



The Marine Stratocumulus-topped Boundary Layer across the Pacific-Atacama Desert transition

Vicente Espinoza^{1,2}, Oscar Hartogensis¹, Felipe Lobos-Roco^{2,3}, and Jordi Vilà-Guerau de Arellano¹

¹Meteorology and Air Quality group, Wageningen University, Wageningen, The Netherlands

²Centro UC Desierto de Atacama, Pontificia Universidad Católica de Chile, Santiago, Chile

³Facultad de Agronomía y Sistemas Naturales, Pontificia Universidad Católica de Chile, Santiago, Chile

Correspondence: Vicente Espinoza (vicente.espinoza@wur.nl)

Abstract. A semi-permanent stratocumulus-topped boundary layer (STBL) is advected daily from the Southeast Pacific toward the Atacama Desert, producing coastal fog that represents a potential water input for ecosystems and communities in this hyper-arid region ($\sim 1 \text{ mm yr}^{-1}$). The STBL is maintained by the balance between synoptic-scale subsidence and entrainment of dry, warm air, driven by cloud-top radiative and evaporative cooling, and surface fluxes over sea and land. These processes are well studied over the open ocean, but their evolution across the ocean-to-desert transition remains poorly understood. Here, we quantify cloud cover fraction (CCF) and STBL height tendency across the Pacific-Atacama Desert transition (from $\sim 500 \text{ km}$ offshore to $\sim 50 \text{ km}$ inland) using three years (2022–2024) of GOES satellite and ERA5 reanalysis, with a zero-order discontinuity approach and dry-cloud formulation. Our results show that CCF decreases from offshore (0.71) to inland (0.12), with higher values in winter (0.12–0.75) than in summer (0.09–0.57). Oceanic CCF variability spans days to weeks, whereas inland CCF variability is dominated by a strong diurnal cycle. The STBL height tendency shifts from near-equilibrium between processes offshore, with a weak positive tendency ($\sim 0.5 \text{ cm s}^{-1}$) driven mainly by cloud-top radiative cooling ($\sim 58\%$), to a strongly perturbed inland regime ($\sim 3 \text{ cm s}^{-1}$), where surface fluxes contribute mainly to entrainment in summer ($\sim 43\%$) and to subsidence in the other seasons ($\sim 47\%$). These results show how subsidence and entrainment govern STBL persistence and breakdown across the coastal desert, with implications for ecosystems and regional climate.

1 Introduction

In the coastal Atacama Desert, where annual rainfall is $\sim 1 \text{ mm}$ (Dirección Meteorológica de Chile [DMC], 2026), fog is the main source of water for ecosystems and can provide supplementary water for local communities (Cereceda et al., 2008; García et al., 2021). Fog originates from a semi-permanent stratocumulus-topped boundary layer (STBL) over the Southeast Pacific Ocean (Bretherton et al., 2010), which is advected daily towards the coastal mountain range of the Atacama Desert (Pallauta Pérez et al., 2025). This transition from the marine STBL over the Pacific Ocean to coastal fog in the Atacama Desert extends from several hundred kilometers offshore to up to 50 kilometers inland, influencing the regional radiation budget and the coastal climate (del Río et al., 2021; Espinoza et al., 2024; Lobos Roco et al., 2018; Schween et al., 2022; Wood, 2012). The Pacific to Atacama Desert (PAD) transition is characterized by extensive stratocumulus (Sc) decks, with spatial and temporal variations in cloud cover fraction (CCF). Over the Southeast Pacific Ocean, the annual mean CCF typically exceeds 40%



25 and can reach around 60% (Wood, 2012). Regional studies along a 100 km offshore-to-inland transect indicate that CCF decreases from 80% offshore to 20% inland during the austral winter (Espinoza et al., 2024). Both seasonal and diurnal variations are observed: CCF tends to be higher in winter and during the night or early morning, and lower in summer and in the afternoon (Keim-Vera et al., 2024; Muñoz et al., 2016). These variabilities are driven by a combination of multi-scale physical processes. On seasonal timescales, variations in CCF are primarily influenced by the strength of the thermal inversion. This is set by synoptic-scale subsidence, associated with the seasonal cycle of the Southeast Pacific Anticyclone (Ancapichún and Garcés-Vargas, 2015), and by sea surface temperature (Espinoza et al., 2024; Vicencio Veloso et al., 2024). The inversion is strongest over the open ocean and weakens toward the coast (Espinoza et al., 2024; Rahn and Garreaud, 2010a; Wood, 2012). It regulates the efficiency of entrainment at the cloud top, which is enhanced by cloud-top radiative cooling and modulated by surface fluxes, with moisture additionally affecting entrainment through evaporative cooling during the mixing of dry air into the cloud layer (Deardorff, 1976; Lilly, 1968). On diurnal timescales, CCF exhibits a pronounced diurnal cycle associated with the inland intensification of subsidence driven by the diurnal heating of the Andean slope and the resulting sea-land breeze circulation (Lobos-Roco et al., 2021; Rutllant et al., 2003, 2013). This diurnal forcing modulates turbulent mixing and entrainment through variations in surface fluxes (Rutllant et al., 2003). These processes control the evolution and depth of the STBL offshore, forming a balance in which subsidence and entrainment dominate under mean conditions, while horizontal advection becomes important during synoptic disturbances linked to midlatitude systems (Rahn and Garreaud, 2010b).

Previous studies have documented aspects of the CCF variability in the PAD transition, while studies in other regions have shown that entrainment driven by surface fluxes plays an important role in stratocumulus-to-cumulus transitions (Pedruzo-Bagazgoitia et al., 2020). Although the physical processes governing the STBL are generally well understood over the open ocean (Bretherton et al., 2004; Wood et al., 2011), how these processes interact and evolve across the full PAD transition remains poorly characterized. This gap is partly due to the scarcity of in-situ measurements and to the fact that current global climate models do not fully resolve these processes, often underestimating CCF and producing boundary layers that are too shallow (Chammas et al., 2026; Hannay et al., 2009; Lin et al., 2014). A better understanding and quantification of the processes governing STBL evolution is therefore needed to improve their representation in large-scale models (Pedruzo-Bagazgoitia et al., 2020).

50 This study poses the following research questions: 1) How does cloud cover fraction vary across the PAD transition at daily and seasonal timescales? 2) Which physical processes, specifically synoptic-scale subsidence and entrainment driven by the interplay between cloud-top radiative and evaporative cooling, and surface turbulent fluxes, control the evolution of STBL across the PAD transition, and at which scales do they act? To answer these questions, we combine satellite data, ERA5 reanalysis, and the mixed-layer theory to quantify the contributions of these drivers to STBL evolution across the PAD transition.

55 2 Theory

To analyze the daily and seasonal variability of CCF across the PAD transition, we adopted a conceptual framework based on mixed-layer theory (Lilly, 1968; van Zanten et al., 1999), as stratocumulus clouds are typically embedded within a well-mixed



boundary layer. This framework provides a physically based approach to describe STBL evolution and quantify the contributions of the different processes governing the transition.

- 60 In this framework, the STBL height (h), defined here as the cloud-top height, is a key variable controlling CCF because it regulates the depth of the marine boundary layer that sustains stratocumulus clouds and controls the vertical distribution of moisture and turbulence. h therefore influences cloud formation and dissipation, which in turn control the cloud thickness and persistence. Changes in h reflect the combined effects of synoptic-scale subsidence and entrainment, with the latter driven by turbulence generated by cloud-top radiative and evaporative cooling and by surface turbulent fluxes. At the synoptic scale, subsidence associated with large-scale atmospheric circulation sets the overall structure and thickness of the STBL. Compression of the boundary layer by subsidence lowers h and thins clouds, decreasing CCF. At the cloud-top, entrainment regulates the exchange of warm and dry free-tropospheric air across the inversion, influencing STBL growth and cloud evolution (Albrecht et al., 1995; Wood, 2012; Zheng et al., 2011). Cloud-top radiative cooling enhances turbulence and vertical mixing within the cloud layer, while surface fluxes further modulate entrainment through surface buoyancy forcing. Both processes influence the potential temperature and moisture jumps across the inversion. The temporal evolution of the STBL height (h) can be described by the following tendency budget:

$$\frac{\partial h}{\partial t} = \underbrace{w_{\text{sub}}}_{\text{Synoptic}} + \underbrace{w_{\text{ent}}}_{\text{Cloud-top}} + \underbrace{h_{\text{adv}}}_{\text{Advection}} \quad (1)$$

- Each term in h tendency budget represents a contribution to the rate of change of h and is expressed in units of velocity (cm s⁻¹). w_{sub} and w_{ent} represent the velocities associated with synoptic-scale subsidence and cloud-top entrainment, respectively. While h_{adv} represents the horizontal advection of the STBL height by the wind (Rahn and Garreaud, 2010b). This formulation provides a quantitative framework to assess the relative contributions of each term and to link observed CCF variability to the STBL evolution. The following sections provide a detailed description of each component of the equation.

2.1 Boundary layer height

- 80 The STBL height, h , influences CCF because a deeper boundary layer can sustain a thicker and more extensive stratocumulus cloud layer, whereas a shallower STBL tends to reduce cloud thickness and cloud cover (Wood, 2012). Changes in h therefore provide a useful measure of the processes governing cloud formation, persistence, and dissipation across the PAD transition. In the mixed-layer framework, h tendency represents the balance between processes that deepen the STBL, such as cloud-top entrainment, and those that limit its growth, such as synoptic-scale subsidence.
- 85 Since h can be defined and estimated using different approaches (Seidel et al., 2010), including methods based on thermodynamic properties (potential temperature), atmospheric stability (Richardson number), and model-satellite retrievals (Heidinger, 2012), we use complementary methods to characterize its variability across the PAD transition.



2.2 Synoptic-scale subsidence

Synoptic-scale subsidence (w_{sub}), characteristic of the descending branch of the Hadley cell, represents the downward motion
90 of air associated with large-scale circulation (Serpetzoglou et al., 2008). In pressure coordinates, vertical motion is dynamically
linked to the horizontal wind field through the mass continuity equation:

$$\frac{\partial \omega}{\partial p} = -\nabla_h \cdot \mathbf{U}_h \quad (2)$$

where $\mathbf{U}_h = (u, v)$ is the horizontal wind vector, and $\nabla_h \cdot \mathbf{U}_h$ denotes the divergence of the horizontal wind field. The vertical
95 velocity (ω) is obtained directly from ERA5 as a model diagnostic variable in pressure coordinates (Pa s^{-1}). Subsidence
velocity (w_{sub}) is then expressed in geometric coordinates as

$$w_{\text{sub}} = -\frac{\omega}{\rho g} \cdot 100 \quad (3)$$

where the negative sign follows from the hydrostatic relation, such that $w_{\text{sub}} < 0$ indicates subsidence and $w_{\text{sub}} > 0$ indicates
100 upward motion. Here, ρ is the air density (kg m^{-3}), and g is the gravitational acceleration (9.81 m s^{-2}). The factor 100 converts
from m s^{-1} to cm s^{-1} . w_{sub} is evaluated at the second vertical level above h .

Consistent with the prevalence of subtropical high-pressure systems, w_{sub} is predominantly negative over the Southeast Pacific
Ocean. This downward motion strengthens the inversion and limits the growth of the STBL, thereby favoring stratocumulus
formation in a continuous layer and inhibiting the development of deep clouds (Klein and Hartmann, 1993). As a result,
105 subsidence promotes a more persistent cloud layer and higher CCF, while weaker subsidence is associated with a weaker
inversion, enhanced entrainment, and reduced CCF.

2.3 Entrainment

Entrainment is defined as the mixing of air from the free troposphere into the STBL. The entrainment flux is driven by tur-
bulence generated at the surface and at the cloud top, while its efficiency is regulated by the thermodynamic jump across the
110 inversion (Lilly, 1968). Surface sensible and latent heat fluxes generate buoyancy-driven eddies that can penetrate through
the temperature inversion. Similarly, cloud-top cooling due to longwave-radiation and evaporation generates turbulence that
causes mixing across the inversion. Through the erosion of the inversion, entrainment links these processes to cloud evolution
by deepening the STBL. It also imports warm and dry free-tropospheric air, reducing cloud liquid water content and cloud
thickness, enhancing decoupling, and weakening cloud persistence. Together, these processes establish a coupled feedback
115 in which radiative cooling, surface fluxes, and moisture modulate entrainment, which in turn governs cloud cover variability
(Lilly, 1968; Stevens, 2005; Wood, 2012).



The entrainment process can be represented using either a zero-order or a first-order model. The zero-order jump model idealizes the inversion as an infinitesimally thin layer separating the well-mixed STBL from the free troposphere (Lilly, 1968; van Zanten et al., 1999). Under this approximation, the entrainment velocity is related to radiative and evaporative cooling, surface
120 fluxes, and the thermodynamic jump across the inversion. In contrast, a first-order jump model represents the inversion as a layer of finite thickness (van Zanten et al., 1999), allowing the internal thermodynamic structure of the entrainment zone to be resolved. In this study, the zero-order model is adopted as the focus is on the bulk evolution of the STBL rather than on resolving the detailed structure of the entrainment zone.

As a simplified representation of the entrainment process, we first consider the formulation of a radiation-absorbing smoke
125 cloud after Lilly (1968), which we will refer to as the "dry-cloud case", in which no phase changes occur during mixing, i.e. the cloud-top evaporative cooling is deactivated. Next, we consider the "wet-cloud case", which includes the evaporative cooling effect. In the following sub-sections the parameterizations used in the dry- and wet-cloud formulations are outlined.

2.3.1 Dry-cloud entrainment

Following a similar approach as used by Lilly (1968), the dry-cloud entrainment velocity $w_{ent,dc}$ can be expressed as

130
$$w_{ent,dc} = w_{flux,dc} + w_{rad,dc} \quad (4)$$

where $w_{flux,dc}$ represents the entrainment contribution associated with surface fluxes, while $w_{rad,dc}$ represents the contribution associated with cloud-top radiative cooling.

The entrainment buoyancy flux driven by surface fluxes is assumed proportional to the surface buoyancy flux through a
135 dimensionless coefficient β (Lilly, 1968; van Zanten et al., 1999), which is taken as 1, based on the assumption that the heat transported downward from above the inversion balances that transported upward. In this framework, β represents the efficiency with which surface-generated turbulence drives entrainment across the inversion. The corresponding surface fluxes velocity is then expressed as

$$w_{flux,dc} = -\frac{\overline{(w'\theta'_v)_h}}{\Delta\theta_{v,h}} = \frac{\beta \overline{(w'\theta'_v)_s}}{\Delta\theta_{v,h}} \quad (5)$$

140

where $\overline{w'\theta'_v}$ denotes the buoyancy flux evaluated at the boundary layer height (h) and the surface (s), while $\Delta\theta_{v,h}$ is the inversion jump in virtual potential temperature. The buoyancy flux is approximated as $\overline{w'\theta'} + 0.61 \langle \langle \theta \rangle \overline{w'q'} \rangle$ following Stull (1988).

The cloud-top radiative cooling for the dry-cloud case ($w_{rad,dc}$) is the main driver of convection inside the stratocumulus and
145 typically the dominant term in the STBL energy budget, with its magnitude depending on cloud fraction and liquid water content (Wood, 2012). The cloud-top radiative cooling is calculated as



$$w_{\text{rad,dc}} = -\frac{a_c}{\bar{\rho}c_p} \frac{\Delta F_{z,h}}{\Delta \theta_{v,h}} \quad (6)$$

where ρ is the air density, c_p is the specific heat of dry air at constant pressure ($1004.67 \text{ J kg}^{-1} \text{ K}^{-1}$), a_c is an efficiency parameter, which is 1 for the dry-cloud case, and $\Delta F_{z,h}$ is the cooling rate due to longwave radiative divergence in the entrainment zone calculated as

$$\Delta F_{z,h} = F_{\text{LW}}(h^+) - F_{\text{LW}}(h^-) \quad (7)$$

where F_{LW} is the net longwave radiative flux, defined as the downward minus the upward longwave radiation, evaluated at the inversion layer. In a first-order model, the longwave flux divergence across the entrainment zone can be decomposed into contributions from the inversion and the layer below, $\Delta F_{z,h} = \Delta F_i + \Delta F_b$, where ΔF_i and ΔF_b represent the respective contributions. In the zero-order model adopted here, these contributions are not treated separately but are combined into the total flux divergence $\Delta F_{z,h}$.

Although radiative cooling represents a loss of energy at cloud top ($\Delta F_{z,h} < 0$), the cloud-top cooling generates negative buoyancy that enhances turbulent mixing within the boundary layer. As a result, $w_{\text{rad,dc}}$ is typically positive, contributing to the growth of the STBL by promoting turbulent mixing (Nicholls and Turton, 1986; Wood, 2012). This enhanced mixing warms the mixed layer, reducing the potential temperature jump and increasing entrainment, while also reducing the cloud-top radiative cooling, creating a negative feedback that regulates STBL growth.

2.3.2 Wet-cloud entrainment

In the wet-cloud formulation, the cloud-top cooling, which enhances entrainment, is from both radiative and evaporative cooling (Lilly, 1968; Deardorff, 1980; Randall, 1980). This provides a more complete representation of the processes governing entrainment at the expense of introducing more complexity as the terms cannot be treated independently as in the dry-cloud case. Specifically, evaporative cooling changes the temperature at the cloud-top, thereby affecting the longwave radiative cooling and the inversion strength. This feedback has been parameterized by van Zanten et al. (1999), whose formulations we will use for the wet-cloud entrainment velocity, $w_{\text{ent,wc}}$:

$$w_{\text{ent,wc}} = w_{\text{flux,wc}} + w_{\text{ctc,wc}} = -\frac{\overline{(w'\theta'_v)_h}}{\Delta_2} - \frac{a_c}{\bar{\rho}c_p} \frac{\Delta F_{z,h}}{\Delta_2} \quad (8)$$

where $w_{\text{ctc,wc}}$ represents the cloud-top cooling contribution to the entrainment due to both radiative and evaporative cooling. Eq. 8 follows the same decomposition as Eq. 4, with the individual contributions expressed analogously to Eqs. 5 and 6,



175 respectively, with two adjustments to account for the evaporative cooling effect. Following the zero-order model, the total longwave flux divergence $\Delta F_{z,h}$ is used instead of the inversion contribution ΔF_i from the original first-order model (van Zanten et al., 1999), consistent with the assumption of an infinitesimally thin inversion. The first adjustment involves a change in inversion strength, from $\Delta\theta_{v,h}$ in the dry-cloud case to Δ_2 in the wet-cloud case. The second adjustment involves adding the evaporative cooling effect to the radiative cooling entrainment contribution through the coefficient a_c , which incorporates
180 the effect of phase changes on entrainment and depends on the slope of the saturated specific humidity curve, dq_s/dT :

$$a_c = \frac{1 + 1.61(\theta_0 \frac{dq_s}{dT})}{1 + \frac{L_v}{c_p} \frac{dq_s}{dT}} \simeq 0.5 \quad (9)$$

where θ_0 denotes the reference potential temperature of the boundary layer and L_v is the latent heat of vaporization ($2.45 \cdot 10^6 \text{ J kg}^{-1}$). Note that in the dry-cloud case $dq_s/dT = 0$ and therefore $a_c = 1$. The inversion strength, Δ_2 , is characterized
185 considering conserved thermodynamic variables through:

$$\Delta_2 = a_c \Delta\theta_l + b_c \Delta q_t \quad (10)$$

where θ_l is the liquid-water potential temperature, q_t is the total water, and $b_c = a_c \frac{L_v}{c_p} - \theta_0$.

Eq. 8 provides a more physically consistent representation of the thermodynamic contrast across the cloud-top inversion,
190 accounting for both thermal and moisture differences between the cloud layer and the overlying free troposphere. In other words, the wet formulation extends the dry framework by incorporating phase changes into both the radiative and inversion-strength terms.

2.4 Advection

Advection represents the horizontal transport of h by the wind (Rahn and Garreaud, 2010b). This process is expected to
195 be negligible (~ 0) over the open ocean, but may become relevant in coastal and inland zones, where stronger horizontal gradients are present (Aguirre-Correa et al., 2023, 2024). Because h_{adv} is difficult to estimate directly, its influence is evaluated qualitatively based on the closure of h budget. Therefore, any residual should be interpreted with caution, as it may reflect both advective effects and methodological uncertainties associated with the estimation of the other budget terms.

3 Methods

200 3.1 Study area

The study area encompasses the transition between the Pacific Ocean and the hyper-arid Atacama Desert, approximately between 18° S , 76° W and 24° S , 70° W (Fig. 1a). This region is characterized by a persistent STBL, sustained by synoptic-



scale subsidence associated with the descending branch of the Hadley cell beneath the Southeast Pacific Anticyclone (Wood, 2012), and reinforced by the cold sea surface temperatures (Espinoza et al., 2024; Vicencio Veloso et al., 2024). Together, these factors contribute to a strong ocean-inland gradient in moisture and temperature, while the transition from ocean to land also leads to changes in surface properties and associated fluxes across the PAD. Over the open ocean, the STBL remains relatively homogeneous across the 6° of latitude (18° S–24° S) spanned by the study area (Rahn and Garreaud, 2010a), whereas greater spatial variability emerges in the coastal and inland zones, where complex topography modulates the inland penetration of the STBL and fog (Keim-Vera et al., 2024; Lobos Roco et al., 2018). This distinction allows us to focus on cross-shore variability as the primary control on the system’s structure. In addition, the area is influenced by large-scale climate variability associated with ENSO (Falvey and Garreaud, 2009; Park and Leovy, 2004). We selected three years (2022–2024) to have a well-represented climatic CCF conditions. Since ENSO modulates CCF (del Río et al., 2021; Espinoza et al., 2024), we chose this period because each year represents a different ENSO phase: a cold phase in 2022 (Oceanic Niño Index (ONI) = -0.9), a warm phase in 2023 (ONI = 0.91), and a neutral phase in 2024 (ONI = 0.43) (National Climate Prediction Center, 2026). We divided this transition into three main regions (Fig. 1). Region 1 corresponds to the offshore and near coast zones (Zones 1–2), extending from 500 to 30 km offshore. This region represents an ocean-dominated environment largely unaffected by terrestrial influence and is well documented in previous studies (Bretherton et al., 2004, 2010; Jia et al., 2019; Rahn and Garreaud, 2010a, b). Region 2 corresponds to the coastal zone (Zone 3) and encompasses the coastal transition between oceanic and inland regimes, extending from 30 km offshore to 15 km inland. This area is strongly modulated by orography due to the presence of a coastal mountain range (~1000 m a.s.l.) located near the shoreline. Finally, Region 3 corresponds to the inland zone (Zone 4) where the STBL influence is still observed in the form of fog, extending from 15 to approximately 50 km inland (Lobos-Roco et al., 2024).

3.2 Research strategy

Despite the importance of entrainment in regulating the evolution of the STBL, the relative contribution of the processes controlling it remains unclear (de Lozar and Mellado, 2015). This uncertainty is reflected in existing entrainment parameterizations, which integrate radiative and evaporative cooling differently, with some formulations emphasizing the interaction between turbulence and evaporative cooling and others attributing entrainment primarily to radiative cooling and turbulence (de Lozar and Mellado, 2015). The representation of entrainment also remains challenging in models, as comparisons between numerical weather prediction models and large eddy simulations have shown that entrainment fluxes can be underestimated, leading to biases in the simulated thermodynamic and cloud properties of the STBL (Ghonima et al., 2017).

This uncertainty is particularly relevant in regions where the boundary layer evolves across contrasting environmental conditions, as in the PAD transition, where the relative importance of moist processes may vary across the transition. The PAD transition extends from a relatively homogeneous marine environment toward a more heterogeneous inland environment, where strong ocean-to-inland gradients in temperature and moisture, together with the presence of coastal topography, modify the structure and inland penetration of the STBL (Keim-Vera et al., 2024; Lobos Roco et al., 2018; Rahn and Garreaud, 2010a).

Motivated by these uncertainties, we adopt a research strategy centered on the dry-cloud case, in which phase changes are

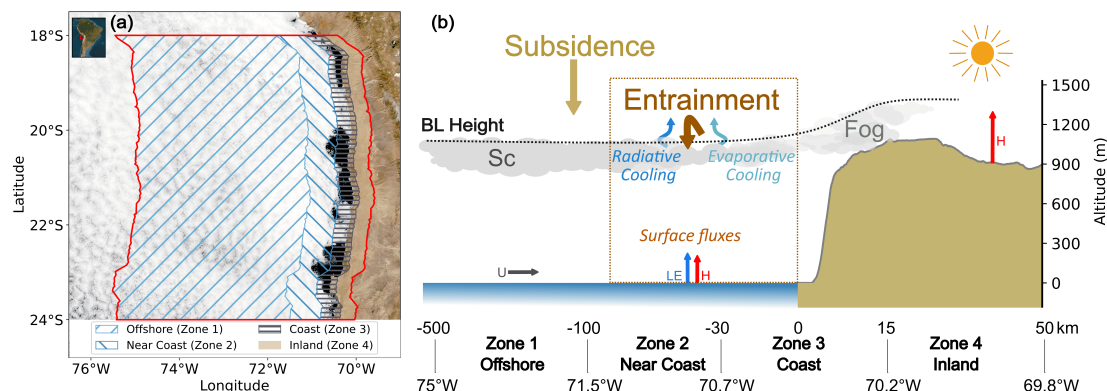


Figure 1. (a) MODIS satellite image from 15 July 2024 at 15:00 LT. The red outline delineates the study area and the hatched pattern indicates its subdivision into zones. (b) Vertical cross section of the STBL across the PAD transition. Arrows indicate key processes contributing to the evolution of the boundary layer height (dashed line). Blue (latent heat flux, LE) and red (sensible heat flux, H) arrows indicate the surface fluxes, which drive local-scale turbulence and influence entrainment at the cloud-top interface. The gray arrow (U) indicates wind direction. Although the prevailing wind is south-southwesterly, consistent with the circulation around the high-pressure system, it is depicted as perpendicular to the coast to represent its dominant cross-shore component.

neglected and the STBL evolution is driven primarily by radiative cooling and surface fluxes (Lilly, 1968). This framework forms the main basis of our analysis, allowing us to examine entrainment without the additional complexity and uncertainties associated with moist processes. However, moisture can also influence entrainment through processes such as evaporative cooling, latent heating, and cloud microphysics. We therefore consider moisture, as part of the discussion, through a sensitivity analysis to assess its effects on entrainment. A full characterization of these interactions between turbulence, radiation, and moist processes, however, would require stronger observational constraints and simulation techniques that explicitly resolve turbulence, such as large-eddy simulations (van der Dussen et al., 2014, 2015; de Lozar and Mellado, 2015; Ghonima et al., 2017).

3.3 Data sources and processing

3.3.1 Data sources

To assess cloud cover fraction (CCF) variability, we used GOES-16 ABI Level 2 Cloud and Moisture Imagery products (Schmit et al., 2017), with 10-min temporal resolution and 2-km spatial resolution, obtained from the Google Earth Engine platform (Gorelick et al., 2017). The dataset provides visible and infrared channels used for cloud detection, as demonstrated in previous studies (del Río et al., 2021; Espinoza et al., 2024; Keim-Vera et al., 2024; Lobos-Roco et al., 2024). The GOES cloud-top height product, obtained from NOAA's Comprehensive Large Array Data Stewardship System (CLASS), was also used to complement cloud vertical structure information (Heidinger, 2012).



To analyze the drivers, we used ERA5 reanalysis (Hersbach et al., 2020), which provides hourly atmospheric variables at 0.25° spatial resolution, including temperature, humidity, vertical velocity, low cloud cover, boundary layer height, and surface fluxes.

In situ observations from Antofagasta Airport (23.45° S, 70.45° W) and the StraToFog field campaign (Lobos-Roco et al., 2026) were also used to estimate and evaluate the boundary layer height (h). The Antofagasta Airport dataset consists of daily radiosonde launches at 12 UTC (corresponding to 08:00-09:00 local time depending on the time of year), providing vertical profiles of temperature, humidity, pressure, and wind. The StraToFog field campaign includes radiosonde and drone observations along an offshore-inland transect conducted on 25 and 26 July 2024 under typical winter stratocumulus conditions, as described in detail in (Lobos-Roco et al., 2026).

The boundary layer height (h) was estimated and evaluated using complementary datasets and methodologies, which are summarized in Table 1

Source	Method	Period	Resolution / Sampling	Reference
ERA5	Bulk Richardson number ($Ri_{\text{bulk}} > 0.25$)	2022–2024	Gridded (hourly)	Hersbach et al., 2020
ERA5	Potential temperature maximum gradient	2022–2024	Gridded (hourly, 137 vertical levels)	Seidel et al., 2010
GOES	Satellite cloud-top height product	2022–2024	Gridded (hourly)	Heidinger, 2012
Morro Moreno Airport	Potential temperature maximum gradient	2022–2024	Point (12 UTC; ~5 m)	Seidel et al., 2010; DMC, 2026
StraToFog campaign	Potential temperature maximum gradient	25–26 July 2024	Points (campaign; ~5 m)	Seidel et al., 2010; Lobos-Roco et al., 2026

Table 1. Datasets and methods used to estimate boundary layer height (h). Resolutions refer to the type of data, temporal resolution (e.g. hourly or specific time) and vertical resolution, where applicable.

3.3.2 Data processing

CCF was derived following del Río et al. (2021), Espinoza et al. (2024), and Keim-Vera et al. (2024), combining daytime and nighttime detection methods, and excluding time-steps with more than 10% missing pixels in any band. The detection method is automatically applied based on illumination and surface temperature conditions.

Daytime detection uses a change detection technique based on visible band reflectance (0.59–0.69 μm), allowing cloud identification under sufficient illumination. Cloud-free mosaics were created monthly at up to four times of day (sunrise, late morning, midday, and sunset), depending on solar illumination conditions (details in A1). Absolute differences between observed images



and cloud-free mosaics were classified using thresholds based on del Río et al. (2021) (details in A1), with values above these classified as cloud presence. Nighttime detection uses the thermal difference between the shortwave ($3.8\mu\text{m}$) and longwave ($10.9\mu\text{m}$) infrared bands, allowing cloud identification when illumination is insufficient for the daytime method. The thermal difference ($3.8\text{--}10.9\mu\text{m}$) was classified using a threshold of -2 K , with values below this classified as cloud presence (adapted from del Río et al., 2021; Lee et al., 1997). In both methods, the longwave thermal infrared band ($10.9\mu\text{m}$) was used as a high-low cloud filter with a 273 K threshold: values below 273 K were classified as high clouds, whereas values above this threshold were classified as low clouds following del Río et al. (2021).

To mitigate inconsistencies during day-night transitions (Cermak and Bendix, 2008; Egli et al., 2017; Keim-Vera et al., 2024), daytime and nighttime cloud methods were first applied independently for all available time-steps. The two products were then merged into a continuous 10-minute time series by selecting the maximum pixel value at each time-step, given the complementary validity of daytime and nighttime retrievals. From this merged time series, an hourly mean was computed for each pixel, assigning a value of 1 if cloud presence was detected for more than 30 minutes within an hour, and 0 otherwise. Finally, a 3×3 pixel average was applied to derive hourly cloud cover fractions at a spatial resolution of $\sim 6\text{ km}$.

The detection methods have been previously validated under diurnal and nocturnal conditions across hyper-arid to Mediterranean zones, with reported accuracies ranging from 70% to 96% (del Río et al., 2021; Espinoza et al., 2024). In our study, consistency was assessed through visual inspection, comparison with earlier datasets (Espinoza et al., 2024; Keim-Vera et al., 2024), and airport observations (Diego Aracena and Andres Sabélla Airports). This 10-minute validation yielded an accuracy of 86% and a false alarm rate of 19%, demonstrating that the method reliably captures cloud presence at high temporal resolution in agreement with previous research (Minnis and Harrison, 1984; Muñoz et al., 2016; Rozendaal et al., 1995; Wood, 2012).

Prior to analysis, ERA5 temperature and humidity from the model and pressure levels (137 and 37 levels, respectively), as well as vertical velocity from pressure levels, were obtained using the $(LS)^2D$ python toolkit (van Stratum et al., 2023). The toolkit was used to extract the ERA5 variables and interpolate them from their native vertical coordinates onto a regular height grid with 50 m vertical spacing up to 2500 m . Temperature and humidity were subsequently used to derive radiative quantities. To ensure that only stratocumulus conditions were analyzed, ERA5 low cloud cover was used to filter the data, retaining periods with values ≥ 0.2 . This criterion isolates boundary-layer dominated conditions consistent with stratocumulus regimes.

For both ERA5 and observations, thermodynamic jumps were calculated following the zero-order jump approach. For each thermodynamic vertical profile, the boundary layer mean was obtained by averaging the profile below the diagnosed boundary layer height (h). The above-inversion profile was fitted with a linear regression from $h + 100\text{ m}$ to 2500 m for StraToFog and from $h + 500\text{ m}$ to 2500

,m for ERA5, and the fitted line was extrapolated to h to obtain the above-inversion value at h . The different lower bounds were chosen to account for the different vertical resolutions of StraToFog and ERA5, ensuring that the linear fit is representative of the above-inversion profile. The corresponding jumps, including $\Delta\theta_v, \Delta\theta_l, \Delta q_t, \Delta F_z$, were calculated as the difference between the above-inversion value at h and the boundary layer mean.

The net longwave radiative flux was computed using the Tripleclouds solver of the European Center for Medium-Range



Weather Forecasts (ECMWF) radiation scheme ecRad (Hogan and Bozzo, 2018). Tripleclouds was chosen because it provides deterministic, noise-free heating-rate estimates that are robust to changes in resolution and cloud optical properties (Shonk et al., 2010). Calculations assume fully liquid stratocumulus with no overlying high clouds, justified by typically warm cloud tops and low occurrence of high clouds ($\leq 15\%$ Muñoz et al., 2016). The effective radius was computed from liquid water content assuming a lognormal distribution of 1.34 and a fixed droplet number concentration of 400 cm^{-3} (Jia et al., 2019; Wood, 2012).

Following data processing, extreme values in process-related variables (w_{sub} , w_{rad} , and w_{flux}) were removed using a 0.1-99.9 percentile filter, and all derived variables were computed from the quality-controlled dataset. The PAD transition was then characterized from complementary temporal and spatial perspectives. These analyses were designed to isolate variability across different scales and link observed cloud behavior with boundary layer structure.

For temporal variability, CCF derived from GOES satellite observations was used as the primary variable to characterize stratocumulus variability. Spectral analysis was performed at the grid-cell level using 751 uniformly distributed points in each zone. This number corresponds to the minimum number of available points among the analyzed regions, ensuring consistent spatial sampling across zones. The data were interpolated onto a uniform temporal grid and then processed using a 3-hour moving average, detrending, and normalization. A continuous wavelet transform (CWT) with a Morse wavelet was applied to quantify variability across temporal scales of 0.25-100 days, capturing sub-daily to sub-seasonal variability. The cone of influence was used to minimize edge effects, and the global wavelet spectrum was computed to identify dominant frequencies. Given the relatively short record length (3 years), the full time series was used instead of seasonal subsets to ensure robust spectral estimates. In addition to spectral variability, the mean diurnal cycle was characterized to capture systematic intra-daily changes in cloud cover and the processes governing it. For this analysis, each variable was spatially averaged over all pixels within predefined zones (Fig. 1), and the mean diurnal cycle was computed using local time.

For spatial variability, seasonal mean values were first computed for each variable. Offshore-inland profiles were then constructed by binning the data according to distance from the coastline and computing mean values within each bin. To reduce small-scale variability and emphasize large-scale spatial structures, a rolling mean was applied to the resulting profiles for both observational (CCF) and driver analyses. For the GOES datasets, 1 km bins and a 25-point rolling window were used, whereas for ERA5, 5 km bins and a 5-point rolling window were applied.

Finally, offshore winter conditions are used as a reference state to interpret the evolution of the STBL, providing a baseline representative of relatively undisturbed marine boundary layer conditions. This reference helps contextualize the alongshore evolution of the system, noting that uncertainties increase inland due to more complex topography and reduced observational constraints.

4 Results

In this section, we analyze the variability of CCF and the key drivers of STBL across the PAD transition. First, we characterize CCF variability at daily and seasonal timescales. Second, we analyze h by comparing estimates derived from different

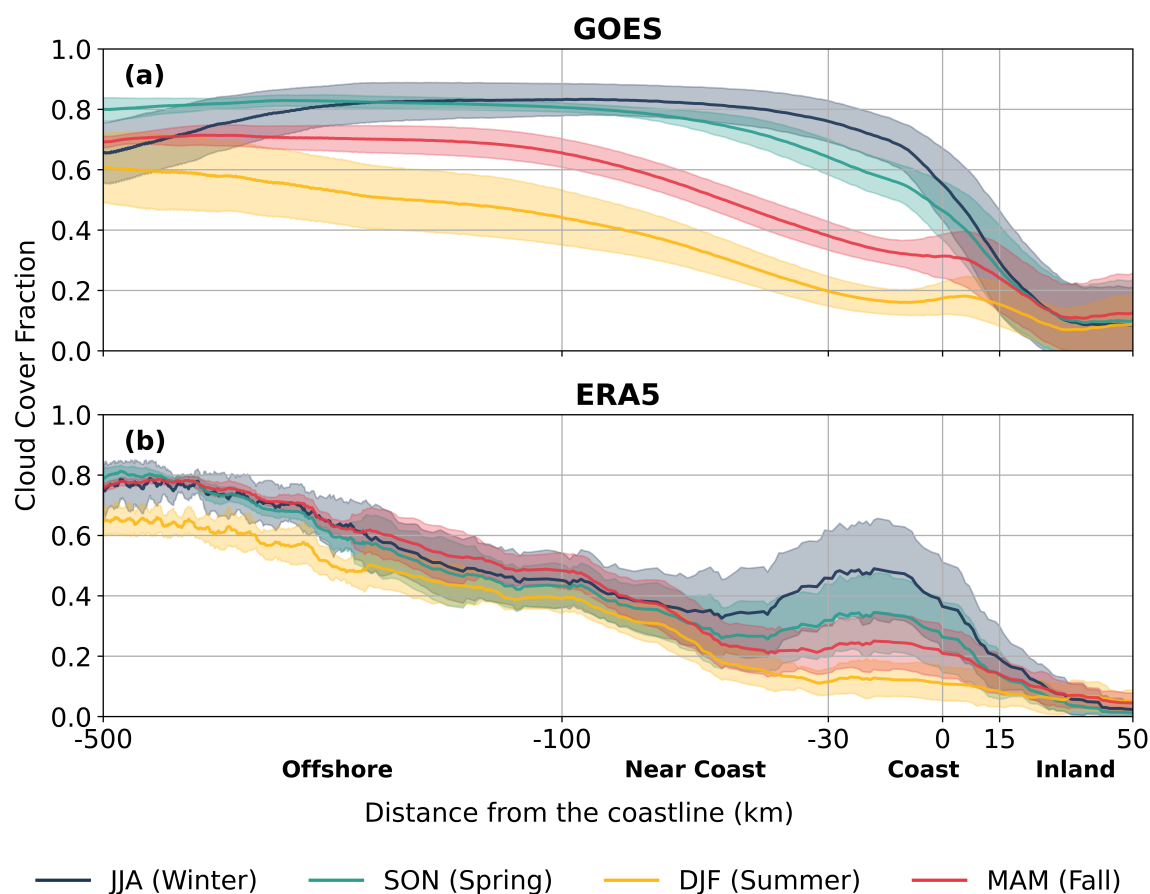


Figure 2. Low cloud cover fraction across the PAD transition from GOES satellite data (a) and ERA5 reanalysis (b). Colored lines show the seasonal mean. Shading indicates the spatial standard deviation across these distance bins for each season.

approaches and available observations. Third, we quantify h tendency budget and its contributions using the mixed-layer
 340 framework in the dry-cloud case to estimate entrainment based on radiative cooling and surface fluxes. In the discussion, we
 first synthesize these results to examine CCF variability and the relative role of the different processes, and then explore the
 wet-cloud case through a sensitivity analysis of moisture's influence on entrainment and its associated uncertainty.

4.1 CCF spatiotemporal patterns

Figure 2 shows the CCF comparison between GOES and ERA5 data across the PAD transition for each season. On average,
 345 CCF decreases from 0.69 offshore to 0.09 inland, consistent with previous studies (del Río et al., 2021; Espinoza et al., 2024),
 with noticeable differences between datasets.

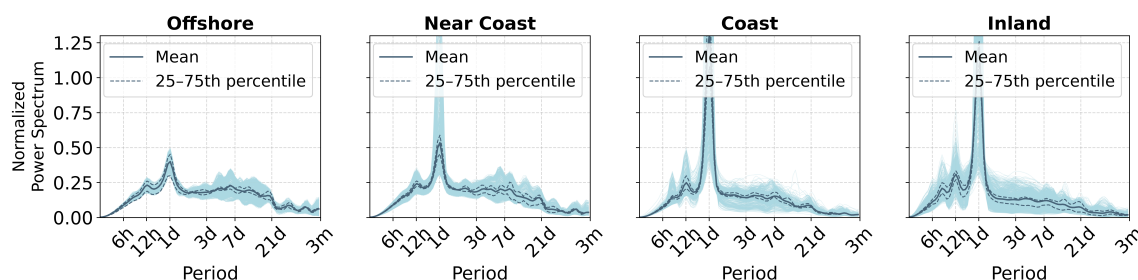


Figure 3. CCF spectral analysis across the PAD transition, scaled to their total variance. Light-blue lines show spectra from individual grid-cells within each zone. The dark-blue line represents the mean power spectrum across all pixels, while the 25th and 75th percentiles indicate mean power spectrum seasonal variability.

GOES shows a clear offshore-to-inland decrease in CCF, with distinct seasonal variations. In austral winter (JJA) and spring (SON), CCF remains relatively high and nearly constant across the offshore and near coast zones (around 0.78) before decreasing toward the coast and inland zones. In summer (DJF) and fall (MAM), CCF decreases gradually from offshore to inland.

350 ERA5, in contrast, shows little seasonal variation and a continuous CCF decrease from 0.68 at 500 km offshore to 0.31 near the coast. A slight increase is observed at the coast, particularly in winter, likely associated with topographic effects that enhance orographic fog formation (Sampurno Bruijnzeel et al., 2006), before CCF decreases again further inland.

Comparing both datasets, ERA5 does not perform well in reproducing the GOES observations. While ERA5 captures the offshore-to-inland decreasing tendency, notable systematic differences in magnitude persist across the different zones. In particular, ERA5 exhibits a consistent bias that varies along the transect, with larger discrepancies occurring in the near coast and coastal zones (bias of -0.29 and -0.13, respectively) compared to offshore and inland areas (bias of -0.03 and -0.06, respectively). Latitudinal variability along the transect is similar in both datasets across all seasons, with standard deviations below 0.15 in all zones. This offshore-to-inland structure in CCF suggests that different physical processes dominate across the transition. We next examine its spectral characteristics to further assess how CCF evolves across the PAD transition.

360 Figure 3 shows the normalized power spectrum of the CCF time series across the PAD transition, revealing a systematic redistribution of CCF variance along the transect. A clear shift in the distribution of spectral power is observed from longer timescales offshore toward progressively shorter timescales in inland environments. Offshore and near coast zones exhibit a broader spectral distribution, between the diurnal (1-day) peak and a wide range of synoptic scales (in order of days to weeks), suggesting the influence of large scale and persistent forcing mechanisms (Bretherton et al., 2010). In contrast, coastal and inland zones become increasingly dominated by higher-frequency variability, with the diurnal cycle (1 day) emerging as the primary contributor. This indicates stronger control by local, rapidly evolving processes in land-influenced environments, consistent with the diurnal forcing associated with sea-land breeze circulations (Schween et al., 2022; Pallauta Pérez et al., 2025).

370 We further examined the diurnal phase of CCF across the PAD transition during winter. As shown in Figure 4, CCF exhibits a clear gradient, with mean values decreasing from offshore to inland, consistent with the spatial pattern in Figure 2. Along this



gradient, the amplitude of the diurnal cycle also increases toward the coast. In the offshore and near coast zones, CCF remains relatively constant throughout the day, with small diurnal variations (0.05 and 0.20, respectively) but high day-to-day variability, broadly similar to the diurnal cycle reported during the EPIC campaign (Bretherton et al., 2004). As one moves toward the coastal zone, this weak cycle intensifies, reaching maximum amplitude (0.47) and developing a well-defined diurnal structure
375 characterized by a nighttime maximum (20:00-08:00 LT), a morning-midday minimum (08:00-16:00 LT), and a secondary increase during the late afternoon (16:00-20:00 LT), consistent with previous studies (Keim-Vera et al., 2024; Schween et al., 2022). Farther inland, the diurnal cycle persists, but its amplitude decreases again (0.21), with reduced variability around midday. This wintertime structure provides a useful reference for assessing seasonal differences in the diurnal cycle.

The diurnal structure is broadly consistent across seasons; however, its magnitude and amplitude vary relative to the winter
380 state. Winter and spring are generally characterized by relatively higher mean CCF values, whereas summer and fall exhibit lower values, in agreement with previous observations (del Río et al., 2021; Keim-Vera et al., 2024; Schween et al., 2022). In addition to these differences, seasonal variability in diurnal amplitude departs from the winter pattern in a spatially confined manner. While offshore and inland regions remain largely similar to winter conditions, most of the seasonal contrasts emerge in the near coast and coast zones. In these regions, diurnal amplitudes are enhanced in winter and spring (0.34 and 0.38,
385 respectively) relative to summer and autumn (0.18 and 0.22, respectively). This enhancement can be associated with seasonal differences in the sea surface temperature gradient, which can lead to a more stable or neutral stratification of the STBL, making it more responsive to diurnal forcing (Schween et al., 2022; Espinoza et al., 2024).

4.2 STBL height tendency

The spatio-temporal patterns of CCF described above are strongly linked to the structure of the STBL. To explore this con-
390 nection, we first evaluated the STBL height (h) using observational and reanalysis datasets. We then compared the observed h tendency ($\partial h / \partial t$) with that inferred from the diagnosed h tendency budget for the dry-cloud case (Eq. 1). This allowed us to assess whether the combined effects of synoptic-scale subsidence, cloud-top radiative cooling and surface fluxes reproduce the observed evolution of h throughout the year.

Figure 5 compares h across the PAD transition for different seasons and datasets. We first evaluate the representation of h
395 during winter, which serves as the reference period due to its well-defined and persistent STBL structure. In winter, h shows a consistent offshore-to-inland structure across all datasets. However, the Richardson number method exhibits a systematic underestimation relative to both the maximum gradient and GOES, which show closer agreement in both magnitude and spatial structure. These differences are particularly pronounced in the near coast and coastal zones, where the Richardson number method yields substantially lower values (e.g., ~ 430 m near coast and ~ 596 m at the coast) compared to the maximum gra-
400 dient (~ 711 m and ~ 841 m) and GOES (~ 843 and ~ 1245 m). Although GOES-derived estimates show overall agreement in magnitude and spatial structure, they are inherently limited by cloud-top height retrieval, which depends on cloud radiative properties and can introduce biases under varying cloud conditions (Heidinger, 2012). In addition, a threshold of 2,000 m was applied to GOES-derived h to filter unrealistically high cloud-top values, which reduces overestimation but further limits the spatial and temporal coverage of the dataset.

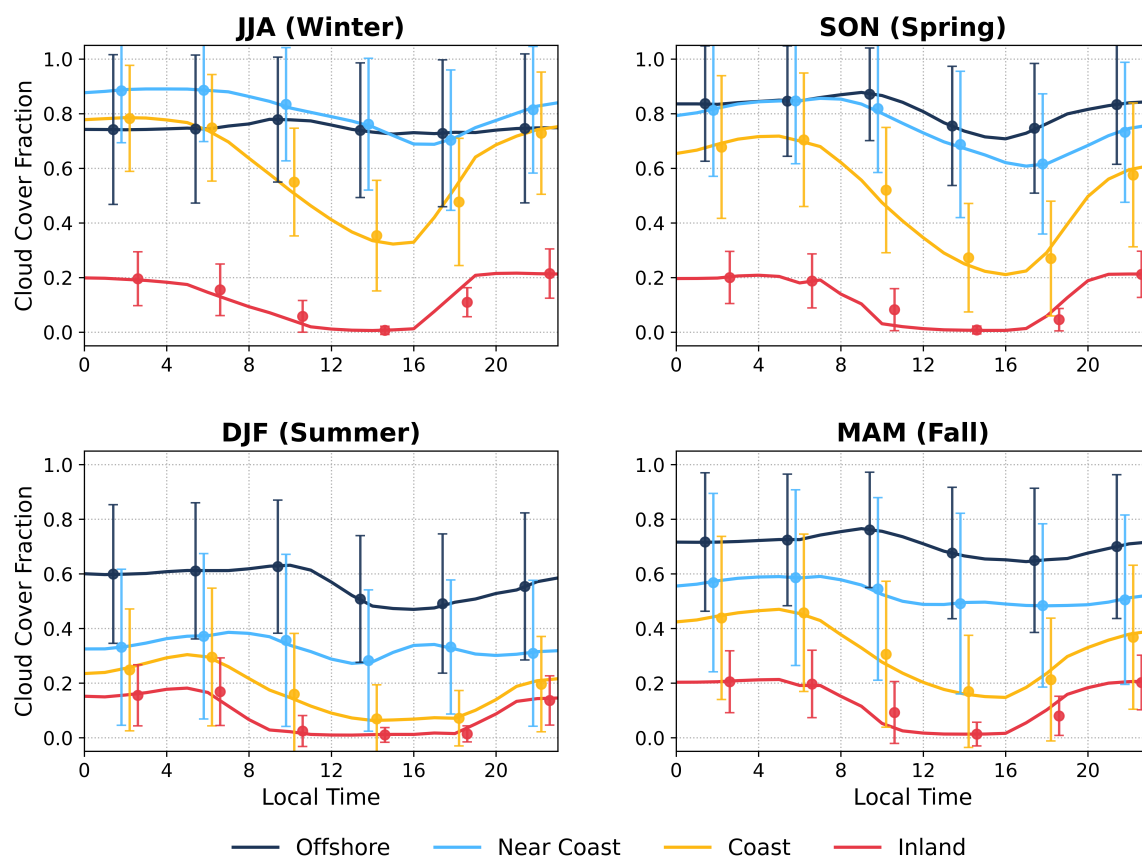


Figure 4. Diurnal cycle of CCF for each season. Lines represent the 3-hour rolling mean of the spatially averaged CCF for each zone, while error bars show the mean $\pm$ standard deviation calculated over all pixels and times within 4-hour groups.

Based on these results and limitations, the maximum gradient method of potential temperature is selected as the reference estimate of h . This choice is further supported by physical considerations, as STBLs are characterized by a sharp potential temperature inversion at the cloud top (Wood, 2012). The maximum gradient method is therefore better suited to capture this inversion structure, which represents a fundamental feature of STBL and controls cloud-top processes. The inter-comparison during winter provides a robust basis for method evaluation. Similar patterns of underestimation and methodological differences are consistently observed across the other seasons, indicating that these biases are not season-specific but rather inherent to the retrieval approaches.

Using the maximum gradient method as the reference estimate, we now focus on the spatio-temporal variability of h across the PAD transition. During winter, h exhibits a coherent offshore-to-inland structure characterized by a decrease from offshore (~ 853 m) toward the coastal zone (~ 841 m), followed by an increase inland (~ 1447 m) associated with rising surface elevation, in agreement with the observations (Fig. 5) and previous studies (Bretherton et al., 2010; del Río et al., 2021; Schween

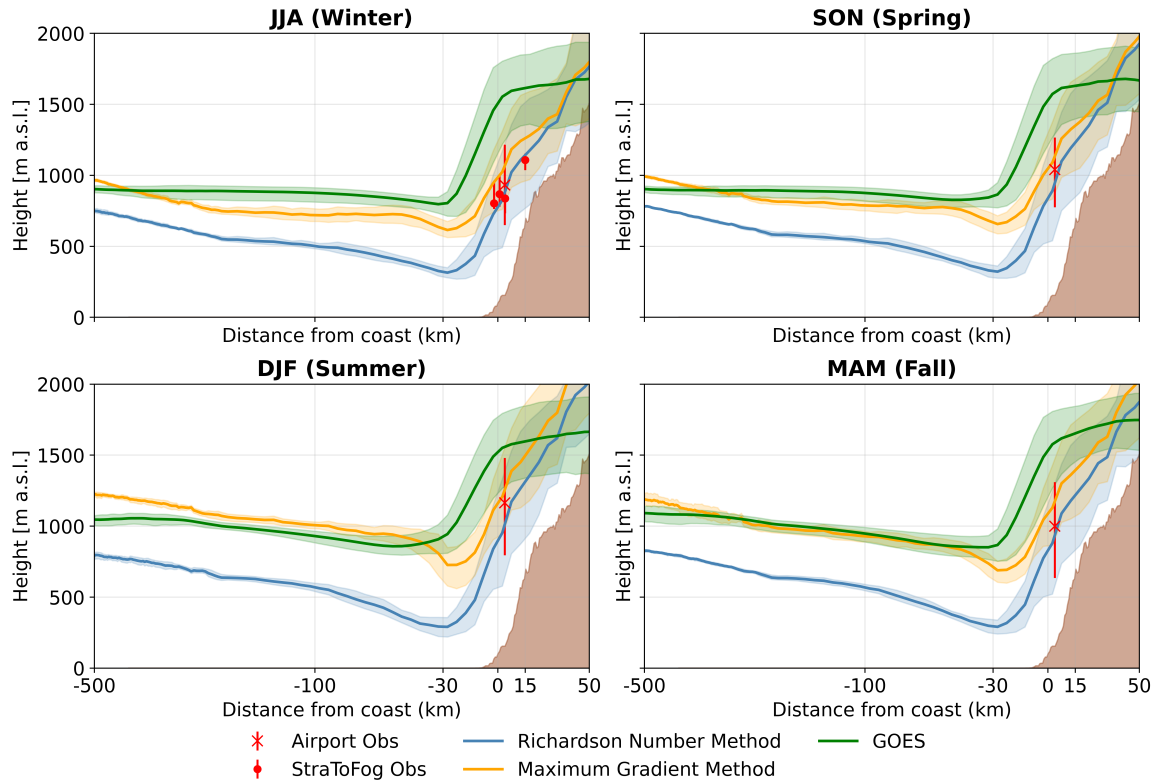


Figure 5. Boundary layer height comparison across the PAD transition for each season. Colored lines indicate the seasonal mean from 2022 to 2024, with shading showing the 10th and 90th percentile across distance bins. Red stars and circles represent observations from the airport radiosondes and the StraToFog field campaign, respectively, and the vertical lines indicate the 10th and 90th percentile of each dataset. CTH GOES is filtered by values below 2,000 m a.s.l. The brown shading is the mean surface elevation based on SRTM digital elevation model.

et al., 2022). Against this winter baseline, seasonal differences become evident across the PAD transition. Spring shows a structure very similar to winter, with mean differences of ~ 73 m across the transition. In contrast, summer and fall exhibit systematically higher values than winter, with mean differences of ~ 256 and ~ 168 m respectively.

We next evaluated whether the dry-cloud h tendency budget reproduces the observed temporal evolution of h . Figure 6 compares the distributions of h tendency ($\partial h / \partial t$) estimated from the budget equation (Eq. 1) with those derived from the temporal evolution of h using the maximum gradient method applied to reanalysis data. Overall, the budget method reproduces the general distribution of $\partial h / \partial t$ across latitudes and distances from the coast, capturing both positive and negative tendencies. The highest density occurs near zero tendency, indicating that small variations in h dominate the temporal variability.

Figure 6 shows that the two estimates are in close agreement offshore and in the near coast zone, but progressively diverge toward the coastal and inland regions. Offshore, the reanalysis tendency is centered near 0 at all latitudes, whereas the budget-based estimate is slightly shifted toward positive values, resulting in a small positive bias (0.49 cm s^{-1}), that indicates a slight

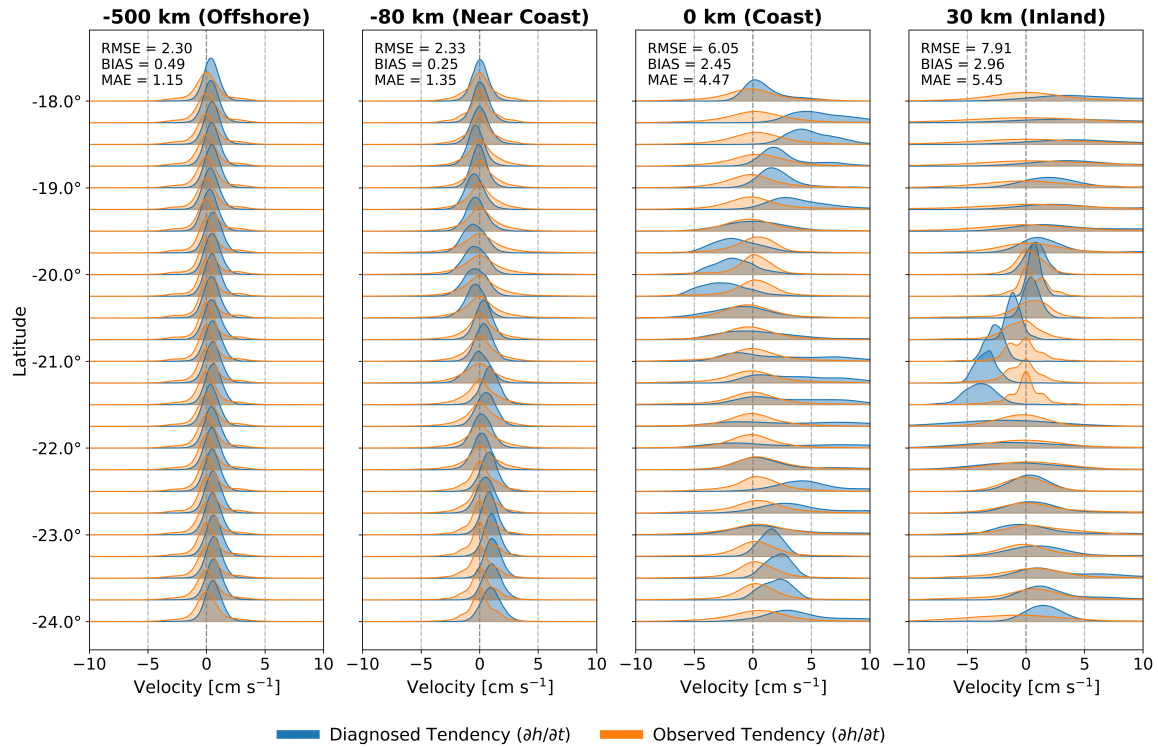


Figure 6. Comparison of the distributions of STBL height tendency at different latitudes and distances from the coast. Colors represent the probability density of the diagnosed tendency (Eq. 1, blue) and the observed tendency estimated using the maximum-gradient method applied to reanalysis data (orange). Each panel corresponds to a specific distance from the coast.

overestimation of the growth of h by the budget. In the near coast zone (-80 km), the reanalysis tendency remains centered near zero but becomes slightly broader, while the budget estimate remains generally close to zero, with a weak latitudinal variability of alternating positive and negative tendencies. In the coastal zone, the two estimates begin to diverge and the errors increase (RMSE = 6.05 and MAE = 4.47); while partial consistency persists between 22° S and 24° S, the budget estimate shows a heterogeneous distribution of positive and negative values across northern latitudes, indicating a growing imbalance of the budget closure of h . Inland, a strong regime shift is observed: distributions widen substantially at the northern (18–20° S) and southern (22–24° S) edges, while in the central latitudes (20–22° S) the consistency between estimates clearly breaks down, with strongly negative tendencies reaching values below (-5 cm s^{-1}).

This progressive disagreement can be interpreted through the definition of h_{adv} in Eq. 1 as a residual term absorbing the mismatch between the resolved budget terms and the reanalysis-derived tendency. Offshore and near the coast, this residual is small and consistent with the advection velocities (-0.5 to 1 cm s^{-1}) reported by Rahn and Garreaud (2010b). Toward the coast and inland, however, the residual grows and becomes more spatially variable, suggesting the increasing influence of local, wind-driven advection not resolved at the synoptic scale. In the coastal zone, this may reflect the "virazon", a diurnal



440 wind system that drives warm air upward along the coastal cliff and locally dissolves the cloud layer (Schween et al., 2020). Further inland, the breakdown at 20–22° S coincides with the Río Loa valley, where nighttime katabatic drainage from the Andean slopes is channeled and intensified by the valley’s orography (Schween et al., 2020). This dry outflow may enhance entrainment at the cloud top, consistent with the positive values needed to compensate the negative diagnosed tendencies. Despite the increasing physical discrepancies described above, RMSE, BIAS, and MAE values (Fig. 6) show that the budget
445 captures the main spatial and temporal variability of the h tendency, allowing the relative contributions of subsidence and entrainment to be compared across the PAD transition, even where local discrepancies remain. This supports the use of the h tendency budget as a framework for attributing the physical drivers of STBL evolution, while also pointing to the coastal and inland zones where the residual term becomes increasingly important.

4.3 STBL height tendency budget

450 To further investigate the mechanisms controlling the STBL height (h), we analyze the tendency budget for the dry-cloud case by decomposing it into its individual processes. This decomposition allows us to quantify the contributions of subsidence, radiative cooling, and surface fluxes to the temporal evolution of h , and to link these tendencies with cloud cover fraction variability across the PAD transition. Figure 7 shows the contribution of each process and the total estimated tendency budget for each season, while Figure 8 summarizes the relative contribution of each process to the total budget. Complementarily,
455 Table 2 provides a summary of the average contributions of each process across different zones.

Overall, the diagnosed h tendency exhibits a quasi-equilibrium over the ocean that progressively weakens toward inland zones, with seasonal differences becoming more evident inland. In winter, subsidence remains negative offshore and near the coast (-0.27 and -0.20, respectively), while radiative cooling (0.64 and 0.49, respectively) and surface fluxes (0.18 and 0.16, respectively) provide positive contributions that remain nearly constant. As a result, the diagnosed tendency stays near zero in both
460 zones (0.55 offshore and 0.45 near coast), with radiative cooling as the dominant process offshore (59% of the total budget; Fig. 8). At the coast, subsidence changes and becomes the dominant term in magnitude (0.98; 53%); its markedly positive values here suggests a local effect (Schween et al., 2020) superimposed on the synoptic-scale subsidence signal. Further inland, subsidence turns negative again, remaining the dominant term in magnitude (-0.36; 44%), consistent with the seasonal position of the subtropical anticyclone (Ancapichún and Garcés-Vargas, 2015). In the coast and inland zones, radiative cooling stays
465 positive but comparatively weak (0.61 and 0.63, respectively), while surface fluxes increase (0.43 and 0.22, respectively). This redistribution produces a coastal maximum in the tendency (2.20) and weak inland values (0.68).

The spatial structure is broadly consistent across the seasons (Fig. 8), with minimal changes offshore and in near coast zones, while the main seasonal variability appears between the coast and inland zones. Subsidence remains the dominant process at the coast throughout the year, although its relative contribution decreases in summer and fall as surface fluxes increase.
470 During summer, surface fluxes increase more strongly in land-influenced zones (gray shaded areas), extending farther inland and becoming the dominant process there (rising from 30% at the coast to 45% inland), consistent with enhanced surface heating during this season. Radiative cooling, in contrast, remains positive, but comparatively weak in coastal and inland zones throughout the year, likely due to the use of cloud-filtered data (cloud cover ≥ 0.2). Together, these seasonal patterns show

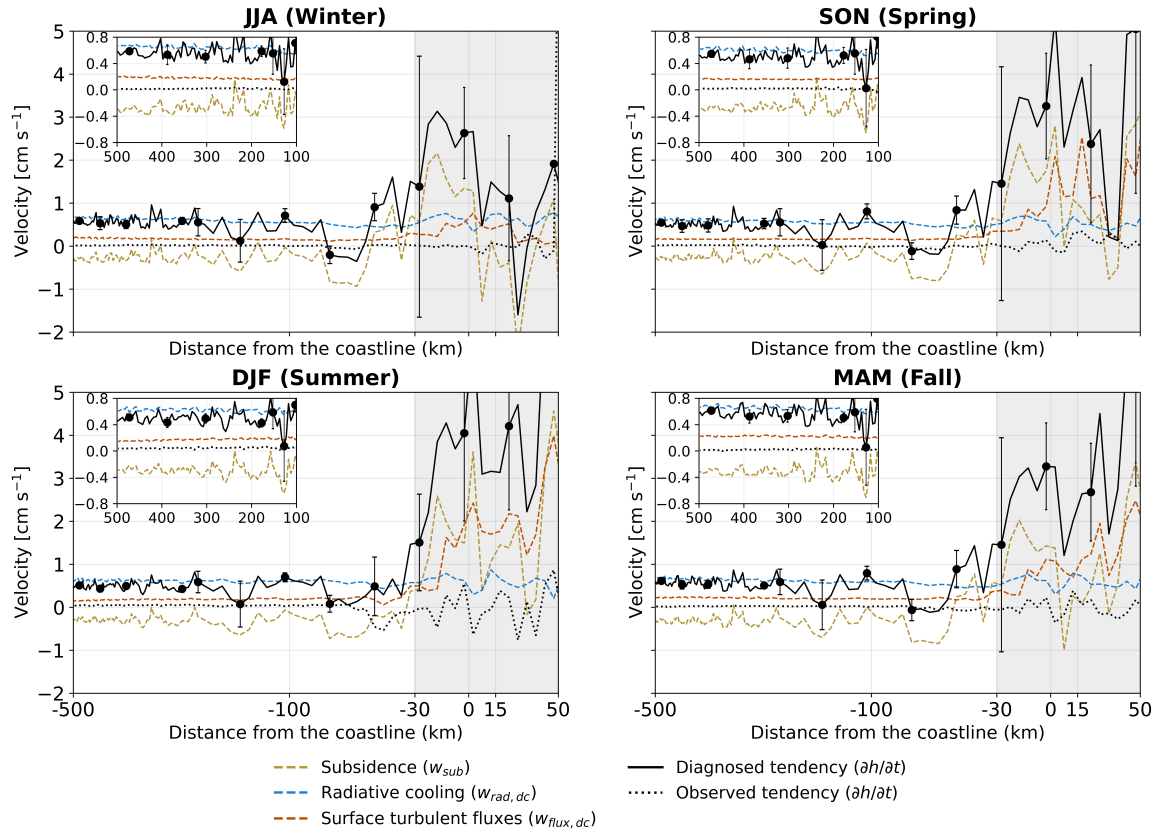


Figure 7. Decomposition of the estimated STBL height (h) tendency budget across the PAD transition for each season. Lines indicate the latitudinally averaged contribution of each process and the h tendencies in distance bins from the coastline. Error bars show the spatial variability of diagnosed tendency. Gray shaded areas mark regions with land influence, based on ERA5 sea-land mask ($> 25\%$ land fraction).

that the inland breakdown of quasi-equilibrium is modulated by both synoptic-scale subsidence and surface fluxes, with their relative importance shifting across seasons.

A mean residual remains in the h tendency budget (Table 2), representing the imbalance between the diagnosed tendency and observed tendency. On average, the residual is small in the offshore and near coast zones (0.50 and 0.49, respectively), but increases sharply at the coast (2.72), and remains large inland (2.70), indicating a systematic budget imbalance across the PAD transition. This imbalance also has a clear seasonal structure (Figure 7): the residual is positive in all seasons, but smaller in winter and larger in summer. Notably, the magnitude of the residual in the offshore zone is consistent with the horizontal advection of h reported by Rahn and Garreaud (2010b), suggesting that horizontal advection is the missing process in the budget offshore. However, the large coastal residual cannot be explained by advection alone, pointing to additional processes or limitations in the calculation of the main terms.



	Offshore	Near Coast	Coast	Inland
Subsidence	-0.29	-0.25	1.27	0.87
Radiative Cooling	0.63	0.52	0.60	0.54
Surface turbulent fluxes	0.18	0.19	0.84	1.34
$\partial h / \partial t_{\text{diagnosed}}$	0.52	0.46	2.71	2.75
$\partial h / \partial t_{\text{observed}}$	0.02	-0.03	-0.01	0.05
Residual	0.50	0.49	2.72	2.70

Table 2. Average contributions of individual processes to the h tendency budget across the PAD transition. Values represent the latitudinally averaged contribution of each process, averaged over all seasons. All values are given in cm s^{-1}

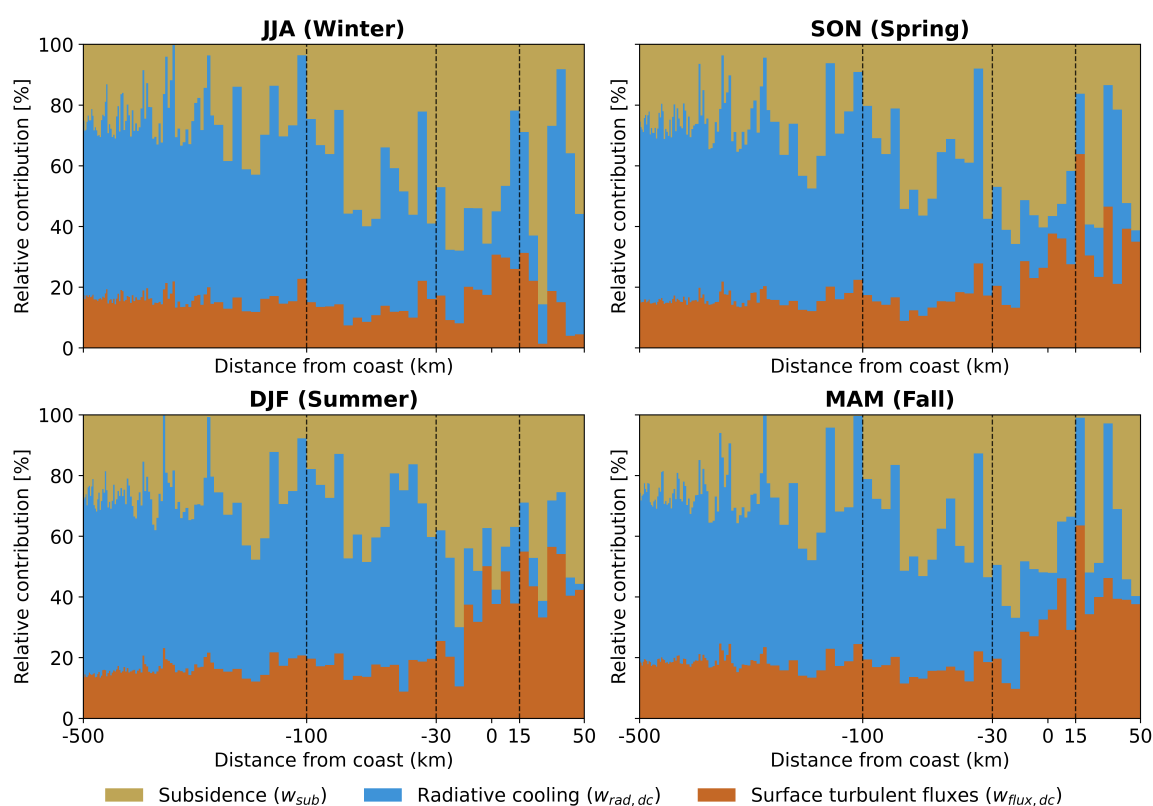


Figure 8. Relative contributions of individual processes to the h tendency budget across the PAD transition. Fractions are calculated from the absolute values of each process to represent their relative magnitude, regardless of sign.

Figure 9 shows the mean diurnal cycle of the h tendency budget across the PAD transition for winter, together with the relative contribution of each process to the total budget. Offshore and near coast zones exhibit a nearly steady diurnal balanced



dominated by radiative cooling (49%), following by subsidence (37%) and surface fluxes (13%): subsidence provides a negative contribution to h throughout the day, while radiative cooling and surface fluxes provide positive contributions, maintaining a quasi-equilibrium state. In the coast zone, this balance is slightly perturbed, resulting in a pronounced diurnal cycle. The diagnosed tendency remains positive, exceeding 4 cm s^{-1} , with minima during night and early morning and maxima during daytime. Subsidence is generally positive in this region, in contrast to offshore conditions, approaching near-zero values in the early morning. The relative contribution of surface fluxes gradually increases during the day, while that of radiative cooling decreases, likely associated with enhanced daytime mixing. In the inland zone, the diurnal cycle is more pronounced, with diagnosed tendencies exceeding 10 cm s^{-1} , negative at early morning and positive during daytime. Subsidence dominates the budget (62% on average) and, together with surface fluxes, follows a similar temporal evolution, strengthening during daytime hours; surface fluxes peak during the day (32%) and diminish at night (2%). Radiative cooling remains the weakest contribution overall although it peaks at night (32%) and diminishes during the day (19%).

The temporal structure is broadly consistent across seasons (details in appendices A2 and A3 for all seasons). From offshore to coastal zones, the diurnal budget shows only minor seasonal differences, with subsidence dominant at night in winter and spring, and radiative cooling remaining secondary, mostly active at night and more pronounced offshore than inland. However, the inland zone exhibits a clear seasonal transition: compared to winter, summer shows higher daytime tendencies, driven mainly by stronger surface fluxes and subsidence, both of which increase specially during daytime in summer and fall. These two processes follow a pronounced diurnal cycle, increasing around 08:00 LT, peaking near 16:00 LT, and decreasing thereafter, consistent with a thermally driven regime. These results demonstrate the diurnal redistribution of processes governing the h tendency across the PAD transition.

5 Discussion

5.1 STBL regions across the PAD transition

Using the spatial framework defined in the study area (Fig. 1), we interpret STBL evolution across the PAD transition through the balance of terms in the boundary layer height tendency equation (Eq. 1). For the dry-cloud case, this balance is governed by the interplay between synoptic-scale subsidence, cloud-top radiative cooling, and surface fluxes, with the latter two jointly driving entrainment of dry and warm air. Its composition shifts systematically across the transect, controlling the spatial variability in cloud cover fraction. Accordingly, we examine each region across the PAD transition.

The first region corresponds to the ocean-dominated environment associated with the offshore and near coast zones of the transect (Fig. 1), where STBL is primarily governed by entrainment driven by cloud-top radiative cooling. In this sector, the STBL remains close to quasi-equilibrium conditions (Fig. 6), characterized by a high and approximately constant boundary layer height h (cloud-top; Fig. 5), similar to the weak h variability reported further offshore during the VOCALS-REx experiment (Bretherton et al., 2010). In that region, the free troposphere was reported to be remarkably dry, which enhanced cloud-top longwave cooling and drove stronger turbulence within the cloud layer, supporting stronger entrainment and a deeper boundary layer offshore. Synoptic-scale subsidence is nearly balanced by entrainment driven by cloud-top radiative cooling and surface

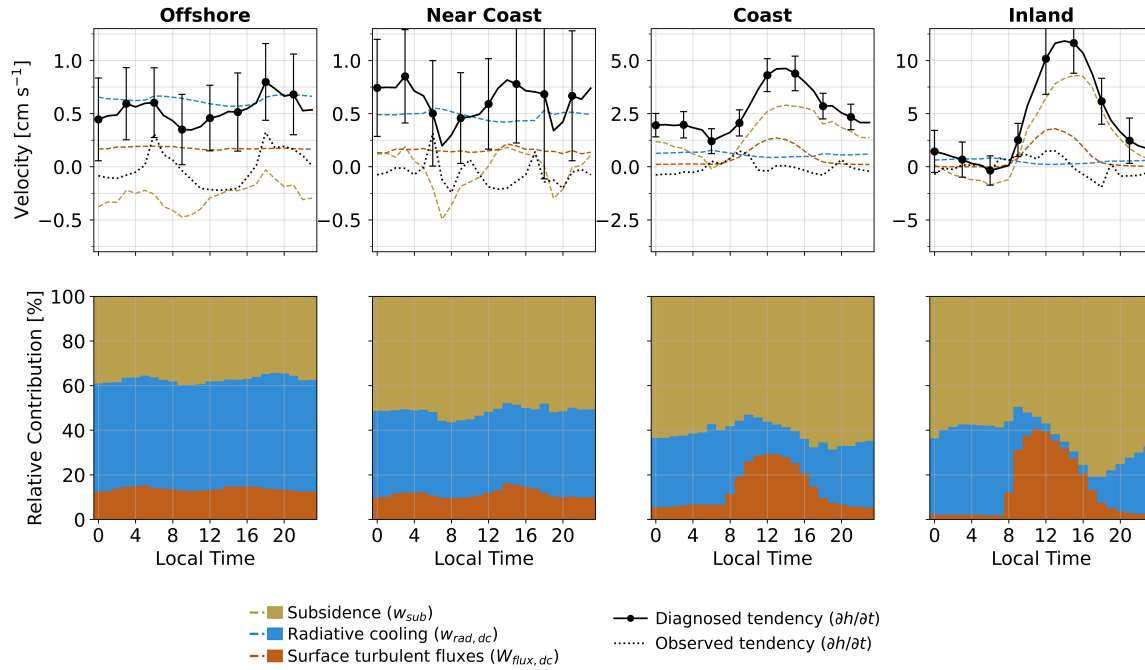


Figure 9. Diurnal cycle of the h tendency budget across the PAD transition during winter. Upper panel: Latitudinally averaged contributions of individual processes and total h tendency; error bars indicate temporal variability in standard deviation. Lower panel: Relative contribution of each process based on absolute tendency values, showing their magnitude regardless of sign.

fluxes, resulting in weak positive tendencies in h (Figs. 7 and 9), with a mean residual of around 0.5 cm s^{-1} , likely associated with horizontal advection contributions similar to those reported by Rahn and Garreaud (2010b). The weak variability of the STBL suggests that synoptic-scale forcing dominates its temporal evolution, consistent with previous descriptions of marine stratocumulus variability further offshore (Bretherton et al., 2010; Vicencio Veloso et al., 2024; Wood, 2012). Accordingly, cloud cover fraction remains persistently high (CCF ~ 0.8 ; Fig. 2), with low-amplitude variability and a broadly distributed spectral power (Figs. 3 and 4), indicating weak diurnal modulation and sustained cloud persistence, characteristic of marine stratocumulus decks (Klein and Hartmann, 1993; Wood, 2012). Together, these characteristics support the interpretation of a radiatively driven STBL in which cloud-top radiative cooling provides the largest contribution to the h tendency balance in this region ($\sim 50\%$).

The second region corresponds to the coastal transition between oceanic and inland areas of the transect, where orographic effects from the coastal mountain range cause the STBL to depart from the quasi-equilibrium conditions observed offshore (Fig. 6). In this region, the boundary layer height increases with the underlying topography (Fig. 5). Unlike the offshore region, orographic lifting of the sea-breeze (virazon wind) circulation against the coastal escarpment modifies the effective contribution of the synoptic-scale vertical motion term, generating locally positive values in the subsidence contribution (Figs. 7-9). Since subsidence is estimated from horizontal wind divergence (Eq. 2), this behavior reflects low-level convergence associated with



the interaction between the virazon circulation and the coastal topography, described by Lobos Roco et al. (2018) as anabatic
winds driven by an afternoon land-side pressure drop and enhanced advection associated with winds from the south-southwest.
This interpretation is supported by QuikSCAT scatterometer observations during VOCALS-REx, which recorded negative
divergence collocated with the positive subsidence values in this region (Rahn and Garreaud, 2010a). ERA5 represents this
terrain-induced convergence and orographic uplift at the large scale (Hersbach et al., 2020; Martinez et al., 2024; Viale et al.,
2013), although its vertical resolution remains too coarse to resolve the associated low-cloud processes (Hannay et al., 2009;
Lin et al., 2014; Neubauer et al., 2014). As a result, synoptic and diurnal variability coexist in this region, producing the largest
variability and strongest cloud cover gradients along the transect (Figs. 2-4), in agreement with previous studies of the coastal
transition zone (del Río et al., 2021; Espinoza et al., 2024; Keim-Vera et al., 2024; Schween et al., 2022). This variability is
also evident in Fig. 1, where the cloud layer intermittently reaches the coastal mountain range, while at other times it remains
detached from it. Together, these results indicate that positive subsidence (wind convergence) provides the largest contribution
to the h tendency balance in this region ($\sim 52\%$).

The third region corresponds to the inland sector of the transect, where the STBL transitions into a thermally driven boundary
layer strongly modulated by the diurnal cycle. In this region, enhanced surface heating promotes turbulent fluxes and mixing,
thereby enhancing entrainment driven by surface fluxes. This term provides the dominant contribution to the h tendency in
summer (43%; Fig. 8), whereas in the remaining seasons subsidence dominates the balance (47%). The relatively shallow
boundary layer enhances coupling between surface forcing and entrainment, leading to pronounced diurnal variability and
coherent 1-day variability in both surface fluxes and subsidence (Fig. 9). This region also exhibits the largest residual in the h
tendency balance along the transect, which may reflect advection of h from which the inland STBL originates. However, since
subsidence retains positive values and a diurnal cycle closely tied to surface fluxes throughout this region, the residual is more
likely dominated by surface influence than by a systematic advective contribution. Unlike the offshore radiatively driven region,
CCF remains low and exhibits strong diurnal modulation (Fig. 4), reflecting reduced cloud persistence and enhanced daytime
cloud dissipation. Despite this transition toward continental forcing, the marine STBL influence persists inland through fog
intrusion extending up to 50 km from the coast (Lobos-Roco et al., 2024). This behavior is consistent with low-cloud regimes
identified in other eastern boundary upwelling systems, including the coastal Namib Desert and California, where enhanced
heating promotes strong diurnal cloud modulation (Dye et al., 2020; Malik et al., 2026; Rojas and Zamora, 2026).
Across these three regions, this characterization is based on the dry-cloud case, as outlined in our research strategy (Sec-
tion 3.2). However, because moisture affects entrainment, its relative importance could vary spatially across the PAD transition.
We therefore next evaluate the contribution of moist processes through a sensitivity analysis of the entrainment formulation.

5.2 Sensitivity of entrainment to moisture effects

Following the entrainment formulations presented in section 2.3, we designed a set of sensitivity experiments to isolate the
contribution of moisture to entrainment velocity (w_{ent} ; Table 3). As described in Section 2.3, the wet-cloud case accounts for
moisture effects through evaporative cooling, which modifies cloud-top radiative cooling and inversion strength. The dry-cloud
case is used as the control configuration, following the formulation in Eq. 4, and is compared against the wet-cloud case to



Experiment	Turbulent flux term	Radiative cooling coefficient (a_c)	Inversion jump term
Dry-cloud case	$\overline{w'\theta'_v}$	1.0	$\Delta\theta_v$
Radiative cooling	$\overline{w'\theta'_v}$	0.5	$\Delta\theta_v$
Inversion strength	$\overline{w'\theta'_v}$	1.0	Δ_2
Wet-cloud case	$\overline{w'\theta'_v}$	0.5	Δ_2

Table 3. Description of the sensitivity experiments. The reference experiment corresponds to the dry-cloud case (in bold), using the zero-order jump model.

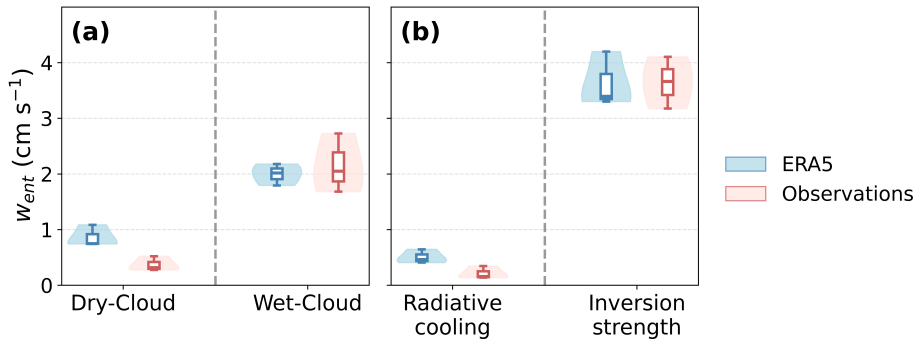


Figure 10. Sensitivity analysis of entrainment velocity using ERA5 and StraToFog data over the coast (20.75° S, 70.25° W). (a) Entrainment velocity for dry-cloud and wet-cloud cases. (b) Sensitivity to the inclusion of moisture in radiative cooling and inversion strength. Box plots show the interquartile range, with violin plots indicating the data distribution. The analysis includes 7 daytime cases ($n = 7$) selected based on observations with presence of clouds ($q_l > 0$), consistent with STBL conditions. ERA5 radiative fluxes are used for both datasets due to the lack of radiative observations during StraToFog.

assess the overall effect of moisture on entrainment. Two additional experiments isolate the specific contribution of each moist term: the radiative cooling experiment isolates the contribution of evaporative cooling (a_c) to cloud-top cooling, while the inversion strength experiment (Δ_2) isolates the effects of moisture and evaporative cooling at the inversion.

Figure 10a shows that w_{ent} increases from the dry-cloud to the wet-cloud case by median factors of 2.40 and 6.71 in ERA5 and observations, respectively. This enhancement is associated with moisture-induced changes in the inversion structure and radiative cooling. Figure 10b separates these effects: the weaker radiative cooling associated with the moist inversion reduces w_{ent} to 0.59 and 0.53 times the dry-cloud value, whereas the reduced inversion jump increases it by factors of 3.83 and 9.07 in ERA5 and observations, respectively. Physically, moisture makes air parcels less dense and more buoyant, allowing them to rise more easily through the weakened inversion and enhancing turbulent mixing across the cloud-top. Thus, the inversion-jump effect dominates, while the weaker radiative cooling partly offsets the enhancement in the wet-cloud case.

Because Δ_2 drives both the largest increase in w_{ent} and the largest discrepancy between ERA5 and observations, we decom-

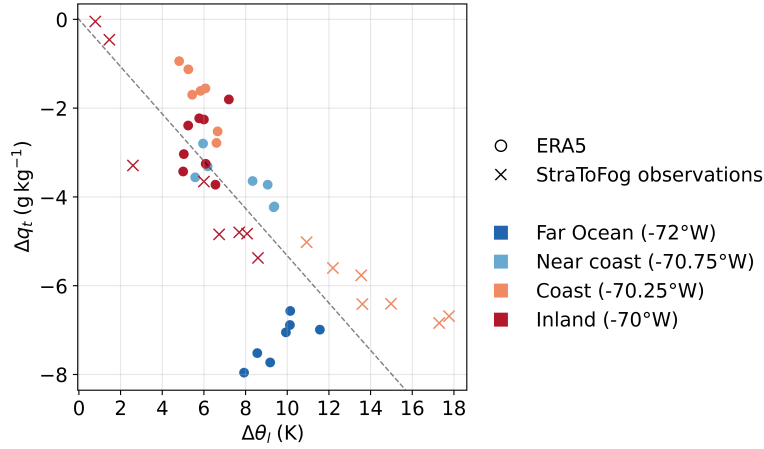


Figure 11. Relationship between $\Delta\theta_l$ and Δq_t along the transect at 20.75° S during daytime. The gray line indicates the buoyancy reversal criterion proposed by Randall (1980) and Deardorff (1980), using $k_{br} = 0.23$. Observations number $n = 7$ in all zones, except inland ($n = 8$)

pose it into its thermal and moisture contributions, $\Delta_2 = a_c \Delta\theta_l + b_c \Delta q_t$. As conserved variables, $\Delta\theta_l$ and Δq_t are commonly
 580 used to characterize mixing at the STBL top (van der Dussen et al., 2014, 2015), with their relative contributions determining whether mixing remains buoyant, favoring cloud persistence, or becomes negatively buoyant, favoring dissipation. The relationship between $\Delta\theta_l$ and Δq_t (Fig. 11; Table 4) reveals a marked contrast between ERA5 and observations. In ERA5, the transition from offshore to coastal conditions is gradual, with points shifting from the left to the right side of the buoyancy-reversal line. In contrast, observations show a sharp coastal transition, consistent with the abrupt reduction in cloud cover
 585 observed in GOES (Fig. 2). Because $a_c \Delta\theta_l$ and $b_c \Delta q_t$ enter Δ_2 with opposite signs, the coastal shift indicates an increasing moisture contribution that offsets the thermal jump and weakens the effective buoyancy jump.

Inland points diverge in two distinct ways from the coastal behavior described above. Most fall on the left side but well removed from the origin, indicating a strong inversion whose partitioning between $\Delta\theta_l$ and Δq_t nonetheless favors thinning (Fig. 11). A smaller subset lies on the right side but close to the origin, reflecting weaker inversions with little resistance to
 590 entrainment. In contrast to ERA5, which retains cloud cover inland, the observations show cloud-free conditions inland. This contrast indicates that the zero-order representation of the inversion alone does not fully explain the observed inland cloud transition, consistent with the broader uncertainty regarding the relative contribution of the processes controlling entrainment (de Lozar and Mellado, 2015).

The zero-order jump model idealizes the inversion as an infinitesimally thin layer with a single thermodynamic jump (Lilly,
 595 1968; van Zanten et al., 1999), which may be too restrictive across the PAD transition, where similar positions relative to the buoyancy-reversal line can represent different inversion structures. A first-order model instead resolves the inversion as a finite-thickness layer and accounts for internal gradients (van Zanten et al., 1999). This distinction also affects radiative cooling: the wet formulation of van Zanten et al. (1999) represents cooling through ΔF_i , which directly cools the inversion layer and enhances w_{ent} without producing TKE, whereas our zero-order model uses $\Delta F_{z,h}$. The more physically consistent repre-



Dataset	Zone	Longitude (W)	$\Delta\theta_l$ (K)	Δq_t (g kg ⁻¹)	Δ_2 (K)
ERA5	Far Ocean	-72.00	10.1 ± 2.2	-5.62 ± 1.23	-0.164 ± 1.141
ERA5	Near coast	-70.75	7.5 ± 1.5	-3.42 ± 0.90	0.568 ± 0.871
ERA5	Coast	-70.25	6.2 ± 1.5	-2.24 ± 0.80	1.029 ± 0.619
ERA5	Inland	-70.00	6.9 ± 1.5	-3.03 ± 1.04	0.647 ± 1.044
Obs	Coast	-70.25	14.3 ± 2.3	-6.11 ± 0.61	1.482 ± 0.685
Obs	Inland	-70.00	5.1 ± 2.8	-3.10 ± 2.03	-0.346 ± 0.831

Table 4. Mean values of $\Delta\theta_l$, Δq_t , and Δ_2 for the selected locations. Values are reported as mean $\pm$ standard deviation. The sample size is $n = 7$ for all zones except inland, where $n = 8$.

600 sentation of radiative cooling in the first-order model may therefore alter the sensitivity of w_{ent} to moisture, as it distinguishes the portion of cooling that directly contributes to entrainment.

These results also extend to coastal fog in the Atacama Desert, which we interpret here as the marine STBL reaching the surface when the advected stratocumulus layer intersects the coastal topography (Keim-Vera et al., 2024). As noted, current models often underestimate cloud cover fraction and produce boundary layers that are too shallow, partly because they do not fully resolve these processes (Hannay et al., 2009; Lin et al., 2014). The moisture dependence of entrainment identified here, particularly the compensating role of $\Delta\theta_l$ and Δq_t in Δ_2 , provide a physical basis for understanding these biases. At the same time, the zero-order jump model used here does not resolve the finite-thickness inversion and the vertical distribution of radiative cooling, which may affect the magnitude of the moisture response. These considerations are particularly relevant for coastal fog, where both the thermodynamic structure of the inversion and entrainment control the persistence of the cloud layer.

6 Conclusions

This study investigates the evolution of cloud cover fraction and cloud-top height, represented by the boundary layer height, across the Pacific-Atacama Desert (PAD) transition. The analysis combines and compares satellite data, in-situ observations from StraToFog campaign, and ERA5 reanalysis to characterize the physical processes governing the stratocumulus-topped boundary layer (STBL). We analyzed in detail how the interplay between synoptic-scale subsidence and cloud-top entrainment, driven by radiative cooling and surface fluxes, controls the transition from an oceanic stratocumulus deck to its breakup over land. This analysis is framed within a mixed-layer approach, using a zero-order model to represent the cloud-top inversion. Cloud cover fraction exhibits strong spatial and temporal variability across the PAD transition. Mean cloud cover fraction decreases from offshore (~ 0.7) to inland (~ 0.1). Offshore and near coast zones are characterized by persistently high cloud cover fraction and weak temporal variability, consistent with a relatively steady STBL state. The coastal zone represents a transition between marine and continental environments and exhibits the largest amplitude in the diurnal cycle, reflecting enhanced cou-



pling with local surface forcing. Inland, cloud cover fraction decreases and variability becomes dominated by a strong diurnal cycle.

The balance between subsidence and entrainment is reflected in the evolution of h , which is a key diagnostic variable integrating the processes governing cloud fraction evolution across the PAD transition. We find that estimating h from the maximum vertical gradient of potential temperature provides robust and physically consistent values across different zones. This gradient-based method performs more consistently than the Richardson number-based approach commonly used in ERA5, which tends to underestimate h , particularly in coastal and topographically complex regions where standard diagnostics may be less reliable. These results indicate the importance of accurately representing boundary layer structure and its role in modulating cloud cover in coastal stratocumulus regions, with implications for the diagnosis and modeling of marine boundary layer clouds.

Analyzing observations and ERA5 reveals a continuous gradient in the STBL across the PAD transition, suggesting a progressive reorganization of the physical controls of the boundary layer. This transition can be interpreted in terms of regions that reflect shifts in the relative balance between synoptic-scale subsidence, cloud-top radiative cooling, and surface fluxes. Within this framework, the STBL evolves from an offshore regime dominated by radiative cooling processes, through a coastal transition zone where topographic effects and thermodynamic forcings interact, and into an inland regime characterized by a thermally driven boundary layer strongly modulated by the diurnal cycle. These changes are consistent with systematic variations in the terms governing h tendency, with near-equilibrium conditions offshore ($\sim 0.5 \text{ cm s}^{-1}$) and a residual increasing to $\sim 2.7 \text{ cm s}^{-1}$ inland, reflecting the increasing influence of advection, the abrupt topographic transition, and the surface fluxes over land. These results extend beyond this regional setting by demonstrating how the relative importance of key STBL processes changes as marine stratocumulus clouds are advected from the open ocean toward an extremely arid continental surface, as occurs in other subtropical coastal desert regions.

Future work should further constrain the processes controlling STBL evolution across the PAD transition through detailed observations and idealized modeling. In particular, campaigns such as StraToFog, which provide observations of inversion structure over the ocean-inland transition, could help assess the relative roles of entrainment and advection. Quantifying subsidence, however, would require a different observational strategy specifically designed to characterize vertical motions and their contribution to boundary-layer evolution. The entrainment sensitivity experiments further show that moisture can strongly modify STBL evolution, although the magnitude of this response may depend on the representation of the inversion. Since this study relies on a zero-order jump model, testing a first-order representation with a finite-thickness inversion enables us to clarify how the inversion structure affects the contribution of cloud-top cooling to entrainment. Overall, our findings demonstrate the importance of multi-scale processes in shaping cloud variability across the Pacific-Atacama Desert transition and provide a physical framework for understanding marine stratocumulus evolution in coastal desert environments.



Appendix A: Supplementary figures

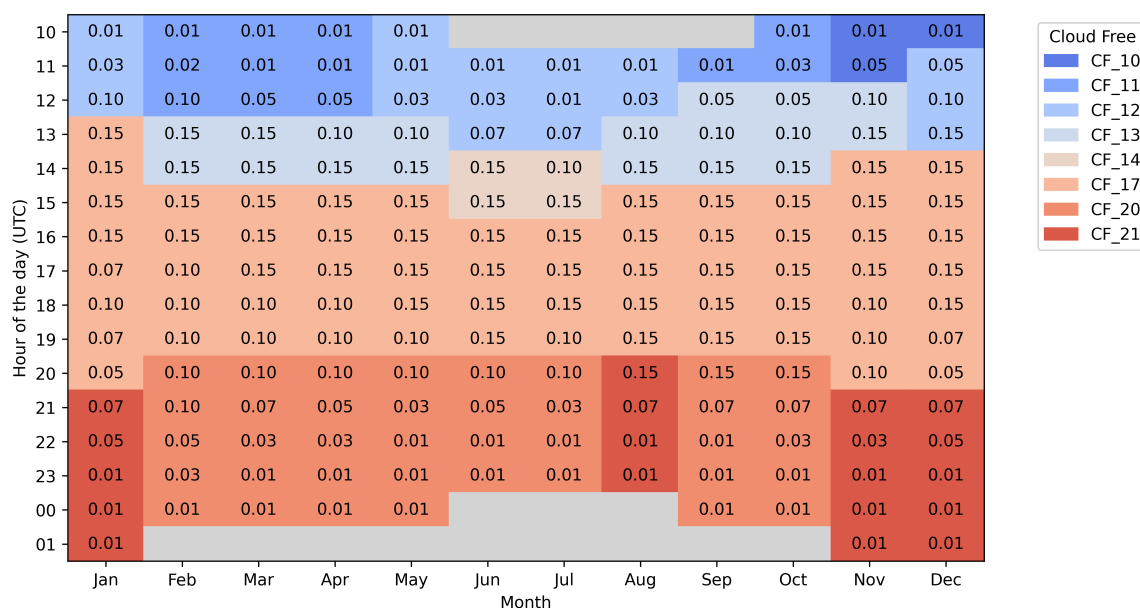


Figure A1. CCF processing details. Daytime classification thresholds shown as numbers, with the applied cloud-free (CF_hh) mosaic represented by color. The hh in CF denotes the hour of images used to generate each mosaic.

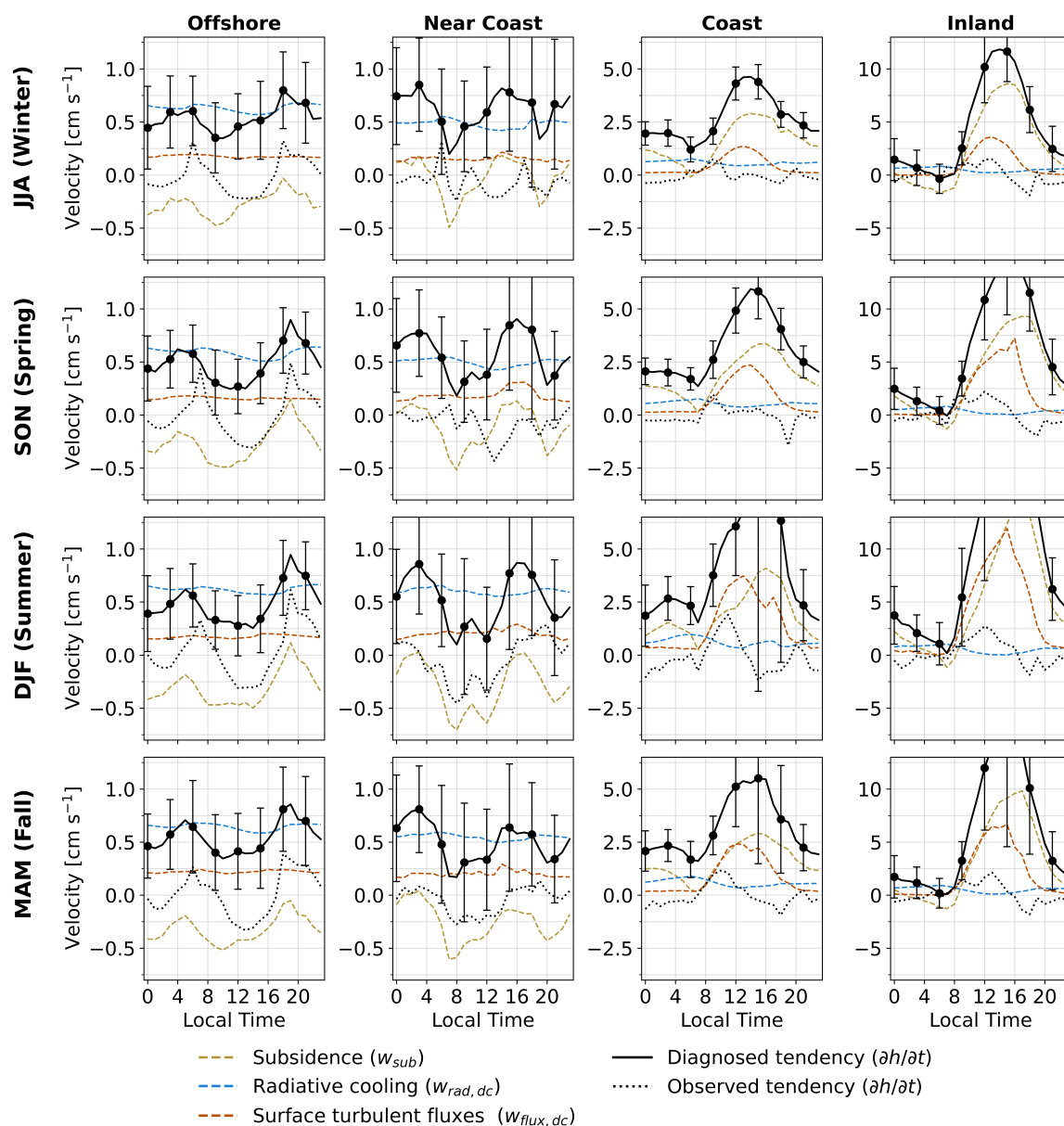


Figure A2. Mean diurnal cycle of the h tendency budget across the PAD transition for each season. Lines indicate the latitudinally averaged contribution of each process and the h tendencies in the different zones. Error bars show the temporal variability.

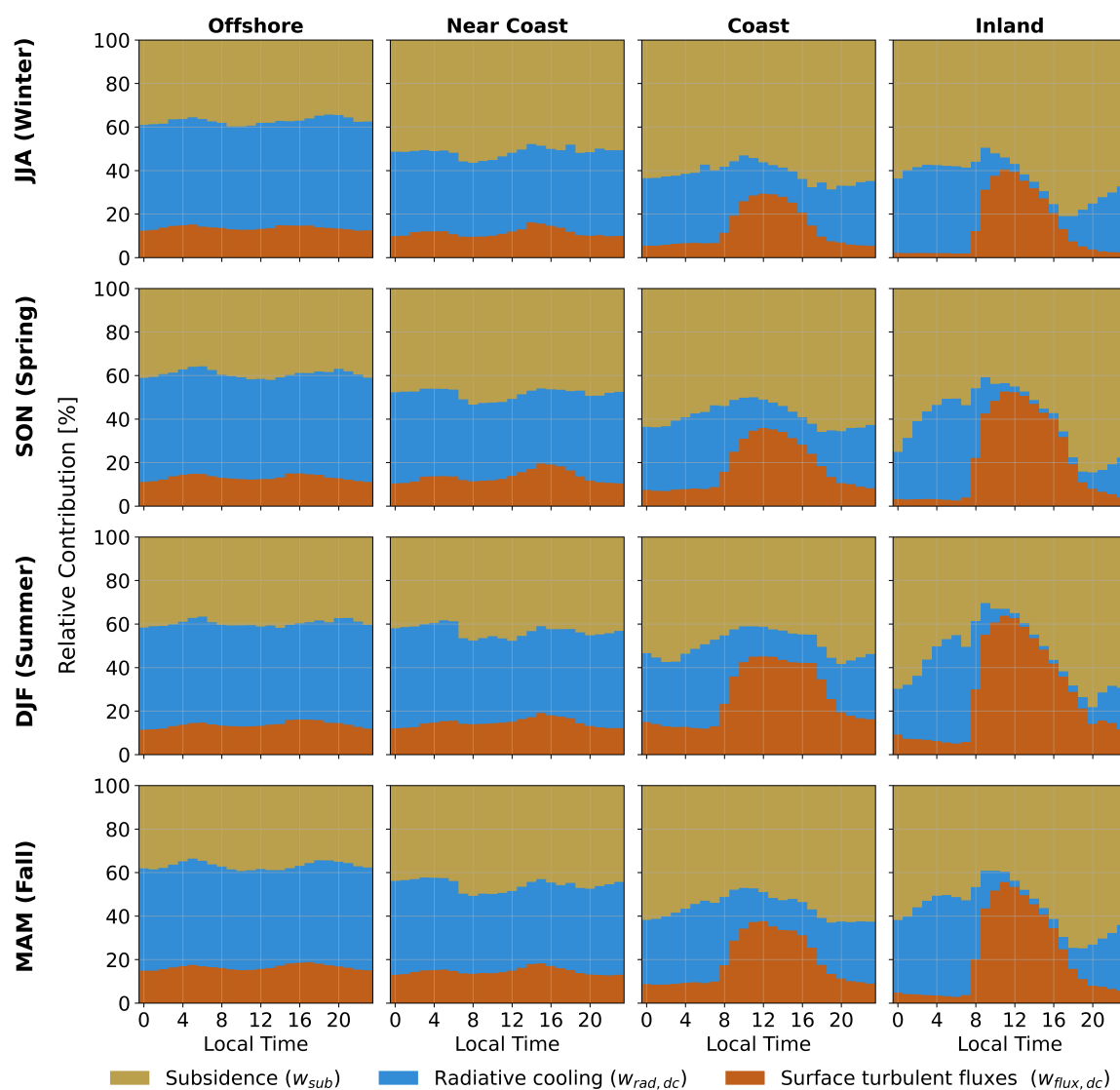


Figure A3. Diurnal cycle of relative contributions of individual processes to the h tendency budget across the PAD transition for each season. Fractions are computed from the absolute values of each process, so that their magnitude is shown regardless of sign.



Data availability. The StraToFog dataset is available from the corresponding author upon reasonable request (Lobos-Roco et al., 2026). Radiosonde observations from Morro Moreno airport were obtained from the University of Wyoming sounding archive.

655 *Author contributions.* V. Espinoza: Writing - review & editing, Writing - original draft, Visualization, Validation, Methodology, Formal analysis, Conceptualization. O. Hartogensis, F. Lobos-Roco and J. Vilà-Guerau de Arellano: Writing - review & editing, Writing - original draft, Supervision, Conceptualization.

Competing interests. The authors declare that they have no conflict of interest.

660 *Acknowledgements.* This research received financial support from the Agencia Nacional de Investigación y Desarrollo (ANID) through the projects ANID/ FONDECYT/11250466, and ANID/FONDECYT/1241176. V. Espinoza also acknowledges financial support from ANID through the PhD scholarship ANID BECAS/ DOCTORADO BECAS CHILE 72240144. Finally, we acknowledge Margreet van Zanten for the valuable discussions.



References

- Aguirre-Correa, F., de Arellano, J. V.-G., Ronda, R., Lobos-Roco, F., Suárez, F., and Hartogensis, O.: Midday Boundary-Layer Col-
lapse in the Altiplano Desert: The Combined Effect of Advection and Subsidence, *Boundary-Layer Meteorology*, 187, 643–671,
665 <https://doi.org/10.1007/s10546-023-00790-5>, 2023.
- Aguirre-Correa, F., de Arellano, J. V.-G., Ronda, R., Lobos-Roco, F., Suárez, F., and Hartogensis, O.: Evaporation Driven by Atmo-
spheric Boundary Layer Processes over a Shallow Saltwater Lagoon in the Altiplano, *Journal of Hydrometeorology*, 25, 1113 – 1134,
<https://doi.org/10.1175/JHM-D-23-0105.1>, 2024.
- 670 Albrecht, B. A., Jensen, M. P., and Syrett, W. J.: Marine boundary layer structure and fractional cloudiness, *Journal of Geophysical Research: Atmospheres*, 100, 14 209–14 222, <https://doi.org/10.1029/95JD00827>, 1995.
- Ancapichún, S. and Garcés-Vargas, J.: Variability of the Southeast Pacific Subtropical Anticyclone and its impact on sea surface temperature off north-central Chile, *Ciencias Marinas*, 41, 1–20, <https://doi.org/10.7773/cm.v41i1.2338>, 2015.
- Bretherton, C. S., Uttal, T., Fairall, C. W., Yuter, S. E., Weller, R. A., Baumgardner, D., Comstock, K., Wood, R., and Raga, G. B.: The Epic
675 2001 Stratocumulus Study, *Bulletin of the American Meteorological Society*, 85, 967 – 978, <https://doi.org/10.1175/BAMS-85-7-967>, 2004.
- Bretherton, C. S., Wood, R., George, R. C., Leon, D., Allen, G., and Zheng, X.: Southeast Pacific stratocumulus clouds, precipitation and boundary layer structure sampled along 20° S during VOCALS-REx, *Atmospheric Chemistry and Physics*, 10, 10 639–10 654, <https://doi.org/10.5194/acp-10-10639-2010>, 2010.
- 680 Cereceda, P., Larrain, H., Osses, P., Farías, M., and Egaña, I.: The spatial and temporal variability of fog and its relation to fog oases in the Atacama Desert, Chile, *Atmospheric Research*, 87, 312–323, <https://doi.org/10.1016/j.atmosres.2007.11.012>, third International Conference on Fog, Fog Collection and Dew, 2008.
- Cermak, J. and Bendix, J.: A novel approach to fog/low stratus detection using Meteosat 8 data, *Atmospheric Research*, 87, 279–292, <https://doi.org/10.1016/j.atmosres.2007.11.009>, third International Conference on Fog, Fog Collection and Dew, 2008.
- 685 Chammas, S., Wang, Q., Schneider, T., Carver, R., Shen, Z., Parker, J. B., Gazen, C., Ihme, M., Chen, Y.-F., and Anderson, J.: High-resolution simulations reveal positive global warming feedback from Pacific low clouds, *Science Advances*, 12, eaec8488, <https://doi.org/10.1126/sciadv.aec8488>, 2026.
- de Lozar, A. and Mellado, J. P.: Evaporative cooling amplification of the entrainment velocity in radiatively driven stratocumulus, *Geophysical Research Letters*, 42, 7223–7229, <https://doi.org/10.1002/2015GL065529>, 2015.
- 690 Deardorff, J. W.: On the entrainment rate of a stratocumulus-topped mixed layer, *Quarterly Journal of the Royal Meteorological Society*, 102, 563–582, <https://doi.org/10.1002/qj.49710243306>, 1976.
- Deardorff, J. W.: Cloud Top Entrainment Instability, *Journal of Atmospheric Sciences*, 37, 131 – 147, [https://doi.org/10.1175/1520-0469\(1980\)037<0131:CTEI>2.0.CO;2](https://doi.org/10.1175/1520-0469(1980)037<0131:CTEI>2.0.CO;2), 1980.
- del Río, C., Lobos, F., Siegmund, A., Tejos, C., Osses, P., Huaman, Z., Meneses, J. P., and García, J.-L.: GOFOS, ground optical fog
695 observation system for monitoring the vertical stratocumulus-fog cloud distribution in the coast of the Atacama Desert, Chile, *Journal of Hydrology*, 597, 126 190, <https://doi.org/10.1016/j.jhydrol.2021.126190>, 2021.
- del Río, C., Lobos-Roco, F., Latorre, C., Koch, M. A., García, J.-L., Osses, P., Lambert, F., Alfaro, F., and Siegmund, A.: Spatial distribution and interannual variability of coastal fog and low clouds cover in the hyperarid Atacama Desert and implications for past and present *Tillandsia landbeckii* ecosystems, *Plant Systematics and Evolution*, 307, 58, <https://doi.org/10.1007/s00606-021-01782-z>, 2021.



- 700 Dirección Meteorológica de Chile [DMC]: Servicios Climáticos, <https://climatologia.meteochile.gob.cl/>, accessed: 2026-03-01, 2026.
- Dye, A. W., Rastogi, B., Clemesha, R. E. S., Kim, J. B., Samelson, R. M., Still, C. J., and Williams, A. P.: Spatial Patterns and Trends of Summertime Low Cloudiness for the Pacific Northwest, 1996–2017, *Geophysical Research Letters*, 47, e2020GL088121, <https://doi.org/10.1029/2020GL088121>, e2020GL088121 2020GL088121, 2020.
- Egli, S., Thies, B., Dröner, J., Cermak, J., and Bendix, J.: A 10 year fog and low stratus climatology for Europe based on Meteosat Second
705 Generation data, *Quarterly Journal of the Royal Meteorological Society*, 143, 530–541, <https://doi.org/10.1002/qj.2941>, 2017.
- Espinoza, V., Lobos-Roco, F., and del Río, C.: Synoptic control of the spatiotemporal variability of fog and low clouds under ENSO phenomena along the Chilean coast (17°–36° S), *Atmospheric Research*, 308, 107533, <https://doi.org/10.1016/j.atmosres.2024.107533>, 2024.
- Falvey, M. and Garreaud, R. D.: Regional cooling in a warming world: Recent temperature trends in the southeast Pacific and along the west coast of subtropical South America (1979–2006), *Journal of Geophysical Research: Atmospheres*, 114,
710 <https://doi.org/10.1029/2008JD010519>, 2009.
- García, J.-L., Lobos-Roco, F., Schween, J. H., del Río, C., Osses, P., Vives, R., Pezoa, M., Siegmund, A., Latorre, C., Alfaro, F., Koch, M. A., and Loehnert, U.: Climate and coastal low-cloud dynamic in the hyperarid Atacama fog Desert and the geographic distribution of *Tillandsia landbeckii* (Bromeliaceae) dune ecosystems, *Plant Systematics and Evolution*, 307, <https://doi.org/10.1007/s00606-021-01775-y>, 2021.
- 715 Ghonima, M. S., Yang, H., Kim, C. K., Heus, T., and Kleissl, J.: Evaluation of WRF SCM Simulations of Stratocumulus-Topped Marine and Coastal Boundary Layers and Improvements to Turbulence and Entrainment Parameterizations, *Journal of Advances in Modeling Earth Systems*, 9, 2635–2653, <https://doi.org/10.1002/2017MS001092>, 2017.
- Gorelick, N., Hancher, M., Dixon, M., Ilyushchenko, S., Thau, D., and Moore, R.: Google Earth Engine: Planetary-scale geospatial analysis for everyone, *Remote Sensing of Environment*, 202, 18–27, <https://doi.org/10.1016/j.rse.2017.06.031>, big Remotely Sensed Data: tools,
720 applications and experiences, 2017.
- Hannay, C., Williamson, D. L., Hack, J. J., Kiehl, J. T., Olson, J. G., Klein, S. A., Bretherton, C. S., and Köhler, M.: Evaluation of Forecasted Southeast Pacific Stratocumulus in the NCAR, GFDL, and ECMWF Models, *Journal of Climate*, 22, 2871–2889, <http://www.jstor.org/stable/26260540>, 2009.
- Heidinger, A.: Algorithm Theoretical Basis Document: ABI Cloud Height, Tech. rep., NOAA NESDIS Center for Satellite Applications and
725 Research, https://www.star.nesdis.noaa.gov/goesr/documents/ATBDs/Baseline/ATBD_GOES-R_Cloud_Height_v3.0_Jul2012.pdf, 2012.
- Hersbach, H., Bell, B., Berrisford, P., Hirahara, S., Horányi, A., Muñoz-Sabater, J., Nicolas, J., Peubey, C., Radu, R., Schepers, D., Simmons, A., Soci, C., Abdalla, S., Abellan, X., Balsamo, G., Bechtold, P., Biavati, G., Bidlot, J., Bonavita, M., De Chiara, G., Dahlgren, P., Dee, D., Diamantakis, M., Dragani, R., Flemming, J., Forbes, R., Fuentes, M., Geer, A., Haimberger, L., Healy, S., Hogan, R. J., Hólm, E., Janisková, M., Keeley, S., Laloyaux, P., Lopez, P., Lupu, C., Radnoti, G., de Rosnay, P., Rozum, I., Vamborg, F., Vil-
730 laume, S., and Thépaut, J.-N.: The ERA5 global reanalysis, *Quarterly Journal of the Royal Meteorological Society*, 146, 1999–2049, <https://doi.org/10.1002/qj.3803>, 2020.
- Hogan, R. J. and Bozzo, A.: A Flexible and Efficient Radiation Scheme for the ECMWF Model, *Journal of Advances in Modeling Earth Systems*, 10, 1990–2008, <https://doi.org/10.1029/2018MS001364>, 2018.
- Jia, H., Ma, X., and Liu, Y.: Exploring aerosol–cloud interaction using VOCALS-REX aircraft measurements, *Atmospheric Chemistry and
735 Physics*, 19, 7955–7971, <https://doi.org/10.5194/acp-19-7955-2019>, 2019.
- Keim-Vera, K., Lobos-Roco, F., Aguirre, I., Merino, C., and del Río, C.: Fog types frequency and their collectable water potential in the Atacama Desert, *Atmospheric Research*, 312, 107747, <https://doi.org/10.1016/j.atmosres.2024.107747>, 2024.



- Klein, S. A. and Hartmann, D. L.: The Seasonal Cycle of Low Stratiform Clouds, *Journal of Climate*, 6, 1587 – 1606, [https://doi.org/10.1175/1520-0442\(1993\)006<1587:TSCOLS>2.0.CO;2](https://doi.org/10.1175/1520-0442(1993)006<1587:TSCOLS>2.0.CO;2), 1993.
- 740 Lee, T. F., Joseph Turk, E., and Richardson, K.: Stratus and fog products using GOES-8-9 3.9- μm data, *Weather and Forecasting*, 12, 664 – 677, [https://doi.org/10.1175/1520-0434\(1997\)012<0664:safpug>2.0.co;2](https://doi.org/10.1175/1520-0434(1997)012<0664:safpug>2.0.co;2), 1997.
- Lilly, D. K.: Models of cloud-topped mixed layers under a strong inversion, *Quarterly Journal of the Royal Meteorological Society*, 94, 292–309, <https://doi.org/10.1002/qj.49709440106>, 1968.
- Lin, J.-L., Qian, T., and Shinoda, T.: Stratocumulus Clouds in Southeastern Pacific Simulated by Eight CMIP5–CFMIP Global Climate
745 Models, *Journal of Climate*, 27, 3000 – 3022, <https://doi.org/10.1175/JCLI-D-13-00376.1>, 2014.
- Lobos Roco, F., Vilà-Guerau de Arellano, J., and Pedruzo-Bagazgoitia, X.: Characterizing the influence of the marine stratocumulus cloud on the land fog at the Atacama Desert, *Atmospheric Research*, 214, 109–120, <https://doi.org/10.1016/j.atmosres.2018.07.009>, 2018.
- Lobos-Roco, F., Hartogensis, O., Vilà-Guerau de Arellano, J., de la Fuente, A., Muñoz, R., Rutllant, J., and Suárez, F.: Local evaporation controlled by regional atmospheric circulation in the Altiplano of the Atacama Desert, *Atmospheric Chemistry and Physics*, 21, 9125–
750 9150, <https://doi.org/10.5194/acp-21-9125-2021>, 2021.
- Lobos-Roco, F., Suárez, F., Aguirre-Correa, F., Keim, K., Aguirre, I., Vargas, C., Abarca, F., Ramírez, C., Escobar, R., Osses, P., and del Río, C.: Understanding inland fog and dew dynamics for assessing potential non-rainfall water use in the Atacama, *Journal of Arid Environments*, 221, 105 125, <https://doi.org/10.1016/j.jaridenv.2024.105125>, 2024.
- Lobos-Roco, F., Espinoza, V., Keim-Vera, K., Muñoz, F., Suárez, F., and Hartogensis, O.: Marine Stratocumulus to Land Fog Transition in
755 the Coastal Mountain-Range of Atacama Desert, *EGUsphere*, 2026, 1–21, <https://doi.org/10.5194/egusphere-2026-2396>, 2026.
- Malik, D., Andersen, H., Cermak, J., Vogt, R., and Adler, B.: Cloud base height determines fog occurrence patterns in the Namib Desert, *Atmospheric Chemistry and Physics*, 26, 681–701, <https://doi.org/10.5194/acp-26-681-2026>, 2026.
- Martinez, J. A., Junquas, C., Bozkurt, D., Viale, M., Fita, L., Trachte, K., Campozano, L., Arias, P. A., Boisier, J. P., Condom, T., Goubanova, K., Pabón-Caicedo, J. D., Poveda, G., Solman, S. A., Sörensson, A. A., and Espinoza, J. C.: Recent progress in at-
760 mospheric modeling over the Andes – part I: review of atmospheric processes, *Frontiers in Earth Science*, Volume 12 - 2024, <https://doi.org/10.3389/feart.2024.1427783>, 2024.
- Minnis, P. and Harrison, E. F.: Diurnal Variability of Regional Cloud and Clear-Sky Radiative Parameters Derived from GOES Data. Part I: Analysis Method, *Journal of Applied Meteorology and Climatology*, 23, 993 – 1011, [https://doi.org/10.1175/1520-0450\(1984\)023<0993:DVORCA>2.0.CO;2](https://doi.org/10.1175/1520-0450(1984)023<0993:DVORCA>2.0.CO;2), 1984.
- 765 Muñoz, R. C., Quintana, J., Falvey, M. J., Rutllant, J. A., and Garreaud, R.: Coastal Clouds at the Eastern Margin of the Southeast Pacific: Climatology and Trends, *Journal of Climate*, 29, 4525 – 4542, <https://doi.org/10.1175/JCLI-D-15-0757.1>, 2016.
- National Climate Prediction Center: Cold & Warm Episodes by Season, https://www.cpc.ncep.noaa.gov/products/analysis_monitoring/ensostuff/ONI_v5.php, accessed: 2026-03-01, 2026.
- Neubauer, D., Lohmann, U., Hoose, C., and Frontoso, M. G.: Impact of the representation of marine stratocumulus clouds on the anthro-
770 pogenic aerosol effect, *Atmospheric Chemistry and Physics*, 14, 11 997–12 022, <https://doi.org/10.5194/acp-14-11997-2014>, 2014.
- Nicholls, S. and Turton, J. D.: An observational study of the structure of stratiform cloud sheets: Part II. Entrainment, *Quarterly Journal of the Royal Meteorological Society*, 112, 461–480, <https://doi.org/10.1002/qj.49711247210>, 1986.
- Pallauta Pérez, G., Muñoz, R. C., Arrué, P., and Zamora Zapata, M.: The Role of Sea Breeze in the Coastal Stratocumulus Dissipation Transition in Antofagasta and San Diego, *Atmosphere*, 16, <https://doi.org/10.3390/atmos16040437>, 2025.



- 775 Park, S. and Leovy, C. B.: Marine Low-Cloud Anomalies Associated with ENSO, *Journal of Climate*, 17, 3448 – 3469, [https://doi.org/10.1175/1520-0442\(2004\)017<3448:MLAAWE>2.0.CO;2](https://doi.org/10.1175/1520-0442(2004)017<3448:MLAAWE>2.0.CO;2), 2004.
- Pedruzo-Bagazgoitia, X., de Roode, S. R., Adler, B., Babić, K., Dione, C., Kalthoff, N., Lohou, F., Lothon, M., and Vilà-Guerau de Arellano, J.: The diurnal stratocumulus-to-cumulus transition over land in southern West Africa, *Atmospheric Chemistry and Physics*, 20, 2735–2754, <https://doi.org/10.5194/acp-20-2735-2020>, 2020.
- 780 Rahn, D. A. and Garreaud, R.: Marine boundary layer over the subtropical southeast Pacific during VOCALS-REx – Part 1: Mean structure and diurnal cycle, *Atmospheric Chemistry and Physics*, 10, 4491–4506, <https://doi.org/10.5194/acp-10-4491-2010>, 2010a.
- Rahn, D. A. and Garreaud, R.: Marine boundary layer over the subtropical southeast Pacific during VOCALS-REx – Part 2: Synoptic variability, *Atmospheric Chemistry and Physics*, 10, 4507–4519, <https://doi.org/10.5194/acp-10-4507-2010>, 2010b.
- Randall, D. A.: Conditional Instability of the First Kind Upside-Down, *Journal of Atmospheric Sciences*, 37, 125 – 130, [https://doi.org/10.1175/1520-0469\(1980\)037<0125:CIOTFK>2.0.CO;2](https://doi.org/10.1175/1520-0469(1980)037<0125:CIOTFK>2.0.CO;2), 1980.
- 785 Rojas, F. and Zamora, M. Z.: Modelling Coastal Stratocumulus Dissipation Under Idealized Variations of Sea Breeze, *Boundary-Layer Meteorology*, 192, 30, <https://doi.org/10.1007/s10546-026-00975-8>, 2026.
- Rozendaal, M. A., Leovy, C. B., and Klein, S. A.: An Observational Study of Diurnal Variations of Marine Stratiform Cloud, *Journal of Climate*, 8, 1795 – 1809, [https://doi.org/10.1175/1520-0442\(1995\)008<1795:AOSODV>2.0.CO;2](https://doi.org/10.1175/1520-0442(1995)008<1795:AOSODV>2.0.CO;2), 1995.
- 790 Rutllant, J. A., Fuenzalida, H., and Aceituno, P.: Climate dynamics along the arid northern coast of Chile: The 1997–1998 Dinámica del Clima de la Región de Antofagasta (DICLIMA) experiment, *Journal of Geophysical Research: Atmospheres*, 108, <https://doi.org/10.1029/2002JD003357>, 2003.
- Rutllant, J. A., Muñoz, R. C., and Garreaud, R. D.: Meteorological observations on the northern Chilean coast during VOCALS-REx, *Atmospheric Chemistry and Physics*, 13, 3409–3422, <https://doi.org/10.5194/acp-13-3409-2013>, 2013.
- 795 Sampurno Bruijnzeel, L., Eugster, W., and Burkard, R.: Fog as a Hydrologic Input, chap. 38, John Wiley & Sons, Ltd, ISBN 9780470848944, <https://doi.org/10.1002/0470848944.hsa041>, 2006.
- Schmit, T. J., Griffith, P., Gunshor, M. M., Daniels, J. M., Goodman, S. J., and Lehair, W. J.: A closer look at the ABI on the GOES-R series, *Bulletin of the American Meteorological Society*, 98, 681–698, 2017.
- Schween, J. H., Hoffmeister, D., and Löhnert, U.: Filling the observational gap in the Atacama Desert with a new network of climate stations, *Global and Planetary Change*, 184, 103 034, <https://doi.org/10.1016/j.gloplacha.2019.103034>, 2020.
- 800 Schween, J. H., del Rio, C., García, J.-L., Osses, P., Westbrook, S., and Löhnert, U.: Life cycle of stratocumulus clouds over 1 year at the coast of the Atacama Desert, *Atmospheric Chemistry and Physics*, 22, 12 241–12 267, <https://doi.org/10.5194/acp-22-12241-2022>, 2022.
- Seidel, D. J., Ao, C. O., and Li, K.: Estimating climatological planetary boundary layer heights from radiosonde observations: Comparison of methods and uncertainty analysis, *Journal of Geophysical Research: Atmospheres*, 115, <https://doi.org/10.1029/2009JD013680>, 2010.
- 805 Serpetzoglou, E., Albrecht, B. A., Kollias, P., and Fairall, C. W.: Boundary Layer, Cloud, and Drizzle Variability in the Southeast Pacific Stratocumulus Regime, *Journal of Climate*, 21, 6191 – 6214, <https://doi.org/10.1175/2008JCLI2186.1>, 2008.
- Shonk, J. K. P., Hogan, R. J., Edwards, J. M., and Mace, G. G.: Effect of improving representation of horizontal and vertical cloud structure on the Earth's global radiation budget. Part I: Review and parametrization, *Quarterly Journal of the Royal Meteorological Society*, 136, 1191–1204, <https://doi.org/10.1002/qj.647>, 2010.
- 810 Stevens, B.: Atmospheric Moist Convection, *Annual Review of Earth and Planetary Sciences*, 33, 605–643, <https://doi.org/10.1146/annurev.earth.33.092203.122658>, 2005.



- Stull, R. B.: An Introduction to Boundary Layer Meteorology, vol. 13 of *Atmospheric Sciences Library*, Kluwer Academic Publishers, Dordrecht, 1988.
- van der Dussen, J. J., de Roode, S. R., and Siebesma, A. P.: Factors Controlling Rapid Stratocumulus Cloud Thinning, *Journal of the Atmospheric Sciences*, 71, 655 – 664, <https://doi.org/10.1175/JAS-D-13-0114.1>, 2014.
- 815 van der Dussen, J. J., de Roode, S. R., Dal Gesso, S., and Siebesma, A. P.: An LES model study of the influence of the free tropospheric thermodynamic conditions on the stratocumulus response to a climate perturbation, *Journal of Advances in Modeling Earth Systems*, 7, 670–691, <https://doi.org/10.1002/2014MS000380>, 2015.
- van Stratum, B. J. H., van Heerwaarden, C. C., and Vilà-Guerau de Arellano, J.: The Benefits and Challenges of Downscaling a Global Reanalysis With Doubly-Periodic Large-Eddy Simulations, *Journal of Advances in Modeling Earth Systems*, 15, e2023MS003750, <https://doi.org/10.1029/2023MS003750>, e2023MS003750 2023MS003750, 2023.
- 820 van Zanten, M. C., Duynkerke, P. G., and Cuijpers, J. W. M.: Entrainment Parameterization in Convective Boundary Layers, *Journal of the Atmospheric Sciences*, 56, 813 – 828, [https://doi.org/10.1175/1520-0469\(1999\)056<0813:EPICBL>2.0.CO;2](https://doi.org/10.1175/1520-0469(1999)056<0813:EPICBL>2.0.CO;2), 1999.
- Viale, M., Houze, R. A., and Rasmussen, K. L.: Upstream Orographic Enhancement of a Narrow Cold-Frontal Rainband Approaching the Andes, *Monthly Weather Review*, 141, 1708 – 1730, <https://doi.org/10.1175/MWR-D-12-00138.1>, 2013.
- 825 Vicencio Veloso, J., Böhm, C., Schween, J. H., Löhnert, U., and Crewell, S.: A comparative study of the atmospheric water vapor in the Atacama and Namib Desert, *Global and Planetary Change*, 232, 104320, <https://doi.org/10.1016/j.gloplacha.2023.104320>, 2024.
- Wood, R.: Stratocumulus Clouds, *Monthly Weather Review*, 140, 2373 – 2423, <https://doi.org/10.1175/MWR-D-11-00121.1>, 2012.
- Wood, R., Mechoso, C. R., Bretherton, C. S., Weller, R. A., Huebert, B., Straneo, F., Albrecht, B. A., Coe, H., Allen, G., Vaughan, G., Daum, P., Fairall, C., Chand, D., Gallardo Klenner, L., Garreaud, R., Grados, C., Covert, D. S., Bates, T. S., Krejci, R., Russell, L. M., de Szoeke, S., Brewer, A., Yuter, S. E., Springston, S. R., Chaigneau, A., Toniazzo, T., Minnis, P., Palikonda, R., Abel, S. J., Brown, W. O. J., Williams, S., Fochesatto, J., Brioude, J., and Bower, K. N.: The VAMOS Ocean-Cloud-Atmosphere-Land Study Regional Experiment (VOCALS-REx): goals, platforms, and field operations, *Atmospheric Chemistry and Physics*, 11, 627–654, <https://doi.org/10.5194/acp-11-627-2011>, 2011.
- 830 Zheng, X., Albrecht, B., Jonsson, H. H., Khelif, D., Feingold, G., Minnis, P., Ayers, K., Chuang, P., Donaher, S., Rossiter, D., Ghate, V., Ruiz-Plancarte, J., and Sun-Mack, S.: Observations of the boundary layer, cloud, and aerosol variability in the southeast Pacific near-coastal marine stratocumulus during VOCALS-REx, *Atmospheric Chemistry and Physics*, 11, 9943–9959, <https://doi.org/10.5194/acp-11-9943-2011>, 2011.