Meteorological Parameter Analysis above Dome C Using Data from the European Centre for Medium-Range Weather Forecasts

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ABSTRACT. We present a characterization of all the principal meteorological parameters (wind speed and direction, pressure, and absolute and potential temperature) that extends up to 25 km from the ground above Dome C, Antarctica, for 2 years (2003 and 2004). The data set is composed of “analyses” provided by the general circulation model (GCM) of the European Centre for Medium-Range Weather Forecasts (ECMWF), and they are part of the MARS catalog. A monthly and seasonal (summer and winter) statistical analysis of the results is presented. The Richardson number is calculated for each month of the year over 25 km in order to study the relative stability of the atmosphere. This allows us to trace a map indicating where and when the optical turbulence has the highest probability of being triggered in the whole troposphere, tropopause, and stratosphere. Finally, we try to predict the best expected isoplanatic angle and wave-front coherence times ($\theta_{0,\text{max}}$ and $\tau_{0,\text{max}}$, respectively) by employing Richardson number maps, wind speed profiles, and simple analytical models of vertical profiles.

1. INTRODUCTION

The Antarctic Plateau has for years been known to be particularly attractive for astronomy (Fossat 2005; Storey et al. 2003). It is extremely cold and dry, and this makes the site an interesting candidate for astronomy in the long-wavelength ranges (infrared, submillimeter, and millimeter), thanks to the low sky brightness and high atmospheric transmission caused by a low temperature and a concentration of water vapor in the atmosphere (Valenziano & Dall’Oglio 1999; Lawrence 2004; Walden et al. 2005). The Antarctic Plateau is situated at high altitudes (the whole continent has an average height of $\sim$2500 m) and is characterized by a quite peculiar atmospheric circulation and a very stable atmosphere, so that the level of optical turbulence ($C_N^2$) profiles in the free atmosphere is predominantly lower than that of any midlatitude site (Marks et al. 1996, 1999; Aristidi et al. 2003; Lawrence et al. 2004). Gillingham (1991) first suggested such a low level of optical turbulence above the Antarctic Plateau. Atmospheric conditions generally degrade with proximity to the coasts. A low level of optical turbulence in the free atmosphere is generally associated with large isoplanatic angles ($\theta_0$). The coherence wave-front time ($\tau_\theta$) is claimed to be particularly large above the Antarctic Plateau, due to the combination of a weak $C_N^2$ and a low wind speed along the entire troposphere. Under these conditions, an adaptive optics system can reach better levels of correction (minor residual wave-front perturbations) than those obtained by an equivalent AO system above midlatitude sites. Wave-front correction at high Zernike orders can be more easily reached over a large field of view, the wave-front corrector can run at reasonably low frequencies and for observations with long exposure times, and this can be done in a closed loop. This could prove to be particularly advantageous for some scientific programs, such as searches for extrasolar planets. Of course, interferometry would also benefit from a weak $C_N^2$ and $\tau_\theta$.

In the last decade, several site-testing campaigns have taken place, first above South Pole (Marks et al. 1996, 1999; Loewenstein et al. 1998; Travaouillon et al. 2003a, 2003b), and more recently, above Dome C (Aristidi et al. 2003, 2005a; Lawrence et al. 2004). Dome C seems to have some advantages with respect to the South Pole:

1. The sky emission and atmospheric transparency are a few orders of magnitude better than above South Pole (Lawrence 2004) at some wavelengths. The sensitivity (depending on the decrease in sky emission and increase in transparency) above Dome C is around 2 times better than above South Pole in near- to mid-infrared regions, and around 10 times better in mid- to far-infrared regions.

2. The surface turbulence layer, which is principally created by katabatic winds, is much thinner above Dome C (tens of meters; Aristidi et al. 2005a; Lawrence et al. 2004) than above
South Pole (hundreds of meters; Marks et al. 1999). The thickness and strength of this layer are indeed tightly correlated to the katabatic winds, which are unique winds that develop near the ground and characterize the boundary layer circulation above the whole Antarctic continent. Katabatic winds are produced by the radiative cooling of the iced surface, which, by conduction, cools the air in its proximity. The cooled air in the proximity of the surface becomes heavier than the air in the upper layers and, because of a simple gravitational effect, moves down along the slope of the ground, with a speed that increases with the slope. Dome C is located on the top of an alitipano in the interior region of Antarctica, and for this reason, the katabatic winds are much weaker above Dome C than above other sites in this continent, such as South Pole, which is situated on a more accentuated sloping region.

At present, not much is known about the typical values of meteorological parameters above Dome C during the wintertime (April–September), which is the most interesting period for astronomers.

The goals of our study are the following:

1. We intend to provide a complete analysis of the vertical distribution of the main meteorological parameters (wind speed and direction, absolute temperature, and pressure) in different months of the year using ECMWF (European Centre for Medium-Range Weather Forecasts) data. Particular attention is paid to the wind speed, a key element in the estimation of the wavefront coherence time \( \tau_{\theta} \). The ECMWF data set is produced by the ECMWF general circulation model (GCM) and is therefore reliable at synoptic scale (i.e., at large spatial scales). This means that our analysis can be extended to the whole troposphere, and even to the stratosphere up to 20–25 km. The accuracy of such data is not particularly high in the first meters above the ground, due to the fact that the orographic effects produced by the friction of the atmospheric flow above the ground are not necessarily well reconstructed by the GCMs.\(^1\) We remind the reader that a detailed analysis of the wind speed near the ground above Dome C, extended over a timescale of 20 yr, was recently presented by Aristidi et al. (2005a). In that paper, measurements of wind speed taken with an automatic weather station (AWS)\(^2\) are used to characterize the typical climatological trend of this parameter. In the same paper, it is noted that estimates of the temperature near the ground are provided by Schwerdtfeger (1984, p. 15).

The interested reader can find in these references information on the value of this meteorological parameter above Dome C and near the surface. Our analysis can therefore complete the picture by providing typical values (seasonal trend and median values) of the meteorological parameters in the high part of the surface layer, the boundary layer, and the free atmosphere. Thanks to the large and homogeneous temporal coverage of ECMWF data, we are able to put forth evidence of the typical features of the meteorological parameters in the summer and winter and the variability of the meteorological parameters in different years. The winter is particularly attractive for astronomical applications, due to the persistence of the “nighttime” for several months. This period is also the most difficult one in which to carry out measurements of meteorological parameters, due to logistical problems. For this reason, ECMWF data offer a useful alternative to measurements for monitoring the atmosphere above Dome C over long timescales in the future.

2. We intend to study the conditions of stability and instability of the atmosphere that can be measured by the Richardson number \( R_i \), which depends on both the gradient of the potential temperature and the wind speed: \( R_i = R (\theta / h, \theta / h) \). The trigger of optical turbulence in the atmosphere depends on both the gradient of the potential temperature \( (\theta / h) \) and the wind speed \( (\theta / h) \); i.e., on the \( R_i \). This parameter can therefore provide useful information on the probability of finding turbulence at different altitudes in the troposphere and stratosphere in different periods of the year. Why this is interesting? At present, we have indications that above Dome C, the optical turbulence is concentrated in a thin surface layer. Above this layer, \( \tau_{\theta} \) is exceptionally large, indicating an extremely low level of turbulence. The astronomical community has so far collected several pieces of evidence certifying the excellent quality of the Dome C site, in addition to different solutions that might be envisaged to overcome the strong surface layer, such as raising a telescope above 30 m or compensating for the surface layer with AO techniques. The challenging question now is how to more precisely establish how much better the Dome C is than a midlatitude site. In other words, what are the “typical” \( \tau_{\theta} \), \( \theta_{\theta} \), and \( \theta_{\theta} \) values that we can expect from this site? What we mean here by “typical” are the values that repeat with statistical relevance, such as a mean or a median value. For example, the gain in terms of impact on instrumental performance and astrophysical feedback can strongly change, depending on how weak the \( C^2_{\nu} \) is above the first 30 m. In spite of the fact that \( C^2_{\nu} = 10^{-14} \), \( C^2_{\nu} = 10^{-15} \), and \( C^2_{\nu} = 0 \) are all small quantities, they can have a different impact on the final values of \( \tau_{\theta} \), \( \theta_{\theta} \), and \( \theta_{\theta} \). Only a precise estimate of this parameter will provide to the astronomical community useful elements to better plan future facilities (telescopes or interferometers) above the Antarctic Plateau and to correctly evaluate the real advantage obtained in terms of turbulence by choosing the Antarctic Plateau as an astronomical site. With the aid of the Richardson number, the wind speed profile, and a simple analytical \( C^2_{\nu} \) model, we try to predict \( \tau_{\theta,\text{max}} \) and \( \theta_{\theta,\text{max}} \) without the contribution of the first 30 m of atmosphere.

3. Data provided by ECMWF can be used as inputs for atmospheric meso-scale models that are usually employed to simulate the optical turbulence \( C^2_{\nu} \) and the integrated atmospheric parameters (Masciadri et al. 2004; Masciadri & Egner 2004, 2005). Measurements of wind speed performed during the summer have been recently published (Fig. 1; Aristidi et

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\(^1\) It was recently shown (Masciadri 2003) that meso-scale models can estimate the near-ground wind above astronomical sites better than GCMs.

al. 2005a). We intend to estimate the quality and reliability of the ECMWF data by comparing these values with measurements from Aristidi et al. so as to have an indication of the quality of the initialization data for meso-scale models. We have planned applications of a meso-scale model (Meso-Nh) to Dome C in the near future. As a further output, this model will be able to more accurately reconstruct the meteorological parameters near the ground, compared to the ECMWF data set.

The paper is organized as follows: In § 2 we present the median values of the main meteorological parameters and their seasonal trend. We also present a study of the Richardson number, tracing a complete map of the instability and stability regions across the whole 25 km range, on a monthly statistical basis. In § 3 we study the reliability of our estimate, comparing the ECMWF analysis with measurements. In § 4 we try to retrieve the typical values of $\tau_{0,\text{max}}$ and $\theta_{0,\text{max}}$ above Dome C. Finally, in § 5 we present our conclusions.

2. METEOROLOGICAL PARAMETER ANALYSIS

The characterization of the meteorological parameters in this paper is done with “analyses” extracted by the MARS (Meteorological Archival and Retrieval System) catalog of the ECMWF. An “analysis” provided by the ECMWF general circulation model (GCM) is the output of a calculation based on a set of spatiotemporal interpolations of measurements provided by meteorological stations distributed on the surface of the whole world, and by satellites and instruments carried aboard aircraft. These measurements are continuously updated, and the model is fed with new measurements at regular intervals of a few hours. The outputs are formed by a set of fields (scalars and/or vectors) of classical meteorological parameters sampled across the world, with a horizontal resolution of 0.5°, corresponding to roughly 50 km. This horizontal resolution is much better than that of the NCEP/NCAR reanalyses, having a horizontal resolution of 2°, so we can expect a more accurate estimate of the meteorological parameters in the atmosphere. The vertical profiles are sampled over 60 levels that extend up to 60 km. The vertical resolution is higher near the ground (~15 m above Dome C) and weaker in the high part of the atmosphere. In order to give an idea of the vertical sampling, Figure 1 shows the output of one data set (wind speed and direction and absolute and potential temperature) from the MARS catalog (extending in the first 30 km) corresponding to

![Vertical profiles of wind speed, wind direction, and absolute and potential temperature](image-url)
levels at which estimates are provided. We extracted from the ECMWF archive a vertical profile of all the most important meteorological parameters (wind speed and direction, pressure, and absolute and potential temperature) at the coordinates 75° south, 123° east, at 00:00 UT each day for the years 2003 and 2004. We verified that the vertical profiles of the meteorological parameters extracted from the nearest four grid points around the Dome C (75°06′25′′ south, 123°20′44′′ east) show negligible differences. This is probably due to the fact that the orography of the Antarctic continent is quite smooth and flat in the proximity of Dome C. Above this site, we can identify on an orographic map a difference in altitude on the order of a few meters over a 60 km surface (Masciadri 2002), roughly the distance between two contiguous grid points of the GCM. The orographic effects on the atmospheric flow are visibly weak at such a large spatial scale across the whole 25 km range. We therefore consider these profiles of meteorological parameters at a macroscopic scale to be good representations of the atmospheric characteristics above Dome C, starting from the first tens of meters, as explained above.

2.1. Wind Speed

The wind speed is among the most critical parameters defining the quality of an astronomical site. It plays a fundamental role in triggering optical turbulence ($C^2_v$), and it is a fundamental parameter in the definition of the wave-front coherence time $\tau_0$:

$$\tau_0 = 0.049\lambda^2 \left( \int V(h)^{3/5} C^2_v(h) \, dh \right)^{-3/5},$$

where $\lambda$ is the wavelength, $V$ is the wind speed, and $C^2_v$ is the optical turbulence strength. Figure 2 shows the median vertical profile of the wind speed obtained from the ECMWF analyses during the years 2003 (Fig. 2a) and 2004 (Fig. 2b). Dotted lines indicate the first and third quartiles (i.e., the typical dispersion at all heights). Figure 2c shows the variability of the median profiles obtained during the 2 years. We can see that from a qualitative (shape), as well as quantitative, point of view (values), the results are quite similar in different years. They can therefore be considered typical of the site. Due to the particular synoptic circulation of the atmosphere above Antarctica (the so-called polar vortex), the vertical distribution of the wind speed in the summer and winter is very different. The wind speed has important seasonal fluctuations above 10 km. Figure 3 shows the median vertical profiles of the wind speed in the summer (left) and winter (right) in 2003 (top) and 2004 (bottom). We can see that the wind speed is quite weak in the first ~10 km from sea level during the whole year, with a peak at around 8 km from sea level (5 km from the ground). At this height, the median value is 12 m s$^{-1}$, and the wind speed is rarely higher than 20 m s$^{-1}$. Above 10 km from sea level, the wind speed is extremely weak during the summertime, but during the wintertime, it monotonically increases with the height, reaching values on the order of 30 m s$^{-1}$ (median) at 20 km. The typical seasonal wind speed fluctuations at 5 and 20 km are shown in Figure 4. This trend is quite peculiar and is different from that observed above midlatitude sites.

In order to give an idea of such differences, we show in Figure 5 the median vertical profiles of the wind speed estimated above Dome C in summer (dashed line) and winter (thick solid line) and above the San Pedro Mártir Observatory (Mexico) in summer (dotted line) and winter (thin solid line) (Masciadri & Egner 2004, 2005). San Pedro Mártir is located in Baja California (31°04′41″ north, 115°45′69″ west) and is taken here as representative of a midlatitude site. Above midlatitude sites, we can see that the typical peak of the wind speed at the height of the jet stream (roughly 12–13 km from sea level) has a strong seasonal fluctuation. The wind speed is higher during the wintertime (thin line) than during the summer (dotted line) in the Northern Hemisphere, and the opposite happens in the Southern Hemisphere. At this height, the wind speed can reach seasonal variations on the order of 30 m s$^{-1}$. Near the ground and above 17 km, the wind speed strongly decreases to low values (rarely higher than 15 m s$^{-1}$). During the wintertime, the wind speed above Dome C at 20–25 km can reach values that are comparable to the highest wind speed values obtained above midlatitude sites at the height of the jet stream (i.e., 30 m s$^{-1}$). On the other hand, one can see that in the first 12 km from sea level, the wind speed above Dome C during the wintertime is weaker than the wind above midlatitude sites in any period of the year.

Figure 6 shows monthly median vertical profiles of the wind speed during 2003 (green line) and 2004 (red line). The different features of the vertical distribution of the wind speed that we have just described and attributed to the winter and summer are more precisely distributed in the year in the following way. During December, January, February, and March, the median wind speed above Dome C at 20–25 km is not higher than 10 m s$^{-1}$. During the other months, starting at 10 km, the median wind speed increases monotonically at different rates. September and October show the steepest wind speed growth rates. It is worth noting that the same wind speed vertical distributions appear in the same months in different years. Only during the month of August is it possible to recognize substantial differences of the median profile in the years 2003 and 2004. This result is extremely interesting, permitting us to predict, in a quite precise way, the typical features of the vertical distribution of wind speed in different months. Figure 7 shows the cumulative distribution of the wind speed at 8–9 km from sea level during each month. We can see that in only 20% of the cases does the wind speed reach values on the order of 20 m s$^{-1}$ during the wintertime. This height (8–9 km) corresponds to the troposphere-tropopause interface above Dome C. As explained below, this is generally one of the places in which the optical turbulence can be more easily triggered, due to the strong gradient and value of the wind speed. We note that similarly to the activity above midlatitude sites with respect to this interface, we find a local peak of the wind speed. In spite of this value,
is much smaller above Dome C than above midlatitude sites. We can therefore expect a less efficient production of turbulence at Dome C than above midlatitude sites at this height.

2.2. Wind Direction

Figure 8 shows monthly median vertical profiles of the wind direction during 2003 (green line) and 2004 (red line). We can see that during the all months, in the low part of the atmosphere, the wind blows principally from the south (∼200°). In the troposphere (1–11 km), the wind monotonically changes its direction from south to west-northwest (∼300°). In the slab characterized by the tropopause and stratosphere (above 11 km), the wind maintains its direction to roughly 300°. Above 20 km during the summer (more precisely, during December, January, and February), the wind changes its direction again to the south. This trend is in excellent agreement with that measured by Aristidi et al. (2005a) (Fig. 6).

2.3. Pressure

Atmospheric pressure is a quite stable parameter, showing small variations during the summer and winter above Antarctica. Figure 9 shows the median pressure during the summer and winter of 2003. In this image, we indicate the values that are associated with the typical troposphere-tropopause interface above midlatitude sites (200 mbar, corresponding to ∼11 km from sea level) and above Dome C (300-320 mbar, ∼8 km from sea level). As explained above, the interface between the troposphere and tropopause corresponds to an area that is prone to triggering optical turbulence.

2.4. Absolute and Potential Temperature

The absolute and potential temperature are fundamental elements defining the stability of the atmosphere. Figures 10 and 11 show the monthly median vertical profiles of the absolute and potential temperatures, respectively, during 2003 (green line) and 2004 (red line).

The value of ∂θ/∂z indicates the level of atmospheric thermal

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3 We note that in spite of the fact that the ECMWF data are not optimized for the surface layer, the difference in the absolute temperature at the first grid point above the ground between the summer and winter is well reconstructed by the GCMs (∼35° as measured by Aristidi et al. 2005a).
stability that is strictly correlated to the production of turbulence. When $\partial \theta / \partial z$ is positive, the atmosphere has high probabilities of being stratified and stable. We can see (Fig. 11) that this is observed in the ECMWF data set, and it is particularly evident during the wintertime. Another way to study the stability near the ground is to analyze $\partial T / \partial z$ (i.e., the gradient of the absolute temperature). When $\partial T / \partial z$ is positive, the atmosphere is hardly affected by advection, because the coldest regions of the atmosphere (the heaviest ones) are already in proximity to the ground. This condition is typical for Antarctica, due to the presence of ice on the surface, but it is expected to be much more evident during the winter, due to the extremely low temperature of the ice. We can see (Fig. 10) that during the wintertime, $\partial T / \partial z$ is definitely positive near the ground, indicating strongly stratified, stable conditions. All this indicates that some large wind speed gradient on a small vertical scale has to take place to trigger turbulence in the surface layer in wintertime. We discuss these results with respect to those obtained from measurements in § 2.5.

A further important feature for the vertical distribution of the absolute temperature is the inversion of the vertical gradient (from negative to positive) in the free atmosphere, indicating the troposphere-tropopause interface generally associated with an unstable region, due to the fact that $\partial \theta / \partial z \approx 0$. We can see that above Dome C, this inversion is located at around 8 km from sea level during all months. In the summertime, the me-

Fig. 3.—Summer (left column) and winter (right column) median wind speed vertical profile estimated in 2003 (top row) and 2004 (bottom row). The first and third quartiles are shown with a dotted line.

Fig. 4.—Seasonal trend of median wind speed estimated with ECMWF data above Dome C at 8 km (thin line) and 20 km (thick line). This seasonal trend shows the effect of the so-called polar vortex.
production of optical turbulence, and it can therefore be an indicator of the turbulence characteristics above a site. The atmosphere is defined as “stable” when \( R_i > 1 \), and it is “unstable” when \( R_i < 1 \). Typical conditions of instability can be set up when, in the same region, \( \partial V/\partial z \gg 1 \) and \( \partial \theta/\partial z < 1 \), or \( \partial \theta/\partial z \sim 0 \). Under these conditions, the turbulence is triggered in strongly stratified shears. These kinds of fluctuations in the atmosphere have typically small spatial scales and can be detected by radiosondes (balloons). When one works with meteorological parameters that are described at lower spatial resolutions, as in our case, it is not appropriate to use a deterministic Richardson number. Following a statistical approach (Vand Zandt et al. 1978), we can replace the deterministic \( R_i \) with a probability density function that describes the stability and instability factors in the atmosphere that are provided by meteorological data at larger spatial scales. This analysis has already been done by Masciadri & Garfias (2001). Figures 12 and 13 show the monthly gradient of the potential temperature \( \partial \theta/\partial z \) and the square of the gradient of the wind speed \((\partial V/\partial z)^2\). Finally, Figure 14 shows for each month the inverse of the Richardson number \((1/R)\) over 25 km. We show \( 1/R \) instead of \( R \) because the first can be displayed with a better dynamic range than the second. From a visual point of view, \( 1/R \) therefore allows us to better provide evidence of stability differences in different months. As explained above, with our data characterized by a low spatial resolution, we can analyze the atmospheric stability in relative terms (in space and time); i.e., identify regions that are less stable or more stable than others. This is quite useful if we want to compare features of the same region of the atmosphere in different periods of the year. The probability that turbulence develops is larger in regions characterized by a large \( 1/R \).

If, for example, we look at the \( 1/R \) distribution in the month of January (middle of the summer), we can see that a maximum is visible at around 2–5 km from the ground, corresponding to the height at which the gradient of the wind speed has a maximum and the gradient of the potential temperature \( \partial \theta/\partial z \sim 0 \) (Fig. 12). The presence of both conditions is a clear indicator of instability. In the same figure, \( 1/R \) decreases monotonically above 5 km, indicating conditions of general stability of the atmosphere in this region. If we compare the value of \( 1/R \) in different months, we can easily identify two periods of the year in which the Richardson number shows similar characteristics.

During the months of December–April, \( 1/R \) has a similar trend over the entire 25 km range. One or two peaks of \( 1/R \) are visible in the [2–5] km region, and a monotonic decrease above 5 km is observed.

During the months of May–November, \( 1/R \) shows more complex features. At [2–5] km from the ground, we find an instability that is similar to that identified in the summertime, but

\[ R_i = \frac{g}{\theta} \frac{\partial \theta/\partial z}{(\partial V/\partial z)^2}, \]

where \( g \) is the gravitational acceleration (9.8 m s\(^{-2}\)), \( \theta \) is the potential temperature, and \( V \) is the wind speed. The stability or instability of the atmosphere is tightly correlated with the

\[ R_i \]
above 5 km, we observe other regions of instability, mainly concentrated at 12 and 17 km from the ground. In a few cases (in September and October above 12 km from the ground), the probability that the turbulence is triggered can be larger than at [2–5] km. The analysis of $R$ (or $1/R$) does not give us the value of $C_n^2$ at a precise height, but it can give us a quite clear picture of “where” and “when” the turbulence has a high probability of being developed over the whole year above Dome C.

Summarizing, we can state that during the whole year, we have conditions of instability in the range of [2–5] km from the ground. We can predict the development of turbulence, which is probably characterized by a strength that is inferior to that observed above midlatitude sites. The wind speed at [2–5] km above Dome C is indeed clearly weaker than the wind speed at the same height above midlatitude sites.

In the high part of the atmosphere ($h > 5$ km) during the summertime, the atmosphere is generally quite stable, and we should expect a low level of turbulence. During the wintertime, the atmosphere is more unstable, and we should expect a higher level of turbulence than during the summer. The optical turbulence above 10 km will be monitored carefully in the future during the months of September and October, to be sure that $\tau_0$ is competitive with midlatitude sites in the wintertime. Indeed, even a weak $C_n^2$ combined with a high wind speed at these altitudes might induce an important decrease of $\tau_0$ with respect to $\tau_0$ found above midlatitude sites. Indeed, as can be seen in Figure 6, the wind speed at this height can be quite strong. On the other hand, we stress that this period does not coincide with the middle part of the winter (June, July, and August), which is the most interesting time for astronomic observations.

We again stress the fact that in this paper, we are not providing the absolute value of the turbulence, but are instead comparing levels of instability in different regions of the atmosphere and in different periods of the year. This state of stability and instability is estimated first from meteorological parameters retrieved from an ECMWF data set. Considering that, as we have proved once more, the meteorological parameters are quite well described by the ECMWF data, the relative status of stability and instability of the atmosphere that is represented by the Richardson number maps provided in our paper is a constraint against which measurements of the optical turbulence need to be compared. We expect that $C_n^2$ measurements agree

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**Fig. 6.**—Seasonal median wind speed vertical profiles for the years 2003 (green lines) and 2004 (red lines).
Fig. 7.—Seasonal cumulative distribution of the wind speed in the 8–9 km range from sea level. This roughly corresponds to the tropopause height above Dome C. The pressure at this altitude is around 320 mbar.

with the stability and instability properties indicated by the Richardson maps. What is the typical seeing above the first 30 m? We should expect that the strength of the turbulence in the free atmosphere is greater in wintertime than during the summer. Are the measurements that have been done so far in agreement with the Richardson maps describing the stability and instability of the atmosphere in different seasons and at different heights?

Some site-testing campaigns have been organized above Dome C (Aristidi et al. 2005a, 2005b; Lawrence et al. 2004) that employ different instruments in different periods of the year. We require measurements provided by a vertical profiler in order to analyze seeing values above 30 m. Balloons measuring vertical $C_\beta$ profiles have been launched during the wintertime (Agabi et al. 2006). Preliminary results indicate a seeing of 0'36 above the first 30 m. Unfortunately, no measurements of the $C_\beta$ vertical distribution during the summertime are available. However, we can retrieve information on the level of activity of the turbulence in the high part of the atmosphere by analyzing the isoplanatic angle. This parameter is indeed particularly sensitive to the turbulence that develops in the high part of the atmosphere. We currently know that the median $\theta_i$ measured with a generalized seeing monitor (GSM) is 6'8 during the summer and 2'7 during the winter.\footnote{We note that some discrepancies were found between $\theta_i$ measured by a GSM (2'7) and balloons (4'7) in the same period (Aristidi et al. 2005b). This should be analyzed more in detail in the future. However, in the context of our discussion, we are interested in a relative estimate (i.e., on a parameter variation between summer and winter). We therefore consider values measured by the same instrument (GSM) in summer and in winter.} This means that during the winter, the level of the turbulence in the free atmosphere is higher than in summer. This matches perfectly with the estimates obtained in our analysis.

On the other hand, a differential image motion monitor (DIMM) placed 8.5 m from the ground measured a median
Fig. 8.—Seasonal median wind direction vertical profiles for the years 2003 (green lines) and 2004 (red lines). Zero degrees corresponds to north.

Fig. 9.—Atmospheric pressure in winter (solid line) and summer (dotted line) during 2003. The figure shows the typical 300–320 mbar pressure associated with the 8 km latitude, and the 190–200 mbar pressure associated with the 11 km altitude.

value of seeing $e_{\text{tot}} = 0^\circ.55$ in the summer (Aristidi et al. 2005b) and $e_{\text{tot}} = 1^\circ.3$ in the winter (Agabi et al. 2006). This instrument measures the integral of the turbulence over the whole troposphere and stratosphere. The large difference of the seeing between the winter and summer is certainly due to a general increase in the turbulence strength near the ground in the summer–winter passage.6 Indeed, measurements of the seeing above 30 m obtained with balloons during the winter (Agabi et al. 2006) give a typical value of $e_{(30 \text{ m}, w)} = 0^\circ.36$. Using the law

$$e_{(0,30 \text{ m})} = (e_{\text{tot}}^{5/3} - e_{(30 \text{ m}, w)}^{5/3})^{3/5},$$

we calculate that during the winter, the median seeing in the first 30 m is equal to $e_{(0,30 \text{ m}), \text{winter}} = 1^\circ.2$. In spite of the fact that we have no measurements of the seeing above 30 m in the summer, we know from the Richardson analysis given in this paper that the seeing in this region of the atmosphere should be weaker in summer than in winter. This means that the seeing

6 This does not mean that one can observe high values of $e$ in some periods of the day in the summertime, as shown by Aristidi et al. (2005b).
above 30 m in the summer should be smaller than 0.36. Knowing that the total seeing in the summer is equal to $S_{tot} = 0.36$, one can deduce that the seeing in the first 30 m should be smaller than 0.55. This means that $S_{(0, 30m), \text{summer}} < 0.55 < S_{(0, 30m), \text{winter}} = 1.2$ and that the turbulence strength of the surface layer is greater during the winter than during the summer. In § 2.4, we noted that during the winter and near the ground, the thermal stability is greater than during the summer. This is what the physics says and the ECMWF data set shows, but it is in contradiction to the seeing measurements. The only way to explain such a strong turbulent layer near the ground during the winter is to assume that the wind speed gradient in the first 30 m is greater during the winter than during the summer. This is difficult to accept if the wind speed is weaker during the winter than during the summer, as stated by Aristidi et al. (2005a). As shown by Masciadri (2003), the weaker the wind speed near the surface, the weaker the gradient of the wind speed. We therefore suggest that a more detailed analysis of this parameter be made near the surface and extended over the whole year. Preferably, this should be done with anemometers mounted on masts or kites. This will also permit the calculation of the Richardson number in the first 30 m during the whole year, as well as observations of differences between the summer and winter. This can certainly be a useful calculation to validate the turbulence measurements. The ECMWF data set does not have the necessary reliability in the surface layer to prove or disprove these measurements.

3. RELIABILITY OF ECMWF DATA

As explained above, measurements obtained recently above Dome C with radiosondes (Aristidi et al. 2005a) can be useful to quantify the level of reliability of our estimates. Aristidi et al. (2005a) provide the median vertical profile of the wind speed measured on several nights during the summer (see their Fig. 4). Figure 1 of Aristidi et al. (2005a) gives a histogram of the time distribution of measurements as a function of month. Most of the measurements were done during the December and January months. Figure 15 (this paper) shows the vertical profile of the wind speed obtained with ECMWF data for the December and January months in 2003 and 2004 (thick line) and the measurements obtained during the same months above
Dome C (thin solid line). E. Aristidi (member of the Laboratoire Universitaire d’Astrophysique de Nice [LUAN] team) kindly selected for us only the measurements related to these 2 months in their sample. We note that the ECMWF data are all calculated at 00:00 UT, although the balloons were not launched at the same hour each day. Moreover, the measurements are related to the 2000–2003 period, while the analyses are related to the 2003–2004 period. In spite of this difference, the two mean vertical profiles show an excellent correlation. The absolute difference remains below 1 m s$^{-1}$, with a mean difference of 0.7 m s$^{-1}$ basically everywhere.

In the high part of the atmosphere (Fig. 15), the discrepancy between measurements and ECMWF analyses is on the order of 1.5 m s$^{-1}$. This is a very small absolute discrepancy, but considering the typical wind speed value of $\sim$4 m s$^{-1}$ at this height, it gives a relative discrepancy on the order of 25%. We calculate that, assuming measurements of the seeing so far measured above Dome C and $C_2^2$ profiles as shown in § 4 (Table 2), this might induce discrepancies in the $T_i$ estimates on the order of 13%–16%. To produce a more detailed study of the accuracy of the ECMWF analyses and measurements, one should know the intrinsic error of measurements and the scale of spatial fluctuations of the wind speed at this height. No further analysis is possible for us above Dome C to improve the homogeneity of the samples (measurements and analyses) and better quantify the correlation between them, because we do not access the raw measurement data. We therefore decided to compare measurements with ECMWF analyses above South Pole in the summer as well as in the winter in order to provide further evidence of the level of reliability of ECMWF analyses above a remote site such as Antarctica. Figure 16 (January [sumertime], 12 nights) and Figure 17 (June and July [wintertime], 12 nights) show the median vertical profiles of wind speed, wind direction, and absolute temperature provided by measurements and ECMWF analyses. We emphasize that in order to test the reliability of ECMWF analyses, we considered all (and only) nights for which measurements are available in the whole 25 km range for the three parameters: wind speed, wind direction, and absolute temperature. It was observed that during the winter, the number of radiosondes providing a complete set of measurements decreases. In this season, measurements were frequently obtained in only the first 10–12 km. Above this height, the balloons burst. To increase the statistics of the

\footnote{See ftp://amrc.ssec.wisc.edu/pub/southpole/radiosonde.}

![](image)
Fig. 12.—Seasonal median $d\theta/dh$ vertical profile calculated with ECMWF analyses of 2003 and 2004 data.

set of measurements extended over the whole 25 km, we decided to take into account nights related to 2 months (June and July) in winter. We can see (Figs. 16 and 17) that the correlation between ECMWF analyses and the measurements is quite good in winter and in the summer for all the three meteorological parameters. We expressly did not smooth the fluctuations characterized by high frequencies of measurements. The discrepancy between measurements and ECMWF analyses is smaller than 1 ms$^{-1}$ in the entire troposphere. It is also apparent that at small scales, the natural typical fluctuations of the measured wind speed is $\sim$1 ms$^{-1}$. We conclude, therefore, that a correlation between measurements and ECMWF analyses within an error of 1 m s$^{-1}$ is a very good correlation, and that this data set can provide reliable initialization data for meso-scale models.

As a further result of this study, we see that during the wintertime, the wind speed above South Pole is weaker than above Dome C, particularly above 8 km from the ground. This fact certainly affects the value of $\tau_0$, placing South Pole in a more favorable position with respect to South Pole. Further measurements are necessary in order to identify which of these two elements (a higher wind speed at high altitudes above Dome C, or a stronger turbulence surface layer above South Pole) affects $\tau_0$ more. Indeed, while typical values of $\tau_0$ (1.58 ms) in wintertime (June, July, and August) above South Pole are already available (Marks et al. 1999), we do not yet have measurements of $\tau_0$ above Dome C that are related to the same period. Of course, if $\tau_0$ above Dome C are revealed to be greater than 1.58 ms, this would mean that the stronger turbulence layer in the surface above South Pole affects $\tau_0$ more than the higher wind speed at high altitudes above Dome C. This study is critical to defining the potential of these sites for future applications of interferometry and adaptive optics.

4. DISCUSSION

We intend here to calculate the value of $\theta_0$ and $\tau_0$ in the slab of atmosphere in the range $[h_{\text{surf}}, h_{\text{top}}]$, using as inputs simple analytical models of optical turbulence $C_n^2$ and the median vertical profiles of the wind speed shown in Figure 3. The superior
limit \(h_{\text{top}}\) is defined by the maximum altitude at which balloons provide measurements before exploding. The inferior limit \(h_{\text{surface}}\) corresponds to the expected surface layer above Dome C. We define \(h_{\text{surface}} = 30 \, \text{m}\) and \(h_{\text{ground}} = 3229 \, \text{m}\) for the Dome C ground altitude. We consider independent models with \(h_{\text{top}} = 25 \, \text{km}\) and \(h_{\text{top}} = 20 \, \text{km}\). Our analysis intends to estimate typical values of some critical astroclimatic parameters \((\theta_0, \tau_0)\) without the contribution of the first 30 m above the icy surface. The wave-front coherence time \(\tau_0\) is defined in equation (1), and the isoplanatic angle \(\theta_0\) is defined as

\[
\theta_0 = 0.049 \left( \int h^{5/3} C_N^2(h) \, dh \right)^{-3/5}.
\]

(4)

Tables 1 and 2 summarize the inputs and outputs of these estimates.

**Models A–F:** The simplest (and less realistic) assumption is to consider the \(C_N^2\) constant over the \([h_{\text{surface}}, h_{\text{top}}]\) range. To calculate the \(C_N^2\), we use three values of reference: \(\varepsilon = 0.27\), \(\varepsilon = 0.2\), and \(\varepsilon = 0.1\). We make the assumption that \(C_N^2\) is uniformly distributed in \(\Delta h\), where \(\Delta h = h_{\text{top}} - h_{\text{ground}} - h_{\text{surface}}\). We then calculate \(C_N^2\) as

\[
C_N^2 = \frac{1}{\Delta h} \left( \frac{\varepsilon}{19.96 \times 10^6} \right)^{5/3}.
\]

(5)

The median vertical profiles of wind speed during the summertime in the years 2003 and 2004 (see Fig. 3) are used for the calculation of \(\tau_0\).

**Models G–N:** As discussed above, the turbulence above Dome C would probably be triggered at around [2–5] km from the ground during the summertime. A more realistic but still simple model therefore consists of taking a thin layer with a thickness of \(h_{\text{top}} = 25 \, \text{km}\) and \(h_{\text{top}} = 20 \, \text{km}\). Our analysis intends to estimate typical values of some critical astroclimatic parameters \((\theta_0, \tau_0)\) without the contribution of the first 30 m above the icy surface. The wave-front coherence time \(\tau_0\) is defined in equation (1), and the isoplanatic angle \(\theta_0\) is defined as

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\]

(5)
assume a realistic $C_n^2$ model for the winter season, and we therefore limit our analysis to the summer season. To calculate the best values of $\theta_0$ and $\tau_0$ that can be reached above Dome C, we consider the realistic minimum values of $C_{n,1}^2 = 10^{-19}$ m$^{-2/3}$ (Marks et al. 1999) given by the “atmospheric noise,” and we calculate the value of the $C_{n,2}$ in the thin layer at 5 km, using equation (5) and the relation

$$C_{n}^2 \Delta h = C_{n,1}^2 \Delta h_1 + C_{n,2}^2 \Delta h_2.$$  (6)

Aristidi et al. (2005c) measured an isoplanatic angle $\theta_0 = 6^\circ 8$ in the summertime. Looking at Table 1, models A–F, we deduce that such a uniform distribution of $C_n^2$ could match these values ($\theta_0 = 6^\circ 8$) only if it was associated with an exceptional seeing of 0.1. In this case, we should expect $\tau_0$ on the order of 30–40 ms. Alternatively, under the assumption of $C_n^2$ that peaks at 8 km from sea level (Table 1, models G–N), a seeing of 0.2 would better match with $\theta_0 = 6^\circ 8$. In this case, we should expect $\tau_0$ on the order of 13–16 ms. Summarizing, we can expect one of the following data sets: ($\varepsilon = 0^\circ 1, \theta_0 = 6^\circ 8, \tau_0 = 30–40$ ms) or ($\varepsilon = 0^\circ 2, \theta_0 = 6^\circ 8, \tau_0 = 13–16$ ms). The second one is much more realistic.

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8 We recall that for values of $C_n^2$ that are smaller than $10^{-19}$ m$^{-2/3}$, we enter into the regime of electronic noise (see Azouit & Vernin 2005; Masciadri & Jabouille 2001).
It is interesting to note that $\tau_o$ can be quite different if one assumes a seeing that is slightly different (0.1"–0.2") under the hypothesis of a distribution of $C_N^2$ as described in this paper. We deduce from this analysis (combined with the discussion in § 2.5) that the seeing above 30 m during the summertime is probably on the order of 0.2" or even smaller. This means that in the free atmosphere, the seeing should be weaker during the summertime than during the winter (average $e = 0.36$; Agabi et al. 2006). This result closely matches our Richardson number maps. However, it would be interesting to measure the seeing in the free atmosphere during the summer in order to better constrain the values of $\tau_o$. This is not evident, due to the fact that radiosondes that have been used to measure the $C_N^2$ cannot be used to measure this parameter during the summer. Measurements are not reliable, due to false temperature fluctuations experienced by the thermic probes in this season (E. Aristidi 2006, private communication). From this simple analysis, we deduce reasonable values of $\theta_{0,\max} \sim 10^\circ$–11° and $\tau_{0,\max} \sim 16$ ms during the summertime under the best atmospheric conditions and with the most realistic distribution of $C_N^2$ in the atmosphere. We remind the reader that some measurements of $\tau_o$ have already been published (Lawrence et al. 2004). Such measurements have been done only in the transition between summer and winter (April–May). Our simple $C_N^2$ model is not set up to compare $\tau_o$ and $\theta_e$ estimates such as those done in this section with those measured by Lawrence et al. (2004). More detailed information on the $C_N^2$ measurements in wintertime will allow future studies to verify measurements done by Lawrence et al. (2004).

5. CONCLUSION

In this paper, we have presented a complete study of the vertical distribution of all the main meteorological parameters (wind speed and direction, pressure, and absolute and potential temperature) that characterize the atmosphere above Dome C, from a few meters from the ground up to 25 km. This study employs the ECMWF “analyses” obtained by general circulation models, which extend over 2 years, 2003 and 2004, and provide a statistical analysis of all the meteorological parameters and the Richardson number in each month of the year. This latter parameter provides us with useful insights into the probability that optical turbulence can be triggered in different regions of the atmosphere and in different periods of the year. The Richardson number indeed monitors the conditions of stability and instability in the atmosphere from a dynamic as well as a statistical perspective.
as thermal point of view. The main results obtained in our study are as follows.

1. The wind speed vertical distribution shows two different trends in the summer and winter due to the "polar vortex" circulation. In the first 8 km above the ground, the wind speed is extremely weak during the whole year. The median value at 5 km, corresponding to the peak of the profile placed at the interface between the troposphere and tropopause, is 12 m s\(^{-1}\). At this height, the third quartile of the wind speed is never higher than 20 m s\(^{-1}\). Above 5 km, the wind speed remains extremely weak (the median value is smaller than 10 m s\(^{-1}\)) during the summertime. During the wintertime, the wind speed increases monotonically with height and at an important rate, reaching median values on the order of 30 m s\(^{-1}\) at 25 km. A fluctuation on the order of 20 m s\(^{-1}\) is estimated at 20 km between the summer and winter.

2. The atmosphere above Dome C shows a quite different regime of stability and instability in the summer and winter. During the summer, the Richardson number indicates a general regime of stability in the whole atmosphere. The turbulence can be triggered with higher probability at [2–5] km from the

### TABLE 1

<table>
<thead>
<tr>
<th>Model</th>
<th>(h_{\text{top}}) (km)</th>
<th>(\epsilon) (arcsec)</th>
<th>(C_{n}^2) (m(^{-2}))</th>
<th>(\theta_0) (arcsec)</th>
<th>(\tau_0) (ms)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Model A</td>
<td>25</td>
<td>0.27</td>
<td>3.53 \times 10^{-18}</td>
<td>1.95</td>
<td>14.00</td>
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<tr>
<td>Model B</td>
<td>25</td>
<td>0.2</td>
<td>2.14 \times 10^{-18}</td>
<td>2.63</td>
<td>18.91</td>
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<tr>
<td>Model C</td>
<td>25</td>
<td>0.1</td>
<td>6.74 \times 10^{-19}</td>
<td>5.26</td>
<td>37.83</td>
</tr>
<tr>
<td>Model D</td>
<td>20</td>
<td>0.27</td>
<td>4.58 \times 10^{-18}</td>
<td>2.53</td>
<td>13.52</td>
</tr>
<tr>
<td>Model E</td>
<td>20</td>
<td>0.2</td>
<td>2.78 \times 10^{-18}</td>
<td>3.41</td>
<td>18.25</td>
</tr>
<tr>
<td>Model F</td>
<td>20</td>
<td>0.1</td>
<td>8.76 \times 10^{-19}</td>
<td>6.82</td>
<td>36.49</td>
</tr>
</tbody>
</table>

### TABLE 2

<table>
<thead>
<tr>
<th>Model</th>
<th>(h_{\text{top}}) (km)</th>
<th>(\epsilon) (arcsec)</th>
<th>(C_{n}^2) (m(^{-2}))</th>
<th>(\theta_0) (arcsec)</th>
<th>(\tau_0) (ms)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Model G</td>
<td>25</td>
<td>0.27</td>
<td>7.46 \times 10^{-16}</td>
<td>4.60</td>
<td>10.17</td>
</tr>
<tr>
<td>Model H</td>
<td>25</td>
<td>0.2</td>
<td>4.40 \times 10^{-16}</td>
<td>6.03</td>
<td>13.87</td>
</tr>
<tr>
<td>Model I</td>
<td>25</td>
<td>0.1</td>
<td>1.25 \times 10^{-16}</td>
<td>10.18</td>
<td>28.34</td>
</tr>
<tr>
<td>Model L</td>
<td>20</td>
<td>0.27</td>
<td>7.51 \times 10^{-16}</td>
<td>4.73</td>
<td>10.15</td>
</tr>
<tr>
<td>Model M</td>
<td>20</td>
<td>0.2</td>
<td>4.49 \times 10^{-16}</td>
<td>6.32</td>
<td>13.75</td>
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<tr>
<td>Model N</td>
<td>20</td>
<td>0.1</td>
<td>1.30 \times 10^{-16}</td>
<td>11.65</td>
<td>28.02</td>
</tr>
</tbody>
</table>

Note.—\(C_{n}^2 = 10^{-18}\) m\(^{-2}\) in all the models.
ground. During the winter, the atmosphere shows a more important turbulent activity. In spite of the fact that the analysis of the Richardson number in different months of the year is qualitative, our predictions are consistent with preliminary measurements obtained above the site in any particular period of the year. Considering the good reliability of the meteorological parameters retrieved from the ECMWF analyses, the Richardson maps shown here should be considered as a reference to check the consistency of further measurements of optical turbulence in the future.

3. With the aid of a simple model for the $C_n^2$ distribution, the Richardson number maps, and the wind speed vertical profile, we calculated a best isoplanatic angle of $\theta_{\text{iso}, \text{max}} \approx 10^{\circ}$–$11^{\circ}$ and coherence time of $\tau_{\text{iso}, \text{max}} \approx 16$ ms above Dome C during the summertime.

4. The vertical distribution of all the meteorological parameters shows a good agreement with measurements. This result is quite promising for employing ECMWF analyses as initialization data for meso-scale models. Furthermore, it opens up perspectives on how to employ ECMWF data for the characterization of meteorological parameters extended over long timescales.

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