Evaporation Driven by Atmospheric Boundary Layer Processes over a Shallow Saltwater Lagoon in the Altiplano

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ABSTRACT: Observations over a saltwater lagoon in the Altiplano show that evaporation $E$ is triggered at noon, concurrent to the transition of a shallow, stable atmospheric boundary layer (ABL) into a deep mixed layer. We investigate the coupling between the ABL and $E$ drivers using a land–atmosphere conceptual model, observations, and a regional model. Additionally, we analyze the ABL interaction with the aerodynamic and radiative components of evaporation using the Penman equation adapted to saltwater. Our results demonstrate that nonlocal processes are dominant in driving $E$. In the morning, the ABL is controlled by the local advection of warm air ($\sim 5$ K h$^{-1}$), which results in a shallow ($\sim 350$ m), stable ABL with virtually no mixing and no $E$ ($\sim 0.5$ W m$^{-2}$). The warm-air advection ultimately connects the ABL with the residual layer above, increasing the ABL height $h$ by $\sim 1$ km. At midday, a thermally driven regional flow arrives to the lagoon, which first advects a deeper ABL from the surrounding desert ($\sim 1500$ m h$^{-1}$) that leads to an extra $\sim 700$-m $h$ increase. The regional flow also causes an increase in wind ($\sim 12$ m s$^{-1}$) and an ABL collapse due to the entrance of cold air ($\sim 2$ K h$^{-1}$) with a shallower ABL ($\sim 350$ m h$^{-1}$). The turbulence produced by the wind decreases the aerodynamic resistance and mixes the water body releasing the energy previously stored in the lake. The ABL feedback on $E$ through vapor pressure enables high evaporation values ($\sim 450$ W m$^{-2}$ at 1430 LT). These results contribute to the understanding of $E$ of water bodies in semiarid conditions and emphasize the importance of understanding ABL processes when describing evaporation drivers.

KEYWORDS: Evaporation; Atmosphere-land interaction; Advection; Complex terrain; Salinity

1. Introduction

Endorheic basins are closed drainage systems isolated from external water bodies (Gao et al. 2018). These basins are predominantly found in semiarid and arid regions, where saline lakes are a common feature (Yapiev et al. 2017; Alcocer and Hammer 1998). While being the main water resource, these saline lakes also serve as the main discharge mechanism from the basins through evaporation $E$ processes (Wang et al. 2019; Yapiev et al. 2017). Therefore, studying $E$ over saline lakes in arid endorheic basins is crucial for water management in dry environments.

The Chilean Altiplano is a region in the Atacama Desert composed by multiple endorheic basins immersed in a complex topography (Fig. 1). These arid basins are predominantly characterized by desert surfaces (>100 km) in which small-scale heterogeneities (~10 km) can be found in the form of salt flats with shallow, saltwater lagoons and wetlands (Lobos-Roco et al. 2022a). The Salar del Huasco, shown in Fig. 1, is an example of these basins. Controlled by groundwater and a strong seasonality of precipitation, these localized environments act as preferential pathways for $E$ (Suárez et al. 2020; de la Fuente and Niño 2010; Johnson et al. 2010). Therefore, understanding the processes that control $E$ in the saltwater lagoons is essential for water balance predictions in the Altiplano as a whole, and it is necessary to understand the impacts of climate change on the region (Blin et al. 2022).

Describing evaporation processes in the Altiplano is challenging as, in addition to the local saline lake and desert conditions, the atmospheric boundary layer (ABL) dynamics and its interaction with large-scale forcing also play a key role in regulating $E$. Based on observations gathered in a field experiment (E-DATA) at the Salar del Huasco (Suárez et al. 2020), the latent heat flux (LE) was reported to be virtually zero during the morning (Fig. 2a) with turbulence as a limiting factor due to the absence of wind (Fig. 2b) (Lobos-Roco et al. 2021). Under these conditions, a shallow boundary layer characterized by an ABL height $h$ of less than 350 m above ground level (AGL) is formed over the lagoon (Fig. 2d). In the afternoon, evaporation is triggered by the entrance of a regional flow that makes LE to go from almost $\sim 0$ to $\sim 450$ W m$^{-2}$ in 2 h. The high $E$ rates are fed in part by the energy provided by the ground heat flux $G$, which becomes negative at 1700 LT (Fig. 2a). This regional flow is a well-studied, thermally driven, and topographically enhanced atmospheric circulation characterized by strong winds that arrive to the Altiplano every afternoon.
Lobos-Roco et al. 2021; Falvey and Garreaud 2005; Rutllant et al. 2003). Simultaneously to the entrance of the regional flow that triggers $E$, the shallow boundary layer over the water grows by more than 1 km into a deep mixed layer ($h \sim 1600$ m AGL, Fig. 2e) similar to the one observed over the surrounding desert. The ABL afterward collapses gradually until the end of the day.

Aguirre-Correa et al. (2023) studied the ABL collapse over the desert surface located upwind the saline lagoon in the Salar del Huasco. They found that the advection of a shallower boundary layer and cold air when the regional flow arrives, plus subsidence motions characteristic of the region, counteract the turbulence that is driven by the high sensible heat flux $H$ over the desert and lead to a decrease in $h$ already at midday. However, no study has been performed to understand how these nonlocal processes influence the ABL dynamics over the water surface and, more specifically, how the ABL processes interact with evaporation. This interaction is especially relevant for $E$ over water bodies in arid regions (Su et al. 2020; Wu et al. 2019; Hamdani et al. 2018; Verburg and Antenucci 2010). We will therefore address the following research question: What are the main drivers of the atmospheric boundary layer and evaporation over a shallow, saltwater lagoon in the Altiplano region?

2. Methods

To answer our research question, we have selected the shallow saltwater lagoon in the Salar del Huasco (Fig. 1), as it is a representative site that encompasses the main characteristics of the interaction between the Pacific Ocean, the lake, and the surrounding desert in the Altiplano. We use a land–atmosphere conceptual model to break down the complexity of the regional processes and quantify their individual contribution to both the ABL and $E$ over the water. The land–atmosphere model requires prescribing the nonlocal processes, such as advection and subsidence, and a well-defined ABL structure that fits in the applied mixed-layer theory. To estimate these terms, we therefore complement the observations from the E-DATA field campaign with the Weather Research and Forecasting (WRF) regional model. To complete the analysis, we evaluate the Penman equation adapted to saltwater conditions to study in

![Fig. 1](image-url)
more detail how the ABL processes independently influence the radiative and aerodynamic components of $E$.

In this section, we first describe the study site and the observations gathered in the Salar del Huasco. Next, we present the conceptual framework, including our hypothesis, that will be applied to analyze the ABL and $E$ using the mixed-layer theory and the Penman equation, respectively. Finally, we present the modeling approach, including the regional and the land–atmosphere model.

a. Study site and observations

Located in northern Chile, the Salar del Huasco is an arid, closed basin that extends over $19^\circ54’–20^\circ27’$S and $68^\circ40’–69^\circ00’$W in the Altiplano region (Fig. 1a). It covers an area of $\sim1470$ km$^2$, and it is placed at an average elevation of $\sim3800$ m above sea level (MSL). Geologically, like most of the adjacent endorheic basins in the Chilean Altiplano, this basin finds its volcanic origin in the Miocene, whose main rock formation in volume and extension is the ignimbrite (Blin et al. 2022). In this geographical setting, the main water input to the basin is the Collacagua River, whose stream from the north sinks a few kilometers before the salt flat (60 km$^2$), where it finally upwells in the Huasco saline lake (Fig. 1b). The lake corresponds to a shallow saltwater column of $\sim10$-cm depth placed over a sediment layer. The saline lake not only represents the principal $E$ pathway (Lobos-Roco et al. 2022a) but also serves as a refuge of endemic flora and fauna (de la Fuente et al. 2021). Although there is a clear surface heterogeneity, the dominant surface in the basin is desert, in which a deep ABL ($h \sim 1800$ m AGL at 1200 LT) is formed due to the high $H$ (Aguirre-Correa et al. 2023).

During the E-DATA field experiment held at the Salar del Huasco between 14 and 23 November 2018, airborne and surface observations were collected (Fig. 1b). Pertinent for our investigation, the study area was equipped with a surface energy balance station consisting of eddy covariance (EC) equipment, a four-component radiometer, and heat flux plates and a water temperature profile to estimate the ground heat flux $G$.

FIG. 2. Observations collected during the E-DATA field experiment (Suárez et al. 2020) over the water surface at the Salar del Huasco (Fig. 1). Diurnal variability of 10 days of (a) net radiation (RN), ground heat flux ($G$), sensible heat flux ($H$), and latent heat flux ($LE$). The $G$ is estimated as the residual of the energy budget (see appendix A); (b) wind speed at 2-m height; (c) temperature at 2-m height; and potential temperature vertical profiles during the (d) morning and (e) afternoon of 21 Nov 2018. Vertical dashed lines in (a)–(c) indicate the time when the regional air mass arrives at the region, while shadows represent the standard deviations of 10-day data. Horizontal dashed lines in (d) and (e) indicate the ABL height $h$ at specific times of the day estimated using the gradient method (Liu et al. 2022) explained later in section 2.
Here, $G$ is the energy that is exchanged with a two-component soil consisting of a water body on top of a sediment layer (see Fig. A1). The dynamic of $G$ is particularly complex. First, as the water body is partially transparent for radiation and it is shallow enough for the radiation to directly heat up the water-sediment interface, a second surface for energy balance partitioning has to be considered. Second, strong winds can make the shallow water layer turbulent. This affects the efficiency of the heat transport in this layer, and it affects the water turbidity through the upwelling of sediments, which in turn reduces the radiation transparency of the layer (de la Fuente and Meruane 2017; de la Fuente 2014). A more in-depth discussion on $G$ and how we estimated it from observations can be found in appendix A. In addition to the surface measurements, radiosonde launches and unmanned aerial vehicle (UAV) flights were executed to describe the state variables in the ABL and the free troposphere. Balloon launches were conducted over the water surface at 0900, 1200, 1500, 1800, and 2000 LT 21 November 2018. UAV flights were performed every 30 min from 0900 to 1200 LT, up to a height of 500 m AGL on the same day. These measurements were used to characterize $h$, the jumps of the conserved variables of potential temperature $\theta$, and specific humidity $q$ at the top of the mixed layer ($\Delta \theta$, $\Delta q$), and the free troposphere lapse rates ($\gamma_\theta$, $\gamma_q$). The observed ABL height was estimated using the gradient method (Liu et al. 2022; Marques et al. 2018; Oke 1989). The method involves analyzing the gradient of the $\theta$ and $q$ profiles, placing $h$ where the mixed layer shows the maximum positive $\theta$ gradient (Sorbian 1989; Garratt 1992) and minimum negative $q$ gradient at the top of the ABL (Ao et al. 2008). For further details regarding the field experiment and the instrumentation used, we refer to Suárez et al. (2020).

As both UAV flights and radiosondes were deployed on 21 November 2018, we use observations gathered during that day to determine the boundary and initial conditions for the land–atmosphere model, as well as to validate the regional model output. Considering that the conditions during the 10-day field campaign were similar, as can be seen by the small uncertainty ranges in Figs. 2a–c, this day is considered representative for all other days at the end of the dry season when $E$ rates are maximum (Lobos-Roco et al. 2022b).

b. Conceptual framework

In this section, we first describe the mixed-layer theory that will be used to characterize the ABL structure over the water surface. We then describe the Penman equation for open water adapted to saltwater conditions (from here on, we will refer to it as the Penman$_{sw}$ equation) that will be used to analyze how the ABL processes interact with the radiative and aerodynamic terms of $E$.

1) ATMOSPHERIC BOUNDARY LAYER: MIXED-LAYER EQUATIONS

Strong eddy currents formed between the surface and the ABL typically result in well-mixed atmospheric variables within the boundary layer. Hence, the ABL evolution can be simplified using the mixed-layer theory in which $\theta$, $q$, and wind components are uniformly distributed in the vertical direction (Tennekes and Driedonks 1981; Tennekes 1973; Lilly 1968). Potential temperature and moisture in the mixed layer can be described by one value representative for the layer estimated as (Vilá-Guerau de Arellano et al. 2015)

$$\frac{d\Psi}{dt} = \frac{(\bar{w} \Psi')_s - (\bar{w} \Psi')_l}{h} - \left(\pi \frac{d\Psi}{d\theta} + \nu \frac{d\Psi}{dy}\right),$$

where $\Psi = \bar{\theta}$ or $\bar{q}$, $(\bar{w} \Psi')_s$ is the surface flux, $(\bar{w} \Psi')_l$ is the entrainment flux, and $\pi$ and $\nu$ are the horizontal wind components. We define $\pi$ positive eastward and $\nu$ positive northward. The second term on the right-hand side of Eq. (1) is the average advection within the ABL ($\overline{\Psi}_{adv}$).

The boundary layer height tendency ($dh/dt$) can be described by

$$\frac{dh}{dt} = w_e + w_z - \left(\pi \frac{dh}{d\theta} + \nu \frac{dh}{dy}\right),$$

where $w_e$ is the entrainment velocity, $w_z$ is the subsidence vertical velocity, and the third term on the right-hand side of the equation is the mass advection $h_{adv}$ (Aguirre-Correa et al. 2023; Kossmann et al. 1998).

2) EVAPORATION: PENMAN EQUATION

The Penman equation is a widely used model to estimate open water evaporation from local measurements (Monteith 1965; Penman and Keen 1948). Differently from other methods [e.g., the Dalton equation (Aldarabseh and Merati 2022)], Penman accounts for both the aerodynamic and radiative contributions to $E$ and allows us to disentangle their relative contributions and those of their subterms. Calder and Neal (1984) modified the Penman equation to estimate the evaporation rates from saline water bodies by considering the decrease that salinity causes in the saturation vapor pressure at the water surface. The decrease is linked to the reduction of the activity of the saline water $a_w$, as dissolved salts reduce the chemical potential of the solution and lower the free energy of water molecules (Mor et al. 2018; Akridge 2008; Asmar and Ergenzinger 1999; Salhotra et al. 1985). The water activity can be estimated from water density, temperature, or salinity (Akridge 2008; Oroud 2001; Asmar and Ergenzinger 1999). We measured a salinity of 0.977. These values are comparable to those of ocean (Rana et al. 2022; Liu et al. 2018; Jungwirth and Tobias 2006; Garrett 2004; Vrbka et al. 2004; Jungwirth and Tobias 2002).

For more details about the estimation of $a_w$ and $r_{salt}$, we refer to appendix B.
The Penman equation adapted to saltwater conditions, Penman$_{sw}$, reads

\[
LE = \frac{\text{radiative contribution}}{s(RN - G)} + \frac{\text{aerodynamic contribution}}{s + \frac{\gamma}{a_w} \left( 1 + \frac{r_{\text{salt}}}{r_{\text{sw}}} \right)},
\]

where $s$ is the slope of the saturated vapor pressure curve, $RN$ is the net radiation, $c_p$ is the specific heat at constant pressure, $r_{\text{sat}}$ is the aerodynamic resistance, $qsat$ is the saturated specific humidity, and $\gamma$ is the psychrometric constant. The psychrometric constant is estimated as $cp/L_v$, where $L_v$ is the latent heat of vaporization. Salt decreases the latent heat of vaporization depending on temperature and salinity (Sharqawy et al. 2010). For the measured salinity at $20^\circ C$, $L_v = 2.36 \times 10^6$ J kg$^{-1}$ (appendix B).

The first term on the right-hand side of Eq. (3) is the radiative contribution to evaporation, which depends on $RN$, $G$, and $s$. To be consistent with the energy balance closure forced in the models, we also estimate $G$ as a residual of the energy budget from the observations ($G_{\text{EBC}}$; Fig. 2a). We refer to appendix A, where we describe the soil heat system and we compare $G_{\text{EBC}}$ with $G$ estimated from observations. The second term on the right-hand side is the aerodynamic contribution, which is controlled by the turbulence intensity described by the aerodynamic resistance and the vapor pressure deficit (VPD). LE is therefore intricately linked to the atmospheric water-holding capacity, which is influenced by the atmospheric temperature and humidity. Consequently, $E$ is controlled by both the properties of ABL and those of the land surface.

c. Modeling approach

In this section, we describe the models and their setup. First, we present the regional model, which is turbulent resolved in the domain around the salt lake, and then, we present the land–atmosphere conceptual model.

1) REGIONAL MODEL

To quantify the effect of the regional forcing and have a better characterization of the ABL that fits in the mixed-layer theory, the WRF Model is used. Figure 3a shows the spatial distribution and dimensions of the four domains, with grid sizes of 12.5 km for domain D01, 2.5 km for D02, and 500 m for D03. The inner domain D04 has a 100 m $\times$ 100 m horizontal resolution and encompasses the study area with 296 $\times$ 346 grid points. By using WRF in large-eddy simulation (LES) mode in D04, we attempt to represent the local atmosphere turbulence as accurately as possible by explicitly solving the dynamical processes in the ABL described by the Navier–Stokes equations (Heus et al. 2010). Land use, and initial and boundary conditions were modified to be in agreement with the 100-m spatial resolution (Figs. 3b,c). For the vertical discretization, physical parameterizations, and how $G$ is treated in the model, the reader is referred to appendix C.

The model was initialized for 1200 UTC 20 November 2018 and a 24-h spinup was applied to analyze 21 November, the day when airborne observations were collected. A time step of 60 s was used, and 30-min fields were stored. Initial and boundary conditions were obtained from ECMWF ERA5 reanalysis data (0.25$^\circ$ spatial resolution) for the domains presented in Fig. 3a. Updates of the large-scale forcing were available every 6 h. A 2-K atmospheric temperature bias in the ERA5 input was corrected as in Aguirre-Correa et al. (2023), which was attributed to the relatively coarse resolution of ERA5 in relation to the marked east–west topography across Chile (Muñoz et al. 2022). To obtain statistically robust estimations from the WRF Model, spatial averages of 300 m $\times$ 300 m were used to avoid single gridcells with possible erratic behavior (Aguirre-Correa et al. 2023). The $\theta$ jump at the top of the ABL is not always as identifiable in models as it is in observations, since it is very localized and tends to get spread out depending on the vertical resolution. We therefore apply the bulk Richardson number method (Vogelezang and Holtslag 1996) to estimate the boundary layer height from WRF, with a critical value of 0.25 (Li et al. 2021; Zhang et al. 2014; Vickers and Mahrt 2004). The method has proven to be consistent under a wide range of atmospheric conditions (e.g., Aguirre-Correa et al. 2023; Min et al. 2020; Zhang et al. 2014; Richardson et al. 2013).

2) LAND–ATMOSPHERE CONCEPTUAL MODEL

In this study, we use the Chemistry Land-Surface Atmosphere Soil Slab (CLASS) model (Vilà-Guerau de Arellano et al. 2015). CLASS captures the fundamental processes that govern the land–atmosphere interactions, encompassing both surface and atmospheric dynamics and their feedback, to simulate the surface fluxes and the ABL development. This property of CLASS facilitates the interpretation of observations as it allows to disentangle the very complex land–atmosphere interactions via the activation and deactivation of different processes. Hence, it allows us to identify the relative importance of various processes to a certain variable (e.g., boundary layer height or evaporation).

To capture the surface and ABL dynamics, CLASS employs parameterizations of the radiation balance, the surface energy balance, the atmospheric surface layer, and the ABL processes (Vilà-Guerau de Arellano et al. 2015). We refer to appendix D for details about the treatment of $G$ in the CLASS model. The ABL is represented using the mixed-layer theory applied to a slab layer. Therefore, Eqs. (1) and (2) are used in CLASS. In the model, the free troposphere is described by a constant lapse rate ($\gamma_0 \times \gamma$) and the entrainment fluxes are estimated as a fraction $\beta$ of the surface fluxes, usually considered 0.2 (Aguirre-Correa et al. 2023; Wouters et al. 2019; Stull 1988; Tennekes 1973). The nonlocal processes of advection in Eqs. (1) and (2), and subsidence in Eq. (2), need to be estimated first (in our case from WRF) and then prescribed in the model. The reader is referred to van Heerwaarden and Teuling (2014) for a comprehensive formulation of the coupled land–atmosphere system in CLASS.
3. Research approach

The research strategy consists of using CLASS to break down the complexity of the regional processes by calculating the individual components of the budget that controls the ABL and $E$. However, we first need to properly define the ABL structure based on the mixed-layer theory and estimate the nonlocal terms that must be prescribed. For the estimation of the prescribed terms (advection and subsidence) and their results, we refer to appendix D. For defining the ABL structure using the mixed-layer theory, Fig. 4 presents the $\theta$ profiles from the WRF Model at three key time steps (1130–1230 LT) when the ABL height increases by more than 1 km (Fig. 2e).

An unstable surface layer with decreasing $\theta$ up to $\sim$50 m AGL is observed at all time steps. Above the surface layer, we recognize a stable layer at 1130 and 1200 LT that reaches up to 300 m AGL, where $\theta$ increases with height. We also detect the positive $\theta$ gradient from the observations (Fig. 2e). On top of the stable layer, we observe a nearly mixed layer separated from the increasing potential temperature in the free troposphere by a small temperature inversion. This layer is typically formed during the evening transition and at night, leading to a residual layer (RL) from the previous day that takes place above the ABL (Ouwersloot et al. 2012). In the RL, the turbulence is typically weak to negligible as it is not fed from the surface. The stable shallow layer and the residual layer are no longer observed at 1230 LT. Instead, a deep mixed layer is found with $h \sim 2000$ m AGL (Fig. 4c). In brief, a shallow ABL with a residual layer on top develops over the water surface during the morning until it suddenly becomes a deep mixed layer at 1230 LT. By using CLASS, we will study the interaction between each atmospheric layer and the processes that control the sharp 1-km growth of the ABL around noon.

According to the WRF profile analysis, we adapted the default land–atmosphere model to include the residual layer recognized during the morning. Figure 5 provides a representation of the ABL structure, along with the primary drivers of the boundary layer and the surface energy balance delineated by CLASS. Based on Fig. 5, we defined four experiments with an increasing level of complexity to analyze the processes that drive the ABL and $E$ (from hereon, we will refer to these as ABL and $E$ drivers, respectively).
Table 1 and Fig. 6 summarize the following experiments:

1) **Base case:** In this first experiment, only land–atmosphere interactions and subsidence motions are considered. The wind divergence that is required to parameterize subsidence in the model is prescribed from Aguirre-Correa et al. (2023) (see appendix D). This experiment resembles the situation of an infinite lake with no connection with the Pacific Ocean and no influence of the direct surroundings (Fig. 6a). As a result, there is no production of turbulence at noon nor mass, temperature, or moisture advection during the day.

2) **Wind:** This experiment adds wind by prescribing it from the observations (Fig. 2b). The situation still resembles an infinite lake with no influence of the surrounding desert, but now, there is a connection with the Pacific Ocean and the arrival of the regional flow that produces an additional mixing during the afternoon (Fig. 6b). As a result, there is no production of turbulence at noon nor mass, temperature, or moisture advection during the day.

3) **Mass advection:** In this experiment, mass advection $h_{adv}$ is incorporated. Now, there is a small lake that interacts with the surroundings, and, consequently, there is advection of an external boundary layer [advected layer (AL) in Fig. 6c] toward the lagoon. Due to the interaction with the surroundings, the model is now initialized with an RL prescribed from the WRF Model profile at 0900 LT. As long as there is a residual layer, $h_{adv}$ influences the RL height. Otherwise, the mass advection acts over the ABL. The mass advection is positive during the morning, indicative of the advection of a deeper ABL, and negative during the afternoon, which means that a shallower boundary layer is advected toward the water surface (see appendix D).

4) **+ θ and q advection:** In this last experiment, we include the advection of temperature and moisture in each layer (Fig. 6d). The ABL and the residual layer are influenced by the advection of warm and dry air in the morning coming from the desert, and cold and dry air when the regional flow arrives. Advection in the upper layer was found to be negligible. The reader is referred to appendix D, where our approach to estimate advection in each layer is presented.

To study the ABL and $E$ drivers, the experiments were run between 0900 and 1900 LT with a time step of 60 s. The results were integrated to a 10-min period to compare them with the observations and the WRF Model results. As previously stated, measurements gathered on 21 November 2018 were used to define the initial and boundary conditions of the model (Table D1 in appendix D). For each experiment, we analyze the sensible heat flux, the ABL height time series, and the $θ$ profiles to determine the ABL drivers. For the $E$ drivers, we evaluate the Penman$_{sw}$ equation [Eq. (3)] to study how the regional processes affect the radiative and aerodynamic terms of LE.

### 4. Atmospheric boundary layer and Evaporation drivers

#### a. Atmospheric boundary layer drivers

Figure 7 presents the sensible heat flux, which is the main driver of the ABL growth, estimated from the experiments in CLASS (Table 1), field observations, and the WRF Model. The morning discrepancy between the observations and the
WRF values is explained by the crude representation of $G$ in the model and its impact on the surface temperature. We refer to appendix C for further details. Differences between the CLASS experiments are only recognized during the afternoon as $H$ remains very close to 0 W m$^{-2}$ during the morning in all the cases due to the lack of turbulence (Fig. 2b). We will therefore focus on the afternoon regime.

When only land–atmosphere interactions and subsidence motions are included in the analysis (Base case, red line), the absence of turbulence continues throughout the day, resulting in small sensible heat flux until 1900 LT ($H < 50$ W m$^{-2}$ in Fig. 7). When wind is added to the experiment (violet line), the turbulent kinetic energy sharply increases from $\sim 0$ to 4 m$^2$ s$^{-2}$ at 1300 LT and sensible heat over the water is strongly enhanced, reaching a maximum value of $H = 175$ W m$^{-2}$ around 1430 LT. When the mass advection is added to the experiment (black line) and the model is initialized with a residual layer on top of the ABL (see Fig. 6c), no significant changes in sensible heat are observed. Later in this section, we will discuss how the ABL dynamics can explain the limited impact of mass advection on $H$. Finally, when temperature and moisture advection are included in the analysis, smaller $H$ is found during the afternoon. We argue that the origin of this lies in the strong warm-air advection observed during the morning ($\sim 5$ K h$^{-1}$ on average, Fig. D2a in appendix D), which considerably increases the air temperature above the water surface (Fig. 2c) and decreases the atmosphere–surface thermal gradient before midday. When the strong mixing starts with the afternoon wind arrival, the fluxes are therefore smaller even though the regional flow is accompanied by a weak cold-air advection ($\sim -2$ K h$^{-1}$, Fig. D2a), which potentially enhances the near-surface temperature gradient. With all processes included, CLASS adequately represents the observations (Fig. 7).

Figure 8 presents the ABL height time series from the experiments defined in CLASS, compared to observations and the WRF Model. To aid the interpretation of Fig. 8 and in general to follow the ABL dynamics for the CLASS experiments, we show potential temperature profiles for three key time intervals before (Figs. 9a–c) and three key time intervals after (Figs. 9d–f) the arrival of the regional flow.

As expected from the small sensible heat flux and the minimal mechanical turbulence over the water surface (Figs. 2a,b), the ABL remains shallow during the morning in all the experiments and the CLASS $h$ estimates agree fairly well with the observed and WRF modeled values (Fig. 8). Note that the morning overestimation of $H$ in the WRF Model (Fig. 7) is not significant enough to impact the

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**TABLE 1. Experiments defined for the sensitivity analysis in CLASS.**

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<tr>
<th>Experiment</th>
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</tr>
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<td>+ Mass advection</td>
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<td>Yes</td>
<td>Yes</td>
<td>No</td>
</tr>
<tr>
<td>+ $\theta$ and $q$ advection</td>
<td>Yes</td>
<td>Yes</td>
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<td>Yes</td>
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**Fig. 5.** CLASS conceptual model scheme for a water body. Main processes and variables involved in the development of the surface and ABL under the mixed-layer theory are presented, including the radiation balance of the longwave ($L_{in}$, $L_{out}$) and shortwave ($S_{in}$, $S_{out}$) components, the surface energy balance, the entrainment fluxes, the advection processes, and the subsidence motions. An RL is also included, for which the subindex RL is used. The AL is represented by $h_{adv}$. In our case, $G$ is estimated as the residual of the energy budget (see appendix D).
ABL growth. During the afternoon, remarkable changes are observed in $h$.

In the Base case (red line), the ABL remains shallow throughout the day with $h < 350$ m AGL until 1900 LT (Fig. 8). In this experiment, there is no mechanism that allows the ABL to grow as the disconnection from the Pacific Ocean also results in small sensible heat flux during the afternoon (Fig. 7). Hence, in the absence of wind and advection processes, there is a very limited ABL growth over the water surface. The ABL height even shrinks after midday because of the ever present subsidence motions ($-8 \times 10^{-5}$ s$^{-2}$ during the afternoon, appendix D).

When wind is added to the model (violet line), the ABL starts growing from 1330 LT onward to $\sim 750$ m AGL. This ABL behavior is directly linked to the wind-enhanced $H$ seen around the same time (Fig. 7). A secondary effect of the enhanced $H$ is an increase in the entrainment flux ($w'\theta'$), and a decrease in the potential temperature jump from $\sim 2$ K at 1300 LT to $\sim 0.5$ K at 1330 LT due to the increase in the air temperature. As a result, the entrainment velocity $w_e$ increases and counteracts the subsidence motions allowing the ABL to grow during the afternoon [see Eq. (2)].

When the mass advection is included in the analysis (black line), we also incorporate an RL to the initial profile (Fig. 6c). In the morning, in the absence of warm-air advection from the surrounding desert and due to the weak sensible heat flux, the shallow ABL is incapable of reaching the residual layer temperature (Figs. 9a–c). It is only when the sensible heat flux
is enhanced at 1300 LT that the ABL rapidly warms and reaches the RL temperature at 1400 LT, leading to the merging of the two layers and an immediate increase in the ABL height (Figs. 8 and 9e). This process has been described as an overshooting (Ouwersloot et al. 2012). The sharp 1.5-km height overshoot observed at 1400 LT is therefore explained by the increase in the air temperature due to the enhanced positive mass advection (Fig. D1) that takes place above the ABL and increases the residual layer height (Figs. 9a–c). Consequently, when the two layers connect, the ABL becomes a deep mixed layer of \( h \approx 1700 \text{ m AGL} \). In this case, adding the mass advection and the interaction with the RL has little impact on the ABL temperature, as the main impact happens in the residual layer. Hence, \( \theta \) in the ABL remains mainly the same as in the previous experiment (Fig. 9). As a result, the land–atmosphere thermal gradient remains undisturbed and no changes are observed in \( H \) when introducing mass advection (Fig. 7).

When the morning advection of warm and dry air from the desert (Fig. D2) is included (orange line), the ABL rapidly increases its temperature until midday (Figs. 9a–c). Although the residual layer also warms due to the \( \theta \) advection, the ABL is able to reach its temperature and the overshooting, the instant merging of both layers into one deep mixed layer, happens already at 1200 LT (Fig. 9c). This process is also observed in the WRF profile (Fig. 4c) but half an hour later, at 1230 LT. In reality, the overshooting does not occur instantaneously and it takes time for the surface thermals to reach the top of the residual layer (Ouwersloot et al. 2012). Therefore, there is a transition period, that WRF accurately simulates, which takes around 30 min and explains the discrepancy between the overshooting timing compared to CLASS. As a consequence of the merging, the ABL height exhibits a sharp increase of 1 km at noon, followed by an extra \( \sim 700 \text{ m} \) increase when the arrival of the regional flow advects a deeper ABL (\( \sim 1500 \text{ m h}^{-1} \) at 1200 LT, Fig. D1). Subsequently, the ABL height gradually collapses until the end of the day, which is in good agreement with the observations and WRF. The collapse was also reported by Aguirre-Correa et al. (2023) for the boundary layer over the desert surface. It is due to a combination of the subsidence, the advection of a shallower ABL (\( \sim 350 \text{ m h}^{-1} \) in Fig. D1), and the arrival of the cold air (\( \sim 2 \text{ K h}^{-1} \) in Fig. D2a) over the course of the afternoon that reduces the entrainment velocity by increasing the potential temperature jump to \( \sim 3 \text{ K} \).

**b. Evaporation drivers**

In this section, we present the evaporation results in the form of latent heat flux (LE) estimated from the CLASS experiments and compared to observations and the WRF Model results (Fig. 10). As observed for \( H \), there is a morning discrepancy between the observations and the WRF Model, which is related to the inadequacy of WRF to represent the soil heat flux in shallow water bodies (see appendix C for further details). To analyze in detail the \( E \) drivers, Fig. 11 presents a breakdown of the Penman-Monti equation terms [Eq. (3)] for all CLASS experiments, field observations, and the WRF regional model.

In the CLASS Base case experiment (red line in Fig. 10), LE shows significantly smaller magnitudes compared to the observations. The underestimation is observed in both the radiative and aerodynamic contributions to \( E \) (Figs. 11a,b). Similarly to what was seen for the sensible heat flux, the small magnitudes in LE are mainly explained by the high aerodynamic resistance (\( \sim 400 \text{ s m}^{-1} \)) due to the lack of wind that in this case inhibits the turbulent component of \( E \) (Fig. 11d). As both \( H \) and LE are low due to the absence of turbulent mixing (Lobos-Roco et al. 2021), a considerable part of the radiation energy is used for \( G \) and there is little energy available for the turbulent fluxes (\( RN - G < 150 \text{ W m}^{-2} \) in Fig. 11e). Last, the absence of warm- and dry-air advection during the morning and the low \( H \) result in lower air temperature, causing a considerable underestimation of the slope of the saturated vapor pressure curve (s, Fig. 11c) and the VPD (Fig. 11f). In all,
the disconnection with the regional flow and the missing interaction with the surrounding desert leads to the underestimation of the radiative and aerodynamic contributions to $E$ when compared to the observations and the WRF modeled values.

When wind is prescribed to the experiment (violet line), LE is triggered at 1300 LT and reaches values up to 380 W m$^{-2}$ at 1500 LT (Fig. 10). The onset of wind in the early afternoon due to the interaction with the Pacific Ocean leads, compared to the Base case (Figs. 11a,b), to a $\sim$285% and $\sim$318% increase in the daily radiative and aerodynamic contributions to $E$, respectively. For the aerodynamic term, this increase is mainly due to the decrease in $r_a$ from $\sim$400 to $\sim$70 s m$^{-1}$ around midday (Fig. 11d) and due to the increase in the VPD (Fig. 11f). The VPD increase is a result of the rise in air temperature caused by the wind-enhanced $H$ within the shallow ABL. The radiative term increase is mainly explained by the feedback between wind and the ground heat flux (Fig. 11e). When the regional flow arrives, it mixes the water body and makes the energy that was stored in the soil system in the morning available to the surface ($G < 0$ W m$^{-2}$ at 1700 LT; see Fig. 2a) where it feeds the turbulent fluxes. This effect has also been observed in other saline lakes (e.g., Hamdani et al. 2018). Compared to the Base case, there is already a much better agreement with the observations and the WRF Model results for the radiative and aerodynamic terms of $E$. Thus, wind explains for 80% the observed evaporation over the saline lake by controlling both the aerodynamic and radiative terms.

Fig. 9. Potential temperature profiles estimated from the CLASS experiments (Table 1) at (a) 1100, (b) 1130, (c) 1200, (d) 1330, (e) 1400, and (f) 1430 LT. Dashed horizontal lines indicate the boundary layer height from the Base case experiment.
of $E$, by decreasing the aerodynamic resistance, but above all releasing the energy stored in the two-component soil. Note that the impact of wind on the magnitude of LE is considerably higher in LE$_{rad}$ ($\sim$200 W m$^{-2}$ increase) than in LE$_{aero}$ ($\sim$40 W m$^{-2}$ increase). That is why $E$ in the Salar del Huasco has been defined as a wind-limited regime in previous studies (Lobos-Roco et al. 2021) in terms of triggering evaporation. However, after wind ceases to be a limiting factor during the afternoon, LE decreases in line with the decrease in the energy and radiation becomes the limiting factor (LE in Fig. 10 decreases from 1500 LT following LE$_{rad}$ in Fig. 11a, while LE$_{aero}$ in Fig. 11b keeps increasing). This makes the radiative term of high relevance when explaining $E$ dynamics, which is well captured by the Penman method.

When we include the interaction with the surrounding desert and the mass advection is added to the experiment (black line), LE remains pretty much the same, like $H$ (Fig. 7). Also, the radiative and aerodynamic terms of $E$ do not present significant changes (Figs. 11a,b).

Finally, when the ABL processes of temperature and moisture advection are included (orange line), LE increases by $\sim$20% (Fig. 10) and approaches the observations and the WRF Model results. Advection has a nearly equal impact on both the radiative (+18%) and aerodynamic (+22%) components of evaporation (Figs. 11a,b) compared to the previous experiment. In the morning, the warm- and dry-air advection from the surrounding desert (Fig. D2) makes $s$ and VPD increase rapidly (Figs. 11c.f). The increase has no direct influence on LE, as evaporation is virtually zero due to the absence of mechanical turbulence. However, warm and dry advection of the morning combined with the mass advection at 1200 LT forms a warm and dry layer with increased $s$ and VPD that act as a buffer to the cold (and dry)-air advection seen after midday (Figs. D2a.b). Therefore, although $s$ and VPD gradually decrease over the afternoon due to the influence of cold-air advection, there are still higher values in comparison with the previous experiments (Figs. 11c.f). Consequently, LE is larger during the afternoon and there is a better resemblance with the observations. We therefore recognize that the main $E$ drivers are the wind and the thermal stability of the air above the water surface controlled by the ABL processes, which can either enhance or suppress evaporation. These dynamics have also been reported in other saline lakes, in which the wind regime and the boundary layer properties dominate the diurnal cycle of $E$ (Hamdani et al. 2018).

5. Summary and conclusions

In this research, we studied the ABL and $E$ drivers in a saltwater lagoon in the Altiplano region using a land–atmosphere conceptual model, field observations, and a regional model. We found that the shallow, stable ABL is primarily governed by the transport of warm air during the morning ($\sim$5 K h$^{-1}$ on average), as well as the advection of a deeper layer from the surrounding desert at 1200 LT ($\sim$1500 m h$^{-1}$). As a result, there is a sharp 1.5-km increase in the ABL height around midday when the ABL reaches the residual layer temperature and an overshooting takes place. Simultaneously, the slope of the saturated vapor pressure curve and the VPD over the water rapidly increase in accordance with the rise in the air temperature driven by the warm-air advection. Despite having the meteorological conditions for evaporation to happen, the lack of turbulence leads to an almost absence of turbulent fluxes ($H$ and LE $< 50$ W m$^{-2}$). Consequently, the energy is stored in the water body and the underlying sediments ($G > 0$ W m$^{-2}$). During the afternoon, the advection of cold air ($\sim$2 K h$^{-1}$) and a shallower ABL ($\sim$350 m h$^{-1}$) when a thermal and topographically enhanced regional flow reaches the region leads to a collapse of the ABL. We confirmed that the increase in the wind speed when the regional flow arrives plays a key role in triggering $E$ (80%). By using the Penman equation adapted to saltwater conditions, we demonstrated that wind does not only decrease the aerodynamic resistance but also release the energy previously stored in the water and sediments ($G < 0$ W m$^{-2}$ during the afternoon). Despite wind being a significant driving force for $E$, the remaining 20% is primarily explained by the advection of air masses.
with different temperatures and moisture that adjust the VPD over the lake and control the atmosphere water-holding capacity. This study emphasizes the relevance of understanding the ABL dynamics to comprehend evaporation regimes, particularly in a complex setting as the Altiplano. For future research, we recommend a continuous monitoring of the ABL with radiosondes to confirm the interactions described in this study.

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Data availability statement. The datasets analyzed in this study are available upon a well-motivated request.

APPENDIX A

Ground Heat Flux Dynamics in the Shallow Lagoon

The ground heat flux $G$ in the saline lake of the Salar del Huasco is characterized as the energy exchange with a two-component soil consisting of a shallow water layer (~10 cm deep) on top of sediments (Fig. A1a). During the E-DATA experiment (Suárez et al. 2020), we had two ground heat flux plates (GHFPs) installed at 2 cm below the water–sediment interface directly measuring the ground flux at that depth ($G_{\text{meas}}$; horizontal bars in Fig. A1a). In addition, we installed temperature profile measurements (dots in Figs. A1a,b) to calculate the energy storage above the GHFPs using the caloric method (Kimball and Jackson 1975). Only thermometers above the GHFPs and within the water layer

FIG. 11. Budget of the evaporation terms based on the Penman$_{\text{eq}}$ equation analysis from the CLASS experiments (Table 1) compared to the observed and WRF Model values for the (a) radiative term, (b) aerodynamic term, (c) slope of the saturated vapor pressure curve, (d) aerodynamic resistance term, (e) available energy, and (f) VPD. The vertical dashed line indicates the time when the regional air mass arrives.

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were considered for the calculations. Storage was estimated as the change in the vertically integrated temperature profile in time (profiles in Figs. A1c,d). We considered an increase (decrease) in $G$ when the integrated value increased (decreased):

$$G = G_{\text{meas}} + C_p \int \frac{dT}{dt} dz,$$

(A1)

where $G_{\text{meas}}$ is the measured heat flux 2 cm below the water–sediment interface, $C_p$ is the volumetric heat capacity, $T$ is the water temperature, and $z$ is the layer depth.

The results for the observed $G$ are shown in Fig. A2 (black crosses) as the 10-day average of the E-DATA field campaign. From our results, we distinguish two regimes in $G$ during daytime (schematized in Fig. A3):

1) During the morning (increasing $G$): Due to the lack of wind, the water is nonturbulent and hence has a maximum transparency (Fig. A3a). Radiation heats up the water–sediment interface more than it does the water–atmosphere interface (de la Fuente and Niño 2010; de la Fuente 2014; de la Fuente and Meruane 2017). Combined with the relatively inefficient mixing (diffusion), this leads to a
strong heating of the water \( G_2 \) and sediment \( G_1 \) from the water–sediment interfaces, whereas \( G_3 \) is small.

2) During the afternoon (decreasing \( G \)): The radiation is at its maximum level, while wind-induced waves mix the water column, efficiently bringing the stored heat from morning in the sediments and water to the surface where it feeds \( H + LE \). After a while, there is sediment resuspension which reduces the transparency of the water (de la Fuente 2014) (Fig. A3b) and the two surface energy partitioning of \( G \) transitions to a single surface.

From these measurements, we found a \( \sim 30\% \) of energy imbalance. The observed imbalance is comparable to other EC campaigns over lakes (Sun and Holmes 2019; Wang et al. 2014; Nordbo et al. 2011; Tanny et al. 2011). The non-closure of the surface energy from the observations is attributed to the estimation of heat storage under conditions of intensive turbulence and horizontal heat transport over the water body, as well as the large advection due to the highly heterogeneous setup. Other possible sources of energy imbalance are instrumentation limitations (Horst et al. 2015), advective fluxes (Moderow et al. 2021), and dispersive contributions to the turbulent fluxes (Morrison et al. 2022).

The green crosses in Fig. A2 present \( G \) estimated as a residual of the energy budget (\( G_{EBC} = RN - LE - H \); Fig. 2a). We observe that even though there is a discrepancy in the magnitudes, the dynamic behaviors of \( G_{EBC} \) and the measured \( G \) are very similar. The energy that is transferred to the two-component soil has positive values and increases until 1200 LT (i.e., energy is used to heat the water–sediment layers and energy is stored in the water column), and then gradually decreases until it becomes negative sometime between 1500 and 1700 LT (i.e., energy stored in the water–sediment layers provides energy to the

---

**Fig. A2.** Ground heat flux from observations (10-day average; black crosses), from observations estimated as the residual of the energy budget (gray crosses), WRF Model (red crosses), and CLASS model (orange line). WRF Model and CLASS model treatment of \( G \) is described in appendixes C and D.

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**Fig. A3.** Ground heat flux in the two-component soil system for (a) low and (b) high wind conditions, and how they connect to the energy balance, where \( T_{aw} \) is the temperature \( T \) at the atmosphere–water interface, \( T_{ws} \) is the \( T \) at the water–sediment interface, \( T_{wat} \) is the \( T \) of the water body, \( T_{sed} \) is the \( T \) of the sediment, \( G_1 \) is the ground heat flux driven by \( T_{aw} - T_{wat} \), \( G_2 \) is the ground heat flux driven by \( T_{ws} - T_{wat} \), and \( G_3 \) is the ground heat flux driven by \( T_{ws} - T_{sed} \).
turbulent fluxes). Since in this research, we aim to analyze the relevance and temporal dependence of the drivers of the turbulent fluxes, we deem the use of \(G_{\text{EBC}}\) acceptable. By doing this, we also provide consistency to our approach as both models and observations equally consider the closure of the energy balance (EBC). The implication of forcing the EBC on the observations through \(G\), however, is that the absolute values of \(H\) and \(LE\) might be uncertain. Still, the relative importance and temporality of the turbulent fluxes would not be affected by the nonclosure of the energy budget, and therefore, the discussions provided in this research remain valid.

Due to the complexity of these processes, we implemented a simplified treatment of \(G\) in WRF and CLASS, explained in appendixes C and D, respectively.

**APPENDIX B**

**The Impact of Salinity on Evaporation**

In this appendix, we provide further details regarding how the influence of salinity on evaporation is considered in our Penman equation adapted to saltwater conditions [Eq. (3)].

*a. Water activity*

The water activity of a solution \((aw < 1)\) relates the vapor pressure over saline water to the vapor pressure over fresh-water (Calder and Neal 1984; Krumgalz and Millero 1982). Salinity decreases the saturated vapor pressure due to dissolved salts reduce the chemical potential of the solution and lower the free energy of water molecules (Mor et al. 2018; Akridge 2008; Asmar and Ergenzinger 1999; Salhotra et al. 1985). Previous studies have suggested that the saltwater density is more significant to determine \(aw\) than temperature or the chemical composition (Groeneveld et al. 2010; Turk 1970; Calder and Neal 1984; Salhotra et al. 1985). Asmar and Ergenzinger (1999) developed the following empirical relationship to estimate the water activity from density:

\[
aw = -30.285 + 77.650 \rho_w - 62.712 \rho_w^2 + 16.300 \rho_w^3, \quad (B1)
\]

where \(\rho_w\) is the fluid density (kg m\(^{-3}\)).

Density was measured from a water sample collected from the saline lake at the Salar del Huasco using a generic densimeter (range 0.990–1.100 g ml\(^{-1}\) at 20°C). A density of 1.038 kg m\(^{-3}\) was found. By applying Eq. (B1), the water activity for the saline lake was estimated as \(aw \approx 0.977\). This value was used in Eq. (3).

*b. Latent heat of vaporization*

Studies on the relationship between salinity and the latent heat of vaporization \(L_v\) often involve empirical correlations and mathematical models that describe the behavior observed in experimental data, which shows that \(L_v\) decreases with temperature and salinity (Generous et al. 2020; Valderrama et al. 2015; El-Dessouky and Ettounney 2002). The weakening of the hydrogen bonding between water molecules (Nucci and Vanderkooi 2008) and changes in the thermal energy due to the presence of salt ions (Fujiyasu and Fahey 2000) might explain this relationship. Sharqawy et al. (2010) developed an equation to estimate \(L_v\) from salinity and temperature. Salinity was estimated from the electrical conductivity of the water sample collected at the salt lake. The electrical conductivity was measured using a multiparameter meter (HACH HQ40d) and an electrical conductivity probe (HACH CDC401). An \(L_v = 2.36 \times 10^6 \text{ kg} \cdot \text{m}^{-2} \cdot \text{s}^{-1}\) was found for a salinity of 37.7 kg \(\text{kg}^{-1}\). The latent heat of vaporization is used to estimate the psychrometric constant [Eq. (3)].

*c. Salt resistance at the surface*

We consider that salinity has a third and last impact on evaporation, which we denoted by \(r_{\text{salt}}\) [Eq. (3)]. Salt ions interact with water molecules at the surface affecting the molecules’ orientation, increasing the surface tension, and reducing the number of water molecules that can occupy the surface, leading to a decreased evaporation rate compared to pure water (Rana et al. 2022; Liu et al. 2018; Jungwirth and Tobias 2006; Garrett 2004; Vrba et al. 2004; Jungwirth and Tobias 2002). Using \(r_{\text{salt}}\) as a closure term with respect to the eddy covariance measurements, we found that \(r_{\text{salt}} = 50 \text{ s} \cdot \text{m}^{-1}\). This additional resistance is only relevant during the afternoon, as during the morning, evaporation is \(-0 \text{ W} \cdot \text{m}^{-2}\) due to the lack of turbulence, independently of any resistance that salt can provide to the surface.

**APPENDIX C**

**Weather Research and Forecasting Model**

Table C1 provides information on the physics schemes and dynamics for each domain of the WRF Model, as well as other information such as the time step and spatial discretizations.

The vertical discretization in the model was set in 79 levels up to 100 hPa, from which 35 levels were defined in the first 2000 m. Regarding the horizontal discretization, the larger domain (D01) has a horizontal resolution of 12.5 km \(\times\) 12.5 km. The inner domains gradually decrease their grid sizes in a ratio of 5 (see Table C1). With this, the smallest domain D04, which closely surrounds the Salar del Huasco, has a 100 m \(\times\) 100 m horizontal resolution. The physical parameterizations are summarized for each domain at the end of Table C1. For the smaller domain D04, we modified the model setting to explicitly solve turbulence using the LES mode. Specifically, the turbulence and mixing option in the model was set to evaluate the mixing terms in space by using the 1.5-order TKE closure eddy coefficient option. We also introduced a Rayleigh damping layer to filter the effect of unrealistic gravity waves caused by the steep topography of the Andes as in Lobos-Roco et al. (2021). We included this layer at 7 km with a damp coefficient of 0.2 (Lobos-Roco et al. 2021; Klemp et al. 2008).

We modified the default land-use map of WRF as shown in Fig. 3c. The salt surfaces were incorporated as the barren surface with a high albedo (0.45 for wet salt and 0.65 for dry salt) as reported in the E-DATA experiment. Moisture was set to 0.5 for the wet salt surface. For the desert surface that surrounds the Salar, the open shrubland and grassland land use was replaced with barren or sparsely vegetated as it
represents better the basin characteristics. We included the water surface as the water category in WRF, with an albedo of 0.12 as obtained from the field campaign (Lobos-Roco et al. 2021; Suárez et al. 2020). We had to represent the lagoon as a deep lake as WRF uses four soil layers defined under a single category (e.g., water or soil). The consequences of not being able to represent the shallow lake and the processes occurring at the water–sediment interface was a morning overestimation of the energy used to heat the lake surface and an important overestimation of the surface temperature $T_s$. Consequently, there is a smaller temperature gradient between the surface and the soil layers that leads to an underestimation of $G$ (Fig. A2), and a higher atmosphere–surface temperature gradient which explains a larger $H$ and LE compared to the observations during the morning (Figs. 7 and 10). During the afternoon, since the lake behaves as a single surface (see appendix A), we found an acceptable agreement during the afternoon when compared to the observations (Fig. A2).

APPENDIX D

Land–Atmosphere Conceptual Model

a. Initial and boundary conditions

Table D1 provides information on the CLASS model settings for representing the water surface in the Salar del Huasco. The table includes general information such as geographical location, day, and time setup for modeling. It also provides the initial and boundary conditions for the free troposphere, the mixed layer, the surface, and the saltwater body.

b. Ground heat flux

CLASS as a conceptual model has very limited resources to represent the complexity of the ground heat flux $G$ (see appendix A). The model only works with an albedo and two soil layers, where everything that is not reflected heats the surface. With the original configuration, we found a poor agreement between the modeled and observed $T_s$ during the whole day. Consequently, we modified the model so CLASS could internally estimate $G$ as the residual of the energy budget ($G = R_N - H - LE$). A good agreement was then found when compared to the observations (Fig. A2). Since the model already constrains the surface energy balance closure, this means that we transferred the EBC from $T_s$ to $G$ due to its very limited resources to represent $G$.

c. Prescribed terms

The regional processes of subsidence and advection must be first estimated and then prescribed in the land–atmosphere conceptual model. We prescribed the wind divergence, used to parameterize the subsidence in CLASS, from Aguirre-Correa et al. (2023). For the mass, temperature, and moisture advection terms, we use the results from the WRF Model developed in this research. The approach to estimate each of the terms and their results are detailed below, together with the values included in the land–atmosphere model.

1) Subsidence

Wind divergence is used to parameterize the subsidence velocity in CLASS. We prescribed wind divergence from Aguirre-Correa et al. (2023) who study the ABL over the desert surface in the Salar del Huasco. They estimated the divergence from WRF runs at 1-km horizontal resolution by averaging the vertical motion in the 1-km layer above the ABL height and using Eq. (7) in their article. They found positive values that reach $\sim 8 \times 10^{-5}$ m s$^{-1}$ during the afternoon, which cause a negative subsidence velocity that reduces the ABL height. For further details, we refer to Aguirre-Correa et al. (2023).
2) MASS ADVECTION

We assume that the mass advection $h_{\text{adv}}$ can influence either the ABL height or the RL height but not both layers at the same time. As long as there is a residual layer, $h_{\text{adv}}$ influences the RL height. Otherwise, the mass advection acts over the ABL.

The WRF profile analysis (section 3) shows that the residual layer is present until 1200 LT. Consequently, the mass advection between 0900 and 1200 LT contributes to changes in the residual layer height. As the WRF Model is unable to separate the ABL and the RL by itself, we include the mass advection over the residual layer by prescribing the RL height as a tendency from the WRF profile analysis. At 1230 LT, the residual layer is no longer recognized and a deep mixed layer is observed. During the afternoon, we therefore estimate the mass advection following the method described in Aguirre-Correa et al. (2023) in which $h_{\text{adv}}$ is estimated from the wind and horizontal gradients. For further details, we refer to Aguirre-Correa et al. (2023). However, due to the abrupt changes observed in the estimation using the gradient method, we also calculate $h_{\text{adv}}$ as a closure term from the mixed-layer equations [Eq. (2)]. By applying the parameterization that CLASS uses for the sub-sidence velocity the boundary layer height tendency ($\frac{dh}{dt}$) becomes

$$ \frac{dh}{dt} = \dot{w} + -\text{Div}(U_h)h + h_{\text{adv}}, $$

where $\dot{w}$ is the entrainment velocity, Div($U_h$) is the divergence of the mean horizontal wind ($U_h$), and $h$ is the boundary layer height. Wind divergence is a prescribed term (section 1). From the WRF Model, we directly obtain the tendency $\frac{dh}{dt}$ and $h$ at a 30-min time step. In the mixed-layer theory, the entrainment velocity can be estimated as (van Heerwaarden et al. 2014)

| Table D1. Initial and boundary conditions for the land–atmosphere conceptual model in CLASS to represent the saltwater surface in the Salar del Huasco. Initial conditions are defined for 0900 LT (1200 UTC). |
|-----------------|-------------------------------------------------|-----------------|
| Variable        | Description and unit                          | Value           |
| Basic settings  |                                                |                 |
| Lat             | Latitude (°)                                  | −20.35          |
| Lon             | Longitude (°)                                 | −68.90          |
| DOY             | Day of the year (—)                           | 325             |
| Time            | Initial time (UTC)                            | 12              |
| $\Delta t$      | Time step (s)                                 | 60              |
| $\gamma_\theta$ | Potential temperature lapse rate (K m$^{-1}$) |                 |
| $\gamma_q$      | Specific humidity lapse rate (g kg$^{-1}$)     |                 |
| Free atmosphere |                                                |                 |
| $\gamma_\theta$ | Potential temperature lapse rate (K m$^{-1}$) | 150 $< z < 500$ m: 0.0076 |
| $\gamma_q$      | Specific humidity lapse rate (g kg$^{-1}$)     | $z > 500$ m: 0.0041 |
| $\gamma_q$      | Specific humidity lapse rate (g kg$^{-1}$)     | 150 $< z < 500$ m: −0.0021 |
| $\gamma_q$      | Specific humidity lapse rate (g kg$^{-1}$)     | $z > 500$ m: −0.0001 |
| Mixed layer     |                                                |                 |
| $h_i$           | Initial boundary layer height (m)             | 150             |
| $\theta_i$      | Initial potential temperature (K)             | 272             |
| $\Delta \theta$ | Initial potential temperature jump (K)        | 2.0             |
| $q_i$           | Initial specific humidity (g kg$^{-1}$)       | 2.1             |
| $\Delta q$      | Initial specific humidity jump (g kg$^{-1}$)   | −0.7            |
| Surface         |                                                |                 |
| $w_\theta$      | Initial surface kinematic heat flux (K m s$^{-1}$) | 0.0093           |
| $w_q$           | Initial surface kinematic moisture flux (g kg$^{-1}$ m s$^{-1}$) | 0.0083           |
| $\nu$           | Surface friction velocity (m s$^{-1}$)         | 0.05            |
| $z_{0m}$        | Roughness length for momentum (m)              | $4 \times 10^{-5}$ |
| $z_{0h}$        | Roughness length for scalars (m)               | $4 \times 10^{-6}$ |
| $\alpha$        | Surface albedo (—)                            | 0.12            |
| $T_s$           | Initial surface temperature (K)               | 275             |
| Saltwater body  |                                                |                 |
| $T_{\text{water}}$ | Temperature top layer (K)                   | 270             |
| $r_{\text{g, min}}$ | Minimum resistance evaporation (s m$^{-1}$)    | 0               |
| $w_g$           | Volumetric water content top layer (m$^3$ m$^{-3}$) | 1.0             |
| $w_2$           | Volumetric water content deeper layer (m$^3$ m$^{-3}$) | 1.0             |
| $w_{\text{sat}}$ | Saturated volumetric water content (m$^3$ m$^{-3}$) | 1.0             |
| $C_{1, \text{sat}}$ | Coefficient force term moisture (—)           | 0.0             |
| $C_{2, \text{ref}}$ | Coefficient restore term moisture (—)        | 0.0             |
where $\beta$ is 0.2, $R_e$ is the gas constant for moist air, $R_d$ is the gas constant for dry air, $\rho$ is the air density, $c_p$ is the specific heat at constant pressure, and $L_v$ is the latent heat of vaporization, which are all near constants. The sensible and latent heat terms of $\dot{H}_s$ and $\dot{H}_l$ contribute to the ABL collapse over the lake. We can therefore solve mass advection from Eq. (D1). However, since the time resolution is too coarse to estimate accurate tendencies, we fit a polynomial function to each variable to decrease the time step from 30 to 1 min and then solve $\dot{w}_a$.

**3) Temperature and Moisture Advection**

The temperature and moisture advection affect simultaneously the ABL, the RL, and the free troposphere (Fig. 5). In the ABL, we estimate the temperature and moisture advection using the gradient method (Aguirre-Correa et al. 2023), as well as a closure term from the mixed-layer equations [Eq. (1)]. By using the definition of the entrainment velocity $w_e$ as the ratio between the entrainment fluxes $(\dot{w}^* \dot{\Psi})_e$ and the jumps $\Delta \Psi$ at the inversion zone, and by replacing the kinematic fluxes for their dynamic form, we have the following expressions for $\dot{\Psi} = \dot{\theta}$ and $\dot{\Psi} = \dot{q}$, respectively:

$$\frac{d\dot{\theta}}{dt} = \frac{H}{\rho c_p} + \frac{w_e \Delta \theta}{h} + \overline{\Psi}_{adv}, \quad (D3)$$

$$\frac{d\dot{q}}{dt} = \frac{LE}{\rho L_v} + \frac{w_e \Delta q}{h} + \overline{\Psi}_{adv}, \quad (D4)$$

Similarly to the mass advection, we can estimate all the terms from the WRF Model to solve temperature and moisture advection from Eqs. (D3) and (D4), respectively.

Figure D2 presents the advection results for $\dot{\theta}$ and $\dot{q}$ (dotted line) in the ABL using the gradient method (WRF$_{\text{grad}}$) and the mixed-layer equations (WRF$_{\text{MLE}}$), along with the values prescribed in CLASS (continuous line). We do not adhere strictly to the WRF values, as the overestimation of the surface fluxes (see Figs. 7 and 10) leads to an overestimation of the surface and entrainment contributions to the budgets. In reality, these contributions are related to advection or other nonlocal processes, which we account for in the $\overline{\Psi}_{adv}$ term. We therefore made minor adjustments to achieve the best fit with $\dot{\theta}$ and $\dot{q}$ observations. Since temperature and moisture advection are included in the last experiment (see Table 1), all other processes are already accounted in the budget before using $\overline{\Psi}_{adv}$ as a closure term. From Fig. D2, three temperature and advection regimes are observed during the day: a morning regime in which warm and dry air is advected from the desert, a transition regime in which warm and moist air is brought to the water surface when the regional air masses arrive, and an afternoon regime in which dry and cold air interacts with the ABL over the lake after a longer influence of the regional flow.
mass advection in which we include the warming/cooling and drying processes in this layer by prescribing the RL characteristics from the WRF profile analysis (section 3). Hence, when temperature and moisture advection are included in the last experiment, the initial profile at 0900 LT is no longer undisturbed in terms of the RL as in the experiment where only mass advection is prescribed.

Finally, in the free troposphere, we followed the method described in Aguirre-Correa et al. (2023) to estimate the advection of temperature and moisture in the upper layer. To provide a more robust estimation, we followed the recommendation of averaging advection in the 1-km layer above the ABL height [for further details, we refer to Aguirre-Correa et al. (2023)]. However, we found that advection in the upper layer is negligible (~0.3 K h⁻¹ and ~-0.2 g kg⁻¹ h⁻¹ on average after midday). Note that this advection would be only relevant during the afternoon, as during the morning, the ABL over the lake is controlled by the interaction with the residual layer and not the free troposphere.

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