# 7      Soil Heat Flux

**THOMAS J. SAUER**
*USDA-ARS National Soil Tilth Laboratory*
*Ames, Iowa*

**ROBERT HORTON**
*Iowa State University*
*Ames, Iowa*

Climatic conditions on the earth's surface are in part a function of varying physical position (elevation, latitude, and aspect) and the influence of large-scale meteorological forces such as air and ocean currents. The density and architecture of plant canopies in natural systems are directly influenced by these climatic factors. By contrast, for agricultural systems, it is the crop canopies that often influence local microclimate. In both instances, the soil plays an important role in affecting climate near the surface. Properties of the surface soil layer including color, water content, texture, and density affect the partitioning of incident radiation and how much energy is used to evaporate water, warm the air above the ground, or warm the soil.

The amount of thermal energy that moves through an area of soil in a unit of time is the soil heat flux or heat flux density. The ability of a soil to conduct heat determines how fast its temperature changes during a day or between seasons. Soil temperature is a key factor affecting the rate of chemical and biological processes in the soil essential to plant growth. Soil heat flux is important in micrometeorology because it effectively couples energy transfer processes at the surface (surface energy balance) with energy transfer processes in the soil (soil thermal regime). This interaction between surface and subsurface energy transfer processes has led to detailed investigations of soil heat flux for a wide variety of agricultural systems.

## SURFACE ENERGY BALANCE AND SOIL HEAT FLUX

In micrometeorology, measurement of soil heat flux is often considered within the context of the surface energy balance

$$R_n - G = LE + H \qquad [1]$$

where $R_n$ is the net radiation, $G$ is the soil heat flux density at the soil surface, and $LE$ and $H$ are the latent and sensible heat flux densities, respectively. All terms in

Eq. [1] have units of J m$^{-2}$ s$^{-1}$ or W m$^{-2}$. Note that in Eq. [1] all fluxes away from the soil surface are defined as positive except for $R_n$. The left side of Eq. [1], ($R_n$ – $G$), represents the available energy while the terms on the right side (*LE* and *H*) are often referred to as the turbulent fluxes. Much of the energy that enters the soil during the day returns to the atmosphere at night through terrestrial longwave radiation. For this reason, $G$ is often the smallest component of the daily surface energy balance and has, in some cases, been ignored; however, there are often significant transfers of energy into and out of a soil during both day- and night-time hours and failure to include $G$ in short-term (i.e., hourly) energy balance determinations can lead to sizeable errors.

Comprehensive surface energy balance studies have been conducted since the 1950s. Lettau and Davidson (1957) and Lemon (1963) are early examples of these types of studies. In recent years, with technical advancement of ground-based and remote sensing instrumentation, surface energy balance measurements have become much more common. The spatial scale of energy balance studies also has expanded with advancing sensor technology. For example, interest in global climate change has prompted several efforts to estimate the earth's annual mean energy budget (e.g., Ohmura & Gilgen, 1993; Kiehl & Trenberth, 1997). This trend is evidenced by the development of the Global Energy Balance Archive (GEBA) for documentation of current climatic conditions and facilitation of the study of past and future climate (Gilgen & Ohmura, 1999).

Examples of measurements of $G$ either alone or as components of energy balance studies can be found for a wide range of agricultural land use practices. Data on $G$ within various micrometeorological investigations have been reported for forests (Stewart & Thom, 1973; McCaughey, 1982; Oliver et al., 1987; McCaughey & Saxton, 1988; Tamai et al., 1998), orchards and vineyards (Fritton et al., 1976; Fritton & Martsolf, 1980; Glenn & Welker, 1987; Heilman et al., 1994), grasslands (Rosset et al., 1997; Bremer & Ham, 1999; Twine et al., 2000), small grains (Lourence & Pruitt, 1971; Choudhury et al., 1987; Kimball et al., 1999), row crops (Brown & Covey, 1966; Ham et al., 1991; Ham & Kluitenberg, 1993), and reclining sheep (*Ovis aries*, Gatenby, 1977). Energy balance studies with measured or estimated $G$ for non-crop surfaces include those for sparse canopies (Tuzet et al., 1997; Verhoef et al., 1999; Kustas et al., 2000), sloping terrain (Oliver, 1992), and bare soils (Fuchs & Hadas, 1972; Rao et al., 1977; Enz et al., 1988). While the number of studies and variety of soils and surface covers examined have resulted in a wealth of data concerning soil heat flux under field conditions, they also attest to the complexity of energy balance relationships. The intricate relationships between terms in Eq. [1], as affected by such factors as soil and air temperature, soil water content, canopy characteristics, residue cover and wind speed, can have significant impact on the direction and magnitude of the fluxes.

The magnitude of $G$ as a component of the surface energy balance varies with surface cover, soil moisture content, and solar irradiance. Daytime peak hourly values of $G$ for a bare, dry soil in midsummer could be in excess of 300 W m$^{-2}$ (Fuchs & Hadas, 1972). By contrast, hourly $G$ for a moist soil beneath a plant canopy, residue layer, or snow cover will often be less than ±20 W m$^{-2}$. Surface soil heat flux typically represents 1 to 10% of $R_n$ for growing crops (Den-

mead, 1969; Szeicz et al., 1973; Brown, 1976; Uchijima, 1976; Baldocchi et al., 1985; Clothier et al., 1986). This percentage can exceed 50% in the fall and spring when $R_n$ is low and the soil is cooling/warming or in arid climates when there is no vegetation (Monteith, 1958; Idso et al., 1975; Choudhury et al., 1987). Figure 7–1 depicts the partitioning of energy balance terms for a corn (*Zea mays* L.) residue-covered soil in central Iowa (Sauer et al., 1998). The data in Fig. 7–1 were obtained in November when the soil and fresh residue layer were dry and the mean surface temperature was 6.8°C. Daytime *G* averaged 14.4% and 16.8% of $R_n$ for Days 310 and 311, respectively.

Measurement or prediction of evaporation is often of great interest as it relates to studies of water balance and water management in agricultural systems. Soil heat flux, as a component of the available energy, is a necessary input for many evaporation measurement and prediction techniques. Evaporation measured with the Bowen ratio energy balance approach (Bowen, 1926), for instance, is dependent on an accurate value for the available energy ($R_n - G$). The impact of errors in *G* on turbulent fluxes determined using the Bowen ratio method is discussed by Malek (1993) and de Silans et al. (1997). Several of the more common equations for predicting evaporation such as the Penman-Monteith (Penman, 1948; Monteith, 1965) and Priestley-Taylor (Priestley & Taylor, 1972), also require available energy as an input. The effect of omitting *G* on evaporation estimates will depend on local climate, soil properties, and cropping system. Although failure to include soil heat flux may introduce relatively small (i.e., $<\sim 10\%$) errors in available energy, exclusion of *G* during the summer can lead to systematic overestimation of available energy and subsequent overestimation of evaporation (Simmers, 1977; Kumar & Rao; 1984; Anadranistakis et al., 1997).
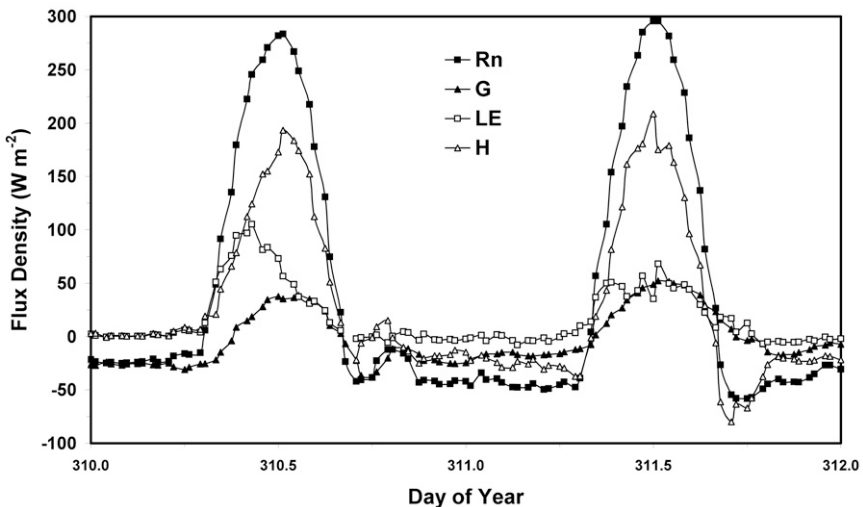


Fig. 7–1. Diurnal patterns of energy balance terms for a no-till corn field in central Iowa (Sauer et al., 1998). The corn residue layer was 0.05 m-thick. Soil heat flux density was measured at 0.05 m using the heat flux plate method and corrected for heat storage in the soil above the plate.

## SOIL THERMAL REGIME AND SOIL HEAT FLUX

In most instances, conduction is the principal mode of energy transport in soils, although radiation and convection in very shallow layers also may transfer energy. Heat flow in soil can be considered analogous to heat flow in a solid to which Fourier's Law is applied

$$G = -\lambda \partial T/\partial z \qquad [2]$$

where $\lambda$ is the thermal conductivity of the soil (W m$^{-1}$ K$^{-1}$) and $\partial T/\partial z$ is the vertical temperature gradient (K m$^{-1}$) of the soil layer. Fourier's Law is defined for solid, homogeneous materials under steady-state conditions with thermal conductivities that are essentially constant over small temperature ranges. Fourier's Law is easily and directly used in many engineering applications, however, in a porous, three-phase medium like soil, use of Eq. [2] is considerably more difficult.

The thermal conductivity of soil varies by composition of the solid fraction (mineral type, particle size, and amount of organic matter), water content, and bulk density (de Vries, 1963; Al Nakshabandi & Kohnke, 1965; Abu-Hamdeh & Reeder, 2000). These properties often vary between soils, spatially at the soil surface for the same soil, between layers within a soil, and over time. Al Nakshabandi and Kohnke (1965), for instance, measured an eight-fold increase in the thermal conductivity of a silt loam soil as its gravimetric water content increased from oven dry to 0.34 kg kg$^{-1}$. A temperature gradient in the soil also induces water flow via evaporation and condensation, which include the concomitant transfer of latent energy (Philip & de Vries, 1957; Cahill & Parlange, 1998). Solar radiation is the ultimate driving force behind soil heat transfer in most field settings. Thus, not only are soil thermal properties dynamic, but superimposed on these complex relationships are annual and diurnal patterns of solar radiation including irregularities in weather patterns.

Agricultural management practices including irrigation, drainage, and tillage have the potential to affect the thermal properties of soils and therefore soil thermal regime. In particular, the effect of tillage and crop residue management on soil heat flux has been the subject of several studies (Allmaras et al., 1977; Pikul et al., 1985; Enz et al., 1988; Azooz et al., 1997; Richard & Cellier, 1998; Sauer et al., 1998). Tillage loosens the surface soil, although some local compaction also may occur. Lower soil bulk density generally translates to lower $\lambda$, thus, lower $G$ has been observed in tilled soil as compared with un-tilled or compacted soil (Azooz et al., 1997; Richard & Cellier, 1998). Crop residue has a low thermal conductivity and, whether lying on the soil surface or incorporated into the soil by tillage, may inhibit heat transfer into the soil. Residue layers also have a shortwave reflectivity that is higher than most soils and provide a barrier to vapor flow (Gausman et al., 1975; Horton et al., 1996; Sauer et al., 1997). Thus, soils with a large proportion of the surface covered by crop residue tend to have higher water contents, lower temperatures, and lower $G$. Such changes in soil thermal regime, of course, have implications for the surface energy balance and evaporation.

# MEASUREMENT TECHNIQUES

Most soil heat flux density measurements have been completed using one of four methods (flux plate, calorimetric, gradient, or combination). One other method that has been developed but not been widely used is the block method (van Wijk, 1967). Measurement of soil heat flux density involves the measurement of soil temperature and one or more soil thermal properties (thermal conductivity or heat capacity), and possibly soil water content. Techniques for measurement or estimation of these parameters are discussed in detail elsewhere (e.g., de Vries, 1963; Goel & Norman, 1990; Smith & Mullins, 1991; Carter, 1993; Topp & Dane, 2002).

An area of current interest in soil heat flux measurement is the spatial variability of $G$ under field conditions. Concern is heightened in energy balance studies where measurements of $LE$ and $H$ by eddy covariance or Bowen ratio techniques are typically representative of an area of 100s of m$^2$. By comparison, $R_n$ (measured with a net radiometer) may be representative of an area of 10s of m$^2$; however, a single measurement of $G$ is representative of a much smaller area, perhaps ~0.1 m$^2$. Several studies (McCaughey, 1982; Ham & Kluitenberg, 1993; Tuzet et al., 1997; Kustas et al., 2000) have shown that spatial variation of $G$ under field conditions can be significant. This is especially true for sites with sparse canopies or uneven surfaces as both shading patterns and microtopography introduce large variation in soil temperature and soil water content. Kustas et al. (2000), found variation in $G$ (measured at 0.08 m depth) between adjacent locations with similar cover in a dune with an uneven surface and partial shrub cover to be greater than 200 W m$^{-2}$. In view of the evidence on spatial variability of $G$, measurement of soil heat flux density at multiple locations is necessary to obtain a spatially-representative value of soil heat flux during energy balance studies for agricultural surfaces, especially partial canopies.

Previous reviews of soil heat flux measurement techniques include Staley and Gerhardt (1957), Carson and Moses (1963), Tanner (1963), Jackson and Taylor (1965), Kimball and Jackson (1979), Fuchs (1986), and Sauer (2002). With the exceptions of Staley and Gerhardt (1957) and Carson and Moses (1963), each of these summaries focused on the measurement of $G$ as it pertains to the characterization of soil thermal regime. Although the soil thermal regime and surface energy balance are inter-dependent, emphasis here will be placed on measurement of $G$ with regard to micrometeorological applications (surface energy balance and evaporation).

## Flux Plate Method

Most recent studies of soil heat flux density have used heat flux plates (also called heat flow meters or heat flow transducers). This trend is probably due to the comparative ease of the flux plate approach. The concept of a soil heat flux plate was adapted from efforts to measure heat transfer in walls of buildings and bulkheads of ships. Falckenberg (1930) is credited as the first to apply this approach specifically for measuring heat transfer in soil. Contributions to the advancements in theory and design of soil heat flux plates have been made by

Dunkle (1940), Deacon (1950), Portman, (1958), Philip (1961), Fuchs and Tanner (1968), and Mogensen (1970).

Soil heat flux plates are small, rigid, wafer-shaped sensors that are placed horizontally into the soil. The plates make a direct measurement of heat flux density that is proportional to the heat flux density in the soil. Most designs employ an encapsulated thermopile, which produces an electromotive force (emf) in response to a temperature gradient across the plate created by vertical heat flow through the sensor body. Two designs of soil heat flux plates do not use a conventional thermopile as the sensing element. One (Weaver & Campbell, 1985) uses a Peltier cooler and the other (Herin & Théry 1992; Robin et al.,1997) uses printed circuit technology. The signal from a plate that has been calibrated under conditions with known heat flow can then be used to determine the soil heat flux density at the depth of plate placement.

In spite of the simplicity and wide acceptance of the heat flux plate method, several areas of concern surround this approach: (i) heat flow convergence/divergence around the plate, (ii) water flow divergence (including water vapor), (iii) thermal contact between the plate and soil matrix, and (iv) accounting for heat storage in the layer(s) between the plate and the soil surface.

The presence of an impervious plate near the soil surface creates concern regarding convergence/divergence of heat flow, thereby biasing estimates of $G$ obtained from the plate readings. Philip (1961) recommended several factors to consider in plate design to minimize perturbations in vertical heat flow based on theoretical analysis of heat flow near a small oblate spheroid of known thermal properties. An equation was derived to predict the ratio of heat flow in the soil to that through the plate:

$$G_m / G = \varepsilon/[1 + (\varepsilon - 1)H] \qquad [3]$$

where $G_m$ is the heat flux density through the plate, $\varepsilon$ is the ratio of the plate thermal conductivity to the soil thermal conductivity ($\lambda_m/\lambda$), and $H$ is an empirical factor related to plate shape:

$$H = 1 - (\beta b) \qquad [4]$$

where $\beta$ is a dimensionless geometric constant dependent on plate shape and $b$ is the plate thickness divided by side length for a square plate or plate thickness divided by plate diameter for a circular plate. Mogensen (1970) tested Eq. [3] in laboratory experiments with a small circular plate and found that the heterogeneous composition of the plate made it difficult to verify Philip's theory. Watts et al. (1990) confirmed and refined Philip's recommendations to conclude that: (i) the plate thickness to width ratio should be small, (ii) dry and saturated sand are suitable calibration media for mineral soils, and (iii) the thermal conductivity of the plate should be greater than 0.5 W m$^{-1}$ K$^{-1}$.

Figure 7–2 illustrates the effect of heat flux plate thickness and thermal conductivity on the ratio of the heat flux through the plate to the heat flux through the soil ($G_m/G$) as predicted by Eq. [3]. The curves in Fig. 7–2 are for hypothetical circular plates that are 50 mm in diameter, 2, 4, or 6 mm-thick, and have ther-
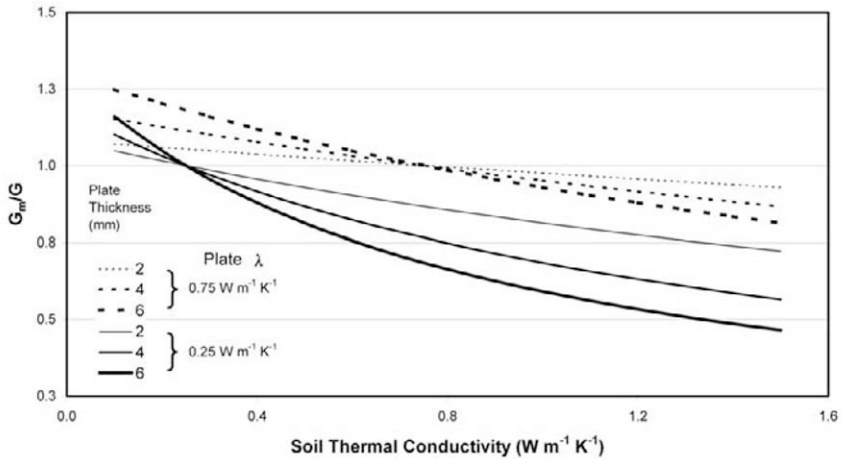
Fig. 7–2. Estimated ratio of soil heat flux density measured with 50 mm-diameter round plates 2-, 4-, and 6-mm-thick with thermal conductivity of either 0.25 or 0.75 W m$^{-1}$ K$^{-1}$ ($G_m$) to actual soil heat flux density in soil ($G$) with λ from 0.2 to 1.5 W m$^{-1}$ K$^{-1}$ as predicted by Eq. [3].

mal conductivity ($\lambda_m$) of 0.25 or 0.75 W m$^{-1}$ K$^{-1}$. As the difference between the plate and soil λ increases, increasingly large errors (up to –50%) in measured $G$ are predicted. Clearly, using a heat flux plate with λ similar to the expected soil λ or within a range common for mineral soils (~0.4–1.2 W m$^{-1}$ K$^{-1}$) would reduce errors associated with heat flow convergence/divergence through the plate. The curves in Fig. 7–2 also indicate that there is less error associated with thinner plates. For example, if a soil had a λ of 1.0 W m$^{-1}$ K$^{-1}$, a 2 mm-diameter plate with a $\lambda_m$ of 0.25 W m$^{-1}$ K$^{-1}$ would have 22% less error in measured $G$ than a 6 mm-diameter plate with the same $\lambda_m$. It should be noted that Philip's analysis assumes that the flux measurement is made across the entire plate area, while several current plate designs include a "guard" area surrounding a thermopile located in the center of the plate. Also, this analysis is only for heat conduction and does not include any provision for the effects of energy transfer due to liquid water movement near a plate.

The presence of an impervious object like a heat flux plate near the soil surface can affect water flow in the soil in several ways. During and after precipitation events, a heat flux plate can impede water movement downward in the soil and result in a higher water content in the soil immediately above the plate and lower water content immediately below. As evaporation dries the soil, the plate also can prevent upward water movement so the soil above the plate can become drier than the surrounding soil while the soil below the plate remains wetter. When soil immediately adjacent to the plate is at a water content that differs from the surrounding soil, it will also have different thermal properties. For a typical silt loam soil, a change in water content from 0.15 to 0.20 kg kg$^{-1}$ can increase its λ from 0.96 to 1.3 W m$^{-1}$ K$^{-1}$ and its volumetric heat capacity ($C$) from 1.75 to 1.96 MJ m$^{-3}$ K$^{-1}$ (de Vries, 1963; Al Nakshabandi & Kohnke, 1965). Unrepresentative soil water content near a heat flux plate will affect the soil thermal proper-

ties and could result in sizable errors in measured $G$ due to altered heat flow in the vicinity of the plate.

If a soil heat flux plate is positioned above the drying front, energy is consumed through the latent heat of vaporization (i.e., evaporative cooling) in the soil below the plate. This loss of energy may not be measured by the plate as the source of the energy may be from deeper in the soil, nor is it accounted for in the calorimetric heat storage correction, which is completed only for the soil layer above the plate. De Vries and Philip (1986), Buchan (1989), and Mayocchi and Bristow (1995) showed that large errors in $G$ (up to 100 W m$^{-2}$) can occur when latent heat loss during periods of high evaporation from bare soil is ignored; however, such large errors are likely only when the plate is positioned near the surface (<0.02 m), evaporation is high, and the drying front is abrupt. If a plant canopy is present, there is a diffuse drying front, or the flux plate is at 0.05 m depth or greater, the error due to latent heat loss should not be significant.

Thermal contact between the plate and soil depends on soil texture, structure, water content, and care in installation of the plate. Fuchs and Hadas (1973) compared laboratory and field calibrations of soil heat flux plates and concluded that contact resistance was the largest source of error in the field measurements. To avoid errors associated with contact resistance, the calibration media should be selected so that the thermal contact is similar to that of the soil where the plate is to be installed. Alternatively, plate calibrations could be completed in situ by comparing $G$ determined with flux plates with $G$ determined by one of the other methods.

As the heat flux plate measures $G$ only at the depth of placement, the calorimetric method (discussed in the next section) has been used to account for the change in heat storage in the soil above the plate (Fuchs & Tanner, 1968). Failure to account for heat storage in the soil above the plate can result in large errors. Mayocchi and Bristow (1995) reported errors as large as 80 W m$^{-2}$ in daytime, half-hour average $G$ when the change in heat storage was ignored in a sugar cane (*Saccharum officinarum* L.) field in Australia. Massman (1992, 1993) concluded that the standard calorimetric correction may itself have errors of ±3 to 10% when assuming that the change in temperature of the soil layer above the plate is well-approximated by the average temperature at the midpoint. de Silans et al. (1997) present an analytical method for determining $G$ at the surface from $G$ measured by a plate and temperature at the soil surface and at the plate depth. This approach requires no knowledge of soil thermal properties but does require harmonic analysis of multiple days of steady, periodic temperature and heat flux waves.

## Sensors

Deacon (1950), Tanner (1963), and Fuchs and Hadas (1973) discuss soil heat flux plate design and construction. In general, plate design is dictated by the requirements that the plate be thin, watertight, have a $\lambda_m$ comparable to the soil being monitored, and that the emf produced is high enough to be easily measured. To assure good thermal contact with the soil, exterior materials with a high thermal admittance are desirable. Thermal admittance ($A_T$) is defined as

$$A_T = (\lambda_m \, C_m)^{0.5} \qquad\qquad\qquad\qquad [5]$$

where $C_m$ is the heat capacity of the plate (Fuchs, 1986). To obtain a high $A_T$, several plate designs have used metal and/or anodized metal shields on their exteriors.

Due to the popularity of the flux plate technique, several sensors of varying dimensions, thermal properties, and sensitivities are now commercially available (Table 7–1). Factors to consider when selecting a soil heat flux plate include the soil thermal and water regimes being monitored, and desired depth of placement. Plates with larger areas and lower $\lambda_m$ positioned at 0.025 to 0.05 m depth may be suitable for $G$ measurements with dry soils in arid environments as the soil $\lambda$ and latent heat loss are likely to be relatively low. Smaller plates with higher $\lambda_m$ positioned at 0.05 to 0.1 m may be preferable at a humid site where more frequent rainfall events keep the soil moist so the soil $\lambda$ is higher and the drying front is above the plate depth.

## Procedure

Figure 7–3 illustrates one arrangement of sensors for application of the heat flux plate method. A shallow excavation to a depth below the desired depth of plate placement is made followed by a horizontal slit slightly smaller than the plate dimensions in a side wall. The plate should be carefully inserted into the slit so that the plate faces are parallel to the soil surface and there is good thermal contact with the soil on all sides. Depth of placement for soil heat flux plates is typically 0.025 to 0.1 m. At depths <0.025 m, there is concern that the soil may crack when dry thereby exposing the plate to solar radiation or creating poor thermal contact between plate and soil. Multiple plates in the same excavation or multiple excavations with single plates may be necessary to adequately represent the site being monitored. Distance between adjacent plates should be at least double the largest dimension (diameter or side length) of the plate face (Watts et al., 1990). Distribution of measurement sites should reflect any spatial heterogeneity

Table 7–1. Specifications of some commercially available soil heat flux plates.

| Model[†] | Shape | Dimensions (L × W or diameter) mm | Thickness mm | Thermal conductivity W m⁻¹ K⁻¹ | Sensitivity µV/ W m⁻² |
|---|---|---|---|---|---|
| CN3[‡] | rectangular | 48 × 29 | 7 | 0.4 | 21 |
| MF-81[§] | rectangular | 110 × 12 | 4 | 0.23 | 26 |
| HFP01SC[¶] | circular | 80 | 5 | 0.8 | 50[¶¶] |
| GHT-1C[#] | square | 52 × 52 | 5.7 | 0.26 | 900 |
| HFT-1[††] | circular | 38 | 3.9 | 1.0 | 24 |
| 610[‡‡] | circular | 25 | 2.6 | 0.33 | 7.5 |
| WS 31S[§§] | circular | 110 | 5 | 0.2–0.3 | 100 |

[†] Names are necessary to report factually on available data; however, the USDA neither guarantees nor warrants the standard of the product, and the use of the name by USDA implies no approval of the product to the exclusion of others that also may be suitable.
[‡] Carter-Scott Manufacturing Pty. Ltd., Brunswick, Victoria, Australia.
[§] EKO Instruments Trading Co., Ltd., Tokyo, Japan.
[¶] Hukseflux Thermal Sensors, Delft, the Netherlands.
[#] International Thermal Instrument Co., Del Mar, CA.
[††] Radiation and Energy Balance Systems, Seattle, WA.
[‡‡] C.W. Thornthwaite Associates, Pittsgrove, NJ.
[§§] TNO Institute of Applied Physics, Delft, the Netherlands.
[¶¶] This sensor can be used in a self-calibrating mode.
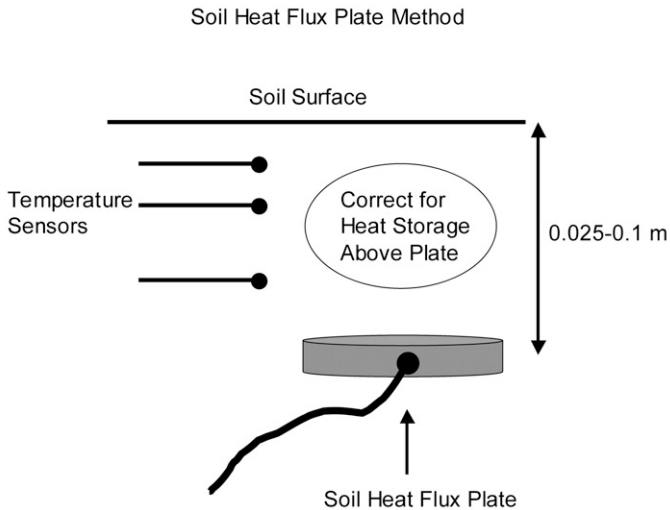
Soil Heat Flux Plate Method



Fig. 7–3. Arrangement of sensors and measurements necessary for application of the heat flux plate method to determine soil heat flux at one location in a homogenous soil.

induced by shading or microtopography. A length of sensor wire near the plate should be buried at the same depth as the plate to reduce the risk of heat conduction down the wire to/from the sensor.

Soil temperature above the heat flux plates is often measured with thermocouples or thermistors to facilitate the calorimetric heat storage correction (Fuchs & Tanner, 1968). In general, one to four measurement depths are used depending on the depth of the plate and uniformity of soil properties. Ideally, the temperature sensor profile should be located adjacent to each plate, however, depending on plate spacing and spatial variability of soil properties, the same temperature profile could be used for the heat storage correction of multiple plates in near proximity. McInnes (2002) provides detailed information regarding soil temperature measurements.

### Soil Heat Flux Plate Calibration

Soil heat flux plates have been calibrated in situ against $G$ determined by the calorimetric (Hanks & Jacobs, 1971; Högström, 1974; Robin et al., 1997) and gradient methods (Kimball et al., 1976; van Loon et al., 1998). An in situ calibration should reduce any errors associated with thermal contact resistance (Fuchs & Hadas, 1973). Such calibrations also provide the best opportunity for obtaining accurate calibration data at varying soil moisture content; however, an in situ calibration involves a considerable investment in time and resources and measurement errors associated with the method used as the standard must be considered. One recent development in sensor technology is a heat flux plate that includes a heater for independent, in situ calibrations (van Loon et al., 1998).

Several other techniques for soil flux plate calibration have been proposed including a radiation technique (Idso, 1972) and several designs of steady-state lab-

oratory apparatus (Fuchs & Tanner, 1968; Biscoe et al., 1977; Howell & Tolk, 1990; Watts et al., 1990; van Loon et al., 1998). The laboratory techniques involve placing the plates in a porous medium (generally quartz sand) inside a well-insulated box where a known, 1-dimensional heat flux is maintained across the media and through the plates. Because establishment of a temperature gradient across moist sand creates a moisture gradient and latent heat transfer, accurate calibrations using this technique can be completed only under dry and saturated conditions.

## Advantages and Disadvantages

The primary advantages of the flux plate method are that the plates are relatively inexpensive, reliable, and can be used continuously in the field for extended periods. For these reasons, the flux plate method has become the most popular technique for measuring $G$, especially in surface energy balance applications. Nonetheless, significant errors in measured $G$ are likely unless certain precautions are taken and/or corrections made:

1.  Plate thermal properties should be matched with the environmental conditions being monitored. As previously noted in Fig. 7–2, errors as large as 50% are estimated from Eq. [3] when using plates with thermal conductivity grossly different than the surrounding soil. Plate thickness and face area along with near-surface soil water dynamics also should be considered when choosing the depth of placement.
2.  Plates should be carefully calibrated and installed. Fuchs and Hadas (1973) measured a 27% difference in sensitivity for a flux plate calibrated both in the field and under controlled laboratory conditions. Most of this error was attributed to differences in thermal contact resistance with the soil.
3.  Sufficient numbers of plates should be installed to obtain a spatially-representative estimate of $G$. Especially with partial canopies and uneven soil surfaces, spatial variation of $G$ can be significant necessitating the careful placement of multiple sensors to obtain an accurate areal average.
4.  Correction for heat storage above the plate must be made. Figure 7–4 shows measured $G$ at 0.05 m (average of three heat flux plates) at 0.5-h intervals with and without correction for heat storage. Although the daily sum of $G$ for the corrected flux is only 3.3% greater than the uncorrected flux, the heat storage correction results in significantly greater $G$ during the daytime and lower $G$ at night. At 10:00 a.m. Central Standard Time (1000), the difference between the corrected and uncorrected $G$ is 52 W m$^{-2}$ (120–68 W m$^{-2}$). If the Bowen ratio energy balance method were being used to determined the turbulent fluxes, failure to correct for heat storage would lead to a 52 W m$^{-2}$ (14%) overestimate of the available energy ($R_n - G = 377$ W m$^{-2}$ instead of 325 W m$^{-2}$). This error would lead to a subsequent underestimate of $LE + H$ by the same amount.
5.  Correction for latent heat loss may be necessary for shallow plates above an abrupt drying front. Buchan (1989) and Mayocchi and Bristow (1995) have shown that errors up to 100 W m$^{-2}$ can occur if latent
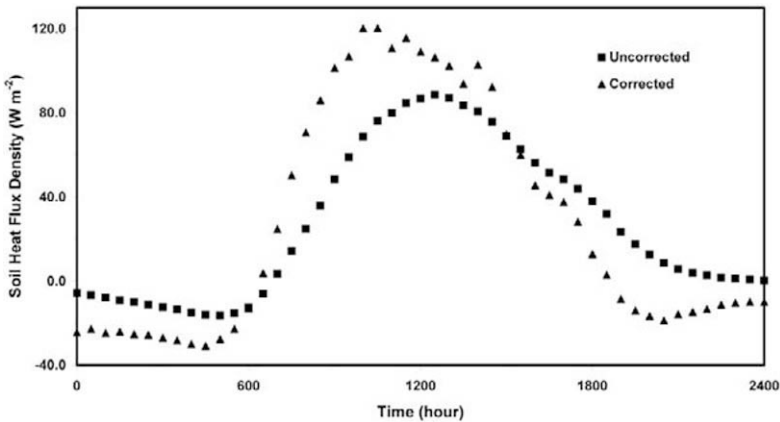
Fig. 7–4. Heat flux density measured at 0.05 m in a silt loam soil in southwestern Minnesota with and without the correction for heat storage in the soil layer above the heat flux plate (T.J. Sauer and N.S. Eash, unpublished data).

heat transfer is neglected under these conditions. This error is probably insignificant when plates are positioned below 0.05 m.

## Calorimetric Method

The calorimetric or temperature integral method is used to determine the average soil heat flux density from the change in heat storage in the soil over a given time interval (Fuchs, 1986):

$$G_{i-1} = \delta z_i C_i (\partial T_i / \partial t) + G_i \tag{6}$$

where $G_{i-1}$ is the heat flux density at the top of a layer, $\delta z_i$ is the layer thickness (m), $C_i$ is the volumetric heat capacity for the layer (J m$^{-3}$ K$^{-1}$), $\partial T_i / \partial t$ is the rate of change of the mean layer temperature (K), and $G_i$ is the heat flux density at the bottom of the layer. If a layer $n$ is found with $\partial T / \partial z = 0$, then by Eq. [2] $G_n = 0$ and the soil heat flux density for layer $j$ is given by

$$G_{j-1} = \sum_{i=j}^{n} \delta z_i C_i (\partial T_i / \partial t) \tag{7}$$

When $j = 1$, Eq. [7] gives $G$ at the soil surface. Alternatively, if $G_i$ is known by some other method at a reference depth ($z_r$), Eq. [6] can be used to calculate $G$ at all other depths. If $G_r$ is calculated using the gradient method, this approach is referred to as the combination method, which is discussed in a later section.

### Sensors

Soil temperature and volumetric heat capacity are the only data inputs necessary for completing the calculations involved with the calorimetric analysis.

Soil temperature must be accurately measured at several depths from near the surface to a depth where $G$ is negligible for the desired measurement interval, generally at ~1 m for discerning diurnal patterns in $G$. Temperature sensors with different characteristics may be desired for the surface layers (i.e., more sensitive thermistors or smaller thermocouples), where temperature changes rapidly, than for deeper in the profile where the amplitude of soil temperature is dampened. Again, McInnes (2002) provides detailed information regarding soil temperature measurements.

Techniques for determining soil heat capacity are discussed by Kluitenberg (2002). Methods for measuring soil heat capacity use calorimetry or heat pulse probes that can be used in laboratory or field settings (Kluitenberg et al., 1993; Bristow et al., 1994). A technique to estimate volumetric heat capacity based on the volume fractions of the mineral, organic matter, and water components of a soil developed by de Vries (1963) also is widely used. To use the de Vries estimation technique, the soil porosity, organic matter content, and volumetric water content must be known. Thus, if bulk density and organic matter content are known for a soil, only measurement of the volumetric water content is necessary to estimate the volumetric heat capacity. Soil water content can be measured by several intrusive or non-intrusive techniques (Topp & Ferré, 2002). Multiple sensors or water content samples collected from each site and depth may be used to improve spatial averaging.

### Procedure

A relatively deep excavation will be required at the site(s) that will allow insertion of sensors into an exposed soil face. Selection of measurement locations should include consideration of the presence or absence of a plant canopy or residue layer, surface roughness, row spacing and orientation, and proximity to anomalous features (e.g., large rocks, tree roots, and compacted areas). Careful selection of depth intervals is essential to successful application of the method. Abrupt discontinuities in soil particle size, mineralogy, or bulk density within layers should be avoided to ensure uniform volumetric heat capacities and linear temperature gradients within each layer. Typically, larger changes in water content and temperature are observed near the surface, which result in progressively smaller contributions to $G$ per depth increment with distance from the surface. Optimal sensor placement in a homogeneous soil should approximate a geometric progression with depth (e.g., 0.01-, 0.02-, 0.04-, 0.08-, 0.16-, 0.32-, 0.64-, 0.96-, and 1.28-m depths).

Measurement frequency and averaging interval of the temperature measurements will depend on the time constant of the temperature and heat capacity sensors, and objectives of the measurement protocol (e.g., hourly flux vs. daily sums vs. seasonal trends). Typically, data signals are averaged across 15- to 60-minute intervals. If heat capacity is being estimated, soil water content and bulk density values (Grossman & Reinsch, 2002) must be available for each layer between the temperature sensors. Identification of a layer with $G = 0$ or determination of $G$ at a reference depth then allows calculation of $G$ at all other depths and at the soil surface using Eq. [6].

## Advantages and Disadvantages

The calorimetric method is constrained by the ability to accurately determine changes in volumetric heat capacity for each soil layer and each time interval. Under conditions with slowly changing soil water content, this may not present a significant challenge; however, for layered soils or when soil water content is changing rapidly (e.g., wetting front advance after a rainfall event), obtaining accurate heat capacity values is difficult. Cellier et al. (1996) provide a recent example of use of the calorimetric method to measure $G$ including the development of an estimation technique using commonly-measured micrometeorological parameters.

Very precise soil temperature measurements (~0.02–0.1 K precision) are required for successful application of this method. This is especially important at deeper depths where the temperature gradient is applied to progressively thicker layers with large volumetric heat capacities. This also is important if data are required for shorter time intervals with smaller temperature changes (Hanks & Jacobs, 1971; Fuchs, 1986). Given the relatively stringent requirements for the temperature measurements, the large number of sensors, and their placement deep in the soil, the calorimetric method is most often used as a reference for comparison with other methods.

## Gradient Method

The gradient method is a direct application of Eq. [2]. A measured temperature gradient ($\partial T/\partial z$) is combined with an estimated or measured thermal conductivity ($\lambda$) to determine $G$. The simplicity of this approach is offset by difficulty in obtaining accurate $\lambda$ measurements under field conditions. Like volumetric heat capacity, soil thermal conductivity changes significantly with soil water content, in this instance due to the large difference in $\lambda$ between water and air (0.57 vs. 0.025 W $m^{-1}$ $K^{-1}$ at 283 K, respectively). As a result, $\lambda$ values for soil layers in the field can change by a factor of two to five across commonly observed changes in water content.

### Sensors

Bristow (2002) provides details concerning the measurement of soil thermal conductivity using both steady-state and transient methods. The single heat probe and heat pulse methods are best suited for in situ field measurements. Also included in the discussion is an approximation by de Vries (1963) for estimating soil thermal conductivity based on a phase mixing model and the weighted volume fraction of soil constituents. Data on soil water content will be required if the de Vries method is to be used for estimating $\lambda$. Thermocouples or thermistors are again suitable temperature sensors for the gradient method as they have sufficient accuracy and are easily logged for continuous measurements.

### Procedure

Due to difficulties in accurately measuring $\lambda$ near the soil surface, the gradient method is generally applied at a reference depth of at least 0.2 m. The calorimetric method can be applied to the surface layers above the reference layer

to determine $G$ at the surface. Coupling the calorimetric and gradient methods is referred to as the combination method and is discussed in the next section. A shallower or deeper reference depth may be appropriate or possible for different soils depending on texture, moisture content, and the amplitude of the temperature wave. Placement of soil temperature, thermal conductivity, and/or soil water content sensors will depend on layering within the soil and overall objectives of the $G$ measurements.

An excavation must be made to allow insertion of sensors at the desired depths and at locations appropriate for accurate representation of the measurement area. Multiple sensors on the same horizontal plane or multiple sensor locations will improve the spatial sampling. Sampling frequency and averaging intervals will depend on sensor characteristics, sampling objectives, and time scale of changes in soil moisture and temperature. The temperature gradient can be obtained by differentiating a smooth curve fit to the temperature data or by simply taking the average temperature difference across each layer over the averaging interval.

## Advantages and Disadvantages

The gradient method is simple to employ, however, accurate measurement of $\lambda$ in situ can be challenging. Recent improvements in thermal conductivity measurement techniques, especially for in situ measurements, now make application of the gradient method more attractive. Nonetheless, accurate estimates or measurements of $\lambda$ near the soil surface are necessary if $G$ is to be determined at the soil surface for use in energy balance or evaporation equations. Kimball et al. (1976) used four variations of the gradient method to measure $G$ in a bare loam soil. They concluded that the method produced acceptable results with $\lambda$ values estimated by de Vries' theory when a 0.2 m reference depth was used but not with a 0.05 m reference depth.

## Combination Method

The coupling of the gradient and calorimetric methods is known as the combination method. In some instances, the coupling of the soil heat flux plate and calorimetric methods has been referred to as the combination method (Fuchs & Tanner, 1968; Massman, 1993). Here, the former will be considered the combination method while the latter is considered a necessary heat storage correction for accurate application of the flux plate method.

The combination of gradient and calorimetric methods takes advantage of the limited measurements required for the gradient method and the accuracy of near-surface $G$ measurements of the calorimetric method. The gradient method is used to determine $G$ at some reference depth (typically 0.2 m) through application of Eq. [2] and the calorimetric method is applied to successive layers between the reference depth and the soil surface using Eq. [6]. The combination method avoids the deep profile measurements of soil temperature and volumetric heat capacity of the calorimetric method and enables the determination of $G$ at the soil surface based on a flux determined from the gradient method applied to one layer.

**Sensors**

Appropriate soil temperature, thermal conductivity, and heat capacity sensors/techniques to be used with the combination method or its null-alignment variant (Kimball & Jackson, 1975) have already been described for the calorimetric and gradient methods.

## Procedure

Only a shallow excavation is necessary as all required sensors will be at or less than approximately 0.25 m. Again, choice of location(s) and number of sensors at each depth will be dictated by the heterogeneity of soil properties and surface cover within the area to be represented by the measurements. The number and vertical spacing of temperature sensors above the 0.2 m reference will be determined by the soil properties although a geometric progression from the surface is still advisable with maximum vertical separation of <0.05 m. The temperature gradient at the 0.2 m reference depth can be determined from sensors placed above and below (e.g., 0.15 and 0.25 m). Unless the heat capacity is directly measured, soil water content and bulk density need to be measured or estimated for each layer concurrent with the temperature measurements to determine the volumetric heat capacity for the calorimetric calculations.

## Advantages and Disadvantages

The heat storage calculations are applied to relatively thin layers near the surface so 0.1 K resolution in soil temperature measurements are acceptable. As the necessary soil temperature measurements are straightforward, success with the combination approach will depend on how accurate the thermal conductivity and volumetric heat capacity of the soil layers can be measured or estimated. Pikul and Allmaras (1984) and de Vries and Philip (1986) compared the null-alignment method with the theory of Philip and de Vries (1957) and reported contrasting results. Pikul and Allmaras (1984) found poor agreement between the null-alignment and mechanistic approaches especially for shallow soil layers and dry soil conditions. De Vries and Philip (1986) concluded that the failure to account for latent heat loss in the upper soil layers resulted in serious underestimation of $\lambda$ and systematic errors with the null alignment technique.

## ESTIMATION AND PREDICTION OF SOIL HEAT FLUX DENSITY

The importance of soil heat flux to surface energy balance and evaporation investigations has encouraged the development of estimation and prediction techniques for use when measured values of $G$ are unavailable. These techniques rely on surrogate micrometeorological data, parameters obtained using remote sensing technology, or information on soil thermal properties, for input into estimation algorithms. One simple and popular technique involves the ratio of $G$ to $R_n$. Table 7–2 lists example values of $G/R_n$ that were obtained for a variety of agricultural surfaces. As expected, $G/R_n$ is relatively high (>~0.15) for bare soils and

Table 7–2. Typical values for the ratio of soil heat flux density to net radiation ($G/R_n$) for various agricultural surfaces.

| Surface/canopy | $G/R_n$ | Location(s) | Comments | Reference |
|---|---|---|---|---|
| Loam soils | 0.22–0.51 | Arizona and Montana | $G/R_n$ increased as soil dried | Idso et al., 1975 |
| Loess soil | 0.34 | Israel | $G/R_n$ unaffected by soil wetness | Fuchs & Hadas, 1972 |
| Silty clay soil | 0.14 | Syria | daylight hours | Oliver et al., 1987 |
| Alfalfa | 0.10 | Arizona | midday values, $G/Rn = 0.3$ for stubble | Clothier et al., 1986 |
| Barley | 0.12 | Syria | daylight hours, < 50% ground cover | Oliver et al., 1987 |
| Grass | 0.04 | California | $R_n > 0$, mown and well-watered | Meek et al., 1989 |
| Maize | 0.06 | New York | daylight hours, 2.5 m-tall canopy | Brown & Covey, 1966 |
| Mixed prairie | 0.19 | Saskatchewan | daylight hours, thick thatch layer | Ripley & Redmann, 1976 |
| Orchard | 0.04 | West Virginia | $G/R_n = 0.07$ with coal dust amended soil | Sharratt & Glenn, 1988 |
| Pasture | 0.10 | The Netherlands | daylight hours | DeBruin & Holtslag, 1982 |
| Pine forest | 0.04 | Australia | daylight hours, 7.5 m-tall canopy | Denmead, 1969 |
| Pine forest | 0.01 | United Kingdom | daylight hours, 9 m-tall canopy | Oliver et al., 1987 |
| Sorghum | 0.11 | Texas | daylight hours | Szeicz et al., 1973 |
| Sugar beet | 0.03 | Nebraska | daylight hours, full canopy | Brown, 1976 |
| Wheat | 0.07 | Australia | daylight hours, 0.32 to 0.55 m-tall canopy | Denmead, 1969 |

sparse canopies and low for fully-developed crop canopies and forests, which have greater attenuation of $R_n$ within the canopy. While the $G/R_n$ ratio has proven to be consistent in some instances and especially during daytime hours (Fuchs & Hadas, 1972), the ratio has been found to be sensitive to changing soil water content and canopy density (Idso et al, 1975; Clothier et al., 1986; Oliver et al., 1987). For these reasons, using $G/R_n$ ratios to estimate $G$ are primarily useful as a first approximation, with the understanding that such estimates may have large errors relating to variation in surface characteristics over time.

In spite of the uncertainty surrounding estimates of $G$ from the $G/R_n$ ratio, it is still reasonable to expect that soil heat flux would be some small percentage of net radiation, a flux that is relatively easy to measure. Plant canopy properties that affect soil heat flux include the height of the canopy, the leaf area index (LAI), and the amount of vegetative cover (Clothier et al., 1986; Yang et al., 1999). Several techniques have been developed to improve estimates of $G$ from $R_n$ or other data by incorporating additional attributes that characterize canopy properties. Anadranistakis et al. (1997) investigated the relationship between $G/R_n$ and LAI for a barley (*Hordeum vulgare* L.) crop at various stages of development in Greece. An exponential relationship was found between LAI and $G/R_n$ for daytime hours. The $G/R_n$ ratio was 0.43 when the LAI was near 0 and approached a limit of 0.1 for large LAI. Clothier et al. (1986) found inclusion of crop height and a spectral vegetation index (ratio of near infrared to red reflectance, NIR/Red) both improved estimates of $G$ from $G/R_n$. Daughtry et al. (1990) and Kustas and Daughtry (1990) used ground-based and remotely-sensed multispectral reflectance data to extend the work of Clothier et al. (1986) to determine a normalized difference vegetation index (NDVI), which is the difference between the near infrared and red reflectance divided by their sum. Use of the NDVI resulted in improved estimates of $G$ from $R_n$ for fields with bare soil and cotton (*Gossypium hirsutum* L.) and alfalfa (*Medicago sativa* L.) canopies at different stages of growth. Kustas et al. (1993) explored nonlinear relationships between $G/R_n$ and vegetation indices (VIs). They concluded that a power function

$$G/R_n = aVI^b \tag{8}$$

where *a* and *b* are fitted constants, was more appropriate than the previously derived linear relationships between $G/R_n$ ratios and VIs such as the NDVI and NIR/Red reflectance.

Another approach for predicting $G$ uses soil temperature measurements in combination with different solutions to soil heat flow equations. Since soil temperature is easily measured and continuous data records for multiple depths are often available, these techniques offer another opportunity to estimate $G$ when direct measurements are lacking. Many of these techniques do, however, involve various assumptions concerning soil thermal properties and may be limited to certain prescribed conditions (e.g., homogeneous soil, constant soil water content, no canopy, sunny days).

Horton and Wierenga (1983) developed a method to estimate $G$ that uses a harmonic analysis of soil temperature at one depth, if soil thermal diffusivity ($\alpha$ =

$\lambda/C$) is known, or at two depths if $\alpha$ is unknown. A shallow soil temperature can be described:

$$T(z,t) = T_t \sum_{n=1}^{M} A_{0n} \sin(n\omega t + \phi_{0n}) \qquad [9]$$

where $T_t$ is the temporal average soil temperature, $M$ the number of harmonics (usually 1 to 3 harmonics are adequate), $A_{0n}$ and $\phi_{0n}$ are the amplitude and phase angle, respectively, of the $n^{th}$ harmonic, and $\omega$ is the radial frequency equal to $2\pi/P$ with $P$ being the period of the fundamental cycle (24 h for diurnal cycles). Fitting Eq. [9] to the observed shallow soil temperature values provides $T_t$, $A_{0n}$, and $\phi_{0n}$. Soil heat flux density can then be described as a function of time and depth:

$$G(z,t) = \sum_{n=1}^{M} \left\{ A_{0n} C \sqrt{n\omega\alpha} \exp(-z\sqrt{n\omega/2\alpha}) \sin\left[ n\omega t + \phi_{0n} + (\pi/4) - z\sqrt{n\omega/2\alpha} \right] \right\} \qquad [10]$$

Equation [10] represents the soil heat flux density, positive downward in a homogeneous soil profile, with the temperature at the surface described by a Fourier series. To calculate $G$ with Eq. [10], one has to know the values for $A_{0n}$ and $\phi_{0n}$ for the temperature at one depth, as well as $\alpha$ and $C$ for the soil. Horton et al. (1983) describe how $\alpha$ can be determined from measurements of soil temperature at two depths. Soil heat flux density estimated with this harmonic method was in good agreement with $G$ measured using the calorimetric method for a clay loam soil in New Mexico. Gupta et al. (1984) also used a harmonic analysis, in this instance, with soil temperature normalized with respect to daily maximum and minimum soil surface temperature. This technique was used to predict $G$ in soils having different tillage and residue cover conditions.

Sharratt et al. (1992) used a finite-difference solution to the transient heat flow equation to estimate $G$ from hourly soil temperature data at three depths. The finite-difference form of the transient heat flow equation for two layers (three nodes) can be written (Campbell, 1985):

$$\lambda_2(\Delta\bar{T}_2)/\Delta z_2 - \lambda_1(\Delta\bar{T}_1)/\Delta z_1 = C(T_2^{j+1} - T_2^{j})\Delta z_3/\Delta t \qquad [11]$$

where the superscripts (j and j+1) indicate time and the subscripts (1, 2, and 3) indicate node and layer number (increasing with depth). Soil volumetric heat capacity was measured once a day and a least-squares solution was used to estimate daily values for $\lambda_1$ and $\lambda_2$, which then enabled calculation of $G$ for each layer using Eq. [2]. The finite-difference method produced values of $G$ that had smaller errors than the harmonic method when estimating $G$ in silt loam soils in West Virginia and Alaska. Horton and Chung (1991) also used a finite difference approach coupled with surface energy partitioning equations to calculate soil heat flux density. Their method required knowledge of soil thermal and hydraulic properties, and calculations were driven by meteorological inputs, e.g., wind-speed, air temperature and humidity, and radiation.

Wang and Bras (1999) applied fractional calculus to establish an explicit relationship between surface temperature and soil heat flux. Although Wang and Bras (1999) assumed that the soil thermal properties were uniform and independent of soil water content and temperature, $G$ estimated with their technique compared well with field data from Kansas and the Amazon.

## SUMMARY

Soil heat flux density is an integral component of the surface energy balance, affecting the amount of energy available for latent and sensible heat transfer. It also represents an energy flow path that couples soil and atmospheric systems. This coupling has important implications for local microclimate and the soil thermal regime, which in turn affect crop performance. The number and diversity of studies devoted to its measurement and prediction demonstrates the importance of soil heat flux to micrometeorological measurements in agricultural systems.

The flux plate method is now the most popular method for determining $G$. Since $G$ is often the smallest term considered in the surface energy balance, the necessity of correcting measured fluxes for heat storage above the plate and/or latent heat loss below have sometimes been discounted as unnecessary. Recent evidence, however, strongly suggests that these corrections may be sizeable under commonly encountered conditions. The large spatial variation of $G$, especially under sparse canopies and with uneven surfaces, also needs to be considered when deploying sensors to measure $G$ during surface energy balance studies. Proper corrective measures to account for these potential errors should lead to better estimates of $G$. More accurate $G$ data will in turn reduce errors in latent and sensible heat fluxes determined from the Bowen ratio energy balance method and improve energy balance closure when eddy covariance or other techniques are used to measure turbulent fluxes.

## REFERENCES

Abu-Hamdeh, N.H., and R.C. Reeder. 2000. Soil thermal conductivity: Effects of density, moisture, salt concentration, and organic matter. Soil Sci. Soc. Am. J. 64:1285–1290.

Allmaras, R.R., E.A. Hallauer, W.W. Nelson, and C.D. Evans. 1977. Surface energy balance and soil thermal property modifications by tillage-induced soil structure. Minnesota Agric. Exp. Stn. Tech. Bull. 306. St. Paul, MN.

Al Nakshabandi, G., and H. Kohnke. 1965. Thermal conductivity and diffusivity of soils as related to moisture tension and other physical properties. Agric. Meteorol. 2:271–279.

Anadranistakis, M., A. Liakatas, S. Alexandris, S. Aggelides, P. Kerkides, S. Rizos, and A. Poulovassilis. 1997. Soil heat flux in the Penman-Monteith evapotranspiration equation. p. 69–74. *In* K.S. Chartzoulakis (ed.) Proc. of the 2nd Int. Symp. on Irrigation of Horticultural Crops, Sept. 9–13, 1996, Chania, Crete, Greece.

Azooz, R.H., B. Lowery, T.C. Daniel, and M.A. Arshad. 1997. Impact of tillage and residue management on soil heat flux. Agric. For. Meteorol. 84:207–222.

Baldocchi, D.B., S.B. Verma, and N.J. Rosenberg. 1985. Water use efficiency in a soybean field: Influence of plant water stress. Agric. For. Meteorol. 34:53–65.

Biscoe, P.V., R.A. Saffell, and P.D. Smith. 1977. An apparatus for calibrating soil heat flux plates. Agric. Meteorol. 18:49–54.

Bowen, I.S. 1926. The ratio of heat losses by conduction and by evaporation from any water surface. Phys. Rev. 27:779–787.

Bremer, D.J., and J.M. Ham. 1999. Effect of spring burning on the surface energy balance in a tall grass prairie. Agric. For. Meteorol. 97:43–54.

Bristow, K.L. 2002. Thermal conductivity. p. 1209–1226. *In* G.C. Topp and J.H. Dane (ed.) Methods of soil analysis. Part 4. Physical methods. SSSA Book Series no. 5. SSSA, Madison, WI.

Bristow, K.L., G.J. Kluitenberg, and R. Horton. 1994. Measurement of soil thermal properties with a dual-probe heat-pulse technique. Soil Sci. Soc. Am. J. 58:1288–1294.

Brown, K.W. 1976. Sugar beet and potatoes. p. 65–86. *In* J.L. Monteith (ed.) Vegetation and the atmosphere. Vol. 2. Academic Press, New York.

Brown, K.W., and W. Covey. 1966. The energy-budget evaluation of the micro-meteorological transfer processes within a cornfield. Agric. Meteorol. 3:73–96.

Buchan, G.D. 1989. Soil heat flux and soil surface energy balance: A clarification of concepts. *In* Proc. of the 4th Australasian Conf. on Heat and Mass Transfer, May 1989, Christchurch, New Zealand. University of Canterbury.

Cahill, A.T., and M.B. Parlange. 1998. On water vapor transport in field soils. Water Resour. Res. 34:731–739.

Campbell, G.S. 1985. Soil physics with BASIC: Transport models for soil–plant systems. Elsevier, New York.

Carson, J.E., and H. Moses. 1963. The annual and diurnal heat-exchange cycles in upper layers of soil. J. Appl. Meteorol. 2:397–406.

Carter, M.R. (ed.) 1993. Soil sampling and methods of analysis. Lewis Publishers, Boca Raton, FL.

Cellier, P., G. Richard, and P. Robin. 1996. Partition of sensible heat fluxes into bare soil and the atmosphere. Agric. For. Meteorol. 82:245–265.

Choudhury, B.J., S.B. Idso, and J.R. Reginato. 1987. Analysis of an empirical model for soil heat flux under a growing wheat crop for estimating evaporation by an infrared-temperature based energy balance equation. Agric. For. Meteorol. 39:283–297.

Clothier, B.E., K.L. Clawson, P.J. Pinter, Jr., M.S. Moran, R.J. Reginato, and R.D. Jackson. 1986. Estimation of soil heat flux from net radiation during the growth of alfalfa. Agric. For. Meteorol. 37:319–329.

Daughtry, C.S.T., W.P. Kustas, M.S. Moran, P.J. Pinter, Jr., R.D. Jackson, P.W. Brown, W.D. Nichols, and L.W. Gay. 1990. Spectral estimates of net radiation and soil heat flux. Remote Sens. Environ. 32:111–124.

Deacon, E.L. 1950. The measurement and recording of the heat flux into the soil. Q.J.R. Meteorol. Soc. 76:479–483.

Denmead, O.T. 1969. Comparative micrometeorology of a wheat field and a forest of *Pinus radiata*. Agric. Meteorol. 6:357–371.

de Bruin, H.A.R., and A.A.M. Holtslag. 1982. A simple parameterization of the surface fluxes of sensible and latent heat during daytime compared with the Penman-Monteith concept. J. Appl. Meteorol. 21:1610–1621.

de Silans, A.P., B.A. Monteny, and J.P. Lhomme. 1997. The correction of soil heat flux measurements to derive an accurate surface energy balance by the Bowen ratio method. J. Hydrol. 188/189:453–465.

de Vries, D.A. 1963. Thermal properties of soils. p. 210–235. *In* W.R. van Wijk (ed.) Physics of plant environment. North-Holland Publ. Co., Amsterdam.

de Vries, D.A., and J.R. Philip. 1986. Soil heat flux, thermal conductivity, and the null-alignment method. Soil Sci. Soc. Am. J. 50:12–17.

Dunkle, R.V. 1940. Heat meters. Bull. Am. Meteorol. Soc. 21:116–117.

Enz, J.W., L.J. Brun, and J.K. Larsen. 1988. Evaporation and energy balance for bare and stubble covered soil. Agric. For. Meteorol. 43:59–70.

Falckenberg, G. 1930. Apparatur zur Bestimmung des momentanen nächtlichen Wärmeaustausches zwischen Erde und Luft. Meteorologische Zeitschrift. 47:154–156.

Fritton, D.D., and J.D. Martsolf. 1980. Soil heat flow under and orchard heater. Soil Sci. Soc. Am. J. 44:13–16.

Fritton, D.D., J.D. Martsolf, and W.J. Busscher. 1976. Spatial distribution of soil heat flux under a sour cherry tree. Soil Sci. Soc. Am. J. 40:644–647.

Fuchs, M. 1986. Heat flux. p. 957–968. *In* A. Klute (ed.) Methods of soil analysis. Part 1. Physical and mineralogical methods. 2nd ed. SSSA Book Series no. 5. SSSA, Madison, WI.

Fuchs, M., and A. Hadas. 1972. The heat flux density in a non-homogeneous bare loessial soil. Boundary-layer Meteorol. 3:191–200.

Fuchs, M., and A. Hadas. 1973. Analysis and performance of an improved soil heat flux transducer. Soil Sci. Soc. Am. Proc. 37:173–175.

Fuchs, M., and C.B. Tanner. 1968. Calibration and field test of soil heat flux plates. Soil Sci. Soc. Am. Proc. 32:326–328.

Gatenby, R.M. 1977. Conduction of heat from sheep to ground. Agric. Meteorol.18:387–400.

Gausman, H.W., A.H. Gerbermann, C.L. Wiegand, R.W. Leamer, R.R. Rodriguez, and J.R. Noriega. 1975. Reflectance differences between crop residues and bare soils. Soil Sci. Soc. Am. Proc. 39:752–755.

Gilgen, H., and A. Ohmura. 1999. The global energy balance archive. Bull. Am. Meteor. Soc. 80:831–850.

Glenn, D.M., and W.V. Welker. 1987. Soil management effects on soil temperature and heat flux in a young peach orchard. Soil Sci. 143:372–380.

Goel, N.R., and J.M. Norman. (ed.) 1990. Instrumentation for studying vegetation canopies for remote sensing in optical and thermal infrared regions. Remote Sensing Rev. 5:1–360.

Grossman, R., and T.G. Reinsch. 2002. Bulk density and linear extensibility. p. 201–228. *In* G.C. Topp and J.H. Dane (ed.) Methods of soil analysis. Part 4. Physical methods. SSSA Book Series no. 5. SSSA, Madison, WI.

Gupta, S.C., W.E. Larson, and R.R. Allmaras. 1984. Predicting soil temperature and soil heat flux under different tillage-surface residue conditions. Soil Sci. Soc. Am. J. 48:223–232.

Ham, J.M., J.L. Heilman, and R.J. Lascano. 1991. Soil and canopy energy balances of a row crop at partial cover. Agron. J. 83:744–753.

Ham, J.M., and G.J. Kluitenberg. 1993. Positional variation in the soil energy balance beneath a row-crop canopy. Agric. For. Meteorol. 63:73–92.

Hanks, R.J., and H.S. Jacobs. 1971. Comparison of the calorimetric and flux meter measurements of soil heat flux. Soil Sci. Soc. Am. Proc. 35:671–674.

Heilman, J.L., K.J. McInnes, M.J. Savage, R.W. Gesch, and R.J. Lascano. 1994. Soil and canopy energy balances in a west Texas vineyard. Agric. For. Meteorol. 71:99–114.

Herin, P., and P. Théry. 1992. Measurements on the thermoelectric properties of thin layers of two metals in electrical contact. Application for designing new heat-flow sensors. Meas. Sci. Technol. 3:495–500.

Högström, U. 1974. In situ calibration of ground heat flux plates. Agric. Meteorol. 13:161–168.

Horton, R., K.L. Bristow, G.J. Kluitenberg, and T.J. Sauer. 1996. Crop residue effects on surface radiation and energy balance: A review. Theor. Appl. Climatol. 54:27–37.

Horton, R. and S.-O. Chung. 1991. Soil heat flow. p. 397–438. *In* J. Hanks and J.T. Ritchie (ed.) Modeling plant and soil systems. Agron. Monogr. 31. ASA, Madison, WI.

Horton, R., and P.J. Wierenga. 1983. Estimating the soil heat flux from observations of soil temperature near the surface. Soil Sci. Soc. Am. J. 47:14–20.

Horton, R., P.J. Wierenga, and D.R. Nielsen. 1983. Evaluation of methods for determining the apparent thermal diffusivity of soil near the surface. Soil Sci. Soc. Am. J. 47:25–32.

Howell, T.A., and J.A. Tolk. 1990. Calibration of soil heat flux transducers. Theor. Appl. Climatol. 42:263–272.

Idso, S.B. 1972. Calibration of soil heat flux plates by a radiation technique. Agric. Meteorol. 10:467–471.

Idso, S.B., J.K. Aase, and R.D. Jackson. 1975. Net radiation-soil heat flux relations as influenced by soil water content variations. Boundary-layer Meteorol. 9:113–122.

Jackson, R.D., and S.A. Taylor. 1965. Heat transfer. p. 349–356. *In* C.A. Black et al. (ed.) Methods of soil analysis. Part 1. Agron. Monogr. 9. ASA, Madison, WI.

Kiehl, J.T., and K.E. Trenberth. 1997. Earth's annual global mean energy budget. Bull. Am. Meteorol. Soc. 78:197–208.

Kimball, B.A., and R.D. Jackson. 1975. Soil heat flux determination: A null-alignment method. Agric. Meteorol. 15:1–9.

Kimball, B.A., and R.D. Jackson. 1979. Soil heat flux. p. 211–229. *In* B.J. Barfield and J.F. Gerber (ed.) Modification of the aerial environment of plants. ASAE Monogr. 2. ASAE, St. Joseph, MI.

Kimball, B.A., R.D. Jackson, F.S. Nakayama, S.B. Idso, and R.J. Reginato. 1976. Soil-heat flux determination: Temperature gradient method with computed thermal conductivities. Soil Sci. Soc. Am. J. 40:25–28.

Kimball, B.A., R.L. LaMorte, P.J. Pinter, Jr., G.W. Wall, D.J. Hunsaker, F.J. Adamsen, S.W. Leavitt, T.L. Thompson, A.D. Matthias, and T.J. Brooks. 1999. Free-air $CO_2$ enrichment and soil nitrogen effects on energy balance and evapotranspiration of wheat. Water Resour. Res. 35:1179–1190.

Kluitenberg, G.J. 2002. Heat capacity and specific heat. p. 1201–1208. *In* G.C. Topp and J.H. Dane (ed.) Methods of soil analysis: Part 4. Physical methods. SSSA Book Series no. 5. Madison, WI.

Kluitenberg, G.J., J.M. Ham, and K.L. Bristow. 1993. Error analysis of the heat-pulse method for measuring the volumetric heat capacity of soil. Soil Sci. Soc. Am. J. 57:1444–1451.

Kumar, K.K., and V.U. Rao. 1984. Significance of soil heat flux in the estimation of potential evaporation. Indian J. Ecol. 11:68–70.

Kustas, W.P., and C.S.T. Daughtry. 1990. Estimation of the soil heat flux/net radiation ratio from spectral data. Agric. For. Meteorol. 49:205–223.

Kustas, W.P., C.S.T. Daughtry, and P.J. Van Oevelen. 1993. Analytical treatment of the relationships between soil heat flux/net radiation ratio and vegetation indices. Remote Sens. Environ. 46:319–330.

Kustas, W.P., J.H. Prueger, J.L. Hatfield, K. Ramalingam, and L.E. Hipps. 2000. Variability in soil heat flux from a mesquite dune site. Agric. For. Meteorol. 103:249–264.

Lemon, E.R. 1963. The energy budget at the earth's surface. Part 1. U.S. Army Electronic Proving Ground Production Res. Rep. no. 71. U.S. Army, Washington, DC.

Lettau, H., and B. Davidson, Ed. 1957. Exploring the atmosphere's first mile. Vol. I. Instrumentation and data evaluation. Pergamon Press, New York.

Lourence, F.J., and W.O. Pruitt. 1971. Energy balance and water use of rice grown in the Central Valley of California. Agron. J. 63:827–832.

Malek, E. 1993. Rapid changes of the surface soil heat flux and its effects on the estimation of evapotranspiration. J. Hydrol. 142:89–97.

Massman, W.J. 1992. Correcting errors associated with soil heat flux measurements and estimating soil thermal properties from soil temperature and heat flux plate data. Agric. For. Meteorol. 59:249–266.

Massman, W.J. 1993. Errors associated with the combination method for estimating soil heat flux. Soil Sci. Soc. Am. J. 57:1198–1202.

Mayocchi, C.L., and K.L. Bristow. 1995. Soil surface heat flux: Some general questions and comments on measurements. Agric. For. Meteorol. 75:43–50.

McCaughey, J.H. 1982. Spatial variability of net radiation and soil heat flux density on two logged sites in Montmorency, Quebec. J. Appl. Meteor. 21:777–787.

McCaughey, J.H., and W.L. Saxton. 1988. Energy balance storage terms in a mixed forest. Agric. For. Meteorol. 44:1–18.

McInnes, K.J. 2002. Temperature. p. 1183–1199. *In* G.C. Topp and J.H. Dane (ed.) Methods of soil analysis. Part 4. Physical methods. SSSA Book Series no. 5. SSSA, Madison, WI.

Meek, D., C. Phene, W. Pruitt, and D. Clark. 1989. The contribution of soil heat flux to real-time evapotranspiration. p. 19. *In* Agronomy Abstracts. ASA, Madison, WI.

Mogensen, V.O. 1970. The calibration factor of heat flux meters in relation to the thermal conductivity of the surrounding medium. Agric. Meteorol. 7:401–410.

Monteith, J.L. 1958. The heat balance beneath crops. Climatology and microclimatology. UNESCO Arid Zone Res. 11:123–128.

Monteith, J.L. 1965. Evaporation and environment. Symp. Soc. Exp. Biol. 19:205–234.

Ohmura, A., and H. Gilgen. 1993. Re-evaluation of the global energy balance. p. 93–110. *In* Interactions between the global climate subsystems: The legacy of Hann. Geophys. Monogr. no. 75. Am. Geophysical Union, Washington, DC.

Oliver, H.R. 1992. Studies of surface energy balance of sloping terrain. Int. J. Climatol. 12:55–68.

Oliver, S.A., H.R. Oliver, J.S. Wallace, and A.M. Roberts. 1987. Soil heat flux and temperature variation with vegetation, soil type and climate. Agric. For. Meteorol. 39:257–269.

Penman, H.L. 1948. Natural evaporation from open water, bare soil, and grass. Proc. R. Soc. London, Ser. A. 193:120–145.

Philip, J.R. 1961. The theory of heat flux meters. J. Geophys. Res. 66:571–579.

Philip, J.R., and D.A. de Vries. 1957. Moisture movement in porous materials under temperature gradients. Trans. Am. Geophys. Union 38:222–232, 594.

Pikul, J.L., Jr., and R.R. Allmaras. 1984. A field comparison of null-aligned and mechanistic soil heat flux. Soil Sci. Soc. Am. J. 48:1207–1214.

Pikul, J.L., Jr., R.R. Allmaras, and S.E. Waldman. 1985. Late season heat flux and water distribution in summer-fallowed Haploxerolls. Soil Sci. Soc. Am. J. 49:1517–1522.

Portman, D.J. 1958. Conductivity and length relationships in heat-flow transducer performance. Trans. Am. Geophys. Union 39:1089–1094.

Priestley, C.H.B., and R.J. Taylor. 1972. On the assessment of surface heat flux and evaporation using large-scale parameters. Mon. Weather Rev. 100:81–92.

Rao, G.R., R.V. Ramamohan, and B.V.R. Rao. 1977. Soil heat flux studies in the bare red laterite soil during winter and summer months at Bangalore. Ann. Arid Zone. 16:5–12.

Richard, G., and P. Cellier. 1998. Effect of tillage on bare soil energy balance and thermal regime: An experimental study. Agronomie 18:163–181.

Ripley, E.A., and R.E. Redmann. 1976. Grassland. p. 349–398. In J.L. Monteith (ed.) Vegetation and the atmosphere. Vol. 2. Academic Press, New York.

Robin, P., P. Cellier, and G. Richard. 1997. Theoretical and field comparison of two types of soil heat fluxmeter. Soil Tech. 10:185–206.

Rosset, M., M. Riedo, A. Grub, M. Geissmann, and J. Fuhrer. 1997. Seasonal variation in radiation and energy balances of permanent pastures at different altitudes. Agric. For. Meteorol. 86:245–258.

Sauer, T.J. 2002. Heat flux density. p. 1233–1248. In G.C. Topp and J.H. Dane (ed.) Methods of soil analysis. Part 4. Physical methods. SSSA Book Series no. 5. SSSA, Madison, WI.

Sauer, T.J., J.L. Hatfield, and J.H. Prueger. 1997. Over-winter changes in radiant energy exchange of a corn residue-covered surface. Agric. For. Meteorol. 85:279–287.

Sauer, T.J., J.L. Hatfield, J.H. Prueger, and J.M. Norman. 1998. Surface energy balance of a corn residue-covered field. Agric. For. Meteorol. 89:155–168.

Sharratt, B.S., G.S. Campbell, and D.M. Glenn. 1992. Soil heat flux estimation based on the finite difference form of the transient heat flow equation. Agric. For. Meteorol. 61:95–111.

Sharratt, B.S., and D.M. Glenn. 1988. Orchard floor management utilizing soil-applied coal dust for frost protection. Part II. Seasonal microclimate effect. Agric. For. Meteorol. 43:147–154.

Simmers, I. 1977. Effect of soil heat flux on the water balance of a small catchment. Hydrol. Sci. Bull. 22:433–443.

Smith, K.A., and C.E. Mullins. (ed.) 1991. Soil analysis: Physical methods. Marcel Dekker, New York.

Staley, R.C., and J.R. Gerhardt. 1957. Soil heat flux measurements. p. 58–63. In H.H. Lettau and B. Davidson (ed.) Exploring the atmosphere's first mile. Vol. 1. Instrumentation and data evaluation. Pergamon Press, New York.

Stewart, J.S., and A.S. Thom. 1973. Energy budgets in pine forest. Quart. J. Roy. Meteorol. Soc. 99:154–170.

Szeicz, G., C.H.M. van Bavel, and S. Takami. 1973. Stomatal factor in the water use and dry matter production by sorghum. Agric. Meteorol. 12:361–389.

Tamai, K., T. Abe, M. Araki, and H. Ito. 1998. Radiation budget, soil heat flux and latent heat flux at the forest floor in warm, temperate mixed forest. Hydrol. Process. 12:2105–2114.

Tanner, C.B. 1963. Basic instrumentation and measurements for plant environment and micrometeorology. Dep. of Soil Sci. Soils Bull. 6. Univ. of Wisconsin, Madison.

Topp, G.C., and J.H. Dane (ed.). 2002. Methods of soil analysis. Part 4. Physical methods. SSSA Book Series no. 5. SSSA, Madison, WI.

Topp, G.C., and T. Ferré. 2002. Water content. p. 417–422. In G.C. Topp and J.H. Dane (ed.) Methods of soil analysis. Part 4. Physical methods. SSSA Book Series no. 5. SSSA, Madison, WI.

Tuzet, A., J-F. Castell, A. Perrier, and O. Zurfluh. 1997. Flux heterogeneity and evapotranspiration partitioning in a sparse crop: The fallow savanna. J. Hydrol. 188–189:482–493.

Twine, T.E., W.P. Kustas, J.M. Norman, D.R. Cook, P.R. Houser, T.P. Meyers, J.H. Prueger, P.J. Starks, and M.L. Wesely. 2000. Correcting eddy-covariance flux underestimates over a grassland. Agric. For. Meteorol. 103:279–300.

Uchijima, Z. 1976. Maize and rice. p. 33–64. In J.L. Monteith (ed.) Vegetation and the atmosphere. Vol. 2. Academic Press, New York.

van Loon, W.K.P., H.M.H. Bastings, and E.J. Moors. 1998. Calibration of soil heat flux sensors. Agric. For. Meteorol. 92:1–8.

van Wijk, W.R. 1967. New method for measuring heat flux density at the surface of soils or of other bodies. Nature 213:214.

Verhoef, A., S.J. Allen, and C.R. Lloyd. 1999. Seasonal variation of surface energy balance over two Sahelian surfaces. Int. J. Climatol. 19:1267–1277.

Wang, J., and R.L. Bras. 1999. Ground heat flux estimated from surface soil temperature. J. Hydrol. 216:214–226.

Watts, D.B., E.T. Kanemasu, and C.B. Tanner. 1990. Modified heat-meter method for determining soil heat flux. Agric. For. Meteorol. 49:311–330.

Weaver, H.L., and G.S. Campbell. 1985. Use of Peltier coolers as soil heat flux transducers. Soil Sci. Soc. Am. J. 49:1065–1067.

Yang, Z-L., Y. Dai, R.E. Dickinson, and W.J. Shuttleworth. 1999. Sensitivity of ground heat flux to vegetation cover fraction and leaf area index. J. Geophys. Res. 104:19 505–19 514.