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Chapter

Field-Scale Estimation of Evapotranspiration

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Abstract

Evapotranspiration (ET) is a major component of the water cycle, which makes it an integral part of water resources management, especially in arid and semiarid environments. ET data are used for water management, irrigation scheduling, various modeling activities, and much more. Some areas of scarce water resources place limitations on water use, which are typically determined from various modeling approaches. As many models use ET as an input, or for validation, accurate ET data is essential to ensure accurate model outputs. In addition, most water management practices are done at the field scale; ET data of a similar scale is needed. Many ET measurement or estimation methods exist and vary widely in approach, instrumentation, complexity, and purpose. A lysimeter is considered the standard for ET measurement and is the most accurate. Other, more portable options are available, such as eddy covariance, scintillometer, Bowen ratio, and remote sensing, all capable of estimating actual field ET within approximately 30% of actual values. Although other methods may not be as accurate as a lysimeter, each has benefits in certain situations. Depending on the purpose, the level of accuracy may be suitable. ET estimation methods are constantly evolving, and accuracy should continually improve further.

Keywords: irrigation scheduling, energy balance, water balance, lysimeter, eddy covariance, scintillometer, remote sensing

1. Introduction

Fresh water is an essential resource that is becoming increasingly limited. In some arid and semiarid regions, groundwater resources are being exhausted with little to no surface water available as an alternate source. Proper water resources management is essential for these areas. In many cases, water management strategies rely on the use of evapotranspiration (ET) to account for some of the water losses. ET is a combined term that represents water lost through evaporation from the soil or plant surface, as well as water lost through transpiration from the plant. In many regions, such as the Texas High Plains, ET is the largest water loss component in the hydrologic budget. This fact makes accurate ET estimates vital for accurately and properly managing crop water. In the Texas High Plains, and the rest of the southern Ogallala Aquifer region, groundwater recharge is very low at ~11 mm yr⁻¹ [1]. With such little recharge, the Ogallala Aquifer is deemed a finite resource. In order to preserve this natural resource for future generations, conserving the remaining water is paramount. The Texas High Plains lies in the Southern Great Plains near the southern end of the Ogallala Aquifer (see **Figure 1**). Agriculture is the predominant land use and irrigated land accounts for the majority of the agricultural production in this region. In the state of Texas, irrigation accounts for 60% of total water use; however, in the Texas High Plains, irrigation accounts for 89% of the total water use [2]. The Texas High Plains is a major corn-, cotton-, wheat-, and sorghum-producing region with much of the agricultural production under irrigation. The vast majority of irrigation water is withdrawn from the Ogallala Aquifer. With limited and sporadic rainfall, the Ogallala Aquifer receives little to no recharge in this region and is essentially being mined; therefore, conservation is an integral part of the regional water plan [3]. The northern and southern parts of the Texas High Plains are similar in size; however, the northern Texas High Plains irrigates over 760,000 ha [4]. In both the northern and southern regions, irrigated crop yields are at least double that of dryland yields (on average).

In the northern Texas High Plains (see **Figure 1**), about 55% of the cropland is irrigated and uses about 1.76 billion m³ (1.43 million ac-ft) of water annually for irrigation [3]. Irrigated winter wheat, grain corn, cotton, and grain sorghum are the predominant crops, comprising 30, 26, 23, and 10% of the total irrigated area, respectively [4]. Corn is a relatively large water use crop, requiring an annual average of over 480 mm (19 in.) of irrigation [3], and all of the corn area in this region requires irrigation. Currently, silage and forage crops are minor crops in the region but are increasing dramatically to meet the demands of new dairy operations that continue to expand into the area.



Figure 1. Ogallala Aquifer and Texas High Plains regions.

In the southern Texas High Plains (see **Figure 1**), cotton is the major crop comprising 65% of the total irrigated area [4]. The popularity of cotton in this area is a reflection of the water resource limitations where the saturated thickness of the Ogallala Aquifer decreases near the southern boundary. Cotton only requires an annual average of 170 mm (6.7 in.) of irrigation in the Texas High Plains [3]. Peanuts are the second most grown crop in the southern region with about 9% of total irrigated area. Grain corn only accounts for 3% of irrigated area with winter wheat and grain sorghum at 7% each [4].

The decline in the saturated thickness of the Ogallala Aquifer has caused some local groundwater conservation districts to begin regulating annual water withdrawals. In Texas, groundwater conservation districts have been granted the authority to regulate water withdrawals to extend the life of the Ogallala Aquifer and meet the goals of regional water plans approved by the state. As part of the Texas State Water Plan, the Panhandle Water Regional Planning Group set the goal of nominally, on average, retaining 50% of current available water in 50 years [5].

Currently, regional irrigation demand is determined by advanced models such as MODFLOW [6] and the Texas A&M-Amarillo [3] model. MODFLOW is a complex model that assesses groundwater resources, which requires ET as an input. In 1999, the Texas A&M-Amarillo (TAMA) model was developed as a new estimation methodology for the region [5]. It was used to accurately estimate irrigation demand in the northern Texas High Plains. The TAMA model estimates the seasonal irrigation demand per crop per county for 21 counties in the northern region of the Texas High Plains. The TAMA model requires inputs of ET, precipitation, and soil characteristics. Accurate ET data and local acreage knowledge beyond USDA-Farm Service Agency values are essential for model accuracy.

Since modeling is one of the main ways regional water plans are developed and assessed, accurate model outputs are highly desired. Many of the models use ET as an input, and the outputs are heavily affected by the accuracy of the inputs. High levels of accuracy are beneficial in regional water planning so that the best decisions are made regarding water allotment and water availability. This creates the need for high levels of accuracy in ET estimation.

1.1 Evapotranspiration

Measuring or estimating ET can be difficult but numerous instruments and methods do exist. A common (and relatively simple) method of estimating ET is using reference ET (ET_{ref} [7]) which uses meteorological data to estimate the water demand of a reference crop, usually a short, clipped grass or alfalfa. To get ET for a specific crop from the reference ET, a crop coefficient (K_c) can be applied to yield potential crop ET or ET_c [7]. When measured ET_c data are available, the K_c values can be obtained by dividing ET_c by ET_{ref} . This approach requires accurate data for ET_c to obtain the best results. K_c values for a wide variety of crops are available throughout the literature [7–9].

Single and dual crop coefficient methods are available. For the single crop coefficient approach, water loss through transpiration is combined with soil evaporation, and a single K_c value is used. In the dual crop coefficient approach, the transpiration and evaporation components are split into a basal crop coefficient (K_{cb}) for transpiration and a soil evaporation (K_e) component [7]. The ET_c from the K_c approach provides the amount of water that would be used by the crop if there is no water limitation. In most cases, ET can be lower than the potential rate due to stresses from water, nutrients, pests, etc. A stress coefficient (K_s) can be applied to the K_c to account for water stress when using ET_{ref} [7]. To account for reduced ET due to stresses, the term actual ET (ET_a) is used. ET_a corresponds to the actual

amount of water lost from a specific field, with a specific crop, under specific environmental conditions. K_s is calculated by

$$K_s = \frac{TAW - D_r}{(1 - p)TAW} \tag{1}$$

(2)

where TAW is the total available soil water, D_r is the root zone depletion (mm), p is the fraction of TAW allowed before the crop experiences water stress (typically 0.50 for most crops [7]), and D_r is typically calculated using a water balance approach. TAW is calculated as

$$TAW = 1000(\theta_{FC} - \theta_{WP})Z_{r}$$

where θ_{FC} is the field capacity water content, θ_{WP} is the wilting point water content, and Z_r is the rooting depth (m). Additional information on calculating K_s is available in [7].

Since ET_{ref} assumes no water limitations, it represents the atmospheric demand for water, which is why most ET_{ref} equations only require weather data. ET_c derived from ET_{ref} provides the potential, or maximum, water use by the crop, assuming no crop water stress, or if the K_s is used, only accounts for water stress. One potential issue with this technique is that crops can typically encounter stress from multiple sources throughout the season, especially in arid and semiarid climates. Another issue is the use of limited crop coefficients during the growth cycle. It is more advantageous to use ET_a in water planning and irrigation scheduling, but acquiring ET_a can be challenging.

To maximize the effectiveness of irrigation scheduling, ET_a is more beneficial than ET_{ref} or even ET_c . Confusion exists regarding what each ET term corresponds to, which can lead to the use of ET_{ref} instead of ET_c . Using ET_{ref} in irrigation scheduling is considered better than no irrigation scheduling, but it can lead to over application of water, as can ET_c . Even though ET_c corresponds to the ET of the specific crop, it does not take ET reduction due to stress into account. One problem with using ET_a in irrigation scheduling is that ET_a can be very difficult to obtain. Where ET_{ref} can be calculated from weather parameters, using a relatively simple weather station on a reference surface, ET_a requires more advanced (and expensive) instrumentation. Current technologies for determining ET_a are described in detail below.

1.2 Water balance

ET is used in production agriculture in the practice of irrigation scheduling. This practice involves tracking ET from the field and applying the water balance. The water balance is based on the equation:

$$\Delta S = P - ET - R - D \tag{3}$$

where ΔS is change in soil moisture, P is precipitation, R is the sum of runoff and run-on, and D is drainage [10]. All units are in mm. The R term is negative when run-on exceeds runoff and positive if runoff is greater. In many arid and semiarid regions, the drainage term is often miniscule. In addition, most current agricultural practices employ measures to control runoff/run-on, such as furrow diking. This practice can make the runoff term minute. In other climates/regions where runoff and run-on can be significant, the values can be estimated from precipitation intensity and infiltration rate [11]. Other methods could also be used, such as from soil moisture sensors or runoff flumes. Drainage, or deep percolation, can be

determined from soil moisture content below the root zone. Deep soil moisture can be measured using soil moisture sensors, neutron probes, or soil cores.

In arid and semiarid regions where precipitation does not meet crop water requirements and is supplemented with irrigation, it is also important to account for the effective addition of water by irrigation. In most cases in areas such as the Texas High Plains (where runoff/run-on and drainage are negligible), the water balance is written as

$$P + I + \Delta S = ET \tag{4}$$

In addition, in these drier climates, soil moisture change between the growing seasons is typically minor, so precipitation and irrigation are the main water inputs. Since precipitation is typically small and highly variable in arid and semiarid regions, irrigation is required for maximum agricultural production. One problem in these areas is that water supplies are rapidly diminishing. This illustrates the importance of maximizing the efficiency of water use. With effective irrigation scheduling, producers can apply only the amount of water required for the respective crop.

In Texas, and especially the Texas High Plains, irrigation is the largest consumer of fresh water, most of which comes from the declining Ogallala Aquifer. In a previous study [2], it was found that reducing irrigation applications by 25 mm (1 in.) over the typical summer growing season for all the irrigated acreage in the northern Texas High Plains would save 92.5 million m³ (75,000 ac-ft) of water, also decreasing pumping costs by over \$6 million. For perspective, that 92.5 million m³ of water equates to over 2.5 months of municipal water use for the city of Houston, TX, with a population over 2 million.

The water balance approach has been widely used to estimate ET. It can be modeled seasonally by obtaining volumetric water content from soil samples at the beginning and end of the growing season. If precipitation and irrigation is measured, the change in soil moisture can be used to calculate seasonal ET. With soil moisture sensors, the same accounting approach can be performed on any time scale. The spatial resolution, however, of the water balance approach depends on the amount and spacing of soil moisture sensors or soil samples. Installing numerous sensors or taking numerous soil samples is often prohibitive due to time and funding constraints. In addition, both sensors and soil samples are specific to the small area of measurement and may not represent the surrounding field, especially in areas with highly variable soils.

1.2.1 Soil moisture measurements

Soil moisture measurements are used to determine ET through the water balance. The soil moisture measurements allow for the determination of ΔS , and with measurements of *P* and I, the water balance can be solved to yield ET.

Time-domain reflectometry (TDR) is a method to determine the soil moisture content and can be used to calculate change in soil water content as a surrogate for ET using the water balance. A TDR instrument consists of multiple probes (typically three) connected to a cable tester. The instrument works on the theory that changes in soil water content change the apparent permittivity of the soil as determined by the probes [10]. The soil moisture status can be calculated by the velocity of an electrical pulse through the probes.

A neutron soil moisture meter, or neutron probe, is an instrument that contains a radiation source that emits high energy neutrons. The high energy neutrons collide with water molecules in soil, and the reflected slower (lower energy) neutrons are counted by the probe counter. The neutron count is related to the soil moisture by a calibration. The amount of lower energy neutrons that is reflected back to the sensor provides an accurate indication to the soil moisture status [12]. In addition, the sphere of influence of the neutron meter is correlated to the soil moisture content with lower moisture contents having larger contributing values.

2. ET measurement

2.1 Energy balance

Many ET estimation methods use or are based on the energy balance. The energy balance concept describes the processes of radiation in the atmospheric boundary layer. Solar radiation is the sole energy input for radiation processes. The incoming shortwave and longwave radiation is either reflected or absorbed by the surface of the earth. The net radiation (R_n) is the amount of radiation absorbed by the earth's surface and is measured by subtracting the reflected radiation from the total incoming radiation. The absorbed radiation contributes to soil heat flux, sensible heat flux, and latent heat flux. Soil heat flux (G) is the amount of radiation gained or lost by the soil surface though conduction. Sensible heat flux (H) is the energy that increases the temperature of the atmosphere causing advection, and latent heat flux (LE) is the energy available for the evaporation of water. Due to the law of thermodynamics, the net radiation must be distributed among the other three fluxes. This yields the basic energy balance equation:

$$R_n = LE + H + G \tag{5}$$

Net radiation can be measured by a variety of instruments where the incoming solar radiation will be measured in addition to the reflected radiation. The soil heat flux can be measured by soil heat flux plates which measure the amount of energy gained or lost by the soil. H and LE require advanced instruments and methods for measurement. Since LE is the energy used for evaporating water, it can be converted to ET by dividing by the latent heat of vaporization. Measuring H and LE is more challenging and requires more sophisticated instrumentation.

2.2 Lysimeter

A lysimeter is considered the most accurate ET_a measurement instrument. A lysimeter consists of a mass of soil in an enclosed container which can be accurately weighed to determine the amount of water lost or gained per unit time. Lysimeters can be very complex and expensive to install and operate but are a direct measurement of soil water storage. Thus, lysimeters are considered the most accurate for ET measurement [13, 14]. Lysimeters are point measurements and only have the measurement area of the container. However, if the surrounding field is properly managed to match the lysimeter, the ET data can represent field conditions. This intensive management is typically only possible at research locations. One example of large continuously weighing lysimeters are those located at the USDA-ARS Conservation and Production Research Laboratory (CPRL) in Bushland, TX. This location houses four lysimeters within a 20 ha (50 acre) field, divided into four quadrants.

Each large weighing lysimeter measures 3 by 3 m on the surface by 2.3 m deep over a fine sand drainage base (see **Figure 2**). It contains an undisturbed monolithic



Figure 2.

Large weighing lysimeters at the USDA-ARS Conservation and Production Research Laboratory (CPRL), Bushland, TX. The view from the surface (a) shows the outline of the lysimeter container during the growing season with a crop planted. A photo taken during installation (b) illustrates the size of the container and the below ground access housing.

Pullman clay loam soil profile with subsurface drip irrigation at 23 cm depth or mid elevation sprinkler irrigation. The soil container rests on a large agronomic scale equipped with a counterbalance and load cell system. Initial design and installation details of the lysimeter are provided by [15, 16]. The lysimeters were later equipped with drainage effluent tanks suspended from the lysimeter by load cells for separate measurement of drainage mass without changing total lysimeter mass. Load cell output is measured and recorded by a precision data logger. Load cell voltage outputs are converted to mass using calibration equations, and 5 minute means are used to develop a base dataset for subsequent processing [17]. Lysimeter mass in kg is converted to a mass-equivalent relative lysimeter storage value (mm of water) by dividing it by the relevant surface area of the lysimeter ($\sim 9 \text{ m}^2$) and the density of water (1000 kg m^3). Equivalent mass values allow for changes in lysimeter mass to be expressed in terms of water flux, defined as mm of water lost or gained per unit time. The lysimeter data logger mass resolution is better than 0.001 mm when converted to equivalent depth of water. Lysimeter accuracy is, however, determined by the RMSE of calibration, which has ranged from 0.05 mm to 0.01 mm [14, 17]. Lysimeter quality assurance and quality control (QA/QC) and data processing techniques are provided by [18].

Calculating ET in units of equivalent depth of water requires that the change in lysimeter mass be divided by the effective evaporating and transpiring area of the lysimeter [13]. The Bushland lysimeter inside surface area is 8.95 m² [14]; however, the area of contribution from captured precipitation or irrigation, as well as ET, is beyond the lysimeter container, resulting in an effective area larger than the physical area of the lysimeter. The reported the outside lysimeter surface area was 9.35 m² [10]. In this case, a correction factor of 1.05 (9.35/8.95 m²) is applied to ET measurements from the lysimeter.

The lysimeter is designed to be the representative of the surrounding field so that measured lysimeter ET closely mimics field ET. Experienced support scientists and technicians are responsible for maintaining lysimeter representativeness as compared to surrounding fields. Careful attention is given to agronomic operations including planting, harvesting, tillage, fertilization, irrigation, and pesticide application such that there should be no distinguishable differences, particularly in height, between the crop grown on the lysimeter and that grown in the surrounding field. To confirm this, multiple neutron probe access sites were located both throughout the field and in the lysimeter to monitor the soil profile water content. Weekly soil water content (SWC) readings from the neutron probes throughout the field are compared to SWC readings from the lysimeter to determine similitude representativeness. In addition to SWC readings, plant mapping and stand counts were periodically taken to ensure the crop growth on the lysimeter approximates the surrounding field. The lysimeter box contains a \sim 50 mm freeboard lip that extends above the soil surface to limit runoff or run-on to the lysimeter. Similarly, furrow dikes are used to limit runoff and run-on for the surrounding field.

The lysimeters at the CPRL are a great example of large weighing lysimeters. However, there are many types and sizes of lysimeters. Some are constantly weighing, such as those at the CPRL, while others are weighed periodically. In addition, lysimeters can vary in size. The large weighing lysimeters at the CPRL are considered highly accurate due to their large size, where the effects of the enclosed space on the plants are minimal. Smaller lysimeters will contain more error, especially if the soil volume is small enough where root growth is impeded. With lysimeters, the accuracy is dependent on the lysimeter design, representativeness, maintenance, and operation. Smaller lysimeters can have value, even if they are not highly accurate. An example of the usefulness of smaller lysimeters is the Soil-Plant-Environment Research (SPER) facility, also at the CPRL (see Figure 3). This facility is equipped with 48 lysimeters, each measuring 1 m by 0.75 m by 2.3 m deep [19]. The 48 lysimeters are comprised of 12 replications each of Ulysses silt loam soil from the Garden City, KS area; Pullman clay loam soil from Bushland, TX; Amarillo sandy loam soil from the Big Spring, TX area; and Vingo fine sand soil from the Dalhart, TX area. These represent the four main soil types of the Southern Great Plains of the United States. The SPER contains an automatically controlled rainout shelter that covers the lysimeters during precipitation events, which allows water additions to be precisely controlled through surface drip irrigation. The size of the



Figure 3.

The Soil-Plant-Environment Research (SPER) facility at the Conservation and Production Research Laboratory (CPRL) in Bushland, TX. This facility contains 48 smaller lysimeters consisting of 12 replications of the 4 main soil types throughout the Southern Great Plains region of the United States. An automatic rainout shelter (seen in the background) covers the lysimeters during precipitation events so that water can be precisely controlled through surface drip irrigation.

lysimeters at the SPER, and the open space between the lysimeters, limits their absolute accuracy. However, this facility can provide good comparisons between treatments and soil types. This illustrates that even though the quantitative measurements from some lysimeters may be lacking, the qualitative data can still be quite valuable. More information regarding lysimeter research at the CPRL can be found in [20].

Even smaller lysimeters can also have value. Temporary "micro lysimeters" have been used to measure soil evaporation on a daily time step (see Figure 4). Lysimeters such as these are useful in research involving partitioning evaporation and transpiration separately. Like the small lysimeters at the CPRL, the micro lysimeters are not perfectly accurate but can still provide meaningful data for certain purposes. Many other lysimeter designs have been used and can be permanent or temporary. Large weighing lysimeters are the most accurate, but other, simpler, and more costeffective options are available. As with most instruments, the accuracy and usefulness of the data will depend on the purpose and management of the lysimeter.

2.3 Bowen ratio

Bowen ratio is a method of partitioning fluxes between latent and sensible heat based on flux-profile relationships for energy and mass exchange [21]. This method assumes flux directions are vertical and no horizontal flux movement occurs. Measurements of air temperature and relative humidity are taken at two different heights in the same location. The relative humidity is used to calculate the vapor pressure. The Bowen ratio is the ratio of sensible heat flux to latent heat flux and can be calculated as

ß

$$\beta = r \frac{\Delta t}{\Delta e}$$
(6)

$$=\gamma \frac{\Delta T}{\Delta e}$$

Figure 4.

An example of a "micro lysimeter" used to determine soil evaporation. The inner soil container can be removed from the outer housing and manually weighed.

where β is the Bowen ratio, γ is the psychometric constant, ΔT is the temperature difference between the two measurements, and Δe is the vapor pressure difference between the two measurements. Using the Bowen ratio, the sensible and latent heat flux are calculated by

$$LE = \frac{R_n - G}{1 + \beta} \tag{7}$$



where *LE* is the latent heat flux, R_n is net radiation, *G* is soil heat flux, and *H* is sensible heat flux [21]. All units are W m⁻². The Bowen ratio method has been shown to contain errors of 25–30% [22, 23].

2.4 Eddy covariance

Eddy covariance (EC) systems are based on the theory that as wind moves, it does not move unidirectionally but in three-dimensional circular patterns, or eddies [24]. In addition, as the air moves, it carries with it molecules of water vapor and other gases such as carbon dioxide, methane, and others. If the speed of these eddies can be determined in all three directions, then the movement of the molecules can be determined. In conjunction, a gas analyzer can be used to measure the amounts of water vapor (or other gases) the air contains at that moment in time. The covariance between the movement of the air mass and the composition of that same air mass can be used to determine the water flux (or fluxes of carbon dioxide and methane), in addition to H, LE, and ET. This is the basis for EC systems (see **Figure 5**), where a three-dimensional sonic anemometer and an infrared gas analyzer (or krypton hygrometer) are used to collect the aforementioned data.



Figure 5.

A typical eddy covariance system consisting of a three-dimensional sonic anemometer (CSAT-3, Campbell Scientific Inc., Logan, UT) and an infrared gas analyzer (LI-7500, LI-COR Biosciences, Lincoln, NE).

The flux for any gas can be calculated from the EC data by

$$F = \overline{\rho} \overline{w^{\,s} s^{\,s}} \tag{9}$$

where $\overline{\rho}$ is the mean air density and \overline{w} and \overline{s} are deviations from the mean for wind speed and dry mole fraction, respectively [24]. The dry mole fraction can be determined for any gas or variable of interest. From this principle, H and LE can be calculated by

and

$$H = \overline{\rho}C_{p}\overline{w^{T}}$$

$$LE = \lambda \frac{M_{w}/M_{a}}{\overline{p}}\overline{\rho}\overline{w^{e}}$$
(10)
(11)

where C_p is the specific heat of air, \overline{T} is deviation from mean temperature, λ is latent heat of vaporization, M_w is the molecular weight of water vapor, M_a is the molecular weight of dry air, \overline{P} is mean atmospheric pressure, and \overline{e} is the deviation from mean vapor pressure.

To detect the fast movements of certain eddies, EC measurements are typically taken at very short intervals, often 10–20 measurements per second (10– 20 Hz sampling rate). A very fine measurement resolution is needed to capture the rapid changes in gas concentration and eddy movements. The quantity of data acquired also provides adequate sample size for the covariance analysis. Although measurement acquisition is very frequent, the data will typically be averaged to 30 minutes for flux computations. The 30 minute time step is comparable to the period of significant, unsteady atmospheric motions [25]. The spatial scale of EC measurements is directly affected by sensor height. A sensor height of 2 m will have a measurement footprint of approximately 150 m, and the footprint will increase with higher sensor heights [24].

Several corrections are typically applied to raw EC data to compensate for instrumentation arrangement and ensure that the assumptions of the EC technique are generally valid [26]. These include corrections for coordinate rotation, air density, frequency-dependent signal loss, and Webb, Pearman, and Leuning (WPL) corrections [27]. The coordinate rotation correction converts the flux data so that the orientation is where fluxes are perpendicular to the surface. The air density correction accounts for density fluctuations due to temperature and humidity fluctuations [26]. Frequency-dependent signal loss corrections account for signal losses in the high and the low frequency ranges [27]. The WPL corrections account for fluctuations in gas concentration due to temperature and humidity fluctuations, which do not contribute to the gas fluxes.

 ET_a can be determined from EC systems where the water vapor flux is calculated. EC systems can be used to determine the energy balance when the R_n and G are also measured. The basic energy balance equation is given in Eq. 4. Based on this equation, the sum of H and LE should equal the difference between R_n and G. It has been acknowledged that EC systems have an issue with energy balance closure where $Rn-G \neq LE + H$ [28]. Previous studies have typically shown that there will be residual energy which is unaccounted. Even with the energy balance closure error, the error of ET calculated from the LE using EC is around 20–30% [29].

2.5 Scintillometry

Scintillometers consist of a transmitter and receiver, separated by a specified path length. Scintillometry uses a beam of electromagnetic radiation of known wavelength transmitted across a relatively large distance (100 m–4.5 km). The beam intensity fluctuates as it encounters gases in the air due to absorption and diffraction. These fluctuations, or scintillations, can be used to determine the structural parameters for temperature and the refractive index of air, which can be used to calculate the H. The calculations to obtain H from scintillometers are based on Monin-Obukhov Similarity Theory (MOST). This theory describes the relationship between the parameters of friction velocity, the temperature scale, and the specific humidity scale, in reference to the process of turbulence, mainly from buoyancy and horizontal shear.

MOST describes the vertical flow and turbulence properties in the lower atmospheric boundary layer or surface layer [30]. This theory provides a set of equations that relate turbulence properties, using dimensionless parameters, to atmospheric processes including H. One of the parameters derived from similarity theory is the Obukhov length, which is the height above the surface that turbulence is caused by wind shear. Above the Obukhov length, turbulence is driven more by buoyancy, or the action of radiant heat moving the air mass upwards. MOST was developed on the idea that turbulence properties, when made dimensionless using friction velocity, temperature scale, and other variables, are a universal function of the Obukhov length [31, 32]. The key parameters of MOST are the friction velocity, u_* ; temperature scale, θ_* ; and the specific humidity scale, q_* [33]. These parameters are calculated as

$$u_* = \left(\frac{\tau_0}{\rho}\right)^{1/2} \tag{12}$$

$$\theta_* = -\frac{H_0}{\rho c_p u_*} \tag{13}$$

$$q_* = -\frac{E_0}{\rho u_*} \tag{14}$$

where ρ is the air density, c_p is the specific heat of air, τ_0 is the turbulent stress at the surface, H_0 is the vertical flux of heat, and E_0 is the vertical flux of water vapor. τ_0 , H_0 , and E_0 can be calculated by

$$\tau_0 = \rho C_D U_r^2 \tag{15}$$

$$H_0 = \rho C_D C_H U_r (\Theta_s - \Theta_r) \tag{16}$$

$$E_0 = \rho C_W U_r (Q_s - Q_r) \tag{17}$$

where U_r is the wind speed at reference height, Θ_r is air temperature at reference height, Q_r is specific humidity at reference height, Θ_s is air temperature at the surface, Q_s is specific humidity at the surface, C_D is the drag coefficient, C_H is the heat transfer coefficient, and C_W is the water vapor transfer coefficient [33]. Monin, Lumley [34] determined that turbulence properties at height z depend on only five quantities: z, ϱ , $\frac{g}{T}$, u_* , and $\frac{q}{c_p\rho}$. From these parameters, one dimensionless parameter, the stability parameter ζ , can be derived. Using ζ , surface flow properties can be described as a function of ζ using dimensional analysis. ζ is calculated as

$$\zeta = \frac{z}{L} \tag{18}$$

where z is the height above the surface and L is the Obukhov length

$$L = -\frac{u_*^3}{k \frac{g}{T_0} \frac{q}{c_p \rho_0}}$$
(19)

where k is the von Karman constant, g is the acceleration due to gravity, ρ_0 is the air density at temperature T_0 , and q is the kinematic heat flux [34].

There are different scintillometer models available, which differ based on the wavelength of the radiation beam and aperture diameter. The aperture diameter determines the path length where a larger aperture will need a longer path length. All models use a transmitter and receiver to send the beam and measure the scintillations. The most common wavelengths are visible (670 nm), infrared (880 nm), and microwave (1 mm to 1 cm). The aperture size for most infrared (large aperture) scintillometers (LAS) is 10–15 cm (see **Figure 6**), while the aperture for visible (surface layer) scintillometers (SLS) is 2.7 mm (see **Figure 7**). Microwave scintillometer aperture sizes can be much larger, up to 30 cm. The SLS is sometimes termed as a displaced-beam small aperture scintillometer (DBSAS) since the SLS beam is split into two parallel beams, displaced by 2.7 mm. Based on the correlation of the intensity fluctuations between the two beams, the inner scale parameter, l_0 , can be determined [31].

The benefits of each scintillometer come from the differences between them. For instance, the visible wavelength scintillometers have a much smaller aperture, which allows for better representation of small eddies and greater sensitivity to smaller changes in temperature and wind fluctuations. The larger apertures can



Figure 6.

Large aperture scintillometer (LAS MKII, Kipp & Zonen, Delft, the Netherlands) with aperture restrictor plate reducing aperture from 15 cm to 10 cm.



Figure 7. Surface layer scintillometer (SLS-20, Scintec AG, Rottenburg, Germany).

detect larger eddies better. Microwave scintillometers are more sensitive to humidity fluctuations than the other wavelengths, providing an inference to accurate ET_a determination [35]. Infrared scintillometers have been used for some time; however, visible and microwave scintillometers are relatively new and have not been used as extensively.

Scintillometers have been considered to be very beneficial for ET remote sensing studies due to their large path length. Specifically, large aperture scintillometers (LAS), which can have path lengths up to a few km, can have a large enough spatial footprint to be similar to most remote sensing data resolution. In addition, the path averaging of the scintillometer provides an integrated benefit in that a homogenous surface is not required to meet any assumptions. This allows the scintillometer to be used across varying terrain and provide an averaged value. The averaging for variable surfaces is similar to that of remote sensing data. The previous points illustrate how the scintillometer can serve as a ground-truthing instrument or as a source of validation data for remote sensing. Since most of the ET remote sensing models are based on the surface energy balance, similarly to scintillometry, measurements other than just ET can be evaluated. The surface fluxes H and LE are determined by both scintillometers and ET remote sensing models, which provide more data for comparison. One benefit scintillometry has over EC is the lack of corrections [36].

One advantage SLS offers over other point source measurements is that the fluxes can be determined over shorter lengths and at heights closer to the surface [37]. In addition, the fluxes can be calculated on shorter temporal scales, as low as 1 minute, compared to EC, for example, which typically uses a 30 minute interval. An advantage the SLS has over the LAS is that the SLS determines the l_0 , which is proportional to the dissipation rate of the turbulent kinetic energy, ε , and C_T^2 ,

which are used to determine the H. Without determining the l_0 , the LAS requires additional measurements to estimate the friction velocity. The SLS has been found to be more accurate than the LAS with errors of 15–30% [29] compared to greater than 30% for the LAS [38].

2.6 Remote sensing

Many ET models are available for use with remote sensing data. In addition, there are a variety of satellite data sources such as Moderate Resolution Imaging Spectrometer (MODIS), Landsat, Advanced Very High-Resolution Radiometer (AVHRR), Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER), and many others. Additional information on satellite sources and available models can be found in [39]. Most remote sensing models are based on the energy balance where the reflectance from remote sensing is used with weather data from nearby weather stations and the four components of the energy balance are calculated. Typically, LE is calculated as the residual of the energy balance and converted to ET at an hourly and daily time step.

The biggest issue with using satellite data for creating ET maps is poor spatial and temporal resolution. Many energy balance-based models such as METRIC [40], SEBS [41], SEBAL [42], and others require thermal data to calculate surface temperature. These models are more limited on available data. ASTER, MODIS, and Landsat are the main data sources available with thermal sensors. ASTER has the highest spatial resolution at 15 m for visible wavelengths and 90 m for thermal wavelengths but has a return interval of 16 days [43]. Landsat data has 100 m resolution for thermal wavelengths and can provide data on an 8 day interval if Landsat 7 and Landsat 8 are both used [44]. MODIS provides daily data but has poor spatial resolution of 1000 m [45]. Although the models typically only provide hourly and daily ET estimates, methods are available to interpolate between satellite passes and for monthly and seasonal sums [46].

The aforementioned remote sensing models not only provide ET_a maps but can also provide estimates of leaf area index, surface temperature, surface albedo, and many others. Although the spatial and temporal resolution of existing satellites limits applications to field-scale agricultural use, the rapid increase in unmanned aerial vehicle (UAV) technology shows vast potential to acquire remote sensing data with spatial resolution at a centimeter scale and as frequent as desired. Satellite-based ET maps typically have accuracy of 20–30% at best [47]; however, the accuracy of using UAV data for ET maps is not currently known.

3. Conclusions

The methods mentioned above can all be used to determine ET; however, there are disadvantages to each one of them. With the soil water balance approach, the drainage and runoff terms can be difficult to determine. Although they are commonly miniscule in arid and semiarid regions, they would still need to be accounted for to obtain the greatest accuracy. Lysimeters are the most accurate but are very expensive and intrusive to install and operate. In addition, they require a high level of knowledge and experience to obtain the best measurements. The Bowen ratio method has been used to determine ET from the energy balance, but it is an indirect measurement. EC is a direct measurement method of turbulent fluxes but is known to have energy balance closure and other errors associated with it. Scintillometers are another indirect measurement method that has been extensively used, but they also have known errors. EC and scintillometers are two of the more common

Advanced Evapotranspiration Methods and Applications

turbulent flux and ET measurement instruments typically used. Remote sensing model studies are also widely available in the literature; however, each method, other than a lysimeter, typically has 20–30% error. For irrigation scheduling and some water resources management, 20–30% error may not be adequate, but for other purposes, such as modeling and land use change monitoring, the error may be acceptable. With many options available, the most suitable instrument/method will be dependent on the purpose and use of the ET data.

Conflict of interest

The author declares no conflict of interest.

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