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A review of models and micrometeorological methods used to estimate wetland evapotranspiration

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Abstract:

Within the past decade or so, the accuracy of evapotranspiration (*ET*) estimates has improved due to new and increasingly sophisticated methods. Yet despite a plethora of choices concerning methods, estimation of wetland *ET* remains insufficiently characterized due to the complexity of surface characteristics and the diversity of wetland types. In this review, we present models and micrometeorological methods that have been used to estimate wetland *ET* and discuss their suitability for particular wetland types. Hydrological, soil monitoring and lysimetric methods to determine *ET* are not discussed.

Our review shows that, due to the variability and complexity of wetlands, there is no single approach that is the best for estimating wetland *ET*. Furthermore, there is no single foolproof method to obtain an accurate, independent measure of wetland *ET*. Because all of the methods reviewed, with the exception of eddy covariance and LIDAR, require measurements of net radiation (R_n) and soil heat flux (G), highly accurate measurements of these energy components are key to improving measurements of wetland *ET*.

Many of the major methods used to determine *ET* can be applied successfully to wetlands of uniform vegetation and adequate fetch, however, certain caveats apply. For example, with accurate R_n and G data and small Bowen ratio (β) values, the Bowen ratio energy balance method can give accurate estimates of wetland *ET*. However, large errors in latent heat flux density can occur near sunrise and sunset when the Bowen ratio $\beta \approx -1{\cdot}0$. The eddy covariance method provides a direct measurement of latent heat flux density (λE) and sensible heat flux density (H), yet this method requires considerable expertise and expensive instrumentation to implement. A clear advantage of using the eddy covariance method is that λE can be compared with $R_n - G - H$, thereby allowing for an independent test of accuracy. The surface renewal method is inexpensive to replicate and, therefore, shows particular promise for characterizing variability in *ET* as a result of spatial heterogeneity. LIDAR is another method that has special utility in a heterogeneous wetland environment, because it provides an integrated value for *ET* from a surface. The main drawback of LIDAR is the high cost of equipment and the need for an independent *ET* measure to assess accuracy. If R_n and G are measured accurately, the Priestley–Taylor equation can be used successfully with site-specific calibration factors to estimate wetland *ET*. The 'crop' cover coefficient (K_c) method can provide accurate wetland *ET* estimates if calibrated for the environmental and climatic characteristics of a particular area. More complicated equations such as the Penman and Penman–Monteith equations also can be used to estimate wetland *ET*, but surface variability and lack of information on aerodynamic and surface resistances make use of such equations somewhat questionable. Copyright © 2004 John Wiley & Sons, Ltd.

KEY WORDS Bowen ratio energy balance; eddy covariance; evapotranspiration; LIDAR; Penman–Monteith equation; Priestley–Taylor equation; surface renewal; wetland

INTRODUCTION

Evapotranspiration (*ET*) constitutes the dominant water loss from many different types of wetlands. The latent heat flux process also represents the chief wetland energy sink (Priban and Ondok, 1985; Wessel and

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Rouse, 1994). The relative importance of *ET* is apparent in its influence over water depth, temperature and salinity (Burba *et al*., 1999) as well as areal extent of water coverage and inundation duration. Because *ET* is inextricably tied to water use and water availability, a considerable literature has accumulated on the subject. The initial interest in wetland *ET* came from agricultural engineers who were concerned with 'consumptive use' of wetland vegetation in arid regions such as southern California, Utah and Colorado in the USA (e.g. Otis, 1914; White, 1932; Blaney *et al*., 1933; Parshall, 1937; Mower and Nace, 1957). Wetlands were seen (and often still are) as wasting water that could better be used for irrigating agricultural crops or supplying domestic needs (Mower and Nace, 1957). For this reason, early information on wetland *ET* was used to make decisions about draining and/or clearing wetlands, and thereby reclaiming them for agricultural production (Linacre, 1976).

More recently, interest in wetland *ET* has been spurred by a wide variety of research and management needs. For example, there is great interest in quantifying water budgets (of which *ET* is often a major component) of both natural and managed wetlands in order to determine wetland water usage requirements and hydrological regimes (Carter, 1986; Rosenberry and Winter, 1997). Such information can then be used to estimate fluxes of contaminants and nutrients that enter, are retained, or leave wetlands (Drexler *et al*., 1999; Guardo, 1999). Related research has focused on differentiating between transpiration rates of particular wetland plants or plant assemblages and the evaporative demand of open water areas in order to better quantify overall consumptive use (Snyder and Boyd, 1987; Koerselman and Beltman, 1988; Pauliukonis and Schneider, 2001). There is also budding interest in using wetland *ET* as a means for phytoremediation of contaminants that enter the transpiration stream of plants (Nietch and Morris, 1999; Vroblesky *et al*., 1999). In addition, wetland *ET* estimates are required for modelling regional groundwater flow and contaminant transport as well as global climate change (e.g. Restrepo *et al*., 1998; Sun *et al*., 1998; Bartlett *et al*., 2002).

Several reviews have addressed the measurement and estimation of wetland *ET*, however, much has changed since they were published. Most of these papers were focused on particular ecosystems such as reedbeds and freshwater marshes (Linacre, 1976), bogs and fens (Ingram, 1983), and arctic tundra (Lafleur, 1990a). Both Linacre (1976) and Ingram (1983) provided important information on the implications of vegetation cover with respect to *ET* (particularly vascular transpiring vegetation), on the debate concerning whether wetland *ET* is greater than open water evaporation, and on the range of methods in use. Perhaps the greatest difference between these two reviews is that Linacre (1976) concluded that wetland vegetation is not an important determinant of *ET*, whereas Ingram (1983) concluded that vascular transpiring plants have a great influence on wetland *ET*. The review by Lafleur (1990a) focused on the importance of canopy or surface resistance and the supply of water as controls over wetland *ET*, two subjects that have received little attention. In addition to these reviews, Crundwell (1986) offered an elegant review of the debate over whether wetland *ET* or open water evaporation is greater in a given system. After a detailed assessment of both sides of the debate, Crundwell (1986) concluded that wetland *ET* generally appears to be greater than open water evaporation (at least during spring and summer months), although which is greater depends strongly on factors such as canopy size, plant species, climate, measurement method and plant density.

Despite considerable research, *ET* and many related physical processes remain poorly characterized for many wetland types (Souch *et al*., 1996). Even within well-studied wetland types, measurements of *ET* are often highly variable, precluding any generalization for particular plant communities or climatic regimes (Campbell and Williamson, 1997). One reason for this may be the variety of techniques that have been used to estimate *ET* and the substantial differences in their relative accuracies. Another reason may be that wetlands are very challenging environments in which to measure *ET* because they lack uniformity in shape, surface cover (e.g. percentage of bare soil, water and vegetation), hydrology and topography. Wetlands may also differ in their relative cover of transpiring and non-transpiring vegetation as well as surface litter. All of these components may have a significant effect on *ET* rates (Ingram, 1983; Campbell and Williamson, 1997). In addition, wetlands may be subject to the confounding influence of both air and water advection. Due to these problems, estimates of wetland *ET* often contain high degrees of error and/or have utility only during certain periods of the year.

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The purpose of this paper is to present and discuss the current state of the science with respect to wetland *ET* measurements and models. We have chosen to focus solely on modelling and micrometeorological methods (including a LIDAR-based technique) because, currently, these are the most common approaches for estimating *ET* at the site level (i.e. the ecosystem scale). Although other approaches exist, including lysimeters, the hydrological balance method and remote sensing, these techniques tend to be less accurate at the scale of an individual wetland (Villagra *et al*., 1995). This review encompasses only wetlands (e.g. marshes, peatlands, swamps, seasonal wetlands, etc.) and not lakes because our intent is to focus on the challenges that are specific to determining *ET* in wetland systems. Although this paper is centred on wetlands, our goal is not to provide a review of the myriad of *ET*-related papers in the wetland literature. Instead we have chosen to assess the difficulties in applying approaches to wetlands that originally were developed for highly managed and uniform agricultural systems and, in so doing, identify strategies for improving estimates of wetland *ET*.

THE WETLAND ENVIRONMENT

Cowardin *et al*. (1979) defined wetlands as '. . . lands transitional between terrestrial and aquatic systems where the water table is usually at or near the surface or the land is covered by shallow water. . . Wetlands must have one or more of the following three attributes: (1) at least periodically, the land supports predominantly hydrophytes; (2) the substrate is predominantly undrained hydric soil; and (3) the substrate is nonsoil and is saturated with water or covered by shallow water at some time during the growing season of each year. This definition encompasses a wide variety of wetland types that, in contrast to managed landscapes such as agricultural fields, often are a mosaic of different habitats covered to varying degrees by vegetation, open water or bare soil'.

Vegetation cover in wetlands may consist of trees, shrubs, macrophytes, bryophytes, algae and submerged and floating aquatic plants. Different functional groups of vegetation may have very different rates of water conductance to the atmosphere, thereby exerting varying degrees of control over *ET* (Koerselman and Beltman, 1988; Lafleur, 1990a). Rooted, emergent vascular plants may be of particular importance with respect to *ET*, because transpiration may progress unimpeded even in wetlands with no standing water as long as roots have access to the groundwater table (Ingram, 1983). The relative cover of bryophyte species such as *Sphagnum* also may be important because *ET* rates in wetlands with a significant bryophyte layer often exceed open water evaporation rates due to the wicking action of the mosses (Nichols and Brown, 1980). Other aspects of vegetation may also influence *ET* including plant density, species diversity, height and roughness of the dominant canopy, number of canopies, leaf characteristics, depth of litter layer and phenology. The albedo of vegetation may have a strong influence over *ET* depending on biochemical properties, orientation of leaves and leaf area index. Albedo of wetlands may change during the year due to emergence and senescence of the dominant vegetation, which in turn changes the relative cover of vegetation, open water and bare ground. Although many of these characteristics may exert only a minute influence over *ET* in a single plant, collectively these factors may significantly affect *ET* rates from a canopy (Penman, 1963; Ingram, 1983; Campbell and Williamson, 1997).

Open water areas in wetlands may vary in depth, expanse and duration depending on hydroperiod, climate, hydrogeomorphological setting and microtopography. The characteristics of open water areas that affect *ET* include depth of water, whether water is standing or flowing and water temperature. These factors influence how much energy the water will ultimately absorb, which in turn affects how much energy is subsequently available for *ET*. Although there has been considerable controversy over whether open water or vegetated areas contribute more to *ET* in wetlands (e.g., Linacre, 1976; Ingram, 1983; Snyder and Boyd, 1987; Pauliukonis and Schneider, 2001) the fact that wetlands are often a complex mixture of both leads to considerable difficulties in measurement. In addition to depth, water quality may also affect *ET*. For example, as salinity increases, *ET* rates decrease as a result of physiological control by plants and a reduction in the saturated vapour pressure (Oroud, 1995).

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The proportion of exposed bare soil within a wetland depends on many factors, including climate, soil nutrient status and salinity. In regions with seasonal precipitation, desiccation and salinity stress may result in large areas of bare soil. This is particularly common in mangrove-dominated areas in Australia and western Africa (Semeniuk, 1983; Hughes *et al*., 2001). Yet, bare soil is also often visible in small patches within particular vegetation types (e.g. cattail marshes) and in wetland types that experience seasonal drawdown of the water table, such as vernal pools, prairie potholes and cypress domes (van der Valk, 1981; Zedler, 1987; Ewel, 1998). If the bare soil surfaces dry out, the *ET* contribution decreases dramatically relative to open water or vegetated areas.

In addition to the above factors, shape and geographical setting of a wetland may also affect *ET*. Wetlands surrounded by surfaces with low evapotranspiration (e.g. bare dry soil) tend to have higher *ET* than those surrounded by forests (i.e. the oasis effect). Furthermore, long narrow wetlands such as riparian zones and marsh fringes around lakes tend to have higher *ET* rates than large expanses of wetlands with greater area-to-perimeter ratios (i.e. the 'clothes line' effect). The oasis effect is caused by advection over areas of variable wetness in a landscape and the clothes-line effect stems from increased evaporation resulting from ventilation through a small, isolated plant canopy (Linacre, 1976).

The above factors clearly demonstrate the complexity of wetland systems. Because *ET* in any landscape is a function of surface characteristics and micrometeorological factors (Penman, 1963), the essential challenge for micrometeorological measurement of wetland *ET* is adequately determining energy balance components in vegetated, open water and bare soil areas with varying microtopography and surface roughness.

ENERGY BALANCE

All of the *ET* measurement methods ultimately are based on the energy balance equation, which accounts for all the sources and losses of energy that are available for vapourizing water. The energy balance equation is

$$R_n = G + H + \lambda E + M + S \tag{1}$$

where R_n is the net radiation, G is the heat flux transfer to and from the soil and water, H is the sensible heat flux density, λE is the latent heat flux, M is the energy flux used for photosynethesis and respiration, and S is the energy transfer into and out of plant tissue. For this review, we will assume M and S are generally small enough to be disregarded. A full discussion of these energy components is contained in the Appendix.

MICROMETEOROLOGICAL METHODS FOR DETERMINING WETLAND *ET*

The most common methods for measuring evapotranspiration in wetlands are the Bowen ratio energy balance (BREB) and eddy covariance (EC) methods. Although less commonly used, the surface renewal (SR) and LIDAR methods show some promise for improving wetland *ET* measurements. Each of these methods has advantages and disadvantages that are presented below. A common assumption in using these methods is that there is adequate 'fetch' (i.e. upwind distance having uniform features) required to ensure that the measurement is representative of the underlying surface and not contaminated by the flux from a distant surface. Because stability changes over the day, the fetch requirements also change. During daylight hours, when there is less stability and more turbulence, the fetch requirement is less than at night when the atmosphere is more stable. However, as evapotranspiration from wet surfaces occurs mainly during the day, fetch requirements tend to be less important for λE measurements than for CO_2 or other gas fluxes that are less dependent on available energy. Generally, 50 m of fetch for each metre that the highest instrument is above the ground should be adequate for λE measurements.

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Bowen ratio energy balance method

A well-known approach to measure evapotranspiration is the BREB method, which was first proposed by Bowen (1926). Starting with the basic energy balance equation (Equation 1), dividing both sides by λE we obtain $(R_n - G)/\lambda E = \beta + 1$, where $\beta = H/\lambda E$ is the Bowen ratio. Solving the equation for λE gives

$$\lambda E = \frac{R_n - G}{1 + \beta} \qquad \text{W m}^{-2} \tag{2}$$

By substituting the expressions for sensible heat flux density (Equation A5, see Appendix) for H and latent heat flux density (Equation A6) for λE, the Bowen ratio becomes

$$\beta = \frac{H}{\lambda E} = \frac{\dfrac{(T_2 - T_1)}{r_a}}{\dfrac{1}{\gamma}\left(\dfrac{e_2 - e_1}{r_a}\right)} = \gamma\left(\frac{T_2 - T_1}{e_2 - e_1}\right) \tag{3}$$

This equation demonstrates that by measuring temperature and vapour pressure at two heights, one can calculate the Bowen ratio (Equation 3). Then, using measurements for R_n and G, together with the calculated β, λE is determined using Equation 2. Large errors in latent heat flux density can occur near sunrise and sunset, when the Bowen ratio $\beta = H/\lambda E \approx -1{\cdot}0$ as the denominator of Equation (2) approaches zero. In such cases, λE can be estimated from surrounding measurement periods. Like other energy balance methods, the BREB is an indirect method to estimate λE. Only the temperature and humidity are measured directly to determine the ratio $H/\lambda E$. Consequently, accurate measurement of R_n and G as well as β are needed to properly estimate λE.

When using a BREB system, the surface is assumed to be horizontally homogeneous, resulting in only vertical energy transport. However, horizontal heterogeneity is common in wetlands, and therefore use of the BREB may lead to error. In addition to having a homogeneous surface, adequate fetch is very important to ensure that the data are collected within the fully adjusted boundary layer. For example, Monteith and Unsworth (1990) recommend a fetch near 100 : 1 for measurements over uniform vegetation. However, Fritschen *et al.* (1983) and Heilman *et al.* (1989) reported field BREB measurements that were relatively insensitive to fetch when small β values were determined. However, when β is small, H is small and $\lambda E \approx R_n - G$ will give results similar to Equation (2) without having to measure the Bowen ratio.

Because λE tends to be high in wetland systems, upward positive H values are likely to be small most of the time. However, in arid environments, extra energy from regional advection can lead to large negative H values over wetland systems in the afternoon and evening (Allen *et al.*, 1992). The BREB method is quite accurate when β is close to zero, but the relative error in λE grows rapidly as the absolute magnitude of β increases (Angus and Watts, 1984).

Early versions of 'psychrometer' BREB systems, which measure dry-bulb and wet-bulb temperature at the two heights, require a mechanical system to periodically interchange the arms to eliminate bias in the temperature and humidity measurements. This is necessary because bias in the temperature and humidity measurements can lead to systematic β measurement errors. Although the early Bowen ratio systems give good results, they are more expensive and less portable. More recently, a Bowen ratio system that alternately draws air samples from inlets at the same two heights as the temperature sensors and uses the same dew point hygrometer to determine the dew point temperature (and vapour pressure) commonly has been used to estimate *ET*. A disadvantage of the hygrometer Bowen ratio system is that air is drawn through PVC tubing to the hygrometer for measuring the dew point. Condensation within the PVC lines at night can be a problem, and one solution is to turn off the pump at night.

Two pairs of high-resolution matched temperature and humidity sensors are needed to determine β for systems with interchangeable arms, and matched temperature sensors are needed for hygrometer-based systems. Both sensor arms should be high enough to be in the zone of the logarithmic wind speed profile. If

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the vegetation is tall, the lower sensor may be high above the ground. The upper sensor must be far enough above the lower sensor to detect temperature and humidity differences. For tall canopies, several hundreds of metres of fetch may be required. As a canopy grows taller during the season, the sensor arms might need to be raised. Therefore, a BREB system should be placed so that there is adequate fetch when the arms are at the highest position during the season.

Possible errors in using BREB measurements to estimate *ET* are discussed in Nie *et al.* (1992). They used a transportable BREB system (a 'rover') in 14 different sites and compared short-term measurements with other BREB systems using a variety of methods to measure R_n and β. In general, the error in β was about 10% between systems with identical instrumentation. They noted that when the measured temperature and humidity gradients were small, variations in β between systems were large. Some of the difference was attributed to variations in the psychrometric constant used and how the half-hour averages were calculated. The mean λE difference between the BREB systems of various designs and the rover system ranged between $-1{\cdot}9$ and $29{\cdot}2$ W m^{-2}. However, the half-hour differences varied between $-112{\cdot}0$ and $97{\cdot}0$ W m^{-2}. Because the correct λE was unknown, the percentage error cannot be determined for the various systems. However, the relative λE difference between the rover and the other BREB systems varied between $4{\cdot}6$ and $-18{\cdot}7\%$. Instrument mounting and site description information were not provided, but the experiment probably was conducted over cropped fields in Kansas, so the canopies were most likely smooth and uniform in contrast to most wetland ecosystems.

Examples of wetland estimates. The BREB is a commonly used micrometeorological method for determining wetland *ET*, so there are many papers on the use of BREB over wetlands. Of particular interest is a suite of studies carried out in tussock tundra of the Hudson Bay Lowland (see Wessel and Rouse (1994), Rouse (1998), and the papers listed therein). In Wessel and Rouse (1994), temperature, vapour pressure and wind speed were measured at several heights to determine the Bowen ratio. Although limited energy balance data were reported, the λE and daily *ET* values matched well with the weighted Penman–Monteith estimates of *ET*. The BREB also has been applied to small wetlands in arid Utah (Allen *et al.*, 1994). Allen *et al.* (1994) measured *ET* over a freshwater marsh dominated by cattails (*Typha* spp.) and bulrush (*Scirpus* spp.) using two hygrometer-type Bowen ratio systems. They reported the highest values for λE (near 800 W m^{-2} during midday) of any papers on wetland *ET*.

Eddy covariance method

Turbulence in the atmospheric boundary layer dominates mixing and the diffusion of sensible and latent heat to and from the underlying surface. Using sonic anemometers, turbulent 'eddy flux' motions are measurable with a high level of precision and with a high degree of spatial and temporal resolution. If the transported energy (i.e. sensible and latent heat) is measured with equivalent precision and resolution, it is possible to monitor the sensible and latent heat flux density using the eddy covariance (EC) method.

The vertical component of the fluctuating wind is responsible for the flux across a plane above a horizontal surface. Because there is a net transport of energy across the plane, there will be a correlation between the vertical wind component and temperature or water vapour. For example, if water vapour is released into the atmosphere from the surface, updrafts will contain more vapour than downdrafts, and vertical velocity (positive upwards) will be positively correlated with vapour content. The covariance of vertical wind speed with temperature and water vapour are used to estimate the sensible and latent heat flux density. The averaging period must exceed the duration of the largest eddy involved in the transport process, so 10–30 min periods are often used.

The fluctuations and the time-averaged components are expressed as $\rho_v = \overline{\rho_v} + \rho_v'$, where ρ_v is the absolute humidity (kg m^{-3}). The overbar signifies a time average over a specified interval of time and the prime indicates a departure from the mean. The vertical velocity component w (m s^{-1}) is treated similarly, resulting in $w = \overline{w} + w'$, where w is the vertical wind speed. By definition, $w'\overline{\rho}_v = 0$ and $\overline{w}\rho_v' = 0$. Therefore, the flux

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density for water vapour E (kg m^{-2} s^{-1}) can be written as $E = \overline{\overline{w}\overline{\rho_v}} + \overline{w'\rho_v'} = 0$. If there is no convergence or divergence of air owing to a sloping surface, the mean vertical velocity ($\overline{w}$) and hence the first term on the right-hand side of the equation equals zero. This simplifies the equation to $E = \overline{w'\rho_v'}$. Therefore, the flux density for water vapour is equal to the mean covariance of the fluctuations of the vertical wind from its mean and the absolute humidity from its mean.

Because of air density perturbations as air parcels move up and down past the sensors, the 'WPL' correction (Webb *et al.*, 1980)

$$E = 1{\cdot}010(1 + 0{\cdot}051\beta_r)E_r \qquad \text{kg s}^{-1}\ \text{m}^{-2} \tag{4}$$

is needed to determine water vapor fluxes (Webb *et al.*, 1980) for open-path sensors. The WPL correction is not needed for closed-path sensors where air is pumped into a closed chamber at a known pressure to measure the water vapour content. In Equation (4), β_r is the uncorrected Bowen ratio ($\beta_r = H/\lambda E_r$) and E_r is the uncorrected vapour flux density.

The EC method requires sensitive, expensive instruments to measure high-frequency wind speeds and scalar quantities. Sensors must measure vertical velocity, temperature and humidity with sufficient frequency response to record the most rapid fluctuations important to the diffusion process. Typically, a frequency of the order of 5–10 Hz is used, but the response-time requirement depends on wind speed, atmospheric stability and the height of the instrumentation above the surface. Outputs are sampled at a sufficient rate to obtain a statistically stable value for the covariance. Wind speed and humidity sensors should be installed close to each other but separated sufficiently to avoid interference. When the separation is too large, an underestimate of the flux may result (Lee and Black, 1994).

High-frequency wind vector data usually are obtained with a triaxial sonic anemometer. In some earlier studies, one-dimensional sonic anemometers were used, but they are problematic because data cannot be corrected post-experiment for sensor or mean streamline tilt. The triaxial instruments provide the velocity vector in all three directions, and, therefore, corrections can be applied for any tilt in the sensor and mean streamline flow. Triaxial sensors with non-orthogonal sound paths perform an internal coordinate rotation to provide signals of three orthogonal velocities from a non-orthogonal transducer path array. If one microphone/transducer in the sensor array malfunctions, the entire velocity vector output is corrupted. Triaxial sensors with orthogonal sound paths do not need internal coordinate rotation, but tend to have probe designs that interfere with the airflow more than those with the non-orthogonal designs. Typically, pulses are repeated up to approximately 200 times per second and the output frequency is between 10 and 20 times per second, which improves the signal to noise ratio. A wide range of humidity sensors have been used in eddy covariance systems including thermocouple psychrometers, Lyman-alpha and krypton hygrometers, laser-based systems and other infrared gas analysers.

Nie *et al.* (1992) compared eddy covariance measurements from three sites with a 'rover' BREB system and found a mean difference in λE of $-8{\cdot}8$ W m^{-2} with a range of half-hour differences varying between -112 and $51{\cdot}8$ W m^{-2}. However, there was no independent measure of λE, so it is unknown whether the BREB or EC method was more accurate. In all three of the EC sites, there was a sizeable residual after comparing $R_n - G$ with $H + \lambda E$ (i.e. lack of energy closure). In general, the daytime λE values from the EC systems averaged 6·8 to 23·5 W m^{-2} (i.e. 2·7 to 9·7%) lower than the rover BREB system. In another comparison study, intensive EC and BREB measurements were taken over a peat bog consisting of a mixture of sphagnum moss, lichen hummocks and black pools (den Hartog *et al.*, 1994). The hummocks had a dry, insulating surface that covered ice down to about 3 m deep. They reported daytime Bowen ratio values near $\beta = 1{\cdot}0$ and eddy covariance $H + \lambda E$ of about 90% of $R_n - G$. During mid-summer, even with high daytime surface temperatures near 40 °C, the insulating materials prevented the ice from melting. The H and λE from the eddy covariance measurements averaged about 0·81 and 0·86 of the Bowen ratio values, respectively. The authors attributed the differences to difficulty in measuring representative R_n and G values over the 'mosaic' surface.

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Examples of Wetland Estimates. Researchers have used EC to measure ET in a variety of wetlands such as subtropical freshwater marshes (Bidlake *et al.*, 1996), a cypress swamp (Bidlake *et al.*, 1996), temperate marshes (Souch *et al.*, 1996; Souch *et al.*, 1998; Bidlake, 2000) and a salt marsh/mangrove complex (Hughes *et al.*, 2001). The EC approach is exceptional in the fact that if other energy components are measured, the EC method self tests for energy balance closure (i.e. when measured $\lambda E =$ measured $R_n - G - H$). However, several challenges still exist in applying EC to wetlands including: (i) adequately accounting for errors in measurements, (ii) achieving good results under changing hydrological conditions and (iii) keeping instruments operational for long periods in order to assess interseasonal variability.

Surface renewal method

Paw U and Brunet (1991) introduced the surface renewal (SR) method, which is based on the premise that air near a surface is renewed by ambient air from aloft. Using this approach, H is estimated based on measurements of high-frequency air temperature fluctuations, and λE is obtained as the residual of the energy budget equation (Equation 1). To estimate H, the air temperature fluctuations, which exhibit rapid increases and decreases (ramp-like structures), are analysed to estimate amplitude and duration of the temperature ramps (Gao *et al.*, 1989; Paw U *et al.*, 1995). The amplitude and duration of the temperature ramps are then used to estimate sensible heat flux density using the following equation

$$H = \alpha \rho C_p \left(\frac{a}{d+s} \right) z \qquad \text{W m}^{-2} \tag{5}$$

where α, a correction for unequal heating below the sensors, depends on z (which is the measurement height, m), on canopy structure and on thermocouple size (Paw U *et al.*, 1995; Snyder *et al.*, 1996; Spano *et al.*, 1997a). The symbol ρ is for the air density (g m^{-3}) and C_p is the specific heat of air at constant pressure (J g^{-1} K^{-1}). The ratio $a/(d+s)$ represents the mean change in temperature (°C or K) with time (s) during the sampling interval (usually 30 min), where a is the mean ramp amplitude and $d+s$ is the sum of the ramp period (d) and the quiescent period (s) between ramps. A more thorough discussion of the surface renewal method can be found in Snyder *et al.* (1996) and Spano *et al.* (1997a,b, 2000).

Surface renewal is a relatively new method to estimate H. Currently, a drawback of the method is that it must be calibrated against a sonic anemometer to account for unequal heating of air parcels below the temperature sensor height and other potential deviations from the assumptions used in formulating surface renewal theory. Once determined, the calibration factor is unlikely to change unless there are significant changes in the vegetation canopy (Paw U *et al.*, 1995; Snyder *et al.*, 1996; Spano *et al.*, 1997a,b, 2000). Therefore, the calibration factor for a particular canopy can be used regardless of the weather conditions. For a non-uniform wetland surface, the α value needs to be determined for each unique surface and may need recalibration if the surface vegetation changes. A great advantage of the method is that measurements can be replicated at a lower cost. Because wetland systems are characterized by variable surfaces, replication to obtain good spatial representation of λE can greatly improve ET estimates.

When estimating H values using the SR method and a sonic anemometer in the EC method, the root mean square errors are typically in the range of 30 to 50 W m^{-2} and 30 W m^{-2}, respectively. Therefore, SR estimates of λE should have nearly comparable accuracy to λE estimated as the residual of the energy balance equation using H from a sonic anemometer. If R_n and G are measured accurately, the SR method should give good estimates of λE.

Examples of estimates. Zapata and Martinez-Cobb (2001) used the SR method to measure λE from an endorreic lagoon, an aquatic environment characterized by short, sparse vegetation with predominantly bare soil. They found a high correlation between surface renewal and eddy covariance H values. The vegetation was mostly less than 0·5 m tall and they reported root mean square errors between eddy covariance and SR estimates of H of the order of 30 W m^{-2} for temperatures recorded at 0·9 m and 1·1 m using an 8 Hz

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sampling rate and a time lag of 0·75 s. Snyder *et al*. (1996) and Spano *et al*. (1997b) reported similar results over uniform, well-watered grass, wheat and sorghum canopies. In addition, Spano *et al*. (2000) found good relationships between eddy covariance and surface renewal measurements over a sparse grape vineyard.

LIDAR and other laser-based methods

A LIDAR (light detection and ranging) system is similar to radar, but consists of a laser used to transmit electromagnetic radiation in the infrared, visible or ultraviolet range. The emitted radiation is backscattered (elastic or inelastic) and then received by an optical telescope using sensitive photomultiplier tubes or other sensors. The wavelength of the radiation and the amount of scattering that occurs between the emitter and receiver can be used to measure temperature, wind and concentrations of various atmospheric constituents. An adaptation of LIDAR called the solarblind Raman water vapour LIDAR has been used to measure concentrations of both nitrogen and water vapour in the atmosphere (Renault *et al*., 1980; Cooney *et al*., 1985). This system is based on Raman (inelastic) scattering. Nitrogen and water vapour measurements are required in order to normalize water vapour concentration by that of nitrogen, the dominant gas in the atmosphere. This procedure corrects for first-order atmospheric transmission effects, variability in the energy of the laser through time, and the overlap of the telescope with the laser beam.

The LIDAR system outputs hundreds of water vapour mixing ratio (i.e. mass of water vapour per unit mass of dry air) gradients over the measurement surface. Micrometeorological measurements at a location within the measurement area are used to estimate the roughness length, surface specific humidity and temperature, specific heat at some height z above the surface, and the friction velocity. These data are used to determine turbulent heat and momentum fluxes over the surface using Monin–Obukhov similarity theory. Linear regression between the water vapour mixing ratio profiles and the similarity functions are used to estimate λE at hundreds of points over the surface. The method to estimate λE from LIDAR data and Monin–Obukhov similarity theory is presented in Eichinger *et al*. (2000). The main advantage of using LIDAR over other methods is that it provides a spatially integrated measure of water vapour flux over mixed terrain and canopy rather than simply a point measurement for a particular area.

Tunable laser diode systems have been used to measure the turbulent changes in gaseous concentrations near the land surface, although because of signal-to-noise limitations on the resolution of these devices, conventional gradient techniques have been used (Simpson *et al*., 1995, 1997). Because of the large expense of these systems, they are currently used for trace gases other than water vapour.

Examples of wetland estimates. Eichinger *et al*. (2000) reported on the use of Raman LIDAR to measure water vapour content of air and estimate evapotranspiration in the upper San Pedro Basin in Arizona, an area consisting of riparian wetland, desert shrub-steppe, grassland, oak savanna and pine woodland. Their LIDAR system emitted a pulsed ultraviolet laser beam and measured the water vapour signal at a wavelength of 273 nm. Eichinger *et al*. (2000) reported that the horizontal range of the LIDAR system was about 700 m and it had a spatial resolution of 1·5 m. Uncertainty in the water vapour mixing ratio was reported to be less than 4%.

In other studies, a scanning Raman LIDAR has been used to measure water vapour mixing ratio profiles over various vegetation surfaces and to analyse for spatial fluxes (Cooper *et al*., 1998, 2002). In their 1998 study, Cooper *et al*. developed spatial maps of latent heat flux density from a riparian cottonwood forest in the San Pedro River Basin in Arizona. The results were compared with evapotranspiration estimates from sap flow measurements. The sap flow estimates of latent heat flux density were approximately 20 W m^{-2} higher than latent heat flux density measured with LIDAR. Based on the spatial variability of the measurements, the authors concluded that using a micrometeorological point measurement of evapotranspiration rather than spatial measurements by LIDAR in the cottonwood forest would be misleading.

In Cooper *et al*. (2002), Raman LIDAR was used to assess advection, edge and oasis effects on latent heat flux density over a riparian wetland along the Rio Grande River in south central New Mexico. Such

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effects often present difficulties in measuring *ET* in a variety of wetlands. In the study, the authors compared measurements when the wind came from a hot, arid desert area and from a cooler, humid area where the air passed over the tamarisk canopy. Cooper *et al*. (2002) showed that LIDAR identifies water vapour profile differences when warm, dry air advects over a canopy, thus providing a method to estimate advection into canopies.

MODELS FOR ESTIMATING WETLAND EVAPOTRANSPIRATION

Empirical equations

In empirical methods, specific relationships are determined between measured environmental parameters and *ET*, usually through the use of least squares regression techniques. The environmental parameters used may be quite elaborate or relatively simple. Empirical models developed by Holdridge (1962), Blaney-Criddle (1950), Linacre (1977) and Thornthwaite (1948) are temperature-based. The latter method has been widely used and is written as

$$PET_i = 1{\cdot}6L_i\left(\frac{n_i}{30}\right)\left(10\frac{T_i}{I}\right)^A \tag{6}$$

Where PET_i is potential monthly evapotranspiration, T_i is monthly mean temperature, L_i is monthly mean day length at the given latitude in units of 12 hours, and n_i is the number of days for the ith month. The exponent A is given by

$$A = (6{\cdot}75 \times 10^{-7} \times I^3 - 7{\cdot}71 \times 10^{-5} \times I^2 + 1{\cdot}792 \times 10^{-2} \times I + 0{\cdot}49239)$$

where $I = \sum_{i=1}^{12}(T_i/5)^{1{\cdot}514}$. Note that the monthly PET_i values from Equation 6 are estimates of the evaporative demand of the atmosphere. PET_i is not an estimate of wetland *ET*. Other empirical models are based on relationships between wetland *ET* and Penman's open water evaporation formula or pan evaporation (Sturges, 1968; Koerselman and Beltman, 1988; Lafleur, 1990b), two measurements that are available throughout much of the world. In addition, other models have been developed that are based on multiple regression of two or more measured components. For example, Dolan *et al*. (1984) used average above-ground live biomass and average saturation deficit to estimate *ET*. Eisenlohr (1966) used wind speed, saturation vapour pressure deficit and a mass transfer coefficient. Rouse (1998) used net radiation and temperature. Snyder and Boyd (1987) developed a model using daily solar radiation, plant height/leaf area index and number of days in a month. Priban and Ondok (1985) estimated *ET* using net radiation and mean relative humidity.

Currently, the most widely used empirical model is the Priestley–Taylor (PT) equation (Priestley and Taylor, 1972)

$$\lambda E = \alpha'\frac{\Delta}{\Delta+\gamma}(R_n - G) \tag{7}$$

where Δ (kPa °C^{-1}) is the slope of the saturation vapour pressure curve at air temperature T (°C), γ is the psychrometric constant (kPa °C^{-1}) and α' is an empirically derived term set to 1·26 by Priestley and Taylor (1972) to estimate λE from well-watered, vegetated canopies and water surfaces. The Tetens equation (Tetens, 1930) can be used to determine values for Δ as: $\Delta \approx 4098e_s/(T + 237{\cdot}3)^2$, where $e_s = 0{\cdot}6108\exp(17{\cdot}27T/(T + 237{\cdot}3))$ is the saturation vapour pressure at temperature T (°C). Simpler polynomial expressions can be found for the vapour pressure and derivatives of these equations are easy to obtain (Paw U and Gao, 1988). Because the PT equation was developed in an empirical manner, it is not clear that α' should be equal to 1·26 for wetland surfaces. Paw U and Gao (1988) note many cases for vegetation where α' is not equal to 1·26. However, if properly calibrated against an independent measure of *ET* (such as the BREB or EC method), the PT equation may work well during certain time periods in

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favourable locations. If the Bowen ratio, $\beta = H/\lambda E$, is incorporated into the PT equation, then the α factor from Equation (7) is computed as

$$\alpha = \frac{\Delta + \gamma}{\Delta(1 + \beta)} \tag{8}$$

Calibration of the α factor in concert with micrometeorological measurements can prove highly useful because then the PT equation can be used to fill in missing data during periods of equipment failure or malfunction.

Examples of wetland estimates. Dolan *et al.* (1984) tested the Thornthwaite (1948) and Linacre (1977) empirical methods for estimating wetland evapotranspiration. They monitored changes in water-table level of a Florida freshwater wetland and reported that neither of the two methods gave consistent estimates of *ET* as estimated with water-table levels. They also compared the measured evaporation rates with pan evaporation and found inconsistent results.

Souch *et al.* (1996) evaluated the use of the PT equation for estimating wetland *ET* at the Indian Dunes National Lakeshore, Indiana. They found that the PT equation with $\alpha' = 1{\cdot}26$ somewhat overpredicted measured *ET*. However, using equilibrium conditions with $\alpha' = 1{\cdot}0$ gave good estimates of wetland *ET*. When $\alpha' = 1{\cdot}0$

$$\lambda E = \frac{\Delta}{\Delta + \gamma}(R_\mathrm{n} - G) = R_\mathrm{n} - G - H \qquad \mathrm{W\ m^{-2}} \tag{9}$$

This situation can occur when the air is saturated ($e_\mathrm{s} - e = 0$) over a wet surface or when H is positive and equal to the fraction of energy heating the air in a diabatic process $H = (\gamma/(\Delta + \gamma)(R_n - G)$, which can happen during cold air advection. The PT equation is meant to estimate the *ET* of a short canopy and the aerodynamic and surface resistance of the wetland surface at the Indiana Dunes National Lakeshore was probably different from a short canopy. Therefore, $\alpha' \neq 1{\cdot}26$ is not unexpected. If the prevailing wind were from Lake Michigan, high humidity and cold-air advection might partially explain the observed lower value for α'. Bidlake (2000) used the PT equation to measure wetland *ET* in Oregon (an arid climate). In that study, Bidlake found that α' values varied, but using $\alpha' = 1{\cdot}0$ gave λE estimates that were highly correlated with λE measured using eddy covariance. He was able to improve the λE estimates by calibrating the α' values for different times of the year. Seasonal changes in canopy and aerodynamic resistance, advection and humidity are the probable explanations for changes in α' during the year.

Combination equations

The Penman (1948, 1963) and Penman–Monteith (PM) equations (Monteith, 1965) are common combination methods, which estimate *ET* by accounting for both radiation and aerodynamic contributions of energy for the vapourization process. Both of these equations use psychrometric concepts to estimate diabatic and adiabatic contributions to evapotranspiration using air temperature and humidity data collected above a canopy. The two equations are derived using flux gradient equations (Equations A5, A6 and A10) and the saturation vapour curve with a first-order Taylor approximation. The PM equation is expressed as

$$\lambda E = \frac{\Delta(R_\mathrm{n} - G) + \rho C_\mathrm{p}\left(\dfrac{e_\mathrm{s} - e}{r_\mathrm{a}}\right)}{\Delta + \gamma^*} \tag{10}$$

where γ^* (Eq. A11) is a modified psychrometric constant that accounts for surface resistance to water vapour flux. The Penman (PE) equation

$$\lambda E = \frac{\Delta(R_\mathrm{n} - G) + \rho C_\mathrm{p}\left(\dfrac{e_\mathrm{s} - e}{r_\mathrm{a}}\right)}{\Delta + \gamma} \tag{11}$$

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is a special case of the PM equation with no surface resistance ($r_s = 0$) and $\gamma^* = \gamma$. This condition is likely when the evaporating surface is wet (e.g. after rainfall). It is not true at night if plant surfaces are dry and the stomata are closed.

Examples of wetland estimates. Several researchers have attempted to use the PM equation to estimate wetland *ET*. Souch *et al*. (1996, 1998) found good agreement among the Penman equation, PM equation, PT equation with $\alpha' = 1{\cdot}0$ and wetland *ET* measured using EC in the Indiana Dunes National Lakeshore, Indiana, USA. Note that Souch *et al*. (1998) used the Penman (1948) PE equation, with a calibrated wind function

$$\lambda E = \frac{\Delta}{\Delta + \gamma}(R_n - G) + \frac{\gamma}{\Delta + \gamma}[6{\cdot}43(1 + 0{\cdot}53u)(e_s - e)] \tag{12}$$

where u is the wind speed at 2 m height. Bidlake (2000), using a fixed surface resistance, also reported a good match between the PM equation and wetland *ET*. When the canopy resistance was varied with time of the year, the PM equation matched measured *ET* even better.

Wessel and Rouse (1994) measured *ET* from a wetland tundra in the Hudson Bay Lowland near Churchill, Manitoba, Canada, using the BREB method and compared the results with the PM equation. The PM equation is based on the assumption of a complete canopy cover and the surface is treated like a 'big' leaf. Wessel and Rouse estimated the canopy resistance as: $r_c = r_s/LAI$, where r_s is the mean stomatal resistance and *LAI* is the leaf area index. The root mean square error between the PM equation and the BREB measurements was about 97·3 W m^{-2}. The inaccuracy in the PM equation was attributed to not accounting for surface resistance of the exposed soil and water surfaces.

Weighted canopy methods

Shuttleworth and Wallace (1985) presented a model (SW) that separates evapotranspiration into soil evaporation and transpiration for estimating evapotranspiration from sparse canopies. Wessel and Rouse (1994) presented a similar method (WR), but accounted for evaporation from water surfaces as well as from the soil. In both models, the Penman–Monteith equation was used. The SW model uses the equation

$$Q_e = Q_c + Q_s \tag{13}$$

and the WR model uses the equation

$$Q_e = LAI \times Q_c + S \times Q_s + W \times Q_w \tag{14}$$

where Q_c, Q_s and Q_w are the PM estimates of *ET* over the canopy, soil and water, respectively, *LAI* is the leaf area index, S is the fraction of exposed soil and W is the fraction of exposed water surface. In the SW model, the canopy *ET* is calculated using

$$Q_c = \frac{\Delta(AE - AE_s) + \dfrac{\rho C_p(e_s - e)}{r_a^c}}{\Delta + \gamma\left(1 + \dfrac{r_c}{r_a^c}\right)} \tag{15}$$

where AE is the diabatic energy to the wetland, AE_s is the available energy to the soil surface, $e_s - e$ is the vapour pressure deficit at mean canopy level, r_c is the bulk stomatal resistance and r_a^c is the boundary layer resistance. The soil evaporation component is given by

$$Q_s = \frac{\Delta AE_s + \dfrac{\rho C_p(e_s - e)}{r_a^s}}{\Delta + \gamma\left(1 + \dfrac{r_s^s}{r_a^s}\right)} \tag{16}$$

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where AE_s is the diabatic energy available at the soil surface, r_s^s is the soil surface resistance and r_a^s is the aerodynamic resistance below the mean canopy level. For the WR model, the three component ET values are calculated as

$$Q_c = \frac{\Delta AE_c + \dfrac{\rho C_p(e_s - e)}{r_a}}{\Delta + \gamma\left(1 + \dfrac{r_c}{r_a}\right)} \tag{17}$$

where AE_c is the available energy $(R_n - G)$ for canopy covered surfaces

$$Q_s = \frac{\Delta AE_s + \dfrac{\rho C_p(e_s - e)}{r_a}}{\Delta + \gamma\left(1 + \dfrac{r_s^s}{r_a}\right)} \tag{18}$$

where AE_s is the available energy $(R_n - G)$ for the bare soil surfaces and r_s^s is the soil surface resistance, and:

$$Q_w = \frac{\Delta AE_w + \dfrac{\rho C_p(e_s - e)}{r_a}}{\Delta + \gamma} \tag{19}$$

where AE_w is the available energy $(R_n - G)$ for the open water surfaces.

Examples of wetland estimates. Wessel and Rouse (1994) compared the Shuttleworth–Wallace (SW) method for estimating ET from wetland tundra in Manitoba, Canada, with the BREB method. They conducted surveys to determine the percentage of hummock, hollow and open water areas in the wetland throughout the season and measured R_n and G over the three surfaces. When compared with BREB measurements, the SW and WR models predicted λE with a root mean square error of 150 W m^{-2} and 38·3 W m^{-2}, respectively. Part of the problem with using the SW method in this study was that the model separates the surface into soil and vegetation areas, but the wetland had bare soil, vegetation *and* standing water. Soil heat flux density was measured over the three surfaces, but how the bare soil and water measurements were combined was not explained.

Canopy cover coefficient method

The crop or canopy cover coefficient (CCC) method is a robust approach used by agronomists to estimate ET from agricultural and horticultural crops (Allen *et al.*, 1994). Evapotranspiration from a crop surface is calculated as a function of K_c, the crop coefficient derived for a particular plant species, and ET_o, a reference evapotranspiration value that characterizes the evaporative demand of a region:

$$ET = ET_o \times K_c \tag{20}$$

Perhaps the most widely accepted measure of ET_o is the American Society of Civil Engineers (ASCE) standard reference evapotranspiration in which ET_o is estimated using the PM equation for a broad expanse of short vegetation with $r_s = 0{\cdot}50$ s m^{-1} during daylight hours and $r_s = 200$ s m^{-1} during nighttime (Walter *et al.*, 2000). In the CCC method, ET_o is a measure of evaporative demand and the K_c value accounts for differences between a particular canopy and ET_o. Typically, wetland ET estimates are achieved by developing monthly K_c values during the growing season.

Several wetland researchers have applied the CCC method to particular wetland types and plant species. Such efforts have met with mixed results for several reasons. First of all, the CCC method works best in wetlands that have a relatively uniform canopy surface, and highly predictable plant growth and development. These conditions may be met only in wetlands with nearly monotypic stands of vegetation such as cattail

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(*Typha* spp.) and bulrush (*Schoenoplectus acutus* or *Scirpus*) marshes, reedbeds (*Phragmites australis* marshes), salt marshes dominated by cord grass (*Spartina* spp.) and treatment wetlands dominated by *Typha* or other emergent macrophytes. Secondly, to use the K_c values, the biological and environmental conditions must also be nearly the same at a particular wetland as the site(s) in which K_c values originally were developed. Lastly, various approaches have been used to determine ET_o resulting in variable K_c values for the same wetland plant species. For this reason, standardization of the ET_o equation would be a major step toward reducing errors and improving the overall accuracy of the CCC method.

Examples of wetland estimates. Table I contains a list of wetland plant species and plant communities for which K_c values have been determined. The range of values for the same species can be attributed to different climate regimes, errors in the *ET* measurements used to develop the K_c values, and differences in the ET_o equations. Clearly, for the CCC method to be broadly applicable, more K_c values are needed for wetland plant species and more data are required on how K_c values change during the growing season. In future research, it would be highly advantageous for researchers to use the same standard equation for ET_o as used in agriculture (e.g. from Walter *et al*., 2000). This would allow for easier comparison and sharing of wetland K_c values.

COSTS OF MEASUREMENT AND ESTIMATION METHODS

Whether or not a researcher should attempt to measure wetland *ET* directly or use one of the combination equations depends to a large extent on the cost of the instrumentation required as well as its accuracy and complexity. Table II provides estimates of the costs for each of the measurement systems and a weather station for measuring parameters needed for combination equations. The cost of net radiation sensors ranges from about \$900 to \$4500. For the purpose of Table II, a conservative estimate of \$1000 was chosen. Each system has different data logger requirements, so the least expensive data logger price that will work for the method was used in the cost estimate. The same costs for two soil heat flux plates and a soil-averaging temperature probe were used for each system. The costs do not include additional supplies and equipment for data transfer or for computers to process the data.

FUTURE DIRECTIONS

Evapotranspiration estimates for wetlands may be improved by numerous technological advances. Until the advent of relatively inexpensive, easy-to-use micrometeorological equipment such as sonic anemometers in the mid-1980s, direct measurement of *ET* was difficult and only more indirect methods such as BREB were used. Future development of inexpensive, but highly accurate humidity sensors, coupled with low-power wireless data networks could allow micrometeorological techniques to be extended to wetlands with limited fetch or high degrees of heterogeneity. Advective components could be measured directly with such sensors. Miniaturization of sonic anemometers, or other wind velocity sensors could allow more detailed analysis of the contributions to *ET* from individual wetland components. If inexpensive laser-based systems could be produced, *ET* could be estimated from a combination of gradient-based measurements or direct eddy covariance. Further development of surface renewal techniques might remove the requirement of initial calibration. The use of radiatively-sensed surface temperatures, and perhaps other remotely sensed variables, in partnership with other micrometeorological measurements, could yield *ET* estimates.

Advanced modelling methods, although more complex than the equations presented here, could be applied to wetland *ET* estimates, with limited measurements required (Paw U and Meyers, 1989; Pyles *et al*., 2000). Isotopic analysis techniques might enable the apportioning of *ET* to each surface type within the wetland.

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Table I. Crop cover coefficient (K_c) values available for wetland plants and plant communities

Study site	Technique	Species	K_c	Researcher
Reedbed, Russia	Lysimeter within stand	*Phragmites communis* (reed)	$K_c = 0{\cdot}4–1{\cdot}25(ET/E_{\text{open water}})$	Gelboukh (1964)
Reedbed, Czechoslovakia	Bowen ratio	*Phragmites communis* (reed)	$K_c > 1{\cdot}0(ET/E_{\text{pan}})$	Vasa (1973) as cited by Smid (1975)
Reedbed and lakes, Poland	Lysimeters (phytometers) within reedbed and on land	*Phragmites australis* (reed), *Typha latifolia* (cattail), *T. angustifolia* (cattail) and *Schoenoplectus lacustris* (bulrush)	$K_c = 0{\cdot}92–1{\cdot}27$ (for the four species; $ET/ET_{\text{lysimeter w/o plants}}$	Bernatowicz *et al*. (1976)
Reedbed, UK	Lysimeter within reedbed	Reed (*Phragmites australis*)	K_c(reed) E_o 1·2–2·0	Gilman and Newson (1983) as cited by Crundwell (1986)
Rice field, Karnal, India (semi-arid climate)	Lysimeter located within row of crop	*Oryza sativa* (rice)	K_c (rice) stages: initial (1·15); crop development (1·23); reproductive (1·14); maturity (1·02)	Tyagi *et al*. (2000)
Louisiana, USA (rice field)	Measuring water level, precipitation, irrigation, seepage and tailwater runoff	Rice (species not specified)	K_c (rice) stages: vegetative (1·39); flowering (1·51); yield formation (1·43)	Shah and Edling (2000)
Southeast coastal India, 9 km west of the Bay of Bengal (dry tropical climate)	*ET* measured by micrometeorological methods and surface temperature measurements	Rice (species not specified)	K_c (rice) stage: whole season (1·25); tillering to heading (1·31); flowering (1·29); maturation (1·14)	Peterschmitt and Perrier (1991)
Murray–Darling Basin, south-east Australia	Lysimeter (in-field)	Rice (species not specified)	K_c (rice) stage: first 2 months of growing season (1·1), next 2 months (1·25), last month (1·0). K_c light to moderate wind conditions (1·14); strong wind conditions (1·21)	Bethune *et al*. (2001)
Northern Utah and southern Florida (freshwater marsh)	Bowen ratio	*Typha latifolia* (cattail)	K_c (cattail) stage: initial period (0·3); mid-season period (1·6); ending period (0·3). Logan, Utah (small stand)	Allen (1995)

(*continued overleaf*)

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Table I. (*Continued*)

Study site	Technique	Species	K_c	Researcher
			K_c (cattail) stage: initial period (0·3); mid-season period (1·15); ending period (0·3). Logan, Utah (large stand)	
			K_c (cattail) stage: initial period (0·6); mid-season period (1·0); ending period (0·6). Southern Florida (large stand)	
		Scirpus lacustris (bulrush)	K_c (bulrush) stage: initial period (0·3); mid-season period (1·8); ending period (0·3). Logan, Utah (small stand)	
Northern Utah (freshwater marsh)	Bowen ratio	*Typha latifolia* (cattail)	K_c (cattail): 9 ha wetland (1·15); 6 m (in width) wetland sizes reported by Allen *et al*. (1992) (1·60)	Allen *et al*. (1994)
The Netherlands (quaking fens)	Lysimeters	*Carex diandra* (sedge)	$K_{c1}(\text{sedge}) = 0{\cdot}81(ET/E_o)$, where E_o is the Penman potential free water evaporation); $K_{c2} = 1{\cdot}68(ET/E_{pan})$	Koerselman and Beltman (1988)
		Carex acutiformis/Sphagnum spp. (sedge/moss mixture)	$K_{c1} = 0{\cdot}74; K_{c2} = 1{\cdot}65$	
		Typha latifolia (cattail)	$K_{c1} = 0{\cdot}76; K_{c2} = 1{\cdot}87$	
Freshwater marsh, central Florida	Lysimeters	*Cladium jamaicense* (sawgrass)	$K_c(\text{May–October}) = 1{\cdot}09$	Mao *et al*. (2002)
			$K_c(\text{November–April}) = 0{\cdot}73$	
			$K_c(\text{annual}) = 0{\cdot}98$	
		Typha domingensis (cattail)	$K_c(\text{May–October}) = 0{\cdot}83$	
			$K_c(\text{November–April}) = 0{\cdot}64$	
			$K_c(\text{annual}) = 0{\cdot}77$ (monthly K_c values also available in paper)	

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Table II. Approximate costs[a] for systems (in 2002 US$) to measure wetland *ET* including Bowen ratio (BREB), eddy covariance (EC) with krypton hygrometer, closed-path infrared gas analyser (IRGA) and open-path IRGA, surface renewal (SR), LIDAR and combination equations (CE), including the Penman and Penman–Monteith equations

System	Main components of system including lowest cost data logger and mounting tower	Total cost
BREB	Interchangeable arms with psychrometers	$15 500 to $24 200
BREB	Fixed arms with dew point hygrometer	$8060
EC	Krypton hygrometer	$15 170 to $31 670
EC	Open path IRGA	$23 670 to $40 170
EC	Closed path IRGA[b]	$22 117 to $38 617
SR	R_n and G plus H from surface renewal	$4310
LIDAR	Laser, telescope, housing, machining, small parts and analysis facilities (not including equipment for micrometeorological parameters)	$1 000 000+
CE	R_n, G, T and RH for Penman type equation	$4530

[a] The same total cost for a net radiometer, two heat flux plates, one soil averaging temperature sensor, a mounting tower, logger enclosure, logger battery, logger solar collector and other miscellaneous items ($2789) was used for all systems except the BREB with interchangeable arms. The BREB system with interchangeable arms includes these items but at the manufacturer's prices. The prices for each system vary depending on components selected. The range in EC cost estimates depends mainly on the choice of sonic anemometer. Three commonly used three-dimensional sonic anemometers cost about $3500, $7700 and $20 000, so the low and high prices were used to determine the range of costs. Data retrieval items, software, etc., are not included in the system costs.
[b] Because of the power requirement for a pump to draw air into the closed path IRGA, it is difficult to operate a closed path IRGA with only battery power.

CONCLUSIONS

In reviewing the various methods to measure or estimate wetland *ET* for this paper, we found no universally accurate model or measurement technique. Instead we found that each measurement and estimation method has advantages and disadvantages based on cost (Table II), theoretical approach, underlying assumptions and calibration and data requirements (Table III). Furthermore, the characteristics of particular wetlands and wetland types (Table IV) may strongly influence the relative success of certain approaches. The reason that no one method proved to be best stems from the fact that wetlands vary greatly with respect to plant communities, hydrology and spatial characteristics, and the different approaches, which originally were developed for uniform agricultural fields, have varying degrees of success in dealing with these complexities. What did stand out in the review is that the best way to improve wetland *ET* estimates is to better account for surface variation by improving the measurement and relative weighting of net radiation (R_n) and conductive (ground or water) heat flux density (G). Both should be measured over each surface type (bare soil, vegetation and open water), and there also should be a unique sensible heat flux density (H) value for each surface, especially in arid environments. Another important result of the review is that, because individual methods have strengths and weaknesses, it seems prudent to use two or more measurement and estimation methods and compare the results.

In general, empirical methods are not universal and require site-specific calibration. For example, the PT equation with site-specific determined α' factors for different times of the year and weather conditions can provide good wetland *ET* estimates regardless of the wetland type. If wind speed (u) and humidity data are available in addition to R_n, G and T, then combination equations such as the PE and PM may further improve estimates. For uniform wetland surfaces, the PE equation (Equation 11) with the proper aerodynamic resistance can work well. However, good estimates of the surface and aerodynamic resistances, which generally are unknown, are needed. Wetland canopies that have a rough, non-uniform surface will have variable aerodynamic resistances over the surface. Therefore, using one aerodynamic resistance for the entire surface can lead to spurious results. Further, most of the empirical and combination equation methods assume that a surface is nearly wet and much of the available energy supply contributes to *ET*. Therefore, these methods may not be suitable for wetlands that have low *ET* rates during certain times of the year due to drought or to

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Table III. Advantages and disadvantages of the various wetland *ET* measurement and estimation methods

Method	Advantages	Disadvantages
Bowen ratio energy balance	The system is quite robust and the instrumentation is less costly than an eddy covariance system. If the wetland is of large extent, relatively smooth, and uniform, the BREB method can give good results	It is assumed that the transfer coefficients for sensible and latent heat are equal. This is generally true during neutral and unstable atmospheric conditions near the surface, but may not be true during stable (inversion) conditions. In the arid west, stable conditions are common during afternoon periods with high *ET*, so some error may occur during crucial periods because of invalid assumptions
Eddy covariance	A clear advantage is that if the water vapour covariance is measured, it provides a direct measure of λE. In addition, if R_n, G and H are measured at the same time, energy balance closure can be computed to provide some verification of the measurements. If all of the energy balance components are measured accurately and there is no horizontal advection, then $R_n - G = H + \lambda E$. None of the other methods have this self-verification	The greatest disadvantages of using eddy covariance are the cost, the complexity and the sensitivity of instruments to damage. Eddy covariance instrumentation generally requires high maintenance to ensure good results
Surface renewal	The advantages are the relatively low cost, portability and ease of maintenance. In addition, the SR method is based on short-term energy transfer between canopy elements and air parcels passing through the canopy, rather than flux gradient theory. Therefore, it is less dependent on fetch, and it may provide a useful method for determining *ET* along the edges of wetlands, in small wetland patches, along narrow wetland corridors and anywhere else with fetch limitations	Currently, the main disadvantage of the surface renewal method is that it must be calibrated against an independent measure of sensible heat flux density. As with all of the energy balance measurements and equations (e.g. BREB, Priestley–Taylor, Penman, Penman–Monteith), the overall accuracy of the method depends on accurate measurement of R_n and G
LIDAR	A great advantage of the LIDAR method is that it provides an averaged, horizontal cross-section of the water vapour content of air. In so doing, it provides a weighted measure of the latent heat flux from the variety of surfaces that may exist within variable terrain and/or plant communities	The main drawbacks of using LIDAR are the complexity and high cost. In addition, because the surfaces of most wetlands are variable, the Monin–Obukov similarity function may vary across the surface, whereas only one similarity function is typically used to calibrate the LIDAR measurements
Empirical equations	Empirical methods require easily measurable parameters for which data are typically available from local climate stations. The convenience and low expense makes this approach highly desirable for managers and researchers	By today's standards, most empirical equations are quite crude, incorporating a high amount of error, especially when applied outside the original climatic area for which they were developed. When *ET* estimates are required

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		for an entire watershed or region, however, empirical methods may be the only practical way to generate estimates for the area needed
Combination equations	A great advantage of using combination equations (i.e. PM and PE) is that the data requirements are minimal and collection at several heights is unnecessary. Combination equations accounts for wind speed, which is an improvement over the PT equation. The PM equation has an advantage over the PE equation in that it accounts for surface resistance	The PM equation assumes that the surface is uniform and 'nearly' wet with known canopy and aerodynamic resistances. In many wetlands, the assumption that the surface is nearly wet is reasonable; however, it may not be true for wetlands that dry up during part of the year. Because wetland canopies tend to be rough due to the presence of a mixture of vegetation, surface and aerodynamic resistances are difficult to determine and are likely to change with vegetation characteristics and weather conditions
Weighted canopy methods	The weighted canopy method has similar advantages and disadvantages as the combination equations. However, an additional advantage is that the method accounts for energy balance and aerodynamic and surface resistance differences over the soil, water and plant canopy surfaces	Although this is theoretically a better approach than treating a wetland like a large, uniform surface (e.g. using the PM equation), it does require aerodynamic and surface resistance estimates and available energy measurements over the canopy, soil and water surfaces. Shading and blockage of wind flow from the lower surfaces make it difficult to obtain representative estimates of available energy and aerodynamic resistance. Because measurements are replicated over several surfaces, the method is more expensive than combination equations and the CCC method
Crop canopy coefficient method	If the biological and environmental conditions are nearly the same at a particular wetland as where the K_c values for particular plant species or plant communities were developed, then using the K_c values to estimate wetland *ET* in the new location is justifiable. Because no on-site measurements are required, the CCC method is the most cost-effective	Wetlands often have variable, non-uniform surfaces and they may have differences in water quality and water temperature from one site to another. Therefore, the CCC method should be used only where accurate reference *ET* data are available and in wetlands with largely uniform stands of vegetation

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Table IV. Descriptions of major wetland types and characteristics most relevant to *ET* estimation and measurement

Wetland type (references used)	General description	Characteristics relevant to *ET*
Freshwater marshes, non-tidal (Weller, 1994; Mitsch and Gosselink, 2000)	A highly diverse group of wetlands having a broad range of geological origins. Specific types include vernal pools, playas, Great Lakes marshes and the extensive subtropical marshes of the Florida Everglades. Vegetation is dominated by grasses, tall reeds, broad-leaved monocots, and/or floating aquatic plants. Sizes range from <1 ha to over 100 000 km^2	Surface cover consists of bare soil, open water and vegetation in varying proportions. Vegetation of emergent plants may be quite tall (up to 2+ m) and form near monocultures (e.g. *Cladium* and *Typha* spp. marshes) or be of generally lower stature and contain high plant diversity, which results in a highly variable surface. Some freshwater marshes may be situated adjacent to large water bodies or may be surrounded by agricultural fields or arid plains
Freshwater marshes, tidal (Odum, 1988; Mitsch and Gosselink, 2000)	A distinct wetland type having strong similarities to both salt marshes and non-tidal freshwater marshes. Although subject to significant tidal fluctuation (up to *c.* 2 m), they exist in the upper reaches of estuaries that do not receive salt water. Mature tidal freshwater marshes have large peat reservoirs and occur in two forms: anchored to substrate or as floating islands. Vegetation consists of submerged aquatics along creek banks, broad-leaved monocots in the low marsh, and a diverse group of annuals and perennials in the high marsh. Sizes range from <1 ha to >1000 km^2	Surface cover may consist of bare soil, open water and vegetation or may be composed almost entirely of dense vegetation (floating islands). Emergent vegetation may reach 1–3 m in height and form near monocultures (e.g. *Typha* spp., *Zizania aquatica* and *Schoenoplectus acutus*) or be of generally smaller stature and much higher diversity. Tidal freshwater marshes are situated adjacent to channels or within channels in the case of floating islands
Freshwater swamps (Ewel, 1990, 1998)	A group of non-tidal, forested wetlands that contain standing water for much of the year and are dominated by mineral soil. They are generally found on coastal plains, embayments of large rivers and along lake edges. In the USA freshwater swamps are dominated by bald cypress–tupelo (*Taxodium distichum–Nyssa aquatica*) and pond cypress–black gum (*Taxodium distichum* var. *nutans*—*Nyssa sylvatica* var. *biflora*). Freshwater swamps typically have a complex understory containing subdominant trees, shrubs, herbs and aquatic vegetation. Sizes range from <1 ha to several thousand square kilometres	Surface cover consists mainly of vegetation and standing water. Depending on the hydroperiod, source waters and nutrient regime, freshwater swamps may be highly productive containing several canopies of lush growth or contain relatively low plant diversity. Tree heights may reach up to 30–40 m (*Taxodium distichum*). The inflows to and outflows from freshwater swamps may include sheet flow, alluvial streams and rivers, and groundwater flow
Mangroves (Tomlinson, 1986; Duke, 1992; Woodroffe 1992)	A tropical, intertidal forested wetland, which forms along protected shorelines, along estuaries and in other areas containing abundant fine-grained sediment. Mangroves are halophytes that typically have low plant diversity, little understory and often have dense looping (e.g., *Rhizophora* spp.) or straight aerial roots (e.g., *Avicennia* spp., *Sonneratia* spp. and *Laguncularia* spp.). Sizes of	Depending on geomorphology and forest maturity, surface cover may consist of an extensive, tall canopy (up to 20+ m) with only small patches of bare ground visible or the canopy may be of much smaller stature (dwarf mangroves <2 m) and have a patchy distribution in which large areas of bare ground/open water are visible. The relative cover of bare ground versus open water

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	mangrove ecosystems range from small lagoons or coves containing only a few hectares to areas covering thousands of hectares or greater	varies according to the tides and precipitation. Mangroves growing along estuaries in arid areas may have a narrow distribution closely following channel borders
Peatlands (Slack *et al.*, 1980; Vitt *et al.*, 1994; Crum, 2000)	Peat-storing wetlands that occur in temperate and cool boreal regions of the world, often in former glacial terrain. They can be classified as bogs or fens: bogs receive most or all their hydrological inflows from precipitation and fens receive inflows from a variety of sources including some groundwater. Bogs and most fens contain substantial moss cover (e.g. *Sphagnum* spp.). Other vegetation in peatlands includes shrubs, sedges, grasses, reeds and various tree species. Sizes of individual peatlands range from <1 ha to >1 million ha	The most distinctive feature with respect to *ET* is the presence of mosses, which are known to insulate the wetland surface as well as provide access to substrate moisture via wicking action. Surface cover of peatlands includes open water, bare soil and vegetation. Vegetation may be highly diverse, consisting of several canopies including mosses, grasses and sedges, and trees. In large expanses of peatlands, peat ridges (strings) may alternate with water-filled depressions (flarks), creating a topographically diverse landscape. In other areas, peatland vegetation may be more uniform consisting of large expanses of sedges, grasses and ericaceous shrubs
Riparian wetlands (Niering, 1989; Mitsch and Gosselink, 2000)	A flood-prone wetland type occurring in the ecotone between stream or river channels and upland areas. In mesic areas such as the eastern USA, they can be found within extensive, low-lying floodplains. In arid areas such as southwestern USA, riparian wetlands occur as narrow bands along ephemeral streams. Riparian wetlands are actually a subset of freshwater swamps. Dominant species include bald cypress (*Taxodium distichum*), cottonwood (*Populus* spp.). alder (*Alnus* spp.) and willow (*Salix* spp.). Riparian wetlands may measure only several metres across or range up to tens of kilometres wide. Their length is dictated by the length of the river or by changing environmental conditions along the river	Surface cover alternates between consisting solely of open water and vegetation to consisting of bare soil or sand, open water and vegetation. Riparian wetlands may be highly productive, containing multiple canopies, or somewhat sparse depending on climate and nutrient regimes. The height of the dominant canopy may reach 30–40 m or be <2 m. The most significant aspect of riparian wetlands with respect to *ET* is that they are long, linear ecosystems directly adjacent to stream channels
Salt marshes (Hackney and de la Cruz, 1982; Bertness and Pennings, 2000)	Highly productive, coastal wetlands occurring in middle and high latitudes in areas where accumulation of sediments can keep pace with land subsidence and where there is adequate protection from high-energy waves and storms. Salt marshes are dominated by salt-tolerant grasses and rushes (e.g., *Spartina, Suaeda* and *Salicornia* spp.) and contain clearly defined vegetation zones in response to frequency of tidal flooding, salinity and other environmental characteristics. Sizes range from narrow fringes <1 ha up to extensive areas many kilometres wide	The relative cover of bare ground versus open water varies according to tides and precipitation. The extent of tidal flooding depends on the elevation of particular vegetation zones, the tidal regime and the hydrodynamics of a particular site. Vegetation zones within salt marshes may contain extensive monotypic stands, especially in the low marsh, which typically contains cordgrasses (*Spartina* spp.) and/or pickleweed (*Salicornia* spp.). The middle marsh and high marsh may be more diverse containing reeds, bulrushes and grasses, however, local conditions often dictate diversity. Plant heights vary depending on particular species. Low marsh vegetation is typically less than 1.0 m, middle marsh vegetation can reach up to 1.5 m+, and high marsh vegetation is variable due to the presence of herbaceous and shrub species

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a substantial litter layer or moss cover that provides a barrier to energy transfer from the wet surface below (e.g. desert playas, prairie potholes, freshwater marshes, peatlands; Table IV).

Perhaps the easiest method of all for estimating wetland *ET* is the canopy cover coefficient method. However, there are several important caveats to using this approach (Table III). Even if all caveats are accounted for, the greatest barrier to using the CCC method is that few K_c values are available for particular wetland plant species and plant communities (Table I). In addition, non-standard calculation of E_o has resulted in various K_c values for the same plant species. Therefore, this approach may potentially develop into a highly suitable method for certain wetland types if the range of K_c values are expanded and if E_o calculation is standardized.

The two most common *ET* measurement techniques, the BREB method and the EC method, have important advantages and disadvantages (Table III). In theory, the BREB method should be used only over a uniform surface with adequate fetch to insure that the data represent fluxes from the surface. The measurement arms should both be within the fully adjusted boundary layer (i.e. within the logarithmic wind speed profile above the canopy) and should be sufficiently far apart to observe differences in temperature and humidity. When the wetland surface is non-uniform (e.g. in forested wetlands or wetlands with high plant diversity) the use of the Bowen ratio is questionable because the β values are likely to be in error and using the BREB method might be less accurate than $R_n - G$ alone. A clear advantage of the EC method in relation to the BREB method is that it is the only technique for which a self-test of accuracy is possible. However, measurement of other energy balance components besides λE, which permits a self-test, is not necessary. This may be advantageous if there is energy advection in the water. Such a situation may occur in several wetland types such as riparian wetlands, tidal freshwater marshes, salt marshes and mangroves (see Table IV for wetland descriptions). Incidentally, few researchers have acknowledged this problem and consequently there is little discussion on how to compensate for advection effects (see Appendix for one method to account for energy advection in water).

To date, much of the literature on the use of the EC method to measure λE in wetlands was conducted using one-dimensional sonic anemometers (to measure vertical wind speed fluctuations) and krypton hygrometers (i.e. to measure water vapour). Recently, however, three-dimensional sonic anemometers and closed path infrared gas analysers have been used in wetlands (e.g. Vourlitis and Oechel, 1997; Soegaard *et al.*, 2001). Three-dimensional sonic anemometers and open path infrared gas analysers have become popular instruments for measurements over forests and crops, and it is likely that their use will increase in wetlands, thus improving results. A drawback is that three-dimensional sonic anemometers and infrared gas analysers are expensive and require careful maintenance and calibration.

The surface renewal method is a relatively new measurement approach that has several advantages (Table III). Most importantly, it offers a low-cost approach for obtaining multiple estimates of *ET* over a variable non-tidal wetland surface (e.g. forested wetlands, freshwater marshes, peatlands). The SR method also can be used in tidal and riparian wetlands if energy advection in the water and the *G* term are properly measured. However, a current drawback for the surface renewal method is that the measurements must be calibrated against a sonic anemometer to account for non-uniform heating under the temperature sensors. Most likely a combination of eddy covariance and surface renewal will provide the coverage and level of accuracy needed to best measure wetland *ET* over uniform and non-uniform surfaces.

LIDAR provides an integrated measure of *ET* over a large surface area, so it may provide an even better estimate of wetland *ET* than using micrometeorological methods. In addition, with the LIDAR method it is unnecessary to account for energy advection in the water and errors in the *G* or R_n measurements. However, to our knowledge, this method has been used only in regions containing riparian wetlands. Therefore, currently little is known about the application of the LIDAR method in different wetland types. LIDAR may have similar problems to other methods in regard to the oasis and clothes-line effects over small or oddly shaped wetlands.

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APPENDIX

The basis of modern microclimatology is the energy balance or energy budget, which considers all incoming, outgoing and stored energy on a surface or within a volume to determine energy distribution. This is a valid approach if most of the energy fluxes are vertical and there is little or no horizontal advection within or above a vegetative canopy. If these criteria are met and the energy balance is measured properly, it provides a method to estimate latent heat flux density, which is converted to water vapour flux density by dividing by the latent heat of vapourization. Water vapour flux density is used to determine the mass of water loss per unit surface area, which is converted to the depth of water loss.

Most objects on Earth's surface are exposed to solar (short-wave) radiation, which is reflected from, absorbed by or transmitted through the surface elements. In addition, a surface receives a certain amount of long wavelength (thermal) radiation from the sky that similarly is reflected, absorbed or transmitted. Because any object with temperature above absolute zero radiates energy, natural terrestrial surfaces also emit thermal radiation. Incoming radiation is given a positive sign because it adds energy to the surface and outgoing radiation is given a negative sign because it subtracts energy from the surface. When the positive incoming radiation and negative outgoing radiation are summed, the result is the net absorbed radiation (i.e. 'net radiation', with a symbol, R_n). Net radiation is the major heating and cooling force on Earth.

Ideal surfaces have no thickness, just as theoretical points have no volume. This means a surface cannot store any of the heat it receives from net radiation; all of the radiant energy must be partitioned into other forms of energy transfer. The other forms of energy transfer include: (i) conduction of energy below the surface (G), (ii) energy used for evaporation or gained from condensation (λE), where λ is the latent heat of vapourization (kJ kg^{-1}) and E is the flux density of water vapour (kg m^{-2} s^{-1}), (iii) sensible heat flux density or energy used to heat the air or energy gained from cooling the air (H), (iv) energy associated with biological processes such as photosynthesis and respiration (M), and (v) energy storage in plant tissues (S). The formal energy balance equation is

$$R_n = G + H + \lambda E + M + S \tag{A1}$$

For plant canopy surfaces, M is small compared with the other terms, so generally it is neglected for energy balance calculations. Depending on the vegetation, S may or may not be important. For example, Tanner (1963) reported that canopy heat storage was negligible during the day and was about 6% of G at night in a 10-t (wet weight) alfalfa-brome crop in Wisconsin. In the scant research on biomass storage in wetlands, Smid (1975) reported that biomass heat storage was not important for a *Phragmites* (reed)-dominated stand, whereas Bidlake *et al*. (1996) demonstrated the importance of including S in ET measurements over freshwater

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swamps. A key difference between these wetlands, and the one that is of utmost relevance to S, is the canopy volume of vegetation. Upland forests, which have tall, large canopies, have been shown to have substantial short-term variability in S (Aston, 1985), and therefore, forested wetlands (swamps) are probably no exception. For this review, we will assume that M and S are generally small compared with other components, however, clearly more research is needed concerning the significance of S in forested wetlands such as bottomland hardwood swamps cypress swamps and mangroves.

If net radiation, sensible heat flux density and ground (and water) heat flux density are measured accurately, then the latent heat flux density is estimated as

$$\lambda E = R_n - G - H \tag{A2}$$

As described above, the difficulty in estimating latent heat flux density from wetlands lies in how to accurately measure these fluxes from a highly variable surface.

Net radiation

Net radiation on a surface is defined as the net sum of radiation striking and leaving a surface, where radiation is positive towards and negative away from the surface. It physically represents the amount of radiant energy that is available to be partitioned into the other forms of surface energy (i.e. sensible, latent, and conductive heat)

$$R_n = (R + R_d)(1 - a_o) + R_{Ld} + R_{Lu} \qquad \text{W m}^{-2} \tag{A3}$$

where R_n is net radiation, R is direct solar radiation, R_d is diffuse (indirect) solar radiation that was scattered or reflected by the sky and clouds, a_o is the albedo (i.e. reflection of solar radiation) from the surface, R_{Ld} is the incoming thermal radiation and R_{Lu} is the emitted and reflected outgoing thermal radiation. The right-hand term (R_{Lu}) is estimated as a function of the surface temperature

$$R_{Lu} = -\varepsilon_o \sigma T_o^4 \qquad \text{W m}^{-2} \tag{A4}$$

where T_o is the surface temperature (Kelvin), ε_o (the fraction of potential radiation emission from a surface) is close to 1·0 for many plant canopies and $\sigma = 5{\cdot}6697 \times 10^{-8}$ W m^{-2} K^{-4} is the Stefan–Boltzmann constant. The albedo of solar radiation for vegetated surfaces typically is between $a_o = 0{\cdot}1$ and $a_o = 0{\cdot}3$. In one of the few studies on albedo in wetlands, Lafleur *et al.* (1987) found $a_o = 0{\cdot}2$ in a subarctic marsh. Because of the non-uniform surface, both R_{Lu} and a_o are likely to vary considerably within a wetland.

Although it is possible to measure the solar and thermal components of net radiation, instrumentation is expensive and often unavailable. Typically, a cheaper but less accurate net radiometer is used. Brotzge and Duchon (2000) compared inexpensive net radiation instruments for long-term field measurements and found that plastic-dome-type and domeless net radiometers each have advantages. In both cases, calibration against a more expensive four-component radiometer was suggested before long-term use in the field. The authors noted that the domeless net radiometer is more influenced by wind speed than the dome-type net radiometers, and a correction for wind speed was suggested. The domeless net radiometer experienced large errors during precipitation events and it may not be suitable for use in high precipitation areas. The plastic-dome-type net radiometer was less affected. The cosine response error of the domeless sensor may be problematic at low solar elevation angles, so use at high latitudes could lead to errors.

All net radiometers should be mounted sufficiently high to obtain a clear view of the underlying surface being measured with minimal influence from the mounting tower, other objects, or surrounding canopy surfaces that might affect albedo or long-wave radiation from the surface being measured. Proper levelling is required to ensure accuracy.

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Sensible and latent heat flux

Several of the methods commonly used to estimate ET (e.g. Penman, Penman–Monteith and Bowen ratio) are based on flux gradient theory, where the flux densities of sensible and latent heat are estimated as functions of the gradient of the parameter of interest. For example, an expression for H is written as

$$H = -\rho C_{\mathrm{p}} \kappa_{\mathrm{h}} \frac{\partial T}{\partial z} = -\rho C_{\mathrm{p}} \left(\frac{T_2 - T_1}{r_{\mathrm{a}}} \right) = -g_{\mathrm{a}} \rho C_{\mathrm{p}} (T_2 - T_1) \qquad \mathrm{W\ m^{-2}} \tag{A5}$$

where ρ is the air density (kg m^{-3}), C_{p} is the specific heat of air at constant pressure (J kg^{-1} K^{-1}), κ_{h} is the thermal diffusivity (m^2 s^{-1}), $\partial T/\partial z$ is the gradient of temperature (°C m^{-1}) and $T_2 - T_1$ is the temperature difference (K) between upper and lower measurements within the fully adjusted boundary layer, r_{a} is the aerodynamic resistance (s m^{-1}) to sensible heat transfer (i.e. $r_{\mathrm{a}} = \partial z/\kappa_h$), and conductance $g_{\mathrm{a}} = 1/r_{\mathrm{a}}$ is the rate in m s^{-1} that the sensible heat concentration (J m^{-3}) moves upwards or downwards. The negative sign is included to make H positive away from the surface.

A similar flux gradient expression is used to estimate water vapour flux density (E) in kg m^{-2} s^{-1}, and multiplying E by $\lambda \approx 2{\cdot}45$ J kg^{-1} converts from mass to energy flux density (W m^{-2})

$$\lambda E = -\frac{\rho C_{\mathrm{P}}}{\gamma} \kappa_{\mathrm{v}} \frac{\partial e}{\partial z} = -\frac{\rho C_{\mathrm{P}}}{\gamma} \left(\frac{e_2 - e_1}{r_{\mathrm{a}}} \right) \qquad \mathrm{W\ m^{-2}} \tag{A6}$$

where $\gamma = \dfrac{PC_{\mathrm{p}}}{\lambda \varepsilon}$ (kPa °C^{-1}) is the psychrometric constant (kPa K^{-1}). In the H and λE equations, although the aerodynamic resistance to sensible and latent heat flux might be different, the same value (r_{a}) is typically assumed. The variable γ depends on the barometric pressure (P), and it can be estimated using the equation

$$\gamma = 0{\cdot}00163 \frac{P}{\lambda} \qquad \mathrm{kPa\ ^{\circ}C^{-1}} \tag{A7}$$

where P (kPa) is estimated using a function from Burman *et al.* (1987):

$$P = 101{\cdot}3 \left(\frac{293 - 0{\cdot}0065 E_{\mathrm{L}}}{293} \right)^{5{\cdot}26} \qquad \mathrm{kPa} \tag{A8}$$

where E_{L} is the elevation in metres above mean sea-level. The variable $\varepsilon = 0{\cdot}622$ is the ratio of molecular weights of water vapour and dry air, κ_{v} is the water vapour diffusivity (m^2 s^{-1}), $\partial e/\partial z$ is the gradient of vapour pressure (kPa m^{-1}), $e_2 - e_1$ is the vapour pressure difference (kPa) between upper and lower measurements within the fully adjusted boundary layer, and $r_{\mathrm{a}} + r_{\mathrm{s}}$ is the sum of aerodynamic and surface resistances (s m^{-1}) to latent heat transfer (i.e. $r_{\mathrm{a}} + r_{\mathrm{s}} = \partial z/\kappa_{\mathrm{v}}$). The negative sign is included to make λE positive away from the surface.

The flux equation for λE is used to estimate the flux density between two levels above a canopy or surface depending on the aerodynamic resistance (r_{a}) between the two levels and it is used to derive the Bowen ratio and Penman equations. However, for the Penman–Monteith equation, the water vapour flux is from the canopy or surface to a level above the canopy rather than between two heights above the surface. Because the canopy or surface also can impart a resistance to water vapour flux, a modified equation for latent heat flux from a surface is given by

$$\lambda E = -\frac{\rho C_{\mathrm{P}}}{\gamma} \left(\frac{e_2 - e_{\mathrm{o}}}{r_{\mathrm{a}} + r_{\mathrm{s}}} \right) \tag{A9}$$

where e_{o} is the apparent vapour pressure (kPa) at the canopy surface and $r_{\mathrm{a}} + r_{\mathrm{s}}$ is the sum of the aerodynamic and surface resistances (s m^{-1}). The surface resistance (r_{s}) represents the resistance of canopy, soil and water surfaces to water vapour transfer from the surface elements to a level in the canopy. The aerodynamic

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resistance (r_a) is between that level and the air above the canopy. If the numerator and denominator are divided by r_a and simplified, then λE is

$$\lambda E = -\frac{\rho C_P}{\gamma}\left[\frac{(e_2 - e_o)/r_a}{(r_a + r_s)/r_a}\right] = -\frac{\rho C_P}{\gamma^*}\left(\frac{e_2 - e_o}{r_a}\right) \qquad \text{W m}^{-2} \tag{A10}$$

where γ^* is a modified psychrometric constant that accounts for the separation of resistance to water vapour transfer into the sum of surface and aerodynamic resistances

$$\gamma^* = \left(1 + \frac{r_s}{r_a}\right)\gamma \tag{A11}$$

The equations for H and λE from a surface (Equations A5, A6 and A10) are used to derive the Penman and Penman–Monteith equations, and the main difference between them is that, in the Penman equation, $r_s = 0$ and therefore $\gamma^* = \gamma$.

Ground (water) heat storage

Soil heat flux determines the ground temperature profile in time and space. When energy reaches the ground, the surface warms, and, if warmer than the ground below, heat conducts downwards from the surface into the soil. By convention, radiation that adds energy to the surface has a positive sign and radiation away from the surface has a negative sign. Therefore, following the convention used in Equation (A1), conduction of energy downward into the soil or water is given a positive sign and conduction upwards towards the surface is given a negative sign.

The soil surface typically heats during the morning and there is positive conduction into the soil. As the net radiation decreases in the afternoon, the surface will cool relative to the soil below and heat is then conducted upwards towards the surface (i.e. negative flux). This negative heat flux continues during the night as soil heat conducts upwards to replace lost energy at the cooler surface.

The ground (conductance) heat flux term G is important for some surfaces and time-scales, but it is negligible in others. For example, on an hourly basis, G can change considerably. When averaged over a diurnal cycle, the positive fluxes are nearly equal to the negative fluxes, so G over a 24-h period typically is close to zero (i.e. providing a good check for measurements). This may not be true during spring or autumn, when large changes in air masses can cause appreciable changes in heat storage or loss within a single day.

There are many papers and books with discussions about the theory and measurement of soil heat flux (e.g. de Vries, 1963; van Wijk and de Vries, 1963; Fuchs and Tanner, 1968; Idso, 1972; Brutsaert, 1982; Jensen *et al.*, 1990; Monteith and Unsworth, 1990). Although accurate measurement of soil and heat flux can be challenging in any environment, measuring G in wetlands is especially difficult because of variations in water depth and non-uniformity of the surface. Several methods to measure soil and water heat flux density are discussed in Brutsaert (1982).

Soil flux heat density (G) is expressed as

$$G = -K_s\frac{\partial T}{\partial z} \qquad \text{W m}^{-2} \tag{A12}$$

where K_s is the thermal conductivity (W m^{-1} K^{-1}) and the second term on the right-hand side is the change in temperature with depth (K m^{-1}), called the thermal gradient. Various methods utilize this equation to estimate G using flux plates, temperature profiles and changes in heat storage (Brutsaert, 1982). Typically, a heat flux plate, which is a thin plate of material with a known thermal conductivity, is inserted horizontally in the soil and the temperature difference between the top and bottom of the plate is used to calculate the heat flux through the plate. Assuming that the plate has a thermal conductivity close to that of the soil and it is not placed too close to the surface, it provides a measure of the heat flux at a point in the soil. When placed

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too close to the surface, a heat flux plate will give inaccurate results owing to heterogeneity, condensation or evaporation from the plate surfaces, and soil cracking, which breaks contact with the soil and permits transfer of sunlight or water to the plate. Brutsaert (1982) recommends that heat flux plates should be buried 0·05 to 0·10 m to avoid these problems.

Generally, heat flux plates with thermal conductivities similar to organic soils are not commercially available. Therefore, the plates need to be constructed. However, construction and calibration of heat flux plates can be difficult (Fuchs and Tanner, 1968; Idso, 1972). Monteith and Unsworth (1990) suggested thermal conductivities of 0·06, 0·29 and 0·50 W m^{-1} K^{-1} for peat soils (80% pore space) with volumetric water contents of 0·0, 0·4 and 0·8, respectively. Sandy and clay soils have thermal conductivities roughly three to five times greater than peat soils.

For energy balance calculations, the energy conducted downward from the surface or upward to the surface is needed. In the case of wetlands, the surface might be soil or water. It is not possible to place a heat flux sensor directly on the soil or water surface to measure the energy fluxes. For soils, the heat flux measurements are taken deeper in the soil and changes in heat storage of the overlying soil are used to estimate the surface value for G. For an inundated surface, energy fluxes into the water are determined by measuring the change in water temperature, and energy fluxes into the soil below the water layer are measured with heat flux plates below the soil surface.

Based on the combination method of C.B. Tanner (Brutsaert, 1982), soil heat flux density (G) at the surface (i.e. depth $z_1 = 0$) can be estimated as

$$G = G_2 + C_V \left(\frac{T_f - T_i}{t_f - t_i} \right) \Delta z \tag{A13}$$

where G_2 is the flux density measured at depth z_2, C_V is the volumetric heat capacity of the soil (J m^{-3} K^{-1}), T_f and T_i are the mean temperatures (K or °C) of the soil layer above the flux plate at times t_f and t_i, which are the final and initial time (s) of the sample period (e.g. for a 30-min sampling period, $t_f - t_i = 1800$ s). The depth of the soil from the surface to the flux plate is Δz (m). Based on de Vries (1963), the formula to estimate C_V (J m^{-3} K^{-1}) is

$$C_V = (1{\cdot}93V_m + 2{\cdot}51V_o + 4{\cdot}19\theta) \times 10^6 \tag{A14}$$

where V_m, V_o and θ are the volume fractions of mineral components, organic matter and water, respectively (Jensen *et al.*, 1990).

For areas of a wetland containing vegetated surfaces, measurements are taken at several locations with differing exposure to sunlight. A mean G value that is weighted proportionally for similar exposure areas is computed. Obtaining an accurate value for G through weighting different areas of a site is one of the most important and, perhaps, most difficult tasks when trying to characterize a wetland.

For shallow water-covered surfaces with minimal flow, G at the surface can be estimated as

$$G = G_2 + C_V \left(\frac{T_{sf} - T_{si}}{t_f - t_i} \right) \Delta z_s + 4{\cdot}19 \times 10^6 \left(\frac{T_{wf} - T_{wi}}{t_f - t_i} \right) \Delta z_w \tag{A15}$$

where T_{wf} and T_{wi} are the final and initial water temperatures over the water depth (Δz_w) and T_{sf} and T_{si} are the final and initial mean soil layer temperatures for the soil depth (Δz_s) between the bottom of the water layer and the heat flux plate. The factor $4{\cdot}19 \times 10^6$ (J m^{-3} K^{-1}) is the volumetric heat capacity of liquid water. The volumetric heat capacity at saturation is used to determine C_V for the soil layer between the bottom of the water and the heat flux plate.

If water flow into the wetland system is appreciable and the temperature of incoming water is different from that in the wetland, then advection of heat in the water may be significant. Such a scenario would require the calculation of heat transfer resulting from horizontal advection. To account for energy advection through horizontal water movement, one must measure the mean temperature over the depth (d_2) of water at some point

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upstream and over the depth (d_1) at the station site and the water flow rates (F_{R2} and F_{R1}) past the station at the same two locations as the depth measurements. The temperature difference divided by the distance between measurements gives the rate of temperature change with distance upstream from the station site (°C m^{-1}). The rate of temperature change with time (°C s^{-1}) at the station site is determined by multiplying by the flow rate (m s^{-1}). Then the energy flux density contributed by advection (A_w) is calculated as

$$A_w = (4{\cdot}19 \times 10^6)\frac{(F_{R2}T_2d_2 - F_{R1}T_1d_1)}{\Delta x} \qquad \text{W m}^{-2} \tag{A16}$$

where T_1 is the mean water temperature over depth d_1 (m) at the station site and T_2 is the mean water temperature over depth d_2 measured at distance Δx m upstream. Then the new energy balance equation is

$$R_n + A_w = G + H + \lambda E \qquad \text{W m}^{-2} \tag{A17}$$

where G, which includes the soil heat flux and energy storage in the water, is determined using Equation (A15) and A_w is determined from Equation (A16). In all of the energy balance methods and combination equations, $R_n + A_w$ should be substituted for R_n to account for advection. Brutsaert (1982) contains further discussion and references on accounting for heat advection resulting from water flow, although his treatment of the subject is limited to lakes and small ponds.

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