Climate dynamics along the arid northern coast of Chile: The 1997–1998 Dinámica del Clima de la Región de Antofagasta (DICLIMA) experiment

José A. Rutllant, Humberto Fuenzalida, and Patricio Aceituno

Departamento de Geofísica, Universidad de Chile, Santiago, Chile

Received 24 December 2002; revised 25 March 2003; accepted 24 April 2003; published 5 September 2003.

[1] The DICLIMA field experiment was designed to test and quantify the hypothesis of an afternoon enhancement of the coastal subsidence in the extremely arid northern Chile because of solar heating over the west slope of the Andes. Ten-day campaigns near Antofagasta (23°S) were carried out in January 1997, July 1997, and January 1998. Significant diurnal cycles in temperature, mixing ratio, and wind from about 1000 to 4000 m above sea level were observed. This layer was decoupled from the marine boundary layer circulation below by the subsidence inversion when its base was under the average height of the coastal mountain range. The solar heating cycle over the Andes and associated circulation resulted in a mean afternoon zonal divergence above the subsidence inversion base of about 30×10^{-6} s^{-1}, exceeding by a factor of 5 typical subtropical west coast divergences. The corresponding early morning convergence was particularly intense during the austral winter experiment. In spite of the very strong El Niño conditions that prevailed during the July 1997 and January 1998 experiments, the overwhelming control that radiation exerts on the daily cycles of the atmospheric circulation over the west slope of the Andes seems to guarantee the general validity of the results. *INDEX TERMS:* 1620 Global Change: Climate dynamics (3309); 3307 Meteorology and Atmospheric Dynamics: Boundary layer processes; 3322 Meteorology and Atmospheric Dynamics: Land/atmosphere interactions; *KEYWORDS:* coastal subsidence, west coast deserts, Atacama Desert, northern Chile

Citation: Rutllant, J. A., H. Fuenzalida, and P. Aceituno, Climate dynamics along the arid northern coast of Chile: The 1997–1998 Dinámica del Clima de la Región de Antofagasta (DICLIMA) experiment, *J. Geophys. Res.*, *108*(D17), 4538, doi:10.1029/2002JD003357, 2003.

1. Introduction

[2] The climate of the arid west coast of South America, extending from about 30°S to 5°S, results from the interaction of large-scale atmospheric systems in the southeast Pacific with the continental margin. There, the Humboldt Current, coastal upwelling and frequent low-cloud cover keep the adjacent sea surface temperature (SST) well below the corresponding latitudinal means. The resulting land-sea temperature difference and the solar heating cycle over the western slopes of the alongshore-oriented mountain ranges must play an important role in shaping the regional-scale circulation along the dry coast in western South America.

[3] The quasi-permanent influence of the southeast Pacific subtropical anticyclone in northern Chile forces year-round equatorward winds along the coast that produce coastal upwelling of colder water, enhancing the land-sea thermal contrast particularly during sunny afternoons when coastal winds reach their maximum speed. Afternoon winds along the narrow coastal strip peak in austral late spring and summer, in contrast with the offshore southeasterly trades that strengthen in early spring in connection with the maximum central pressure of the subtropical anticyclone [e.g., *Vincent*, 1998]. The resulting change in the alongshore wind stress curl expands and contracts the coastal upwelling strip in spring and summer, respectively.

[4] The stratocumulus cloud deck atop the well-mixed marine boundary layer (MBL) below the subsidence inversion is especially persistent during the austral spring [*Klein and Hartmann*, 1993]. Besides contributing to lower surface temperatures at the coastal ocean and adjacent land, the net radiative cooling at their top drives the buoyant component of the MBL mixing and defines an elevated cold boundary surface that enhances the daytime temperature contrast with the inland (Atacama) desert surface.

[5] The search for physical mechanisms explaining the extremely arid climate along western South America has been the aim of a number of studies. They range from descriptions of the climate factors presumably responsible for that extreme aridity, as given by *Trewartha* [1961], *Cornejo* [1970], and *Prohaska* [1973], to modeling studies, as given by *Luzimar de Abreu and Bannon* [1993]. The required extra subsidence along the coast has been attributed to several posible mechanisms: coastal divergence due to

0148-0227/03/2002JD003357$09.00

differential friction of the equatorward alongshore flow (R. A. Bryson and P. M. Kuhn, cited by *Trewartha* [1961]); the combined regional secondary circulation components of the alongshore (southerly) and inland (northerly) low-level jets [*Lettau*, 1976]; boundary layer divergence due to daytime heating above the western Andean slope [*Rutllant*, 1977, 1990]; eddy mass transfer atop the coastal low-level jet [*Lettau*, 1978]; and the latitudinal change in the Coriolis parameter along the coast [*Luzimar de Abreu and Bannon*, 1993].

[6] Field experiments trying to unveil the regional dynamics of the long arid strip along western South America have been carried out at Pampa de la Joya, Peru [*Lettau*, 1967], and at Antofagasta (23°S), Chile [*Rutllant*, 1977; *Rutllant and Ulriksen*, 1979]. The Pampa de la Joya results stressed the relative importance of thermotidal winds against classical land-sea breezes in producing sand dune migration roughly parallel to the coastline [*Lettau*, 1967]. The Antofagasta Field Experiment, while documenting the existence of a strong diurnal cycle in the vertical wind profiles over the Atacama Desert, revealed the tendency of the coastal subsidence inversion base to attain its minimum altitude and its maximum strength in the afternoon [*Rutllant and Ulriksen*, 1979]. The analysis of the very limited inland vertical wind profiles data during that experiment supported an estimated daytime coastal subsidence of the order of 3–4 cm s^{-1} [*Rutllant*, 1977].

[7] The DICLIMA field experiment was primarily designed to test and quantify the hypothesis of a significant afternoon increase in subsidence along the coast, identifying the relevant mechanisms in the extreme seasons [*Rutllant et al.*, 1998]. The experiment was originally planned in two 10-day field experiments at the seasonal extremes in 1997 (January and July). However, the very strong warm ENSO event that developed since March 1997 forced the extension of the field experiment into January 1998 to allow comparison with the cold-warm transition conditions in January 1997. Therefore measurements in July 1997 cannot be compared with similar observations in the opposite phase of the ENSO cycle.

[8] Section 2 describes the experimental setup, the location of the principal measuring sites, the data obtained, the procedures to enhance diurnal-cycle signals and the method to estimate vertically integrated zonal divergences above the subsidence inversion base. Results presented in section 3 include an assessment of the large and regional-scale meteorological conditions, a description and interpretation of the observed daily cycles over the coast and the associated zonal mass divergences. Results are summarized and discussed in section 4.

2. DICLIMA Experiments: Data and Methods

[9] The three 10-day DICLIMA field experiments were carried out on 20–30 January 1997 (austral summer 1997 (S97)), 14–24 July 1997 (austral winter 1997 (W97)), and 19–28 January 1998 (austral summer 1998 (S98)). The principal observing site was located on the coast at Michilla, 100 km to the north of the city of Antofagasta and 30 km north of the Mejillones peninsula (Figure 1, top panel). The cross-shore terrain profile (Figure 1, bottom panel) shows a narrow coastal plain and a coastal range with altitudes ranging from 1000 to 1500 m above sea level. Eastward, the Atacama Desert extends inland between the coastal range and the Andes. In the high-Andes plateau or Altiplano, at about 4000 m altitude, the Andes peaks reach altitudes well in excess of 5000 m. Baquedano, at 80 km to the northeast of Antofagasta (about 50 km from the coastline) is located over the inland (Atacama) desert at 1020 m altitude (Figure 1).

[10] AIR-GPS rawinsondes were released at Michilla every 3 hours in the last two experiments. In the first survey rawinsondes were launched every 6 hours. A few tethersonde observations were also made on board the R/V *Abate Molina* while participating in a joint oceanographic survey, but difficulties with the balloon during navigation and in the operation of the system in strong winds resulted in poor time and altitude coverage. At Baquedano pressure-temperature-humidity radiosondes were launched at the extremes of the average daily cycle of the wind regime (0800–0900 LT: 8–9 AM and 1700–1800 LT: 5–6 PM). These upper air observations were complemented with pilot balloon wind measurements with an optical theodolite every 3 hours, from 0600 LT until midnight, taking advantage of the permanent clear skies there.

[11] Surface observations with automatic meteorological stations were performed at Baquedano and Michilla to support upper air measurements and to document local diurnal cycles of air temperature, relative humidity, wind and atmospheric pressure. An automatic meteorological station was deployed on a long-term basis at Caleta Constitución (Figure 1, top panel) to monitor coastal upwelling-favorable winds. Another automatic station was located on a 30-m communications tower at Tres Amigas (Figure 1) on top of the coastal range (around 1000 m altitude), close to the mean position of the subsidence inversion base.

[12] To enhance diurnal-cycle signals, the 24-hour-average vertical profiles of temperature, mixing ratio and wind (zonal and meridional components) were subtracted from individual vertical profiles to obtain individual diurnal-cycle anomaly profiles. Diurnal-cycle anomaly profiles corresponding to the same local time were then averaged for each field experiment and also for the three experiments altogether, to obtain mean diurnal-cycle anomalies. Height scales were normalized to a unit subsidence inversion base height (Z = 1) to effectively separate the MBL from the free atmosphere over Michilla. Actual heights above sea level of the 24-hour-average subsidence inversion base ranged from 1000 to 1100 m in January 1997 (S97), from 800 to 900 m in July 1997 (W97), and from 1300 to 1400 m in January 1998 (S98), reflecting the extremes of the annual cycle and the deepening of the marine boundary layer in austral summer during the El Niño event (S98). Therefore the normalized height scale (Z) approximately corresponds to kilometers above sea level.

[13] Data from rawinsondes at Michilla together with standard radiosonde and pilot balloon wind observations at Baquedano were individually validated and interpolated to 50-m intervals. From them, individual wind profiles at 5, 8 and 11 AM for each campaign were averaged to obtain a morning mean AM wind profile. A similar treatment was applied to the 2, 5 and 8 PM data to obtain a mean PM wind

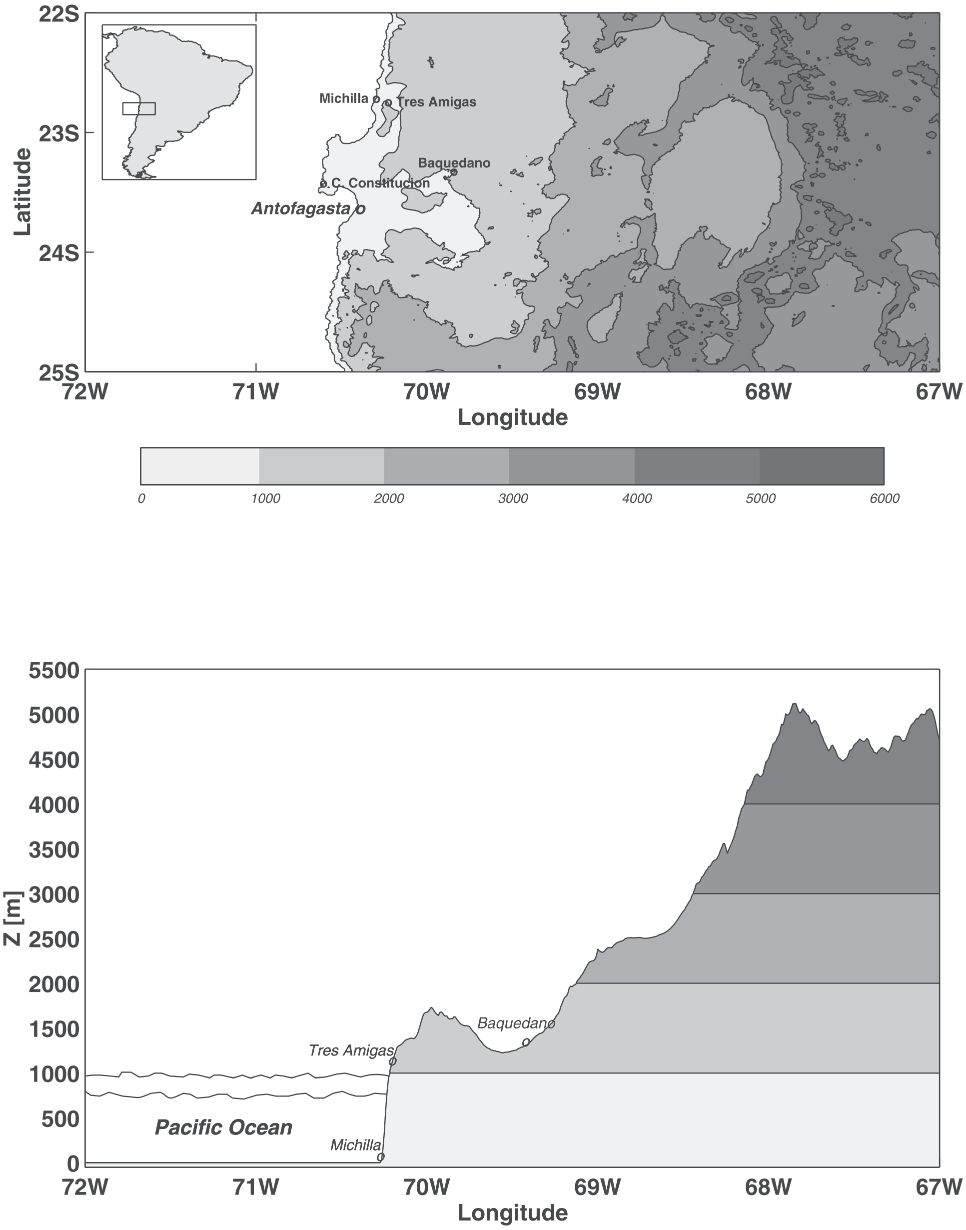

Figure 1. Observation sites and orographic features around the study area. A smoothed west-east terrain cross section is depicted in the bottom panel. Contours every 1000 m are shaded.

profile. The procedure to calculate the mean zonal flux divergence is detailed in Appendix A.

3. Results

3.1. Large and Regional-Scale Meteorological Conditions

[14] The S97 campaign developed during the final stage of the cold phase of the ENSO cycle. Consistent with this large-scale condition, an enhanced subtropical anticyclone and poleward displaced westerlies with associated storm tracks were observed in the SE Pacific [*National Centers for Environmental Prediction*, 1997, 1998]. However, sea level pressure (SLP) at Antofagasta did not significantly depart from normal, though wind speed and low-cloud cover means for January 1997 presented values below the corresponding long-term averages (Universidad Católica del Norte (UCN), Climatological summary, report, Antofagasta, Chile, 1998) (hereinafter referred to as UCN, 1998).

[15] Concurrent anomalous conditions were also observed along the coastal ocean during S97. Satellite imagery and local observations revealed a shallow surface layer with positive SST anomalies of about 1°C on top of the cold waters (negative temperature anomalies) that characterize the cold phase of the ENSO cycle. Those positive SST anomalies appeared early in November 1996 at about 38°S and progressed equatorward along the coast [*Peña et al.*, 1998] constituting a particular regional transition feature to the 1997–1998 El Niño event. On the basis of an outgoing long-wave radiation index (OLRI), defined by *Garreaud and Aceituno* [2001], only moderate precipitation occurred in the Andean highlands (Altiplano) during S97 (OLRI

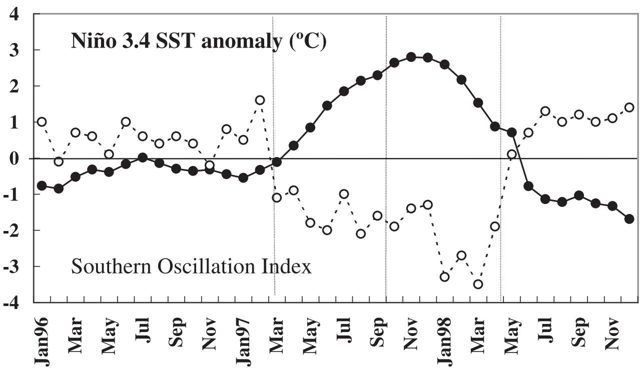

Figure 2. Sea surface temperature anomalies in the Niño 3.4 region and Southern Oscillation Index from December 1996 to February 1998. The observational periods (S97, W97 and S98) are indicated by vertical lines.

above 200 W m^{-2}) with a dry spell (OLRI above 220 W m^{-2}) from January 29 until the end of the experiment. In spite of that, the rainy season in the Altiplano was quite active from mid December 1996 to mid January and again in February 1997.

[16] The strong 1997–1998 El Niño event dominated the tropical Pacific ocean and global atmosphere conditions during W97 and S98, as reflected in the positive sea surface temperature (SST) anomalies at the central equatorial Pacific (Niño 3.4 region) and the negative Southern Oscillation Index (SOI) (Figure 2). The SLP anomaly at Antofagasta was −1.7 hPa in July 1997, while air temperatures stayed 3°C above normal (UCN, 1998), following the positive SST anomalies. Positive wind speed departures of 1 to 2 m s^{-1} were observed in downtown Antofagasta (UCN, 1998) when compared with El Niño Julys of 1987, 1991, 1992, 1993 and 1994, as identified by *Trenberth* [1997].

[17] The average zonal wind profile at Michilla for January 1998 shows westerlies increasing with height in the upper troposphere, reaching more than twice the speed observed in January 1997 at the 200 hPa level. This anomalous strengthening of the upper level westerlies reflects an enhanced subtropical jet centered at about 30°S off the South American west coast [*National Centers for Environmental Prediction*, 1998], a characteristic feature of the mature stage of strong ENSO events [e.g., *Arkin*, 1982]. Owing to strong submonthly-scale variability these enhanced westerlies did not result into suppressed rainfall in the Altiplano during S98, though relatively dry conditions prevailed in December 1997 and February 1998. The average mean temperature of the air column from the surface to 450 hPa over Michilla in January 1998 was 3°C to 4°C higher than in January 1997. Consistently, the subsidence inversion base in 1998 was 300 to 400 m higher than in 1997. The SLP anomalies at Antofagasta attained −3.0 hPa while air temperatures averaged to 3.0 degrees above normal in January 1998 (UCN, 1998). Surface winds for January 1998 were not substantially different from those during El Niño January's of 1987, 1988, 1992 and 1995 [*Trenberth*, 1997], remaining strong and upwelling favorable.

3.2. Diurnal Cycles

[18] Consistent diurnal cycles in air temperature, mixing ratio and winds over a significant depth of the troposphere were found at Michilla in all the field experiments. Mean diurnal-cycle anomalies (see section 2) for temperature, mixing ratio and horizontal vector winds up to about 6 km altitude at Michilla are shown for each campaign and for the whole data set in Figure 3. Besides the similar shape of these diurnal-cycle anomaly patterns, a near-simultaneous phase at all levels is apparent, the extremes occurring around 3 AM (minimum) and 3 PM (maximum) in the air temperature, about noon (minimum) and 9 PM (maximum) in the mixing ratio and around 6 AM and 6 PM in the wind.

[19] During the austral summer experiments (S97 and S98) an amplitude of about 2°C in the diurnal cycle of temperature (Figure 3, left panels) extended roughly between 2 and 4 km altitude. In W97 a stronger nocturnal cooling (longer nights and drier air aloft) produced an equivalent daily swing, with somewhat larger amplitudes at about the same altitude interval (Figure 3, top right). Diurnal cycles in the mixing ratio ranged from 1.5 to 3 g kg^{-1}, with maximum amplitudes within and below the subsidence inversion, except in S98 (Figure 3, bottom left) when large amplitudes are also found approximately between 1.5 and 2 km above the subsidence inversion base (Z between 2.5 and 3). A drying process below the inversion, more pronounced during the summer experiments (Figures 3, left panels), culminated early in the afternoon with a progressive increase in the mixing ratios thereafter. Maximum values tended to concentrate from late evening to early morning. Individual mixing-ratio vertical profiles invariably indicated a well-mixed coastal MBL. The drying of the MBL from dawn to early afternoon will be discussed after due consideration of advections by the mean diurnal-cycle circulation.

[20] Diurnal-cycle anomalies of the horizontal wind vectors are also depicted in Figure 3, where directions are represented in the conventional form and speeds according to the arrow length at the bottom of each panel. Consistent diurnal-cycle wind anomalies are found above the subsidence inversion base (Z = 1) and in the bottom half of the MBL, with SW anomalies in the afternoon and NE ones at dawn. The corresponding return flow of the principal cell, above the subsidence inversion base, is located at about 3 to 4 km height. The MBL cell only presents a distinct return flow in W97 consisting of southeasterlies in the afternoon and northwesterlies in the early morning. This complete circulation cell was possible since the inversion base height (900 m) remained well below the average height of the coastal mountain range (see Figure 8, bottom panel). During that period the daytime phase (sea breeze) was present from 08 AM until midnight, reaching maximum strength around 4 PM. In summer there are long time intervals when the whole MBL moves inland (from 5 PM to 03 AM in January 1997 and from 07 AM to 08 PM in January 1998) with no offshore return flow under the subsidence inversion (see Figure 8, top panel). This can only happen if the incoming flow can surmount the coastal summits, as in January 1998 when the inversion base was at 1300 m altitude on the average. An alternatively explanation for the January 1997 afternoons, when the inversion base was slightly below the top of the coastal summits, is the erosion of the inversion through solar heating (chimney effect), allowing the onshore flow to gain enough buoyancy to overcome the inversion resistance.

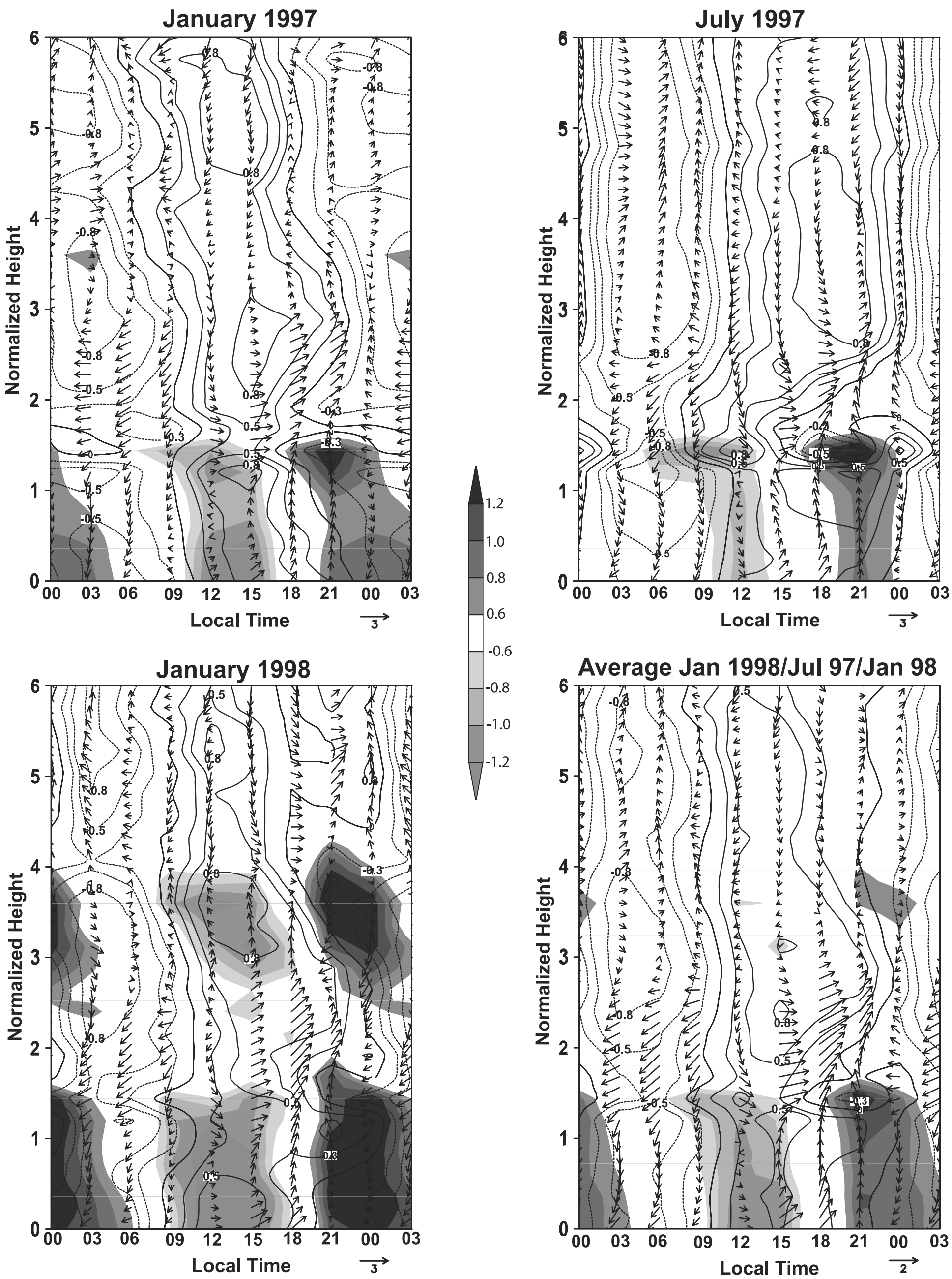

Figure 3. Mean diurnal-cycle anomalies for temperature (°C), mixing ratio (g kg^{-1}) and horizontal vector winds (m s^{-1}) at Michilla up to about 6 km altitude are shown for each campaign: (top left) S97, (top right) W97, (bottom left) S98 and (bottom right) overall mean for the three campaigns. Air temperature anomalies (contours), mixing ratio anomalies (shading) and horizontal vector wind (conventional direction and arrow lengths at the bottom right of each panel) anomalies are represented with a height scale normalized by the inversion base height (Z = 1), corresponding approximately to 1 km above mean sea level.

[21] The principal circulation cell, located between the subsidence inversion base and up to about 3000 to 4000 m altitude, presents an afternoon inland flow that peaks between 3 and 6 PM, roughly from 1500 to 2500 m altitude, with a gradual phase lag with height. At nighttime the direct flow is from the NE. This diurnal cycle in the regional wind direction, deviating from the along-slope direct circulation by the Coriolis effect, was also a significant feature observed at Baquedano during the Antofagasta Field Experiment [*Rutllant and Ulriksen*, 1979].

[22] Statistical significance of these diurnal-cycle anomalies was tested with the local t test for the whole data set (S97 + W97 + S98) separated into a nocturnal subset (3 and 6 AM) and a daytime one (3 and 6 PM) consisting of 37 soundings each. The 95% confidence interval is significantly different from 0 (no diurnal-cycle signal) at all levels in the

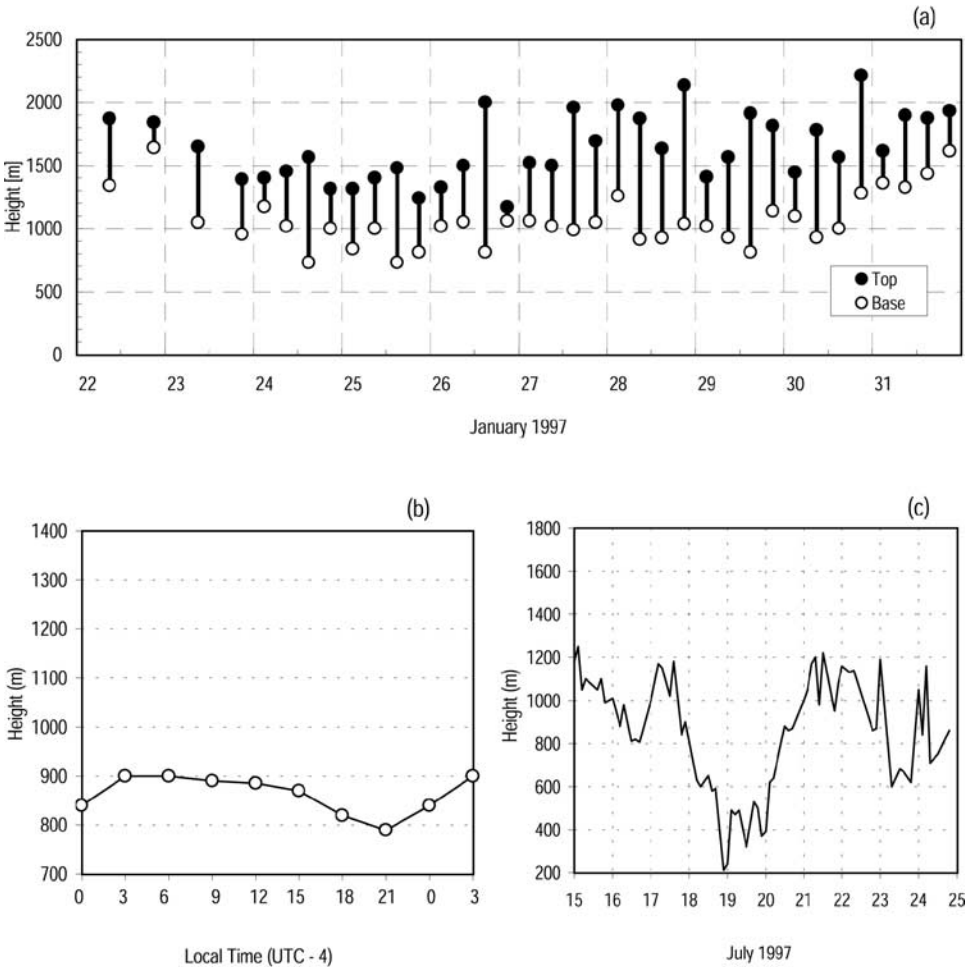

Figure 4. (a) Heights of the subsidence inversion base and top at Michilla during S97 (January 1997). (b) Average diurnal cycle of the subsidence inversion base height during W97 (July 1997). (c) Height of the subsidence inversion base during W97. Figures 4b and 4c are adapted from *Rutllant et al.* [1998].

temperature and the zonal wind, except across the transition through the subsidence inversion (temperature) and the transitions from direct to return flow within the MBL and the principal circulation cells. Mixing ratio diurnal-cycle anomalies are significant within the MBL and between approximately 2.5 to 3 km height (Z = 2.5 to 3), while meridional wind anomalies are only significant within the MBL with a marginal significance at about 3 to 4 km height.

[23] Zonal and meridional components of the vertical wind profiles, averaged for each observing hour (not shown) reveal that prevailing regional wind directions are roughly from the SW and NE, except close to Z = 5 where westerlies begin to dominate (except during S98, in spite of strong average westerlies in the upper troposphere) independently of the sign of the meridional component. Above the principal circulation cell actual winds from the NNW were strong in W97 at about 3500 m (not shown), reflecting an effective blocking of the westerly flow by the Andes (barrier wind effect [e.g., *Parish*, 1982; *Rutllant*, 1983]. Above that level winds veer to westerlies at 4000 m and to south westerlies aloft. In summer winds near the mid troposphere show a NW direction, particularly in S97. The vertical wind shear across the subsidence inversion (not shown) consisted of a SW to NW change from bottom to top, particularly evident in S98 and to a lesser degree in S97. A wind direction shift from SW to SE (bottom to top) across the inversion was observed in W97.

[24] When interpreting the MBL moisture diurnal cycle it is important to recall that the source of moisture below the average altitude of the Andes is the Pacific Ocean. Consequently, easterly (downslope) winds should be in general dryer than their westerly (upslope) counterpart over the coastal free atmosphere below 3500 m. During the first stage of this drying process, easterly advection sets in just below the inversion base (Z = 1) progressing downward while merging with the MBL return flow in the afternoon, particularly in the W97 and S97 campaigns. Dry air entrainment over the coastal plain could happen later in connection with the onset and early development of the sea breeze circulation, helping in the dissipation of the stratus deck together with the easterly advection at the cloud level. As the day progresses, a strong southerly flow component

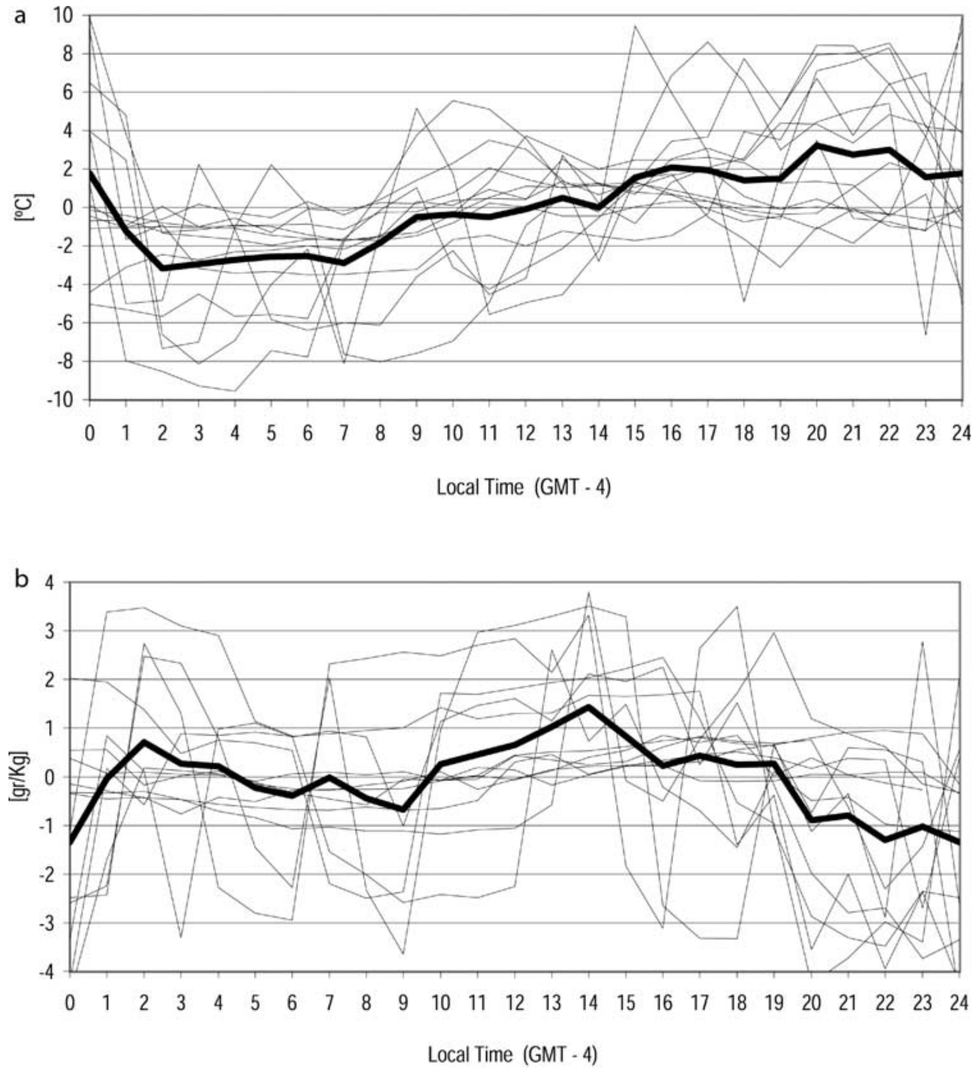

Figure 5. Individual and average diurnal-cycle anomalies of (a) air temperature and (b) mixing ratio at Tres Amigas during W97 (July 1997).

takes over in the afternoon favoring evaporation in the adjacent ocean and inland moisture advection, particularly in summer when the mean zonal flow within the MBL is onshore. Such moistening process is consistent with small cumulus clouds that usually appear against the coastal range late in the afternoon.

[25] Another early afternoon minimum in the mixing ratio above the inversion and up to 3500 to 4000 m altitude was observed in all the experiments, in particular in S98 (Figure 3, bottom left). A plausible explanation here refers to dry-air advection from the SE at those heights corresponding to the average position of the return branch of the daytime surface inland flow over the sloping desert. Subsequent vertical advection near the coast generated by the regional-scale divergence of the zonal flow over the sloping terrain (see next section) can explain the drying process within and above the subsidence inversion. In any case, this particular behavior in the mean daily cycle of the mixing ratio within the coastal MBL and above the subsidence inversion base deserves further study, particularly the "entrainment-drying" process [*Mahrt*, 1991] of the morning MBL, and the role of low-cloud overcast in the moisture budgets below and above the subsidence inversion base.

[26] The average diurnal cycle in the height of the coastal subsidence inversion base shows a consistent trend toward lower heights in the afternoon during all the field experiments. In S97 (Figure 4a), when radiosondes were released every 6 hours, the minimum height of the inversion occurs at 2 PM. As the subsidence inversion base descends the inversion top goes up, in agreement with the 3 PM maximum diurnal-cycle temperature anomaly in the layer between 2 and 4 km (Figure 3). For the W97 experiment, Figure 4b illustrates again this tendency with a 120 m difference between the 5 AM maximum and the 9 PM minimum. This diurnal cycle in the subsidence inversion base is consistent with results reported by *Rutllant and Ulriksen* [1979] and with those from the FIRE experiment at the stratocumulus-topped MBL off the California coast [*Betts*, 1990]. Nevertheless, individual daily cycles show a

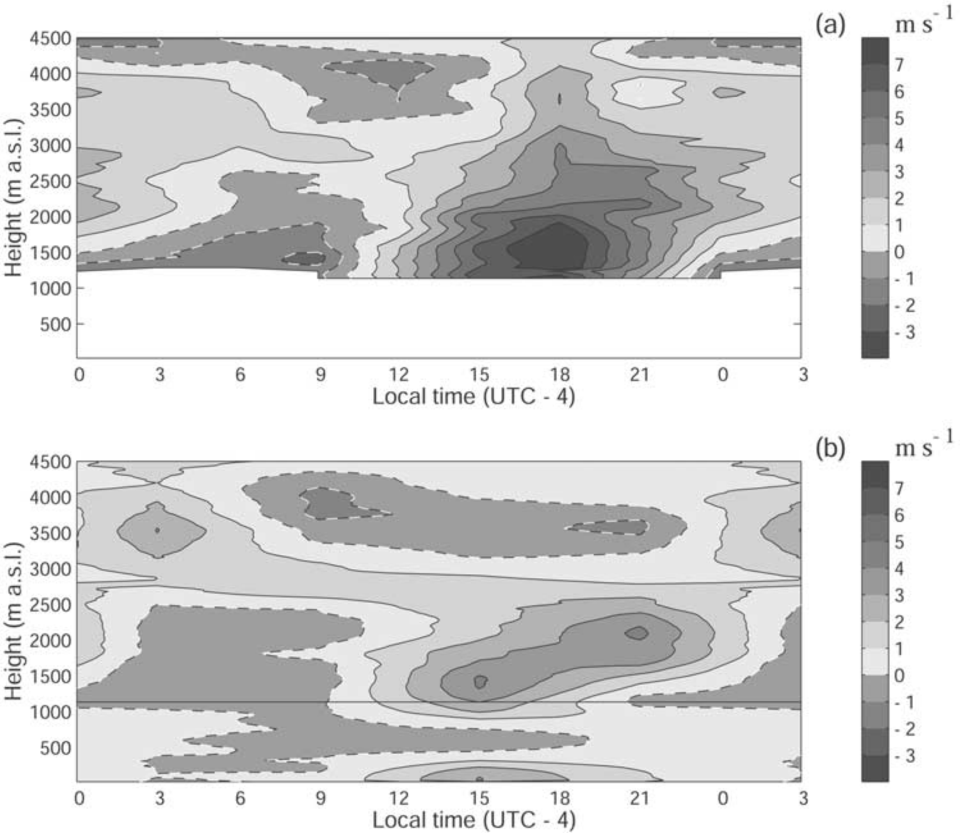

Figure 6. Time-height cross sections of the average diurnal cycles of the zonal wind component at (a) Baquedano and (b) Michilla up to 4500 m above sea level for S97 (January 1997).

large dispersion due to occasional synoptic and subsynoptic-scale disturbances, as evidenced in the day-to-day changes of the subsidence inversion base height at Michilla in W97 (Figure 4c), and in the individual diurnal-cycle anomalies of the air temperature and mixing ratio at Tres Amigas for July 1997 (Figure 5). There, away from surface influences and close to the subsidence inversion base, both the increase in temperature and decrease in the mixing ratio from 2 PM to 8–10 PM are consistent with the descent of the inversion base illustrated in Figure 4b.

3.3. Average Zonal Wind Divergences

[27] Figures 6 and 7 represent time-height cross sections of the mean diurnal cycles of the average zonal wind components at the coastal site (Michilla, sea level: bottom panels) and at the 50 km inland site (Baquedano, 1000 m altitude: top panels) during S97 and W97, respectively. The amplitude of the diurnal cycle at Baquedano is significantly larger than the Michilla one, with stronger PM westerlies in S97 (Figure 6) and stronger AM easterlies in W97 (Figure 7), anticipating AM convergence and PM divergence over the 50-km coastal strip roughly between 1000 and 2500 m altitude. The increasing phase lag with height of the PM component of these cycles could be reflecting the gradual delay in the response to the solar heating as the Andes slope becomes steeper with inland distance (Figure 1).

[28] Assuming alongshore (meridional) homogeneity and a negligible net vertical mass exchange across the inversion layer, the extreme AM/PM convergences/divergences above the subsidence inversion base between Baquedano and Michilla are calculated. For this purpose the layers encompassing the low-level direct flow and the associated upper level return flow at Michilla (Table 1) are considered separately. Although the cores of the PM inland flow approximately coincide at both locations, the stronger Baquedano one in the summer (S97) afternoon (Figure 6, top panel) penetrates deeper into the layer where easterlies (return flow) concentrate over Michilla (Figure 6, bottom panel), enhancing the mean zonal PM divergence below 4000 m height. The zonal wind divergence within 500-m-depth layers centered on the direct and the return zonal flow cores above the subsidence inversion base (Table 1) is calculated according to the scheme described in Appendix A and is presented in Table 2. There, results from the S97 and S98 campaigns have been merged into a single column as "Summer," while those from the W97 campaign have been named as "Winter."

[29] The strongest divergence is found in Summer PM in both the direct (inland, westerly low-level) and the return (seaward, easterly high-level) flows. As mentioned before, the 3000- to 3500-m layer corresponding to the return flow at Michilla presents average easterlies of around 2 m s^{-1} at Michilla and mostly westerlies of the same average speed at Baquedano. In W97 high variability in the return zonal flow during the 10-day campaign (position of the central axis and sign changes), also reflected in the meridional wind component, was due to enhanced synoptic-scale variability in the mid troposphere westerlies associated with the El Niño.

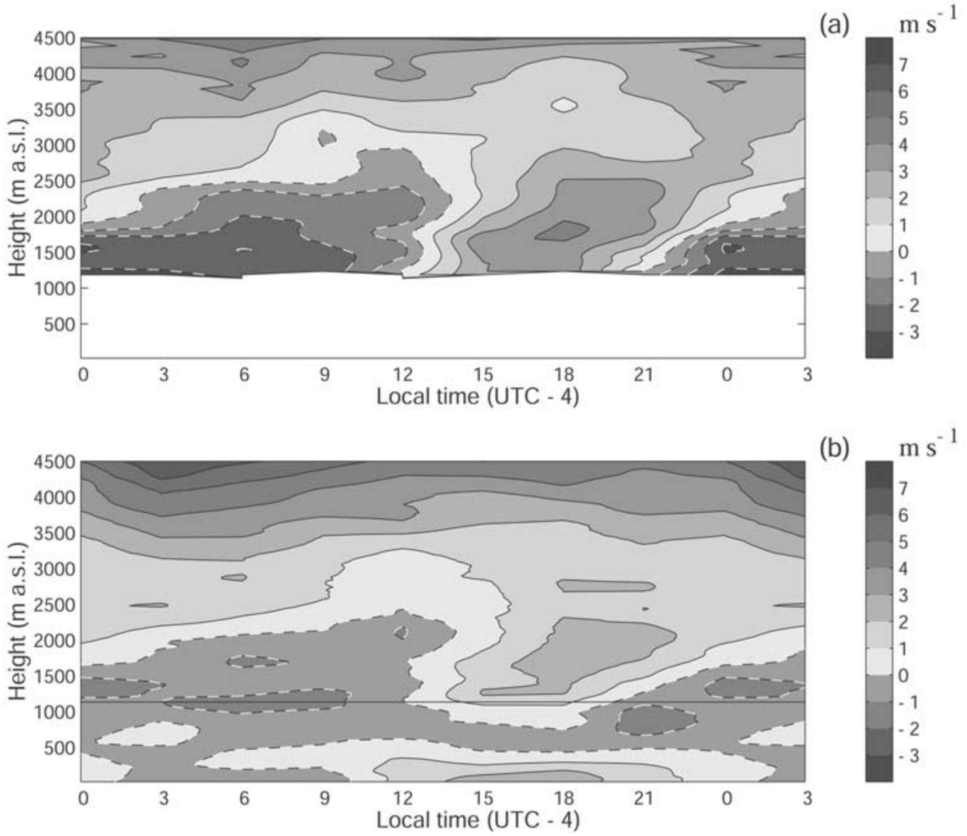

Figure 7. Same as 6 but for W97 (July 1997).

Zonal mean convergence in the AM easterly low-level flow was stronger in W97 due to enhanced katabatic flow down the west Andean slope produced by nocturnal radiative cooling. As a reference for the calculated PM divergences, typical values of 5 to 6 (10^{-6} s^{-1}) are often associated with the subtropical anticyclone just off the California coast [e.g., *Hastenrath*, 1985]. In the afternoon MBL, meridional (alongshore) wind components gradually dominate over the zonal ones. However, except for S98 when the subsidence inversion base was above the coastal escarpment (Figure 8, top panel), the cores of the onshore and offshore flows were centered at about 150 and 600 m, respectively (Table 1), yielding a quasi-balanced net mass flow across the coastline (Figure 8, bottom panel), as expected from theoretical considerations [*Lettau*, 1978]. Furthermore, the summer afternoon onshore flow against the coastal escarpment would contribute with convergence to the mean zonal divergence over the 50 km coastal strip, as discussed in section 4. Therefore the mean zonal divergence cycle within the coastal MBL does not significantly contribute to the daytime mass budget across the coastline.

[30] It may be concluded that the diurnal cycle of the zonal divergence over the 50-km-wide land strip along the coast and above the subsidence inversion layer, forced by the diurnal heating cycle over the west slope of the Andes, can produce afternoon regional-scale divergences 5 to 6 times as large as those typically associated with the eastern rim of subtropical anticyclones.

Table 1. Average Altitude of the Cores of the Direct Zonal Flow (Westerly PM, Easterly AM) and of the Associated Return Flow (Easterly PM, Westerly AM) Above (Upper Layer) and Below (Lower Layer) the Subsidence Inversion Base at Michilla for the Three DICLIMA Experiments

		Upper Layer, m asl[a]		Lower Layer, m asl[a]	
Campaign	Zonal Wind	AM	PM	AM	PM
January 1997	return	2900	3500		600
(DICLIMA I)	direct	1500	1700		150
July 1997	return	3000–4000	3000–3700[b]		600–700
(DICLIMA II)	direct	1000–2000	1500–1800		150
January 1998	return	3000[b]	3500		600
(DICLIMA III)	direct	1800	2000		150

[a]Definition: asl, above sea level.
[b]Change in wind direction within the campaign.

4. Summary and Discussion

[31] Results of the DICLIMA experiments have shown two distinct circulation cells above and beneath the subsidence inversion base. These cells are largely uncoupled

Table 2. Average Zonal Wind Divergences Within 500-m Layers Centered at the Cores of the Direct and Return Flows Above The Subsidence Inversion Base Over Michilla[a]

	Summer AM	Summer PM	Winter AM	Winter PM
Return flow	−8	+35	−12	÷0
Direct flow	÷0	+35	−25	+25

[a]See Table 1 and Appendix A. Wind divergences are given in 10^{-6} s^{-1}.

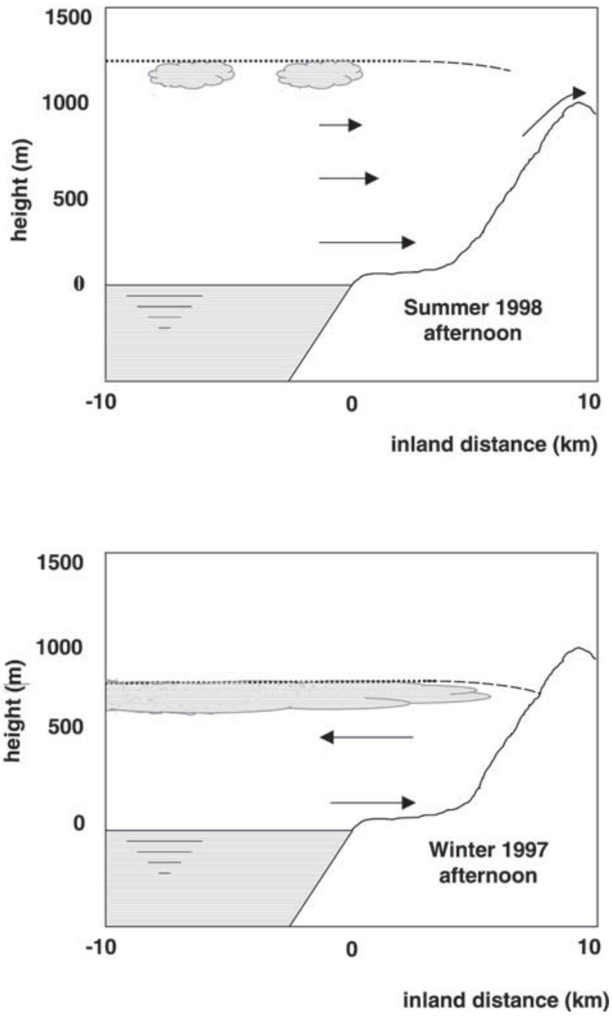

Figure 8. Schematic representation of the afternoon circulation within the coastal marine boundary layer in northern Chile (see Figure 9 caption). (a) January 1998 (austral summer), with a high inversion base associated with warm waters (peak of the 1997–1998 ENSO event). (b) July 1997 (austral winter), with a lower inversion base.

when the inversion base height stays below the coastal summits; otherwise some vertical exchange occurs over the coastal slope (Figure 8, top panel). The most energetic cell is above the inversion where the circulation is forced by the thermal contrast between the top of the MBL and the solar heating cycle over the arid western slope of the Andes. Upslope afternoon winds above 1000 m are stronger in austral summer, without significant changes between S97 and S98 in spite of the large interannual swing in the large-scale ocean and atmospheric conditions associated with the El Niño 1997–1998 event. Downslope winds are stronger in winter. Over the coastal plain, below the subsidence inversion, southwesterly winds dominate in the afternoon and early evening, vanishing thereafter.

[32] The ENSO signal was clearly reflected in the average mean temperature of the air from the surface to 450 hPa over Michilla, being in S98 (El Niño) 3°C to 4°C higher than in S97 (La Niña). Consistently, the subsidence inversion base in S98 was located 300 to 400 m higher than in S97. Actual heights of the 24-hour average subsidence inversion base ranged from 1000 to 1100 m in January 1997, from 800 to 900 m in July 1997 and from 1300 to 1400 m in January 1998, reflecting the extremes of both the annual and the ENSO cycles.

[33] From average zonal wind divergence estimates above the subsidence inversion base, it may be concluded that the regional circulation forced by the diurnal heating/cooling cycle over the west slope of the Andes can produce afternoon divergences 5 to 6 times larger than those typically associated with the eastern rim of the subtropical anticyclones. Other sources of subsidence over this area could include the effect of the South American tropical heat sources [*Gandu and Silva-Dias*, 1998], including those associated with the latent heat release at the South Atlantic and Intertropical Convergence Zones to the east of the main continental heat source. Model results from these authors show maximum subsidence in the middle troposphere at about 85°W, 20°S with a magnitude of about 10% of the typical subsidence associated with the subtropical anticyclone (C. Bretherton, personal communication, 2003). With similar arguments *Rodwell and Hoskins* [2001] ascribe the adiabatic descent west of tropical heat sources to the Rossby wave response interacting with the midlatitude westerlies, suggesting that this mechanisms could explain the strong adiabatic descent off the Chilean coast at 28°S and 79°W at a rate in excess of 3 hPa h^{-1}, considering orography, the South American heating and the southeastern Pacific cooling sources. Therefore the extreme aridity of the desert there can be ascribed to an enhanced afternoon coastal subsidence within and above the subsidence inversion associated to the Andes west slope plus a strong subtropical anticyclone forced by the Rodwell and Hoskins mechanisms. Consistent with this result, the average diurnal cycle in the coastal subsidence inversion shows a lower base and a higher top in the afternoon, the latter in connection with the 3 PM maximum diurnal-cycle temperature anomaly observed in the layer between 2 and 4 km. The strong convergence at dawn above the subsidence inversion base is a new feature that could explain the frequent penetration of dense fogs (locally known as ‘‘camanchacas’’) at that time in the western rim of the Atacama Desert and the nighttime preference for the episodic rains over that region [e.g., *Garreaud and Rutllant*, 1996].

[34] The average daytime decrease in the height of the subsidence inversion base over Michilla, associated with the afternoon zonal divergence above the inversion is not a regular feature, as shown in the individual diurnal cycles depicted in Figure 5. In fact, the growth of the MBL over the coastal plain due to solar warming and entrainment during the light breeze period from mid to late morning would tend to raise the inversion base [e.g., *Garratt*, 1992]. This tendency is further enhanced by afternoon mechanical mixing within the MBL when alongshore winds blow with their maximum intensity. Conversely, at dawn and early morning the regional convergence above the MBL is expected to produce upward motion tending to raise the inversion base, a tendency that could be partially overridden by the nighttime surface cooling and calm conditions within the coastal MBL. Altogether these factors seem to contribute to the dispersion of the individual diurnal cycles in Figure 5.

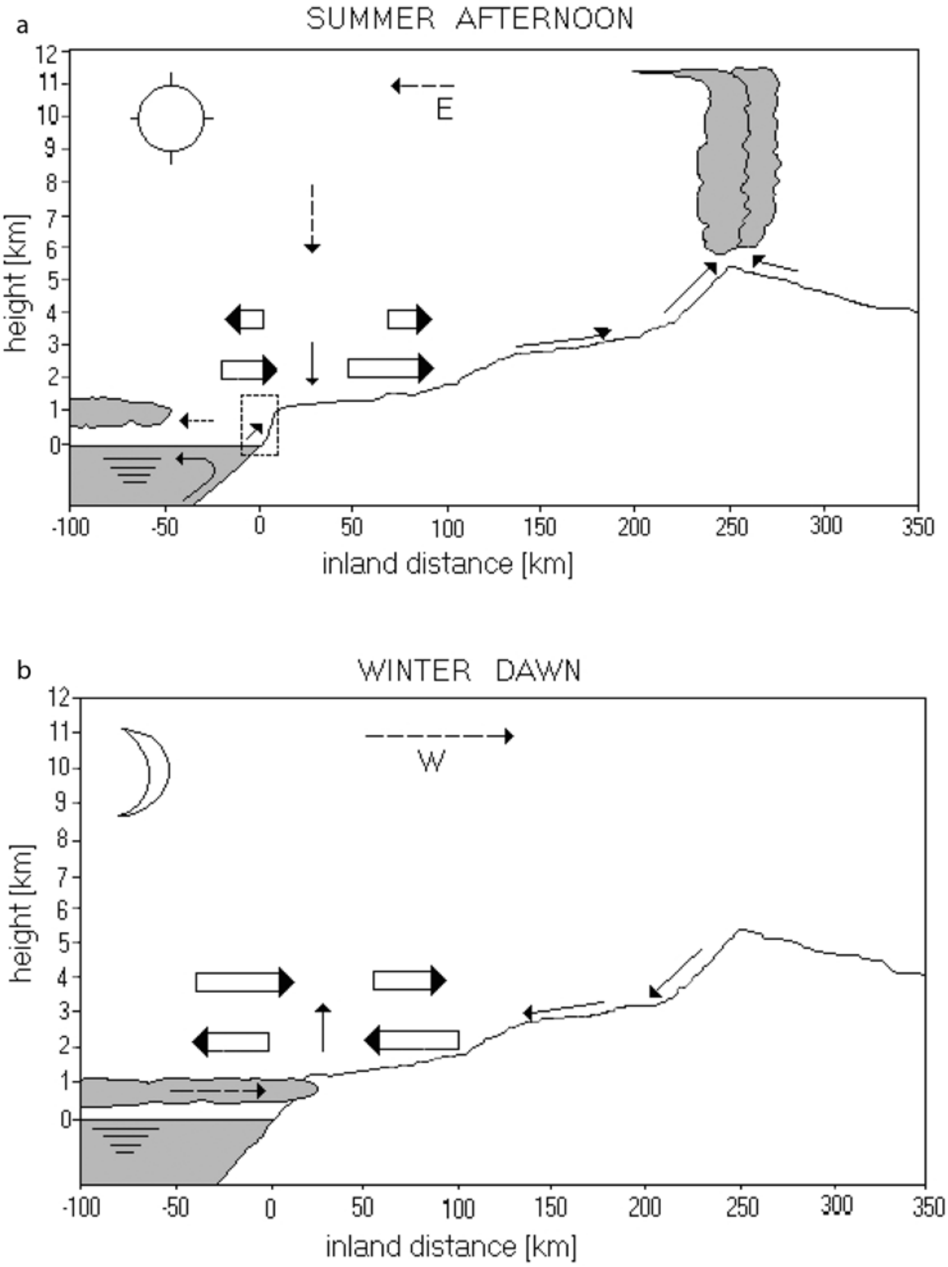

Figure 9. Schematic diagrams of the zonal mass flux (thick arrows) and zonal flow (thin arrows) across the arid northern coast of Chile: (a) austral summer afternoon conditions and (b) austral winter early morning conditions. The dashed rectangle in Figure 9a represents the cross section depicted in Figure 8.

[35] The enhanced convective activity over the Altiplano during S98 (OLRI below 200 W m^{-2}) was a singular period within a weak rainy season there, as usually observed during strong ENSO events [e.g., *Garreaud and Aceituno*, 2001]. A possible influence of this Altiplano convective period on the free atmosphere over Michilla, especially during the first part of the field experiment, could be reflected in the somewhat larger amplitudes on the diurnal cycles of the temperature and mixing ratio observed at about 4000 m altitude associated with an enhanced easterly flow. Figure 9a schematically depicts an austral summer PM circulation (thin arrows) and zonal mass transport (thick arrows) when wet convection is occurring at the western edge of the Altiplano. The average enhanced easterlies near the middle troposphere over Michilla in S98 (not shown) contrast with the average westerlies at that level over Baquedano, as discussed in connection with the highly divergent return flow above the inversion. However, most of the individual zonal wind profiles at Baquedano between 4000 and 6000 m reveal enhanced easterlies, particularly near the top of the layer. This apparent mismatch is due to the averaging process since the few westerly wind events that occurred there were much stronger, probably in connection with enhanced convergence forced by the moist convection over the western Altiplano. It could then be postulated that, westward of the Andean slope, easterly momentum is brought down by sinking motion favored by convergence in the upper troposphere of the subtropical

southeast Pacific where 200 hPa winds blow from the NW [e.g., *Schwerdtfeger*, 1976]. The opposite winter AM conditions are schematically illustrated in Figure 9b when strong westerlies dominate the upper half of the troposphere [e.g., *Schwerdtfeger*, 1976].

[36] Although the 10-day time span of each campaign was thought to average out synoptic-scale disturbances (to be analyzed in a separate contribution), strong 30–40 days intraseasonal variability could result in aliasing of those samples. Studies on the austral summer intrasesasonal variability over the South American Monsoon System [e.g., *Jones and Carvalho*, 2002; *Garreaud*, 1999] show that westerly (easterly) zonal wind anomalies in the upper troposphere characterize dry (wet) spells over the Altiplano region. Dry spells there seem to be associated with the Madden-Julian Oscillation (MJO) through a wave train from the area of enhanced convection in the central equatorial Pacific toward South America, as inferred from *Paegle et al.* [2000], producing equivalent barotropic circulation anomalies similar to the Pacific-South America (PSA) [e.g., *Mo and Higgins*, 1998] pattern. *Jones and Carvalho* [2002] also document this circulation pattern consisting of an anticyclonic anomaly over the southeastern Pacific and a cyclonic one over southern South America at both 850 and 200 hPa levels. The opposite pattern characterizes wet periods in the Altiplano. Therefore the amplitude and phase of the MJO during the three campaigns was assessed through a MJO index (K. Weickmann, personal communication, 2002). Although from October 1996 to June 1997 this MJO index presents large amplitudes in the 30–60 days band, the S97 campaign coincided with a node (not shown) between the convective and the subsident phases of the MJO around the dateline over the central equatorial Pacific. The passage of the subsident phase of the MJO around the dateline resulted in strong rains over the Altiplano in February 1997. Much smaller amplitudes in the MJO index immersed in higher frequency variability dominated both the W97 and S98 campaigns. Significant intraseasonal variability could not be found in the MBL off Michilla. In fact, time series of coastal alongshore winds at Caleta Constitución and subsidence inversion bases at Antofagasta (not shown) only revealed synoptic-scale variability during the three field experiments. In spite of the very strong El Niño conditions that prevailed during the July 1997 and January 1998 experiments, the overwhelming control that radiation exerts on the daily cycles of the atmospheric circulation over the west slope of the Andes seems to guarantee the general validity of the results.

Appendix A

[37] Figure A1 schematically illustrates a budget box of unit width and length L used in the calculation of the mean zonal flux divergence over 500 m-depth layers (H). These layers are centered at the average altitudes of the AM (downslope) and PM (upslope) flows and their corresponding return flows aloft (see Table 1). In order to represent the extremes of the diurnal cycle, individual vertical wind profiles at Michilla and Baquedano were grouped around 8 AM and 5 PM and then averaged over each campaign. Average wind profiles at those hours were then interpolated every 50 m from 1020 m (Baquedano surface altitude) to 4070 m.

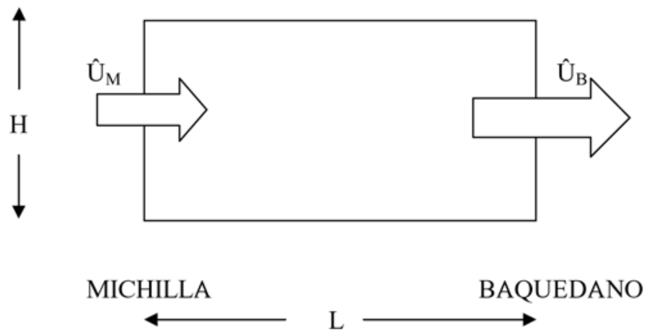

Figure A1. Budget box of unit width and length L used in the calculation of the mean zonal flux divergence over 500-m-depth layers (H) centered on the AM and PM axes of the direct and return flows over the Andean slope. $\hat{U}_M$ and $\hat{U}_B$ are the height-averaged zonal speeds over H at Michilla (M) and Baquedano (B), respectively.

[38] With $\hat{U}_M$ and $\hat{U}_B$ being the height-averaged zonal speeds over H at Michilla (M) and Baquedano (B), respectively and assuming a negligible contribution of density changes over those 500-m-depth layers, the budget equation becomes

$$\left(\hat{U}_B - \hat{U}_M\right)H = \Delta w\ L, \tag{A1}$$

with Δw being the difference of the length-averaged vertical speeds between the top and the botton of the box. If L = 50 km and H = nΔz with Δz = 50 m,

$$\begin{aligned}\Delta w[\mathrm{cm\ s^{-1}}] &= 2 \times 10^{-3} \times 50 \sum_{i=n}^{i=n+10} (U_B - U_M)_i \\ &= \sum (U_B - U_M)/10\end{aligned} \tag{A2}$$

in which n indicates the bottom of the corresponding 500-m layer. In terms of the zonal mean divergence ZDIV we have

$$\mathrm{ZDIV}[10^{-6}\mathrm{s^{-1}}] = \sum (U_B - U_M)/10H \tag{A3}$$

with H in [cm].

[39] **Acknowledgments.** Participants in different aspects of the DICLIMA field experiment from Universidad de Chile (UCH), Universidad de Concepción (UC) and Universidad Catolica del Norte (UCN) were instrumental in the success of the research project. Site/activity, university and researchers/assistants were as follows: Baquedano/radiosoundings, UC, Juan Inzunza/Manuel Aguayo and Carlos Figueroa; Baquedano/pilot balloons, UCH, Hugo Salinas/Gabriel Vargas; Michilla/radiosoundings and surface energy budget, UCH, Humberto Fuenzalida and Patricio Aceituno/Aldo Montecinos, Roberto Rondanelli, Laura Reyes, Rainer Schmitz, Luis Sanchez and Ricardo Bustos; R/V *Abate Molina*/tethersoundings, Jose Rutllant/Rodrigo Sanchez; logistic support, UCN, Ricardo Zuleta/Patricio Zuleta and Alejandro Galleguillos; data processing, UCH, Aldo Montecinos, Hugo Salinas and Rodrigo Sanchez/Zaida Salinas, and UC, Juan Inzunza. Thanks are also due to the captain and crew of the R/V *Abate Molina*. Our warmest thanks to the Municipalidad de Sierra Gorda and Municipalidad de Mejillones for allowing us the use of public schools as operation bases. Additional data were kindly provided by the Universidad Católica del Norte, Antofagasta; Dirección Meteorológica de Chile and the Servicio Hidrográfico y Oceanográfico de la Armada de Chile. This work was supported by Project FONDECYT 5960020 under Program "Biomas y Climas Terrestres y Marinos del Norte de Chile." Additional support for DICLIMA III came from the Department of Research and Development and the Marine Sciences Development Program of the Universidad de Chile under Project I 3667 and

from the Marine Natural Resources Program (SAREC-CONICYT). The first author also benefited from the contribution of the FONDAP-Humboldt, from the Program in Atmospheric Dynamics and Climate (PRODAC) at the Universidad de Chile, and from fruitful discussions with Rene Garreaud, Ricardo Muñoz and David Enfield in the preparation of an early version of this contribution. The anonymous reviewers significantly improved the final version of the manuscript.

References

Arkin, P. A., The relationship between interannual variability in the 200 mb tropical wind field and the Southern Oscillation, *Mon. Weather Rev.*, *110*, 1393–1404, 1982.

Betts, A. K., Diurnal variation of the California coastal stratocumulus from two days of boundary layer soundings, *Tellus, Ser. A*, *42*, 302–304, 1990.

Cornejo, A., Resources of arid South America, in *Arid Lands in Transition*, edited by H. E. Dregne, *Publ. 90*, pp. 345–380, Am. Assoc. for the Adv. of Sci., Washington D. C., 1970.

Gandu, A. W., and P. L. Silva-Dias, Impact of tropical heat sources on the South American tropospheric upper circulation and subsidence, *J. Geophys. Res.*, *103*, 6001–6015, 1998.

Garratt, J. R., *The Atmospheric Boundary Layer*, Cambridge Univ. Press, New York, 1992.

Garreaud, R. D., Multiscale analysis of the summertime precipitation over the central Andes, *Mon. Weather Rev.*, *127*, 901–921, 1999.

Garreaud, R. D., and P. Aceituno, Interannual rainfall variability over the South American Altiplano, *J. Clim.*, *14*, 2779–2789, 2001.

Garreaud, R. D., and J. Rutllant, Análisis meteorológico de los aluviones de Antofagasta y Santiago de Chile en el período 1991–1993 (in Spanish), *Atmósfera*, *9*, 251–271, 1996.

Hastenrath, S., *Climate and Circulation of the Tropics*, D. Reidel, Norwell, Mass., 1985.

Jones, C., and L. M. Carvalho, Active and break phases in the South American monsoon system, *J. Clim.*, *15*, 905–914, 2002.

Klein, S. A., and D. L. Hartmann, The seasonal cycle of low stratiform clouds, *J. Clim.*, *6*, 1587–1606, 1993.

Lettau, H., Small to large-scale features of boundary layer structure over mountain slopes, *Publ. 122*, Dep. of Atmos. Sci., Colo. State Univ., Fort Collins, 1967.

Lettau, H., Dynamic and energetic factors which cause and limit aridity along South America's Pacific coast, in *World Survey of Climatology*, vol. 12, *Climates of Central and South America*, edited by W. Schwerdtfeger, chap. 4 (Appendix 1), pp. 188–192, Elsevier Sci., New York, 1976.

Lettau, H., Explaining the world's driest climate, in *Exploring the World's Driest Climate*, edited by H. Lettau and K. Lettau, *IES Rep. 101*, chap. 12, Univ. of Wis. Press, Madison, 1978.

Luzimar de Abreu, M., and P. Bannon, Dynamics of the South American coastal desert, *J. Atmos. Sci.*, *50*(17), 2952–2964, 1993.

Mahrt, L., Boundary layer moisture regimes, *Q. J. R. Meteorol. Soc.*, *117*, 151–176, 1991.

Mo, K. C., and R. W. Higgins, The Pacific South American modes and tropical convection during the Southern Hemisphere winter, *Mon. Weather Rev.*, *126*, 1581–1596, 1998.

National Centers for Environmental Prediction, Climate Diagnostics Bulletin, *97/1*, Natl. Weather Serv., Natl. Oceanic and Atmos. Admin., Camp Springs, Md., 1997.

National Centers for Environmental Prediction, Climate Diagnostics Bulletin, *98/1*, Natl. Weather Serv., Natl. Oceanic and Atmos. Admin., Camp Springs, Md., 1998.

Paegle, J. N., L. E. Byerle, and K. C. Mo, Intraseasonal modulation of South American summer precipitation, *Mon. Weather Rev.*, *128*, 837–850, 2000.

Parish, T., Barrier winds along the Sierra Nevada Mountains, *J. Appl. Meteorol.*, *21*(7), 925–930, 1982.

Peña, H., A. Grechina, and D. Arcos, Factores oceanográficos que afectan la distribución de los ejemplares juveniles de jurel en la región de Chile centro-sur, paper presented at XVIII Congreso de Ciencias del Mar, Univ. Arturo Prat, 4–8 May 1998.

Prohaska, F., New evidence on the climatic controls along the Peruvian coast, in *Coastal Deserts: Their Natural and Human Environments*, edited by D. H. Amiran and A. W. Wilson, chap. 11, pp. 97–104, Univ. of Ariz. Press, Tucson, 1973.

Rodwell, M. J., and B. J. Hoskins, Subtropical anticyclones and summer monsoons, *J. Clim.*, *14*, 3192–3211, 2001.

Rutllant, J., On the extreme aridity of coastal and Atacama Deserts in northern Chile, Ph.D. thesis, 175 pp., Dep. of Meteorol., Univ. of Wis., Madison, 1977.

Rutllant, J., Coastal lows in central Chile, in *First International Conference on Southern Hemisphere Meteorology*, pp. 344–346, Am. Meteorol. Soc., Boston, Mass., 1983.

Rutllant, J., Natural desertification mechanisms along the arid west coast of South America, in *Proceedings of the Conference on Sand Transport and Desertification in Arid Lands*, edited by F. El-Baz, I. El-Tayeb, and M. Hassan, pp. 235–252, World Sci., River Edge, N. J., 1990.

Rutllant, J., and P. Ulriksen, Boundary layer dynamics of the extremely arid northern part of Chile: The Antofagasta Field Experiment, *Boundary Layer Meteorol.*, *17*, 41–55, 1979.

Rutllant, J., H. Fuenzalida, R. Torres, and D. Figueroa, Interacción océano-atmósfera-tierra en la Región de Antofagasta (Chile, 23°S): Experimento DICLIMA, *Rev. Chil. Hist. Nat.*, *71*, 405–427, 1998.

Schwerdtfeger, W., Introduction, in *World Survey of Climatology*, vol. 12, *Climates of Central and South America*, edited by W. Schwerdtfeger, chap. 1, pp. 1–11, Elsevier Sci., New York, 1976.

Trenberth, K., The definition of El Niño, *Bull. Am. Meteorol. Soc.*, *78*, 2771–2777, 1997.

Trewartha, G., *The Earth's Problem Climates*, Univ. of Wis. Press, Madison, 1961.

Vincent, D. G., Pacific Ocean, in *Meteorology of the Southern Hemisphere*, *Meteorol. Monogr.*, vol. 27, no. 49, edited by D. J. Karoly and D. G. Vincent, chap. 3B, pp. 101–117, Am. Meteorol. Soc., Boston, Mass., 1998.

P. Aceituno, H. Fuenzalida, and J. A. Rutllant, Departamento de Geofísica, Universidad de Chile, Casilla 2777, Santiago, Chile. (jrutllan@dgf.uchile.cl)