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Estimating Incident Terrestrial Radiation under all Atmospheric Conditions

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Abstract. To apply a complete energy balance to the computation of evaporation or snowmelt, radiation data are necessary. At present there is a reasonably adequate network of incident solar radiation stations. However, a network of all-wave radiometers does not exist. It is therefore evident that to apply the energy budget technique to hydrological problems, incident long-wave radiation has to be estimated. An empirical equation is developed by which incident long-wave radiation can be computed from observations of surface air temperature, vapor pressure, and incident solar radiation. Based on comparisons of estimated and observed incident long-wave radiation at six locations, the following conclusions seem warranted: (1) The equation should give results comparable within a few per cent on a long-term basis (six months or longer); (2) For any specific atmospheric condition, there seems to be no tendency for the equation to over or under compute or to give increased scatter; (3) When data for periods of six months or more were analyzed, correlation coefficients of approximately 0.90 were obtained between computed and observed daily radiation values. (Key words: Radiation; evapotranspiration; hydrology; meteorology)

INTRODUCTION

To apply a complete energy balance to the computation of evaporation or snowmelt, radiation data are necessary. For the majority of evaporation computations, incident all-wave radiation data would be sufficient since the reflectivity of a water surface is nearly constant (.03 to .06 [*Anderson,* 1954]) throughout the electromagnetic spectrum. In the case of snowmelt, there are large differences in the reflectivity of the snow in the portion of the spectrum less than 4 microns (short-wave or solar) as compared with the remainder of the spectrum (long-wave or terrestrial) and, therefore, both incident short-wave and long-wave radiation are necessary. For simplicity, incident long-wave radiation will be called 'atmospheric radiation' hereafter.

In the conterminous United States there is a fairly adequate network of incident solar radiation stations (69) to provide data for the operational use of energy balance computations. However, a network of all-wave radiometers does not exist. Although there are a limited number of net all-wave radiometers at various research installations, these data are of little use over the surrounding area, because net all-wave radiometers values are only applicable to the type of surface over which the instrument is installed. If the instrument is installed over a standard surface (water would be excellent), the observed net radiation could be corrected for back radiation, if the temperature of the water surface is known, to give a value of incident minus reflected all-wave radiation. Incident all-wave radiation could then be computed by assuming an average all-wave coefficient of reflection. Only a few research installations measure incident all-wave radiation directly. It is not likely that such a network will be established in the near future.

The Weather Bureau is now testing an experimental insulated evaporation pan (designated X-2) that, if observations of water, dew point, and air temperature are made concurrently with pan evaporation, will provide estimates of incident all-wave radiation, using a method similar to that described in the *Lake Hefner Report* [1952] for the Cummings Radiation Integrator. The method is as follows:

1. Net radiation for the insulated pan can

be computed as the residual of the energy balance;

2. From the surface water temperature, emitted long-wave radiation can be computed, thus providing a value of incident minus reflected all-wave radiation;

3. Using this value 2 and an average all-wave coefficient of reflection, incident all-wave radiation can be calculated. If incident solar radiation is measured, incident atmospheric radiation can be obtained.

A network of these insulated pans, if established in the future, could provide a measure of incident all-wave radiation during the period of the year that the pans are operable. These pans would not be operable during the snowmelt season. As stated previously, in the application of the energy budget to computation of rate of snowmelt, incident solar radiation and atmospheric radiation must be treated separately. Therefore, for purposes of computation of both evaporation and snowmelt, it would be necessary to have a method for estimating atmospheric radiation.

The purpose of this study was to develop a general equation to estimate atmospheric radiation at any location. The equation should estimate atmospheric radiation from surface observations. The equation was developed to provide a lower radiation limit (clear sky) and an upper radiation limit (cloudy sky). The clear sky portion of the final equation was developed mainly from radiosonde data from four radiosonde stations: El Paso, Texas; Las Vegas, Nevada; Ely, Nevada; and Santa Monica, California. The cloudy sky radiation limit was set by Stefan-Boltzmann's equation. The proportioning in between clear sky and cloudy sky radiation was determined from radiometer data from the U. S. Army Cold Regions Research and Engineering Laboratory (CRREL), Lebanon, New Hampshire. It was hoped that the development of the equation would be based as little as possible on radiometer measurements, even though it would be necessary as a check to compare some of these measurements with the results of the equation. Radiometer measurements were eliminated from the development because of their unknown quality. Although it is not the purpose of this paper to go into a detailed discussion of radiometer measurements, a few pertinent comments should be made.

The accuracy of all-wave radiometers under the conditions of calibration is estimated to be within a few per cent. In the field, however, this accuracy, even with the most careful observational techniques, would be difficult to obtain, owing to such things as unequal response to different wavelengths, nonuniform specular response, changes in the absorption characteristics of the receiving surface, and responses caused by sources other than radiative energy (such as wind and precipitation). Also, since an instrument to measure only long-wave radiation has not been developed, measured values of solar radiation must be subtracted from all-wave radiation to obtain long-wave radiation. Thus any error in the measurement of solar radiation also affects the long-wave radiation. In the past the temperature dependence of pyranometers has been ignored in most cases. The effect of this temperature dependence can be quite significant. Pyranometers are now available that greatly reduce this temperature dependence.

BACKGROUND

A blackbody emits with maximum possible intensity at all wavelengths at a given temperature. The total energy (E) emitted by a blackbody can be determined by Stefan-Boltzmann's law

$$E = \sigma T^4 \tag{1}$$

where σ is the Stefan-Boltzmann constant and T is the temperature of the body in °K. A grey body is one that, at a given temperature, emits a fixed proportion of the blackbody radiation in all wavelengths. Considering conditions of clear skies, the Earth's atmosphere is neither a black nor a grey body. Rather it absorbs and emits radiation to varying degrees, dependent on wavelength. At certain wavelengths the atmosphere acts almost as a blackbody, whereas at others it absorbs only a small portion of the radiation. Only two of the gases in the atmosphere, carbon dioxide and water vapor, have an appreciable effect on the long-wave portion of the spectrum. The over-all effect of carbon dioxide is less important in the radiative exchange than water vapor, because of its lesser quantity and fewer absorption bands. Since the proportion of carbon dioxide in the atmosphere is practically constant, its effect may be considered as fixed for our purposes. The amount of water vapor in the atmosphere, however, exhibits wide variations. Thus it is the controlling variable in determining the amount of atmospheric radiation for a given temperature and with clear skies. The long-wave radiation to the Earth's surface is a result of radiation from all levels of the atmosphere. It is dependent upon the moisture content and temperature distribution of the entire atmosphere. *Elsasser* [1942] has advanced a method to determine the atmospheric long-wave radiation if these distributions are known. This method, however, is quite complex and requires that upper air soundings be made. Since the layers of the atmosphere near the Earth have the greatest moisture content and highest temperature and display the greatest variability, they generate a majority of the atmospheric radiation received at the surface. Based on this fact, *Angstrom* [1919], *Robitzsch* [1926], *Elsasser* [1942], and *Brunt* [1944] have proposed equations to estimate clear sky atmospheric radiation from surface air temperature and vapor pressure. Because of the high degree of correlation normally found between air temperature and vapor pressure at a given location, *Swinbank* [1963] proposed an equation utilizing surface air temperature alone. These equations all contain coefficients that must be evaluated from radiation observations at the given location or otherwise must be estimated based on evaluations conducted by others. A summary of some of these evaluations is given in *Snow Hydrology* [1956].

When clouds are present the amount of atmospheric long-wave radiation increases. The absorption spectrum for liquid water is similar to that for water vapor; however, the magnitude of the absorption is much greater. Clouds can effectively be considered to be blackbodies with respect to atmospheric radiation. *Angstrom* [1919], *Anderson* [1954], *Lamoreux* [1959], and *Koberg* [1964] have proposed equations to estimate the atmospheric radiation under all conditions from surface air temperature, surface vapor pressure, and an index to the extent and/or height of the cloud cover. The coefficients for these equations are based on radiometer measurements of atmospheric radiation.

CLEAR SKY RADIATION

A logical first step in the derivation of an equation to estimate atmospheric radiation is to determine upper and lower limits for any set of conditions. This section discusses the methods used to obtain the lower limit, i.e., the atmospheric radiation from clear skies.

To determine the parameters needed to estimate clear sky atmospheric radiation reasonably, this quantity was computed from theoretical considerations, using a digital computer program developed by *Myers* [1966]. This program is basically a computerized application of methods described by *Elsasser* [1960]. Myers uses Elsasser's generalized transmissivities, except that lower water vapor transmissivities from *Wark et al.* [1962] are substituted for wavelengths of from 8.6 to 13 microns.

Atmospheric radiation under clear skies was computed by Myers' program, using mean monthly radiosonde data from four stations varying in elevation and climatic conditions. The months used were chosen to give a good range and distribution of surface level temperatures and because of relatively consistent clear sky conditions throughout. These data are summarized in Table 1. Monthly data were used to reduce the very considerable time that would have been required for daily analysis.

The results of the program were plotted against surface level air temperatures and relative humidities. This plot is shown in Figure 1. It was decided that the scatter was more than could be tolerated. To determine if the scatter could be reduced, the humidity profiles were made constant for all stations, and the effect of upper air temperature variations was determined (i.e., relative humidity assumed constant at all levels). Then the temperature profiles were held constant, and the effect of upper-air water vapor variations was determined. The results of these tests are shown in Figures 2 and 3. As can be seen, the variation in upper-level temperatures is mainly responsible for the scatter of Figure 1.

To reduce the scatter, a station adjustment term is used. This term is a function of the long-term relationship between air temperatures at the surface and one upper level. To determine the station adjustment term, a mean daily upper air temperature profile, typical of the majority of stations for clear sky conditions,

TABLE 1. Summary of Radiosonde Data

Location	Elevation above Mean Sea Level, feet	Months Used, Mean Monthly Air Temperature °C, Relative Humidity, and Per Cent Sunshine			
		2/64	4.6°	25%	91%
El Paso,		2/61	9.0°	38%	91%
Texas	3918	3/63	13.4°	28%	92%
		4/62	19.6°	22%	91%
		6/63	27.3°	31%	93%
		4/62	21.1°	13%	92%
Las Vegas,		12/63	7.0°	30%	94%
Nevada	2162	7/63	32.0°	10%	96%
		6/63	26.1°	20%	92%
		7/63	19.0°	20%	99%
Ely,		9/62	15.0°	23%	93%
Nevada	6257	10/62	9.0°	35%	94%
		12/62	−1.2°	50%	90%
Santa Monica,		7/64	19.6°	71%	95%
California	30	2/64	13.2°	39%	96%

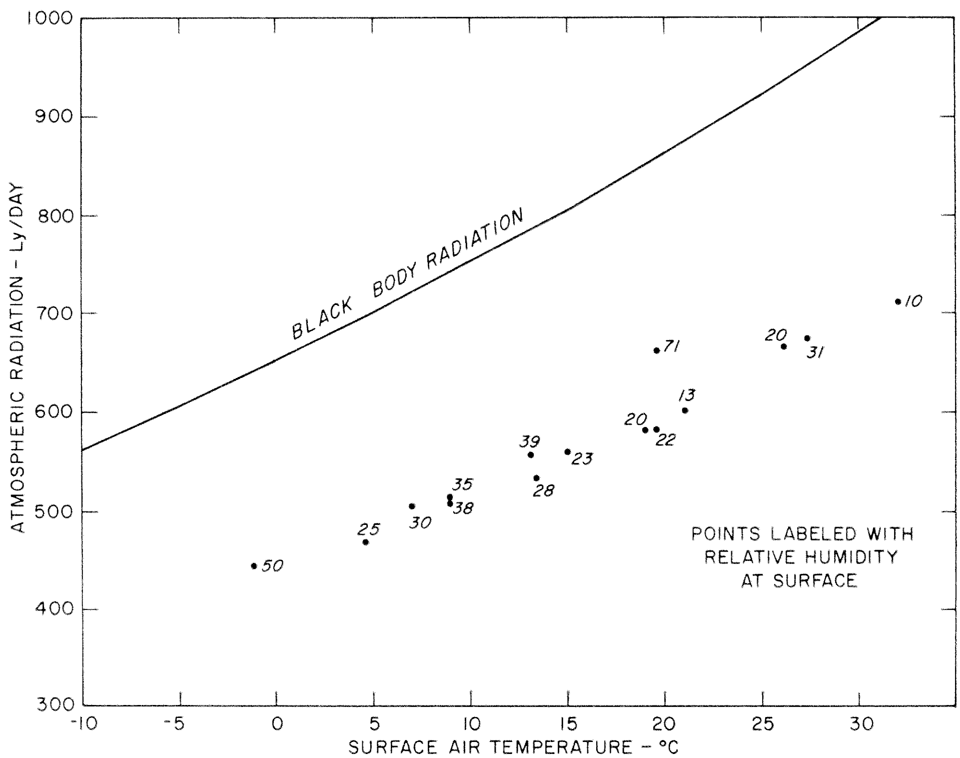

Fig. 1. Atmospheric radiation from Myers plotted against surface air temperature and relative humidity.

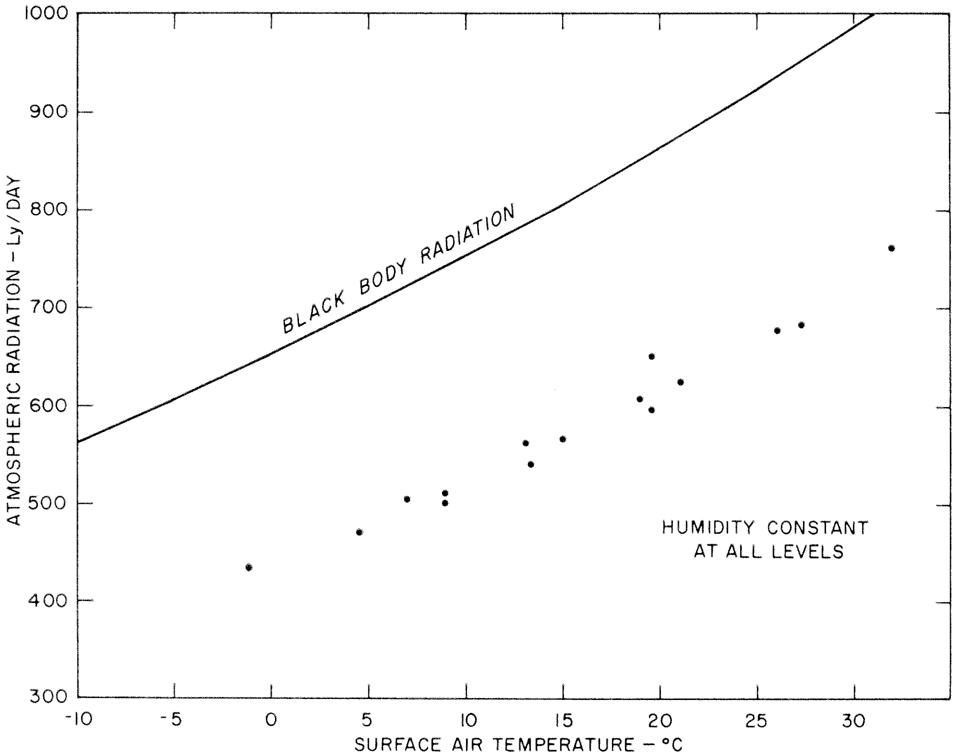

Fig. 2. Effect of temperature. Humidity profiles held constant at all stations.

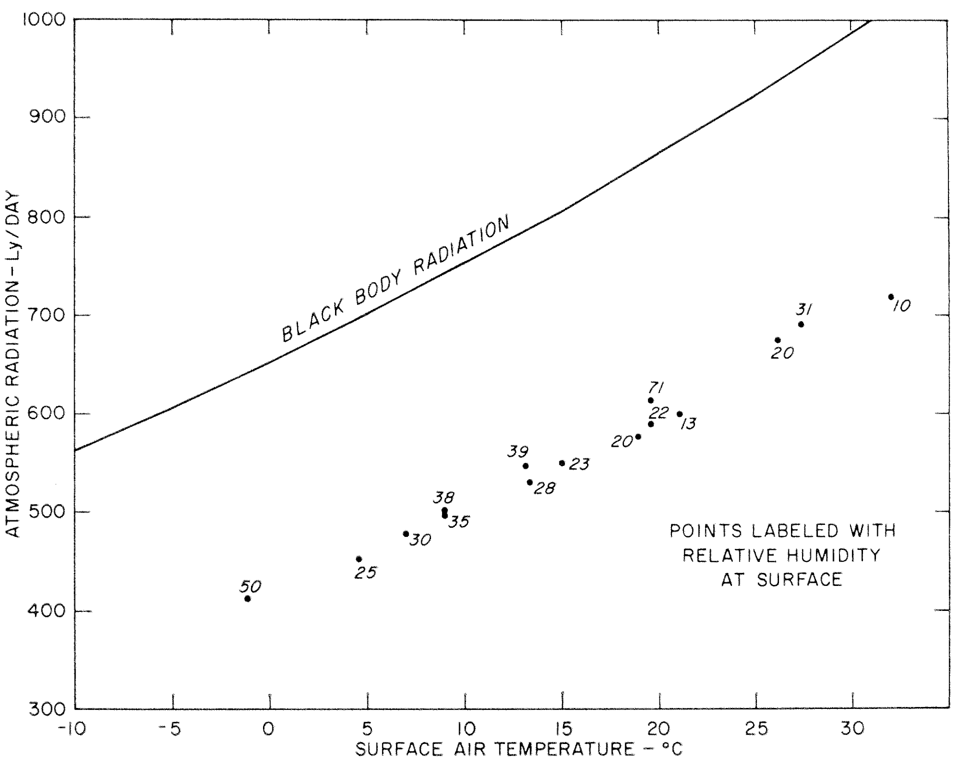

Fig. 3. Effect of water vapor. Air temperature profiles constant at all stations.

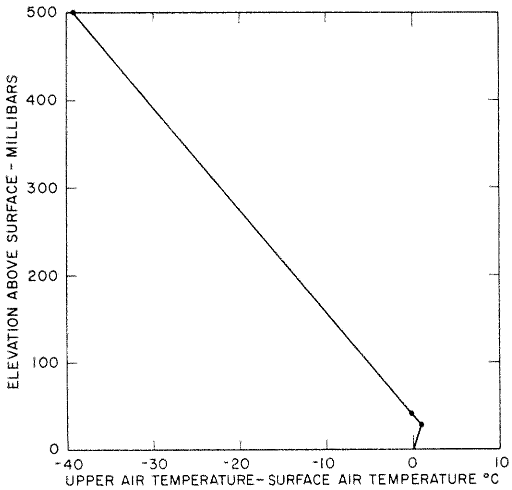

Fig. 4. Typical clear sky air temperature profile.

was selected. This typical profile is shown in Figure 4. (This is also the profile used to compute the values shown in Figure 3.)

The adjustment term was determined by plotting the atmospheric radiation difference between the typical and actual profiles versus the temperature difference between the two profiles at the surface and a given level. In choosing the upper level to use, it is best to pick a level high enough not to be dominated by surface conditions and yet low enough to be at a level that significantly affects the atmospheric radiation received at the surface. This would suggest using a level in the range of 50 to 200 millibars above the surface (75 mb and 150 mb were selected). The radiative and temperature differences between the typical and actual profiles are shown in Figure 5. Assuming a linear relationship, the best fit of the plotted

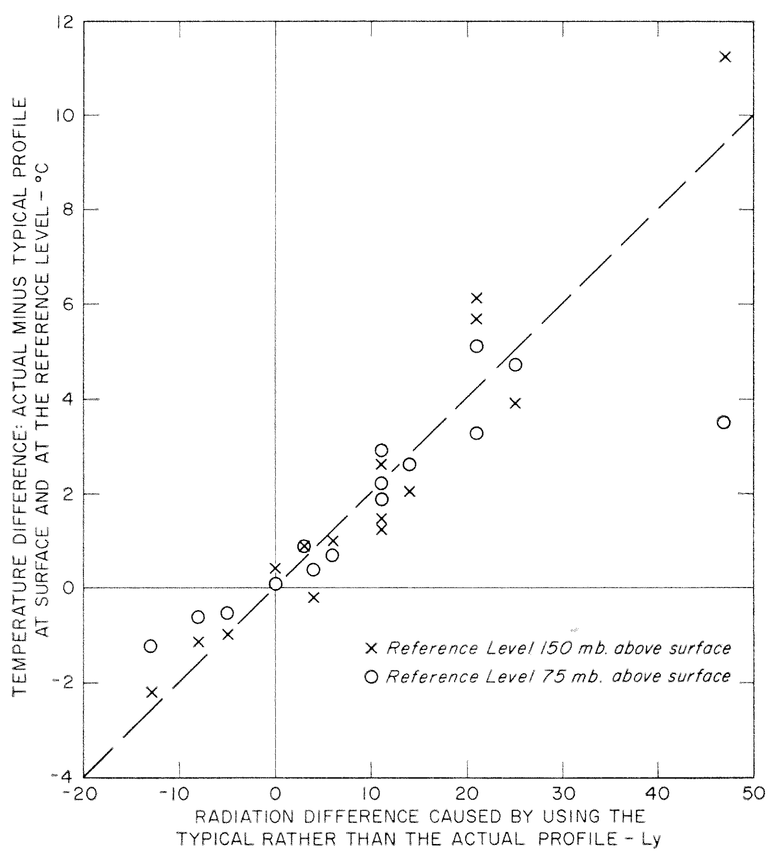

Fig. 5. Difference between using the actual and typical air temperature profiles for the computation of clear sky atmospheric radiation.

data for 150 mb above surface gives an adjustment term of

$$A = 5.0(T_{ua} - T_{ut}) \qquad (2)$$

where A is the station adjustment term in langleys per day, T_{ua} is the difference between actual upper-air and surface temperatures as determined from the long-term relationship for the site in °C, and T_{ut} is the same difference for the typical temperature profile in °C (Figure 4). It might be expected that the deviations from the typical profile would not be linear with temperature; however, in the limited amount of data used such a tendency was not apparent.

To compute the clear sky atmospheric radiation from the typical upper air temperature profile, utilizing only surface air temperature and vapor pressure, an empirical relationship was derived to fit the points in Figure 3. The resulting fit is shown in Figure 6, and the equation is

$$Q_{act} = \sigma T_a^{\,4} - 228.0 - 11.16[\sqrt{e_s} - \sqrt{e_a}] \qquad (3)$$

where Q_{act} is the atmospheric clear sky radiation for the typical upper air temperature profile in langleys per day, σ is the Stefan-Boltzmann constant (11.71×10^{-8} langleys/day/degree^4), T_a is surface air temperature in °K, e_s is the saturated vapor pressure at T_a in millibars, and e_a is the surface vapor pressure in millibars. Clear sky atmospheric radiation can be computed for any station by applying the station adjustment term to equation 3.

CLOUDY SKY RADIATION

To calculate atmospheric radiation during cloudy conditions some means of determining the extent of cloudiness and its effect on atmospheric radiation is necessary. One approach is to use amount of cloudiness. Amount of cloudiness has the advantage that it can be observed both day and night, although nighttime values are generally of doubtful quality. However, it

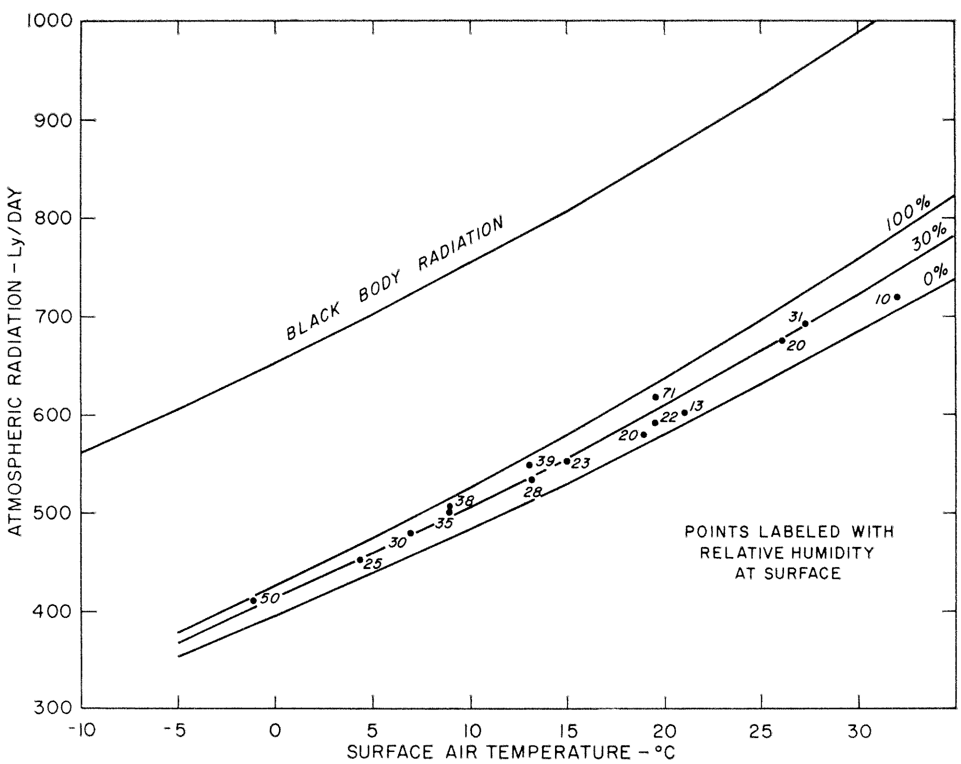

Fig. 6. The fit of equation 3 to the data of Figure 3.

has the disadvantage that equal amounts of cloudiness, as reported, do not always refer to the same degree of cloudiness. Per cent sunshine could also be used and contains essentially the same disadvantage as using amount of cloudiness. Per cent sunshine, however, is only an index to cloudiness during the day. The ratio of observed solar radiation to clear sky solar radiation is another index to the degree of cloudiness. It has the advantage that it is a good index to the mass of liquid water in the atmosphere. This ratio has the disadvantage that it is only an index during the day and is weighted toward the hours of maximum solar intensity. In this study, basically because incoming solar radiation would be needed in most energy balance computations, it is assumed to be available. Therefore, the ratio of incoming observed solar radiation Q_s to the clear sky solar radiation Q_{sc} (referred to as 'the ratio' for simplicity throughout the remainder of this text) was selected as the index to the degree of cloudiness.

Having made this decision, the assumption was made that when the ratio equaled zero (i.e., observed solar radiation is zero) the atmospheric radiation would be equal to blackbody radiation at the existing surface air temperature. Thus an upper limit for atmospheric radiation was set.

To use the ratio as an index, it is necessary to determine the clear sky solar radiation for the particular location. Two possible methods are as follows:

1. If sufficient data are available, a plot of observed incoming solar radiation versus time of year can be constructed. Assuming that the highest values occur on clear days and considering that the amount of clear sky solar radiation is not only a function of time of year but also of the amount of water vapor and suspended liquid and solid particles, a curve can be drawn through these highest values that would then provide a satisfactory estimate of clear sky solar radiation.

2. If sufficient data are not available to define reasonably the clear sky curve of method 1, atmospheric transmission under clear sky conditions (a dust-free atmosphere) can be estimated by means of a chart prepared by *Kimball* [1930]. Since in this study total solar radiation rather than direct solar radiation (for which the chart is constructed) is used, a correction has to be made for the portion of the direct solar radiation that has been scattered and diffusely reflected to the Earth. The total depletion by atmospheric scattering and diffuse reflection(s) can be expressed as

$$s = 1 - a_1 \qquad (4)$$

where a_1 is the transmission coefficient after scattering by air and water vapor molecules (from Kimball's chart). Diffuse radiation can be computed from s by allowing for the variability of both the angle of incidence and distribution in the celestial hemisphere of sky radiation. Since this would be very difficult, a simpler approximation that has been used by *Klein* [1948] and *Fritz* [1949] and was stated first by *Kimball* [1935] '. . . about half the radiation lost from the incoming rays through scattering and diffuse reflection is finally received at the ground as diffuse radiation from the sky,' has also been used in this study. Thus a simple expression for the total incoming solar radiation is obtained

$$Q_{sc} = Q_{se}(a + 0.5s) \qquad (5)$$

where Q_{se} is the extraterrestrial solar radiation and a is the transmission coefficient, which takes into account moisture absorption and molecular scattering, from Kimball's chart for transmission of direct solar radiation through moist air. These computed values of clear sky solar radiation should be compared with the observed incoming solar radiation data that are available. If it is assumed that, even in a relatively short period of record, the clearest days are in fact clear from the standpoint of atmospheric radiation, the computed and observed values should agree. This may not be the case. This disagreement can be caused by the pyranometer reading high or low, by inaccuracies in the method of computing Q_s, or by depletion due to dust (which can scatter a significant amount of solar radiation but has a negligible absorption). Thus a linear adjustment should be made, so that the observed solar radiation on the clearest days corresponds to the computed values.

To determine the relationship between the ratio and atmospheric long-wave radiation, values from the CRREL observational site at Lebanon Airport, New Hampshire, were used. These were judged to be the best data available.

A plot of the ratio versus the observed atmospheric radiation minus the lower limit value (clear sky) in per cent of range between upper and lower limits is shown in Figure 7. On several days with heavy fog, air temperature greater than 32°F, and snow on the ground, the atmosphere radiated at a rate exceeding that for a blackbody at the existing surface air temperature. The most likely explanation is that the shelter temperature was lower, because of the presence of the snow, than the air above, which controlled the incoming atmospheric radiation. Assuming the relationship to be in the form of the ratio raised to a power (n), the best fit of the plotted data was obtained using a value of n of approximately 2.0.

By combining the results of the preceding two sections, the empirical equation derived for estimating atmospheric radiation (Q_a) under all conditions is

$$Q_a = \sigma T_a^4 - \{228.0 + 11.16 \cdot [\sqrt{e_s} - \sqrt{e_a}] - A\} \bullet \{Q_s/Q_{sc}\}^n \qquad (6)$$

RESULTS

Equation 6 was tested at six locations where input data of known quality were available. These six locations are shown in Table 2. The comparisons of observed versus estimated values are shown in Figures 8 to 13. The estimated values in these figures were computed using method 2 for determining clear day solar radiation. Precipitable water, for use in Kimball's chart, was estimated from surface vapor pressure by an equation suggested by *Reitan* [1963].

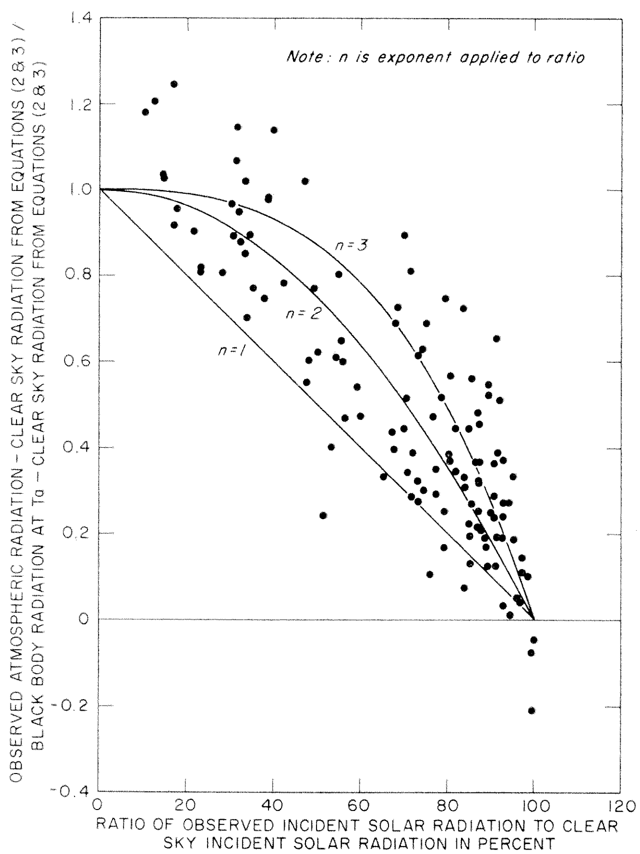

Fig. 7. Effect of observed incident solar radiation to clear sky incident solar radiation ratio on atmospheric radiation.

TABLE 2. Summary of Test Sites

Location	Elevation above Mean Sea Level, feet	Upper Air Station and Level Used To Determine Mean Relationship	Air Temperature and Vapor Pressure Measurements	Solar Radiation Measurements	Observed* Atmospheric Long-Wave Radiation	Period Tested
U. S. Army Cold Regions Research & Engineering Laboratory observation site; Lebanon Airport; Lebanon, New Hampshire	562	Average of Albany, N. Y., and Caribou, Maine	Lebanon FAA Airport	50-Jct., temp-compensated, Eppley pyranometer	Beckman Whitley total hemispherical radiometer minus solar radiation.	Sept., Oct. 1963 Nov. 1964 to Apr. 1965
Lake Hefner at Oklahoma City, Oklahoma	1200	Oklahoma City, Oklahoma	Barge Station	10-Jct., non-temperature compensated, Eppley pyranometer	Nighttime values measured with Gier & Dunkle total hemispherical radiometer. Daytime values estimated from nighttime.	July 1950 to Aug. 1951
Lake Mead, Arizona-Nevada	1180	Las Vegas, Nevada	Boulder Basin Barge and Boulder Island stations			Mar. 1952 to Sept. 1953
Central Sierra Snow Laboratory, Calif.	7170	Donner Summit, California	Lower Meadow Station 3			Apr., May 1954
ESSA-Weather Bureau, Test and Evaluation Laboratory, Sterling, Virginia	280	Washington, D. C.	WMO Evaporimeter Comparison site	10-Jct., non-temperature compensated, Eppley pyranometer correction of $-0.18\%/°C$ applied for energy balance calculations; reference temperature = 29.4°C.	Computed by energy balance calculations as outlined in the introduction.	Aug. to Nov. 1965
Silver Hill Observatory, Maryland	280	Washington, D. C.	Evaporimeter Comparison site			July 1957 to Sept. 1960

* Except at Lebanon Airport these are not actually observed values. However, for simplicity they will be referred to as such in the remainder of the text.

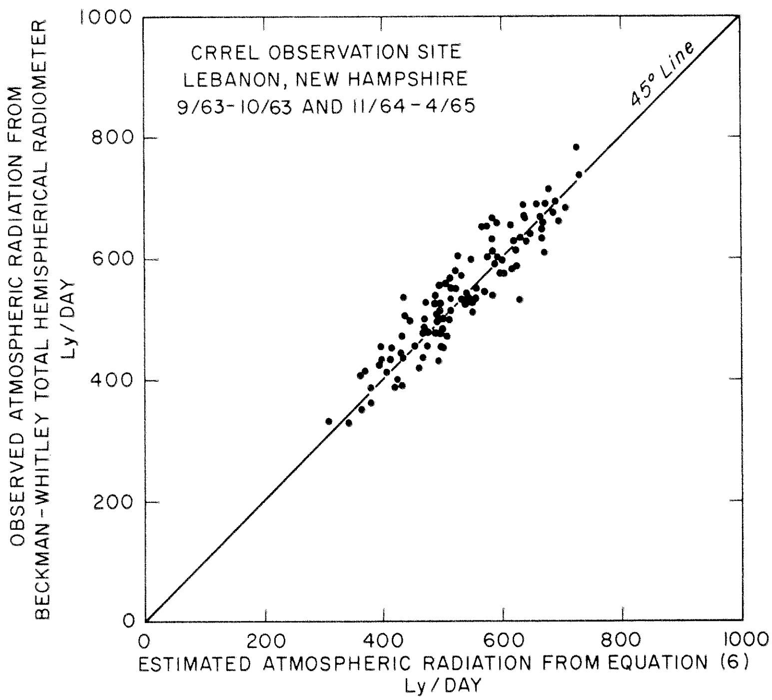

Fig. 8. Comparison of observed and estimated atmospheric radiation at Lebanon, New Hampshire.

Values for Lebanon Airport were from *Bolsenga's* [1965] evaluation of Reitan's equation. Estimated values for Lebanon Airport, Sterling, and Lake Hefner were also computed using method 1, with no significant change in the results. Values for Lake Mead were computed using both Boulder Basin Barge and Boulder Island temperatures, as given in the Lake Mead Studies [*Harbeck et al.*, 1958]. There were no significant differences, as the higher air tem-

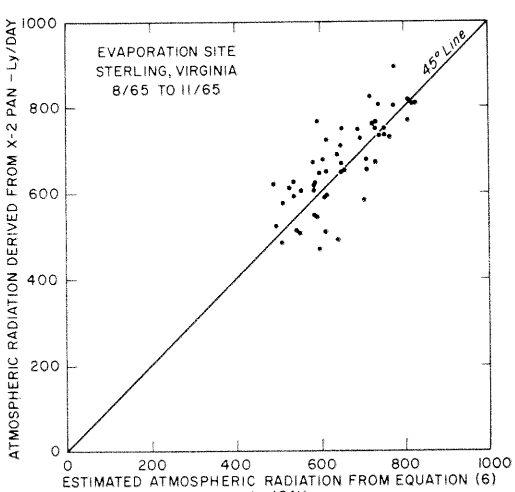

Fig. 9. Comparison of derived and estimated atmospheric radiation at Sterling, Virginia.

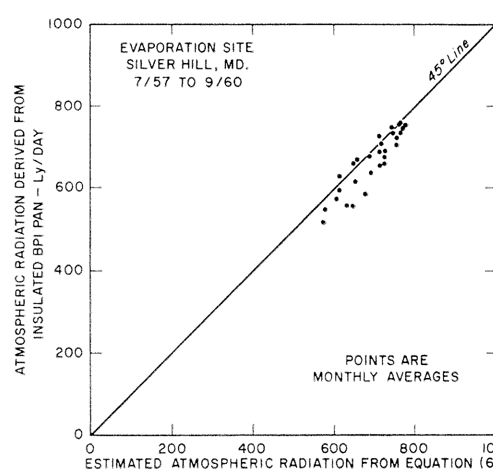

Fig. 10. Comparison of derived and estimated atmospheric radiation at Silver Hill, Maryland.

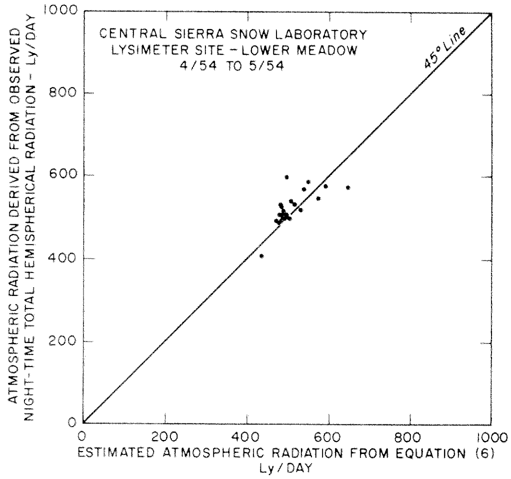

Fig. 11. Comparison of derived and estimated atmospheric radiation at Soda Springs, California.

perature at Boulder Island than at the Boulder Basin Barge was balanced by a lower dewpoint temperature.

A summary of these comparisons is given in Table 3. The over-all bias between estimated and observed values varied from +4% at Silver Hill to −3.3% at Sterling. The correlation of estimated and observed values on a daily basis resulted in correlation coefficients that varied from 0.71 to 0.92.

CONCLUSIONS

On the basis of the results obtained, the following conclusions concerning the reliabilty of equation 6 for estimating atmospheric radiation seem warranted:

1. The equation should give results comparable with observed values within a few per cent on a long-term basis (6 months or longer). A comparison with absolute values is impossible at the present time. Even with the most careful observational program, currently available radiometers will not provide atmospheric radiation data of accuracy within 5%.

2. For any specific atmospheric conditions, there seems to be no tendency for the equation to over or under compute or to give increased scatter. For example, the comparisons of observed and estimated values are similar at high and low temperatures and under overcast and clear conditions.

3. When long periods are compared (6 months or longer) a correlation coefficient of approximately 0.90 can be expected between observed and estimated daily radiation values. Correlation much higher than 0.90 is probably not possible using the present input data. For further improvements either additional upper-air data (on a daily basis), necessitating much more involved calculation, and/or a better estimate of the degree of cloudiness would need to be incorporated.

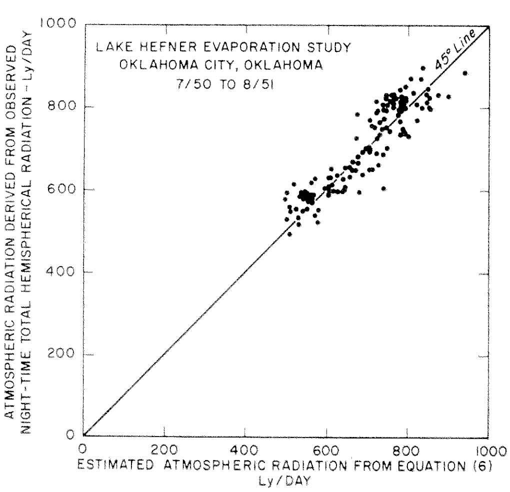

Fig. 12. Comparison of derived and estimated atmospheric radiation at Oklahoma City, Oklahoma.

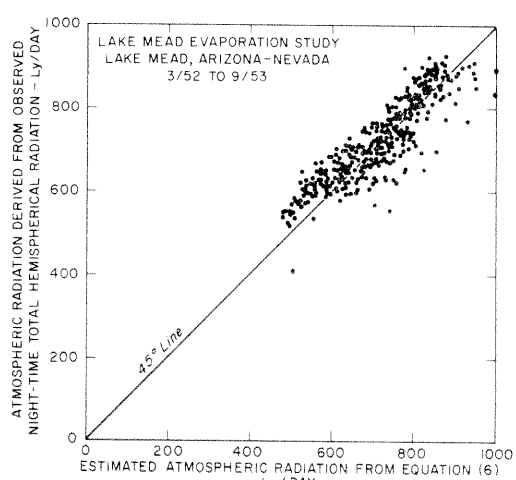

Fig. 13. Comparison of derived and estimated atmospheric radiation at Lake Mead.

TABLE 3. Summary of Comparisons

Location	Difference between Estimated and Observed Values for the Period of Record	Daily Correlation Coefficient
Lebanon Airport	−2.4%	0.92
Lake Hefner	−1.4%	0.91
Lake Mead	−1.9%	0.89
Central Sierra Snow Laboratory	−1.3%	0.71
Sterling	−3.3%	0.75
Silver Hill	+4.0%	0.73

4. The scatter in the results from the two sites (Silver Hill, Maryland, and Sterling, Virginia) where insulated evaporation pans were used to compute atmospheric long-wave radiation was greater than at other sites, where observed values were based on radiometer measurements. This should be expected from the Silver Hill, Maryland, site for two reasons: (a) Minimum water temperatures were used to estimate the change in heat storage on a daily basis, because measurements at time of observation were not available; and (b) the assumption used in the energy balance calculations of negligible heat flow through the sides and bottom of the pan seems to be doubtful, because there is a seasonal shift.

The data from X-2 experimental pan at Sterling, Virginia, are for a relatively short period of record and also include several days when the water temperature was near or below 39°F. On these days water temperature at one level in the pan may not be sufficient to estimate the change in heat storage due to layering. Further analyses will be made of subsequent X-2 pan data at Sterling and also for the X-2 pan installed at Lake Mead, Arizona-Nevada, and the University of California at Davis, California.

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