Estimating Evaporation in a Prairie Landscape under Drought Conditions

Robert W. Armstrong, John W. Pomeroy, and Lawrence W. Martz

Abstract: Physically-based atmospheric models of evapotranspiration (ET) that consider the Penman combination energy balance and aerodynamic approach have achieved acceptance as useful tools for obtaining estimates of actual ET from land surfaces. These models have been made applicable to the case of non-saturated conditions through either surface resistance formulations (e.g., Penman-Monteith) or by application of the complementary evaporation theory of feedback between the atmosphere and surface moisture states (e.g., Granger-Gray). Their application becomes complicated under conditions of drought, when extremely low soil moisture availability severely restricts ET from the soil and plants. Under such severe conditions, consideration for the surface water balance and interactions with the balance of available energy and aerodynamic principles are important for accurately estimating actual ET. A modelling application is demonstrated using the Cold Regions Hydrological Model (CRHM) platform to examine the estimation of ET under drought conditions. CRHM allows users to assemble hydrological models by linking a suite of modular physically-based algorithms that describe the individual processes. In this case, the models assembled consider infiltration, evaporation, and soil moisture accounting and are applied to a mixed prairie located at Lethbridge, Alberta, Canada under drought conditions during the growing period in 2000 and 2001. Near surface meteorological and ecological observations used as model input and for evaluating model performance were obtained through the Ameriflux network and the Agriculture and Agri-Food Canada (AFC) Lethbridge Research Centre. Results show that consideration for the effective rooting zone depth of the mixed-prairie at the site is important for estimating actual ET using the Penman-Montieth and Granger-Gray models during severe moisture stress. Relative differences in ET estimates provided by the models are discussed in the context of their contrasting theoretical approaches.

Résumé : Les modèles atmosphériques physiques de l'évapotranspiration (ET) qui tiennent compte de la formule de Penman combinant le bilan d'énergie et l'approche aérodynamique ont été reconnus en tant qu'outils pratiques pour l'obtention d'estimations de l'évapotranspiration réelle des surfaces terrestres. Ces modèles ont été rendus applicables aux conditions non saturées, soit au moyen des formulations de la résistance de surface (p. ex. Penman-Monteith), soit au moyen de l'application de la théorie de l'évaporation complémentaire par rétroaction entre l'état atmosphérique et l'état d'humidité superficielle

Robert W. Armstrong[1], John W. Pomeroy[1], and Lawrence W. Martz[2]

[1] Centre for Hydrology, University of Saskatchewan, Saskatoon, SK S7N 5C8
[2] Graduate Studies and Research, University of Saskatchewan, Saskatoon, SK S7N 5A2

Submitted May 2009; accepted March 2010. Written comments on this paper will be accepted until December 2010.

(p. ex. Granger-Gray). Leur application devient compliquée dans des conditions de sécheresse, lorsque la disponibilité de l'humidité du sol est extrêmement faible et nuit gravement à l'évapotranspiration par le sol et par les plantes. Dans des conditions aussi extrêmes, il est important de prendre en considération le bilan hydrique des surfaces et les interactions avec le reste de l'énergie disponible ainsi que les principes d'aérodynamique afin d'estimer avec précision l'évapotranspiration réelle. Une application de modélisation est démontrée à l'aide de la plateforme du Modèle hydrologique pour les régions froides / en anglais le « Cold Regions Hydrological Model » (CRHM) pour l'examen de l'estimation de l'évapotranspiration dans des conditions de sécheresse. Le modèle CRHM permet aux utilisateurs d'assembler les modèles hydrologiques en reliant une série d'algorithmes modulaires qui décrivent les processus individuels. En l'occurrence, les modèles assemblés tiennent compte de l'infiltration, de l'évaporation et de l'humidité du sol et sont appliqués à une prairie mixte située à Lethbridge, en Alberta, au Canada, et ce, dans des conditions de sécheresse pendant la saison de croissance en 2000 et en 2001. Des observations météorologiques et écologiques à faible profondeur servant de données d'entrée au modèle et à l'évaluation du rendement du modèle ont été obtenues par l'entremise du réseau AmeriFlux et du Centre de recherches de Lethbridge d'Agriculture et Agroalimentaire Canada (AAC). Les résultats révèlent qu'il est important de considérer la profondeur efficace de la rhizosphère de la prairie mixte au site visé afin d'estimer l'évapotranspiration réelle à l'aide des modèles Penman-Montieth et Granger-Gray lors des épisodes de tension hydrique intense. Les différences relatives dans les estimations de l'évapotranspiration fournies par les modèles sont abordées dans le contexte de leurs approches théoriques opposées.

Introduction

The estimation of evaporative losses from land surfaces is required for a variety of water resource management, agricultural, hydrological, and climatological purposes. Evapotranspiration (ET) is of great importance to water resources management and planning (Gowda *et al.*, 2008), and particularly during drought when it can dominate regional water balances. Obtaining consistently reliable estimates of actual ET, however, can be a difficult task on a practical basis for operational water management. Typically, estimates of actual ET are obtained from estimates of the potential ET as a function of water availability through surface and soil moisture accounting in hydrological models (e.g., Leavesley *et al.*, 1983; Kite, 1995; Kouwen, 2001). In general, the numerous methods available for calculating potential ET give a range of estimates which only adds to the uncertainty in estimating actual ET for the purpose of hydrological modelling.

Nonetheless, physically-based Penman-type combination models that consider a combination of energy balance and aerodynamic principles provide practical and useful tools for estimating actual ET from the land surface. Penman's original equation (Penman, 1948) is commonly referred to as the Penman potential evaporation equation which is applicable for grassed surfaces with an abundant supply of water (Granger, 1989a). Monteith (1965) and Granger and Gray (1989) amongst others have shown that the Penman equation provides a convenient and consistent estimate of the potential evaporation based on the general climate conditions. Diverging theoretical approaches have led to the development of resistance-type and complementary relationship formulations for estimating actual ET from non-saturated surfaces (e.g., Monteith, 1965; Granger, 1989a). These models are generally applicable for estimating ET from a wide variety of terrestrial land covers under various climate conditions (Jacobs *et al.*, 2002).

The Penman-Montieth (Montieth, 1965) model, designated as PM herein, considers the control exerted by plants on evaporative losses from stomata by introducing a canopy resistance term in addition to aerodynamic resistance. The canopy resistance is used to regulate the transfer of water vapour from stomatal cavities to the atmosphere during the process of photosynthesis in response to changing environmental conditions (e.g., light, soil moisture, humidity, temperature, CO_2, etc). Plant photosynthesis is strongly linked to soil moisture conditions as there typically is more than enough light and CO_2 for adequate plant growth over a growing season. As such, incorporation of a canopy resistance term generally increases model complexity because descriptions of physical characteristics such as canopy height and leaf area of the dominant plant species are needed. These are often obtained through field measurements over manageable areas or derived from remote sensing observations over large regions.

By contrast, a complementary theory approach to estimating ET simply assumes that evaporative losses can be described as a function of surface-atmosphere feedbacks. This eliminates the need to describe complex surface resistances which are exceedingly difficult to verify over large areas. The implicit assumption of complementary theory is that, as near surface moisture availability decreases, the dryness of the atmosphere increases. In other words, potential evaporation increases with surface drying but the rate of actual evaporation decreases due to reductions in moisture availability. Granger and Gray (1989) and Granger (1989b) integrated this theory with the Penman equation by introducing a 'relative evaporation' term. The relative evaporation, *G (ratio of actual to potential ET)* is assumed to be inversely related to the relative drying power of the air, *D* which is further assumed to be a function of both the available energy and humidity deficit; designated GD herein to reflect the GD relationship.

An interesting consequence of determining the relative evaporation is that the model is not always dependent on soil moisture information for operational purposes. Nor is there a need for plant characteristics such as leaf area, whilst the PM surface resistance is typically quite dependent upon such information (Jarvis, 1976). Nevertheless, the PM and GD models have been applied to a variety of land surfaces with reasonable success (e.g., Jarvis, 1976; Granger and Pomeroy, 1997; Wever *et al.*, 2002; Armstrong *et al.*, 2008).

Soil Moisture and Effective Rooting Zone Depth

The relative physical complexity of a given hydrological model is a fundamental consideration for practical applications. Comprehensive and accurate descriptions of interactions between the soil-vegetation-atmosphere continue to be a challenge for environmental modelling in general. Just as with plant controls for regulating water losses, soil moisture availability is also critical for determining water and energy balances. During periods of reduced water availability and particularly under drought conditions, ET from the land surface may be greatly reduced but may also dominate the water balance.

From an ecohydrological perspective, a key to estimating actual ET at a given location or region under such conditions is knowledge of physiological characteristics of the dominant plant species; this includes growth both above and below the ground surface. The availablity of soil moisture accessible by the root systems of plants has long been considered the main limiting factor affecting prairie grasses and crop yields (Weaver, 1925; Weaver, 1926; Weaver and Clements, 1938). As a result, conditions of prolonged drying and drought strongly reduce evaporative losses from soils and plants as soil moisture reserves become depleted. Still, the magnitude to which ET is impacted is generally dependent on physiological adaptations of the plant species to handle periods of water stress.

Extensive research on grasslands and crops and their rooting habits within the Great Plains by Weaver (1925; 1926) and Weaver and Clements (1938) indicated that short to mid height grasses typically root to depths of approximately 1 m to 1.5 m. Tall grass species can extend to depths of 2 to 3 m, and legumes such as alfalfa to greater than 4 m. During drought conditions, however, short to mid height grasses invest more energy in increasing the area of their roots but extend their depth very little. Above ground the plant is typically dwarfed and may even go dormant, only to become revitalized when adequate moisture becomes available.

Plant roots and their distribution within the soil profile are also important with regard to the water storage potential of the soil (England, 1975). A review by Jackson *et al.*, (1996) showed the largest fraction of roots for terrestrial plants and trees tend to be located in the upper layers of soil. They found that on a global average, 75% of the root profile is located in the top 40 cm of the soil. For grasses, 80% of the roots are within the top 20 cm to 75 cm of the soil but also depends on plant species and biome. Therefore it is evident that plant root development and effective rooting depth is an important factor when considering the soil moisture available to plants in the terrestrial environment. However, modelling the dynamics of rooting zone development with moisture availability would be a difficult and likely unnecessary challenge for the purpose of operational water management.

Under normal to wet environmental conditions, the largest changes in soil moisture due to ET have been shown to occur in these uppermost layers of the soil profile; typically at depths less than 50 cm. Observational evidence is provided from data obtained for the 2005 growing season at St. Denis National Wildlife Area (SDNWA), located in south-central Saskatchewan, Canada. Here, Lungal (2008) found large and rapid variations in soil moisture over time at depths less than 50 cm from the soil surface at several locations surrounding a wetland. Smaller variations over time were found at depths between 50 cm to 100 cm. At the same location under similar rainfall and moisture conditions, it was shown that reasonable estimates of actual ET could be obtained by applying the PM and GD models without any restriction for continuity in surface and soil moisture (Armstrong *et al.*, 2008).

In other words, when near surface moisture was not limiting, ET could be estimated from the general atmospheric conditions without the need for coupling the models to the soil moisture balance thereby reducing modelling complexity. This is not possible, however, under drought conditions where soil moisture becomes the critical limitation and continuity must be enforced as moisture reserves become depleted. This brings into question the main considerations for modelling actual ET under drought. For example, how adequately can ET be estimated during drought conditions given the general characteristics of the physical environment and physiology of the plants such as their effective rooting depth? Also, is the effective rooting depth an important consideration even in the case of plants that typically root in shallower layers?

Objective and Rationale

The main objective of this paper is to present a simple modelling application to examine the effects of soil depth or the effective rooting depth on cumulative estimates of actual ET for a Canadian Prairie environment during severe drought. The modelling is performed within the Cold Regions Hydrological Model (CRHM) platform (Pomeroy *et al.*, 2007). Within CRHM a hydrological model is assembled to estimate actual evaporation from a mixed prairie at Lethbridge, Alberta in 2000 and 2001. The application presented here is relevant to better understanding and modelling of ET processes during the 1999-2004 drought period that affected various regions of Canada and is the focus of the Drought Research Initiative (Stewart *et al.*, 2008).

Study Site

The modelling techniques used herein are applied to archived data from an Ameriflux mixed prairie study site located at Lethbridge, Alberta. The Lethbridge grassland monitoring site is part of a much larger biospheric global monitoring network (FLUXNET). A crucial benefit of this monitoring network is a wealth of data obtained by eddy covariance, remote sensing and in situ means for better understanding of water, energy, and carbon fluxes (Running *et al.*, 1999). A comprehensive overview of FLUXNET objectives, monitoring sites and data availability is provided by Baldocchi *et al.* (2001). Stockli *et al.* (2008) have demonstrated the value of using such data for validating models, examining model deficiencies, and potential applications for better understanding physical processes driving surface fluxes.

Wever *et al.* (2002) have previously reported considerable variation in interannual ET estimates for the years 1998–2000 for the Lethbridge mixed landprairie site. They observed a large reduction in the peak evaporation rates in 1999 and 2000 compared to those of 1998; 3 mm per day under drought conditions *versus* 4.5 mm per day before the onset of drought. Similar rates have been summarized in the range from normal to dry conditions at other grassland sites in the Canadian Prairies (Armstrong *et al.*, 2008). Such information is potentially valuable when considering transitions from normal to drought conditions for regions where water management and planning under extreme conditions is a concern.

Wever *et al.* (2002) and Flanagan *et al.* (2002) describe the site as relatively flat with dark-brown chernozem soils that are clay loam to clay in texture. The site is characterized by a mix of short to mid height grasses and forbs. The dominant grasses (the focus of this study) include wheatgrass, commonly known as thickspike (Agropyron *dasystachyum)* and western

(Agropyron *smithii). Other short grass species include needlegrass (Stipa comata) and blue grama (Bouteloua gracilis*). Physiological observations provided by Weaver (1926), such as the rooting habits of grass and crop species under various environmental conditions provide valuable information that can be used for parameterizing hydrological models on a physical basis.

The focus of the present application is on actual ET estimates during the growing season period (May 1–September 30) for the years 2000 and 2001 which were extreme drought years for the Lethbridge site. Archived Environment Canada data (DAI Team) indicate that growing season rainfall was extremely low for these years with only 111 mm recorded in 2000 and 83 mm recorded in 2001 over the same period. By comparison, much higher rainfall was recorded during 1998 (240 mm) and 1999 (200 mm) with a wet period occurring in 2002 (350 mm). Flanagan and Johnson (2005) reported that the average plant height during 2001 was measured to be approximately 19 cm compared to 34 cm in 2002. Ecological measurements such as leaf area index (LAI) and the timing of maximum LAI were also observed to be appreciable different during this period. In late May of 2000, LAI reached a maximum of 0.43 and was 0.5 in 2001 in late June. In late July of 2002 LAI was much greater at 1.22.

General Environmental Conditions: Measured Precipitation, Evaporation and Near Surface Soil Moisture

For instructive purposes it is useful to consider the actual evaporative losses during the 2000 and 2001 drought years in the context of relevant water balance components. Precipitation (irrigation not a factor) represents the only input, and in the present case runoff can reasonably be neglected due to lack of any heavy convective rainfall events during these drought years. This allows the water budget to be simplified down to a vertical problem. As a result, only precipitation and evaporation are considered and the change in storage can be simply determined from the difference between them, $\Delta S = P - E$.

As shown in Figure 1 the measured total evaporation is much higher than precipitation for the period of May 1–September 30 for both 2000 and 2001. In 2000 measured ET (188 mm) is approximately 1.6 times larger than the total precipitation of 113 mm observed at the study site and in 2001 measured ET (164 mm) is nearly 2.5 times larger than the total precipitation of only 67 mm observed at the study site. These differences suggest rather large changes in soil moisture storage for these years of approximately -75 mm and -100 mm, respectively.

It is also noteworthy from Figure 1 that there is marked change in the slope of the cumulative evaporation curve in both years which occurs in early July and is more pronounced in 2001. By July 2 of 2000 the cumulative total was 105 mm (1.67 mm per day on average) or 56% of the total observed ET. By comparison only 37% of the total precipitation had been received by this date (Figure 1). The daily average ET dropped abruptly thereafter to 0.91 mm for the remainder of the period. For 2001 the difference was even greater; by July 4 the cumulative total was 119 mm (1.83 mm per day on average) or 72% of the total ET whilst 86% of the total rainfall had been received. Thereafter, the decline in the daily average ET (0.51 mm per day) was even larger than that in 2000.

As is to be expected, relative changes in observed ET tend to coincide with changes in soil moisture (Figure 2). ET (mm/day) lags slightly behind observed changes in volumetric soil moisture as clearly shown for the simpler case in 2001. For example, 27 mm of rainfall was received over a two day period in early June thereby increasing the soil moisture content from 0.21 to 0.31. This was quickly followed by a peak period of ET that was maintained through most of June. By early July both the volumetric soil moisture (~0.18) and ET (generally <1 mm) had declined dramatically, signalling severe moisture limitations.

Methods

Cold Regions Hydrological Model Platform

Pomeroy *et al.* (2007) provided an overview of CRHM and various modules available. For this application, the CRHM platform permits linking ET algorithms to precipitation, runoff and soil moisture accounting to impose limits on ET under moisture stressed conditions. CRHM itself is not a model but rather is a structured platform that allows users to generate

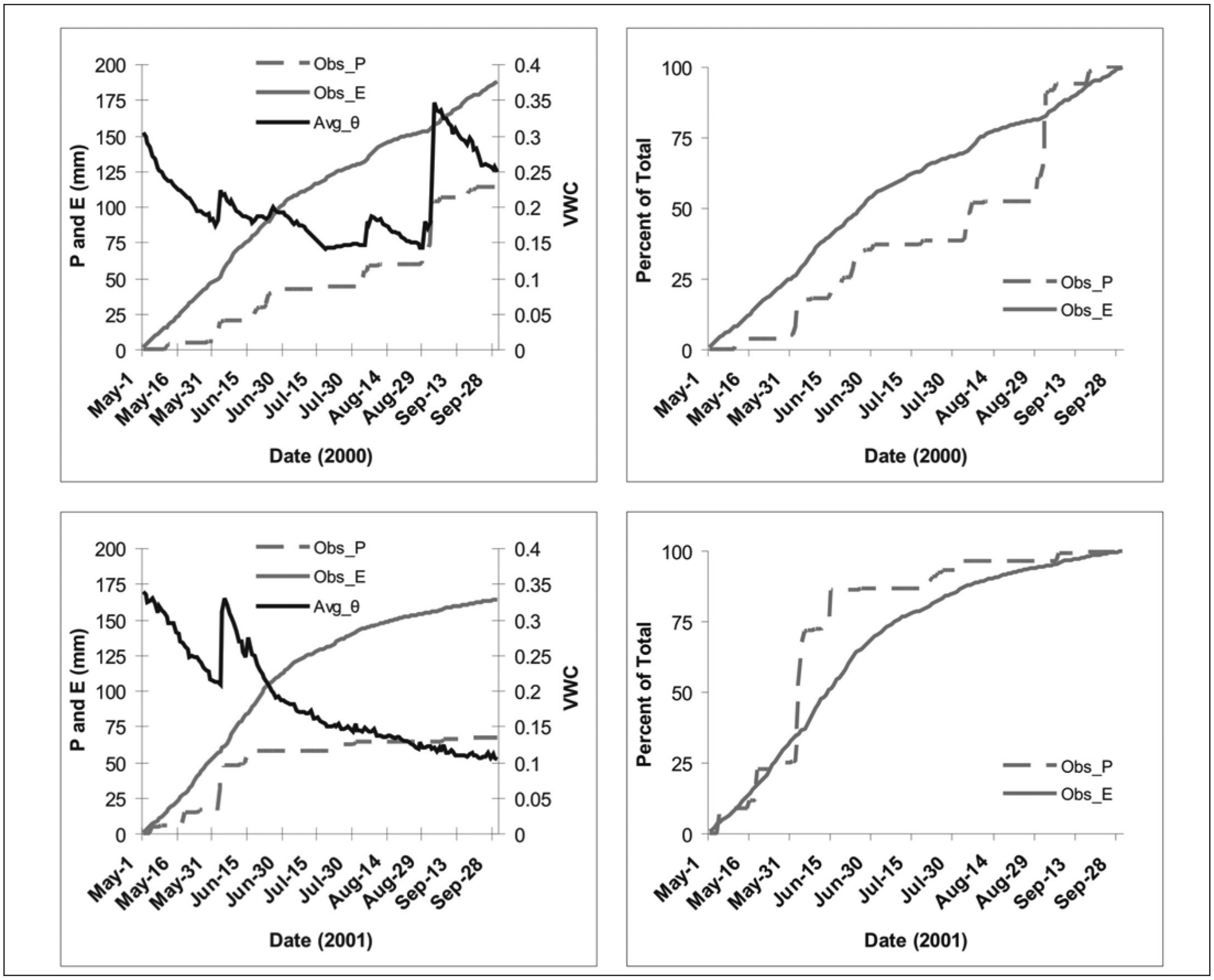

Figure 1. Daily cumulative totals of measured precipitation (Obs_P) and ET (Obs_E), measured volumetric water content of the upper 15 cm profile (Avg_θ), and precipitation and ET as a percentage of the total over the period of May 1–September 30 for 2000 and 2001.

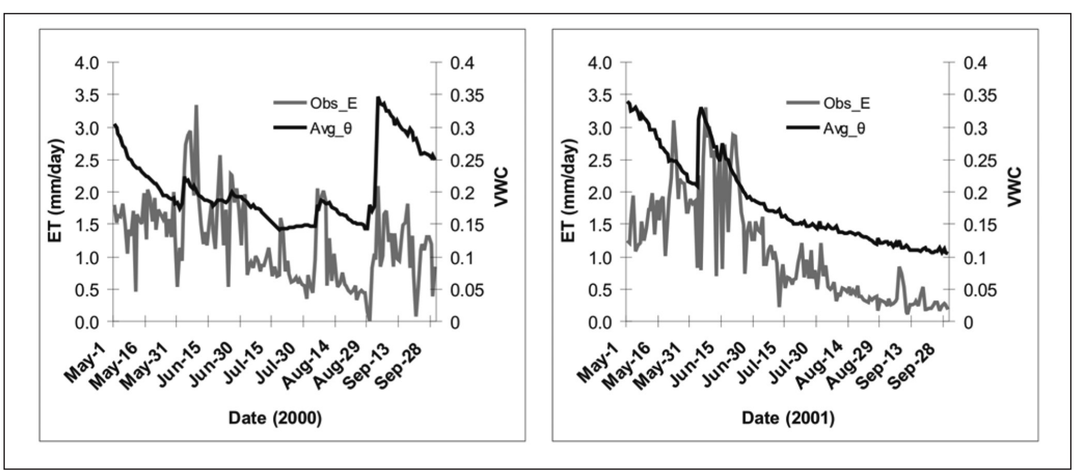

Figure 2. Eddy covariance measured evaporation and near surface soil moisture over 15 cm profile depth at Lethbridge grassland site from May 1–Sept 30 for 2000 and 2001.

specific hydrological models (Pomeroy *et al.*, 2007). Models are created within the CRHM platform by linking physically-based algorithms (modules). The modelling environment is flexible in that users may alter algorithms or create their own. This allows the user to develop a model suited to a specific problem given the available information. It also provides a convenient diagnostic tool for comparing different algorithms designed for the same purpose.

CRHM treats spatial arrangements of elements in a basin as hydrological response units (HRU). Energy and mass balances are applied to an HRU, each of which represents a single biophysical landscape unit with a distinct set of parameters, location in the flow network and driving meteorology (Pomeroy *et al.*, 2007). A landscape unit that does not contribute surface or sub-surface runoff to a particular stream or river is also a valid HRU. This makes the CRHM platform and assembled hydrological models relevant for use with a large portion of the Canadian prairie region.

The model assembled for this application uses near-surface observations of temperature, humidity, wind speed, rainfall, net radiation, and solar radiation to generate the appropriate driving meteorological data. Ecological data include plant heights for determining surface roughness and an estimate of LAI. ET can either be computed with or without soil moisture limitations (where conditions permit such application). When continuity is considered, ET is limited to that water available as interception, depressional storage, near-surface soil moisture or rooting zone soil moisture. For simplicity, depressional storage and interception storage are not considered in this application.

Within CRHM the soil column is considered as the sum of two layers whose total soil moisture content is θ and total maximum soil moisture content is θ_{max}. The upper layer is a recharge zone representing the top soil from which both soil evaporation and transpiration losses can occur. The lower layer represents the maximum extent of the rooting zone which supplies water for transpiration. Any rainfall infiltrating to the soil is supplied to the recharge layer first, and once filled, percolates into the lower layer. Excess water from both soil layers contribute to the ground water and subsurface flows.

Previous treatment of the soil column in CRHM was more conceptual than physical which permitted diagnostic assessments of model behaviour. Recently a more physical component was added which allows for parameterization of the plant available water of specific soil texture classes to be strictly enforced. This is done via look-up table for specific soil textures from combined field observations of depth of available water per meter of soil as reported by Stefferud (1955) and Scherer *et al.* (1996). The upper limit of plant available water is simply the difference between the maximum available holding capacity and the wilting point. The maximum capacity is obtained by specifying the soil column depth (i.e., effective rooting zone depth) and the soil texture. This means that for the clay-loam soils at the Lethbridge site the maximum capacity is 367 mm and wilting point is 150 mm which results in 167 mm of soil moisture available to the plants per 1 m of soil.

Effective infiltration capacities are also treated via a look-up table which is based on field experiments for pastures, crops and forests as summarized in Ayers (1959). Previously the Green-Ampt (Green and Ampt, 1911) algorithm was the major description of infiltration into unfrozen soils but is somewhat limited for use with heavier soils by disregarding the effects of decaying surface materials and plant roots. Any infiltration excess due to either saturated conditions or extreme rainfall events is considered surface runoff and is removed from the HRU.

Estimating Moisture Limited ET

A complete description of the PM and GD model parameterizations can be found in Armstrong *et al.*, (2008). The general equation for the PM model may be written as

$$E = \frac{\Delta \dfrac{(Q_n - Q_g)}{\lambda} + \left(\rho C_p \dfrac{\left(e_a^* - e_a\right)}{r_a} \right)}{\Delta + \gamma \left(1 + \dfrac{r_c}{r_a} \right)} \qquad (1)$$

where Δ is the slope of the saturation vapour pressure curve (kPa °C^{-1}), Q_n and Q_g are net radiation and ground heat flux (W m^{-2}), respectively, λ is the latent heat of vaporisation (kJ kg^{-1}) which can be found as $2501 - 2.361(T)$ where T is in °C. ρ is the air density (kPa), Cp is the specific heat of air (1.005 kJ kg^{-1}), e^*_a

and e_a are the saturated and actual vapour pressures of the air (kPa), γ is the psychometric constant (kPa °C^{-1}), and r_a and r_c are the aerodynamic and canopy resistances (m day^{-1}).

The general equation for the GD method can be written as

$$E = \frac{\Delta G \frac{(Q_n - Q_g)}{\lambda} + \gamma G E_A}{\Delta G + \gamma} \tag{2}$$

where E_A represents the drying power and is calculated using a Dalton-type formulation, $E_A = f(u)(e^*_a - e_a)$, and f(u) is a vapour transfer function derived from extensive field observations (Pomeroy *et al.*, 1997)

$$f(u) = 8.19 + 0.22z_0 + (1.16 + 0.08z_0)u \tag{3}$$

where z_0 is the aerodynamic roughness length (cm) and is a function of vegetation height, and u is the wind speed (m s^{-1}). *G,* the relative evaporation parameter (dimensionless) is given by

$$G = \frac{1}{0.793 + 0.2e^{4.902D}} + 0.006D \tag{4}$$

and is assumed to be related to the relative drying power of the air, *D* (dimensionless), which is a function of the humidity deficit, turbulent transfer and available energy

$$D = \frac{E_A}{E_A + \frac{(Q_n - Q_g)}{\lambda}} \tag{5}$$

The PM and GD models were linked to the infiltration and soil moisture accounting algorithms to estimate the hourly moisture limited actual ET. This moisture limited rate is designated as E_L in CRHM. E_L is calculated as a function of the maximum allowable ET, *E* (calculated from Equation 1 or Equation 2) over each hour and the wetness of the soil column. When soil moisture becomes limiting, restrictions are applied depending on the general soil texture class (sand, loam, and clay). In the case of the Lethbridge site restriction functions for a clay soil class are used.

These simple functions are based on the developments of Zahner (1967) for three general soil classes under forest cover and later modified by Leavesley *et al.* (1983). This approach requires that the soil wetness ratio, R_θ, be calculated. R_θ is simply the ratio of current plant available soil moisture, θ, to maximum available plant water holding capacity of the soil, θ_{max}. The functions show that at certain fractions of available soil water step changes in actual ET occur. Under moist conditions soil moisture tension is low and moisture may be depleted simply at the rate, *E*. For example, when R_θ is above 0.67 (67%) for the clay-loam soils at the Lethbridge site, the calculated maximum allowable actual ET can meet the atmospheric demand and is simply

$$E_L = E \tag{6}$$

where *E* is calculated using Equation 1 or Equation 2. As the fraction of available soil water falls to between $0.67 > R_\theta > 0.33$ (67% and 33%) under drying conditions, moisture tension increases and soil water depletion becomes restricted. The effects of this moisture stress on actual ET can be described as a linear function of the fraction of available water content and is calculated as

$$E_L = R_\theta E \quad . \tag{7}$$

Finally, when the fraction of available soil water becomes critical under severely drying conditions (i.e. drought) and $0.33 > R_\theta$, soil moisture tension increases more rapidly and soil water depletion becomes more severely restricted. At this stage actual ET is greatly reduced and can be found as

$$E_L = 0.5\, R_\theta E \quad . \tag{8}$$

Modelling Assumptions, Derived Variables, and Varying Rooting Depths

The depth of the recharge layer was set to 15 cm (approximately six inches) from which both soil evaporation and ET can occur. This is also the profile depth of available soil moisture measurements at the site. The initial wetness of the entire soil column for all model runs was set to 0.75 or 75% for simplicity since no soil moisture measurements are available to a depth of greater than 15 cm. The initial wetness is considered reasonable for this application since the observations indicate the volumetric soil moisture of the upper 15 cm was fairly similar on May 1 (JD 121) for both 2000 and 2001 (0.31 and 0.34). For a clay-loam soil with a porosity of approximately 0.476, as obtained from the look-up table according to Stefferud (1955) and Scherer *et al.* (1996), this results in a soil wetness of 65% and 71%, respectively. Given that volumetric soil moisture in this upper layer had peaked at 0.37 on JD 104 (2000) and 0.39 on JD 99 (2001), and was reducing slowly prior to JD 121 it is more than likely that the profile had a higher moisture content at depths below the upper shallow layer since surface layers dry more rapidly.

For the PM model a reference minimum value of the resistance is needed for determining evaporative losses under optimal conditions. This value is used for estimating the effective canopy resistance using a multiplicative approach following that introduced by Jarvis (1976). Canopy resistance is also adjusted early in the season taking into consideration the reduced LAI. Under conditions where soil moisture and other contributing environmental factors are non-limiting, a reference value ≈ 50 s m^{-1} is reasonable for grasses and used for all model runs. In the case of the GD model, there are no initial parameters to set and it is driven by meteorological forcing data.

Changes in canopy height were derived based on measurements near mid season reported in the ecological dataset for the Lethbridge Ameriflux site. This was done using a simple linear interpolation from the earliest date of measured minimum LAI to the date of the measured canopy height; taken simply to be the maximum average canopy height. From the heights, LAI was then estimated from Verseghy *et al*. (1993); estimating LAI as function of the current canopy and maximum canopy heights, and measured values of minimum and maximum LAI; also available in the ecological dataset.

For the purpose of examining the effects of rooting depth on cumulative ET several model runs were performed with each one having a different maximum soil column depth (maximum effective rooting depth). For each run the depth was incremented by 0.2 m (200 mm) starting at 0.8 m (800 mm). The upper limit was taken to be approximately 1.4 m (1400 mm). These depths are considered to be reasonable lower and upper limits to which the grasses at Lethbridge might be rooted based on 50 years of research conducted by Weaver (1925; 1926) and others.

Results

Moisture Limited Evapotranspiration During Drought

Results for the modelled cumulative ET from May 1-September 30 for the 800, 1000, 1200, and 1400 mm rooting zone depths are shown in Figure 3 for 2000, and Figure 4 for 2001. From the figures it is clear that the contrasting ET algorithms produce different totals at each depth in both years. For example, Figures 3 and 4 show the cumulative ET total for the PM model is consistently higher than the GD model, and the difference tends to increase with increases in soil depth. Also, the graphs show a logical progression from the models clearly underestimating ET at shallower rooting depths to overestimating ET at deeper rooting depths.

In both 2000 and 2001 the poorest agreement between the modelled and observed cumulative ET was for the 0.8 m rooting depth. By way of comparison, the PM modelled cumulative ET was 86% (-26 mm) of the observed total whilst the GD was 79% (-39 mm) for the period in 2000. In 2001, the PM modelled cumulative ET was 88.6% (-18.6 mm) of the observed total whilst the GD was 83% (-27 mm). The best agreement between the modelled and observed cumulative ET, however, was different for both years. In 2000, the total for PM was nearly equal to the observed at the effective rooting depth of 1.2 m. The ET curve for the 1 m rooting depth appears to be in better agreement with the observed for longer though the cumulative total is only 93% (-12 mm) when compared to the observed. For the GD model the best agreement (-3 mm) was for the 1.4 m rooting depth and the relative differences between the modelled and observed totals over the season appear to simply cancel out (Figure 3); whereas the PM model overestimated the total slightly by 7% (14 mm) for this depth.

The observed ET curve for 2001 presents a somewhat more simple case where cumulative ET increases steadily and then begins to level off around the start of July. In this case, the cumulative ET curve for the PM model agrees surprisingly well over the season for a rooting depth of 1 m and only underestimates

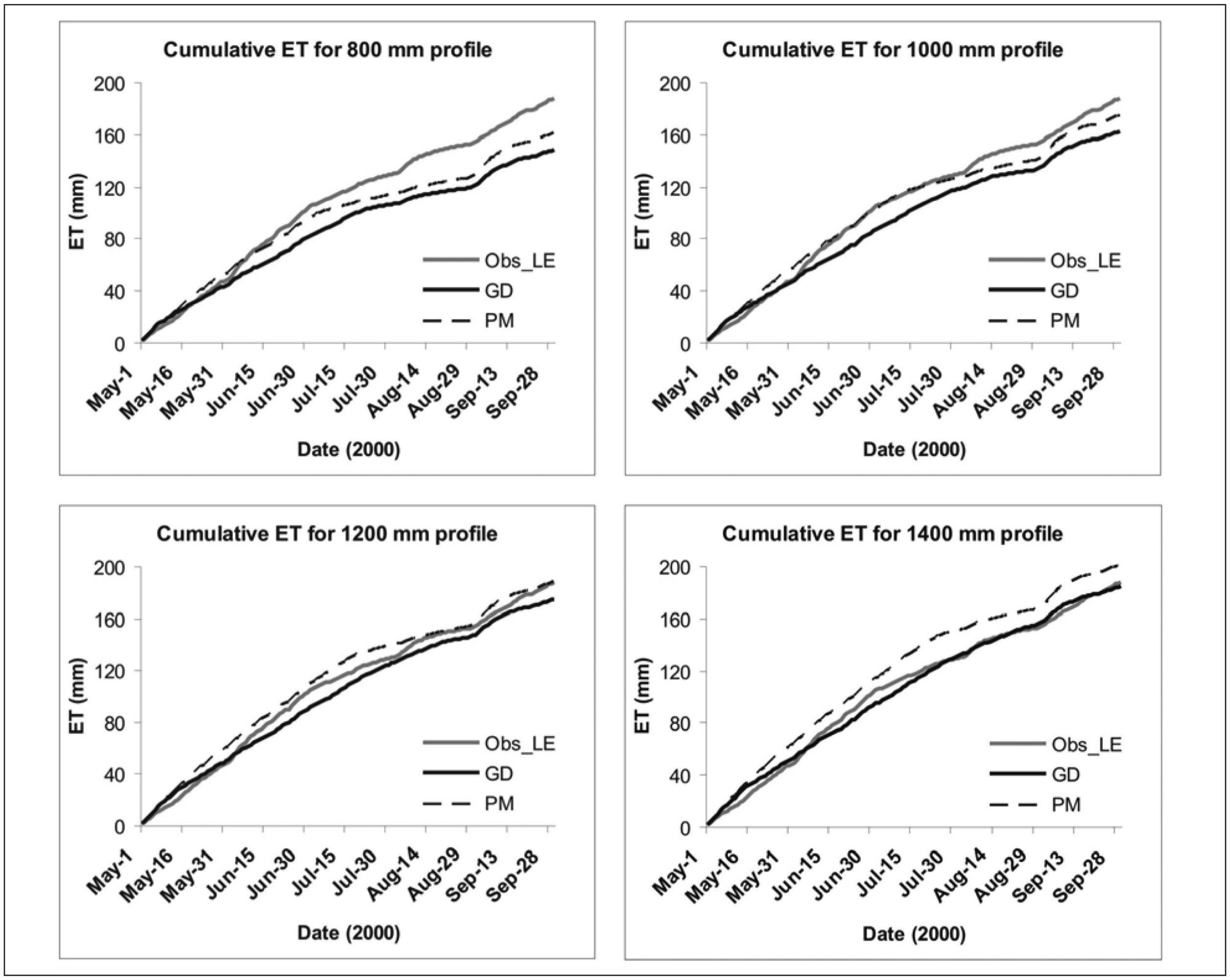

Figure 3. Model results for the 2000 drought period for soil profile depths of 800, 1000, 1200 and 1400 mm.

the observed total by 2.5% (-4 mm); whereas the GD model produces an underestimate of 10% (-15 mm). The GD shows considerable improvement compared to the observed (-3 mm) at a rooting depth of 1.2 m whilst the PM method overestimates slightly by 6% (10 mm). Similar to the results for the rooting depth of 1.4 m for 2000 (Figure 3), the GD cumulative ET deviates abruptly from the observed but still produces a similar final total.

Discussion

In general, reasonable agreement between the modelled and observed cumulative ET could be achieved with both the PM and GD models. In the case of the PM model during drought for the grasses at Lethbridge, setting a rooting depth of 1 m to 1.2 m provided the best results. This appears to be in agreement with the rooting habits of these grass species according to the extensive research of Weaver (1925; 1926) and Weaver and Clements (1938). In the case of the GD model, a depth of 1.2 m to 1.4 m provided the best agreement for the cumulative total although slight deviations between modelled and measured values are observed over the period prior to the final date.

The differences between the cumulative totals of the PM and GD models and apparent lack of a common rooting depth can be attributed to the difference in theoretical approaches. In contrast to the PM model applied here, the GD model does not consider the physical characteristics of plants such as LAI which is very low in early spring and over the period in general. The reduction in LAI increases the

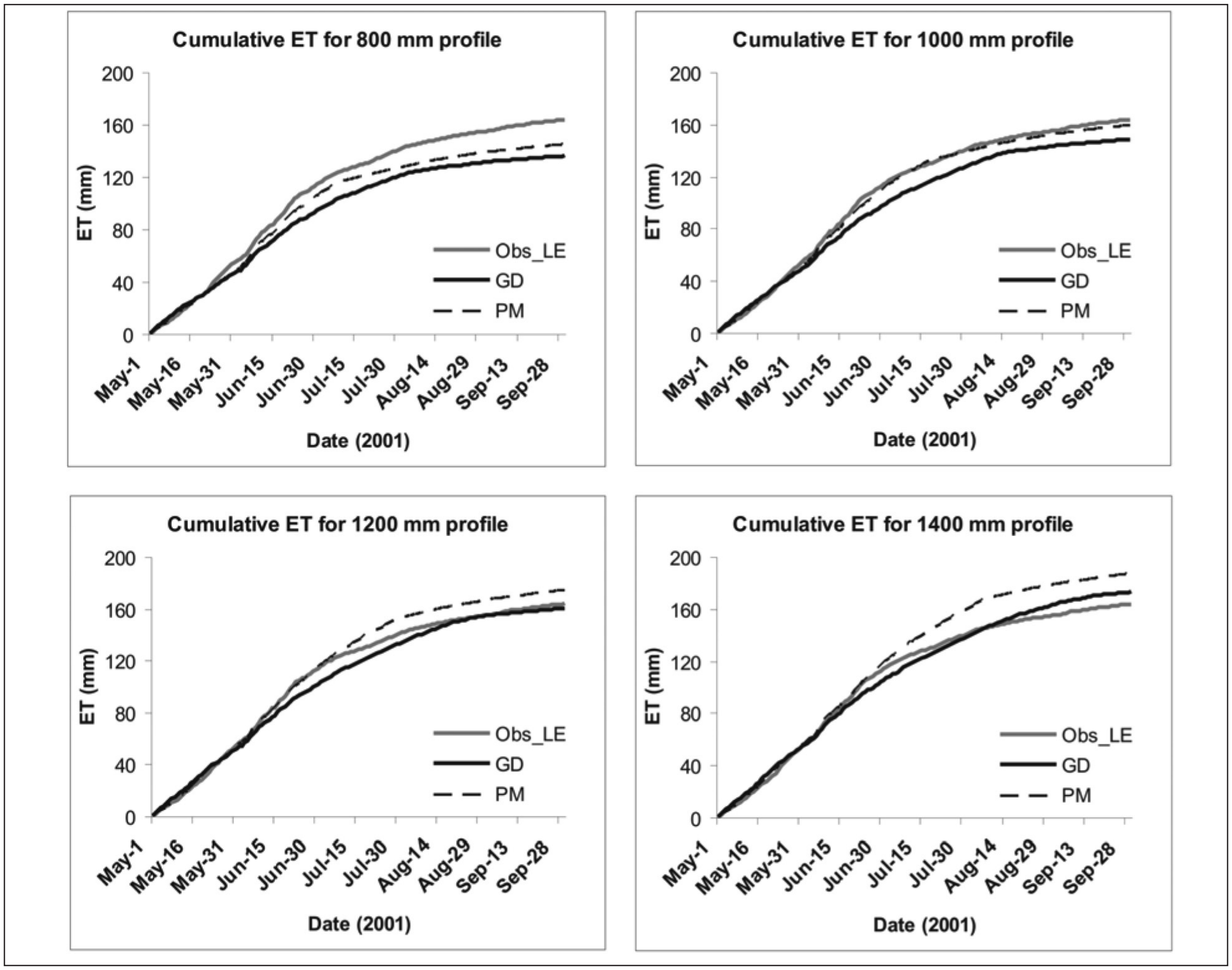

Figure 4. Model results for the 2001 drought period for soil profile depths of 800, 1000, 1200, and 1400 mm.

effective resistance of the PM model which reduces ET estimates and subsequently can deplete the stored moisture less rapidly than does the GD model. Also, the humid deficit is generally high for an extended period due to the drought which results in an increase in the drying power term, thereby impacting the relative evaporation more strongly than the effects of the canopy resistance term does the PM model. As a result of this combination, the GD model produces an underestimate when compared to that of the PM model.

As the rooting depth is increased for subsequent model runs, the initial contents of the stored water also increases which generally results in larger differences between the model totals. This can also be attributed in part to the minimum resistance (50 s m^{-1}) used for the PM model runs. As further soil moisture becomes available, ET increases more appreciably relative to the ET that would result if a larger minimum value were used. In the case of the GD model there is no provision to treat plant phenology and thus overestimates of ET are more likely to occur early in the season under severe drying conditions.

Conclusions

A simple application has been demonstrated that examines the influence of effective rooting depth on actual ET estimates during drought. Two separate hydrological models consisting of theoretically contrasting ET formulations (PM resistance and GD complementary) were assembled using a suite of physically-based model algorithms provided with the CRHM platform. Under conditions when soil moisture is considered to be non-limiting, the actual ET is calculated without restriction. Continuity is enforced to limit actual ET estimates as drying progresses and moisture reserves become depleted further under drought conditions such as experienced at the Lethbridge site in 2000 and 2001.

Both ET models were shown to provide reasonable agreement with the measured cumulative ET values over the course of the growing period May 1–September 30 for 2000 and 2001. The PM model provided good agreement at specified rooting depths of 1 m and 1.2 m for 2000 and 2001, respectively. This would appear to be in agreement with research by Weaver (1925; 1926) and others in the case of the rooting habits of shorter, shallow rooting grasses and mid-height grasses such as wheatgrass under drought conditions. Under more favourable conditions, however, it is likely that the rooting depth of the wheatgrass would be much greater. The GD model provided the best agreement at a rooting depth of 1.2 m in 2000 and 1.4 m.

The results of this simple application suggest the relative complexities of the PM resistance-based approach is generally not a problem with soil moisture accounting to limit actual ET estimates over the growing period examined here. Rather, the initial parameterization of the model is based on observations and the model behaviour generally agrees well with the cumulative ET measurements over the period. In the case of the GD model, a lack of a physical description of vegetation characteristics would make it difficult to suggest physically meaningful improvements to correct noticeable deviations between the modelled and measured values during the period other than adjusting soil depth or the initial soil moisture conditions. As such, the results suggest that for the continuous model simulation presented here, which considers soil moisture continuity there would appear to be no apparent benefit in choosing the complementary model over the resistance based model for estimating cumulative ET over the growing period.

Acknowledgements

Funding support for this study was provided by the Drought Research Initiative (Canadian Foundation for Climate and Atmospheric Sciences) and the Canada Research Chairs programme. The authors wish to extend a special thanks to Larry Flanagan at the University of Lethbridge for providing access to the Ameriflux data for the mixed prairie site. Thanks also to Hugh McLean at the Lethbridge Research Centre for providing meteorological data (incoming solar radiation specifically) and Tom Brown for his continued model development at the Centre for Hydrology, University of Saskatchewan, Saskatoon. The authors would like to acknowledge the Data Access Integration (DAI, see http://quebec.ccsn.ca/DAI/) team for providing the data and technical support. The DAI data download gateway is made possible through collaboration among the Global Environmental and Climate Change Centre (GEC3), the Adaptation and Impacts Research Division (AIRD) of Environment Canada, and the Drought Research Initiative (DRI). The Ouranos Consortium (in Québec) also provides IT support to the DAI team. We acknowledge Environment Canada (EC) for the access of observed station data, retrieved from the National Climate Data and Information Archive, operated and maintained by EC.

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