An Analysis of the Effect of Local Heat Advection on Evaporation over Wet and Dry Surface Strips

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ABSTRACT

The effect of local advection on evaporation and Bowen ratio over alternating crosswind infinite dry-warm and wet-cool surface strips (patches), by redistribution of surface heat, is analyzed. The analysis shows that evaporation over the region is enhanced by local advection, as observed experimentally by other researchers. Results suggest that evaporation can be enhanced more than 10% or 20% by local advection on the area covered by patches smaller than 10t_μ U_μ m or 5t_μ U_μ m, respectively, where t_μ is a unit time and U_μ the mean wind velocity in the internal boundary layer. This means that the traditional first-order approximation of surface energy balance, without considering local advection, might lead to significant error for such small-scale inhomogeneity. It may, however, be applicable to large patches (≥10t_μ U_μ m), where energy partition may then be directly approximated from area-weighted averages of properties derived from extended wet and dry surfaces, according to the patchy area fractions involved.

1. Introduction

Surfaces such as northern wetlands are often covered by alternating wet-cool puddles and dry-warm lichen beds during clear summer days, at patch size scales from about a hundred meters down to a few meters. The wet patches are frequently observed to be completely saturated like open water with surface temperature around 20°C, and the dry patches completely dry with surface temperature around 40°C. Such extensive highly contrasted small-scale surface temperature and wetness inhomogeneity might be expected to affect the local (and in a cumulative way the regional) heat and moisture transfer processes and balances, particularly evaporation rates and Bowen ratio (ratio of sensible to latent heat transfer). In the presence of advective wind velocity, a horizontal heat flux between the wet-cool and dry-warm patches can be generated, which would reallocate a portion of sensible heat from the dry to the wet patches, enhancing evaporation from the wet patches.

The above picture presents a challenge insofar as it is not clear at what patch size the overall energy partitioning over an extensive surface with temperature and wetness inhomogeneity can be approximated from area-weighted averages of properties derived from extended wet and dry surfaces, which may be assumed to be in equilibrium with the surface boundary layer. In other words, it is not clear at what patch size the Bowen ratio of a patchy surface deviates nonnegligibly from an area-weighted average of Bowen ratios of extended wet and dry areas, according to the area fractions involved. This problem is particularly acute in questions about the effects of subgrid-scale surface heterogeneity on atmospheric processes (Wetzel and Chang 1988; Avissar and Pielke 1989; Avissar and Verstraete 1990). This problem is also of interest in the issue of water-use efficiency for “strip- and block-farming” systems where narrow, fallow strips alternate with plowed cultivated ones. There is concern in that case that sensible heat advection may reduce the anticipated benefits to the soil moisture budget (Clothier 1989).

The small-scale horizontal transports, that is, local advection, of heat and moisture resulting from surface temperature and wetness inhomogeneity has been extensively studied in the literature, as reviewed partially by Brutsaert (1982) and Garratt (1990). The results have been applied to diverse cases such as grassland discontinuity (Rider et al. 1963), small water surfaces (Brutsaert and Yeh 1969), inhomogeneous irrigation (McNaughton 1981), and marginal ice zone (Clausen 1991), etc. Most of the studies emphasized the development of boundary-layer profiles with special attention to the transport processes in the atmosphere. The methods of describing atmospheric transport have been developed to a high degree of sophistication, which includes similarity (Philip 1959), self-preservation (Townsend 1965), first-order closure (Taylor 1970,

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turbulent kinetic energy closure (Panchev et al. 1971), higher-order closure (Rao et al. 1974), and Karman's integral [Novak (1990), extension of Elliott (1958)]. By comparison, the surface component of the problem drew less attention. Although the surface moisture balance (or surface water availability) has been treated in a potentially realistic way by Rao et al. (1974) and McNaughton (1976), the surface energy balance is still limited by the assumption that the surface inhomogeneity only affects the partition of the available net radiation between sensible and latent heat (Rider et al. 1963; Rao et al. 1974). This assumption essentially excludes the presence of the horizontal heat flux over the surface that is responsible for transporting a portion of heat between patches, so that the surface energy balance treatment currently used in local advection problems may still be in its first-order approximation. When models with this surface treatment are extrapolated to the calculation of surface fluxes, the calculated values disagree with the measurements (Novak 1990). Although there might be many reasons for this discrepancy as discussed by Novak, the above-mentioned exclusion of the horizontal heat flux might be a contributing factor. This was confirmed by Hares and Novak (1992a,b) when they compared their soil model predictions, under a surface energy balance treatment that excludes horizontal advection, against field measurements in strip tillage experiments.

Oke (1979) was probably one of the first to experimentally test such discrepancy in surface energy balance. From the measurements made on an irrigated suburban lawn (area approximately 160 m²) and its surrounding hotter and drier surfaces, he attributed the increase in energy used in evapotranspiration by the lawn over its net radiation budget on both hourly and daily basis, to advection of sensible heat from surrounding surfaces. A similar argument was earlier used by Shaw and Decker (1977) in the case of reduced canopy density, where sensible heat produced at bare soil surfaces is advected over the canopy and used to enhance transpiration beyond that permitted by the available net radiation. Some other experimental results supporting the above argument were later given by Nullet (1987) and Nullet et al. (1990) in the case of ocean–land discontinuity, and by Gay (1991) in the case of drip-irrigation in arid regions. Hares and Novak (1992b) recently presented more quantitative supporting results from their experiments with dry–hot strips of straw mulch (30 cm wide) alternating with wet–cool strips of bare soil (10 cm wide). They found that the measured evaporation rate from the wet strips could be 30% larger than that from an adjacent completely bare-wet plot, and the largest evaporation enhancement by the strips occurred when the strips were arranged perpendicular to the wind direction. This enhancement of evaporation was attributed to the local horizontal transport of warmer and drier air from the straw strips to the bare soil.

According to these experimental results, we may argue that the small-scale surface temperature and wetness inhomogeneity may not only change the partition of the available net radiation between sensible and latent heat, but also cause the redistribution of the available energy between patches. The effect of the latter is pertinent to the questions raised above, which by and large have not yet been explored theoretically and numerically. This study should then be viewed as such an attempt.

We shall explicitly study the energy budgets over small-scale heat-wetness inhomogeneous surfaces for the following simplifications only: The surface is considered to be composed of alternating crosswind-infinite wet–cool and dry–warm strips ("wetland" and "dryland") with sharp discontinuities between them but uniform conditions within each of them, as shown in Fig. 1. The crosswind horizontal transport is neglected. Practical application of this analysis is then limited to situations with significant crosswind dimension. Dryland surfaces are assumed to be completely dry with infinite surface resistance, and wetland surfaces completely saturated like open water surfaces with zero surface resistance. Vapor density in near-surface air is assumed to be close to saturation, so as to eliminate the influence of the large-scale (regional scale) advection described by Brakke et al. (1978), Brutsaert (1982, p. 218), McNaughton and Spriggs (1986), and Steyn (1990). These two approximations appear to be frequently satisfied by many parts of Northen wetlands during the early summer. The region is further considered to be covered by the same weather system, with large-scale inhomogeneities of the dynamic and thermodynamic properties ignored. Surfaces are considered

![Diagram](image_url)

**Fig. 1:** Schematic diagram of the dryland and wetland strips: (a) side view and (b) angled view from above; \( x_1 \) is the width of dryland and \( x_2 \) is the width of wetland.
"ideal," both dynamically and thermodynamically, with uniform surface roughness and stationary local variables.

As a preliminary test of our analysis, the modeled results are finally compared against some existing experimental measurements by other researchers. Since our approach at this stage contains significant simplifications that may not match the reported experimental conditions, the comparison presented here should, strictly speaking, be seen as only qualitatively meaningful.

2. The surface energy budget

Without considering local advection, the energy budgets at the dryland (subscript \(d\)) and wetland (subscript \(w\)) surfaces can be expressed, respectively, as (Oke 1987; Arya 1988):

\[
\vec{R}_d = \vec{R}_w - \vec{G}_d - \vec{S}_d = \vec{H}_d \tag{1a}
\]

\[
\vec{R}_w = \vec{R}_w - \vec{G}_w - \vec{S}_w = \vec{H}_w + \vec{Q}_w. \tag{1b}
\]

Here \(\vec{R}, \vec{R}_d, \vec{G}, \vec{S}, \vec{H}, \) and \(\vec{Q}\) are average flux densities per width of strip of available surface energy, net radiation, ground heat, change in subsurface heat storage, sensible heat, and latent heat, respectively.

By introducing the Bowen ratio \(\beta_w\) and the difference \(\Delta \vec{R}\) in available surface energy

\[
\beta_w = \frac{\vec{H}_w}{\vec{Q}_w},
\]

\[
\Delta \vec{R} = \vec{R}_w - \vec{R}_d
\]

\[
= (\vec{R}_w - \vec{R}_w) - (\vec{G}_w - \vec{G}_d) - (\vec{S}_w - \vec{S}_d)
\]

\[
= \Delta \vec{R}_n - \Delta \vec{G} - \Delta \vec{S}, \tag{2}
\]

the sensible and latent heat flux densities on the dryland and wetland can be expressed as

\[
\vec{H}_d = (\vec{R}_w - \Delta \vec{R}) \tag{3a}
\]

\[
\vec{H}_w = \frac{\beta_w}{1 + \beta_w} \vec{R}_w \tag{3b}
\]

\[
\vec{Q}_w = \frac{1}{1 + \beta_w} \vec{R}_w. \tag{3c}
\]

Due to local advection, the surface heat would be transported across both dry–wet and wet–dry borders. Under the conditions that the dryland is completely dry and hotter than the wetland, and that no condensation occurs in the surface boundary layer, a net amount of the dryland sensible heat will be "lost" into the wetland in the process. This amount of sensible heat would then become a net advected heat flux input to the wetland, with its average flux density \(\Delta \vec{H}\) per strip width and total heat \(x_2 \Delta \vec{H}\), where \(x_2\) is the width of the wetland strip. As a result, the loss of the average dryland sensible heat flux density would be \((x_2/x_1) \Delta \vec{H})\), where \(x_1\) is the width of the dryland strip. Assuming that \(\Delta \vec{H}\) will be primarily partitioned into an addition to the wetland average sensible heat flux density (\(\Delta \vec{H}_w\)) and an addition to the latent heat flux density (\(\Delta \vec{Q}_w\)), in a ratio \(B\) of \(\Delta \vec{H}_w\) to \(\Delta \vec{H}\), that is,

\[
\Delta \vec{H} = \Delta \vec{H}_w + \Delta \vec{Q}_w \tag{4}
\]

with

\[
B = \frac{\Delta \vec{H}_w}{\Delta \vec{H}} \tag{5}
\]

we have

\[
\Delta \vec{H}_w = B \Delta \vec{H} \tag{6a}
\]

and

\[
\Delta \vec{Q}_w = (1 - B) \Delta \vec{H}. \tag{6b}
\]

Therefore, with consideration of local advection, the total average sensible and latent heat flux densities on the dryland and wetland would become

\[
\vec{H}_d = \vec{H}_d - \frac{x_2}{x_1} \Delta \vec{H}
\]

\[
= (\vec{R}_w - \Delta \vec{R}) - \frac{x_2}{x_1} \Delta \vec{H} \tag{7a}
\]

\[
\vec{H}_w = \vec{H}_w + \Delta \vec{H}_w = \vec{H}_w + B \Delta \vec{H} \tag{7b}
\]

\[
\vec{Q}_w = \vec{Q}_w + \Delta \vec{Q}_w = \frac{1}{1 + \beta_w} \vec{R}_w + (1 - B) \Delta \vec{H}. \tag{7c}
\]

The following constraint should be satisfied for surfaces subject to local advection in the case that the dryland is hotter than the wetland:

\[
\vec{H}_w \leq \vec{H}_d.
\]

According to (7a) and (7b), after rearrangement, we have

\[
\Delta \vec{H} \left( B + \frac{x_2}{x_1} \right) \leq \frac{1}{1 + \beta_w} \vec{R}_w - \Delta \vec{R}. \tag{8}
\]

3. Evaluation of \(\Delta \vec{H}\)

The average net advected heat flux density input \(\Delta \vec{H}\) into the wetland can be estimated as follows.

Under the assumption that the effects of downwind and crosswind diffusion and the net radiation divergence are negligible, which is justifiable except for very
low wind velocity, the simplified steady-state two-dimensional mean thermodynamic equation in the internal boundary layer can be written as (Philip 1958; Rider et al. 1963; Rao et al. 1974; McNaughton 1976; Novak 1990)

\[
\rho c_p U(x, z) \frac{\partial \theta(x, z)}{\partial x} = -\rho c_p \frac{\partial w' \theta'}{\partial z} = -\frac{\partial H(x, z)}{\partial z}. \tag{9}
\]

Here \( \rho \) is the mass density of air, \( c_p \) the specific heat of air, \( U(x, z) \) the mean wind velocity, and \( \theta(x, z) \) the mean potential temperature; \( w' \) and \( \theta' \) are the fluctuations of the vertical wind and potential temperature, respectively, and \( H(x, z) \) is the heat flux density.

The mutual dependence of temperature and mean wind poses a potential problem in the integration of (9) over \( x \). But if mean wind along \( x \) does not change significantly, the integration with a constant average mean wind velocity may be considered a reasonably good approximation. In this case, integrating (9) over the wetland, with respect to \( x \) from \( x = 0 \) to \( x = x_2 \), yields

\[
\frac{\bar{U}(x_2, z)}{x_2} \rho c_p \left[ \theta(0, z) - \theta(x_2, z) \right] = -\frac{\partial \bar{H}(x_2, z)}{\partial z}. \tag{10}
\]

Here \( \bar{U}(x_2, z) \) and \( \bar{H}(x_2, z) \) are horizontally averaged mean wind velocity and heat flux density over the wetland, at any height \( z \) and \( \theta(0, z) \) and \( \theta(x_2, z) \) are mean potential temperatures at upwind and downwind ends of the wetland at height \( z \), respectively.

Integrating (10) with respect to \( z \) from the surface to the average internal boundary-layer height \( h_i \) gives

\[
\frac{1}{x_2} \int_0^{h_i} \rho c_p \bar{U}(x_2, z)[\theta(0, z) - \theta(x_2, z)]dz = \bar{H}(x_2, h_i) - \bar{H}(x_2, 0) = \Delta \bar{H}. \tag{11}
\]

Here \( \bar{H}(x_2, 0) \) and \( \bar{H}(x_2, h_i) \) are respectively, the horizontally averaged heat flux densities over the wetland at the surface and at the top of the internal boundary layer. Their difference approximately accounts for the net advected heat flux input \( \Delta \bar{H} \) into the wetland, according to heat balance in the control volume of the internal boundary layer. The integration (11) is, however, only an approximate estimate of the heat flux difference which is handled essentially in von Kármán's integral method (Elliott 1958; Novak 1990).

The heat flux change in the internal boundary layer over the wetland \( \int_0^{h_i} \rho c_p \bar{U}(x_2, z)[\theta(0, z) - \theta(x_2, z)]dz \) can be expressed as

\[
\int_0^{h_i} \rho c_p \bar{U}(x_2, z)[\theta(0, z) - \theta(x_2, z)]dz = A \int_0^{x_1} \tilde{H}_d dx. \tag{12}
\]

Here \( A \) is the ratio of the net advected heat input into the wetland to the "effective" heat of the dryland \( \int_0^{x_1} \tilde{H}_d dx \). The latter represents the portion of dryland sensible heat that could potentially be swept into the wetland per unit time. It is defined by

\[
\int_0^{x_1} \tilde{H}_d dx = x_1 \tilde{H}_d \quad \text{for} \quad (x_1 \leq t_u \bar{U}_d) \tag{13a}
\]

\[
\int_0^{x_1} \tilde{H}_d dx = t_u \bar{U}_d \tilde{H}_d \quad \text{for} \quad (x_1 > t_u \bar{U}_d). \tag{13b}
\]

Here \( t_u \) is a unit time, introduced only for dimensional consideration because we are dealing with the heat flux density. The above equations would be seen dimensionally "unbalanced" without \( t_u \). Further, \( \bar{U}_d \) is the mean wind velocity averaged through the entire internal boundary layer above the dryland. In practice, \( t_u \) can be chosen as 1 second, 1 minute, or 1 hour depending on the time period one is interested in. Correspondingly, \( \bar{U}_d \) must be given in meters per second, per minute, or per hour.

Substituting (12) and (13) into (11), with consideration of (7a), we have

\[
\Delta \bar{H} = \frac{A}{1 + A} \left( \bar{R}_w - \Delta \bar{R} \right) \frac{x_1}{x_2} \quad \text{for} \quad (x_1 \leq t_u \bar{U}_d) \tag{14a}
\]

\[
\Delta \bar{H} = \frac{A t_u \bar{U}_d}{x_1 + A t_u \bar{U}_d} \left( \bar{R}_w - \Delta \bar{R} \right) \frac{x_1}{x_2} \quad \text{for} \quad (x_1 > t_u \bar{U}_d). \tag{14b}
\]

Substituting (14) into (8), the constraints for \( x_2 \) are derived as

\[
x_2 \geq f_1(x_1) \quad \text{for} \quad (x_1 \leq t_u \bar{U}_d) \tag{15a}
\]

\[
x_2 \geq f_2(x_1) \quad \text{for} \quad (x_1 > t_u \bar{U}_d) \tag{15b}
\]

where

\[
f_1(x_1) = \frac{(1 + \beta_w)(\bar{R}_w - \Delta \bar{R})AB}{(1 - A \beta_w)\bar{R}_w - (1 + \beta_w)\Delta \bar{R}} x_1 \tag{16a}
\]

\[
f_2(x_1) = \frac{(1 + \beta_w)(\bar{R}_w - \Delta \bar{R})AB t_u \bar{U}_d}{(x_1 - A \beta_w t_u \bar{U}_d)\bar{R}_w - (1 + \beta_w) x_1 \Delta \bar{R}} x_1 \tag{16b}
\]

When \( x_2 \) approaches its minimum values in (15), \( \Delta \bar{H} \) will reach its maximum values in (14), that is,
\[ \Delta \tilde{H} = \frac{(1 - A \beta_w) \tilde{R}_w - (1 + \beta_w) \Delta \tilde{R}}{B(1 + A)(1 + \beta_w)} \]

for \( (x_2 = f_1(x_1), x_1 \leq t_u \tilde{U}_d) \) \quad (17a)

\[ \Delta \tilde{H} = \frac{(x_1 - A \beta_w t_u \tilde{U}_d) \tilde{R}_w - (1 + \beta_w)x_1 \Delta \tilde{R}}{B(x_1 + A t_u \tilde{U}_d)(1 + \beta_w)} \]

for \( (x_2 = f_2(x_1), x_1 > t_u \tilde{U}_d) \) \quad (17b)

For \( x_2 \) smaller than \( f_1(x_1) \) or \( f_2(x_1) \), \( \Delta \tilde{H} \) should stay at its maximum values. Overall, \( \Delta \tilde{H} \) is evaluated as

\[ \Delta \tilde{H} = \frac{A}{1 + A} (\tilde{R}_w - \Delta \tilde{R}) \frac{x_1}{x_2} \]

for \( (x_2 \geq f_1(x_1), x_1 \leq t_u \tilde{U}_d) \) \quad (18a)

\[ \Delta \tilde{H} = \frac{(1 - A \beta_w) \tilde{R}_w - (1 + \beta_w) \Delta \tilde{R}}{B(1 + A)(1 + \beta_w)} \]

for \( (x_2 > f_1(x_1), x_1 \leq t_u \tilde{U}_d) \) \quad (18b)

\[ \Delta \tilde{H} = \frac{A t_u \tilde{U}_d}{x_1 + A t_u \tilde{U}_d} (\tilde{R}_w - \Delta \tilde{R}) \frac{x_1}{x_2} \]

for \( (x_2 > f_1(x_1), x_1 > t_u \tilde{U}_d) \) \quad (18c)

Here \( \Delta \tilde{H} \) would increase as \( A, \tilde{U}_d, \) and \( \tilde{R}_w \) increase, and as \( B \) and \( \Delta \tilde{R} \) decrease. For a given dryland width \( x_1, \Delta \tilde{H} \) would decrease with increasing wetland width \( x_2 \), and, reversely, for a given wetland width \( x_2, \Delta \tilde{H} \) would increase with increasing dryland width \( x_1 \).

4. Changes of evaporation and Bowen ratio

a. The evaporation change

Under the assumption that air over the wetland is close to saturation in water vapor, a simplified Penman equation based directly on the latent heat should be adequate for determining the average evaporation density over the wetland (Slatyer and McIlroy 1961). Without considering local advection, according to (3c), the average evaporation density \( \tilde{E}_w \) over the wetland should be determined as

\[ \tilde{E}_w = \frac{\tilde{Q}_w}{L_e} = \frac{1}{1 + \beta_w} \frac{\tilde{R}_w}{L_e}, \] \quad (19)

while the average evaporation density \( \tilde{E} \) over the whole area, composed of both wetland and dryland, should be evaluated as

\[ \tilde{E} = \frac{x_2}{x_1 + x_2} \tilde{E}_w, \] \quad (20)

where \( L_e \) is the latent heat of water evaporation.

With consideration of local advection, according to (7c), the average evaporation density \( \tilde{E}_w \) over the wetland should be given by

\[ \frac{\tilde{E}_w}{L_e} = \frac{\tilde{Q}_w}{L_e} = \frac{1}{1 + \beta_w} \frac{\tilde{R}_w}{L_e} + (1 - B) \frac{\Delta \tilde{H}}{L_e} \] \quad (21) 

and the average evaporation density \( \tilde{E} \) over the whole wetland and dryland area by

\[ \tilde{E} = \frac{x_2}{x_1 + x_2} \tilde{E}_w. \] \quad (22)

Due to local advection, the increase rate \( r_{ew} \) of average evaporation density, relative to that with zero local advection, over the wetland is then determined, according to (19) and (21), as

\[ r_{ew} = \frac{\tilde{E}_w - \tilde{E}_e}{\tilde{E}_w} = (1 - B)(1 + \beta_w) \frac{\Delta \tilde{H}}{\tilde{R}_w}. \] \quad (23)

Considering (18), we have

\[ r_{ew} = (1 - B)(1 + \beta_w) \frac{A}{1 + A} \left( 1 - \frac{\Delta \tilde{R}}{\tilde{R}_w} \right) \frac{x_1}{x_2} \]

for \( (x_2 > f_1(x_1), x_1 \leq t_u \tilde{U}_d) \) \quad (24a)

\[ r_{ew} = (1 - B)(1 + \beta_w) \frac{1}{1 + A} \frac{\Delta \tilde{R}/\tilde{R}_w}{B(1 + A)} \]

for \( (x_2 > f_1(x_1), x_1 > t_u \tilde{U}_d) \) \quad (24b)

\[ r_{ew} = (1 - B)(1 + \beta_w) \frac{A t_u \tilde{U}_d}{x_1 + A t_u \tilde{U}_d} \left( 1 - \frac{\Delta \tilde{R}}{\tilde{R}_w} \right) \frac{x_1}{x_2} \]

for \( (x_2 > f_2(x_1), x_1 > t_u \tilde{U}_d) \) \quad (24c)

\[ r_{ew} = (1 - B)(1 - A \beta_w x_1) \frac{x_1 - A \beta_w x_1 \tilde{U}_d - (1 + \beta_w) x_1 \Delta \tilde{R}/\tilde{R}_w}{B(x_1 + A t_u \tilde{U}_d)} \]

for \( (x_2 < f_2(x_1), x_1 > t_u \tilde{U}_d) \) \quad (24d)

These results suggest that \( r_{ew} \) increases with increasing \( A \) and \( \tilde{U}_d \), and decreasing \( B \) and \( \Delta \tilde{R} \). For a given dryland width \( x_1, r_{ew} \) decreases with increasing wetland width \( x_2 \), and for a given wetland width \( x_2, r_{ew} \) increases with increasing dryland width \( x_1 \).

According to (20), (22), and (23), the relative increase rate of the average evaporation density over the whole wetland and dryland area due to local advection is given by

\[ r_e = \frac{\tilde{E} - \tilde{E}_w}{\tilde{E}_w} = \frac{\tilde{E}_w - \tilde{E}_e}{\tilde{E}_w} = r_{ew}. \] \quad (25)

The fact that the increase rate \( r_e \) over the whole area is equal to the rate \( r_{ew} \) over the wetland simply reflects the fact that under the assumptions of the model the drylands are characterized by negligible evaporation.
b. The Bowen ratio change

Without considering local advection, the Bowen ratio over the whole area should be estimated directly from the fractional area distributions (widths) of the dryland and wetland as

$$\beta = \frac{x_w \bar{H}_w + x_d \bar{H}_d}{x_w \bar{Q}_w}.$$  \hspace{1cm} (26)

Considering (3), we have

$$\beta = \beta_w + (1 + \beta_w) \left( 1 - \frac{\Delta \bar{R}}{\bar{R}_w} \right) \frac{x_1}{x_2}. \hspace{1cm} (27)$$

With consideration of local advection, however, the Bowen ratio over the whole area would become

$$\bar{\beta} = \frac{x_w \bar{H}_w + x_d \bar{H}_d}{x_w \bar{Q}_w}. \hspace{1cm} (28)$$

Substituting (7) into (28), after rearrangement with (23) and (27), we have

$$\bar{\beta} - \frac{\beta - (1 - \beta)}{1 + (1 - B)(1 + \beta_w) \Delta \bar{R}/\bar{R}_w} = \frac{\beta - r_{ew}}{1 + r_{ew}}. \hspace{1cm} (29)$$

According to (27) and (29), the decrease rate \(-r_b\) of the Bowen ratio relative to that with zero local advection, over the whole area, due to local advection is determined as

$$-r_b = \frac{\beta - \bar{\beta}}{\beta} = \frac{1 + \beta_w}{\beta} \frac{r_{ew}}{1 + r_{ew}}$$

$$= \frac{1 + \beta_w + (1 + \beta_w)(1 - \Delta \bar{R}/\bar{R}_w) x_1/x_2}{\beta_w + (1 + \beta_w)(1 - \Delta \bar{R}/\bar{R}_w) x_1/x_2} \frac{r_{ew}}{1 + r_{ew}},$$

which increases with increasing \(r_{ew}\).

5. Estimation of \(\beta_w, B, A,\) and \(\Delta \bar{R}\)

a. The value of \(\beta_w\)

The term \(\beta_w\) is the Bowen ratio over the wetland with zero local advection, which can be considered equivalent to the Bowen ratio over an extensive saturated wetland. Under the assumption that air over the wetland is close to saturation in water vapor, \(\beta_w\) can be estimated as (Slater and McIlroy 1961)

$$\beta_w = \frac{\gamma}{\Delta} \hspace{1cm} (31)$$

with the psychrometric constant defined as

$$\gamma = \frac{c_p P}{0.622 L_e} \hspace{1cm} (32)$$

and the rate of change of the saturated vapor pressure \(e_s\) against air temperature \(T\) defined as

$$\Delta = \frac{de_s}{dT}, \hspace{1cm} (33)$$

where \(P\) is atmospheric pressure.

According to the integral form of the Clausius-Clapeyron equation, considering the fact that \(e_s = 6.11\) mbar when \(T = 273\) K, we have (Hess 1959)

$$e_s = 6.11 \exp \left( \frac{19.85(T - 273)}{T} \right) \hspace{1cm} (34)$$

whose derivative with respect to \(T\) is

$$\frac{de_s}{dT} = 6.11 \exp \left( \frac{19.85(T - 273)}{T} \right) \frac{5419.05}{T^2}. \hspace{1cm} (35)$$

For \(T = 300\) K and \(P = 1000\) mbar, \(\gamma\) and \(\Delta\) are evaluated, according to (32) and (35), as

$$\gamma = 0.646 \text{ mb K}^{-1}$$

$$\Delta = 2.188 \text{ mb K}^{-1}$$

so that

$$\beta_w \approx 0.3. \hspace{1cm} (36)$$

It is perhaps interesting to note that this value is in general agreement with observations by Lafleur and Rouse (1988) over extensive marshland covered with standard surface water under offshore wind conditions, when air temperature is about 293 K.

b. The value of \(B\)

According to the definition (5), the ratio \(B\) of the sensible heat compensation part to the total advected heat input into the wetland should vary within a range of positive values less than or equal to 1. A bigger value of \(B\) means that the advected heat is less efficiently used for evaporation on the wetland. When \(B = 1\), no advected heat is used for evaporation.

As the value of \(B\) decreases, the efficiency for the advected heat to be used for evaporation on the wetland increases. However, a minimum limit \(B_{\min}\) may exist at which the advected heat would be maximally used for evaporation. Therefore, the value of \(B\) should actually vary within the following range:

$$B_{\min} \leq B \leq 1. \hspace{1cm} (37)$$

When \(B = B_{\min}\), the Bowen ratio of the sensible heat compensation part to the total advected heat input into the wetland may be reasonably approximated by \(\beta_w\).

According to (4) and (5), we have

$$B_{\min} = \frac{(\Delta \bar{H}_w)_{\min}}{(\Delta \bar{Q}_w)_{\max}}. \hspace{1cm} (38)$$

If

$$\frac{(\Delta \bar{H}_w)_{\min}}{(\Delta \bar{Q}_w)_{\max}} = \beta_w.$$

then

\[ B_{\text{min}} = \frac{\beta_w}{1 + \beta_w}. \]  

(39)

For \( T = 300 \text{ K} \) and \( P = 1000 \text{ mb} \), according to (36), we have

\[ B_{\text{min}} \approx 0.23. \]  

(40)

c. The value of \( A \)

Generally, the ratio \( A \) of the net advected heat input into the wetland to the "effective" heat of the dryland [Eqs. (12), (13)] must be linked to the ratio of the horizontal advective component to the sum of horizontal and vertical components of wind velocity across the dryland and wetland border, within the internal boundary layers. As a very rough approximation, we may use the latter to estimate \( A \) under the assumption that the effective sensible heat of the dryland is primarily transported by the horizontal advective and the vertical convective motions. The horizontal advective transport would contribute to the evaporation on the wetland and the vertical convective transport to the development of thermals. We then have

\[ A \approx \frac{\bar{U}_{\text{dw}}}{\bar{U}_{\text{dw}} + \bar{W}_{\text{dw}}}, \]  

(41)

when \( \bar{U}_{\text{dw}} \) and \( \bar{W}_{\text{dw}} \) are, respectively, the mean advective velocity and the average mean updraft convective velocity, averaged in the internal boundary layers above the dryland and wetland border.

The term \( \bar{U}_{\text{dw}} \) is primarily caused by horizontal pressure gradients. Its value is influenced by surface roughness characteristics and boundary-layer stratification; \( \bar{W}_{\text{dw}} \) is mainly caused by buoyancy. Its value is mainly influenced by the boundary-layer thermal stability.

According to (41), \( A = 0 \) when horizontal wind velocity is zero, which means that no sensible heat of the dryland will be transported into the wetland. In other words, all sensible heat on the dryland will contribute to the development of the thermals. As horizontal wind velocity increases, and thermal instability decreases, the value of \( A \) increases. This means that more sensible heat on the dryland will be transported into and then used for evaporation over the wetland, and that less sensible heat on the dryland will contribute to the development of thermals. Under stable thermal boundary-layer conditions with nonzero horizontal wind, \( A \) may approach 1, which means that the sensible heat on the dryland will be transported into the wetland at the maximal rate allowed by horizontal wind velocity. Generally, the value of \( A \) should vary within the range

\[ 0 \leq A \leq 1. \]  

(42)

d. The value of \( \Delta R \)

According to (2), the difference \( \Delta R \) in surface available energy between the wetland and dryland should generally be a function of the differences in albedo, surface emissivity, subsurface thermal conductivity, and subsurface heat capacity between the two types of surfaces. It should in principle be possible to deduce \( \Delta R \) from surface observations. Depending on weather conditions and type of surface medium, its value may be positive or negative.

6. The maximal effect of local advection

In this section, we investigate the maximal evaporation increase rate and Bowen ratio decrease rate by local advection, and compare the results with some existing experimental observations, then try to answer the question we raised in the Introduction. According to the above analysis, these maximal rates occur when \( B \) approaches its minimum and \( A \) approaches its maximum. For simplicity, we ignore the difference \( \Delta R \) in available energy between wetland and dryland, as generally assumed by many others (Rider et al. 1963; Rao et al. 1974), that is,

\[ \Delta R \approx 0 \text{ W m}^{-2} \]

\[ \beta_w \approx 0.3 \]

\[ B_{\text{min}} \approx 0.23 \]

\[ A_{\text{max}} = 1. \]

Substituting these values into (16), we have

\[ f_1(x_1) = 0.427 t_u x_1 \]

(43a)

\[ f_2(x_1) = \frac{0.299 t_u \bar{U}_d x_1}{x_1 - 0.3 t_u \bar{U}_d}. \]

(43b)

Substituting these values into (24), (25), and (30), the maximal evaporation increase rate \( (r_e)_{\text{max}} \), and the maximal Bowen ratio decrease rate \( (-r_B)_{\text{max}} \) are calculated as follows:

\[ (r_e)_{\text{max}} = \frac{0.5 x_1}{x_2} \quad \text{for} \quad (x_2 \geq 0.427 t_u x_1, x_1 < t_u \bar{U}_d) \]  

(44a)

\[ (r_e)_{\text{max}} = \frac{1.17}{x_2} \quad \text{for} \quad (x_2 < 0.427 t_u x_1, x_1 \leq t_u \bar{U}_d) \]  

(44b)

\[ (r_e)_{\text{max}} = \frac{t_u}{x_1 + t_u \bar{U}_d} \quad \text{for} \quad \left( x_2 \geq \frac{0.299 t_u \bar{U}_d x_1}{x_1 - 0.3 t_u \bar{U}_d}, x_1 > t_u \bar{U}_d \right) \]  

(44c)

\[ (r_e)_{\text{max}} = \frac{3.35}{x_1 + t_u \bar{U}_d} \quad \text{for} \quad \left( x_2 < \frac{0.299 t_u \bar{U}_d x_1}{x_1 - 0.3 t_u \bar{U}_d}, x_1 > t_u \bar{U}_d \right) \]  

(44d)
narrower wetland strips, the value of \((r_e)_{\text{max}}\) is big and dramatically increases with increasing dryland width. Both the value of \((r_e)_{\text{max}}\) and the dryland width influence decrease with increasing wetland width \(x_2\). They eventually vanish for big \(x_2\).

For small dryland strips with \(x_1 \leq 1\) m, the high-wind regime does not help to increase \((r_e)_{\text{max}}\) since the supply of the advected heat becomes the limiting factor. For \(x_1 > 5\) m, the increase of \((r_e)_{\text{max}}\) with increasing \(x_1\) in the high-wind regime is larger than that in the low-wind regime. In this case, a wider wetland is influenced by local advection to have more significant evaporation increase in the high-wind regime than in the low-wind regime. For example, the width of wetland with \((r_e)_{\text{max}} \geq 10\%\) is about 10 m when \(\bar{U}_d = 1\) m s\(^{-1}\), but about 50 m when \(\bar{U}_d = 5\) m s\(^{-1}\).

Generally, the width of wetland with \((r_e)_{\text{max}} \geq 10\%\) or \(20\%\) is about 10\(t_u\bar{U}_d\) m or 5\(t_u\bar{U}_d\) m, respectively. This suggests that estimating regional evaporation without considering local advection only causes an error of \(<10\%\) when the wetland width exceeds 10\(t_u\bar{U}_d\) m, but may lead to an error of \(>20\%\) when the wetland width becomes less than 5\(t_u\bar{U}_d\) m.

The development of the Bowen ratio decrease rate \((-r_b)_{\text{max}}\), with both \(x_1\) and \(x_2\) for the two wind regimes, is shown in Fig. 3. Here \((-r_b)_{\text{max}}\) always has positive value, reflecting the reduction of the Bowen ratio over the whole area by local advection. However, \((-r_b)_{\text{max}}\) generally decreases with the increase of the wetland width.

Fig. 2. The development of the evaporation increase rate \((r_e)_{\text{max}}\) with dryland width \(x_1\) and wetland width \(x_2\) for two wind regimes: (a) \(\bar{U}_d = 1\) m s\(^{-1}\), and (b) \(\bar{U}_d = 5\) m s\(^{-1}\).

\[
(-r_b)_{\text{max}} = \frac{x_1 + x_2}{x_1 + 0.23x_2} \left(\frac{x_1}{x_1 + 2x_2}\right)
\]
for \((x_2 \geq 0.427x_1, x_1 \leq t_u\bar{U}_d)\) (45a)

\[
(-r_b)_{\text{max}} = \frac{0.54x_1 + 0.54x_2}{x_1 + 0.23x_2}
\]
for \((x_2 < 0.427x_1, x_1 \leq t_u\bar{U}_d)\) (45b)

\[
(-r_b)_{\text{max}} = \frac{x_1 + x_2}{x_1 + 0.23x_2} \left[\frac{t_u\bar{U}_d x_1}{x_1 x_2 + t_u\bar{U}_d(x_1 + x_2)}\right]
\]
for \((x_2 \geq 0.299t_u\bar{U}_d x_1, x_1 > t_u\bar{U}_d)\) (45c)

\[
(-r_b)_{\text{max}} = \frac{x_1 + x_2}{x_1 + 0.23x_2} \left(\frac{0.77x_1 - 0.23t_u\bar{U}_d}{x_1}\right)
\]
for \((x_2 < 0.299t_u\bar{U}_d x_1, x_1 > t_u\bar{U}_d)\). (45d)

Applying (44) and (45) to the two assumed wind velocity regimes of \(\bar{U}_d = 1\) m s\(^{-1}\) (or \(t_u\bar{U}_d = 1\) m) and \(\bar{U}_d = 5\) m s\(^{-1}\) (or \(t_u\bar{U}_d = 5\) m), respectively, yields the following results.

Figure 2 shows the development of \((r_e)_{\text{max}}\) with both dryland width \(x_1\) and wetland width \(x_2\) for the two wind velocity regimes. Here, \((r_e)_{\text{max}}\) always exhibits a positive value, which means that evaporation over the wetland is generally enhanced by local advection. For
For narrow dryland strips with \( x_1 \leq 1 \text{ m} \), \((-r_h)_{\text{max}}\) generally increases with increasing dryland width. In this case again, the role of the supply of the advected heat as a limiting factor in the high-wind regime is apparent. In analogy with the evaporation increase, the change in \((-r_h)_{\text{max}}\) for \( x_1 \geq 1 \text{ m} \) in the high-wind regime is bigger than that in the low-wind regime for \( x_2 > 1 \text{ m} \). Similarly, the wetland width leading to a significant change in Bowen ratio over the whole area in the high-wind regime is again larger than that in the low-wind regime.

The wetland width leading to a change of \((-r_h)_{\text{max}}\) 
\[ \geq 10\% \] \( \) is about \( 20\tau_u \tilde{U}_D \text{ m} \). Compared with \( 10\tau_u \tilde{U}_D \text{ m} \) in \((r_e)_{\text{max}}\) in Fig. 2, the effect of local advection on the Bowen ratio seems to be more persistent than on evaporation. This phenomenon can also be seen in Fig. 4.

Figure 4 shows the developments of both \((r_e)_{\text{max}}\) and \((-r_h)_{\text{max}}\) with equal dryland and wetland width for the two wind regimes. When \( x_1 \) and \( x_2 \) have values < \( \tilde{U}_D \text{ m} \), the effect of local advection on both evaporation and Bowen ratio becomes independent of wind velocity and widths \( x_1 \) and \( x_2 \). In this case, the rates of evaporation increase \((r_e)_{\text{max}}\) and Bowen ratio decrease \((-r_h)_{\text{max}}\) become constant at their biggest magnitude of 50\% and 55\%, respectively. When the strip width exceeds \( \tilde{U}_D \text{ m} \), \((r_e)_{\text{max}}\) and \((-r_h)_{\text{max}}\) will decrease with increasing width of strips. For strips \( > 1 \text{ m} \), both \((r_e)_{\text{max}}\) and \((-r_h)_{\text{max}}\) are always larger in the high-wind regime than in the low-wind regime. The differences in \((r_e)_{\text{max}}\) and \((-r_h)_{\text{max}}\) between the two wind velocity regimes decreases with increasing width of strips, however. These differences will eventually vanish for very wide dryland and wetland strips. Again, the patches with \((r_e)_{\text{max}} \geq 10\% \) or 20\% are about the sizes of \( 10\tau_u \tilde{U}_D \text{ m} \) or \( 2\tau_u \tilde{U}_D \text{ m} \), respectively.

Shaw and Decker (1977), Oke (1979), Nullet (1987), Nullet et al. (1990), Gay (1991), and Hares and Novak (1992b) have reported that, according to their experiments, local heat advection enhanced evaporation or evapotranspiration. Among these, Hares and Novak’s experimental design appears closest to our analytical situation in the microscale range. In their experiments, alternating parallel dry–warm straw strips of 0.3-m width and wet–cool bare soil strips of 0.1-m width were oriented along north–south, northeast–southwest, and east–west directions. With wind from the southwest, they observed evaporation from the north–south and east–west plots enhanced by local advection up to 30\% compared with an adjacent completely bare–wet plot. Using the same strip widths and combination, our analysis predicts an evaporation enhancement around 50\% (Fig. 2) for both low- and high-wind regimes. The reason why the high-wind regime does not contribute more to evaporation enhancement in this situation is that the supply of the advected heat from dry–warm strips to wet–cool strips becomes the limiting factor for this microscale range when wind speed exceeds 0.3 m s\(^{-1}\). Although Hares and Novak did not give the value of wind velocity for their experiments, we guess that it exceeded 0.3 m s\(^{-1}\) under the reported windy conditions. The quantitative difference of evaporation enhancement between Hares and Novak’s experiment and our analysis could be explained by many reasons. The major one is the significant simplifications in our approach, which at this stage is limited to ideal conditions, from which the maximal effect of local advection is examined. The analysis may then, to a certain degree, overestimate their experimental results, which would not match the conditions for maximal effect. Our analysis is, at least, qualitatively encouraging in predicting a maximal effect of local advection which looks credible in light of the reported observations.

Our analysis not only supports the experimental findings of enhanced evaporation under conditions of local heat advection, but also predicts that such effect should be nonnegligible for small patches with dimensions of the order of < \( 10\tau_u \tilde{U}_D \text{ m} \). In this situation, estimations of evaporation, Bowen ratio, and energy partition over a region, without considering local advection, could lead to significant errors. This result is particularly important for northern wetlands where patch sizes are frequently found with dimensions less than a few meters. As patch sizes increase, however, the effects of local advection quickly decrease. For large patches with dimensions \( \geq 10\tau_u \tilde{U}_D \text{ m} \), the evaporation enhancement by local advection would become neg-
ligibly small. In this situation, evaporation, Bowen ratio, or energy partition over the region can be approximated from the area-weighted averages of properties derived from extended wet and dry surfaces, according to the patchy area fractions involved.

This analysis is based on the assumption that air near wetland surface is close to saturation in water vapor for the purpose of eliminating the influence of the large-scale (regional scale) advection (e.g., Brutsaert 1982, p. 218). The approximation is easily satisfied by many parts of northern wetlands during the early summer. If this assumption deviates from reality, however, the effect of local advection on evaporation would be underestimated by the present analysis, because of the omission of the atmospheric drying power to which local advection should contribute. As the near-surface air becomes drier, the underestimation may become more serious. This influence should be examined in a future study.

7. Conclusions

The energy balance on surfaces with small-scale heat and wetness inhomogeneities may have been traditionally treated in its first-order approximation, limited by the assumption that surface inhomogeneity only affects the partition of the available net radiation between sensible and latent heat (Rider et al. 1963; Rao et al. 1974; McNaughton 1976). This assumption would essentially exclude the local horizontal heat flux above the surface that is responsible for redistribution of surface energy between patches.

Our study particularly analyzes the effect of such horizontal advection on evaporation and Bowen ratio, over alternating crosswind-infinite dry–warm and wet–cool land surface strips. It is based on simplifications such as negligible large-scale inhomogeneity of dynamic and thermodynamic properties, and assumption of ideal surface conditions. Results suggest that local heat advection enhances evaporation and hence reduces Bowen ratio over the region, and agree qualitatively with the experimental measurements reported by many others (Shaw and Decker 1977; Oke 1979; Nullet et al. 1990; Gay 1991; Hares and Novak 1992b).

According to this analysis, the contribution to evaporation by local advection generally increases with increasing boundary-layer wind velocity and dryland width but decreasing wetland width, under given land–atmosphere conditions. When the patches are small, the effect of local advection can be very big, so that the traditional first-order approximation of surface energy balance may lead to significant errors. For example, evaporation would be enhanced more than 10% or 20% by local advection over the region with patches less than 10\(t_u U_d\) m or 5\(t_u U_d\) m, respectively (\(t_u\) a unit of time and \(U_d\) the mean wind velocity in the internal boundary layer). On the area covered by large patches \(\gg 10_t U_d\) m, however, the effect of local advection becomes negligibly small. In such case, energy partition over the region may be reasonably estimated by the traditional first-order approximation.

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APPENDIX

List of Symbols

\[
\begin{align*}
A & \quad \text{ratio of net advected heat input into the wetland to } \int_0^{x_l} \dot{H}_d dx \\
B & \quad \text{ratio of } \Delta \dot{H}_w \text{ to } \Delta \dot{H} \\
B_{\text{min}} & \quad \text{minimum of } B \\
c_p & \quad \text{specific heat of air} \\
\varepsilon_s & \quad \text{saturated vapor pressure} \\
\bar{E} & \quad \text{average evaporation density per whole-area width including wetland and dryland without local advection} \\
\bar{E}_w & \quad \text{average evaporation density per whole-area width including wetland and dryland with local advection} \\
\bar{E}_d & \quad \text{average evaporation density per wetland width without local advection} \\
\bar{E}_d & \quad \text{average evaporation density per wetland width with local advection} \\
\bar{G}_d & \quad \text{average ground heat flux density per dryland width} \\
\bar{G}_w & \quad \text{average ground heat flux density per wetland width} \\
\Delta \bar{G} & \quad \text{difference between } \bar{G}_w \text{ and } \bar{G}_d \\
h_i & \quad \text{average internal boundary layer over wetland} \\
H(x, z) & \quad \text{heat flux density in the internal boundary layer} \\
\tilde{H}(x_2, 0) & \quad \text{horizontally averaged heat flux density over the wetland at surface} \\
\tilde{H}(x_2, h_i) & \quad \text{horizontally averaged heat flux density over the wetland at } h_i \\
\tilde{H}(x_2, z) & \quad \text{horizontally averaged heat flux density over wetland} \\
\tilde{H}_d & \quad \text{average surface sensible heat flux density per dryland width without local advection} \\
\tilde{H}_d & \quad \text{average surface sensible heat flux density per dryland width with local advection} \\
\tilde{H}_w & \quad \text{average surface sensible heat flux density per wetland width without local advection} \\
\tilde{H}_w & \quad \text{average surface sensible heat flux density per wetland width with local advection} \\
\Delta \tilde{H} & \quad \text{average net advected heat flux density input to unit width of wetland}
\end{align*}
\]
\( \Delta H_w \) average wetland sensible heat flux density compensation by local advection

\((\Delta H_w)_{\text{min}} \) minimum of \( \Delta H_w \)

\( L_e \) latent heat of water evaporation

\( P \) atmospheric pressure

\( \dot{Q}_w \) average surface latent heat flux density per wetland width without local advection

\( \dot{Q}_w \) average surface latent heat flux density per wetland width with local advection

\( \Delta \dot{Q}_w \) average wetland latent heat flux density compensation by local advection

\((\Delta \dot{Q}_w)_{\text{max}} \) maximum of \( \Delta \dot{Q}_w \)

\(-r_b \) average whole-area Bowen ratio decrease rate due to local advection

\((-r_b)_{\text{max}} \) value of \(-r_b\) due to maximal effect of local advection

\( r_e \) average whole-area evaporation increase rate due to local advection

\((r_e)_{\text{max}} \) value of \( r_e \) due to maximal effect of local advection

\( r_{w_e} \) average wetland evaporation increase rate due to local advection

\((r_{w_e})_{\text{max}} \) value of \( r_{w_e} \) due to maximal effect of local advection

\( \bar{R}_d \) average surface available energy flux density per dryland width

\( \bar{R}_{d_d} \) average net radiation flux density per dryland width

\( \bar{R}_w \) average surface available energy flux density per wetland width

\( \bar{R}_{w_w} \) average net radiation flux density per wetland width

\( \Delta \bar{R} \) difference between \( \bar{R}_w \) and \( \bar{R}_d \)

\( \Delta \bar{R}_d \) difference between \( \bar{R}_{w_w} \) and \( \bar{R}_{d_d} \)

\( \bar{S}_d \) average subsurface heat storage change rate per dryland width

\( \bar{S}_{w} \) average subsurface heat storage change rate per wetland width

\( \Delta \bar{S} \) difference between \( \bar{S}_w \) and \( \bar{S}_d \)

\( t_p \) unit time

\( T \) air temperature

\( U(x, z) \) mean wind velocity in the internal boundary layer

\( \bar{U}(x, z) \) horizontally averaged mean wind velocity over wetland

\( \bar{U}_d \) average mean wind velocity over dryland

\( \bar{U}_{d_w} \) average mean horizontal wind velocity over dryland–wetland border

\( w' \) fluctuation of vertical wind velocity

\( \bar{W}_{d_w} \) average mean vertical wind velocity over dryland–wetland border

\( x \) streamwise coordinate along mean wind direction

\( x_1 \) wetland width

\( x_2 \) dryland width

\( z \) vertical coordinate

\( \beta \) Bowen ratio over whole area without local advection

\( \beta \) Bowen ratio over whole area with local advection

\( \beta_w \) Bowen ratio over wetland without local advection, equivalent to that over an extensive wetland

\( \rho \) mass density of air

\( \theta' \) fluctuation of \( \theta \)

\( \theta(0, z) \) mean potential temperature at upwind end of wetland

\( \theta(x, z) \) mean potential temperature in the internal boundary layer

\( \theta(x_2, z) \) mean potential temperature at downwind end of wetland

\( \gamma \) "effective" sensible heat of dryland

\( \Delta \) change rate of \( e \) against \( T \)

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