

LONG-TERM SURFACE ENERGY BALANCE OF THE GREAT MIAMI RIVER VALLEY¹

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Abstract. Estimates of the monthly average values of the components of the surface energy balance were calculated from climatological data observed at the municipal airport near Dayton, OH, using a method proposed by Sellers (1965). Based on these calculations the Great Miami River valley had an annual total net radiation of 52 kilolangleys. Latent heat flux to the air from the surface is 37 kilolangleys yr⁻¹ and a sensible heat flux of 15 kilolangleys yr⁻¹. Heat flux into and out of the soil ranged from about 27 langleys day⁻¹ in May and June to -27 langleys day⁻¹ in November and December.

OHIO J. SCI. 79(2): 59, 1979

Biological survival at or near the surface of the earth depends on energy exchange by radiation, conduction, convection and evapotranspiration. These mechanisms are controlled by both surface and atmospheric environments. The climatology of the layer of air near the ground, then, is an important component of an ecological system. One approach to the problem is to follow the flows and conversions of energy in the system during a specified interval of time.

Formulating the climatology is not without its own special problems because its elements may vary considerably in space and time. The variations are related to changes in the physical condition of the earth's surface, stability of the air, transparency of the air to radiation of different wavelengths and to the phase transformations of water. Some insight about different forms and flows of energy (available to the biological component) can be gained by studying the energetic elements of the climate. In this regard, the scope of the present study was limited to the valley of the Great Miami River in southwestern Ohio. The objective was to characterize the energetics of the regional climate near the ground.

THEORY

The ultimate source of energy for ter-

restrial ecological systems is the sun. Its energy reaches the surface as direct and diffuse short-wave radiation ($Q+q$). The surface absorbs a fraction $(1-\alpha)$ of the solar energy, where α (a number between 0 and 1) is the albedo of the surface. The earth radiates energy $I\uparrow$ in the long-wave infrared to the atmosphere and to space through the atmosphere's ozone, CO₂, and water vapor. The atmosphere in turn emits some of its energy earthward as counter-radiation $I\downarrow$ in the infrared. In general $I\uparrow$ exceeds $I\downarrow$. The difference $I\uparrow - I\downarrow$ is the effective outgoing (infrared) radiation I from the earth's surface. Furthermore, solar energy absorbed at the surface usually exceeds the effective outgoing radiation, which gives the surface a positive net radiation R described by:

$$(1) \quad R = (Q+q)(1-\alpha) - I$$

in suitable units of energy flux density, usually langleys per day (ly day⁻¹) where a langley = one calorie/cm².

Prolonged and unrelieved radiative surplus is inimical to life but there are some very efficient mechanisms for disposing of the surplus. A small part G is conducted into the ground as sensible heat but the dominant mode of dissipation in a humid climate is the vaporization of water in the evapotranspiration process. Latent heat LE bound in the water vapor diffuses into the air above the surface because the air is usually thermally and mechanically

¹Manuscript received 8 November 1976 and in revised form 25 August 1978 (#76-87).

turbulent. Turbulence also transfers sensible heat H in smaller amounts upward from the surface. In a dry climate the lack of water would, of course, force an increase in H or G or both. The net radiation (R) in ly day^{-1} can be balanced by writing:

$$(2) \quad R = G + LE + H$$

Direct observations of the parameters in equations (1) and (2) at the same place during the same time period are rare but some methods for estimating monthly average values from climatological observations are available. The methods suggested by Sellers (1965) have had reasonable success and were used for my calculations with some slight modification.

DATA SOURCES

Air temperature T , mean wind speed $u_{13.3}$ and sky cover n were monthly climatological normals for the period 1941–1970 from observations at the National Weather Service observatory at Cox Municipal Airport, Vandalia, OH. The time-weighted mean height of the anemometer is 13.3 m above the ground; hence this subscript is attached to the symbol for mean wind speed. Monthly mean air pressures p at station height (approximately 1000 feet amsl) are 3-year averages, and relative humidities f are averages of 4 daily observations over 12 years. Monthly average solar irradiation ($Q+q$) was obtained from Indianapolis, IN (the closest observatory of record) from a 4-year record of daily values published in monthly issues of *Climatological Data* (1966–1969).

Certain other primary data had to be derived to meet demands of the model. Wind speed at 2 m above ground, in m sec^{-1} , was obtained from the monthly mean wind speed by assuming a "power law" wind profile of the form

$$(3) \quad u_{2.0} = u_{13.3} \left[\frac{13.3}{2} \right]^{-a}$$

where $0.1 \leq a \leq 0.5$; the higher values of a hold in the colder months when the lowest layer of air was most stable.

Solar radiation under clear skies

$(Q+q)_0$ was derived from the mean and annual terms of a Fourier series fit to 12 maximum daily values of insolation, one from each month during 1969 at Indianapolis. The estimate in ly day^{-1} is given by:

$$(4) \quad (Q+q)_0[t] = 488.29 + 282.15 \sin(\omega t - 1.40)$$

where ω is equal to $2\pi/365.25$ and t takes on the julian date of each of the 12 mid-month days. In this and other similar expressions, the phase angles are in radians.

The shortwave albedo α was obtained by harmonic analysis of values typical of the Miami Valley taken from maps prepared by Kung *et al* (1964). The albedo was estimated from

$$(5) \quad \alpha[t] = 0.22 - 0.08 \sin(\omega t - 1.33) + 0.03 \sin(2\omega t + 0.98) + 0.02 \sin(3\omega t + 0.09)$$

Crouch (1976) arrived at virtually the same annual variation from a consideration of area-weighted land use in the Miami Valley, and from albedos of typical surfaces cited by List (1966). The data are summarized in Table 1.

ANALYSIS

Radiant Energy Budget. Since temperatures at the surface are not usually measured, equation (1) needs to be rewritten to account for radiation emitted by the layer of air between screen height and surface. The net radiation in ly day^{-1} becomes:

$$(6) \quad R = \tilde{R} - 4\epsilon\sigma T^3(T_s - T)$$

where emissivity ϵ is equal to 1 in this and another case to follow. The surface temperature is T_s and σ is the appropriate value of the Stefan-Boltzmann constant. The following quantity in ly day^{-1} ,

$$(7) \quad \tilde{R} = (Q+q)(1-\alpha) - \tilde{I}$$

is a first estimate of the net radiation R and in ly day^{-1}

$$(8) \quad \tilde{I} = I_0(1 - kn^2)$$

is an estimate of the effective outgoing radiation at screen height. The quantity I_0 ly day^{-1} is the effective outgoing radia-

TABLE 1
Climatological Data Representative of the Miami Valley.

Variable*	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
T°C	-2.2	-0.9	3.9	10.8	16.4	21.8	23.7	22.8	19.1	13.1	5.4	-0.6
p mb	982.4	980.9	978.6	979.5	977.6	979.1	980.7	982.3	982.7	984.3	981.9	981.6
f	0.71	0.70	0.70	0.65	0.66	0.67	0.67	0.71	0.73	0.69	0.74	0.75
u _{1.3m} sec ⁻¹	5.3	5.4	5.5	5.3	4.5	4.1	3.6	3.3	3.7	4.0	5.1	5.1
n	0.79	0.84	0.81	0.63	0.67	0.62	0.47	0.63	0.64	0.56	0.61	0.76
r cm	7.54	5.66	8.74	9.14	9.78	10.31	9.30	7.87	7.52	6.17	7.11	6.58
F cm	4.12	3.66	4.98	4.47	3.15	2.21	1.40	1.00	0.64	0.66	1.09	2.08
(Q+q) ly day ⁻¹	180.0	253.0	329.0	418.0	487.0	535.0	534.0	509.0	374.0	279.0	148.0	125.0
u _{2.0m} sec ⁻¹	2.2	2.1	2.3	2.5	2.6	2.8	2.8	2.7	2.9	2.8	2.8	2.4
(Q+q) ₀ ly day ⁻¹	232.0	323.0	455.0	600.0	715.0	769.0	747.0	654.0	517.0	373.0	259.0	208.0
α	0.33	0.29	0.20	0.15	0.15	0.16	0.16	0.18	0.22	0.25	0.26	0.30

*T°C=Temperature, Celsius degrees; p mb=Pressure, millibars; f=Relative humidity; u_{1.3m} sec⁻¹=Wind speed at 13.3 meters, meters per second; n=Sky cover; r cm=Precipitation, centimeters; F cm=Runoff, centimeters; (Q+q) ly day⁻¹=Total shortwave irradiation, langley per day; u_{2.0m} sec⁻¹=Wind speed at 2 meters, meters per second; (Q+q)₀ ly day⁻¹=Total clear-sky shortwave irradiation, langley per day; α=Surface albedo.

tion under clear skies. It is related to the relative humidity by

$$(9) \quad I_o = \epsilon(237.6 - 110.7f)$$

In equation (8),

$$(10) \quad k \cong 0.56 + 0.46 \sin [2\pi(0.56v) + 1.22]$$

where

$$(11) \quad v = 1 - 1/n \left[1 - \frac{(Q+q)}{(Q+q)_e} \right] = \frac{(Q+q)_e}{(Q+q)}$$

which is the ratio of the solar radiation received under completely overcast skies (Q+q)_e ly day⁻¹ to that received under clear skies. The two quantities k and v are related because each depends on cloud

height and type. Table 2 shows typical monthly values of the principal components of the radiative energy budget derived up to this point.

Heat Flux into the Ground, G. This component is described with sufficient accuracy by

$$(12) \quad G = 1.3(\Delta T)(\lambda C\omega)^{1/2}$$

$$\sin [\omega(t - 105) + \pi/4]$$

ly day⁻¹, where ΔT is the amplitude (half-range) of the annual temperature curve at screen height. The factor 1.3 modifies screen level ΔT to approximate ground surface ΔT. The quantity (λC)^{1/2} is the so-called "thermal property" of the soil, taken here to be approximately 12.9 ly °C⁻¹day^{-1/2}, appropriate for a clay-loam soil. The phase angle π/4 in equation (12) insures that the maximum (positive) flux of heat into the ground is attained

TABLE 2
Components of the Radiative Energy Budget and Heat Flux into the Ground.

Component*	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
(Q+q) (1-α)	120.0	183.0	262.0	356.0	414.0	448.0	409.0	292.0	211.0	109.0	109.0	87.0
I _o	159.0	160.0	160.0	165.0	165.0	164.0	163.0	159.0	157.0	161.0	150.0	154.0
v	0.72	0.76	0.66	0.52	0.52	0.51	0.39	0.63	0.57	0.55	0.30	0.48
k	0.31	0.25	0.39	0.61	0.60	0.63	0.80	0.44	0.53	0.56	0.92	0.68
I	128.0	132.0	119.0	125.0	120.0	124.0	134.0	131.0	123.0	132.0	103.0	94.0
R	-8.0	52.0	143.0	231.0	294.0	324.0	313.0	278.0	169.0	78.0	6.0	-7.0
G	-20.0	-7.0	7.0	20.0	27.0	27.0	20.0	7.0	-8.0	-20.0	-28.0	-27.0

*(Q+q) (1-α)=Total shortwave irradiation absorbed at surface; I_o=Effective outgoing radiation under clear skies; I=First estimate, effective outgoing radiation; R=First estimate, net radiation; G=Heat flux into the soil. The quantities v and k are dimensionless ratios; all other units are ly day⁻¹.

about 1½ months prior to the occurrence of maximum surface temperatures. According to Sellers (1965, p. 138), this behavior has been verified by observation. The last line of Table 2 indicates the annual course of G , negative values signal heat flux out of the ground.

The Flux of Latent Heat, LE. This term is by far the most difficult to evaluate. Any significant error in the final energy balance can probably be traced to this point in the analysis. What follows here is a very brief summary of the procedure worked out by Sellers (1965, pp. 175-177).

The amount of water E cm mo^{-1} lost to the air from the surface through evapotranspiration is to be found by constructing a water balance. The first step is to find monthly values of potential evapotranspiration, E_o cm mo^{-1} . The Budyko-Penman formulation has been adopted. It is:

$$(13) \quad E_o = d/L \cdot \frac{\tilde{\Delta}(R-G) + \gamma LE_a}{\Delta + \gamma}$$

cm mo^{-1} , where d is the number of days in the month being considered, and L is the latent heat of vaporization of water ($L = 597 - 0.56T$ cal gm^{-1} with air temperature T in ° Celsius). The elements \tilde{R} and G in equation (13) retain their values from table 2. The weighting factors Δ and γ mb deg^{-1} are the slope of the saturation vapor pressure curve in the neighborhood of the air temperature, and a psychrometric-radiative term, respectively. The aerodynamic term in equation (13) is defined by:

$$LE_a = \left(0.622 \frac{\rho LD}{p} \right) [(1-f)e_{sa}]$$

ly day^{-1} . In this expression, ρ gm cm^{-3} is the density of the air at screen temperature and pressure, and e_{sa} mb is the saturation vapor pressure at air temperature. The saturation vapor pressure is calculated according to the Goff-Gratch formulation (List 1966, p. 350) for air over a plane surface of pure ordinary water. For this study, the turbulent

transfer coefficient is defined as:

$$D = (0.60 + 1.73u_{2.0}) \cdot 10^4$$

cm day^{-1} , when $u_{2.0}$ is in meters per second.

The field capacity w_{max} is assumed to be 10 cm. A critical value of soil moisture w_k cm is set equal to $0.75 w_{\text{max}}$. Now the amount of soil moisture w_1 cm at the beginning of April is set equal to w_{max} . At the end of the month, after E cm of water have been lost through evapotranspiration, the soil moisture falls to w_2 cm. The latter value is taken as w_1 for the following month, and so on through the year, back to April. The monthly mean moisture content of the soil is defined as:

$$\bar{w} \cong 0.5(w_1 + w_2)$$

cm, and the soil moisture lost by evapotranspiration (or gained by recharge) is

$$\delta w = w_2 - w_1$$

cm mo^{-1} .

For the monthly value of E cm mo^{-1} , the following relations hold:

$$(14) \quad \begin{aligned} &\text{if } \bar{w} \geq w_k, E = E_o \text{ cm } \text{mo}^{-1}; \\ &\text{if } \bar{w} < w_k, E = \bar{w} E_o / w_k \text{ cm } \text{mo}^{-1}. \end{aligned}$$

The monthly surplus is $S = r - E - \delta w$ cm mo^{-1} , and estimated runoff according to a recursion formula is:

$$\tilde{F}_m = 0.5(\tilde{F}_{m-1} + S_m)$$

cm mo^{-1} , where $1 \leq m \leq 12$ is a monthly index number. If observed runoff data are available, the difference between observed and estimated runoff provides a crude check on the water balance calculations.

Pertinent results of the water budget calculations are summarized in table 3 where the last line of the table is the flux of latent heat recovered by the arithmetic replacement

$$(15) \quad LE = L/dE$$

ly day^{-1} , where E on the right is in centimeters per month as defined in equation (14).

Correction of \tilde{R} . The first estimates of

TABLE 3
Components of the Soil Moisture Balance and the Flux of Latent Heat.

Component*	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
E_o	1.3	1.9	4.2	7.5	10.8	13.2	14.0	12.5	8.3	5.5	2.5	1.5
r	7.5	5.7	8.7	9.1	9.8	10.3	9.3	7.9	7.5	6.2	7.1	6.6
w_1	8.9	9.5	9.5	9.4	7.5	5.2	4.6	4.0	3.9	4.9	5.6	7.5
E	1.3	1.9	4.2	7.5	9.6	9.0	8.5	7.0	5.2	4.0	2.3	1.5
S	5.6	3.8	4.7	3.5	2.5	1.9	1.3	1.0	1.3	1.4	2.9	3.7
\tilde{F}	4.3	4.0	4.3	3.9	3.2	2.5	1.9	1.5	1.4	1.4	2.2	2.9
F	4.1	3.6	4.9	4.5	3.2	2.2	1.4	1.0	0.6	0.6	1.1	2.0
LE	29.0	45.0	80.0	148.0	183.0	176.0	160.0	131.0	101.0	77.0	46.0	32.0

*Soil moisture w_1 in units of cm, latent heat flux LE in ly day^{-1} ; all other units, cm mo^{-1} . E_o =Potential evapotranspiration; r=Precipitation; w_1 =Soil moisture, beginning of month; E=Actual evapotranspiration; S=Surplus moisture; \tilde{F} =Estimated runoff; F=Observed runoff; LE=Latent heat flux.

monthly net radiation can now be corrected. A correction factor c is defined by: $c = \tilde{R} - G - LE \text{ ly day}^{-1}$, such that when:

$$(16) \quad c \geq 0, R = \tilde{R} - \frac{c}{4};$$

$$c < 0, R = \tilde{R} - \frac{c}{2}$$

The corrected values of $R \text{ ly day}^{-1}$ are listed in the first row of table 4. The second and third rows are G and LE repeated from tables 2 and 3 for comparative purposes.

The Flux of Sensible Heat, H. The final component of the energy budget is found as a residual in the energy balance equation. That is:

$$(17) \quad H = R - G - LE$$

ly day^{-1} . The procedure is risky, but an independent check on the magnitude of H is more difficult than the estimation of the LE component. Values of H from equation (17) are listed in the fourth row of table 4.

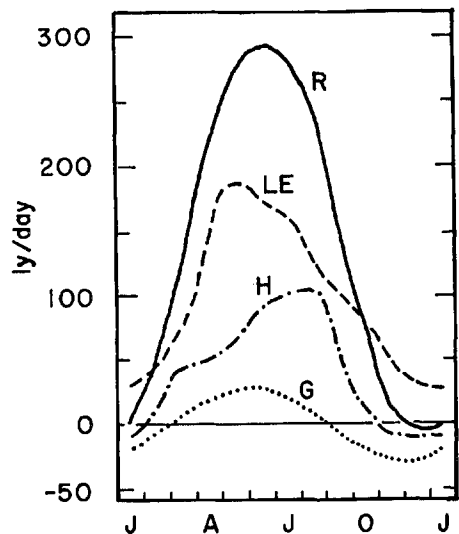


FIGURE 1. Annual variation of surface energy balance components for the Miami Valley. The net radiation, latent and sensible heat flux, and the heat flux into the ground are represented by curves labeled R, LE, H, and G, respectively. The months of January, April, July and October are designated on the abscissa by J, A, J and O, respectively. Energy fluxes are in langley per day.

TABLE 4
Final Estimates of Components of the Surface Energy Balance, Miami Valley.

Component*	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
R	1	48	129	215	273	294	280	243	150	73	13	-1
G	-19	-7	7	20	27	27	20	7	-8	-20	-27	-27
LE	29	45	80	148	183	176	160	131	101	77	46	32
H	-9	10	42	47	63	90	100	104	56	16	-6	-6

*All units, ly day^{-1} . R=Net radiation; G=Heat flux into soil; LE=Latent heat flux; H=Sensible heat flux.

RESULTS AND CONCLUSIONS

The annual variations in the 4 components of the surface energy budget of the Miami Valley are shown in figure 1. The annual course of net radiation R compares favorably with that shown by Sellers (1965, p. 106) for Madison, WI, where net radiation was only slightly higher in summer and slightly lower in winter. Madison's higher latitude produces longer summer days (winter nights) with an attendant increase (decrease) in solar radiation received at the ground, all other elements being roughly equal. Annually, net radiation in the Miami Valley is 52 kly yr⁻¹. Sellers (1965) cited amounts of 60 kly yr⁻¹ for the entire latitude zone of the earth from 30° to 40°N, 45 kly yr⁻¹ for the zone from 40° to 50°N and for North America, 40 kly yr⁻¹. Thus the monthly and annual estimates for net radiation we produced seem to be in accord with Sellers' results for a region whose climatology is similar to that of the Miami Valley.

The flux of latent heat LE is the most effective mechanism for disposing of the radiative surplus in the Miami Valley. This is not an unexpected result, since large areas of the valley are devoted to agriculture, wood lots and forests. Annually, LE amounts to 37 kly. The flux peaks in May, but never quite reaches zero even in the colder winter months. The latter phenomenon may be accounted for by a combination of circumstances. First, winter temperatures are not significantly below freezing in the mean. Furthermore, the aerodynamic term is relatively vigorous throughout the winter and finally high relative humidities are sustained by prevailing winds from the south during the colder months, thus causing the intermittent snow on the ground to melt more rapidly than if humidities were lower. The occasional warm spell then furnishes enough heat energy to evaporate the melted frost and snow. It should be reported at this point that calculations of the water balance were tried with values of field capacity w_{max} up to 20 cm without significantly affecting the overall annual course of actual evapotranspiration presented here.

The reader should also note that the annual variations of estimated and observed runoff are relatively synchronous; annual totals show estimated runoff larger than observed by only 4.3 cm (standard error <0.7 cm). Heat flux into the ground is quite predictable because of the sinusoidal character of its model. It would have been more variable had changes in soil moisture content been considered. At any rate, heat flux into the soil is the smallest of the three dissipative components having an annual average of zero.

Sensible heat flux to the air rises steadily from its winter minimum of -9 ly day⁻¹ (heat flux toward the colder surface) to over 100 ly day⁻¹ in August, when LE is declining sharply. The annual total is slightly greater than 15 kly. Based on the annual totals for H and LE , the Bowen ratio H/LE is about 0.41. Sellers estimates the Bowen ratio for North America as 0.74, and for the entire latitude zone 40° to 50°N, about 0.88 over land. The annual course of net radiation in the Miami Valley seems to be typical for its location. The latent heat flux is perhaps a little larger, and the sensible heat flux a little smaller than expected from corresponding latitudinal and continental averages cited by Sellers (1965).

Acknowledgments. The analytic parts of this study were carried out on the IBM 360-65 computer in the Research and Instruction Computation Center at Wright State University. The cooperation of its director, Dr. D. J. Schaefer, and his staff is gratefully acknowledged. The monthly average values of runoff for the Great Miami Valley were determined from data furnished to me by Mr. Keith Pastor of the Miami Conservancy District, Dayton, OH.

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