysis:

With reliable independent estimates of evapotranspiration it is possible to abstract much hydrological information from the law of conservation of mass or the water balance. The water balance equation for a drainage basin states that the precipitation equals the sum of the runoff, evapotranspiration and change in storage and in the past it has been possible to routinely observe or estimate only two of these four quantities, i.e. the precipitation and runoff. The measurement of changes in storage on a weekly, monthly, seasonal or even an annual basis is too expensive to contemplate, except in a research situation, because of the heterogeneity of topography, soils, vegetation and snow cover in most drainage basins. Therefore the water balance has been useful only to provide some idea of what happens over a period of many years when the change in storage becomes relatively insignificant and the evapotranspiration can be assumed equal to the precipitation less runoff. However, with the operational estimates of evapotranspiration provided by the current CRAE model, the water balance can be used to estimate and study the changes in storage.

Fig. 5 provides some examples of this capability. It is a graphical representation of the mean seasonal water balances for the basins of Creighton Tributary to Bad Lake near Rosetown, Saskatchewan; Upper Rainy River on the Ontario-Minnesota Boundary; Perch Lake Inlets 4 and 5 near Chalk River, Ontario; Big Fossil Creek near Fort Worth, Texas;

North River near Des Moines, Iowa; and W-5 Watershed of Sleepers River near Danville, Vermont; The seasonal water balances in Fig. 5 are based on monthly mean values of observed precipitation, observed runoff and the CRAE model estimates of evapotranspiration for five or more integral years. The top solid line is the accumulated precipitation since the beginning of September; the middle long-dash line is the accumulated precipitation less evapotranspiration since the beginning of September; and the bottom short-dash line is the accumulated precipitation less evapotranspiration less runoff since the beginning of September. Thus the upper line represents a mass curve of runoff; the vertical distances in the horizontally hatched area represent the ordinates of a mass curve of evapotranspiration; the vertical distances in the stippled area represent the ordinates of a mass curve of runoff; and the vertical distances in the vertically hatched area represent the ordinates of a mass curve of storage changes. The vertical distances on the right hand edge represent the annual totals for an average year. The very small annual totals for the storage changes may represent an average of the storage change for five or more integral years or may represent the effects of errors in estimating or measuring the other components of the water balance.

The most obvious use for the kind of analysis presented in Fig. 5 is to provide general comparisons of seasonal water balances for river basins in differing environments. However it can be used to answer more specific questions. For example, why is the accumulated storage change for Upper Rainy River, a large basin in the Precambrian Shield with myriads of lakes, only 60 percent of that for Perch Lake Inlets 4 and 5, a small basin in the Precambrian Shield with no lakes. Fig. 5 provides the following partial explanations:

(1) The winter runoff for the basin of Upper Rainy River is much greater than the winter runoff for the basin of Perch Lake Inlets 4 and 5 and this provides an indication that the lake storage is being depleted at the same time as the snowpack storage is being augmented. In fact it is apparent that the accumulation of snow on the lake ice would tend to squeeze an equivalent amount of water out of the lake.

(2) The spring runoff for the basin of the Upper Rainy River is much less than the spring runoff for the basin of Perch Lake Inlets 4 and 5. This indicates that the lake storage is increasing at the same time as the snowpack storage is decreasing.

(3) The mass curve of precipitation for the Upper Rainy basin is typical of a continental climate where the precipitation is high during the summer when evapotranspiration is high and low during the winter when evapotranspiration is low and this reduces the need to draw on stored water. On the other hand the precipitation mass curve for the basin of the Perch Lake Inlets shows little seasonal variability.

The North River near Des Moines, Iowa provides the most extreme example in Fig. 5 of a continental type of precipitation mass curve. It demonstrates how heavy summer rainfall can reduce the need for large reductions in soil moisture storage during an average year. However the storage depletion during the dryest of the five years was almost three times greater than that for the average year.

Fig. 6 presents the average water balances for two basins in the Sahel region of northern Cameroon. It has been prepared in the same way as Fig. 5 except that the mass curves begin at the beginning of July. The data for the Kallaio basin were published by Nouvelot (1973) while the data for the basin at Station 7 in the Sanguéré were published by Casenave (1978). The precipitation mass curves are flat during the long rainless season.

Fig. 6 contains a hydrological anomaly. Although the two basins are separated by less than one and a half degrees of latitude and half a degree of longitude, the runoff from the basin with the lower precipitation exceeds the runoff from the basin with the higher precipitation by approximately 150 percent. Until now this kind of anomaly could be explained only by assuming that there is groundwater runoff that flows out of the basin without being measured at the streamflow gauging site. Casenave (1978) studied the water balance of the Sanguéré basins using this assumption and concluded that the unmeasured groundwater outflow, which could be exploited without depleting the reserves, was approximately 15 percent of the precipitation and approximately 150 percent higher than the measured runoff from the basin at Station 7. However, with reliable independent estimates of evapotranspiration, there is no need to make such an assumption. Thus Fig. 6 shows that the evapotranspiration from the Sanguéré basin is much higher than the evapotranspiration from the Kallaio basin and this difference explains why the higher precipitation produces a much lower runoff. According to this explanation the large exploitable groundwater resource does not exist.

The model estimates indicate that the evapotranspiration from the Sanguéré basin exceeds the evapotranspiration from the Kallaio basin by 348 mm/year. This excess may include an error of 54 mm/year, the difference between the annual values of accumulated changes in basin storage, and does include the effects of the 206 mm/year difference in precipitation. The remainder of 88 mm/year is attributable to the way that Sanguéré basin provides conditions favouring evapotranspiration at the expense of runoff in comparison with the Kallaio basin. In order of increasing plausibility the possible reasons for this difference are:

(1) The surface of the Sanguéré basin may be more pervious than the Kallaio basin so that more water infiltrates into soil moisture storage. This could account for the difference between the values of maximum storage accumulation but is not consistent with the low runoff from the Kallaio basin during the first three months of the wet season.

(2) The net radiation is higher for the Sanguéré basin than for the Kallaio basin and the increased availability of energy for evapotranspiration may have increased the availability of water for evapotranspiration.

(3) The water table in the Sanguéré basin may be close enough to the surface to supply water for significant soil evaporation during the long rainless season. However there is some doubt that the saturated zone would continue to supply water for evapotranspiration in excess of rainfall during the first three months of the wet season as shown in Fig. 6.

(4) Some vegetation in the Sanguéré basin may be deeper rooted than any vegetation in the Kallaio basin and may therefore draw water from greater depths. This is compatible with the tendency in Fig. 6 for the evapotranspiration from the Sanguéré basin to exceed rainfall during the dry season and the first three months of the wet season while the evapotranspiration from the Kallaio basin for the same period remains equal to the rainfall. Note that the change in storage from the end of October to the end of June for the Kallaio basin is negligible whereas for the Sanguéré basin it is approximately equal to the difference in annual evapotranspiration between the two basins less the difference in annual precipitation (i.e. 142 mm/year).

Application to Forecasting:

The only realistic basis for hydrological forecasts is an accurate estimate of the antecedent moisture or the amount of water stored in the soil, rock, snowpack, swamps and lakes of the basin at the beginning of the forecast period. The water stored in snowpacks can be estimated from snow surveys and the water stored in the lakes can be estimated from water level gauges but the financial and human resources required to monitor the water stored in the soil, rock and swamps of even a small basin are too horrendous to contemplate. However the variations in total basin storage can be followed on a real time basis by combining the water balance with independent estimates of areal evapotranspiration. The basin storage is computed by selecting an arbitrary initial value and then accumulating the water balance estimates of storage change, i.e. precipitation (P) less runoff (Q) less areal evapotranspiration (E_T).

Fig. 7 has been prepared to demonstrate how the concept of basin storage can provide a basis for real-time forecasting. The data are for the Turtle River near Mine Centre, a river that flows into Rainy Lake on the Ontario-Minnesota boundary. The areal evapotranspiration was estimated from data at Atikokan, Ontario, the same station used for the Upper Rainy estimates in Table I. The circular points in the upper half of Fig. 7 show the month-to-month variations in basin storage for a 9 year period. The vertical distances in the shaded area under the line joining these points represents the water stored in solid form in the snow cover as estimated from snow surveys, and the line under the shaded area represents the basin storage with the snow cover water content removed. The lack of any significant upward or downward trend over the 9 year period indicates that consistent errors in the precipitation, runoff or areal evapotranspiration are either very small or compensatory.

Equations that relate the runoff for seasonally defined periods of from one to four months to the total basin storage at the beginning of the period (adjusted empirically for any evident time trend) and the precipitation less evapotranspiration during the period usually have regression coefficients that are highly significant. However, when the equations are used in the forecasting mode, using long-term average values of precipitation less evapotranspiration as input, the results are disappointing. Part of the problem is the regression approach with its requirements for fixed seasonal time periods and for many more years of data than there are years of record. Holecek (1982) reports good results in predicting runoff responses to specific precipitation and snowmelt events for Spring Creek (see Table I), using a relationship between effective drainage area and total basin storage. The relationship is highly non-linear and this relates to antoher major problem; i.e. that the total basin storage includes solid storage in the snow, tension storage in the unsaturated soil and gravity storage in the lakes and in the saturated groundwater zone, and that each of these different types of storage have different effects on the runoff. A graphical technique for analyzing these different effects has been presented by Klemes (1982) and is discussed briefly in subsequent paragraphs.

The circular points in the lower half of Fig. 7 are the accumulated values of precipitation (P) less evapotranspiration (E_T) less the long-term average runoff ($\bar{Q}$ = total runoff for 9 years divided by 108 periods) starting from an arbitrary initial value. The vertical distances in the shaded area under the line joining these points represents the water stored in the snow pack as estimated from snow surveys. Thus the line under the shaded area and through the circular points of the snow-free season represents a residual mass curve of the net liquid inputs to basin storage with respect to the long term average runoff. Similarily the line through the triangular points represents the mass curve of outputs from basin storage (Q = runoff) with respect to the long-term average runoff ($\bar{Q}$). The slopes of the mass curves represent the rates of net input or output with respect to Q and the differences between the mass curves of net input and output are the basin storage (without snowcover water content) to a datum specified by the difference in initial values. Note that the prime reason for including $\bar{Q}$ is to keep the mass curves horizontal over the 9 year period.

The difference between the initial values of the net input and the output mass curves has been selected by trial-and-error to aid in discriminating between tension and gravity storage. Thus during most of the 9 years it can be noticed that:

(1) An increase in net input when the net input mass curve is above the output mass curve produces an increase in output because it augments gravity storage.

(2) A decrease in net input when the net input mass curve is above the output mass curve has an insignificant effect on the output because it depletes tension storage.

(3) A decrease in net input when the net input mass curve is below the output mass curve produces a decrease in output because it depletes gravity storage.

(4) An increase in net input when the net input mass curve is below the output mass curve has an insignificant effect on the output because it augments tension storage.

(5) As suggested by Klemes (1982) the difference between the initial values of the two mass curves is approximately equal to the average basin storage (without snow cover water content), as defined by the line in the upper half of Fig. 7, and the arbitrary initial starting point.

Tendencies (1) to (4) inclusive are masked to some extent by surface runoff from small saturated areas near the stream channels, by error in the snow cover water content and by error in the precipitation, evapotranspiration and runoff inputs.

Fig. 8 has been prepared in the same way as Fig. 7. The basin storage and the mass curves are for the Humber River at Elder Mills, a basin that is located a short distance west of Toronto, Ontario (see Table I). The deductions made from Fig. 7 apply also to Fig. 8 although the results for the Humber basin are complicated to some extent by snowmelt during the winter months and by the small but noticeable response of runoff to the fall rains when the output mass curve is higher than the net input mass curve. This latter response, which is believed due to surface runoff from small saturated areas near the stream channels, is most noticeable during the autumn of 1972.

The kind of analysis presented in Figs. 7 and 8 has obvious applications. Thus the amount by which the output mass curve exceeds the net input mass curve represents the amount of storage that must be replenished before the output (runoff) will respond in a significant way to increases in net input (precipitation less evapotranspiration) and the amount by which the input mass curve exceeds the output mass curve represents the amount of storage that must be depleted before the output will respond in a significant way to decreases in net input. Thus the distance between the two mass curves provides an excellent estimate of antecedent moisture which, when combined with average seasonal values of precipitation less evapotranspiration, can provide the basis for real-time forecasts. Applications of this potentiality would be more useful and informative if based on weekly time periods rather than monthly time periods.

The graphical approach has two additional practical advantages. These are: (1) that forecasting can begin on a tentative basis with only a few years of data and, (2) that it is easy to detect and correct for errors in the input data near the time when they occur. Klemes (1982) has provided an example of the latter capability and Fig. 8 provides another. Thus there must have been an error exceeding 50 mm during 1978 because, with the input mass curve higher than the output mass curve during the autumn of 1977 and the autumn of 1978, the output responded to an increased net input during the former period and did not respond to an increased net input during the latter period. A possible explanation was the loss of a key precipitation station during 1978 and its replacement by another some distance away.

Fig. 9 demonstrates the most serious impediment to the use of basin storage for runoff forecasts, i.e. the accumulation of small consistent errors in the precipitation, evapotranspiration or runoff inputs to produce a trend that prevents realistic year-to-year comparisons of basin storage. The trace for the Humber River at Elder Mills is the same as that shown in top half of Fig. 8 and the lack of any significant upward or downward trend in basin storage over the 14 years indicates that consistent errors in the precipitation, runoff and areal evapotranspiration components are either very small or compensatory. Therefore the seasonal and annual variations in basin storage should provide a realistic basis for runoff forecasts. The Conestogo River at/above Drayton is in the snowbelt of southwestern Ontario so that the seasonal variations in basin storage, which includes snowpack storage, are much greater than those for the Humber River. There is a distinct upward trend for the Conestogo basin storage prior to 1972 which could have been caused by an error of less than 4 percent in precipitation, 6 percent in evapotranspiration or 9 percent in runoff. That the error was in the runoff component is evident in the lack of trend in the basin storage during the five years following December, 1972, when the streamflow gauging station was moved upstream out of the backwater from the Conestogo Reservoir.

The sensitivity of basin storage to minor errors in any one of the water balance components is the most serious hindrance to its use in runoff forecasts, agricultural forecasts or fire hazard forecasts. However, its potentialities are so great that the time and effort needed to solve the problem would be well worthwhile. The most obvious and expensive solution would be to make an annual correction that is based on a concentrated late summer survey of the water stored in the soil, aquifers, lakes and swamps of the basin. Further investigations are required for more economic alternatives but at the present time the best approach seems to be the kind of graphical detection and correction suggested in the discussion of Figs. 7 and 8.

Detection of Change:

The discussion of Fig. 9 dealt with the sensitivity of computed basin storage to small consistent errors in measuring or estimating the precipitation, runoff or areal evapotranspiration components of the water balance. Although the negative aspects were stressed there are positive aspects as well. For example, changes in the trend of basin storage that are similar to those shown for the Conestogo basin in Fig. 9 can be used to alert hydrometric or meteorological organizations to the development of possible errors or inconsistencies in their data. Furthermore, they may be used also to detect the effects of changes in land use in the basin.

Fig. 10 shows changes in basin storage over a 14 year period for three rivers in the western outskirts of Toronto. The areal evapotranspiration estimates used in the computation of basin storage are the same for all three basins, being based on temperature and humidity observations at Toronto International Airport. The trace of basin storage for the Humber River at Elder Mills is the same as that shown in Fig. 9 although drawn to a different scale. The basin above Elder Mills is still largely rural and the trace of basin storage is included in Fig. 10 merely to demonstrate that there were no significant changes in the areal evapotranspiration estimates at Toronto International Airport or in the water balance of the basin between 1965 and 1979. The Mimico Creek and Etobicoke Creek basins were still largely rural in 1965 and

1966 although there was development for some distance upstream of the hydrometric stations. However during the 1970s the basin of Mimico reek was transformed completely and the basin of Etobicoke Creek was transformed partially by urban development.

The basin storage for Mimico Creek has no trend prior to 1971. This indicates that the areal evapotranspiration estimated at Toronto International Airport was compatible with the water balance of the basin. After 1971, however, there was a strong negative trend in basin storage. The trend may have been due to consistent errors in the precipitation or runoff components of the water balance but this seems unlikely (the runoff would need to be almost 40 percent too high to account for the trend). A more probable explanation is that urbanization increased the runoff and decreased the evapotranspiration without affecting the evapotranspiration estimated at Toronto International Airport. An undetected reduction in evapotranspiration of approximately 20 percent could account for the trend and this compares with the previously mentioned reduction of 15 percent between the evapotranspiration for a rural area estimated at Edmonton International Airport and the evapotranspiration for an urban area estimated at Edmonton Industrial Airport.

The trace of basin storage for Etobicoke Creek in Fig. 10 is more difficult to explain. Thus the trend remained horizontal until 1970-71 and then started a downward trend similar to that for Mimico

Creek. Unlike Mimico Creek, however, the trend became horizontal again in 1972-73. This means that there were two lengthy periods with balanced water balances separated by a two-year transition period. If the transition was real, i.e. if it was not caused by an error in precipitation or runoff that was later corrected, it may have been due to some permanent reduction in stored water associated with urban pressures, such as the drainage of large swamps or the lowering of the water table. As the decrease in basin storage was approximately 200 millimetres the volume permanently abstracted from storage would have been about 40 million cubic metres.

The trace of basin storage for the Castor River near Ottawa, Ontario, is not shown. It has a similar appearance to that for Mimico Creek in that a strong negative trend developed after about five years with no trend. There is a good chance that the negative trend after 1972 was due to a change in land use in the basin from pasture and hay production to corn cultivation with tile drainage. However, the development of error in the evapotranspiration estimates cannot be ruled out as the humidity inputs during the period of negative trend appear to be too high.

A Hydrological Curiosity:

A comparison of the results shown in Table I for the Waikato River in New Zealand and the James River in Missouri is very interesting from a hydrological point of view. The two basins are roughly equidistant from the equator and there is an approximate equality in both the model and the water budget estimates of basin evapotranspiration. This seems peculiar in the light of a difference of more than 700 mm/year between the precipitation for the Waikato River and the precipitation for the James River. Thus the evapotranspiration, which is roughly the same for both basins, is only 44 percent of the precipitation for the Waikato basin compared with 77 percent for the James basin. The more mountainous nature of the Waikato basin provides a partial explanation because it would favour runoff at the expense of evapotranspiration. However the main reason for the anomaly is that the Waikato basin has a maritime climate with low radiation and a winter precipitation maximum whereas the James basin has a continental climate with high radiation and a summer precipitation maximum. The difference in radiation is of most interest because the precipitation on the James River basin during the six months closest to the summer solstice is only about 80 percent of that for the Waikato River basin despite the difference in precipitation patterns.

Figure 11 is a plot of the monthly mean CRAE model outputs for the Waikato and James basins. The solid line is the areal evapotranspiration, the dashed line is the potential evapotranspiration and the dotted line is the evaporation equivalent of the net radiation that would have occurred with the surface at air temperature. The time scale for the Waikato basin is offset by six months to make the seasonal variations comparable with those of the James basin in the opposite hemisphere. In the wet maritime climate of the Waikato basin the areal evapotranspiration is only slightly lower than the net radiation and potential evapotranspiration whereas in the drier continental climate of the James basin it is well below the net radiation and much below potential evapotranspiration. The approximate equality in distance from the equator simplifies comparison and analysis. According to the complementary relationship, one half of the sum of the potential and areal evapotranspiration is the wet environment areal evapotranspiration, or the energy available for evapotranspiration from a large area, and the difference between the wet environment and the actual areal evapotranspiration is the areal water deficit, or the reduction in areal evapotranspiration caused by lack of water. On an annual basis the energy available for evapotranspiration from the James basin exceeds that for the Waikato basin by 33 percent and this presumably is the most important factor counteracting the effects of the 41 percent difference in annual precipitation. Furthermore, the areal water deficits for the James and Waikato basins were 350 and 90 mm/year respectively so that a precipitation difference of 725 mm/year produced an evapotranspiration difference of 260 mm/year. The latter difference would be larger if both basins had a continental precipitation pattern, which would increase slightly the Waikato basin evapotranspiration; or if both basins had a maritime precipitation pattern, which would decrease significantly the James basin evapotranspiration.

Other Opportunities:

The foregoing discussion has demonstrated how operational estimates of areal evapotranspiration provided by the latest version of the CRAE models can be used to revitalize the science and practice of hydrology. This will be due mainly to the increase in knowledge and understanding that can result from reliable independent estimates of what has been, until now, the largest and most intractable unknown in the water balance. However it will also result from the use of such estimates to provide a reliable basis for hydrological, agricultural and fire hazard forecasts; for detecting the development of errors in hydrometeorological records; and for detecting and monitoring the effects of land use changes. There are other more obvious applications that have not been discussed. Thus independent estimates of evapotranspiration can be used to estimate groundwater recharge and to estimate the long-term average runoff from climatological records in regions where hydrometric records are not available. Furthermore they could be incorporated into watershed modelling systems where they would not only be more reliable than those produced by the conventional techniques but would also permit a more rigorous appraisal of the other model components by substantially reducing the number of coefficients that must now be determined by fudging.

The use of the complementary relationship to provide operational estimates of lake evaporation and net reservoir evapotration from routine climatological observations in the land environment is presented in a companion paper (Morton, 1982e).

Probably the greatest human constraint to the use of the complementary relationship and the CRAE models is that the apparent intractability of the evapotranspiration problem has caused education and research in hydrology to be focused on a few narrow fields of interest so that most engineers and hydrologists, including the author, find it difficult to visualize the sudden expansion in topics that are now open for study and research with a reasonable probability of success.

The major long-term conceptual limitation to the use of complementary relationship models is that the evapotranspiration is estimated from its effects rather than its causes. Therefore, although they can be used to detect and monitor the hydrometeorological effects of climatic or land-use changes, they cannot be used to predict them. To make such predictions would require knowledge of the complex processes and interactions in the soil-plant-atmosphere continuum and it seems unlikely that such knowledge will be available in useable form within the next generation (Morton, 1982a). In this context it would appear that the greatest long-term contribution of the complementary relationship models will be to supply the frame of reference and the reliable estimates of actual evapotranspiration that are needed for research on the processes and interactions and for the development of causal models.

CONCLUSIONS

The conceptual and factual bases for the complementary relationship between potential and areal evapotranspiration and the superiority of the relationship to other hydrometeorological concepts have been documented elsewhere (Morton, 1982a, 1982b). The formulation of a model that uses the complementary relationship to provide operational estimates of areal evapotranspiration from routine climatological observations with no local optimization of coefficients has been documented in another companion paper (Morton, 1982c) and the results of a test procedure that compares model estimates with water budget estimates for river basins in a wide variety of environments are presented herein. The preceding section indicates how the ability to make reliable operational estimates of evapotranspiration can be a big factor in overcoming the current sterility of hydrologic research and practice and in providing the frame of reference and information needed to transform hydrology from a descriptive to a predictive science.

It has been noted that the use of the complementary relationship and the CRAE models is constrained by the failure of hydrologists and engineers to understand how the science and practice of hydrology have been stunted by the lack of reliable operational estimates of evapotranspiration and to visualize how the availability of such estimates can suddenly expand the horizons of hydrology and the range of problems that can be addressed, studied and solved. A related problem is

Morton, F.I., 1982e. Operational estimates of lake evaporation, This issue.

Morton, F.I., Goard, R. and J. Piwowar, 1980. Programs REVAP and WEVAP for estimating areal evapotranspiration and lake evaporation from climatological observations, NHRI Paper No. 12, Inland Waters Directorate, Environment Canada, Ottawa.

Nouvelot, J.F., 1973. Hydrologie des mayos du Nord-Cameroun, Cah. O.R.S.T.O.M., Ser. Hydrol., X(3), pp. 211-301.

TITLES FOR FIGURES

Fig. 1 Comparison of evaporation rates across irrigated cotton fields on December 27, 1963 (Davenport and Hudson, 1967).

Fig. 2 Areal evapotranspiration for the southern United States during the five years ending September 30, 1965

Fig. 3 Areal evapotranspiration for the part of Canada to the east of the Pacific Divide during the five years ending December 31, 1969

Fig. 4 Comparison of model with water budget estimates of areal evapotranspiration for 143 river basins in North America, Africa, Ireland, Australia and New Zealand.

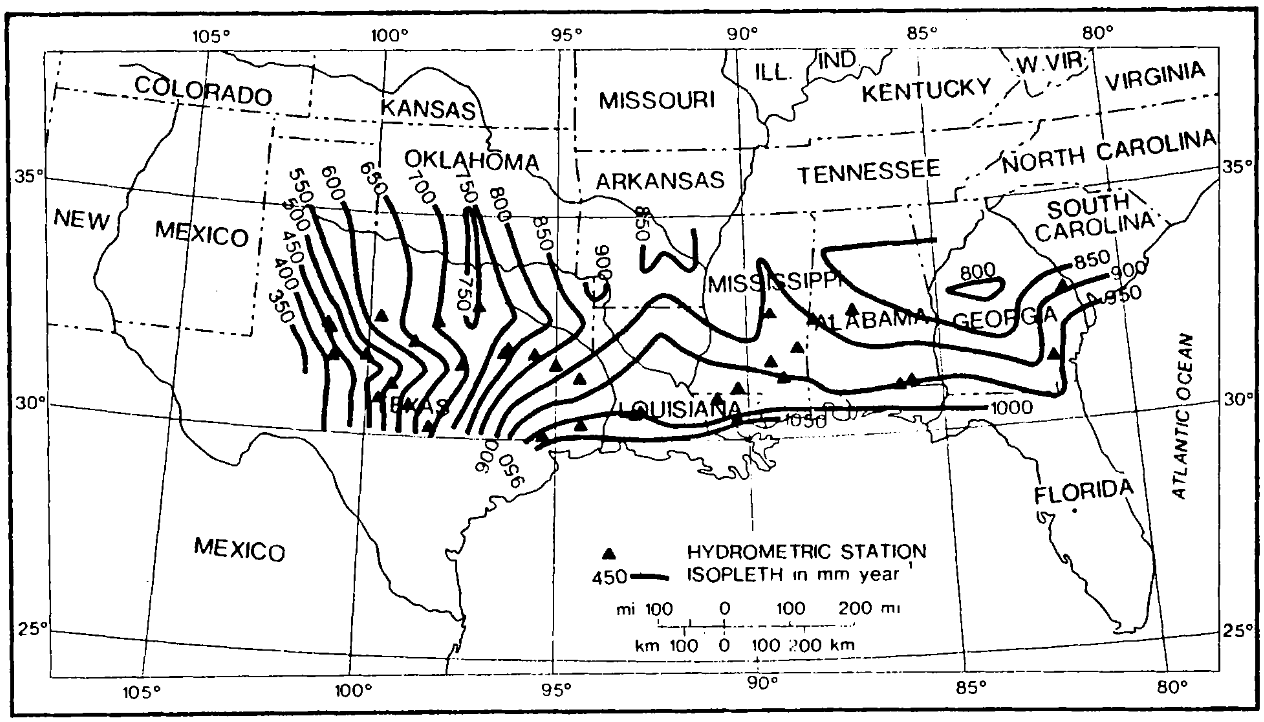
Fig. 5 Seasonal water balance for six basins in North America

Fig. 6 Seasonal water balances for two basins in Africa

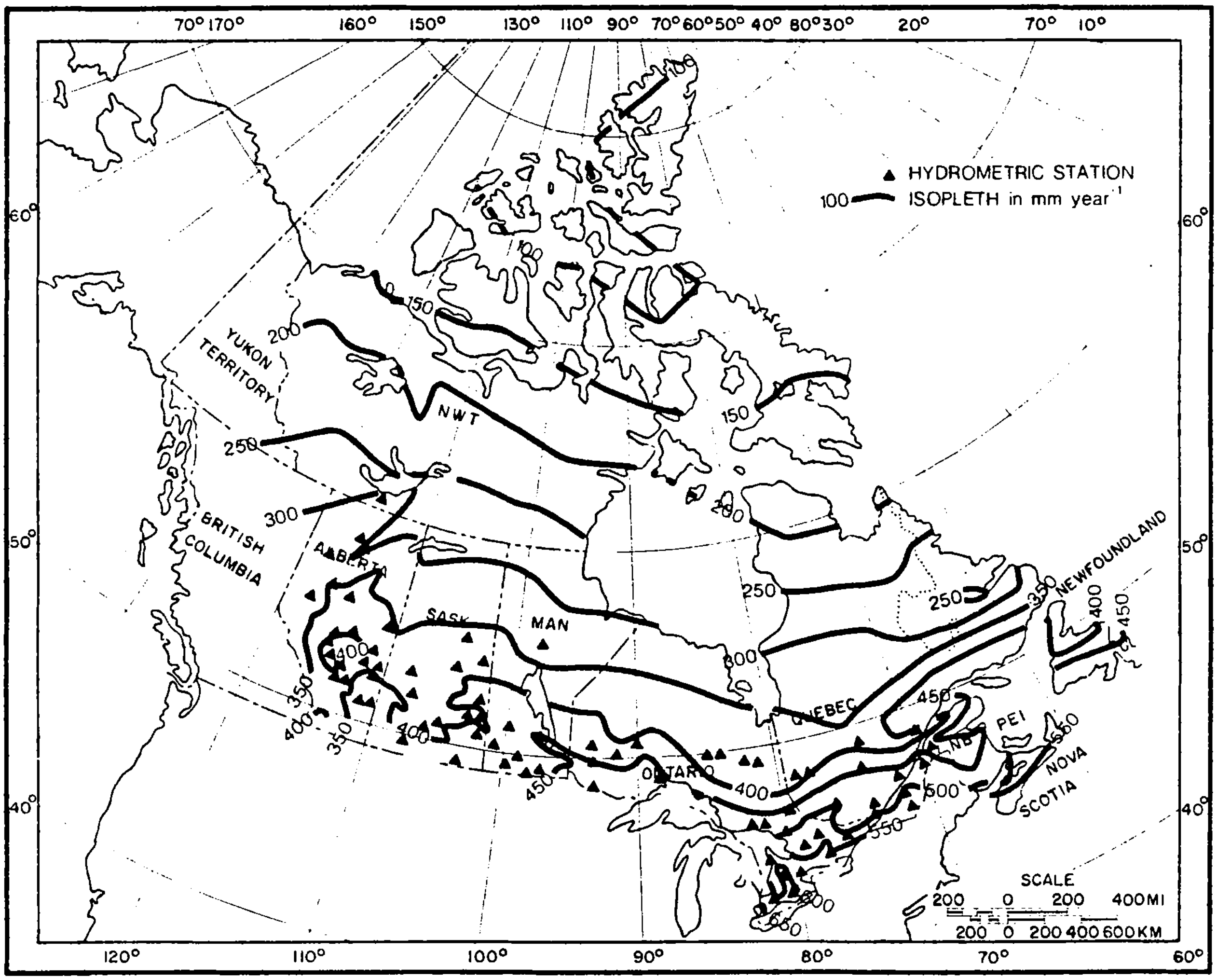
Fig. 7 Basin storage and mass curves of input and output for Turtle River.

Fig. 8 Basin storage and mass curves of input and output for Humber River.

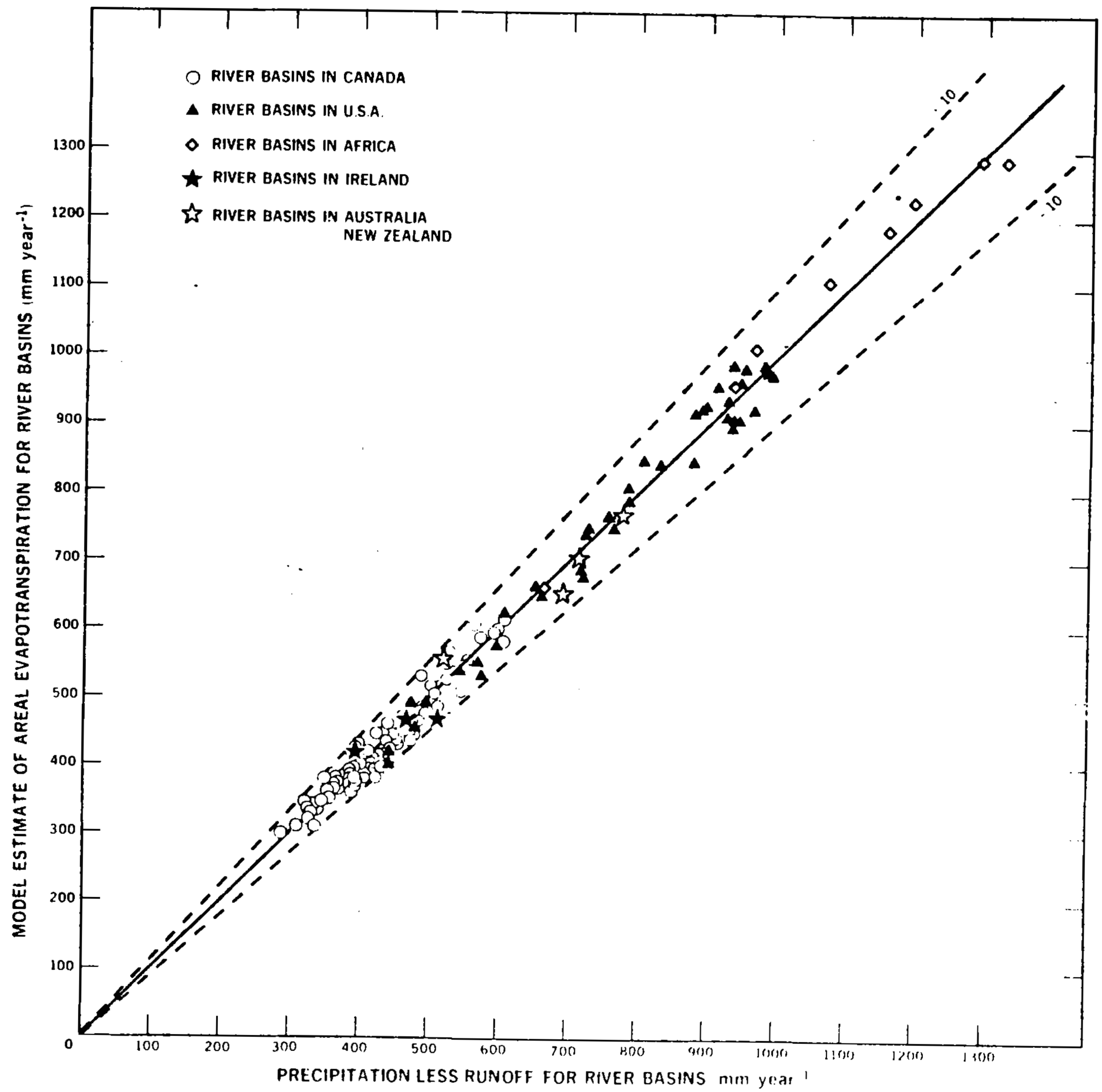
2



COLORADO
KANSAS
MISSOURI
ILL.
IND.
KENTUCKY
W.VIR.
VIRGINIA
OKLAHOMA
ARKANSAS
TENNESSEE
NORTH CAROLINA
SOUTH CAROLINA
NEW MEXICO
MISSISSIPPI
ALABAMA
GEORGIA
TEXAS
LOUISIANA
FLORIDA
MEXICO
ATLANTIC OCEAN
HYDROMETRIC STATION
450 ISOPLETH in mm year⁻¹
mi 100 0 100 200 mi
km 100 0 100 200 km

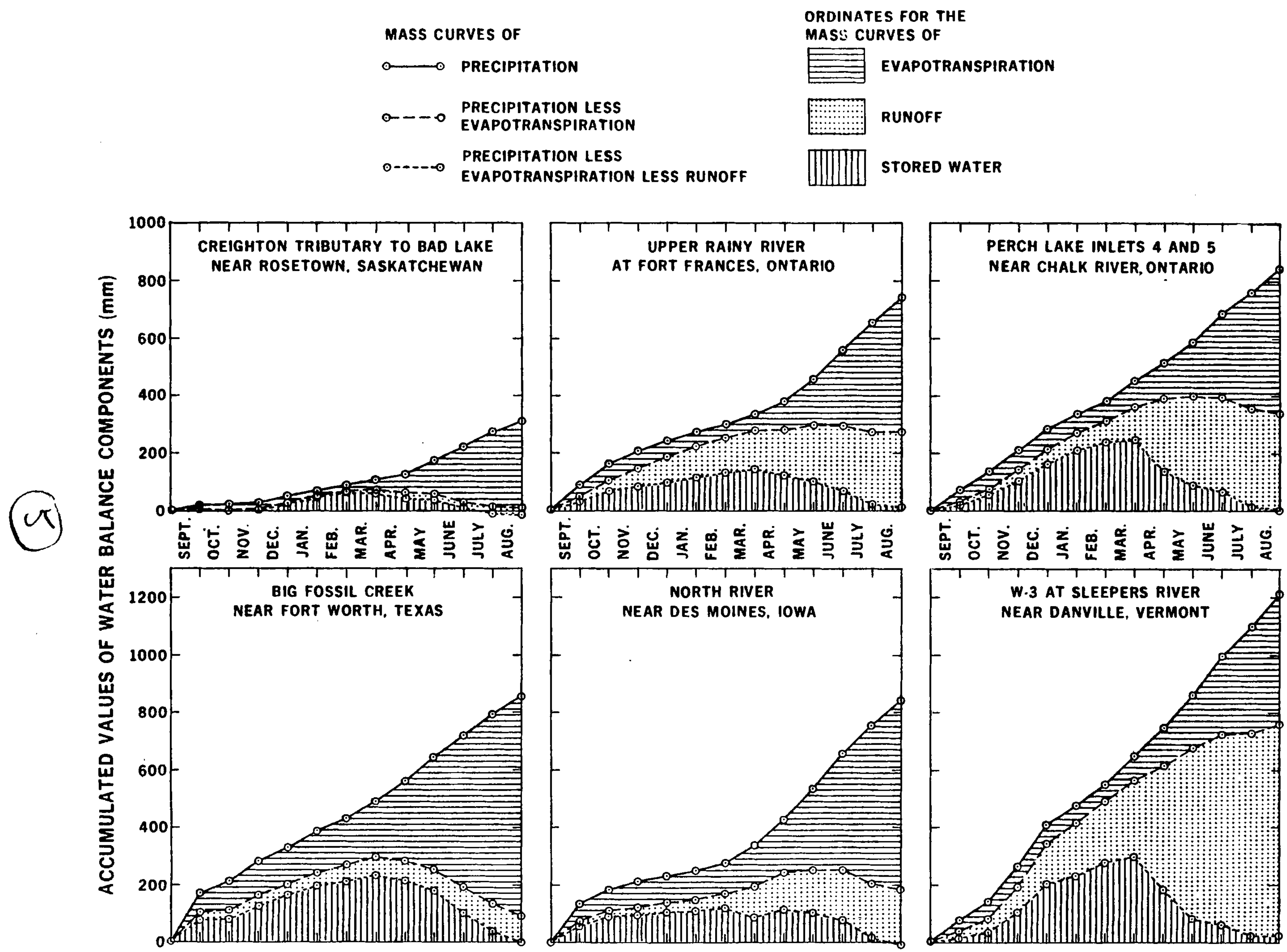

70° 170° 160° 150° 130° 110° 90° 70° 60° 50° 40° 80° 30° 20° 70° 10°
HYDROMETRIC STATION
100 ISOPLETH in mm year⁻¹
YUKON TERRITORY
NWT
BRITISH COLUMBIA
ALBERTA
SASK
MAN
ONTARIO
QUEBEC
NEWFOUNDLAND
NB
PEI
NOVA SCOTIA
SCALE
200 0 200 400MI
200 0 200 400 600KM
60° 50° 40°
120° 110° 100° 90° 80° 70° 60°

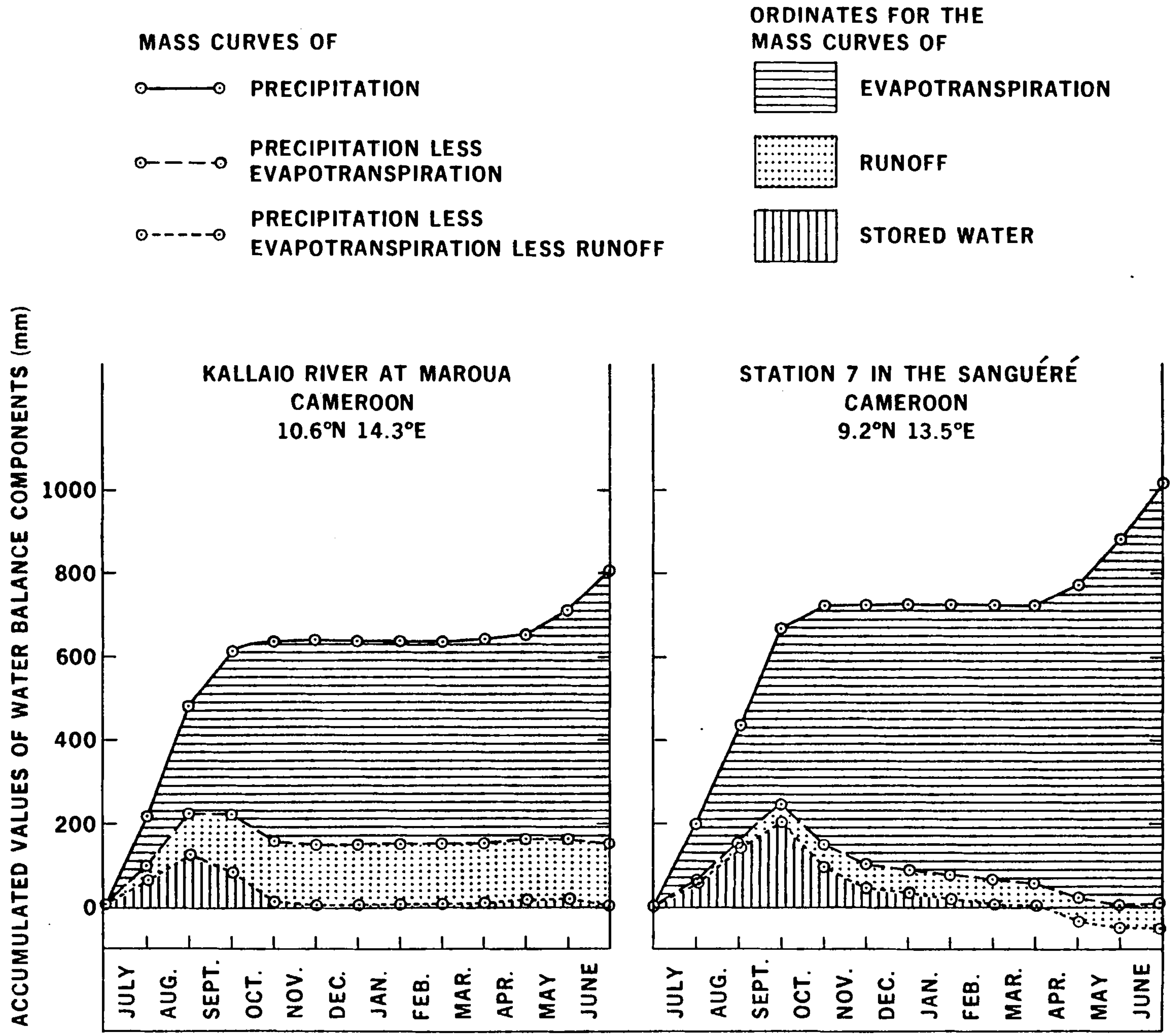
3



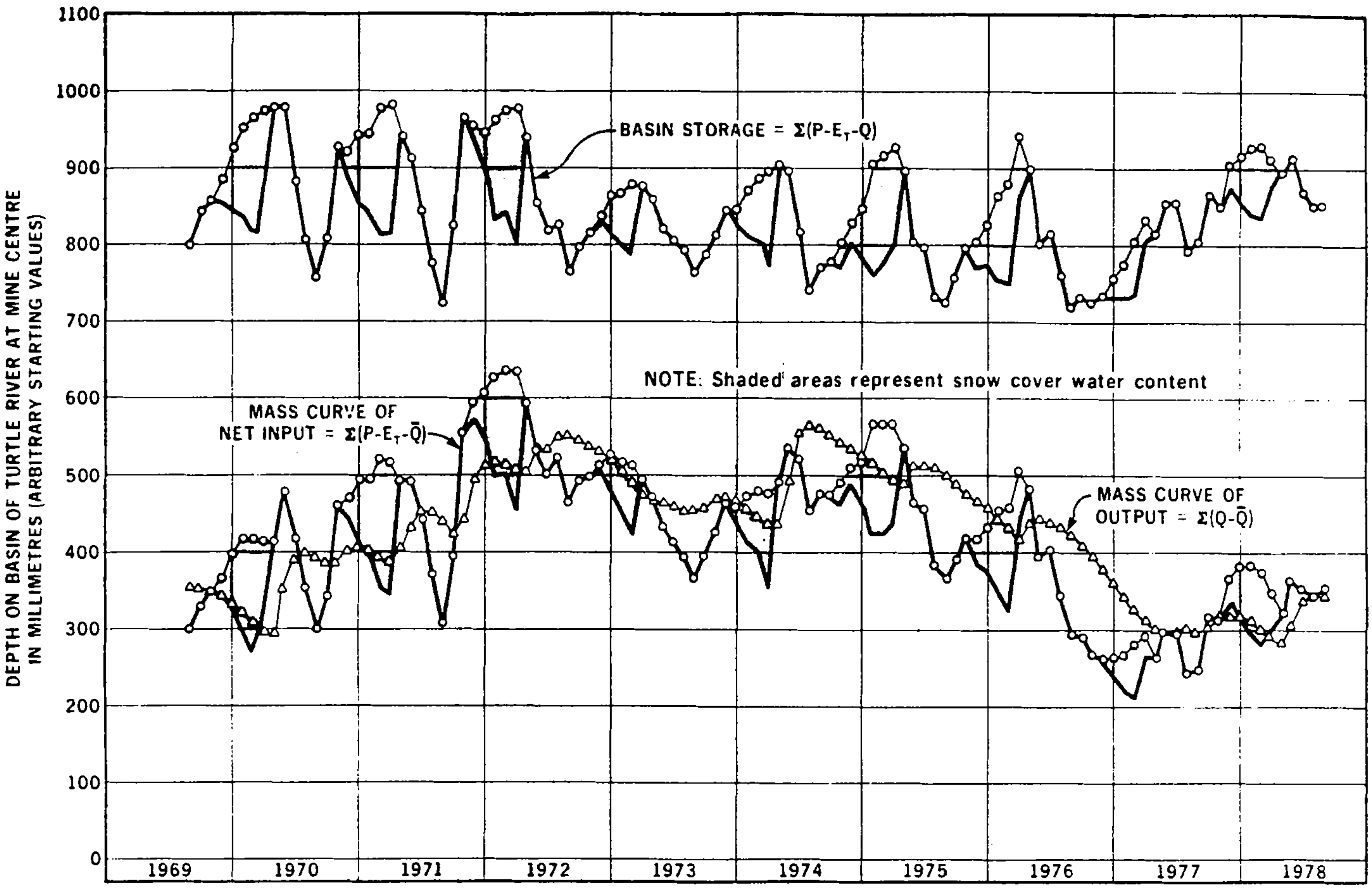
RIVER BASINS IN CANADA
RIVER BASINS IN U.S.A.
RIVER BASINS IN AFRICA
RIVER BASINS IN IRELAND
RIVER BASINS IN AUSTRALIA NEW ZEALAND
.10
.10
0
100
200
300
400
500
600
700
800
900
1000
1100
1200
1300
MODEL ESTIMATE OF AREAL EVAPOTRANSPIRATION FOR RIVER BASINS (mm year⁻¹)
PRECIPITATION LESS RUNOFF FOR RIVER BASINS mm year⁻¹

4


MASS CURVES OF
PRECIPITATION
PRECIPITATION LESS EVAPOTRANSPIRATION
PRECIPITATION LESS EVAPOTRANSPIRATION LESS RUNOFF
ORDINATES FOR THE MASS CURVES OF
EVAPOTRANSPIRATION
RUNOFF
STORED WATER
ACCUMULATED VALUES OF WATER BALANCE COMPONENTS (mm)
CREIGHTON TRIBUTARY TO BAD LAKE NEAR ROSETOWN, SASKATCHEWAN
UPPER RAINY RIVER AT FORT FRANCES, ONTARIO
PERCH LAKE INLETS 4 AND 5 NEAR CHALK RIVER, ONTARIO
BIG FOSSIL CREEK NEAR FORT WORTH, TEXAS
NORTH RIVER NEAR DES MOINES, IOWA
W-3 AT SLEEPERS RIVER NEAR DANVILLE, VERMONT
SEPT. OCT. NOV. DEC. JAN. FEB. MAR. APR. MAY JUNE JULY AUG.
0 200 400 600 800 1000 1200

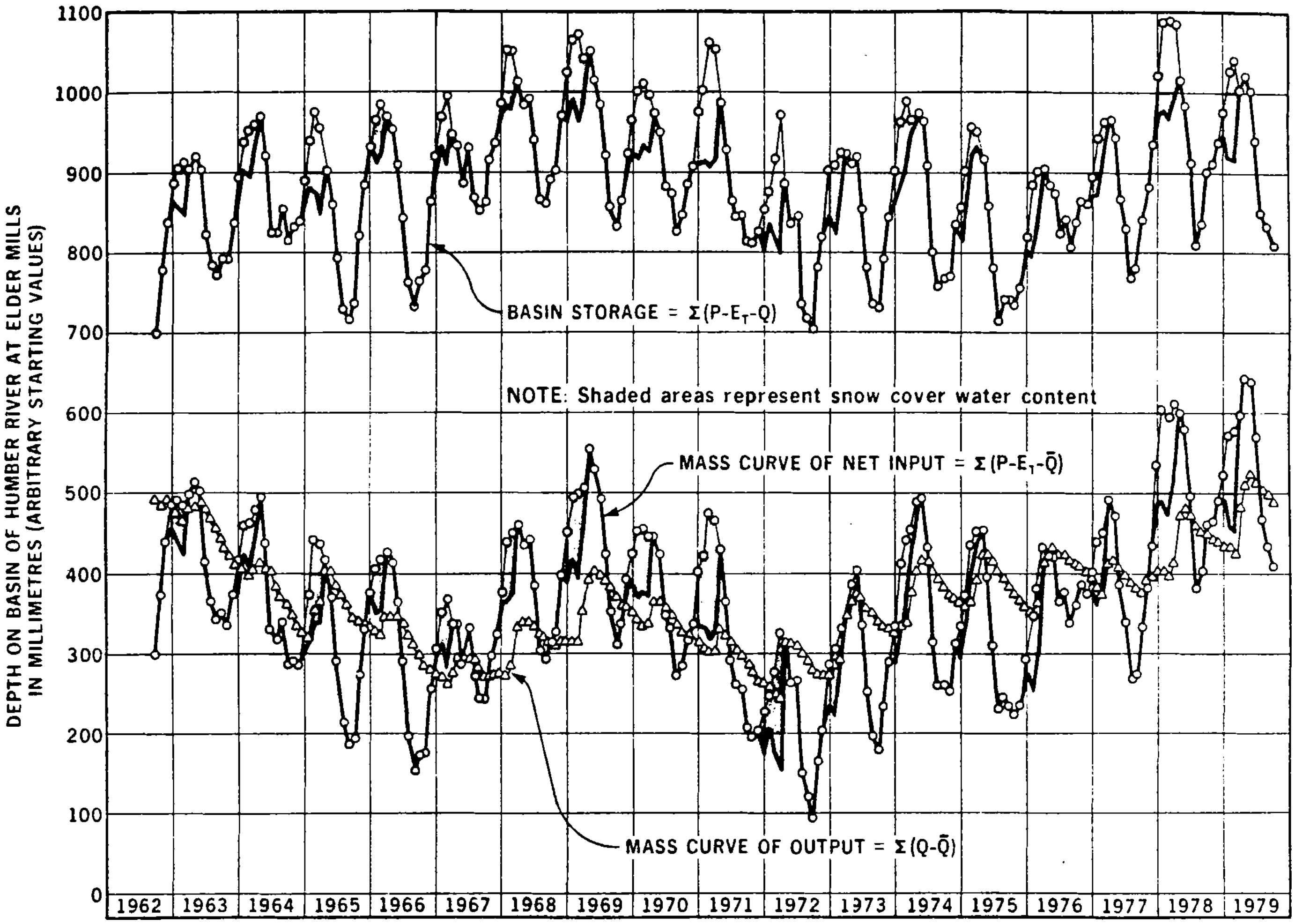


MASS CURVES OF
PRECIPITATION
PRECIPITATION LESS EVAPOTRANSPIRATION
PRECIPITATION LESS EVAPOTRANSPIRATION LESS RUNOFF
ORDINATES FOR THE MASS CURVES OF
EVAPOTRANSPIRATION
RUNOFF
STORED WATER
ACCUMULATED VALUES OF WATER BALANCE COMPONENTS (mm)
KALLAIO RIVER AT MAROUA
CAMEROON
10.6°N 14.3°E
STATION 7 IN THE SANGUÉRÉ
CAMEROON
9.2°N 13.5°E
1000
800
600
400
200
0
JULY
AUG.
SEPT.
OCT.
NOV.
DEC.
JAN.
FEB.
MAR.
APR.
MAY
JUNE

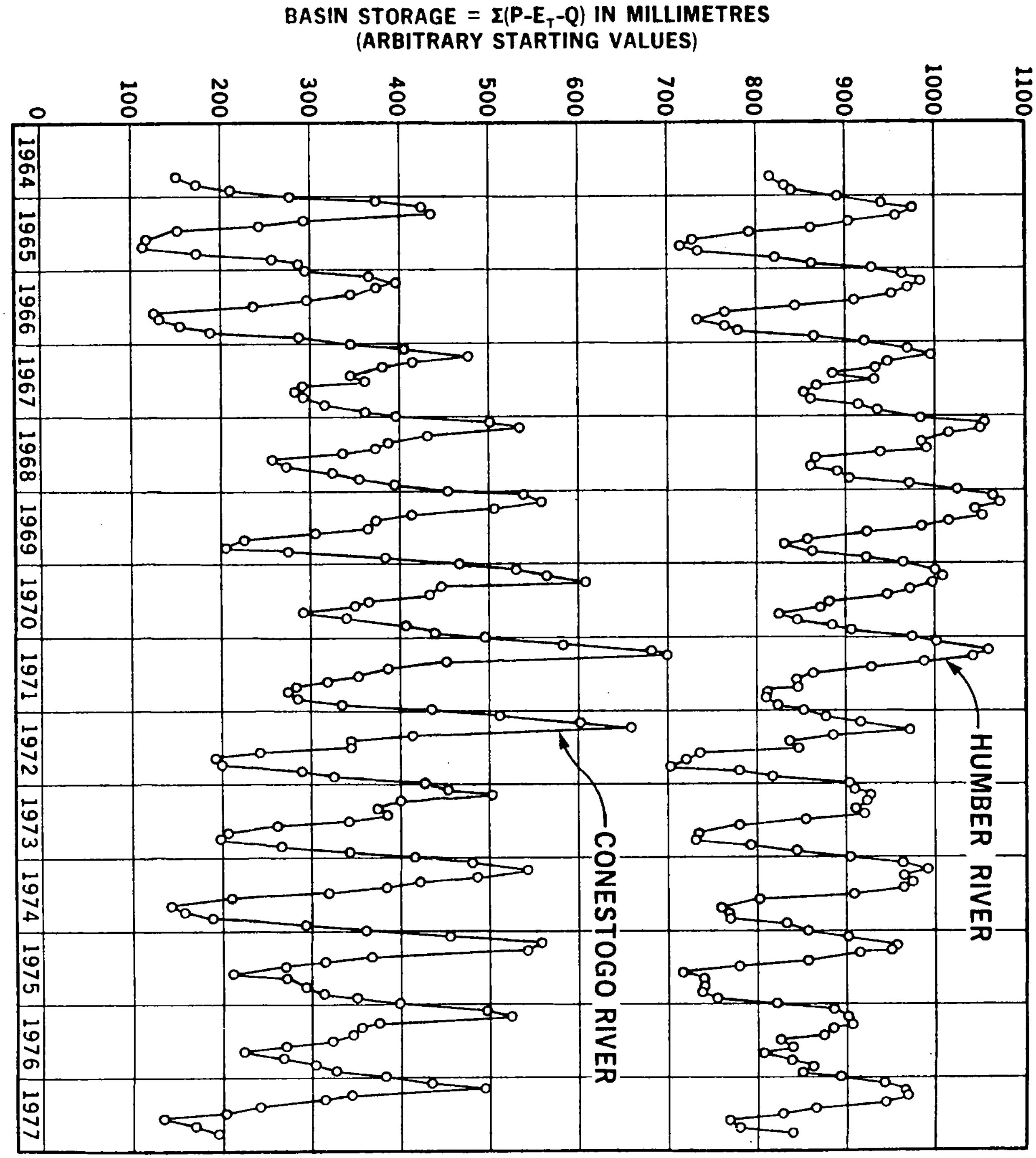


DEPTH ON BASIN OF TURTLE RIVER AT MINE CENTRE
IN MILLIMETRES (ARBITRARY STARTING VALUES)
1100
1000
900
800
700
600
500
400
300
200
100
0
1969
1970
1971
1972
1973
1974
1975
1976
1977
1978
BASIN STORAGE = $\Sigma(P-E_T-Q)$
NOTE: Shaded areas represent snow cover water content
MASS CURVE OF
NET INPUT = $\Sigma(P-E_T-\bar{Q})$
MASS CURVE OF
OUTPUT = $\Sigma(Q-\bar{Q})$

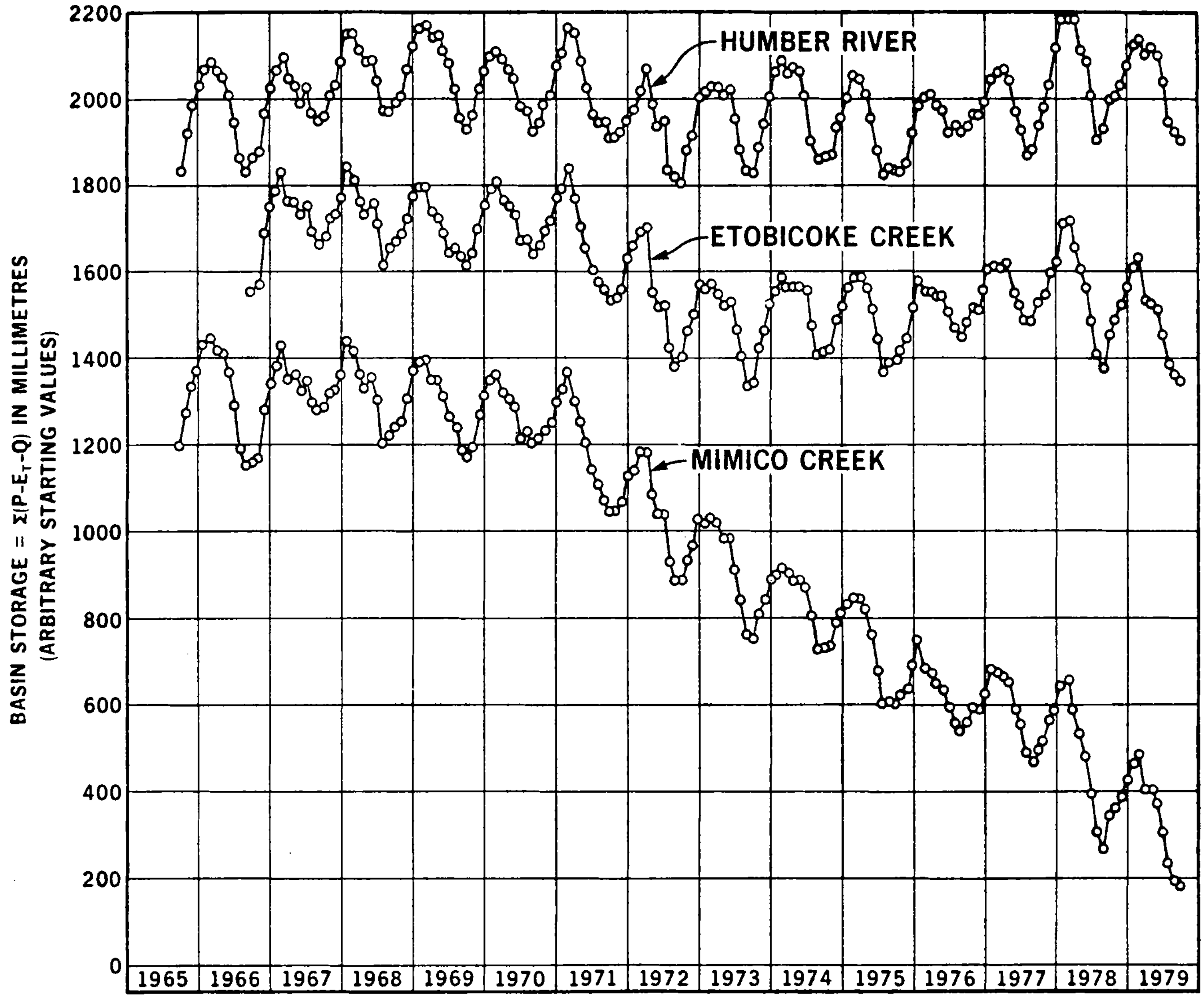
7

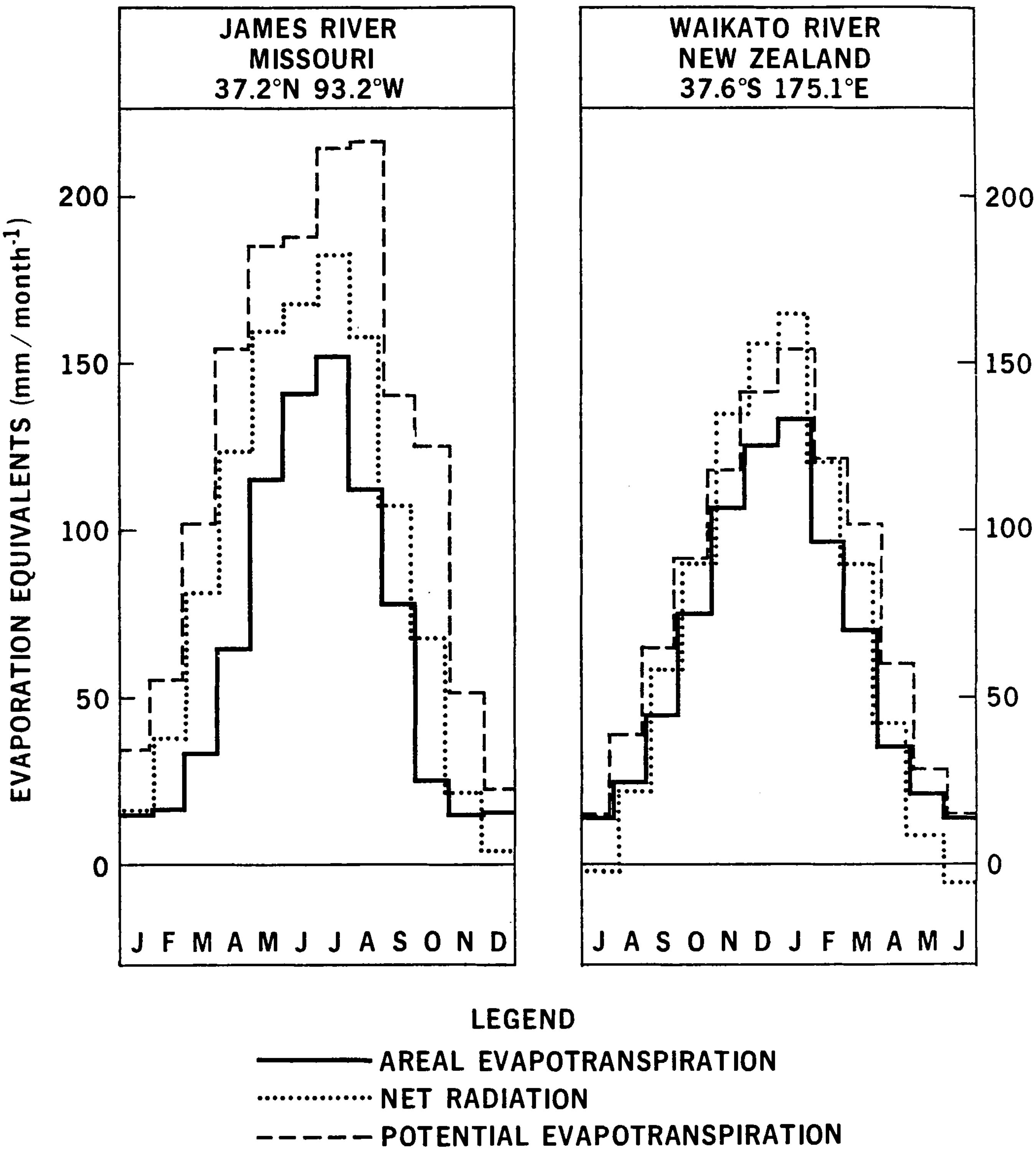


DEPTH ON BASIN OF HUMBER RIVER AT ELDER MILLS IN MILLIMETRES (ARBITRARY STARTING VALUES)
1100
1000
900
800
700
600
500
400
300
200
100
0
1962 1963 1964 1965 1966 1967 1968 1969 1970 1971 1972 1973 1974 1975 1976 1977 1978 1979
BASIN STORAGE = Σ(P-E_T-Q)
NOTE: Shaded areas represent snow cover water content
MASS CURVE OF NET INPUT = Σ(P-E_T-Q̄)
MASS CURVE OF OUTPUT = Σ(Q-Q̄)


BASIN STORAGE = Σ(P-E_T-Q) IN MILLIMETRES
(ARBITRARY STARTING VALUES)
0
100
200
300
400
500
600
700
800
900
1000
1100
1964
1965
1966
1967
1968
1969
1970
1971
1972
1973
1974
1975
1976
1977
HUMBER RIVER
CONESTOGO RIVER

9


HUMBER RIVER
ETOBICOKE CREEK
MIMICO CREEK
BASIN STORAGE = Σ(P-E_T-Q) IN MILLIMETRES
(ARBITRARY STARTING VALUES)
2200
2000
1800
1600
1400
1200
1000
800
600
400
200
0
1965
1966
1967
1968
1969
1970
1971
1972
1973
1974
1975
1976
1977
1978
1979


JAMES RIVER
MISSOURI
37.2°N 93.2°W
WAIKATO RIVER
NEW ZEALAND
37.6°S 175.1°E
EVAPORATION EQUIVALENTS (mm / month-1)
200
150
100
50
0
J F M A M J J A S O N D
J A S O N D J F M A M J
LEGEND
AREAL EVAPOTRANSPIRATION
NET RADIATION
POTENTIAL EVAPOTRANSPIRATION

11