A Literature Review on Vegetation-Atmosphere Interaction Research for Carbon Cycle and Energy Balance in Terrestrial Ecosystems

Hiroki Ikawa

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It is known that the physical and biological processes that occur at the surface of terrestrial vegetation interact with the atmosphere. While research tends to focus on either the impact of the atmospheric processes on the ecosystem or vice versa, there is a profound need to study ecosystem processes in the vegetation surface-atmosphere coupled system. This paper aims to review a certain subject matter where mutual influences between vegetation and atmosphere are expected with respect to carbon cycle and energy balance to better understand this need to investigate this coupled system. Finally, derivations of basic equations for a clear-sky one-dimensional (1D) planetary boundary layer (PBL) model and a regional atmospheric model are introduced as they are useful tools to investigate the interactions between vegetation and the atmosphere.

1. Background

More than two decades ago, Entekhabi (1995) wrote in his review article that scientists still use Lewis F.
attention in the past few decades to link ecosystem and environment studies with different spatial and temporal scales. For example, a received energy on Earth’s surface is distributed to sensible and latent heat fluxes that in turn heat or moisten the planetary boundary layer (PBL). Furthermore, a sensible heat flux (more accurately buoyancy flux) modifies the PBL height, which affects the budget of heat, vapor, and other gases in the atmosphere that in turn regulates surface fluxes. These studies have proved that more appropriate boundary conditions must be considered, namely, the top of the PBL rather than the interface between the atmosphere and vegetation (Margulis and Entekhabi, 2001a). Here, I define a coupled system as an integrated system where the atmosphere and an ecosystem interact. Nonetheless, it is useful to assume a simple system for investigating each process within the boundary layer. Therefore, a majority of recent studies still implicitly rely on the assumption of an uncoupled system where the atmospheric and ecosystem processes occur separately. The difficulty of investigating the coupled system in an individual study has been a significant challenge in the advancement of vegetation-atmosphere interaction research. In this paper, I aim to review a certain subject matter where mutual influences between the atmospheric forcing and vegetation feedback are evident. By so doing, I hope to better understand the importance of investigating vegetation-atmosphere interactions in coupled systems in future research. Note that from a perspective of the land surface, I use forcing as an atmospheric effect on vegetation and feedback as vegetation and entrainment effects on the PBL. Our knowledge constrains this subject matter review to ecosystem- and regional-scale studies on energy balance and carbon dioxide (CO2) in terrestrial ecosystems. It is important to note that relevant and important studies have also been conducted at the global scale (Manabe and Wetherald, 1967; Saito et al., 2004). Marine systems also affect the atmosphere over the land (Sato and Sugimoto, 2013), and other trace gases and volatile organic compounds (e.g., Miyazaki et al., 2016; Mochizuki et al., 2015; Vilà-Guerau de Arellano et al., 2011) are also important considerations in vegetation-atmosphere research.

2. Hydrological Cycle

Water flows continuously among different reservoirs on Earth’s surface and in the atmosphere, whereas water flow and the state of each reservoir are strongly interrelated to each other and influenced by the local energy balance. When estimating evapotranspiration from the land surface, not only the moisture status of the vegetation surface, but also the net impacts on the humidity level in the atmosphere and the resultant change in the partition of net radiation must be considered.

Sensitivity analysis with relatively simple land surface and atmospheric models has successfully identified a number of unique hydrological processes. Margulis and Entekhabi (2001b) developed an adjoint framework for a simple land surface and boundary layer model. This adjoint framework is useful in efficiently identifying the sensitivities of state variables (e.g., evapotranspiration) to both temporally fixed and variable parameters and their pathways, which are the components of the net effect. With the developed adjoint model, a sensitivity analysis has been conducted for latent and sensible heat fluxes with and without accounting for boundary layer feedback (Margulis and Entekhabi, 2001a). Their sensitivity demonstrated that the results were clearly different between the coupled and the uncoupled cases. Specifically, the net effects of a perturbation on the canopy temperature and ground temperature were dampened when feedbacks to the atmosphere were considered. They noted, however, that it is still challenging to obtain a clear consensus regarding the important parameters that control the feedback loops based on case studies.

A combination of the Penman-Monteith equation and an atmospheric PBL model (e.g., McNaughton and Spriggs, 1986), is one of the simplest modeling approaches used to investigate vegetation-atmosphere interactions. van Heerwaarden et al. (2010) quantitatively analyzed forcing and feedback effects on the diurnal patterns of evapotranspiration in two contrasting temperature environments. A single time derivative equation of the Penman-Monteith equation coupled with the PBL model was employed. Thus, the sensitivity of the dependent variable (e.g., evapotranspiration) to each
independent variable in the equation can be evaluated by the ratios of their derivatives with respect to time. Their model simulations showed that even though the total effect of boundary layer feedback (temperature and moisture fluxes from the land surface and entrainment) was similar between the two environments, the effect of adding or removing moisture had a more significant effect in cooler environments compared to the effect in warmer environments, which is reasonable according to the Clausius-Clapeyron relationship.

Similar model-based approaches in the framework of a regional atmospheric model have been conducted by Santanello and his colleagues (Santanello et al., 2015, 2013, 2011, 2009, 2007). Santanello et al. (2013) evaluated the performance of the NASA Unified Weather Research and Forecasting model (NU-WRF; Peters-Lidard et al., 2015) by coupling three land surface model (LSM) schemes and three PBL schemes plus offline LSM spin-ups (nine combinations) during dry and wet conditions in the southern Great Plains region of the United States. NU-WRF was run with a high spatial (~1 km) and temporal (~5 s) resolution for one week with LSMS spun up offline to provide initial land surface conditions. Santanello et al. (2013) compared surface heat fluxes (sensible, latent, and ground heat fluxes), air temperature, humidity, PBL height, and the lifting condensation level (LCL) deficit across the various simulations and observations.

Their results included the following:

• Surface heat fluxes varied more with the choice of LSM than with the PBL schemes.
• The difference in the surface heat fluxes between the coupled and the offline models was greater during wet regimes because shortwave radiation was overestimated owing to the inadequate representation of cloud formation.
• Both LSM and PBL schemes impacted the air temperature, humidity, and LCL deficit, but the impact of the PBL was greater during wet regimes.
• Model outputs such as reanalysis data are useful for investigating land-atmosphere interactions.

3. Climate Change

According to the classical theory of ecological research (e.g., Connell and Sousa, 1983), when the intensity of perturbation exceeds a certain threshold in a particular ecosystem, an alternative stable state, if it exists, appears as the integrative results of different processes across the atmosphere and biosphere. It is, therefore, important to understand the processes that are often overlooked owing to other competing factors in order to accurately understand the future trajectory of the vegetation-atmosphere coupled system under changing climate conditions.

One example where an alternative stable state likely occurs under climate change as the result of changing equilibrium among competing factors can be seen in the case of different species compositions with different interactions with the atmosphere (Baldocchi et al., 2000; Ikawa et al., 2015; Kobayashi et al., 2018; Nagano et al., 2018; Tsuyuzaki et al., 2008). Eddy covariance is a popular technique that is used to quantitatively understand ecosystem responses to the environment. With the aid of ecosystem models, eddy covariance data have been utilized to delineate a particular process that contributes to surface fluxes (Katul and Albertson, 1999; Lai et al., 2002; Ono et al., 2013; Ueyama et al., 2016). Carbon and oxygen isotope techniques can also be used to understand the role of different ecosystem components (e.g., Murayama et al., 2010; Wei et al., 2017, 2015). Eddy covariance can also be useful for targeting a particular ecosystem composition (e.g., forest understory) (Baldocchi et al., 2000; Black et al., 1996; Falk et al., 2005; Helbig et al., 2016; Iida et al., 2009; Ikawa et al., 2015).

Ikawa et al. (2015), for example, conducted eddy covariance flux measurements for a sporadic black spruce forest in Alaska’s interior (Fig. 1) as well as the understory compartment, which accounts for more than half of the areal fraction estimated by a flux footprint analysis. The major findings included the following: (1) the understory contributed to about half (40–80%) of the ecosystem CO₂ fluxes and an even greater fraction (50–98%) of surface energy fluxes and (2) the ecosystem and understory fluxes exhibited different responses to vapor pressure deficit. The results suggest that the
understory with the current species composition may be more vulnerable to extreme wet or dry conditions than black spruce trees, at least in the short term (e.g., a season).

Similarly, Helbig et al. (2016) conducted an eddy covariance measurement for a landscape of a jack pine forest in southern Taiga in Canada and for a permafrost-free wetland, which has been expanding in the forest. Their observations indicated that the wetland exhibited a higher albedo in the snow cover season, a greater latent heat flux, and a lower sensible heat flux compared to the landscape. They further investigated the potential impact of the conversion of a mixed boreal forest to a homogeneous wetland on the potential temperature and water vapor in the PBL using a clear-sky PBL model (McNaughton and Spriggs, 1986). They utilized observed surface fluxes as the boundary conditions of the model, and the model simulations indicated that the potential air temperature was lower, the water vapor pressure was greater, and the PBL height was lower for the wetland compared to the mixed boreal forest. The simulation also indicated that the greatest cooling effect was simulated during the snow cover period, likely because the high albedo resulted in a low sensible heat flux. They noted that their simulation was limited because the feedback effects to the atmosphere on the surface fluxes were not considered.

4. Land Use Distribution: Irrigated Fields

Land use distribution and its impact on the atmospheric environment have been investigated in the context of land management, environmental change, and the combination of the two (Baldocchi and Ma, 2013; Baldocchi et al., 2016; Bonan et al., 1992; Hemes et al., 2018; Law et al., 2018). Aside from greenhouse gas effects, the atmospheric impacts of a vegetation surface are primarily determined by the surface heat fluxes (Santanello et al., 2013). However, the partition of energy and its atmospheric impact are complex. For
example, increasing the aerodynamic conductance generally lowers the Bowen ratio, resulting in a cooling effect (Hemes et al., 2018; Chapter 6 in Kondo, 1994); on the other hand, a high rate of heat transfer could warm the atmosphere when the availability of moisture is limited (Baldocchi and Ma, 2013; Chapter 6 in Kondo, 1994). A cooler (warmer) land surface does not necessarily indicate a cooling (warming) effect on the atmosphere, and the mixing between the surface and the atmosphere must be considered (e.g., Okada et al., 2014).

The partition of energy and its impact on the atmosphere in the context of land use distribution have been investigated for irrigated agricultural fields. The impacts of irrigated fields on the atmosphere have long been recognized (Holmes, 1970). In particular, rice paddy fields distribute a large fraction of the received energy to evapotranspiration instead of a sensible heat flux (e.g., Ikawa et al., 2017). Ikawa et al. (2017) estimated the fraction of latent heat flux in net radiation to be 0.73 on average for 13 crop seasons on the basis of the data collected at the Mase Rice Paddy Site, one of the oldest continuous eddy covariance sites in Japan (Fig. 2) (Iwata et al., 2018; Mano et al., 2007; Miyata et al., 2005; Ono et al., 2015, 2013, 2008; Saito et al., 2005). Low air temperatures are, therefore, often recorded in areas near rice paddy fields (Dong et al., 2016; Kuwagata et al., 2014; Yokohari et al., 2001). However, Chen and Jeong (2018) reported that the average daily air temperature was higher in the irrigated areas compared to nonirrigated ones during the dry seasons on the North China Plain owing to warm temperatures at night.

The spatial distribution of air temperature and humidity near agricultural fields and other land uses have been reported in Japan (Kuwagata et al., 2018, 2014; Oue et al., 1994; Sakakibara et al., 2006; Yokohari et al., 2001). One of the important uses of such information is in the evaluation of whether and to what extent publicly available data from local meteorological stations are useful for agricultural management. With a meteorological data set obtained from one of the meteorological stations of the Japan Meteorological Agency (JMA) and from an adjacent crop field in the city of Kumagaya (known for its high temperature in summer), Kuwagata et al. (2014) compared air temperature between the two sites for three years. The result revealed that air temperature in the rice paddy field was lower than that at the JMA site, and the difference was more pronounced in the daily maximum temperature (>1°C) when solar radiation was high compared to the daily minimum temperature. Interestingly, the air temperature differ-
ence between the JMA site and a nearby forest in Kumagaya was known to be more pronounced in the daily minimum value than in the maximum value. Different local effects other than vegetation difference between the rice paddy field and the forest were not ruled out for a reason of the temperature difference compared to the JMA site; however, these results suggested that plant function types play an important role in determining the meteorological conditions over vegetation fields.

Kuwagata et al. (2018) further continued their study and compared the air temperature and vapor pressure between a rice paddy field and another JMA site in the city of Tsukuba, where surface energy flux data were available. Their main findings included the following: (1) air temperature was lower in the rice paddy field than at the JMA site, and the difference was primarily explained by the sensible heat flux; (2) the vapor pressure was higher in the rice paddy field than at the JMA site, and the difference was primarily explained by the latent heat flux; and (3) the daily minimum temperature difference in the fallow season was likely related to radiative cooling.

For agriculture, consideration of the vegetation-atmosphere coupled system is particularly important for estimating the plant body temperature. Yoshimoto et al. (2011) constructed an energy balance model (IM²PACT) to estimate the panicle temperature of rice plants in the region of Kanto in Japan. The model was performed for the year 2007 when unusually high air temperatures were recorded with the atmospheric forcing data estimated from a regional atmospheric model with the model biases corrected using the data obtained from local meteorological stations. They demonstrated that the panicle temperature was a better indicator for heat-induced spikelet sterility than the air temperature. Although the atmospheric forcing data used in this study may not perfectly represent the condition of agricultural fields, this study successfully demonstrated the importance of estimating the thermal environment of plant canopy by considering the energy exchange between the atmosphere and vegetation and not solely by air temperature.

The integrative effects of atmospheric forcing and land surface feedback ultimately affect crop production. Yoshida et al. (2012) simulated rice crop yields under decreasing rice paddy field areas for two decades (1987-2006) for the island of Shikoku in Japan with a rice growth model (Iizumi et al., 2009) and a nonhydrostatic regional atmospheric model (Saito et al., 2007). Their simulation suggests that air temperature increased more in the grid cells dominated by rice paddy fields than others by five times owing to the decrease in the fraction of the area of rice paddy fields. The daily maximum air temperature in the farm lands increased more than the daily minimum air temperature over the simulation period. The resultant decrease in the rice yield was 0.27% (1,196t of rice production), which was likely due to the shortened growth period with a smaller amount of total absorption of solar radiation. They also indicated that their crop growth model and the regional model were driven separately and suggested that considering the fully coupled system with an accurate evaluation of the latent heat flux would improve their model simulations.

Even among irrigated agricultural lands, differences in crop types and varieties alter the interaction between vegetation and the environment (Ikawa et al., 2018; Le et al., 2011). Based on the results of the Tsukuba free-air CO₂ enrichment experiment (Hasegawa et al., 2016; Nakamura et al., 2012), Ikawa et al. (2018) estimated evapotranspiration for a high-yielding rice cultivar (Takanari) and a commonly grown cultivar (Koshihikari) under changing atmospheric CO₂ concentrations using an LSM that consisted of submodels for photosynthesis (de Pury and Farquhar, 1997) and energy balance (Maruyama and Kuwagata, 2010; Maruyama et al., 2017; Watanabe, 1994). Our primary finding was that Takanari had 4-5% greater evapotranspiration than that of Koshihikari, but the increase was countered by an increase in CO₂ expected in 50 years. We further coupled the LSM with the clear-sky PBL model (McNaughton and Spriggs, 1986), which simulates the PBL height and mean scalar quantity (e.g., temperature, humidity, and CO₂) in the PBL. These parameters estimated by the PBL model were fed in the LSM, and surface fluxes by the LSM were used for the PBL model. The computation was made by an explicit numerical method with a small time step (1 min). The model results suggested that the change in the cultivars or the
CO₂ conditions has a clear impact on the canopy temperature and air temperature in the PBL (Fig. 3).

Topography and aerodynamic roughness modulate vegetation-land interactions at an extensive spatial scale. With a hydrostatic mesoscale atmosphere model, which was developed by Kimura and Arakawa (1983), Watanabe and Shimoyama (2015) simulated wind vectors over the Kanto Plain in Japan. The model simulation successfully regenerated a wind convergence near the city. However, the model did not simulate the wind convergence when the same aerodynamic roughness was assigned for all land types. The value of such a mesoscale model is used as a tool to implement a control experiment over a large spatial scale. Watanabe and Shimoyama (2015) further conducted a numerical experiment by hypothetically reducing the capacity of evapotranspiration from the forest by half. One of the most noteworthy results of the hypothetical simulation is that basins displayed a particular increase in air temperature despite the absence of forests. The increased sensible heat flux in the mountains that surround the basins enhanced the valley breeze during the daytime. The valley breeze was compensated by the subsidence, which supplies adiabatically warmed air over the basin.

5. Cloud Formation and Surface Energy Partitioning

The net available radiation that a land surface ecosystem receives is redistributed to sensible and latent heat fluxes. These fluxes, in turn, play a key role in determining the turbulence in the atmospheric boundary layer, as well as the formation and characteristics of boundary layer clouds. Moderate levels of cloud shading may increase photosynthesis in some plants by increasing the diffuse radiation (Freedman et al., 2001; Mercado et al., 2009; Pedruzo-Bagazgoitia et al., 2017). The cloud shading determined by the cloud optical depth has nonlinear effects on surface fluxes and, therefore, ecosystem feedback to the atmosphere.

The majority of the advancements in cloud-vegetation interactions have been accomplished using the large-eddy simulation technique (Horn et al., 2015; Vilà-Guerau de Arellano et al., 2014). In their seminal work, Vilà-Guerau de Arellano et al. (2014) performed a systematic numerical simulation to investigate how cloud shading affects energy partitioning of C3 and C4 plants and in return affects cloud formation under two competing factors (namely, the higher (lower) the buoyancy surface flux, the higher (lower) the convection, and the lower (higher) the evapotranspiration). A particular emphasis was also placed on analyzing the short time behaviors of stomatal conductance according to the meta-analysis reported by Vico et al. (2011) and
how to mimic the perturbations of evapotranspiration measured by the scintillometer technique (van Kesteren et al., 2013). Sikma et al. (2017) further elaborated on this work by reporting that the pattern of such cloud formation depends on the atmospheric structure, which was primarily determined by background wind speeds.

Vegetation-atmosphere interactions and the resultant cloud formation also interplay with climate change. In a model simulation study, Viñà-Guerrau de Arellano et al. (2012) suggested that elevated CO₂ decreases evapotranspiration via stomatal closure, which negatively affects cloud formation.

Synoptic-scale meteorology also interacts with the land surface and affects cloud formation. On the basis of the simulation over an idealized terrain using a two-dimensional (2D) cloud-resolving model (Ogura and Yoshizaki, 1988), Shinoda and Uyeda (2002) demonstrated that rice paddy fields in eastern China supply additional water vapor to the wet monsoon and generate shallow convective clouds. Shallow convective clouds further supply the moist to the troposphere, resulting in a condition that inhibits evaporation cooling of uplifted air and favors the formation of deep convective clouds.

6. Upper-Atmosphere Processes

I mentioned earlier that the top of the PBL is an appropriate boundary condition that must be considered for vegetation-atmosphere interaction research. It is important to be aware of how sensitive the processes within the PBL are to the prescribed boundary conditions in the upper atmosphere.

Based on eddy covariance measurements on the land surface and from aircraft (Electra) within the daytime PBL over boreal forests with the framework of the Boreal Ecosystem-Atmosphere Study (BOREAS) (Sellers et al., 1997), Davis et al. (1997) reported flux divergences for potential temperature and water vapor. In their study, flux divergence was defined by “the difference of fluxes measurements at two levels spaced over a significant fraction of the convective boundary layer.” One of their findings was that the flux divergence of water vapor was positive throughout the day, which indicates that the drying effect of the entrainment flux at the top of the PBL was greater than the moistening effect caused by evapotranspiration. Despite the drying air in the PBL, the effect of increased humidity outweighed and formed clouds as the PBL developed. They estimated the ratio of entrainment to surface flux of water vapor to be 1.57. Davis et al. (1997) also displayed an image of the cross section of the aerosol-laden PBL height, which provides us a visual sense of the PBL top characterized by the entrainment air and convective cells.

Upper atmospheric processes also affect the CO₂ budget. Combe et al. (2015) compared the performances of two coupled land surface and atmosphere models and used the one with a better agreement with observations to perform a sensitivity analysis of subsidence and soil moisture with respect to meteorological variables, intrinsic water use efficiency (intrinsic WUE, ratio of photosynthesis and stomatal conductance), and evaporation fraction (EF). The sensitivity analysis was performed by either increasing subsidence or decreasing soil moisture, both of which are characteristics of a drought period in the study region. Their findings showed that both changes in subsidence and soil moisture had a similar level of impact on the EF and WUE in the short term (one day). In the case of changed subsidence, the atmospheric CO₂ concentrations within the boundary layer were strongly modulated by a decrease in the height of the boundary layer. Based on these results, they emphasized the importance of the upper-atmosphere processes in the vegetation-atmosphere system. They also added an insightful discussion that although day-to-day variations of boundary layer growth and entrainment are less important from an atmospheric state perspective, these processes can be important for a specific period of crop development when plants are sensitive to heat and water stresses.

7. Clear-Sky 1D Planetary Boundary Model as a Tool to Analyze Vegetation—Atmosphere Interactions

One of the difficulties in vegetation-atmosphere interactions is that it is often not possible to obtain information on all relevant processes based on observation. Models are useful to estimate missing information at best. Models are often useful in analyzing the impact
of each process on the overall picture of vegetation-atmosphere interactions.

The mixed-layer theory is one of the most powerful tools used to investigate vegetation-atmosphere interactions, and the theory is utilized in the aforementioned PBL models. Although its application is mostly limited to 1D conditions with a developed PBL, it makes it possible to relate land surface processes to the state of the PBL with simple formulations (Baldocchi and Ma, 2013; Combe et al., 2015; Helbig et al., 2016; Ikawa et al., 2018; van Heerwaarden et al., 2010). Equations associated with the PBL model based on the mixed-layer theory were developed in the period from 1960 to 1980 (McNaughton and Spriggs, 1986; Tennekes and Driedonks, 1981). Here, I introduce basic equations used in a clear-sky PBL model (McNaughton and Spriggs, 1986).

A budget equation of a scalar ($c$) (e.g., temperature, humidity, and CO$_2$) averaged over the mixed-layer ($c_m$) can be written after the Reynolds decomposition as follows:

$$\frac{dc_m}{dt} = \frac{\overline{\omega c'}}{h} + \overline{(\omega c')}_e$$

(1)

where $h$ is the boundary layer height (m), $\overline{\omega c'}$ (unit of scalar $\cdot$ m s$^{-1}$) is a flux of $c$, and $s$ and $e$ denote the surface and the entrainment, respectively. Here, the equation relates the land surface process ($\overline{\omega c'}_s$) to $c_m$ in a very simple way.

When the land surface flux, $\overline{\omega c'}_s$, is provided either by observation or by an LSM, additional equations are needed for $\overline{\omega c'}_s$ and $h$. In order to obtain the equation for $h$, the first derivative of the difference in $c$ at the capping inversion ($c_c - c_m$) over time is introduced as follows:

$$\frac{d(c_c - c_m)}{dt} = \frac{dc_c}{dt} - \frac{dc_m}{dt}$$

(2)

The first term in the right-hand side (RHS) of Eq. (2) can be rewritten as

$$\frac{dc_c}{dt} \approx \frac{dc_c}{dz} w_e = rw_e$$

(3)

where $w_e$ is the entrainment velocity and $r$ is the vertical gradient of $c_c$ (unit of scalar $\cdot$ m$^{-1}$). The entrainment velocity ($w_e$) is defined as the balance between the boundary growth ($dh/dt$) and the mean vertical velocity ($w_v$). Equation (2) is, therefore, rewritten as

$$\frac{d(c_c - c_m)}{dt} = \gamma \left( \frac{dh}{dt} - w_v \right) - \frac{dc_m}{dt}$$

(4)

Under an ideal condition with a constant jump between the boundary layer and the entrainment ($c_e - c_m = \text{const}$.) and no mean vertical velocity ($w_v = 0$), combining Eqs. (1) and (4) yields

$$\frac{dh}{dt} = \frac{\overline{\omega c'}_s}{\gamma h}$$

(5)

Equation (1) is then reduced to

$$\frac{dc_m}{dt} = \frac{\overline{\omega c'}_s}{h}$$

(6)

Ikawa et al. (2018) used Eq. (6) for virtual potential temperature, the mixing ratio of water vapor, and the mixing ratio of CO$_2$ concentration, whereas Eq. (6) was used for virtual potential temperature, assuming that the entrainment fluxes were zero and $\gamma$ of the virtual potential temperature is 0.004 K m$^{-1}$. For some details and useful interpretations, I recommend Chapters 2 and 4 of Vilà-Guerau de Arellano et al. (2015) and Chapter 11 of Stull (1988).

8. Governing Equations for a Regional Atmospheric Model

8.1 General Equations

Here, I introduce how governing equations often used in a regional atmospheric model can be derived. Starting with the conservation of momentum (e.g., Stull, 1988),

$$\frac{\partial U_j}{\partial t} + U_i \frac{\partial U_i}{\partial x_j} = -\delta_{ij} + f_\epsilon \epsilon_{ij} - \frac{1}{\rho} \frac{\partial P}{\partial x_i} + \nu \frac{\partial^2 U_i}{\partial x_j^2}$$

(7)

where $U(U, V, W)$ is the wind vector on the cartesian coordinate ($x, y, z$), $g$ is the acceleration due to gravity (m s$^{-2}$), $f$ is the Coriolis parameter, $\rho$ is the air density (kg m$^{-3}$), $P$ is the pressure (Pa = kg m$^{-1}$ s$^{-2}$), $\nu$ is the kinematic viscosity (m$^2$ s$^{-1}$), $\delta$ is the Kronecker delta, and $\epsilon$ is the alternating unit tensor.

The Boussinesq approximation assumes that the air density varies linearly with temperature and that the density variation is important only in the buoyancy term (see also Eqs. 17 and 18). With that assumption, Eq. (7) can be expanded into mean and turbulent components:
\[
\frac{\partial (U_t + u_t)}{\partial t} + \left( U_t + u_t \right) \frac{\partial (U_t + u_t)}{\partial x_t} = -\delta_a g + f_e \varepsilon \left( U_t + u_t \right) - \frac{1}{\rho} \frac{\partial P_g}{\partial x_t} + \frac{\partial (U_t + u_t)}{\partial x_t^2} \rho \frac{\partial \theta_g}{\partial x_t} \tag{8}
\]

where \( \theta_g \) is the virtual potential temperature (K).

The Reynolds average yields

\[
\frac{\partial U_I}{\partial t} + U_I \frac{\partial U_I}{\partial x_t} = -\delta_a g - \delta_a \varepsilon + f_e \varepsilon U_I - \frac{1}{\rho} \frac{\partial P}{\partial x_t} + \frac{\partial U_I}{\partial x_t^2} \rho \frac{\partial \theta_g}{\partial x_t} \tag{9}
\]

Assuming that the fourth term in the RHS is much smaller than the fifth term,

\[
\frac{\partial U_I}{\partial t} + U_I \frac{\partial U_I}{\partial x_t} = -\delta_a g + f_e \varepsilon U_I - \frac{1}{\rho} \frac{\partial P}{\partial x_t} - \frac{\partial U_I}{\partial x_t^2} \rho \frac{\partial \theta_g}{\partial x_t} \tag{10}
\]

I rewrite the third term of the RHS in Eq. (10), introducing the spatiotemporal averaged potential temperature (\( \Theta \)) and the Exner function (\( \pi = T/\Theta = (P/P_0) \Theta \)). \( T \) is air temperature (K), \( P_0 \) is reference pressure (=1000 hPa), \( R \) is the gas constant of dry air (J kg\(^{-1}\) K\(^{-1}\)), and \( C_\rho \) is the specific heat at constant pressure (J kg\(^{-1}\) K\(^{-1}\)). The average air density (\( \rho \)) and the average Exner function \( \bar{U} \) vary vertically as \( \rho = P_0/(RT_0) \) \((T_0 = \Theta - z g / C_\rho\) and \( \bar{U} = T_0 / \Theta \). Taking a derivative of \( \pi \) with respect to \( P \)

\[
\frac{d\pi}{dP} = \frac{C_\rho P}{R \pi} \left( \frac{P}{P_0} \right) \tag{11}
\]

Rearranging Eq. (11),

\[
dP = \frac{C_\rho P}{R \pi} \frac{d\pi}{dP} \tag{12}
\]

Using \( 1/\rho = RT/P = R \theta/P \) and substituting Eq. (12) into the third term of the RHS in Eq. (10) yields

\[
\frac{1}{\rho} \frac{\partial P}{\partial x_t} = \frac{R \rho}{P} \frac{C_\rho P}{R \pi} \frac{d\pi}{dP} \tag{13}
\]

Here, I indicate departures from the reference point by \( dP \). The RHS of Eq. (13) hence becomes

\[
C_\rho \frac{d\pi}{dP} = C_\rho \frac{\partial (\bar{U} + \bar{\varepsilon})}{\partial x_t} = C_\rho \frac{\partial (\bar{U} + \bar{\varepsilon})}{\partial x_t} \tag{14}
\]

Finally, Eq. (10) can be written as

\[
\frac{\partial U_I}{\partial t} + U_I \frac{\partial U_I}{\partial x_t} = -\delta_a g + f_e \varepsilon U_I - C_\rho \frac{\partial (\bar{U} + \bar{\varepsilon})}{\partial x_t} \frac{\partial \pi}{\partial x_t} \tag{15}
\]

### 8.2 Hydrostatic Assumption

Although nonhydrostatic models are more common in recent atmospheric research (Saito et al., 2007), the hydrostatic assumption is still useful for a wide range of research questions with relatively easy computations (Kimura and Arakawa, 1983; Kuwagata et al., 1994; Watanabe and Shimoyama, 2015). For the z-component, assuming a hydrostatic condition \( \frac{\partial P}{\partial z} = -\rho g \),

\[
\frac{1}{\rho} \frac{\partial P}{\partial z} = C_\rho \frac{\partial \pi}{\partial z} + \frac{\partial \pi}{\partial z} = -g \tag{16}
\]

Rearranging and multiplying \( \rho = \rho_0 (1 + \rho/\rho_0) \) by Eq. (16) yields

\[
\rho_0 \left( 1 + \frac{\rho}{\rho_0} \right) \left[ \frac{\partial \pi}{\partial z} + \frac{\partial \pi}{\partial z} \right] = -\rho_0 \frac{\partial (\bar{U} + \bar{\varepsilon})}{\partial z} \frac{\bar{\pi}}{C_\rho \bar{U}} \tag{17}
\]

\( \rho_0 \) is air density of average The Boussinesq approximation neglects buoyancy \( \rho_0/\rho_0 \) in the LHS and assumes \( \rho_0/\rho_0 = \theta_0 / \Theta \). Therefore,

\[
\frac{\partial \pi}{\partial z} + \frac{\partial \pi}{\partial z} = -\frac{\partial (\bar{U} + \bar{\varepsilon})}{\partial z} \frac{\bar{\pi}}{C_\rho \bar{U}} \tag{18}
\]

For each \( x, y, \) and \( z \)-component, Eq. (15) with the hydrostatic assumption can be written as

\[
\frac{\partial U_I}{\partial t} + U_I \frac{\partial U_I}{\partial x_t} + V \frac{\partial U_I}{\partial y} + W \frac{\partial U_I}{\partial z} = -f_u V - C_\rho \frac{\partial \pi}{\partial x_t} \frac{\partial \pi}{\partial x_t} \tag{19}
\]

\[
\frac{\partial \pi}{\partial z} = -\frac{g \pi}{C_\rho \bar{U}} \tag{19}
\]

\[
\frac{\partial \pi}{\partial z} = -\frac{g \pi}{C_\rho \bar{U}} \tag{19}
\]

### 9. Final Remark

In this work, I reviewed certain topics where mutual influences between vegetation and atmospheric forcing are evident. Vegetation-atmosphere coupled models have been implemented in the area of hydrological cycles (Margulis and Entekhabi, 2001a; Santanello et al., 2013; van Heerwaarden et al., 2010), and these works have developed effective tools to investigate vegetation-atmosphere interactions. However, a significant challenge still remains in disentangling the complex interactions beyond very limited conditions where model simulations have been performed. It is evident that climate changes affect vegetation feedback to the atmosphere, particularly in sensitive areas, such as the northern high latitudes (Helbig et al., 2016; Ikawa et al., 2015), and future investigations are anticipated to quantitatively evaluate the magnitude of the changes in
the vegetation feedback in such sensitive areas under climate change conditions. Crop yield, water use, and heat environment are interrelated sensitively to the feedback in areas like Southeast Asia where irrigated crop fields account for a major fraction of the land surface (Ikawa et al., 2018; Kuwagata et al., 2018, 2014; Yoshida et al., 2012; Yoshimoto et al., 2011). Typically, heterogeneous land surface characteristics make a fully coupled model simulation challenging in these areas. The continuous collection of meteorological and yield data is important. Finally, a vegetation-atmosphere coupled system is also influenced by the conditions at the top of and synoptic meteorology beyond the PBL (Combe et al., 2015; Davis et al., 1997; Shinoda and Uyeda, 2002; Sikma et al., 2017; Vilà-Guerau de Arellano et al., 2014).

A challenge still remains in the fact that the spatial and temporal scales of observations for the atmosphere are often much greater than those for the Earth surface, and their representations are mismatched. The majority of recent studies on vegetation-atmosphere interaction research are, therefore, based on models. However, our recent work suggests that buoyancy surface flux is very sensitive to even slight changes in plant water use (Ikawa et al., 2018), and further reinforcement of the current LSM based on observations may be necessary in order to accurately evaluate the effects of vegetation feedback to the atmosphere. With the recent development of an observation network, such as FLUXNET, long-term eddy covariance sites (e.g., Figs. 1 and 2) that accommodate interdisciplinary fields of research will play an important role in future research on vegetation-atmosphere interaction research.

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