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Key Points:
- Atmospheric boundary layer stability above lakes varies predictably across temporal and spatial scales
- The atmospheric boundary layer above lakes is more often unstable toward the tropics annually and above smaller lakes during summer
- The effects of latitude and lake size on lake-atmosphere interactions have implications for heat loss from lakes and the hydrologic cycle

Supporting Information:
- Supporting Information S1

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Latitude and lake size are important predictors of over-lake atmospheric stability

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Abstract Turbulent fluxes across the air-water interface are integral to determining lake heat budgets, evaporation, and carbon emissions from lakes. The stability of the atmospheric boundary layer (ABL) influences the exchange of turbulent energy. We explore the differences in over-lake ABL stability using data from 39 globally distributed lakes. The frequency of unstable ABL conditions varied between lakes from 71 to 100% of the time, with average air temperatures typically several degrees below the average lake surface temperature. This difference increased with decreasing latitude, resulting in a more frequently unstable ABL and a more efficient energy transfer to and from the atmosphere, toward the tropics. In addition, during summer the frequency of unstable ABL conditions decreased with increasing lake surface area. The dependency of ABL stability on latitude and lake size has implications for heat loss and carbon fluxes from lakes, the hydrologic cycle, and climate change effects.

1. Introduction

Temperature gradients between lake water surfaces and air influence the exchange of heat, gas fluxes, and evaporation from lakes. Lake surface temperatures are governed by the interactions with the overlying atmosphere and radiative fluxes [*Schmid et al.*, 2014]. Lake surface temperature exceeding surface air temperature tends to produce an unstable atmospheric boundary layer (ABL), which is the lower layer of the troposphere that is influenced directly by the Earth's surface [*Garratt*, 1994]. In general, the higher the lake surface temperature is relative to the air temperature, the more unstable the ABL, although other factors play a role as well. In turn, the thermodynamic interaction of lakes with the atmosphere is influenced by the stability of the ABL.

Unstable ABL conditions can result in enhanced heat loss (where heat loss is defined as heat transport from lakes to the atmosphere) by sensible and latent heat transfer from the lake surface to the atmosphere [*Brutsaert*, 1982], thereby influencing the local climate [*Bonan*, 1995; *Lofgren*, 1997], altering hydrological cycles at catchment scales [*Rouse et al.*, 2005], and enhancing lake evaporation and water level decline [*Gronewold and Stow*, 2014], with consequences for water security [*Vörösmarty et al.*, 2000] and supply [*Brookes et al.*, 2014]. This is of significance to lakes which are an important supply of national water demand and can have considerable consequences for water management strategies [*Vörösmarty et al.*, 2000; *Immerzeel et al.*, 2010; *Vörösmarty et al.*, 2010]. The turbulent heat and moisture fluxes can also enhance the gas transfer coefficient and thereby the emission of carbon dioxide and methane from lakes [*Polsenaere et al.*, 2013; *Podgrajsek et al.*, 2015].

Unstable ABL conditions have been shown to persist above lakes for long periods [*Rouse et al.*, 2003], resulting in enhanced turbulent heat loss. ABL conditions have been reported in lake studies from around the world, in tropical [*Verburg and Antenucci*, 2010] and temperate regions [*Derecki*, 1981; *Lofgren and Zhu*, 2000; *Laird and Kristovich*, 2002; *Rueda et al.*, 2007] but no study has compared ABL stability across a range of latitudes and other lake attributes.

In view of increasing temperature gradients reported between lake surfaces and air in the past few decades as a result of climate warming [*O'Reilly et al.*, 2015; *Woolway and Merchant*, 2017], and the acceleration expected in the hydrologic cycle [*Wentz et al.*, 2007], an understanding of the extent, causes, and effects of variations in the stability of the ABL above lakes is of great interest. However, until recently, in situ high-frequency measurements at the air-water interface of wind speed, water temperature, air temperature, and relative humidity, that are required to examine patterns in the stability of the ABL, have not been widely available, preventing a comprehensive comparison across lakes.

The recent establishment of scientific networks dedicated to the collaborative analysis of high-frequency lake buoy data has provided an opportunity for a large-scale analysis. We have collated data from 39 lakes from around the globe to quantify variability and identify patterns in annual, seasonal, and diurnal ABL conditions above the water surface and analyze how these patterns vary across different lake types and climatic gradients. To understand the overarching controls on ABL stability, we examine the influence of four variables that might be expected to have an effect, latitude, altitude, lake surface area, and lake depth. Latitude determines variation in net radiation flux and thus has a strong influence on lake temperature and the seasonal temperature cycle [*Straskraba*, 1980] and ultimately on ABL stability [*Verburg and Antenucci*, 2010]. Altitude can influence air-water temperature relationships via differential lapse rates [*Livingstone et al.*, 1999] and can thus be expected to influence ABL stability [*Rueda et al.*, 2007]. Lake area has been shown to affect lake temperatures at diurnal timescales [*Woolway et al.*, 2016]. Lake depth influences the thermal capacity of water and thus air-water temperature relationships [*Balsamo et al.*, 2010; *Woolway and Merchant*, 2017].

2. Materials and Methods

We collected high-frequency observations (measurement interval range from 4 min to 1 h) of lake surface temperatures and meteorological conditions from 39 lakes on 5 continents, ranging in surface area between 0.001 km^2 and 32,900 km^2, in altitude between 0 m above sea level (asl) and 1897 m asl, in latitude between 38.8°S and 71.24°N, and in maximum depth between 2 m and 1470 m (Table S1 in the supporting information). Instrumented buoys measured near-surface water temperatures (T_0, K) at an average depth of about 0.5 m, always within the surface mixed layer. In this study, we assume that the measured water temperature equals that at the surface, although it must be noted that under certain conditions (e.g., under intense heating conditions) temperatures can vary considerably within the upper meter of a lake [*Tedford et al.*, 2014]. Meteorological conditions including wind speed (u_z, m s^{-1}), air temperature (T, K), and relative humidity (RH, %) were measured on average 2.9 m (range 1.3 to 10 m) above the lake surface. Fourteen lakes had observations available throughout at least 1 year (i.e., 12 months of observations). All lakes had observations for the summer months (defined as July to September in the Northern Hemisphere and January to March in the Southern Hemisphere) for at least 1 year. Each lake had measurements at one location, except for Lake Tanganyika (two locations) and Lake Tahoe (four locations). We analyzed the data from each monitoring station in these two lakes independently before combining the results.

Stability of the atmospheric boundary layer is characterized by the nondimensional atmospheric stability parameter, determined from surface measurements which we obtained from the lake buoy monitoring stations. It is given by $\zeta = zL^{-1}$ [*Brutsaert*, 1982], which is the ratio of the sensor measurement height above the water surface (z; in meters) to the Monin-Obukhov stability length

$$L = \frac{-\rho_a u_{*a}^3 T_v}{\kappa g \left(\frac{Q_h}{C_{pa}} + 0.61 \frac{T Q_e}{L_v} \right)} \tag{1}$$

where the absolute value of L is the height above ground level above which buoyancy-generated turbulence dominates wind shear-generated turbulence, ρ_a is air density (kg m^{-3}), u_{*a} is the air friction velocity (m s^{-1}), $T_v = T(1 + 0.61q_z)$ is virtual air temperature (K), which we assume to be approximate to the virtual potential temperature and thus are neglecting the influence of adiabatic expansion and compression on T_v [*MacIntyre et al.*, 2002; *Hipsey et al.*, 2014], q_z is the specific humidity (kg kg^{-1}) calculated from relative humidity, air temperature, and air pressure, κ = 0.41 is the von Karman constant, g = 9.81 m s^{-2} is the gravitational acceleration, Q_h and Q_e are the sensible (equation (2)) and latent (equation (3)) heat fluxes, respectively, positive when heat flux is from the lake surface to the atmosphere (W m^{-2}), C_{pa} = 1005 J kg^{-1} K^{-1} is the specific heat of air at constant pressure, and L_v is the latent heat of vaporization (J kg^{-1}). A negative ζ indicates an unstable ABL, a convective atmosphere above the lake, and increased potential for heat loss from lakes via turbulent heat fluxes. The opposite is the case when ζ is positive and the ABL is stably stratified. ABL conditions are neutral when $\zeta = 0$. We analyzed the incidence of an unstable ABL as the percentage of time in which $\zeta < 0$, using hourly data, and examined the incidence against lake attributes using a multiple regression model.

Air pressure was not measured on all instrumented buoys, and since local variability in air pressure has negligible effect on the estimation of ABL stability, transfer coefficients, and turbulent fluxes [*Verburg and Antenucci*, 2010], a constant air pressure was assumed for each site, based on the altitude of the lake [*Woolway et al*., 2015a]. With the exception of air pressure, all data used to estimate the stability of the ABL were measured directly above each lake, as opposed to over land. Previous work has often relied on overland weather data [e.g., *Derecki*, 1981; *Croley*, 1989; *Rueda et al*., 2007].

Sensible heat and latent heat fluxes were estimated with bulk aerodynamic methods [*Brutsaert*, 1982]

$$Q_h = \rho_a C_{pa} C_h u_Z (T_o - T) \tag{2}$$

$$Q_e = \rho_a L_V C_e u_Z (q_s - q_z) \tag{3}$$

where q_s is the specific humidity at saturation (kg kg^{-1}). The turbulent transfer coefficients for heat (C_h) and humidity (C_e), which regulate Q_h and Q_e energy exchanges, were assumed to be equal and adjusted for ABL stability by applying stability functions [*Verburg and Antenucci*, 2010]. Estimates of the surface turbulent fluxes of Q_h and Q_e taking into account varying ABL conditions were compared with estimates that assume ABL neutrality to elucidate the role of ABL stability in regulating air-water exchanges.

The air-water surface temperature difference ($\Delta T = T - T_0$) has been used as a measure of ABL stability in previous studies [*Croley*, 1989; *Derecki*, 1981]. However, stability of the ABL is more accurately related to the virtual air-water temperature gradient [*Verburg and Antenucci*, 2010], which is estimated as the difference between T_v and the virtual temperature of saturated air at the water surface (T_{0v}, K): $\Delta T_v = T_v - T_{0v}$. When $\zeta < 0$ (unstable ABL), then $\Delta T_v < 0$ is also true (Figure S1), while $T - T_0$ can be several degrees higher [*Verburg and Antenucci*, 2010]. Therefore, in addition to ζ, we use ΔT_v as an alternative measure of ABL stability and estimate the incidence of an unstable ABL as the percentage of time in which $\Delta T_v < 0$. As the meteorological measurement heights vary among lakes, we converted T and q_z to a surface elevation of 10 m (T_{10} and q_{10}, respectively) following *Zeng et al*. [1998], using the algorithms of *Woolway et al*. [2015a], prior to calculating ΔT_v. In addition, to ensure that ABL conditions that were close to neutral were not inferred as unstable as a result of measurement uncertainty, we included an uncertainty threshold to our estimates of percent unstable conditions by determining the prevalence of unstable ABL conditions as the proportion of time during which $\Delta T_v < -0.5$ K.

All statistical analyses in this study were performed in *R* [*R Development Core Team*, 2014], including bilinear regressions computed with the "Segmented" package [*Vito and Muggeo*, 2008], used to relate the frequency of unstable ABL conditions to lake surface area.

To evaluate further the influence of latitude on ABL stability, we also examined incoming solar radiation (i.e., shortwave; Q_{sw}) and thermal radiation (i.e., longwave; Q_{lw}). A number of lakes (n = 28) had in situ measurements of Q_{sw}, but to be consistent among all lakes, we used Q_{sw} and Q_{lw} estimates from the ERA-Interim reanalysis product [*Dee et al*., 2011], available at a spatial resolution of 0.75°. Data were extracted for the grid point situated closest to the center of each lake, and the reanalysis data were verified using in situ measurements when available. To estimate net incoming longwave radiation, $Q_{lwin} = (1 - \alpha_{lw})Q_{lw}$, 3% of Q_{lw} was assumed to be reflected at the lake surface (α_{lw} = 0.03) [*Brutsaert*, 1982]. Emitted, outgoing longwave radiation was estimated as $Q_{lwout} = 0.972\sigma T_0^{\,4}$, where σ is the Stefan-Boltzmann constant (= 5.67 × 10^{-8}). The shortwave albedo (α_{sw}), estimated from Fresnel's equation [*Woolway et al*., 2015a], was used to account for the amount of Q_{sw} reflected at the lake surface and to estimate the net incoming shortwave radiation: $Q_{swin} = (1 - \alpha_{sw})Q_{sw}$. These surface radiative fluxes were then used to calculate net radiation, as $Q_{net} = Q_{swin} + Q_{lwin} - Q_{lwout}$.

3. Results

The ABL over lakes was unstable ($\zeta < 0$ and $\Delta T_v < 0$) 82% of the time on an annual basis (n = 14) averaged across this globally distributed set of lakes (Figure 1 and Table S1). However, the stability of the ABL varied at both seasonal and diurnal timescales and differed between lakes. Unstable ABL conditions were most prevalent in summer (91% of time) and least common in spring and winter (74% and 70% of time, respectively; Figures 1b and S2). Since an unstable ABL can enhance surface heat loss considerably [*Verburg and Antenucci*, 2010] this suggests seasonal differences in heat transfer efficiency [*Lofgren and Zhu*, 2000; *Rouse et al.*, 2003; *Woolway et al.*, 2015b]. Unstable ABL conditions were most common during the late evening to early morning hours (Figures 1c and S3), when air temperatures tend to be cooler while the surface temperature of the lake retains daytime heat longer. In addition, at night, winds are often less intense, producing less mixing of the lower ABL (Figures 1c and S4). The magnitude of the diurnal cycle in ABL stability also varied among the seasons, similar to that of T_0 [*Woolway et al.*, 2016] with greatest diurnal variability in spring (Figure S3).

For lakes with data available throughout the year (n = 14), a multiple linear regression model including latitude, altitude, lake surface area, and lake depth explains 94% of the variation in annual mean frequency of occurrence of unstable ABL conditions ($\zeta < 0$, $\Delta T_v < 0$, Table S2 and 95% for $\Delta T_v < -0.5$, Table S3). Latitude was the most important covariate (R^2 = 0.81, p = 0.001), and lake surface area was also significant ($p < 0.1$). The ABL was most commonly unstable near the equator (Figure 2a), with $\zeta < 0$ more than 99% of the year in Lake Tanganyika. Unstable conditions decreased with increasing latitude, to $\zeta < 0$ occurring about 72% of the year for lakes situated in the English Lake District. Q_{net} increases with decreasing latitude (R^2 = 0.93, $p < 0.001$). Q_{net} correlated significantly with T_{0v} (R^2 = 0.88, $p < 0.01$) and also with ΔT_v (R^2 = 0.69, $p < 0.01$), with a more negative ΔT_v at higher Q_{net}. ΔT_v decreased (i.e., became increasingly negative) at a rate of 0.28 K per 10 W m^{-2} increase in Q_{net}. Similarly, ΔT_v decreased at a rate of 0.49 K per 10° decrease in latitude, meaning that while T_v increases toward the tropics, the increase in T_{0v} is larger, as a result of higher Q_{net} received by tropical lakes.

When restricting the analysis to summer, latitude and lake surface area were statistically significant predictors as well of the frequency of unstable ABL conditions (n = 39, Tables S2 and S3), with latitude being the stronger factor. The relationship between latitude and ABL stability was strongest among the larger lakes (>10 km^2) during summer (R^2 = 0.70, n = 18; Figure 2b) with, similar to the annual data, a higher incidence of unstable ABL conditions at lower latitude.

Assuming neutral turbulent transfer coefficients to estimate the transfer efficiency to and from the atmosphere of vapor and heat resulted in a mean annual (n = 14) underestimate of 8.6 W m^{-2} (12% difference) and 3.4 W m^{-2} (28% difference) in estimates of Q_e and Q_h, respectively, compared to transfer coefficients adjusted for ABL stability. This bias in Q_e increased with decreasing latitude (R^2 = 0.65, $p < 0.001$; Figure 2c). The bias was even greater, an average 14.0 W m^{-2} and 4.4 W m^{-2} for Q_e and Q_h, respectively, if a constant, such as 1.35×10^{-3} [*Hicks*, 1972, 1975] was assumed for the transfer coefficients C_e and C_h, as opposed to transfer coefficients accounting for ABL stability. Moreover, in summer (n = 39), there was a mean difference of 12.7 W m^{-2} (16% difference) and 4.1 W m^{-2} (28% difference) for Q_e and Q_h, respectively, between estimates assuming neutral transfer coefficients and those which account for ABL stability, with largest differences at low latitudes (R^2 = 0.56; Figure 2d). Similar to the annual data, the bias was greater, 21.5 W m^{-2} and 5.5 W m^{-2} for Q_e and Q_h, respectively, in summer if constant transfer coefficients were assumed. A difference of 27 W m^{-2} in Q_e and Q_h combined equates to a 22% change in the surface energy budget as given by the summer average Q_{net} among the study lakes (121.9 W m^{-2}).

The transfer coefficients for momentum (C_d), and water vapor and heat ($C_e = C_h$), which describe the transfer efficiency to and from the atmosphere, are higher when the ABL is unstable [*Verburg and Antenucci*, 2010] and when wind speeds are low, such as during summer (Figures S4 and S5). Mean wind speed (u_{10}) in summer correlated significantly with lake surface area (R^2 = 0.36, $p < 0.001$), but not with latitude or altitude ($p > 0.1$), suggesting an effect of lake fetch. A strong lake size dependence also existed for the mean transfer coefficients (Figure 2e and Table S4). We also found a weak but statistically significant relationship (R^2 = 0.12, p = 0.04) between summer average wind speed and the percentage of time in which the ABL was unstable ($\zeta < 0$ and $\Delta T_v < 0$; Figure S6). The greatest change in the transfer coefficients following accounting for ABL

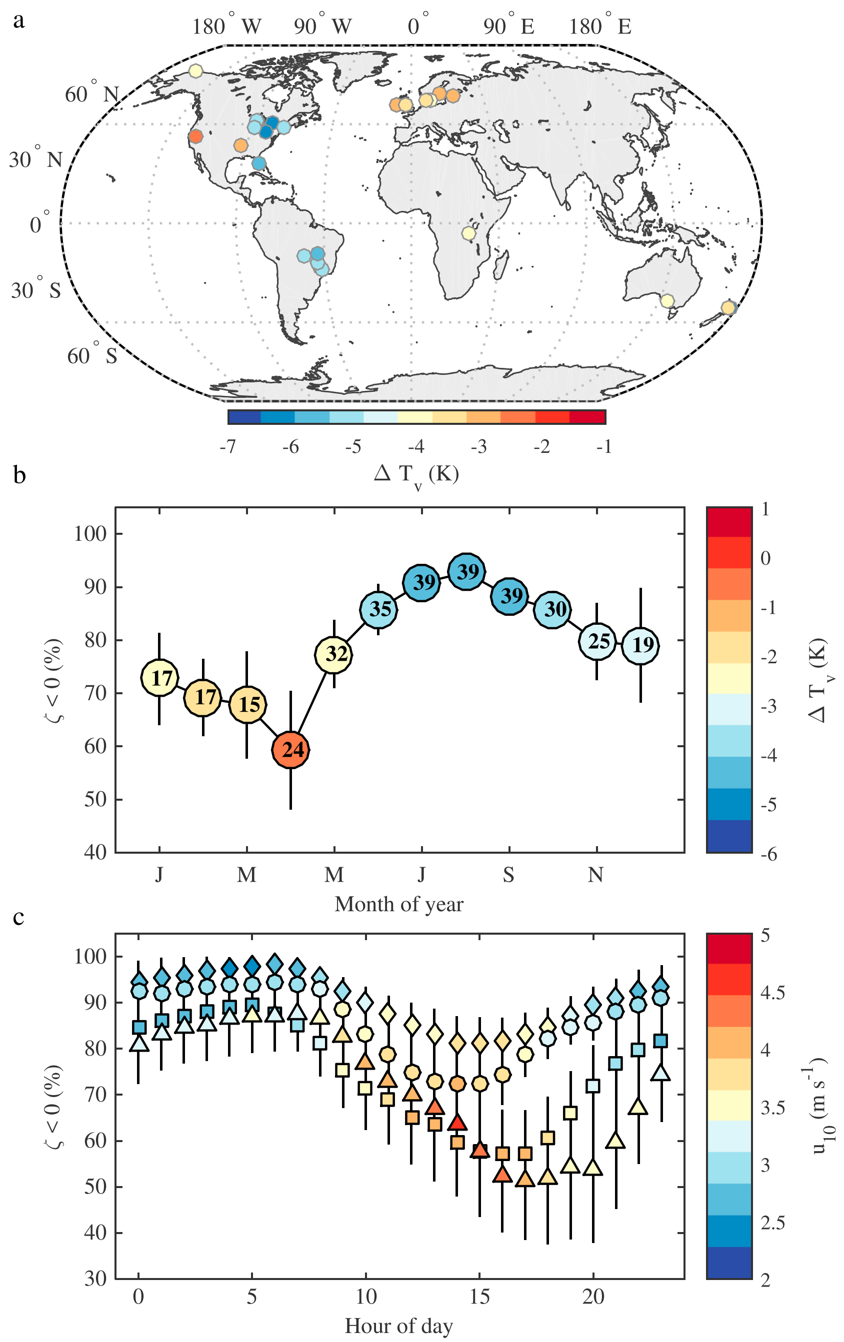

Figure 1. Comparisons of atmospheric stability above lakes across temporal and spatial scales. (a) Location of each lake, colored according to the summer (July–September in the Northern Hemisphere and January–March in the Southern Hemisphere) average air-water virtual temperature difference ($\Delta T_v = T_v - T_{0v}$). (b) Monthly averaged percentage of time in which $\zeta < 0$, with confidence intervals, as well as with colors indicating mean ΔT_v. The number of lakes with available data for each month contributing to the average is indicated. Monthly average percentage of time in which $\zeta < 0$ in lakes with data available throughout the year ($n = 14$) is shown in Figure S2. (c) Average diurnal cycles of the percentage of time in which $\zeta < 0$, for winter (January, February, March (JFM), triangles, $n = 16$), spring (April, May, June (AMJ), squares, $n = 30$), summer (July, August, September (JAS), diamonds, $n = 39$), and autumn (October, November, December (OND), circles, $n = 24$) months, with 95% confidence intervals. Colors represent the average diurnal cycles in surface wind speed adjusted to a height of 10 m (u_{10}). Southern Hemisphere lakes were shifted by 182 days.

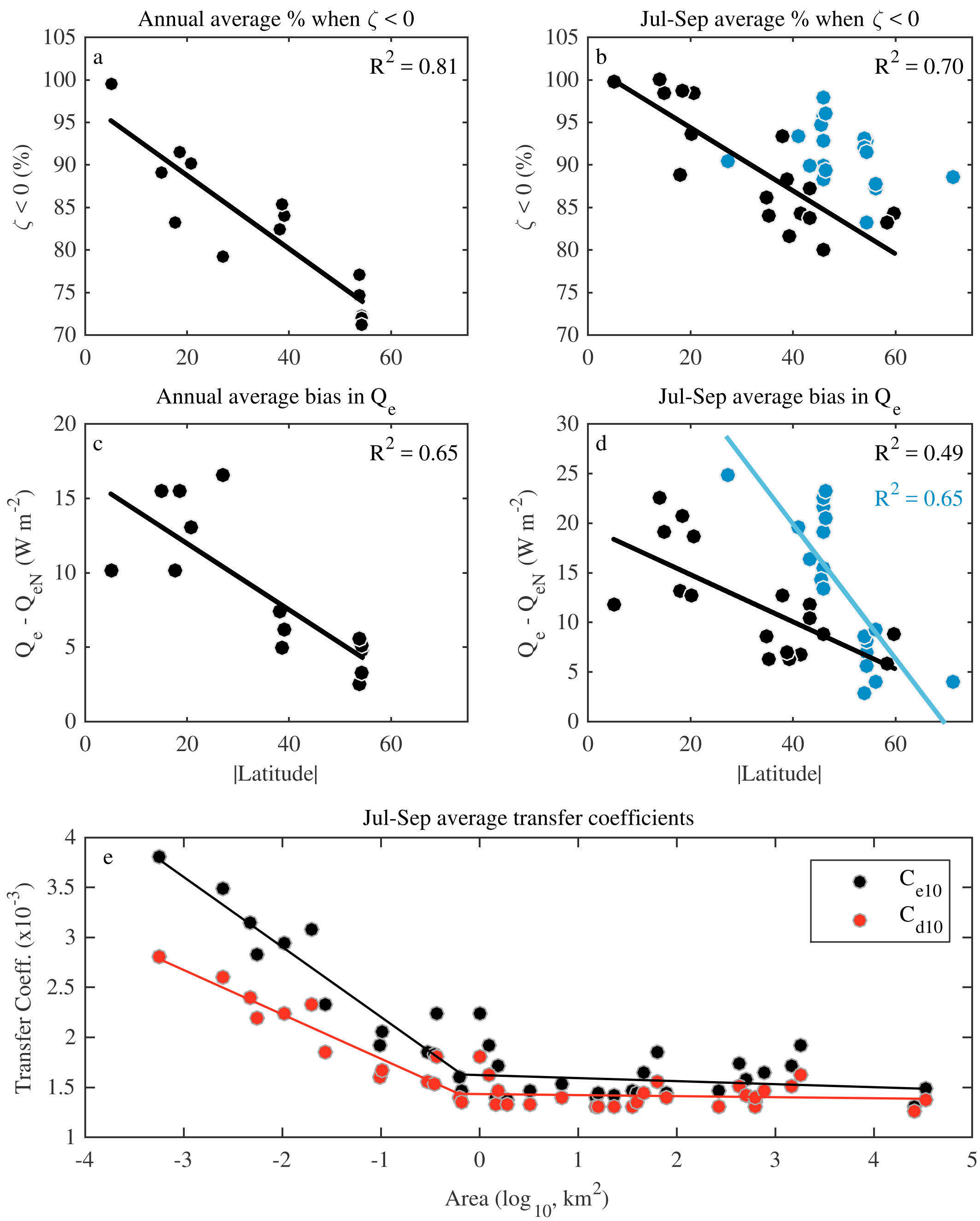

Figure 2. Relationships between lake location and size, atmospheric stability, and the accurate estimation of turbulent energy fluxes. Shown are the relationships between latitude (shown as absolute latitude) and (a) the annual percentage of time in which $\zeta < 0$ ($n = 14$), (b) the summer percentage of time in which $\zeta < 0$ ($n = 39$), (c) the annual average difference in the latent heat flux (Q_e) which arises by not accounting for atmospheric stability in the estimation of transfer coefficients and (d) the summer average difference in Q_e which arises by not accounting for atmospheric stability in the estimation of transfer coefficients. For the summer data (Figures 2b and 2d), a linear fit is shown for the lakes (black) with surface area > 10 km^2 ($p < 0.1$). A linear fit is also shown for the smallest lakes (<10 km^2; blue circles) where significant ($p < 0.1$). R^2 values are shown for each linear fit. (e) Relationship between lake surface area and the transfer coefficients (adjusted to a height of 10 m) for the latent heat flux (C_{e10}) and momentum (C_{d10}), accounting for atmospheric stability. All values are shown as summer averages. A bilinear trend is shown. Coefficient values are in Table S4.

stability was observed in lakes $< \sim 1\ \mathrm{km}^2$ (Figure 2e). These results demonstrate that lake surface area and its effect on ABL stability affects C_e and C_d and that the stability of the ABL has a greater influence on the transfer coefficients in smaller lakes during summer. Latitude showed no significant effect on the transfer coefficients ($p > 0.1$). However, after removing the lake area effect (i.e., the bilinear trends in Figure 2e), C_{e10} (adjusted to a height of 10 m, $R^2 = 0.19$, $p = 0.006$) and C_{d10} ($R^2 = 0.16$, $p = 0.01$) were both significantly correlated with latitude, consistent with the aforementioned effect of Q_{net} on ABL stability.

4. Discussion

This global-scale analysis has identified latitude and lake surface area as significant predictors of the variability in ABL stability over lake surfaces. Previous studies have shown evidence of a high incidence of unstable ABL conditions and its influence on the turbulent energy fluxes in individual lakes [*Verburg and Antenucci*, 2010; *Lorenzzetti et al*., 2015; *Yusup and Liu*, 2016]. This paper investigated the effects of geographic location and lake morphometry on ABL stability, the turbulent transfer coefficients, and on energy fluxes across the air-water interface. High-frequency data recorded on lakes showed variability in ABL stability on diurnal, seasonal, and annual timescales. Our results reveal that the ABL was unstable on average $> 80\%$ of the year across the examined lakes and that a strong latitudinal and lake size dependence exists for the incidence of unstable ABL conditions. An unstable ABL is more common above low-latitude lakes on seasonal and annual timescales and above small lakes during summer.

Lake size can affect lake temperatures (and the overlying atmosphere) and thereby the ABL and lake-air interactions. Small lakes are often well sheltered from the wind and experience lower wind speeds than large lakes with greater fetch [*Hondzo and Stefan*, 1993], resulting in less mechanical mixing of the lower ABL and, consequently, a higher incidence of unstable ABL ($\zeta < 0$) conditions. Moreover, small lakes tend to have higher dissolved organic carbon concentrations [*Hanson et al*., 2007], resulting in more stratified distributions of thermal energy, higher lake surface temperatures, and more negative ΔT_v due to higher attenuation of light in the water column [*Read and Rose*, 2013]. The effects of high dissolved organic carbon concentrations in small lakes on lake surface temperature and on ABL stability would be especially notable in summer when the lake water column is stratified. This is consistent with the relationship between lake size and the frequency of unstable conditions, which was negative during summer but not the rest of the year. The transfer coefficients C_d and C_e increased sharply as lake surface area fell below 1 km^2. This lake size threshold is similar to that reported for transitions in the relative contribution of convective mixing in the gas transfer coefficient [*Read et al*., 2012] and in the magnitude of diurnal heating and cooling in surface waters [*Woolway et al*., 2016].

Net radiation received by lakes is higher in the tropics, due to higher Q_{swin} and Q_{lwin}, and lower shortwave albedo. The latter is a result of a smaller zenith angle at low-latitude lakes [*Cogley*, 1979], thus increasing the proportion of shortwave radiation being absorbed annually. Higher Q_{net} likely explains why, while lake water temperatures are typically higher than air temperatures in most lakes, they are more so in tropical lakes, on an annual average. ΔT_v becomes increasingly negative toward the tropics indicating a more frequently unstable ABL.

Comparisons of turbulent energy fluxes calculated without taking ABL stability into account with those calculated by applying stability functions demonstrate substantial differences. Assuming neutral transfer coefficients instead of accounting for ABL stability in the estimation of turbulent transfer coefficients resulted in evaporation being underestimated on average by 16% during summer. Note that this is not a comparison with turbulent energy fluxes in a truly neutral ABL ($\zeta = \Delta T_v = 0$), because then Q_e and Q_h are close to zero. Many lake models that use a physically based approach [*Imberger et al*., 1978; *Imberger and Patterson*, 1981; *Hipsey et al*., 2014] apply fixed transfer coefficients [*Fischer et al*., 1979] when simulating lake temperatures at daily or longer timescales, as it is often assumed that only at diurnal timescales is the thermal inertia of water too great to use constant values. However, our results illustrate that on seasonal and annual timescales C_e and C_d can be over twice as large as assumed fixed values, particularly for small lakes (Figure 2e). This challenges the validity of neglecting the effect of ABL stability on transfer coefficients by using neutral coefficients or, more generally, of applying constant transfer coefficients at seasonal and longer timescales. The inclusion of variable ABL conditions in the formulation of turbulent transfer coefficients has consequences for the simulation of heat transfer, and thus lake thermal dynamics, which are fundamental to understanding lake biogeochemistry and ecology and the role of lakes in the global carbon cycle.

Variable ABL conditions, and the accurate estimation of their effect on the turbulent energy fluxes, are likely to influence estimates of rates of gas transfer, in particular, at lower latitudes [*Erickson*, 1993], where lakes generally have persistent carbon dioxide supersaturation [*Cole et al.*, 1994; *Marotta et al.*, 2009; *Marotta et al.*, 2014]. In addition, small lakes, which have recently been demonstrated to be important in the study of global and regional processes including carbon and nutrient cycling [*Cole et al.*, 2007], vastly outnumber large lakes globally [*Downing et al.*, 2006; *Verpoorter et al.*, 2014]; thus, our findings of the effect of lake size on ABL stability are important when evaluating the role of lakes in global cycles.

Given the importance of ABL stability in influencing the exchange of water and energy at the air-water interface, our findings highlight the importance of lake location and size for evaluating the role of lakes in the Earth's hydrologic cycle, which is expected to accelerate with climate change [*Wentz et al.*, 2007; *Wu et al.*, 2013]. Our results show that water surface temperatures are on average higher than air temperatures. While air temperature increases toward the tropics, the increase in lake surface temperature is larger. Climate warming will likely increase the destabilization of the ABL, as suggested by the observation that temperatures in many lakes have increased more than air temperatures in the past few decades [*Desai et al.*, 2009; *O'Reilly et al.*, 2015; *Woolway and Merchant*, 2017]. Such enhanced lake-air temperature gradients by climate warming are likely to result in enhanced heat loss, gas fluxes, and evaporation from lakes.

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