NET RADIATION – SOIL HEAT FLUX RELATIONS AS INFLUENCED BY SOIL WATER CONTENT VARIATIONS*

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Abstract. Net radiation, soil heat flux, incoming and reflected solar radiation, and soil water content were measured during several clear day periods following approximate 10-cm applications of water to loam soils at Phoenix, Arizona, and at Sidney, Montana. The regression of soil heat flux on net radiation changed significantly as the soil dried, with the difference between them being a linear function of the volumetric soil water content of the uppermost 2 to 4 cm of soil. The net radiation-soil heat flux difference for soil in an air-dry state was only about one-half of what it was on the day after irrigation. Techniques discussed allow evaluation of what the net radiation-soil heat flux difference would be under conditions of no surface saturation deficit at any time of year from measurements of net solar radiation, soil water content, and air temperature, thereby improving the utility of many evaporation models. The data also indicate that water content measurements may be replaced by more easily measured soil albedo.

1. Introduction

Net radiation (R_N) and soil heat flux (G) are two important components of the Earth's surface energy balance. The difference between them ($R_N - G$) is the primary term of most formulations for potential evaporation (Penman, 1948; Slatyer and McIlroy, 1961; Van Bavel, 1966; Tanner and Fuchs, 1968; Priestley and Taylor, 1972). Indeed, for a wet surface of effectively infinite fetch, the potential evaporation is a direct function of ($R_N - G$). In terms of actual post-potential evaporation (E), the quantity ($R_N - G$) is of great significance too, since many approaches to evaporation modeling formulate E in terms of E_p (Penman and Schofield, 1951; Penman, 1961; Monteith, 1963, 1965; Idso *et al.*, 1975b). A common problem thus shared by all of these approaches is the proper evaluation of the term ($R_N - G$), so that E_p may be evaluated and the ratio E/E_p used to obtain actual values for E.

In applying basic evaporation formulas, the standard practice is to use measured values of ($R_N - G$) to determine E_p. However, as Tanner and Fuchs (1968) and Fuchs *et al.* (1969) have emphasized, if a dry surface is wetted, the surface and equilibrium meteorological conditions may change as a result of wetting, and a new E_p result. Thus, a major problem shared by these post-potential evaporative modeling approaches

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reduces to inferring what $(R_N - G)$ would be under wet surface conditions compared to measurements of $(R_N - G)$ obtained under conditions of some surface saturation deficit so that correct E values may be determined from E/E_p ratios. In this paper we present results of investigations into this subject carried out over a number of years at different seasons and at two different locations.

2. Experimental Sites and Procedures

The primary site of our experimental work was Phoenix, Arizona, with some additional data being acquired at Sidney, Montana. At both sites the ground surface was a smooth bare soil. At Phoenix the soil was an Avondale loam (fine-loamy, mixed (calcareous), hyperthermic, Anthropic Torrifluvent), and at Sidney it was a Williams loam (fine-loamy, mixed, Typic Argiboroll).

The primary data were net radiation, soil heat flux, and soil water content. Net radiation was obtained from Fritschen (1963, 1965) net radiometers, located about 25 cm above the soil surface for optimal results (Idso and Cooley, 1971, 1972). Soil heat flux was obtained from National Instruments Laboratory* Model HF-1 heat flow discs calibrated by the procedure of Idso (1972) and placed at a depth of 1 cm. Soil water contents were obtained by gravimetric sampling, as described in detail by Jackson (1973) and Jackson *et al.* (1973).

Ancillary data acquired during our experiments were air temperature and incoming and reflected solar radiation, from which albedo was obtained. This latter quantity was needed to infer water content values on those days that direct measurements of this parameter were not made (Idso *et al.*, 1975a).

Six major experiments were carried out in July, 1970; March, 1971; and May, September, and December, 1973, at Phoenix; and in September, 1973, at Sidney. The first two of these experiments did not include soil heat flux data; and the last one did not include direct soil water content measurements.

In the first five experiments, the bare soil was flooded with about 10 cm of water and then allowed to dry, while in the last one just over 9 cm of water was supplied by rain. Beginning the day following these water additions, net radiation, soil heat flux, air temperature, and albedo data were acquired at 20- to 30-min intervals for periods ranging from one to four weeks.

3. Soil Heat Flux vs Net Radiation

A common way of comparing soil heat flux and net radiation has been to plot instantaneous values of the former parameter as a function of the latter for different individual days. We have done this for the concurrent September 1973 experiments at Phoenix and Sidney in Figures 1 and 2, where the slopes of the G vs R_N plots are seen

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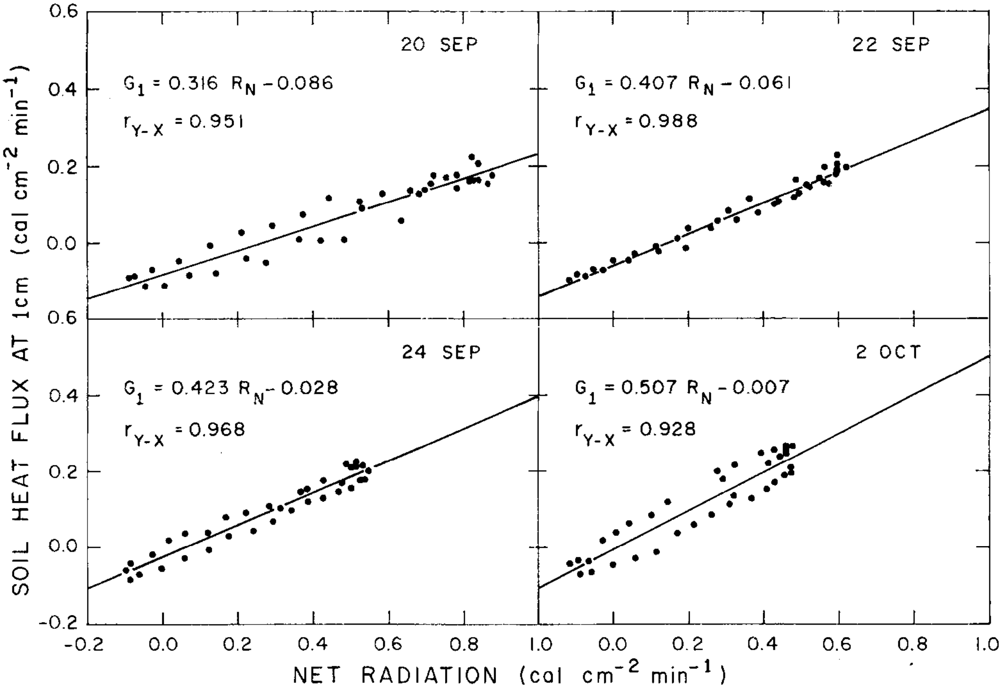

Fig. 1. Regressions of soil heat flux measured at 1-cm depth in a smooth bare field of Avondale loam at Phoenix, Arizona, in September, 1973, vs net radiation measured at 25 cm above the field, following an irrigation of approximately 10 cm of water applied on 17 September.

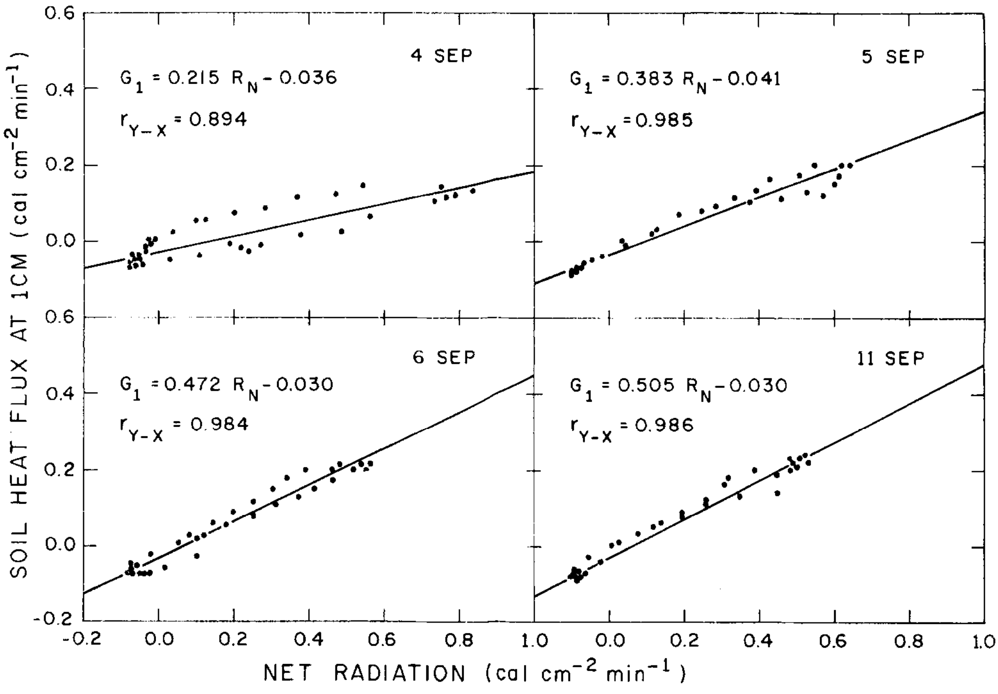

Fig. 2. Regressions of soil heat flux measured at 1-cm depth in a smooth bare field of Williams loam at Sidney, Montana, in September, 1973, vs net radiation measured at 25 cm above the field, following rains supplying a total of 9.1 cm of water on 1, 2, and 3 September.

essentially to double in going from wet to dry conditions. Fuchs and Hadas (1972) have presented similar plots for a wet and dry loessial soil of the northern Negev that exhibit no differences in slope. Apparently, their less variable results were due to their wet condition occurring fully two weeks after irrigation, whereas ours occurred from only one to three days afterwards. Indeed, the volumetric water content of the 0–4 cm

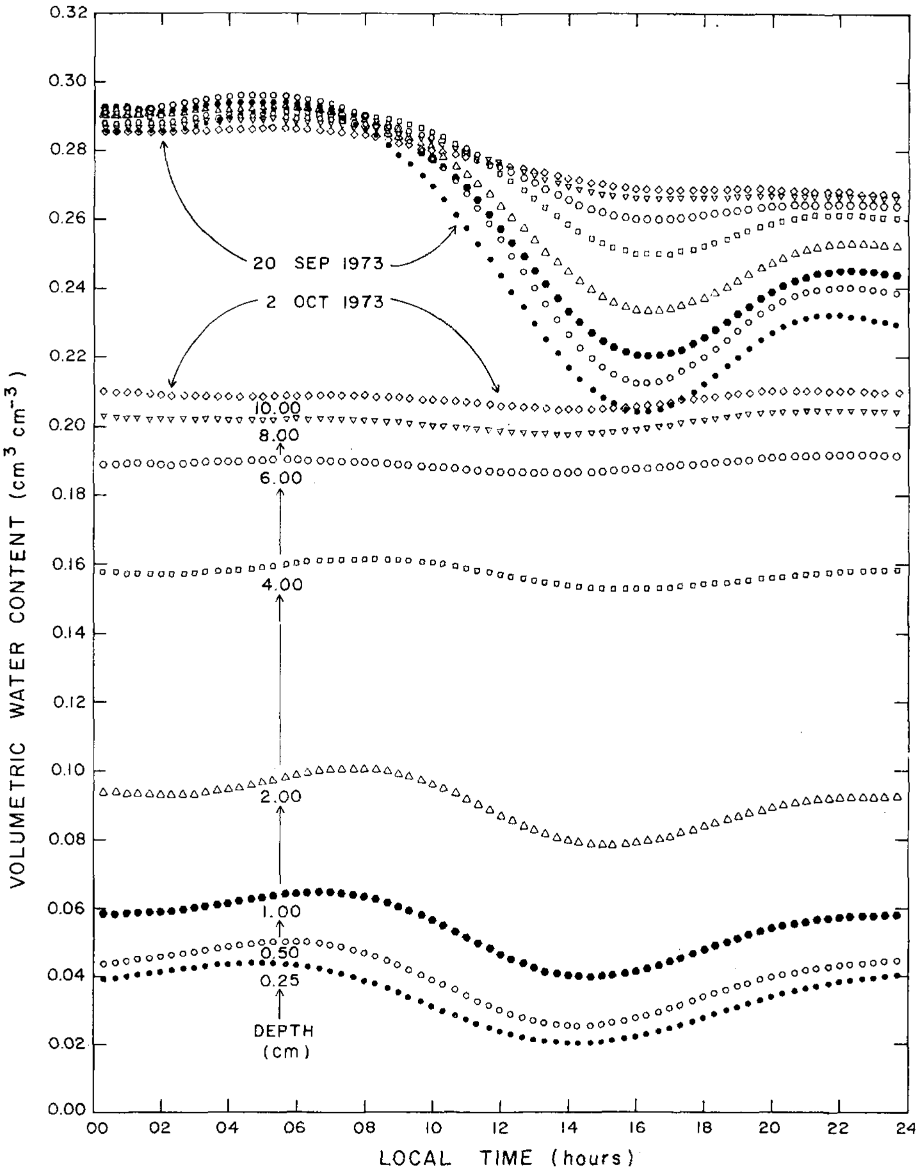

Fig. 3. Diurnal variations in volumetric soil water content at several depths in an Avondale loam on 20 September and 2 October, 1973, following a 10-cm irrigation on 17 September.

layer of our Avondale loam ranged from about 0.29 to 0.22 on our wet day (Figure 3), while theirs ranged from about 0.15 to 0.09, fully 50% less. Fuchs and Hadas apparently worked with soils well past the first stage of drying and possibly into the third stage of drying to obtain their similar results for both 'wet' and 'dry' soils (Idso *et al.*, 1974).

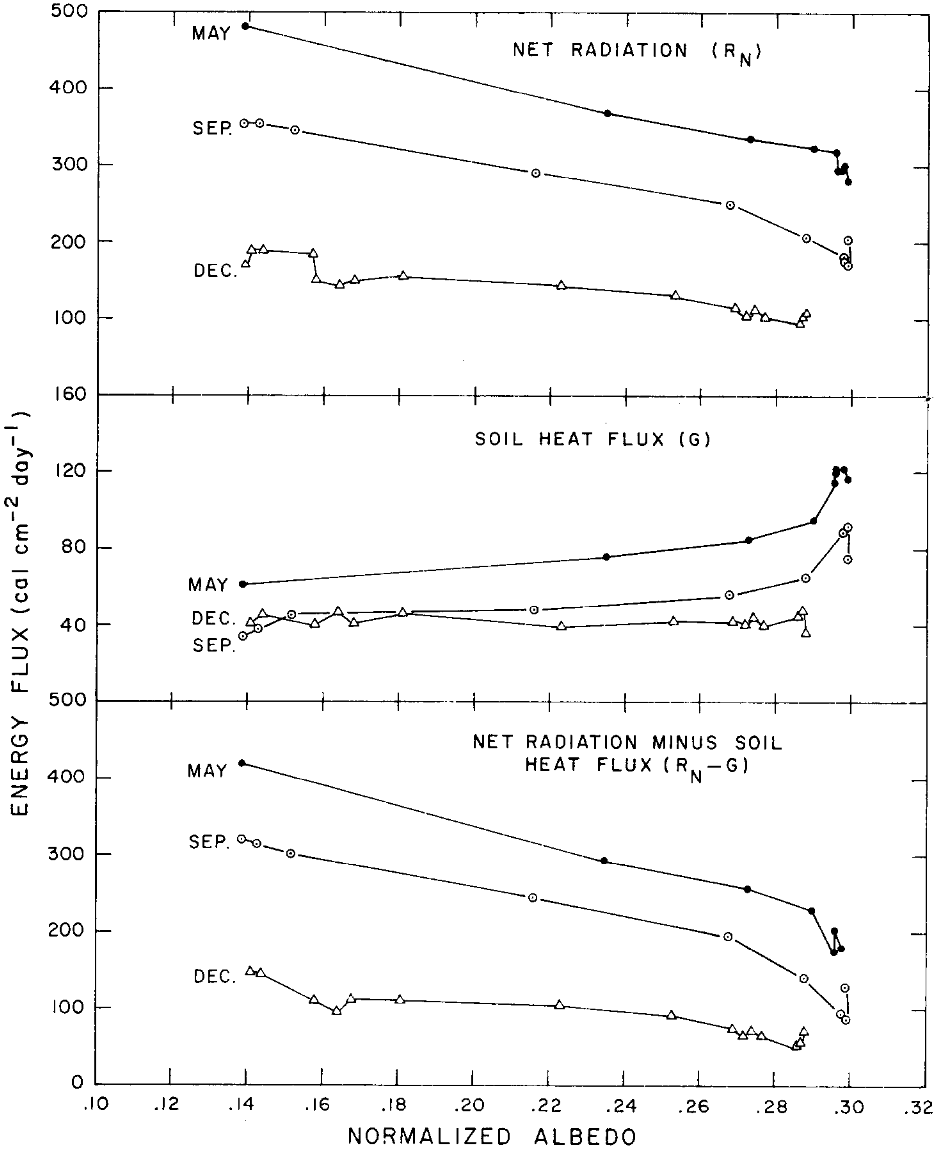

Fig. 4. Daylight totals of net radiation, soil heat flux, and their difference vs mean daylight normalized albedo following 10-cm irrigations of an Avondale loam at Phoenix, Arizona, at three different times of year.

4. Net Radiation Minus Soil Heat Flux

Figure 4 depicts plots of daylight totals of R_N, G, and $(R_N - G)$ vs soil albedo (normalized to remove solar zenith angle effects) as the soil dried during the three 1973 Phoenix experiments. Results for the three months are all different, due to the fact that external boundary conditions vary among those periods. The two boundary conditions of prime importance are the solar radiation and atmospheric thermal radiation. The former energy flux varies with the changing solar altitude and daylength through the year; and the latter energy flux varies with the changing effective emissivity of the atmosphere through the year.

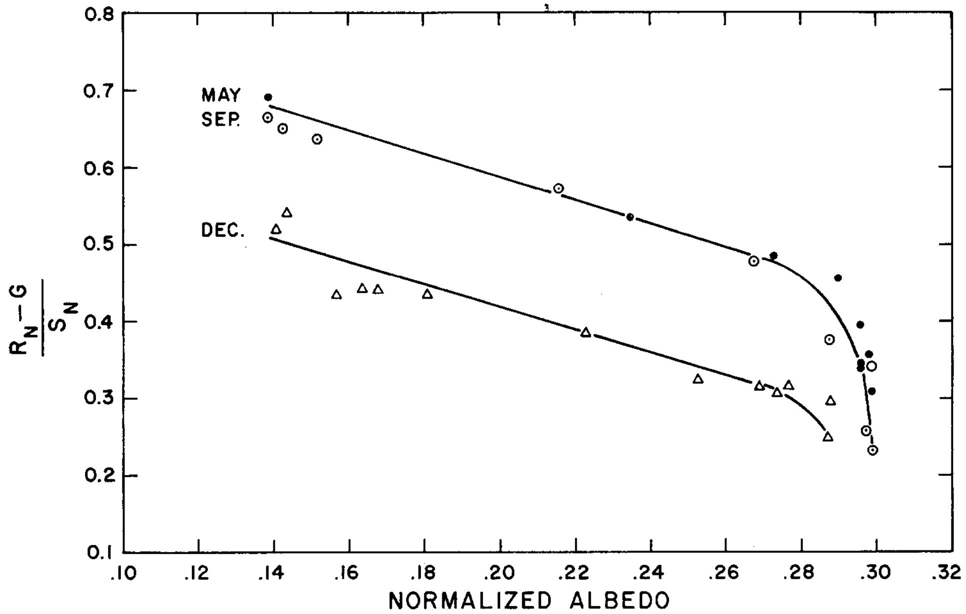

Fig. 5. The daylight totals of net radiation-soil heat flux differences divided by net absorbed solar radiation vs mean daylight normalized albedo.

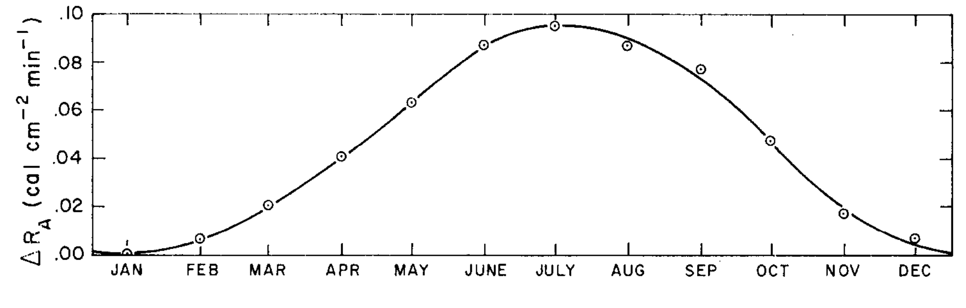

Fig. 6. The mean daylight instantaneous excess of atmospheric thermal radiation in relation to the rate of receipt in January, computed by the equation of Idso and Jackson (1969).

We normalize the results of Figure 4 for the variable solar radiation forcing function by dividing $(R_N - G)$ by S_N, with results as shown in Figure 5. To normalize these results for the variable atmospheric thermal radiation, we first use the Idso and Jackson (1969) formula for clear sky atmospheric thermal radiation and the mean daylight temperature of each month at Phoenix to develop the graph of Figure 6. Using January, the coldest month, as a base, the graph gives for all other months, the mean instantaneous excess rate of energy (ΔR_A) received from the atmosphere via this means.

In viewing Figure 6 it may be seen that May and September receive about equal amounts of atmospheric thermal radiation. December, however, receives considerably less. Thus, the difference between the December R_A value and that of the average of the other two months was taken and multiplied by the length of the mean December daylight period. This extra amount of energy was added to each of the December $(R_N - G)$ values in Figure 4 to express all results in terms of an equal R_A component in R_N. With these adjustments for independently varying boundary conditions, the results for all three seasons were equivalent (Figure 7).

The final step to be taken involved relating the normalized $(R_N - G)/S_N$ curve of Figure 7 to soil water content. Comparison of the curve of Figure 7 with the family of volumetric water content vs normalized albedo curves developed by Idso *et al.* (1975a) for this soil indicated that the curves integrating volumetric soil water content over

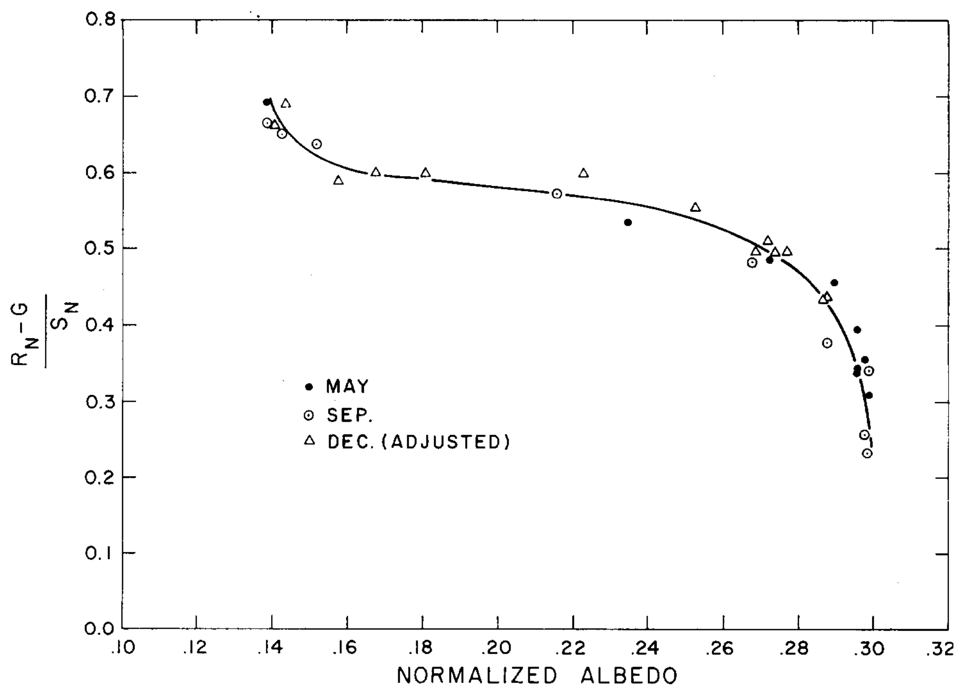

Fig. 7. A plot similar to Figure 5, but with the December data adjusted to make up the deficiency in atmospheric thermal radiation received in December relative to that received in May and September.

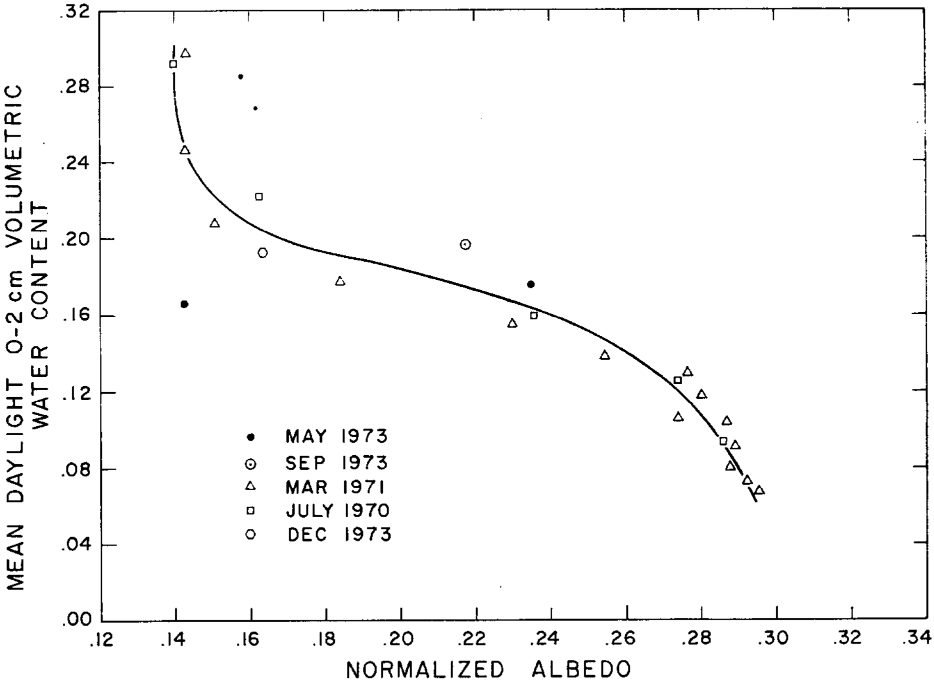

Fig. 8. Mean daylight volumetric water content of the uppermost 2 cm of an Avondale loam vs mean daylight normalized albedo of the soil surface.

the 0–2 and 0–4 cm depth intervals looked quite similar. Thus, since soil heat flux was measured at a depth of 1 cm and since R_N is related to surface soil properties, a new relation was developed in Figure 8 relating the mean daylight 0–2 cm volumetric soil water content to normalized albedo. The water contents used in its development were obtained from total daylight integrations of direct measurements made every 20 or 30 min.

Corresponding values of $(R_N - G)/S_N$ and volumetric soil water content were next obtained from Figure 7 and 8 for every normalized albedo interval of 0.01 starting at 0.14 and extending to 0.29. The $(R_N - G)/S_N$ values were then plotted vs the volumetric soil water contents (Figure 9), and found to describe a straight line. The line drawn through the points was determined from a linear regression analysis. A similar analysis utilizing 0–4 cm soil water contents yielded a regression equation $X = 0.225 + 1.54\,Y$ with correlation coefficient $r_{Y-X} = 0.986$.

5. Discussion

The results of Figure 9 indicate the great importance of the comments of Tanner and Fuchs (1968) and Fuchs *et al.* (1969) cited in the introduction. Wetting of a dry surface does indeed drastically alter conditions related to the calculation of potential (and,

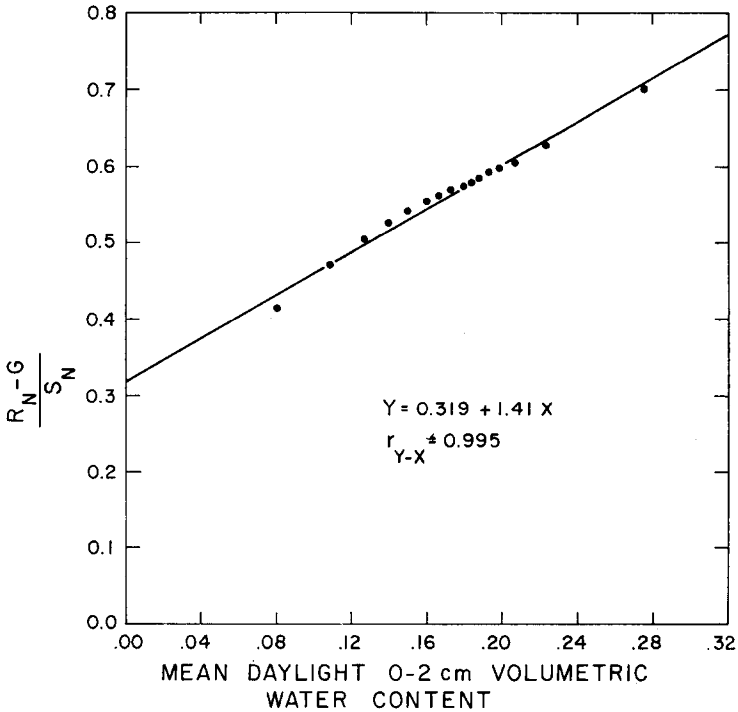

Fig. 9. The net radiation-soil heat difference divided by net absorbed solar radiation vs mean daylight volumetric water content of the uppermost 2 cm of an Avondale loam as determined from Figures 7 and 8.

hence, actual) evaporation. For the Avondale loam we have investigated, the dominant $(R_N - G)$ term is reduced approximately to half in going from an air dry to a condition of no saturation deficit in the uppermost 2 to 4 cm. Using the procedures outlined here, however, it is possible to infer what $(R_N - G)$ would be under such a wet condition at any time of year from measurements of S_N and soil water content. Our data also indicate that the water content measurement may be by-passed in favor of the more easily measured soil albedo.

It should be noted that although our procedure allows water content dependent changes in $(R_N - G)$ to be calculated, it does not treat the problem of potential changes in the near-surface air temperature and vapor pressure. In humid areas, such changes are negligible, but in arid areas they may be significant. Burman *et al.* (1975) recently measured variations in these parameters at 2-m height over irrigated farmlands at Twin Falls, Idaho, and the adjacent desert. The *maximum* variations they encountered created 20% reductions in potential evapotranspiration calculations for the irrigated farmlands compared to a similar surface in the desert environment. In this situation, when the energy and aerodynamic terms of combination type evaporation equations may be equal, the effect we treated is 2.5 times more important than the one we have neglected. In humid areas, where the energy term may account for practically all of the evaporation, the effect we treated is the only one of significance. Thus, the procedure we have developed will account for 80 to 100% of the variability in potential evaporation calculations caused by variations in the input parameters induced by soil water content reductions below that sufficient to sustain potential evaporation.

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