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Impact of an unstressed canopy conductance on the Bouchet-Morton complementary relationship

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[1] Indirect methods of quantifying evapotranspiration (λE_a) are sought since regional estimations of λE_a require prohibitive instrumentation or highly parameterized and data-intensive land surface models (e.g., involving temporally and spatially varying soil moisture, soil hydraulic properties, and vegetation properties). Complementary relationship (CR) models, based on Bouchet's hypothesis, are one such method of estimating λE_a from routinely measured meteorological variables. Bouchet's CR states that given a change in regional surface moisture availability (M_A), changes in λE_a are reflected in changes in potential evapotranspiration (λE_p), such that $\lambda E_a + \lambda E_p = 2\lambda E_0$, where λE_0 is an assumed equilibrium condition at which $\lambda E_a = \lambda E_p = \lambda E_0$ given sufficiently large M_A. Whereas λE_p conceptually includes a transpiration component, the treatment of vegetation in existing CR applications varies from neglecting it to indirectly accounting for it through recalibration of Penman's empirical wind function. We utilize the First International Land Surface Climatology Field Experiment (FIFE) data set to demonstrate that inclusion of a maximum (i.e., unstressed) canopy conductance ($g_{c,\max}$) in a Penman equation with stability-corrected atmospheric conductance (i.e., replacing the Penman equation with an unstressed Penman–Monteith equation) significantly improves both CR convergence and symmetry. Inclusion of $g_{c,\max}$ results in more accurate λE_a estimates than are found with the Penman equation (using either Monin–Obukhov atmospheric conductance or using empirical wind functions in the literature). The proposed method also performs better than the 1992 advection-aridity CR method, modified to include atmospheric stability effects, which attributes noncomplementarity to horizontal advection and corrects for it by adjusting λE_0 by the magnitude of the potential sensible heat flux ($|H_p|$), found from the surface energy balance of the λE_p calculation. In the proposed method a similar energy balance adjustment occurs naturally because the finite canopy conductance causes an increase of surface temperature and sensible heat flux in the λE_p energy balance calculation. The proposed method is consistent with CR explanations that rely on feedbacks with the boundary layer vapor pressure deficit since the impact of such changes on both λE_p and λE_a would be modulated by stomatal conductance.

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1. Introduction

[2] The accurate quantification and modeling of actual evapotranspiration (hereafter λE_a) is required to estimate and predict hydrologic budgets, climate dynamics, and ecosystem productivity. Unfortunately, due to the difficulty in directly measuring λE_a, there exist a limited number of point-based λE_a measurements worldwide, resulting in the pursuit of methods of estimating λE_a from more routinely measured meteorological variables. Complementary relationship (CR) models are one such method of indirectly estimating λE_a. They are based on *Bouchet*'s [1963] hypothesis that λE_a and potential evapotranspiration (hereafter λE_p) are codependent in such a way that they respond to changes in moisture availability (hereafter M_A) with equal but opposite changes in flux, i.e.,

$$\lambda E_p + \lambda E_a = 2\lambda E_0. \tag{1}$$

In equation (1) λE_0 is an evapotranspiration rate at a highly sufficient M_A at which λE_p and λE_a converge (such that $\lambda E_p = \lambda E_a = \lambda E_0$).

[3] *Brutsaert and Stricker*'s [1979] advection-aridity (AA) model and *Morton*'s [1983] complementary relationship areal evaporation (CRAE) model are the two primary working interpretations of equation (1) to indirectly estimate λE_a. In both the AA and CRAE models, λE_p estimates are founded on the *Penman* [1948] combination equation (hereafter $\lambda E_{p,Pen}$), in which Penman developed an empirical

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'wind function' to account for atmospheric conductivity for latent heat flux. *Parlange and Katul* [1992] replaced this empirical wind function in the original AA model with *Brutsaert*'s [1982] turbulent profile based atmospheric conductance term. In this paper, we focus on *Parlange and Katul*'s [1992] revisions of the AA model, as the excessive use of fitted parameterization in the original AA and also in *Morton*'s [1983] CRAE formulation leave them more site-specific and make Bouchet's hypothesis more difficult to test and interpret theoretically. Our study therefore utilizes *Parlange and Katul*'s [1992] modified AA model as a physical framework upon which to investigate the role of canopy conductance in the CR framework for land surfaces dominated by plant transpiration.

[4] We hypothesize that in areas where plant transpiration is a major constituent of λE_a, λE_a and $\lambda E_{p,Pen}$ cannot physically converge (as required by CR) due to the additional conductance regulating λE_a, i.e., the canopy conductance (g_c). This nonconvergence problem can be solved by replacing $\lambda E_{p,Pen}$ with a Penman–Monteith [*Penman*, 1948; *Monteith*, 1965] flux in which the canopy conductance is held constant at a "potential," unstressed value (hereafter $g_{c,\max}$). Figure 1 illustrates how the lack of $g_{c,\max}$ in $\lambda E_{p,Pen}$ would conceptually result in an overestimation of actual λE_p, and thus the failed convergence of λE_p and λE_a at sufficiently high M_A. Note that the stressed g_c, represented schematically in the lower (λE_a) curve is presumed to be some unknown function of moisture availability, vapor pressure deficit (VPD), and other environmental and/or physiological factors [*Ball et al.*, 1987; *Jarvis*, 1976], while the unstressed $g_{c,\max}$ in the middle curve is the (single) constant value of g_c evaluated (conceptually) under conditions principally characterized by unlimited M_A and zero VPD stress.

[5] While Penman-type empirical wind functions can be calibrated to correct for the $\lambda E_{p,Pen}$ overestimation of true λE_p, such calibrations serve no direct use in theoretical investigations of the role of canopy conductance in the CR hypothesis. For example, recent long-term, large-scale implementations of CR have found *Penman*'s [1948] empirical wind function, and recalibrations thereof, to yield relatively accurate indirect λE_a estimates using the original AA model [*Hobbins et al.*, 2001a; *Ramírez et al.*, 2005]. While these studies have obvious merit, their reliance upon empirical wind functions do not help advance CR away from a heuristic hypothesis. The purpose of this study is not to challenge the performance of Penman-type empirical wind functions, but rather to investigate the role of canopy conductance in CR.

[6] *Parlange and Katul* [1992] proposed that the noncomplementarity of estimated potential evaporation with actual evaporation (of similar nature to that shown in Figure 1, i.e., λE_p too large and steep to be complementary to λE_a), as measured in a bare soil field experiment at the Campbell Tract at the University of California, Davis, could be attributed to the horizontal advection of dry air. They argued that this would increase $\lambda E_{p,Pen}$ by creating a situation with a negative Bowen ratio (i.e., net radiation and sensible heat flux both act as a source of energy for potential evaporation), while leaving the *Priestley and Taylor* [1972] estimate of λE_0 unaffected. Therefore, in the case where the surface energy balance of $\lambda E_{p,Pen}$ results

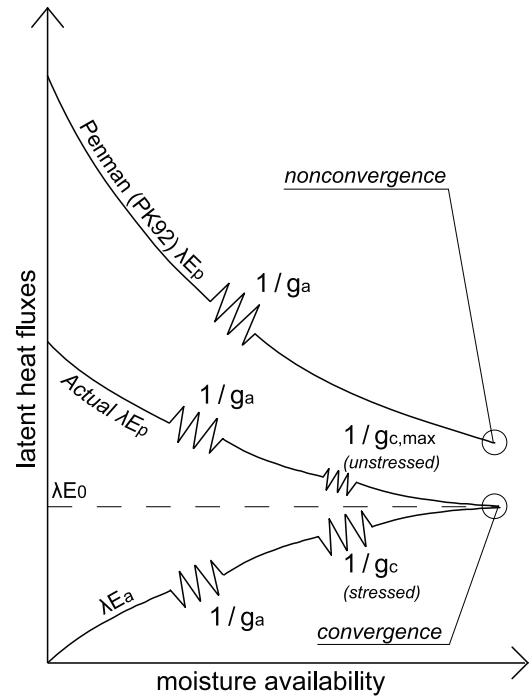

Figure 1. Conceptual role of stomatal conductance in the complementary relationship. Note that the *Penman* [1948]–*Brutsaert* [1982] estimate of potential evaporation is too large to be complementary with actual evapotranspiration (λE_a) but that the addition of a stomatal conductance ($g_{c,\max}$) in series with the atmospheric conductance (g_a) improves both the dry end symmetry and wet end closure of the CR.

in a 'potential' sensible heat flux (H_p) < 0, *Parlange and Katul* [1992] suggest that noncomplementarity can be corrected for by adding the absolute value of the (negative) H_p to λE_0 in equation (1), such that $\lambda E_p + \lambda E_a = 2(\lambda E_0 + |H_p|)$. In this study, therefore, we evaluate the two following versions of *Parlange and Katul*'s [1992] modified AA model: (1) no H_p correction to λE_0 (hereafter PK92v1), and (2) H_p correction to λE_0, when $H_p < 0$ (hereafter PK92v2).

[7] In summary, we demonstrate how the replacement of $\lambda E_{p,Pen}$ with an unstressed Penman–Monteith flux (constant $g_{c,\max}$) in the modified AA method (PK92v1) results in *Bouchet*'s [1963] assumed convergence of λE_a and λE_p at the wet end of the CR relationship and also better symmetry (i.e., better complementarity) at the dry end. The proposed method performs better than *Parlange and Katul*'s [1992] modified AA model for the data set analyzed, and maintains the attractive simplicity of CR, i.e., quantifying λE_a without estimating complex land surface parameter responses to nonpotential (e.g., water limited) evapotranspiration conditions. Further, our results yield insight into the role of canopy conductance in regulating CR for vegetated regions.

2. Theory

2.1. Derivation of CR

[8] *Bouchet*'s [1963] derivation of equation (1) is based on the assumption that actual evapotranspiration (λE_a) and potential evapotranspiration (λE_p) converge at an equilibrium

environmental state characterized by sufficiently high M_A. This convergence point is referred to as an equilibrium evapotranspiration (λE_0) flux rate. Upon departure from this convergence point with a decrease in M_A, a change in λE_a w.r.t. a change in M_A is proportional, but opposite in sign to a change in λE_p w.r.t. the same change in M_A, such that we can define a constant "Φ" as

$$\Phi \equiv -\frac{\partial \lambda E_a}{\partial M_A} \Big/ \frac{\partial \lambda E_p}{\partial M_A}. \tag{2}$$

Next, equation (2) is separated and integrated to yield

$$\Phi \lambda E_p = -\lambda E_a + c_1. \tag{3}$$

Assuming the equilibrium convergence point at which $\lambda E_p = \lambda E_a = \lambda E_0$, equation (3) can be rewritten as

$$\Phi \lambda E_0 = -\lambda E_0 + c_1. \tag{4}$$

Solving equation (4) for c_1 and substituting into equation (3) yields the following more general form of CR:

$$\Phi \lambda E_p + \lambda E_a = (1 + \Phi)\lambda E_0. \tag{5}$$

By assuming perfect symmetry (and thus $\Phi = 1$), equation (5) takes the form of equation (1). *Bouchet* [1963] assumed $\Phi = 1$ in his derivation of equation (1), i.e., given a decrease in λE_a below λE_0, a quantity of energy (c_1 in equations (3) and (4)) directly increases λE_p. While he speculated that certain climatic conditions may yield asymmetry ($\lambda E_a + \lambda E_p \leq 2\ \lambda E_0$), equation (1) and the corresponding assumption of perfect symmetry ($\Phi = 1$) are standard expressions of *Bouchet*'s [1963] heuristic hypothesis [*Brutsaert*, 2005].

[9] *Granger* [1989] derived Φ to be equal to γ/Δ, where γ is the psychrometric constant and Δ the slope of the saturation vapor pressure curve, by assuming all three fluxes in equation (5) take the form of vapor pressure gradients resisted by a diffusivity parameterized by a common generalized wind speed function, f(u). Additionally, by assuming that the Bowen ratio (β, the ratio of sensible heat flux to latent heat flux) is equal to negative one as λE_a converges to λE_0, *Granger* [1989] derived the following alternate expression for CR:

$$\frac{\gamma}{\Delta}\lambda E_p + \lambda E_a = \left(1 + \frac{\gamma}{\Delta}\right)\lambda E_0. \tag{6}$$

Granger's [1989] derivation implicitly accounts for asymmetry of CR for cases where $\gamma/\Delta \neq 1$. It should be noted that *Granger*'s [1989] derivation of equation (6) does not directly take into account transpiration processes and thus the addition of a scaled canopy conductance acting on λE_a. In this derivation, Granger also assumed that the vapor pressure (and air temperature) at reference height does not change as λE_a decreases below λE_0. In Granger's derivation, complementarity results from an increase in surface temperature (as M_A and λE_a decrease) inducing an increased vapor pressure gradient from a saturated surface and thus an increased λE_p. Interestingly, *Szilagyi* [2001] made an almost opposite assumption to derive a complementary relationship. Specifically, he assumed that the surface temperature of the evaporimeter (e.g., an evaporation pan) remains constant, but that the humidity of the air decreases as M_A and λE_a decrease, again increasing the humidity gradient over the pan and increasing λE_p, but for a mechanistically different reason. Neither of these sets of assumptions is likely to be found generally true, though both appear to validate *Bouchet*'s [1963] hypothesis. With such contradictory explanations for CR, it not surprising that *Brutsaert* [2005, p. 138] states, "Although the complementary approach shows specific promise as a practical tool, it still awaits a definitive physical analysis to make it fully effective."

2.2. Advection-Aridity Model Parameters

[10] *Brutsaert and Stricker*'s [1979] advection-aridity (AA) model estimates λE_0 using the *Priestley and Taylor* [1972] partial equilibrium flux ($\lambda E_{0,PT}$):

$$\lambda E_{0,PT} = \alpha \frac{\Delta}{\Delta + \gamma} Q_n. \tag{7}$$

Q_n in equation (7) is the net radiation (R_n) minus the soil heat flux (G). Further, α in equation (8) is a constant whose suggested value has been found to range from 1.05 to 1.32 [see, e.g., *Hobbins et al.*, 2001b].

[11] *Brutsaert and Stricker*'s [1979] original AA model estimates λE_p using the *Penman* [1948] combination equation ($\lambda E_{p,Pen}$):

$$\lambda E_{p,Pen} = \frac{\Delta}{\Delta + \gamma} Q_n + \frac{\gamma}{\Delta + \gamma} E_{a,DP}. \tag{8}$$

In equation (8), $E_{a,DP}$ has the general form:

$$E_{a,DP} = f(U_2)(e_a^* - e_a). \tag{9}$$

In equation (9), f(U_2) is some function of the mean wind speed U_2 at a reference level (i.e., two meters above the ground surface) and e_a^* and e_a are the saturated and actual vapor pressure of the air, respectively. *Brutsaert and Stricker* [1979] original AA model uses *Penman*'s [1948] empirical wind function in equation (9), the general form for which can be expressed as

$$f(U_2) = a(b + cU_2). \tag{10}$$

In equation (10), a, b, and c are empirical constants whose values are calibrated to specific land surfaces, vegetation type and cover, climate, season, etc. Penman's original [1948] wind function (calibrated for open water) is commonly used in CR evaluations [*Brutsaert and Stricker*, 1979; *Hobbins et al.*, 2001a], in which the constants a = 0.0026, b = 1.0, and c = 0.54 are used, given vapor pressure and $E_{a,DP}$ units in equation (9) of Pa and mm day^{-1}, respectively. *Brutsaert and Stricker* [1979] also evaluated *Penman*'s [1954, 1956] "Rome" wind function, in which the calibration coefficient b = 0.5 is used to account for an aerodynamic roughness parameter (z_o) approximately equal to 0.5 mm [*Thom and Oliver*, 1977]. *Parlange and Katul* [1992] replaced equations (9) and (10) with *Brutsaert*'s

[1982] "drying power of the air" ($E_{a,DP}$), which can be expressed as:

$$E_{a,DP} = g_a \rho_a \lambda \left(q_a^* - q_a\right). \tag{11}$$

In equation (11), q_a^* is the saturated specific humidity of the air at air temperature (T_a), q_a is the specific humidity of the air, λ is the latent heat of vaporization, ρ_a is the density of the air, and g_a, the aerodynamic conductance for latent heat flux, takes the following, less empirical form, incorporating *Monin and Obukhov* [1954] surface similarity theory [*Brutsaert*, 1982]:

$$g_a = \kappa u_* \left[\ln\left(\frac{z - d_0}{z_{0v}}\right) - \psi_v\left(\frac{z - d_0}{L}\right)\right]^{-1}. \tag{12}$$

In equation (12), z is the measurement height, d_0 is the displacement height, κ is the von Karman constant (equal to 0.41), u_* is the friction velocity, $\psi_v()$ denotes the stability correction function for water vapor, obtained from the *Businger* [1988] and *Dyer* [1974] formulations, and L is the Obukhov length defined by

$$L = \frac{-u_*^3}{\kappa g \left[\frac{H_v}{\rho_a c_p T_a}\right]}. \tag{13}$$

In equation (13), g is acceleration due to gravity, H_v is the virtual sensible heat flux ($H_v = H + 0.61 T_a c_p E_a$), and c_p is the dry air isobaric specific heat. Further, z_{0v} in equation (12) is the scalar roughness length for water vapor, whose relation to the scalar roughness length for momentum, z_{0m}, is often symbolized as $kB^{-1}|_v = \ln(z_{0m}/z_{0v})$ and is typically assumed to be equal to the relationship between scalar roughness lengths for momentum and sensible heat (z_{0h}), i.e., $kB^{-1}|_h = \ln(z_{0m}/z_{0h})$. *Brutsaert* [1982] demonstrates that, whereas kB^{-1} for surfaces with bluff roughness elements may vary considerably, in the case of admixture in "densely placed permeable roughness obstacles" (p. 109), kB^{-1} is much less variable. Therefore, as the purpose of this study is not to calibrate kB^{-1} equations but to test the sensitivity of CR to canopy conductance, we have assumed a $kB^{-1}|_v = 2$, per *Brutsaert*'s [1982] recommendation of an acceptable, first approximation for practical applications. *Qualls and Brutsaert* [1996] applied *Brutsaert*'s [1979] $\ln(z_{0h})$ formulation for turbulent admixture in a dense, uniform vegetation canopy to FIFE, and derived an empirical relationship, i.e., $\ln(z_{0h}) = -7.02/LAI - 1.79$, using a representative vegetation height (z_{veg}) = 0.4 m for FIFE, where LAI is the leaf area index. Assuming $z_{0h} = z_{0v}$, evaluations of *Qualls and Brutsaert*'s [1996] empirical formulation for $\ln(z_{0h})$ using measured leaf area index data from FIFE yield $kB^{-1}|_v$ values in the range of 1.5 to 5.

[12] As discussed in the introduction, *Parlange and Katul* [1992] suggested that in some cases the noncomplementarity between λE_a and $\lambda E_{p,Pen}$ may be due to overestimation of the potential evaporation due to the horizontal advection of dry air, increasing both λE_a and $\lambda E_{p,Pen}$, while leaving $\lambda E_{0,PT}$ unaffected. Specifically, when the surface energy balance implicit in the calculation of $\lambda E_{p,Pen}$ results in $H_p < 0$, the PK92v2 model holds that the CR relationship need to be corrected by adding the absolute value of the negative H_p to λE_0 in equation (1), such that

$$\lambda E_{p,Pen} + \lambda E_a = \left\{\begin{array}{ll} 2\left(\lambda E_{0,PT} + \left|Q_n - \lambda E_{p,Pen}\right|\right) & \text{if}\left(Q_n - \lambda E_{p,Pen}\right) < 0 \\ 2\lambda E_{0,PT} & \text{if}\left(Q_n - \lambda E_{p,Pen}\right) \geq 0 \end{array}\right\}. \tag{14}$$

2.3. Accounting for an Unstressed Canopy Conductance in $\lambda E_{p,Pen}$

[13] Whereas λE_p conceptually includes a transpiration component [*Thornthwaite*, 1948], Penman-type combination equations have been used to estimate λE_p in CR tests, e.g., using equations (9)–(12) without account of stomatal conductance (and thus implicitly representing open water, wet soil, or wet vegetation canopies). For example, *Parlange and Katul* [1992] applied such a formulation over a bare soil lysimeter (which by definition has no stomatal resistance), but *Crago and Crowley* [2005] applied the same formulation over four partially vegetated sites, and *Hobbins et al.* [2001a, 2001b] applied equation (10) to long-term water balances for numerous, minimally impacted basins across the conterminous United States. It is well established, however, that stomatal conductance, and the respective g_c, prohibits plant transpiration even when soil and atmospheric conditions are at optimum levels [*Jarvis*, 1976]. Further, as the concept of λE_p is founded upon unstressed (e.g., water unlimited) conditions, a potential (unstressed) g_c can be conceptualized which reflects the decrease in physiological stress due to increased root zone water availability for transpiration, the absence of VPD stress, and other environmental and/or physiological factors. In a Jarvis-type stomatal resistance model, this potential canopy conductance is referred to as "$g_{c,\max}$," reduced through a multiplicative series of environmental stress functions [*Jarvis*, 1976].

2.4. Objectives

[14] In this paper we utilize 147 days (July and August 1987–1989) of daily averaged flux data at the First International Land Surface Climatology Field Experiment (FIFE) to explore how the absence of a $g_{c,\max}$ in $\lambda E_{p,Pen}$ results in an overestimation of λE_p, thereby violating the assumptions of equilibrium convergence and symmetry between λE_a and λE_p manifested in equation (1). The resulting noncovergence and asymmetry of $\lambda E_{p,Pen}$ using $E_{a,DP}$ in equation (11) will result in poor estimates of λE_a when evaluating CR in areas where plant transpiration is a major evapotranspiration constituent. We next demonstrate how the addition of a $g_{c,\max}$ to $\lambda E_{p,Pen}$ (essentially replacing the $\lambda E_{p,Pen}$ in the AA method with a Penman–Monteith λE_p characteristic of unstressed, potential conditions) results in both convergence of λE_a and λE_p when soil moisture is not limiting, and better symmetry when soil moisture is limiting. Both of these improvements to *Parlange and Katul*'s [1992] modified AA model (PK92v1 and PK92v2) yield more precise λE_a estimates, maintain the attractive simplicity of CR, and add useful insight into previously unexplained error in CR evaluations. Finally, we show that *Penman*'s [1948] wind function (i.e., calculated with coefficients a = 0.0026, b = 1 and c = 0.54), even though

it was calibrated for open water, predicts lower potential evaporation rates than are predicted using equation (12), and thus yields better CR estimates of λE_a than *Parlange and Katul*'s [1992] modified AA (with or without H_p correction), although still far less precise than the proposed Penman–Monteith method.

3. Methods

3.1. FIFE Data

[15] The FIFE experiment was an intensive, land-surface-atmosphere experiment conducted on the Konza prairie near Manhattan, Kansas in the late 1980s, the details of which are described by *Sellers et al.* [1992]. The FIFE site is predominately grassland of moderate topography. Ten portable automatic meteorological (PAM) stations were used to collect surface data from 1 May 1987 to 10 November 1989. The dominant C_4 grassland species include *Andropogon gerardii* (big bluestem), *Sorghastrum nutans* (Indiangrass), and *Panicum virgatum* (switchgrass) [*Kim and Verma*, 1991]. We first evaluated CR over 147 days at FIFE days using *Parlange and Katul*'s [1992] modified AA method in which $\lambda E_{p,Pen}$ and $\lambda E_{0,PT}$ (α = 1.26) fluxes were calculated at a 30-min frequency between 0800 and 1800, July–August 1987–1989. Next, $\lambda E_{p,Pen}$ was calculated at each interval using the *Penman*'s [1948] empirical wind function for open water (i.e., a = 0.0026, b = 1.0, and c = 0.54) and also his later [*Penman*, 1954, 1956] "Rome" wind function (i.e., a = 0.0026, b = 0.50 and c = 0.54). The resultant $\lambda E_{p,Pen}$ and $\lambda E_{0,PT}$ fluxes were then averaged for each day. Following *Brutsaert* [1982], we estimated u_* and L in equations (12) and (13) using an iterative solution in which u_* and L are corrected until measured horizontal wind speed (U) at FIFE (z = 5.4 m) converges with the calculated wind speed using the following *Monin and Obukhov* [1954] surface similarity profile equation [*Brutsaert*, 1982]:

$$U = \frac{u_*}{\kappa}\left[\ln\left(\frac{z - z_d}{z_{0m}}\right) - \psi_m\left(\frac{z - d_0}{L}\right)\right]. \tag{15}$$

In equation (15), $\psi_m()$ denotes the stability correction function for momentum, obtained from the *Businger* [1988] and *Dyer* [1974] formulations, and z_{0m} is assumed equal to $0.1z_{veg}$ [*Dingman*, 2002].

3.2. An Unstressed Penman-Monteith λE_p Estimate

[16] The original *Penman* [1948] and *Monteith* [1965] equation (hereafter $\lambda E_{a,PM}$) is advantageous in that unlike the empirical Penman wind function, it incorporates a canopy conductance (g_c) in conjunction with g_a, i.e.:

$$\lambda E_{a,PM} = \frac{\Delta Q_n + \rho_a c_p g_a\left(e_a^* - e_a\right)}{\Delta + \gamma(1 + g_a/g_c)}. \tag{16}$$

In equation (16), e_a^* is the saturated vapor pressure of the air, and e_a is the vapor pressure of the air. As $\lambda E_{a,PM}$ in the form of equation (15) is a function of a stressed g_c, $\lambda E_{a,PM}$ represents the stressed λE_a function in Figure 1.

[17] We propose that, for use in CR, a more appropriate estimate of λE_p [cf. *Thornthwaite*, 1948] than the Penman wind function based model can be derived by replacing g_c in equation (15) with the potential, maximum (unstressed) $g_{c,\max}$:

$$\lambda E_{p,PM} = \frac{\Delta Q_n + \rho_a c_p g_a\left(e_a^* - e_a\right)}{\Delta + \gamma\left(1 + g_a/g_{c,\max}\right)}. \tag{17}$$

As such, the flux in equation (17) is an unstressed Penman–Monteith λE_p estimate (hereafter $\lambda E_{p,PM}$) which we propose should replace $\lambda E_{p,Pen}$ in PK92v1 for vegetated land surfaces. The Penman–Monteith equation (as opposed to methods that recalibrate the wind function) allows for a more theoretical investigation of the role of stomata in CR, particularly in vegetated land surfaces.

[18] In addition to calculating $\lambda E_{p,Pen}$ (using E_a from both equations (9) and (11)) and $\lambda E_{0,PT}$ (α = 1.26) at each 30-min interval over our 147 day study period at FIFE, we evaluated $\lambda E_{p,PM}$ with incremental $g_{c,\max}$ values (e.g., 1/50, 1/100, 1/250, 1/500 m s^{-1}) to both test the overall concept of an unstressed $\lambda E_{p,PM}$ in the CR framework and to estimate the $g_{c,\max}$ that produces λE_p estimates whose relationship to λE_a satisfies the assumptions of convergence and symmetry (see equation (1)). Further, to evaluate PK92v2, H_p was calculated from Q_n measured data and calculated $\lambda E_{p,Pen}$. Equation (14) was then evaluated using $\lambda E_{0,PT}$ (α = 1.26).

[19] All flux and meteorological data were taken from the Betts' site-averaged flux and AMS data sets, available online at Oak Ridge National Laboratory National Archive Center, Oak Ridge, Tennessee (http://www.daac.ornl.gov). Further, we utilized the Betts site-averaged gravimetric soil moisture data set (5–10 cm depth) as an indicator of M_A [*Betts and Ball*, 1998]. To construct standard CR plots (Figure 1), λE_a, $\lambda E_{0,PT}$, $\lambda E_{p,Pen}$ and $\lambda E_{p,PM}$ daily averaged fluxes were binned into 2% gravimetric soil moisture groups; bin means and standard deviations were then calculated.

4. Results and Discussion

[20] Figure 2a illustrates the results of the modified advection-aridity CR methods over 147 days in July–August 1987–1989 at FIFE. While a CR-type relationship between λE_a and $\lambda E_{p,Pen}$ (using $E_{a,DP}$ from equation (11)) exists, two fundamental violations of *Bouchet*'s [1963] assumptions are apparent. First, the modified AA $\lambda E_{p,Pen}$ and λE_a fail to converge at high M_A conditions, an assumption central to the derivation of equation (1). Second, the use of *Parlange and Katul*'s [1992] $\lambda E_{p,Pen}$ results in a dramatically asymmetrical CR relationship. Both of these deviations from *Bouchet*'s [1963] assumptions have significant impact on the precision of the CR method to accurately estimate λE_a indirectly from λE_p and λE_0 estimations (discussed below). Further, the PK92v1 $\lambda E_{0,PT}$ with an α = 1.26 (uncorrected for the H_p shown in Figure 2b) appears to slightly overestimate the actual λE_0, as demonstrated in Figure 2a. Figure 2b illustrates the PK92v2 H_p, the magnitude of which is added to $\lambda E_{0,PT}$ when $H_p < 0$ (see equation (14)). The impact of the H_p correction on estimated evaporation is discussed below, after the role of vegetation is explored.

[21] Figure 3 illustrates the resultant $\lambda E_{p,PM}$ fluxes with contour labels indicating the constant (single value) $g_{c,\max}$

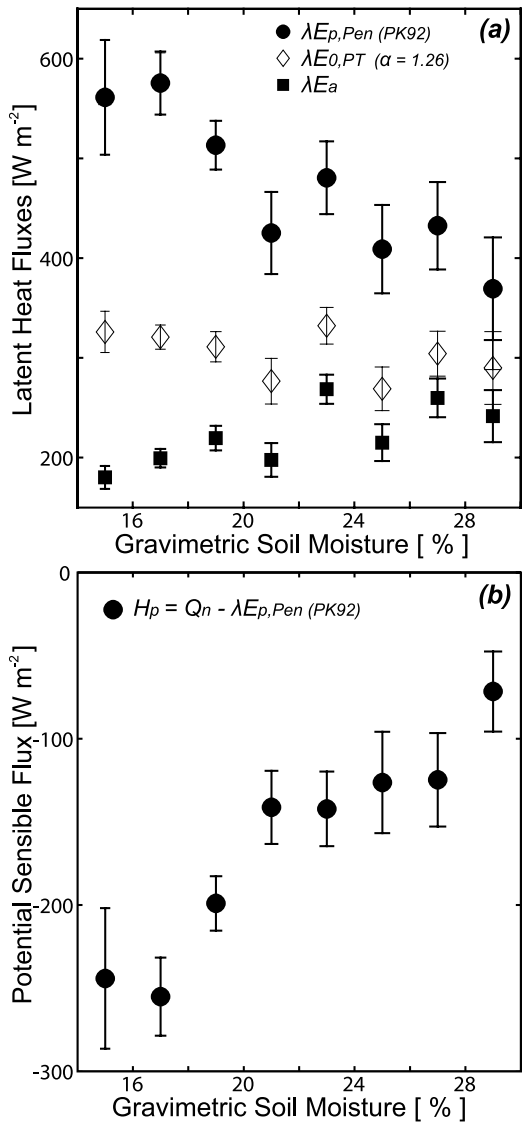

Figure 2. (a) Complementary relationship results at FIFE for July–August 1987–1989 study period (147 days included) using binned daily (0800–1800 CSDT) averaged fluxes. Error bars represent one standard deviation about the bin mean. (b) Binned daily averages of H_p calculated from $\lambda E_{p,Pen}$ per *Parlange and Katul* [1992], the magnitudes of which are evaluated in equation (14) when $H_p < 0$.

in equation (16). Our results indicate that the lack of observed convergence between the $\lambda E_{p,Pen}$ open water flux and λE_a at high moisture conditions can be attributed to the lack of a $g_{c,\max}$ that would resist vapor diffusion and lower λE_p, as consistent with *Thornthwaite* [1948]. Further, the problem of CR asymmetry ($\Phi \neq 1$) is improved upon when replacing $\lambda E_{p,Pen}$ in the modified AA framework (PK92v1) with $\lambda E_{p,PM}$ evaluated with a constant $g_{c,\max}$. That is, including a $g_{c,\max}$ both shifts λE_p downward and also alters its slope w.r.t. M_A, especially at the dry end. As further demonstrated in Figure 3, the *Penman* [1948] wind function (diamond symbols) appears to overcorrect for this effect, yielding convergence at the wet end but lack of complementary at the dry end. The resulting $\lambda E_{p,Pen}$ fluxes using

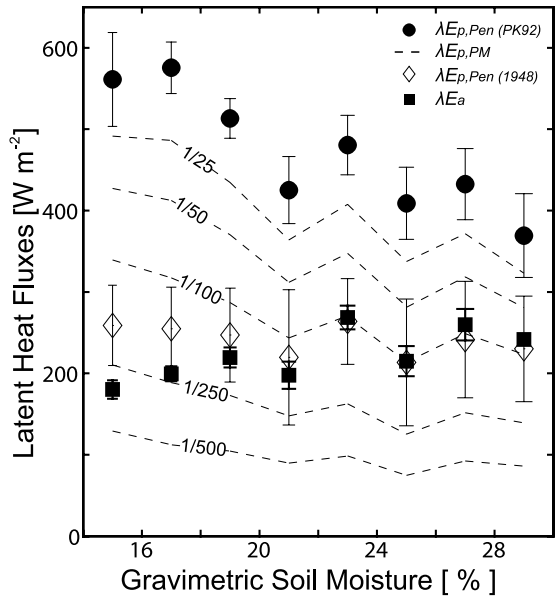

Figure 3. Complementary relationship plot study period at FIFE with contours of the modeled flux results with incremental $g_{c,\max}$ values in $\lambda E_{p,PM}$. Note that a $g_{c,\max}$ = 1/80 m s^{-1} in $\lambda E_{p,PM}$ results in best convergence and symmetry.

Penman's [1956] "Rome" wind function were indistinguishable from those using *Penman*'s [1948] f(U_2), similar to the comparison made by *Brutsaert and Stricker* [1979].

[22] Figure 4 illustrates how, in contrast to Figure 2a, an $\lambda E_{p,PM}$ with a $g_{c,\max}$ of 1/80 m s^{-1} (best fit from Figure 3) results in a CR relationship characteristic of more complete convergence of λE_p and λE_a at high M_A conditions. Additionally, evaluation of $\lambda E_{0,PT}$ with an α = 1.10, as

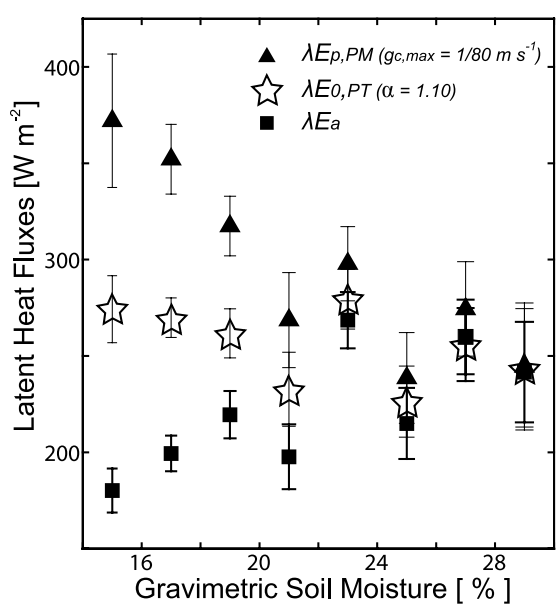

Figure 4. Complementary relationship results when using $\lambda E_{p,PM}$ with a $g_{c,\max}$ = 1/80 m s^{-1} (best fit with respect to convergence from Figure 3). Stars represent $\lambda E_{0,PT}$ with α = 1.10, as opposed to the α = 1.26 shown in Figure 2a.

opposed to either the standard α = 1.26 (see Figure 2a) or *Hobbins et al.*'s [2001b] recalibrated α = 1.318, results in convergence at adequately high M_A, as seen in Figure 4. Further, an unstressed $\lambda E_{p,PM}$ (evaluated with $g_{c,\max}$) maintains the symmetry about $\lambda E_{0,PT}$ (α = 1.10) as *Bouchet* [1963] intended, and upon which the derivation of equation (1) is founded. Moreover, as the finite canopy conductance causes an increase of surface temperature and sensible heat flux in the λE_p energy balance calculation, an energy balance adjustment occurs naturally, thus negating the need for an H_p correction similar to that proposed by *Parlange and Katul* [1992] in equation (14). As such, the proposed method is consistent with CR explanations that rely on feedbacks with the boundary layer vapor pressure deficit, since the impact of such changes on both λE_p and λE_a would be modulated by stomatal conductance.

[23] These improvements to λE_p estimation in *Parlange and Katul*'s [1992] modified AA model (PK92v1, uncorrected for $|H_p|$) not only clarify CR theory, but result in more precise indirect λE_a estimations. As seen in Figure 5, the use of $\lambda E_{p,Pen}$ (using $E_{a,DP}$ in equation (11)) and $\lambda E_{0,PT}$ (α = 1.26) in equation (1) per PK92v1 results in an underestimation of λE_a (root mean square error, RMSE $\approx$ 133 W m^{-2}); moreover, this underestimation of λE_a increases with decreasing M_A (consistent with Figure 2a). Alternately, use of an $|H_p|$ corrected $\lambda E_{0,PT}$ (α = 1.26), as suggested by *Parlange and Katul* [1992] (i.e., using PK92v2), increases error in λE_a estimations, i.e., RMSE $\approx$ 299 W m^{-2}. The replacement of *Parlange and Katul*'s [1992] $\lambda E_{p,Pen}$ with $\lambda E_{p,PM}$ (evaluated with $g_{c,\max}$) in the modified AA method (PK92v1) results in much more precise indirect λE_a estimations, i.e., RMSE $\approx$ 48 W m^{-2} (see Figure 5). Further, Figure 5 demonstrates that equation (1) evaluated with *Penman*'s [1948] empirical wind function and $\lambda E_{0,PT}$ (α = 1.26) also overestimates λE_a, but results in lower RMSE than either PK92 method, i.e., use of Penman wind function λE_p in equation (1) results in RMSE $\approx$ 158 W m^{-2}.

[24] In summary, as the derivation of equation (1) is founded upon equilibrium convergence and perfect symmetry (i.e., Φ = 1), and as the relationship between λE_a and $\lambda E_{p,PM}$ (evaluated with an unstressed $g_{c,\max}$ instead of a stressed g_c) more accurately reproduces this behavior, more precise estimates of λE_a should be expected. Therefore our results suggest that an estimation of $g_{c,\max}$ is fundamental to accurate estimations of λE_a per equation (1). Complementarity is only preserved when the two major controls on vapor diffusion, stomatal conductance and stability-corrected atmospheric conductance, are each incorporated in the definition of λE_p. In a similar fashion, *Ozdogan and Salvucci* [2004] and *Ozdogan et al.* [2006] found that stability corrections were necessary to preserve complementary over an irrigated agricultural site in semiarid Turkey.

[25] While our choice of $g_{c,\max}$ increments was arbitrary, a $g_{c,\max}$ = 1/80 m s^{-1}, when scaled to leaf-level stomatal conductance via leaf area index, is within the range of maximum stomatal conductance values applied in Jarvis-type models for FIFE [*Stewart and Verma*, 1992; *Kelliher et al.*, 1995; *Niyogi and Raman*, 1997]. Additionally, the Food and Agricultural Organization (FAO) Penman–Monteith reference crop formulation suggests surface resistance

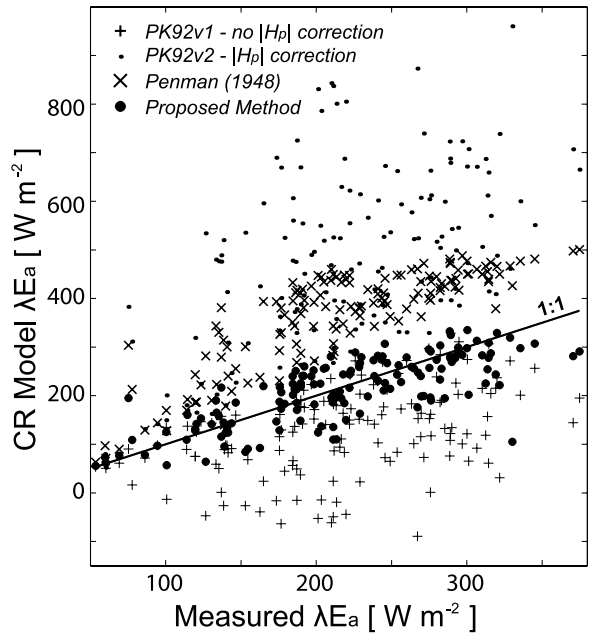

Figure 5. Poor results of both PK92v1 (no $|H_p|$ correction to $\lambda E_{0,PT}$, α = 1.26) and PK92v2 ($|H_p|$ correction per equation (14) to $\lambda E_{0,PT}$, α = 1.26) modified advection aridity λE_a estimation using *Penman* [1948]–*Brutsaert* [1982] $\lambda E_{p,Pen}$. More precise λE_a estimates are shown when using $\lambda E_{p,PM}$ ($g_{c,\max}$ = 1/80 m s^{-1}) instead of $\lambda E_{p,Pen}$ in modified AA methods. The λE_a estimates are shown using $\lambda E_{p,Pen}$ with *Penman*'s [1948] wind function, as used in the original AA formulation.

should be fixed to 1/70 m s^{-1} [*Allen et al.*, 1998]. While FAO's agricultural-oriented methodologies may necessitate a global (constant) surface resistance equal to 1/70 m s^{-1}, we do not suggest that our $g_{c,\max}$ = 1/80 m s^{-1} should necessarily be applied elsewhere. Rather, the purpose of our study is to highlight the importance of including a spatially scaled canopy conductance in both theoretical and application-oriented CR investigations. Further, while we do not suggest methodologies for estimating scaled $g_{c,\max}$ for regions, recent studies have addressed this topic in greater quantitative detail [*Isaac et al.*, 2004; *Kelliher et al.*, 1995].

[26] *Crago and Crowley* [2005] compared the *Granger* [1989] CR method (equation (6)) with the modified AA (PK92v1) over FIFE at 10–30 min time steps. They also found the AA method significantly underpredicted λE_a, while Granger's method was relatively unbiased. They did not address the issue of stomatal conductance in the AA framework. It is interesting that either the AA method with a canopy conductance term (as conceptually described in Figure 1 and demonstrated in Figure 4), or the *Granger* method [*Crago and Crowley*, 2005] (Figure 2a, top left) each explain the FIFE data, despite being built on fundamentally different assumptions (see paragraph 9).

5. Conclusions

[27] We demonstrate that at the FIFE site, the replacement of $\lambda E_{p,Pen}$ in *Parlange and Katul*'s [1992] modified AA

model with an unstressed $\lambda E_{p,PM}$, in which the canopy conductance is held at a maximum "potential" value ($g_{c,\max}$), results in *Bouchet*'s [1963] assumed convergence of λE_a and λE_p at the wet end of the CR relationship and symmetry at the dry end. These improvements to the modified AA model yield more precise λE_a estimates for the FIFE temperate grassland while maintaining the attractive simplicity of CR, i.e., quantifying λE_a without estimating complex land surface parameter responses to nonpotential (unstressed) evapotranspiration conditions. Our results suggest that the *Penman* [1948] wind function, as used in *Brutsaert and Stricker*'s [1979] original AA, improves convergence at the wet end, but does not display complementary-type behavior at the dry end (see Figure 3). Furthermore, the empiricism of such wind functions does not help theoretically advance the poorly understood CR hypothesis. Alternately, utilization of $\lambda E_{p,PM}$ allows for sound incorporation of, and investigations with respect to, the interactive role of canopy conductance and stability-corrected atmospheric conductance in Bouchet's heuristic hypothesis.

[28] Further, *Parlange and Katul*'s [1992] proposed $|H_p|$ correction to $\lambda E_{0,PT}$ to account for advection-induced shifts in the complementarity between λE_a and $\lambda E_{p,Pen}$ appears, in this case, to introduce additional error to the CR method (see Figure 5). Alternately, incorporation of $g_{c,\max}$ results in a natural adjustment to the λE_p energy balance, thus negating the need for an H_p correction. Therefore, since the impact of such changes on both λE_p and λE_a would be modulated by stomatal conductance, the proposed method is consistent with CR explanations that rely on feedbacks with the boundary layer vapor pressure deficit.

[29] In conclusion, our results suggest that theoretical explanations of *Bouchet*'s [1963] heuristic hypothesis should not neglect the degree to which canopy conductance regulates the complementarity between λE_a and λE_p. Further, when plant transpiration is a significant constituent of evapotranspiration, the $\lambda E_{p,Pen}$ flux, i.e., using *Brutsaert*'s [1982] $E_{a,DP}$, will overestimate true λE_p due to the absence of an unstressed $g_{c,\max}$ in $\lambda E_{p,Pen}$. Finally, while Penman-type wind functions may be calibrated to account for $g_{c,\max}$ and thus prove useful in applications, such lumped incorporations of complex atmospheric and vegetative conductances do not allow detailed investigation of the mechanisms underlying *Bouchet*'s [1963] hypothesis.

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