Rapid changes in ozone mixing ratios at Cerro Tololo (30°10'S, 70°48'W, 2200 m) in connection with cutoff lows and deep troughs

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[1] Rapid changes in surface ozone mixing ratios (up to 15 ppbv over 12 hours) at the Cerro Tololo (CT) station (70°W, 30°S, 2200 m) over the period 1996–2000 are analyzed. These changes explain most of the wintertime variability of the ozone data at CT. Since the wintertime data show no significant diurnal cycle, local circulations and in situ photochemical processes cannot account for the sudden changes in ozone. Rather, a synoptic-scale forcing of the changes is described on the basis of composite analyses of reanalysis data. Synoptic configurations associated with sudden changes in ozone were separated into W and D cases, according to the local evolution of humidity. W cases are characterized by humidity values in excess of 4 hPa lasting one or two days before the ozone maximum. D cases show no significant change in humidity before the ozone maximum. Both in W and in D cases, stratospheric air appears to be advected from polar latitudes to midlatitudes and subtropical latitudes during the development of cutoff lows and deep troughs. The origin of the air arriving at CT during a W and a D case study was assessed using mesoscale numerical simulations. The model calculations for the W case show a deep stratospheric intrusion reaching the lower troposphere in subtropical latitudes during the amplification and mature stages of a cutoff low. For D cases the stratospheric intrusion is mostly confined to midlatitudes and to an oceanic area far from the site of study. Analyses of trajectories show that in both cases, high ozone mixing ratios at the surface occur in connection with air parcels having a stratospheric history.

INDEX TERMS: 3362 Meteorology and Atmospheric Dynamics: Stratosphere/troposphere interactions; 0365 Atmospheric Composition and Structure: Troposphere—composition and chemistry; 3329 Meteorology and Atmospheric Dynamics: Mesoscale meteorology; 9360 Information Related to Geographic Region: South America; KEYWORDS: stratosphere-troposphere exchange, ozone, cutoff lows


1. Introduction

[2] The spatial and temporal distribution of ozone is crucial in the understanding of tropospheric chemistry. Model calculations have been used to establish the relative importance of different sources and sinks in the tropospheric budget of ozone suggesting that on the global scale, the dominant fluxes are related to photochemical processes [Crutzen et al., 1999, and references therein]. However, the influx of stratospheric air may explain the origin of nearly 40% of tropospheric ozone especially in the sub-tropics [Roelofs and Lelieveld, 1997]. Crutzen et al. [1999], on the basis of results of a three-dimensional (3-D) chemical transport model, suggest that the main region of influx of ozone from the stratosphere is near 30° of latitude in both hemispheres, associated with breaks of the subtropical tropopause. The majority of regional studies concerning stratosphere-troposphere exchange (STE) processes related to ozone have focused on the midlatitudes of the Northern Hemisphere [e.g., Wakamatsu et al., 1989; Ebel et al., 1991; Beekmann et al., 1994; Ancellet et al., 1994; Davies and Schuepbach, 1994; Cox et al., 1995]. In these works, subsidence in connection with cutoff lows (closed cyclonic midtropospheric circulation generated from a precursor...
Moreover, tropopause foldings in the Southern Hemisphere are believed to be less deep than the Northern Hemisphere. Observational evidence of stratospheric intrusions influencing ozone mixing ratios, low humidity and high stability of the STE [e.g., Baray et al., 2000; Meloen et al., 2001]. It is placed in a transition zone between the hyper-arid conditions of the Atacama Desert (to the north of 25°S) and the Mediterranean climate of Central Chile [Miller, 1976]. All the precipitation at CT (less than 100 mm per year) takes place in winter during a few episodes (frontal systems and cutoff lows) whose frequency and intensity have a large

Figure 1. (a) Hourly ozone mixing ratios during 1997. (b) Ozone mixing ratios (thick line) and water vapor pressure measured at CT during July, August and September 1997. Enclosed in boxes labeled W or D are the sudden increases episodes recorded during this period. (c) A closer view of a W episode registered around July 13, 1997.
interannual variability [Rutllant and Fuenzalida, 1991; Pizarro and Montecinos, 2000].

[4] One of the objectives of this work is to contribute to the overall characterization of the CT station that belongs to the Global Atmospheric Watch (GAW) program developed since the mid 90’s in Chile by the World Meteorological Organization (WMO). Preliminary work by Gallardo et al. [2000] established that CT appears to be representative of background conditions of the subtropics of the Southern Hemisphere (SH). The seasonal cycle of ozone at CT has an early spring maximum (35 ppbv October average) and a late summer minimum (25 ppbv March average). The relatively small amplitude of the diurnal ozone cycle during winter (2 ppbv) suggests that in situ photochemical processes are weak at least in the scale of days. In summer, there is a more pronounced diurnal cycle linked to the strengthening of the local thermally driven circulations, which appear to bring ozone-poor air to the station in the afternoon hours. The reader is referred to Kalthoff et al. [2002] for a characterization of the local circulation and its influence on the ozone measurements at CT in summer.

[5] The features earlier described are illustrated in Figure 1. Figure 1a shows the time series of ozone for 1997. In summer, increases and decreases in ozone (~20 ppbv) occur in timescales that span from days to weeks, with ozone and humidity values at the CT station anti-correlated (not shown). Low humidity values generally reflect high static stability and free tropospheric conditions, while high values reflect transport from the marine boundary layer (MBL), which is, on average, poorer in ozone than free tropospheric air. The air within the MBL has typically suffered ozone depletion due to photochemical destruction that takes place in absence of significant sources of nitrogen oxides over the ocean. During the rest of the year, the stronger static stability and the weaker thermally driven circulation restrict the transport from MBL up to CT to episodes of strong mid tropospheric ascent, as will be seen in the next sections.

[6] During winter and spring, rapid changes in ozone are observed (Figure 1b), with increases as large as 15 ppbv in 12 hours. Some of these rapid increases are preceded by high values of humidity and we refer to these episodes to as W cases. Those increases with no previous change in humidity are referred to as D cases. We hypothesize that in both W and D cases the increase in ozone mixing ratios are related with STE processes, as documented for other regions of the world. Davies and Schuepbach [1994] show a case study of an intense stratospheric intrusion in a mountain site at Junfraujoch (46°49’N, 07°59’E, 3580 m.a.s.l.) on the Swiss Alps, reaching ozone mixing ratios of 90 ppbv after the passage of a cold front and an upper level trough. Also, an ozone increase of about 40 ppbv in 12 hours in connection with subsidence in the southwestern part of an upper level trough has been documented by Tsutsumi et al. [1998] for a station at the summit of Mount Fuji (35°21’N, 138°44’E, 3776 m.a.s.l.) in Japan.

[7] The paper is organized as follows. Data sets are described in section 2. An objective definition of the ozone changes or episodes is discussed in section 3. To describe common features and differences between the W and D cases in the synoptic scale, compositing analyses of mete-
orological fields are presented in section 3. To further describe the transport mechanism at play, high-resolution numerical simulations of typical W and D cases were performed. These results are analyzed in section 4. Finally in section 5 summary and conclusions are presented.

2. Data

[8] Since late 1995 the Chilean Meteorological Bureau, within the context of the GAW program under the WMO, has carried out continuous measurements of ozone at CT. Mixing ratios of ozone are measured with an ozone analyzer sensor (TECHO 49-003) whose principle of operation is the attenuation of ultraviolet (UV) light at 254 nm due to the ozone content in the sample. Two photometers operate synchronously measuring this attenuation in the air sample and in a reference gas. The ozone mixing ratio is obtained through the Beer-Lambert law, corrected by internal temperature and pressure. The instrument has a sensitivity of ±1 ppbv, a precision of 2 ppbv and a response time of 20 seconds to reach 95% of response [Thermo Environmental Instruments (TEI), 1994]. The quality of the collected data is regularly checked following the procedures of WMO for GAW stations.

[9] Together with the ozone instrumentation, a complete automatic weather station is operated. It measures atmospheric pressure, temperature, humidity, total radiation, precipitation and wind speed and direction. The meteorological and the ozone data are recorded as 15-minute averages. In this work the data are analyzed for the period between March 1996 and May 2000. In addition, we consider a short period of data during September 2001.

[10] A reanalysis data set from the National Centers for Environmental Prediction/Atmospheric Research (NCEP/NCAR) is used to characterize the large-scale meteorological features. The reader is referred to Kalnay et al. [1996] and Kistler et al. [2001] for details of the reanalysis data set. Spatial resolution is 2.5° latitude × 2.5° longitude in the horizontal and 12 levels in vertical between 1000 and 100 hPa. The temporal resolution is six hours. The fields used in this study are geopotential height, wind, temperature and omega velocity (dp/dt). Although data are in isobaric vertical coordinates, for some analyses, they will be interpolated to isentropic surfaces.

3. Composite Analyses

3.1. Rapid Increases in Ozone

[11] To objectively define the existence of a rapid increase of ozone, two criteria had to be met: (1) The ozone mixing ratio increases more than 15 ppbv over 12 hours. (2) The ozone maximum must be higher than the corresponding
monthly average ozone mixing ratio plus one standard deviation of the daily average.

[12] Since the amplitude of the diurnal cycle is small (2 ppbv in winter and 5 ppbv in summer), the criteria filter out typical diurnal variations related to the local thermal circulation or in situ photochemical activity. The events were further divided according to water vapor pressure, a proxy for MBL air, in two groups. A first group (called W) is composed by cases with high humidity values (up to 4 hPa in water vapor pressure) lasting one or two days before the ozone maximum. A second group (called D) is composed by cases showing low humidity values between two days before and one day after the ozone maximum. Three W cases and two D cases are shown in Figure 1b. In 4 years of data 11 W cases, and 6 D cases were found: 9 in winter, 5 in spring, 3 in autumn and none in summer. W cases are more frequent during winter and D cases during spring. Visual screening of the 500 hPa height field during these episodes reveals a number of common synoptic-scale features among them, so a compositing analysis of W and D cases is used to highlight their mean features. The compositing analysis consists in the average of time series or spatial fields for the selected episodes sorted in time with respect to the time of the ozone maximum, referred to as day 0. In the case of the reanalysis data, available at 00, 06, 12 and 18 Z, the ozone maximum time of each episode was approximated by the nearest of the four.

[13] Figures 2a and 2b show the composite time series of anomalies of ozone mixing ratio and water vapor pressure for W and D cases respectively. The anomalies are calculated with respect to the average values during an episode (an episode spans from four days before to one day after the maximum in ozone). Figures 2c and 2d show the composite time series of omega velocity at 500 and 200 hPa in a grid point located at 30°S and 70°W. In W cases, high humidity (one or two days before the ozone maximum) are associated with marked ascent in the middle and upper troposphere. In the D cases, very weak midlevel ascent occurs before the ozone maximum, and humidity remains almost constant during the whole period. In both cases, the initial increase in ozone occurs about 12–18 h before the maximum, when the vertical motion is still upward, indicating that the ozone increment is not merely explained by descent of midtropospheric air over CT. Two or three days after the maximum, ozone mixing ratios remain higher than the previous values, especially during W cases.

3.2. Geopotential Height and Omega Velocity

[14] To further describe the synoptic-scale circulation associated with the rapid increases in ozone we have performed a compositing analysis of relevant reanalyzed fields: 500 hPa geopotential height and omega vertical velocity. Statistically significant regions of low geopotential height anomalies at 5% confidence level were defined through a Monte Carlo method. This method consists in the generation of a large number (1000 in the present study) of groups composed by N spatial samples randomly chosen from the 5 year reanalysis data (N = 11 for W cases and N = 6
for D cases). For each one of the one 1000 N-elements groups, the average is calculated obtaining the empirical distribution of the average at each grid point. Then, a probability of occurrence is assigned comparing the actual average with the empirical distribution.

3.2.1. W Cases

The composite pattern of 500 hPa geopotential height and omega velocity for the W cases is presented in Figure 3. By day $d = 4$ a deep trough is observed over the Pacific Ocean and it is oriented in the northwest-southeast direction. Strong descent (in excess of 0.1 Pa/s) is observed upstream of the trough axis. As the trough approaches the continent, the magnitude and the extent of the descending vertical motion increases. By day $d = 2$ a closed low in the geopotential field is observed, with its center near $35^\circ S$ and $75^\circ W$ over the ocean. By day $d = 1$ the trough axis is still to the west of the Andes together with a symmetric structure in the vertical motion field. The area of descending motion reaches CT by day $d = 0$, with values near 0.1 Pa/s or 42 hPa/12h. By day $d = 1$ the trough has moved to the east of the subtropical Andes, but relative vorticity and vertical motion have considerably weakened. This weakening is found in most individual cases and it could be associated with the release of latent heat in regions of deep convection over the slope of the Andes. This process may also contribute to irreversible exchange of air between troposphere and stratosphere [e.g., Price and Vaughan, 1993].

In brief, the composite picture of the W cases suggests that the ozone maximum at CT is related to the development of a cutoff low reaching its mature stage a couple of days before this maximum. High values of ozone could be associated with the subsidence induced by the cutoff cyclone. However, the resolution of reanalysis data precludes a description of the mesoscale structures typically associated with cutoff lows such as upper level fronts and tropopause foldings. Mesoscale modeling of a case study will be helpful to clarify this point (see section 4.1). High humidity and low ozone values, registered two or three days before the ozone maximum, are explained by the upward motion at low and middle levels ahead of the incoming midtropospheric cyclonic perturbation (Figure 3, panels $d = 2$ and $d = 1$). The uplift of MBL is further discussed in section 4. This MBL air mixing mechanism could be extrapolated to the events in which ozone decreases associated with humidity increases are observed without a subsequent rapid ozone increase (e.g. at the end of July 1997 in Figure 1b).

3.2.2. D Cases

Figure 4 shows the 500 hPa composite patterns for D cases. By day $d = 4$ a midlatitude long-wave can be observed with a trough over the Pacific Ocean and a ridge over the

![Figure 5. Composite map for W cases at the 320-K isentropic level. PV in PVU, geopotential height in meters and wind in m/s interpolated to 320-K level are shown in shaded areas, solid contours and arrows, respectively. The approximate position of CT is indicated in each panel by a black dot. A reference vector of 40 m/s is shown below the panels. Only wind speeds in excess of 20 m/s are drawn.](image-url)
continent. Approaching the day of maximum ozone, the trough amplifies and further tilts in the northwest-southeast direction. By day $-2$ the strongest subsidence at 500 hPa occurs in a large region far from the continent (110°W to 90°W). By day $-1$ the southern part of the trough has moved towards the Atlantic Ocean and a subtropical cutoff low appears located near 25°S and 90°W. In the following days (0, +1) the cyclonic perturbation weakens over the ocean, without a significant influence in the vertical motion over CT. In contrast to the W cases, the axis of the midtropospheric trough in subtropical latitudes remains far from the coast, leading to a very weak upward motion over this region. Therefore no significant transport of humidity to CT occurs in D cases.

3.3. Potential Vorticity

In the absence of vertical profiles of ozone in the area of study, STE processes become difficult to assess. However, dynamical and chemical fields available may be used as tracers of stratospheric air. The Ertel-Rossby’s potential vorticity (PV) has been widely used as a dynamical proxy of atmospheric motions [e.g., Hoskins et al., 1985]. With some approximations, the PV defined in isentropic coordinates can be written as:

$$PV(x, y, 0) = -g(\zeta_\theta + f)(\partial \theta / \partial p)$$  \hspace{1cm} (1)

where $\zeta_\theta$ is the relative vorticity at isentropic surfaces, $f$ is the planetary vorticity, $\zeta_\theta + f$ is the absolute vorticity at isentropic surfaces, $\theta$ is the potential temperature, $p$ is the atmospheric pressure and $\partial \theta / \partial p$ is a measure of the stability of the atmosphere. The most salient property of PV is its conservation in absence of friction and diabatic processes, established by the Ertel’s theorem. Many studies have confirmed the agreement between PV and other stratospheric tracers such as specific humidity, radioactivity and ozone [e.g., Danielsen, 1968]. A detailed discussion about the so-called PV-chemical analogy and its limitations is given by Haynes and McIntyre [1990].

Figures 5 and 6 show the composite patterns for PV, horizontal wind and geopotential height interpolated to the 320 K isentropic level for cases W and D, respectively. The 320 K isentropic surface is a cross-tropopause level located in the troposphere for low and subtropical latitudes and in the stratosphere in midlatitudes and high latitudes [Hoskins et al., 1985]. The shaded values of PV shown in Figure 5 are smaller than $-1.5$ potential vorticity units (PVU, 1PVU = $10^{-4} \text{m}^2 \text{s}^{-1} \text{Kkg}^{-1}$), which are found in the stratosphere of the Southern Hemisphere.

3.3.1. W Cases

Four days before the ozone maximum at CT, high values of $|PV|$ are intruding from the so-called polar stratospheric reservoir into the southern part of South America (Figure 5). This intrusion becomes narrower and reaches 30°S by day $-2$ when a closed contour of $-2$ PVU is observed over Central Chile. During days $-2$ and $-1$ a clear inversion of the climatological gradient of PV at 70°W
between 50°S and 35°S can be identified. The wind blows nearly parallel to the isopleths of geopotential height except in the northwestern extreme of the PV anomaly. In this part of the system, the isentropic wind is crossing the height contours indicating that adiabatic descent is occurring. By day 0, the PV anomaly crossed the Andes, and by day +1 it has disappeared and the zonal configuration resembles the climatological distribution of the depicted variables.

### 3.3.2. D Cases

[21] For the D cases, the composite shows an anomaly of PV in the form of a tongue oriented from southeast to northwest that reaches 30°S with stratospheric values of PV during day −3 (Figure 6). This tongue seems to be eroded rather than advected (from day −4 to day −2) suggesting that nonconservative processes, such as convection and small-scale turbulence, may have taken place. Again, southerly isentropic winds (not strong enough to be shown in Figure 6) cross the height contours in the northwestern part of the PV anomaly coinciding with the area of descending motion seen in the omega field (compare Figure 3). Finally, only tropospheric values of PV (i.e., \(|PV| < 1.5\) PVU) are found north of 40°S from day −1 to day +1. Nevertheless, traces of the cyclonic anomaly remain in the height and wind fields. Unlike the W cases, no significant

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**Figure 7.** (a) Satellite image of GOES-8 water vapor absorption band for 1200 UTC, July 11, 1997. The gray scale is arbitrary, but darker areas indicate lack of humidity in the midtroposphere and upper troposphere. Letter C indicates the center of the cutoff low derived from the model for this time. (b) Model derived 320-K PV field and isentropic winds for the time in Figure 7a. Only wind vectors with speed in excess of 20 m/s are drawn. The vertical line indicates the position of the cross section shown in Figure 8.
vertical transport near CT can be inferred from this composite. However, regions of mesoscale descent could be masked due to the coarse resolution of reanalysis data. Besides, air descending during the decaying stages of the system off the coast of South America could be subsequently transported to CT. The strong subtropical jet stream over CT during these episodes could also influence the vertical motion and the transport down to the lower troposphere. For instance, the breaking of gravity waves excited by the concurrence of the topography and the jet stream may lead to an irreversible exchange of air from the stratosphere to the troposphere, as observed in other mountain regions [Lamarque, 1996]. As in the W cases, these hypothesis derived from the coarse resolution reanalysis data will be tested in our numerical simulation of a typical D case.

4. Mesoscale Simulations

The composite patterns shown in section 3 suggest that stratospheric air is advected from polar to subtropical latitudes together with the development of cutoff cyclones (in W cases) and deep troughs (in D cases), with a frequency of four to five episodes per year mainly during winter and early spring. However, the way in which ozone-rich air could be transported to the lower free troposphere is still unclear. The temporal and spatial resolution of the reanalysis data is not enough to calculate accurate trajectories or to observe mesoscale features such as fronts or tropopause foldings. For this reason, we explore in detail the transport of ozone-rich air for both cases by means of sub-synoptic numerical simulations.

Non-hydrostatic numerical simulations were performed using the Fifth-Generation Pennsylvania State University-National Center for Atmospheric Research Mesoscale Model (MM5) (for details, see Grell et al. [1995]). The selected domain covers the region 95°W–45°W, 45°S–15°S centered at CT, with a horizontal resolution of 60 km and 37 vertical sigma (σ) levels from σ = 0.995 to σ = 0.05. Model outputs were interpolated to 14 pressure levels between 1000 and 100 hPa. The physical parameterizations used in this simulation include simple ice microphysics, the...
Kain-Fritsch convective scheme, the Blackadar boundary layer scheme and a cloud-radiation scheme [Dudhia, 1989; Kain and Fritsch, 1993; Blackadar, 1979]. Initial conditions and time-dependent boundary conditions were obtained by interpolating global NCEP surface and upper air analysis available every 12 h. MM5 simulations using similar physical parameterizations, and spatial resolution over this region have been successfully performed in earlier works [e.g., Garreaud, 1999].

4.1. W Case

The selected W case took place between July 9 and 13 1997. Figure 1c shows the ozone mixing ratios and the water vapor pressure during that episode. Ozone mixing ratios increase from 25 ppbv on July 12 00Z to about 50 ppbv on July 13 12Z. This case was selected because it is representative of a strong enhancement of ozone and also an intense upper level frontogenesis, as shown later.

The main features observed in this particular case coincide with those observed in the composite maps of W cases (compare Figures 3 and 5), namely, the development of a deep trough in the geopotential field, an upper level cyclone passing over CT area and the weakening of the mid tropospheric system downstream the Andes. The comparison between the water vapor (WV) satellite images and their corresponding isentropic PV and wind maps is illustrative of the capability of MM5 in reproducing the cutoff low structure. An example of this correspondence is shown in Figure 7. The darkest zone in the northern part of the cyclone in Figure 7a coincides closely with the area in which the stratospheric PV values are found deeper in the troposphere. The dark strip crossing the southern part of South America and linking the precursor trough with the cutoff low (Figure 7a) has also its counterpart in Figure 7b as a high |PV| band or “streamer” along 37°S (see Appenzeller et al. [1996] for a complete discussion of the atmospheric transport patterns inferred from WV images). These features of the simulation suggest a realistic representation of the synoptic-scale processes instrumental in the STE.

Meridional cross sections of PV and humidity derived from the model confirm the existence of a deep intrusion of stratospheric air, surrounding the core of the cyclone and being deeper in its equatorial side (Figure 8). In Figure 9, The time evolution of vertical profiles of PV, wind and humidity over a grid point near CT is shown in Figure 9. Small cyclonic values of PV and northwesterly winds are found in the troposphere before July 12 00Z. Later, low humidity and stratospheric values of PV can be found at middle and low levels in the troposphere. Dry and stable air is associated with southwesterly winds on July 13 15 00Z, concurrently with the increase in ozone at CT. The downward intrusion of stratospheric PV values is observed to occur by July 13 between 00Z and 12Z. Noteworthy, the time of |PV| maximum at 800 hPa (close to CT level) coincides with the time in which the ozone maximum mixing ratios were registered at CT.

From the model outputs, 3-D 48h-backward trajectories arriving at CT were calculated (Figure 10). Trajectories...
are labeled according to the date of arriving at CT. For instance, the label 00Z 11 indicates that the trajectory that has its origin at July 9 00Z in the point indicated by the arrow and, it arrives at CT at July 11 00Z. Before July 12 06Z, the trajectories originate to the north of CT, over the ocean and nearly at the same altitude of the station (~2000 m). After July 12 06Z, the trajectories originate in the midtroposphere and upper troposphere near 45⁰S. The average values of PV along the trajectories are presented in Figure 11. They show a close correspondence between the time of maximum ozone and the maximum of average $|PV|$. Thus, air parcels with a more stratospheric history (in the sense of the average PV) are positively correlated with high ozone values measured at CT (see Figure 11). Because of the relatively deep extent of the stratospheric intrusion, it is not expected a correspondence between the height of the trajectories and the measured ozone mixing ratios. For instance an air parcel originated at 3000 m crossing through the intrusion is expected to be richer in ozone than a parcel starting at 8000 m without crossing the intrusion.

[28] The maximum ozone registered during the July 13 event (50 ppbv) is in the lower range of observed ozone mixing ratios associated with stratospheric intrusions, typically higher than 80 ppbv [e.g., Keyser and Shapiro, 1986]. This suggests that the stratospheric air reaching CT could be diluted by small-scale turbulence below the intrusion or subject to photochemical depletion processes.

[29] In sum, the rapid changes in ozone observed during W cases have their origin in air transported from the stratosphere. Three stages of this transport could be identified. First, transport along the cross tropopause isentropic surfaces which lead air from the polar stratospheric reservoir into midlatitudes and subtropical latitudes in the form of a deep trough. Second, the subsequent formation of a cutoff low and a deep stratospheric intrusion reaching the midtroposphere and low troposphere. And finally a third stage represented by the smaller-scale processes that finally bring the ozone enriched air to CT during a relatively short period of time.

4.2. D Case

[30] In this case, the simulated period spans from September 16 00Z to September 21 18Z, 2001. The maximum ozone mixing ratio was registered on September 20 01Z (51 ppbv). The main synoptic features during this episode have a close resemblance to those observed in the composite of D cases, including a deep trough becoming tilted with a mid tropospheric cyclone not well resolved and relatively far from the continent, a rather stationary behavior of the tilted trough and erosion of the PV tongue.

[31] About 3 days before the ozone maximum, a weak cutoff low crossed over CT, resulting in no major changes in ozone or water vapor at the surface. Later a trough reaching subtropical latitudes, can be observed as a dark tongue and high $|PV|$ values in the Figures 12a and 12b, respectively. Again, there is a good correspondence between the synoptic-scale patterns as observed from WV images and model derived PV.

[32] Figure 13 shows a cross section along 40⁰S illustrating the vertical structure of the streamer. We can observe that this intrusion is narrower and less deep than the W case intrusion (compare Figure 8), however the low values of humidity and the core of high $|PV|$ values near 700 hPa and 80⁰W suggest that stratospheric air is being transported downward and irreversibly near this intrusion. As it was hypothesized in previous sections, the origin of ozone in D cases appears to be associated with the STE occurring in a region farther to the south from CT. Indeed, the time evolution of vertical profiles of PV over CT for this case shows that at this subtropical latitude the high $|PV|$ values remain above the 350 hPa level.

[33] The trajectory analysis reveals that the ozone-rich air intruding from higher latitudes descends anticyclonically to...
CT level from the intrusion depicted in Figure 13. At the beginning of the episode the trajectories are coming from the south and mainly from the same level of CT (Figure 14). As the ozone at CT begins to increase, the trajectories come from higher altitudes and near the area of the intrusion. The average $|\mathcal{PV}|$ along the trajectories in Figure 15 also shows a good correlation with the ozone at the surface although in this case the maximum in ozone occurs three hours after the maximum of average $|\mathcal{PV}|$ along the trajectories. In this case, the height of origin is a better indication of the ozone content of the air parcels than in W cases. Although the heights of the trajectories 24 hours before the ozone maximum do not show significant differences (Figure 14), there is much better correspondence between the initial height of the trajectories 48 hours before they arrive to CT and the ozone at CT. Why height is a better proxy of ozone than PV in D cases can be explained by the less local character of the intrusion as compared with the W cases. In the D case, the ozone-rich air has a longer distance to travel before to reach CT from its midlatitude stratospheric intrusion. Consequently, non-conservative processes and photochemistry have more time to act limiting the utility of the PV-ozone analogy. This is an important difference between W and D cases that also helps to explain the differences in the observed correspondence between ozone and average PV along the trajectories. In the W case, the PV-ozone analogy is expected to work more adequately since the high $|\mathcal{PV}|$ values come from a deep and localized stratospheric intrusion at low and middle levels that occurs over a relatively short time period (taking only a few hours to reach CT from the core of the intrusion).

[34] In this D case, moist MBL air does not arrive to CT before the ozone maximum, in contrast with the previous simulation (compare Figures 9 and 16). The cause of this

**Figure 12.** Same as Figure 7, except for 0000 UTC, September 18, 2001. Horizontal line in Figure 12b indicates the position of the cross section shown in Figure 13.
difference can be attributed to the different forcing of the MBL by the synoptic-scale circulations. In order to compare these forcings, average vertical velocities at 850 hPa two days before the maximum of ozone, over a grid point near CT, were calculated. The results are 0.052 m/s (marked ascent) for the W case and -0.004 m/s (weak subsidence) for the D case. Moreover, during the W case strong, tropospheric-deep ascent is found ahead of the cutoff low, which tracks very close to CT. In the D case, an area of ascent and relatively high humidity values is also found ahead of the trough, but it does not affect subtropical latitudes. Thus, the magnitudes of the midlevel vertical velocity and their impact upon the humidity at CT for these simulations are in agreement with our previous findings from the compositing analysis.

5. Summary

[35] Rapid increases in ozone mixing ratios (up to 15 ppbv in 12 hours) are observed at Cerro Tololo (30°10'S, 70°48'W, 2220 m.a.s.l.) mainly during winter and spring. These increases can be divided into two groups, the so-called W and D cases. W cases are characterized by water vapor pressure values in excess of 4 hPa and relatively low ozone values lasting one or two days before the ozone maximum. D cases show no significant change in humidity or ozone before the rapid ozone increase.

[36] Compositing analyses of large-scale meteorological data suggest that rapid increases in ozone mixing ratios at CT during W cases are related with a cutoff low development in the Subtropical Pacific in front of the coast of South America, and its subsequent crossing of the Andes at subtropical latitudes (that is, near CT). The D cases are associated with deep troughs that drift slowly over the subtropical Southeast Pacific. In these cases, the maximum ozone at the surface is registered when the axis of the trough is still to the west of the subtropical Andes. Deep troughs and cutoff low systems energetic enough to reach subtrop-

Figure 13. Zonal cross section along 40°S of PV (shaded) and specific humidity (continuous line) on September 18, 2001, at 00 UTC. The black area indicates the topography.

Figure 14. Same as Figure 10, except for September 18 00Z to September 20 18Z, 2001.
ichal latitudes are able to transport stratospheric ozone-rich air from polar latitudes. These events appear to be the main explanation for the observed synoptic-scale variability of ozone measurements at CT during winter and early spring.

[37] The sub-synoptic simulations of two case studies confirms the overall picture derived from the reanalysis data. In the W case study the ozone-rich air is transported from an intense stratospheric intrusion surrounding the core of a cutoff low. The ascent in the leading side of the cutoff low forces the uplift of the MBL, increasing the humidity values and lowering the ozone mixing ratios at CT. At the time when the cyclone’s axis passes over CT, stratospheric PV values are observed to intrude to the lower troposphere near CT level. Thus, transport of rich ozone air from the stratosphere explains the rapid change in ozone registered in this case study. This result is extrapolated to cutoff lows and deep troughs reaching the western coast of subtropical South America as suggested by the coherence of the composite fields derived from reanalysis data.

[38] The simulation of a D case study exhibit also the presence of a stratospheric intrusion associated with a deep trough which is narrower and less deep than the W intrusion, and confined mainly to midlatitudes. The ozone-rich air is originated in a midlatitude stratospheric intrusion

**Figure 15.** Same as Figure 11, except for D case study.

**Figure 16.** Same as Figure 9, except for D case study.
and subsequently advected down- and southwestward, eventually reaching CT. During this case, the trough remains over the ocean, without a significant influence in the synoptic-scale forcing over CT. This explains the absence of major changes in humidity during D cases.

[29] Since only about four years of measurements are available and the area is characterized by a large interannual variability, explained to some extent by the El Niño/Southern Oscillation, the influence of stratospheric intrusions in the seasonal variation of ozone cannot be fully addressed. The events show a preferred occurrence during winter and early spring, coinciding with the seasonal frequency of cutoff lows in the western coast of South America [Pizarro and Montecinos, 2000] and with the seasonal behavior of stratospheric intrusions at several mountain sites in the Northern Hemisphere [e.g., Beekmann et al., 1997].

[40] Additional measurements such as beryllium 7 isotope and ozone vertical profiles would be helpful to better assess the regional features of these intrusions. Furthermore, ozone profiles could be used to establish correlations between modeled PV and ozone, in order to obtain a quantitative estimate of ozone transport during these events [e.g., Ancellet et al., 1994]. Also, further modeling efforts are required, for instance to assess the role of topography. In fact, tropopause foldings might be deep enough to be obstructed by the high Andes. Then, turbulence and the latent heat release by forced ascent might contribute both to the exchange between the troposphere and stratosphere and to the weakening of the midtropospheric cyclonic system.

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