MEAN VERTICAL PROFILES OF TEMPERATURE AND ABSOLUTE HUMIDITY FROM A TWELVE-YEAR RADIOSOUNDING DATA-SET AT TERRA NOVA BAY (ANTARCTICA)

by

Claudio TOMASI [(1)], Alessandra CACCIARI [(2)], Vito VITALE [(1)], Angelo LUPI [(1)], Christian LANCONELLI [(1)], Andrea PELLEGRINI [(3)] and Paolo GRIGIONI [(4)]

[(1)] Institute of Atmospheric and Climatic Sciences (ISAC), Consiglio Nazionale delle Ricerche (CNR), via Gobetti 101, I-40129 Bologna (Italy)
[(2)] Carlo Gavazzi Space S. p. A. - ISAC-CNR, via Gobetti 101, I-40129 Bologna (Italy)
[(3)] ENEA – Antarctic Project, C. R. Casaccia, via Anguillarese 301, I-00060 S. Maria di Galeria , Rome (Italy)
[(4)] ENEA Climate Section, C. R. Casaccia, via Anguillarese 301, I-00060 S. Maria di Galeria , Rome (Italy)

ABSTRACT

A set of 1330 radiosoundings were performed at the Italian station of Terra Nova Bay in Antarctica during twelve measurement campaigns from 1987 to 1998. These measurements were analysed separately for the various ten-day periods from mid-October to mid-February, to determine the mean vertical profiles of air pressure and temperature in the troposphere and lower stratosphere, and those of moisture parameters in the troposphere. The temperature data were corrected for the errors due to radiation and heat exchange processes and for the lag errors of the sensor. Due to temperature dependence and other dry bias effects, the humidity errors were also taken into account. The tropospheric temperature was found to present average values of its vertical gradient varying between −5.4 and −6.3 °K/km, while its minimum height associated with the tropopause gradually lowered from 13.7 to 7.9 km. Total water vapour content increased correspondingly from 0.12 to 0.36 g cm^{-2} during the first two months and decreased to 0.26 g cm^{-2} in the following period. Stratospheric temperature was observed to increase during the first two-months by about 15 °K in the region below the 24 km height and to appreciably decrease at upper levels, maintaining almost stable features during the subsequent months.

Keywords: Antarctic atmosphere, Radiosounding data analysis, Temperature vertical profiles, Absolute humidity in the troposphere, Moisture conditions in the low stratosphere.

1.- INTRODUCTION

The atmospheric spectrum of incoming solar radiation presents a great number of water vapour absorption bands in the visible and near-infrared wavelength range, with intensity features generally increasing with wavelength. As shown by Goody (1964), six relatively weak absorption bands, all due to ground-state transitions, are located within the spectral interval from 0.543 to 0.652 μm. Another eight bands (commonly identified by Greek letters) appear at longer wavelengths from 0.67 to 2.08 μm, all presenting strong absorption features (Kondratyev, 1969). Some of them exhibit absorption characteristics of so marked an intensity in their middle part as to cause extensive spectral regions of complete absorption in the solar radiation spectrum measured at sea-level. Another group of strong water vapour bands, forming the so-called χ band, covers the spectral range from 2.27 to 3.57 μm, producing total absorption of incoming solar radiation within the 2.52–2.85 μm range, also in cases where the water vapour mass W distributed along the atmospheric sun-path is equal to a few tenths of g cm^{-2}. As a result of such strong absorption by water vapour, the solar irradiance turns out to be considerably attenuated during its passage through the atmosphere. Our calculations of the global solar radiation flux reaching the ground, performed with the 6S computer code for the Subarctic Summer model (Vermote et al., 1997), indicate that about 12% of the incoming solar radiation is absorbed by W = 2.1 g cm^{-2}, 15% by W = 4.2 g cm^{-2}, 17% by W = 6.3 g cm^{-2}, and 20% by W = 10.5 g cm^{-2}. Therefore, atmospheric water vapour plays an important role in the solar radiation transfer processes occurring in the Antarctic atmosphere, even if it presents generally low columnar content values.

Moreover, water vapour very strongly absorbs the thermal radiation emitted upward by both terrestrial surface and atmosphere. In fact, the absorption of infrared radiation by atmospheric water vapour is simultaneously produced by two different processes: (i) selective absorption mainly due to the strong vibro-rotational band ν_2 occupying the wavelength range from 4.88 to 8.70 µm and a numerous group of rotational bands causing very strong absorption features beyond 16 µm, as pointed out by Goody (1964), and (ii) continuum absorption, due to both foreign- and self-broadening of the absorption lines (Bignell, 1970). Consequently, absorption and emission processes of thermal radiation occurring in the atmosphere are strongly influenced by the presence of water vapour, to such an extent as to modify considerably both the upwelling flux of infrared radiation emitted toward space and the downwelling flux of infrared radiation emitted by the atmosphere toward the terrestrial surface.

Atmospheric water vapour absorption of long-wave radiation strongly depends on air temperature, air pressure and water vapour partial pressure. Its radiative effects on the thermal radiation balance of the atmosphere cannot be neglected in the Antarctic regions, even though the atmospheric columnar content of water vapour is in general appreciably lower here than in other areas of our planet. Precise calculations of the mean short-wave and long-wave terms giving form to the radiation balance of the atmosphere are very useful in Antarctic climate studies, requiring the knowledge of the mean vertical distribution curves of air pressure, air temperature and absolute humidity in the troposphere and lower stratosphere. Moreover, precise evaluations of solar radiation extinction due to Rayleigh scattering and gaseous absorption can be obtained at the visible and infrared wavelengths, only if the main average vertical distribution features of temperature and moisture parameters are known with accuracy. Other remote sensing problems can be more thoroughly investigated by taking into account the water vapour absorption effects in various correction procedures of field data, such as those given by backscattering measurements carried out with lidar techniques or those from cloud images given by microwave radiometers mounted aboard satellites.

Considering the above remarks, we decided to examine a twelve-year data-set of radiosounding measurements taken at the Italian base of Terra Nova Bay (74° 42' S; 164° 07' E) in Antarctica during the period from January 1987 to February 1998. Among the 16 WMO upper air stations in Antarctica, Terra Nova Bay is one of the two stations in the Ross Sea area, where radiosoundings are taken regularly during the local summer months, the other being the US station of McMurdo (77° 31’ S; 166° 24’ E). For this reason, the present data provide useful information on the thermodynamic conditions of the Antarctic atmosphere in the Ross Sea coastal region. The primary aim of the present study is to determine the mean vertical profiles of pressure, temperature and absolute humidity in the lower part of the atmosphere through the analysis of the radiosounding measurements taken at Terra Nova Bay over various ten-day periods from the end of October to mid-February, with a resolution in height suitable for radiative transfer calculations. Considering that precise values of both air pressure and temperature can be obtained from these radiosounding data-sets up to the stratospheric height of 32 km, while humidity measurements appear to be fully reliable at tropospheric levels only, we decided to complete the definition of such atmospheric models within the low stratosphere, by analysing a set of mixing ratio measurements available in the literature, performed by means of radiometric techniques taken from satellites and balloons, with the main purpose of determining the vertical profiles of the moisture parameters in the stratosphere, from the tropopause region up to the 32 km level.

2.- THE RADIOSOUNDING MEASUREMENTS WITH FACTORY CORRECTIONS

An overall number of 1330 radiosounding measurements were taken during the twelve campaigns held at the Terra Nova Bay station (55 m a.m.s.l.) by the meteorologist group of the Antarctic Project (ENEA, C. R. Casaccia, Rome) using the RS-80-A radiosondes manufactured by Vaisala (Helsinki, Finland). The measurements were provided by the following three sensors:

(i) air pressure p was measured using a capacitive aneroid (called BAROCAP) with a measurement range from 1060 to 3 hPa, resolution of 0.1 hPa and accuracy of ±0.5 hPa;

(ii) air temperature T was measured using the THERMOCAP sensor, which is a small capacitive bead in glass encapsulation, with a measurement range from 333 to 183 °K, resolution of 0.1 °K and accuracy (standard deviation) of ±0.2 °K; and

(iii) air relative humidity f was measured using a capacitive thin film humidity sensor, called HUMICAP, model A, with a sensitivity range from 2% to 100%, as declared by the manufacturer, resolution of 1%, declared accuracy of less than ± 3% and calibration repeatability of ± 2%.

During the ascent, the measurements of parameters p, T and f were sent by the transmitter to the receiver at the ground station using the nominal frequency of 403 MHz. The three sensor signals were transmitted to the ground station at the sampling rate of 2 samples every 10 seconds during the first campaign in 1987, 5 samples every 10 seconds during the second campaign in 1988 and one sample only every 10 seconds in the subsequent ten campaigns. A radiosonde usually ascends at a rate of 5÷6 m/s: consequently, the significant levels of each radiosounding were found to be distributed along the vertical path in steps of 25÷30 m in 1987, 10÷12 m in 1988 and 50÷60 m in the other campaigns, starting from the ground-altitude of 55 m above mean sea-level.

For each triplet of meteorological parameters p, T and f, the corresponding height z was calculated from the values of air pressure p and virtual temperature T^* by integrating, step by step from one level to the subsequent one, the differential term

$$dz = -(R\,T^*/g)\,d(\ln p) \qquad (1)$$

given by the well-known hydrostatic equation, where R is the gas constant and g is the gravitational acceleration, provided that the initial values of pressure and height are known with good precision, as recommended by Lally (1985). Thus, the error in altitude depends mainly on the error in virtual temperature T^*, which is related to the air temperature T through a linear relationship with slope coefficient proportional to the ratio of water vapour partial pressure e to total air pressure p. This implies that the errors made in the Antarctic atmosphere (where relatively dry air conditions are generally observed) are expected to be of no more than 5 m at 5 km height, 10 m at 10 km altitude and about 20 m at 20 km, in all cases where the systematic error in T^* does not exceed 0.2 °K.

As pointed out by Luers and Eskridge (1995), important errors can also be made in determining the atmospheric temperature, due to contamination by heating from sources other than the air itself (i.e. solar and infrared irradiation of the sensor, heat conduction to the sensor from its attachment points, and radiation emitted by the sensor). They first developed a temperature correction model for the Vaisala RS-80 sonde, taking into account all significant environmental processes that can influence heat transfer to the sensor, and then validated the model by comparing the corrected temperature profiles obtained for the RS-80 sonde with those derived from the NASA multithermistor radiosonde. Potential reference radiosonde (PREFRS) tests were conducted in Crawley, England, throughout February and March 1992, by means of balloons carrying five different types of radiosondes, during night and day balloon flights, for both clear and overcast sky conditions, and for solar elevation angles varying between 13° and 35°. The results showed that (i) solar and infrared irradiation causes an appreciable heat gain, (ii) convection produces a heat loss of comparable magnitude, and (iii) emission and conduction processes yield less important cooling and heating effects, respectively. Thus, the overall correction was estimated by them to be of no more than 0.5 °K at altitudes below 10 km, and to slowly increase from about 1 to 2 °K on the average, as the height increases from 10 to 30 km, in close agreement with the factory correction adopted by Vaisala.

Therefore, the above THERMOCAP data were corrected following the factory Vaisala RS-80 correction procedure, which appears suitable also for the Terra Nova Bay radiosounding data, since the solar elevation angle was observed not to exceed 39° during the period from October 20 to February 20 of each year at this Antarctic station, covering a range very similar to that of the PREFRS campaign. From these calculations and corrections performed for each radiosounding measurement, we obtained the preliminary vertical profiles of *p*, *T* and *f*. Depending on the sampling rate of the transducer mounted on the radiosonde and the vertical ascent velocity of the radiosonde balloon, each vertical profile of temperature was found to include at least 130 significant levels in the troposphere, a comparable number of levels from the tropopause level (mainly situated at altitudes ranging between 8 and 11 km) to the 15 km height, and many other levels within the upper stratospheric region up to the highest altitude reached by the radiosonde. The top-levels of the radiosondes generally varied between 15 and 25 km, but in many cases were higher than 30 km. When the radiosonde passes from the troposphere to the stratosphere, the air relative humidity decreases very sharply to values of only a few percent and, then, very often lower than the repeatability level of 2% established by the manufacturers. Moreover, the radiosonde measurements of relative humidity are expected to be often unreliable at cold temperatures and, hence, also in the high troposphere. In order to improve the reliability of our radiosounding data, we examined all the above measurements following some correction procedures proposed for reducing the lag errors of both THERMOCAP and HUMICAP-A sensors. Other additional criteria were adopted in order to reduce the errors caused by occasional freezing episodes of the HUMICAP-A sensor, taking place during the crossing of clouds and/or in the presence of very low temperature conditions at high tropospheric levels.

3.- FURTHER CORRECTIONS FOR SENSOR ERRORS

As pointed out by Huovila and Tuominen (1991), the air temperature measurements given by the capacitive bead THERMOCAP and those of relative humidity provided by the HUMICAP-A capacitive sensor are usually affected by important lag errors. The lag effects on the temperature measurements substantially depend on both air density (and, hence, total air pressure *p*) and ventilation speed (related to the ascent rate of the radiosonde balloon). These errors in height commonly vary between 10 and about 30 m at tropospheric levels, the thermometer lag being closely related to the total air pressure *p*. Because of such lag errors, the mean values of air temperature directly determined from the radiosounding measurements are generally evaluated to be (i) underestimated by no more than 0.1 °K at all the tropospheric levels, (ii) substantially accurate at all levels from the tropopause to 15 km height, and (iii) slightly overestimated, by no more than 0.1 °K, at the upper stratospheric levels. Since the evaluations of air temperature were obtained with standard deviations of no more than ±0.2

°K, the systematic errors due to lag effects turn out to be relatively unimportant at all levels. However, on the basis of both theoretical remarks and field measurements carried out with the RS-80 radiosondes, Huovila and Tuominen (1991) defined an empirical series of nine values of the thermometer time-lag coefficient α_1, gradually decreasing as a function of total air pressure p (measured in hPa) throughout the range $10 \le p \le 1000$ hPa. Examining these values in terms of the analytical form,

$$\alpha_1 = a_1 \; p^{-b_1} \qquad , \qquad (2)$$

we determined the best-fit values a_1 = 35.15 s and b_1 = 0.3877, with regression coefficient equal to −0.999. Using these best-fit values in eq. (2), we applied the procedure proposed by Vitale and Tomasi (1994) and calculated the true measurement-times at all the significant levels of each radiosounding, which were then assumed to correspond to the values of T given by the THERMOCAP sensor. Correspondingly, we calculated the values of virtual temperature T^* at all significant levels and determined the values of height z according to eq. (1).

Huovila and Tuominen (1991) also found that significant errors can frequently be caused by the time-lag on the air humidity measurements, due to temperature, relative humidity and ventilation effects. They estimated that the hygrometer lag coefficient α_2 depends on temperature, humidity and ventilation, and varies as a function of the air temperature T, following an approximately exponential dependence form characterized by an average slope coefficient $d(\ln \alpha_2)/dT = -0.08$ $(°K)^{-1}$ throughout the range of T from 223 to 293 °K. A more thorough examination of such errors was conducted by Miloshevich *et al*. (2001) in order to improve the analysis of HUMICAP-A measurements taken at cold temperatures. They made use of both (i) a statistical analysis of simultaneous measurements taken with Vaisala radiosondes and NOAA hygrometers, and (ii) laboratory measurements performed at Vaisala. On examining the latter data-set, they found that the relative humidity evaluations were generally affected by various errors, depending on sensor time-response, air temperature and other bias effects. The time-response errors were evaluated on the basis of laboratory results, using the radiosonde temperature measurements to calculate the time-lag values in terms of a power of 10 with exponent depending on air temperature, according to the NIST data provided by Vaisala. This procedure yielded values of the time-constant increasing as a function of decreasing temperature, which were evaluated as equal to 7 s at T

= −20 °C, 27 s at T = −40 °C, 108 s at T = −60 °C, and 215 s at T = −70 °C. On this basis, a correction procedure was proposed by Miloshevich *et al*. (2001) for minimizing the errors due to the temperature dependence of the sensor, which makes use of a correction curve calculated in terms of a non-linear relationship between the sensor responsivity and air temperature T, giving correction factors of about 1.1 at T = −35 °C, 1.4 at T = −50 °C, 1.8 at T = −60 °C and 2.5 at T = −70 °C. Other important bias errors were also identified and evaluated by them, using a statistically derived function of the air temperature.

Chemical contamination and temperature dependence errors have been further indicated by Wang et al. (2002) to cause the predominant errors in the HUMICAP-A measurements: appropriate correction methods for both RS-80-A and RS-80-H sensors were developed during the TOGA/COARE experiment, showing that alongside the above errors, other less important errors can be made when measuring the air humidity conditions with the A- and H-type sensors, due to the use of the basic calibration model, ground check, sensor aging and sensors-arm-heating. Both chemical contamination and temperature dependence were estimated to produce dry bias effects. In fact, the chemical contamination error of the HUMICAP-A sensor is due to the use of a polymer as sensor dielectric material, where water vapour can be absorbed or desorbed, consequently modifying the characteristics of the capacitor. The source of contaminating molecules is the sonde packaging material that outgasses after the sonde has been vacuum-sealed in its mylar foil bag. Thus, the magnitude of the dry bias is a function of sonde age (i.e. the amount of outgassing) and the humidity sensor type. In the case of the HUMICAP-A sensor, the polymer is less sensitive to contamination than the one used in the HUMICAP-H, because of its larger selectivity to water. Therefore, the contamination errors of the A-type humidity sensor are estimated to be appreciably lower than those of the H-type sensor for all the sonde ages. They were evaluated to be (i) lower than 1% for relative humidity f < 10% and around 2% for higher values of f , in the 1-year age A-type sensors; (ii) lower than 2% for f < 30% and around 3% at higher values of f for the 2-year sensors; and (iii) lower than 3% for f < 30% and ranging between 3% and 5% at higher values of f (with a maximum of 5% at f = 70%) for the 3-year age sensors. Thus, realistic corrections of such errors can be made only if the age of each sonde is known with good precision. However, considering that only 1-year and 2-year age sensors were used at Terra Nova Bay, we can realistically assume that contamination errors did not exceed 3% even for the highest relative humidity conditions: this means that the contamination errors made in all our cases are in practice of comparable magnitude with the accuracy declared by the manufacturers.

The temperature dependence errors of the HUMICAP-A sensor strongly predominate over those due to chemical contamination and other causes at temperatures below –40 °C: Wang et al (2002) have evaluated that these errors need to be corrected by means of a correction factor which considerably increases as the air temperature decreases from –40 °C to –80 °C. These findings agree very well with the results of Miloshevich *et al*. (2001). Thus, we decided to use the procedure of the latter authors to remove confidently the temperature dependence errors. The other errors due to minor sources were neglected according to the estimates of Wang et al. (2002), who established that such dry biases are usually smaller than sensor accuracy.

Thus, for each significant level of each radiosounding, we calculated the precise value of height z, together with the corresponding values of air pressure p, air temperature T and relative humidity f, following (i) the procedure of Vitale and Tomasi (1994) based on the Huovila and Tuominen (1991) relationship for the THERMOCAP corrections, and (ii) the procedure proposed by Miloshevich *et al*. (2001) for the temperature dependence HUMICAP-A corrections. The chemical contamination errors leading to dry bias effects were not corrected, being only of a few percents, due to the short sonde age of 1 or 2 years in all cases. Then, for each level and each set of the corresponding three thermodynamic parameters p, T and f, we determined the water vapour partial pressure e by multiplying f by the value of saturation vapour pressure $E(T)$ in the pure phase over a plane surface of pure water, this quantity being evaluated in terms of the general formula proposed by Bolton (1980),

$$E(T) \; = \; 6.112 \;\; \exp\,[17.67\;(T - 273.16)/(T - 29.66)]\;, \qquad (3)$$

where T is measured in °K. The results provided by eq. (3) were found by Tomasi and Deserti (1988) to agree very well with the values obtained using the Goff-Gratch (1946) formula, and those determined by List (1966) within the 190÷310 °K temperature interval. Moreover, for each value of e found in terms of eq. (3), we calculated the corresponding value of dew-point T_d, using the inverse formula of eq. (3), as obtained by Bolton (1980).

Thereupon, for each set of values of p, T, e and T_d, we calculated the values of absolute humidity q, using the values of T and e in the well-known equation of state for water vapour:

$$q = 216.685\, e/\, T\,, \tag{4}$$

where q is measured in g m^{-3}, e in hPa and T in °K.

Consequently, realistic vertical distribution curves of the six above-mentioned meteorological parameters were obtained for the whole set of 1330 radiosounding measurements taken at the Terra Nova Bay station during the twelve campaigns. The long and careful analysis of the various physical parameters enabled us to define the most reliable vertical profiles of air pressure p, air temperature T, dew-point T_d, water vapour partial pressure e, relative humidity f and absolute humidity q for each radiosounding measurement. Fig. 1 shows a comparison between (i) the vertical profiles of temperature T and dew-point T_d, directly obtained from the original data-sets given by the two radiosoundings performed on November 24, 1990 at 12:00 GMT and January 31, 1991 at 00:00 GMT, and (ii) the vertical profiles of T and T_d, correspondingly found following the present correction procedures, which take into account the lag effects and temperature dependence errors of the THERMOCAP and HUMICAP-A sensors, respectively, from the ground-level up to 12 km height for temperature T and up to the highest tropospheric level measured by the HUMICAP-A sensor for dew-point T_d , respectively. As can be seen, both vertical curves of temperature T turn out to be only slightly modified by the lag corrections, while those of dew-point T_d change appreciably as a result of the corrections for the hygrometric errors, leading to mean variations in altitude of about 10 m in the first case, within the altitude range from 6 to 10 km, and of less than 10 m in the second case, within the altitude range from 5 to about 7 km. Thus, the values of relative humidity f were found to increase considerably as a result of the above corrections, by 1% to 8% in the above cases, within the 6-10 km altitude range in the first case (with the greatest change at 7.7 km) and within the 5.2-7.4 km height range in the second (with the maximum at 6.9 km).

4.- SELECTION OF HUMIDITY DATA FOR CLOUDLESS CONDITIONS

Examining the relative humidity measurements taken in cirrus clouds, Miloshevich *et al*. (2001) pointed out that “the radiosonde sensors in general cannot measure ice-supersaturation because the surface of the sensor acts as a nucleation site upon which vapour condenses, so the sensor itself is actually exposed to air that is at ice-saturation”. In practice, the sensor measures 25÷30% relative humidity below ice-saturation when ice-supersaturation conditions actually occur. This implies that, within the ice-supersaturated region of the cloud, the error relative to the ice-saturation curve is a dry bias whose magnitude increases with decreasing temperature, presumably in the range of T below -40 °C. Moreover, clouds are expected to cause another kind of dry bias, since "during balloon ascent in the real atmosphere, humidity changes quickly and the sensor does not have time to drift upward unless it is in a cloud for a long period of time", as remarked by Wang et al. (2002).

For these reasons, we paid special attention when dealing with the HUMICAP measurements performed at high tropospheric levels, where cirriform clouds may be often present and parameter f usually assumes very low values, close to the lower limit of its sensitivity range for cloudless conditions. Considering the above evaluations of the accuracy and repeatability constants of the HUMICAP sensor, we decided to reject all the measurements of $f \leq 4\%$, so as to reduce the dry bias errors occurring in slightly humid air masses. Moreover, since the presence of clouds generally implies higher values of relative humidity f at all the tropospheric levels, where ice-saturation conditions can result to be underestimated by the sonde because of the above dry-bias errors, we decided to test the whole set of HUMICAP-A measurements performed at the Terra Nova Bay station in order to attempt to distinguish the humidity conditions of the cloudless atmosphere from those occurring inside the clouds.

Among the overall set of 1330 radiosoundings, we selected a set of 84 radiosoundings, taken on days when simultaneous sun-photometric measurements (Tomasi et al., 1989; Vitale and Tomasi, 1990; Tomasi et al., 1992; Cacciari et al., 2000) were performed, showing that clear-sky conditions were present above the Terra Nova Bay station during the radiosounding time-periods. The aerosol optical depth measurements carried out on those days with multispectral sun-photometers yielded relatively low values of these spectral quantities at all the visible and near-infrared window-wavelengths. Therefore, we examined separately the temperature and dew-point data obtained from the radiosoundings on the above-selected 84 days (of certain cloudless conditions and with low columnar contents of aerosol particles) and the corresponding data-set including all the 1330 radiosoundings, in order to define the values of f given by each radiosounding at the following 189 fixed levels from ground-level (55 m a.m.s.l.) to 12 km height: (i) 38 levels besides the ground-level, taken in steps of 25 m from the 75 m to 1 km altitude; (ii) 80 levels in steps of 50 m from 1 to 5 km; and (iii) 70 levels in steps of 100 m from 5 to 12 km. More precisely, we determined the values of T and T_d at all the levels by linear interpolation in height between the values of these two quantities obtained at the significant levels of the radiosoundings. The values of f were then calculated following the procedure described above. The relative frequency histograms of *f* were subsequently defined at all the above levels, for both data-sets, after which the values of the three quartiles, the 90^{th} percentile and the upper limit of *f* were determined. Fig. 2 presents the vertical profiles of the five relative frequency parameters found for the clear-sky data-set, together with those of the three quartiles for the overall data-set. The comparison indicates that the clear-sky quartiles assume considerably lower values than those of the overall data-set, with relative differences of around 20% at all altitudes from ground-level to 9 km. Moreover, the vertical profile of the 90^{th} percentile for the clear-sky data-set was found to be almost overlapping that of the third quartile of the overall data-set, throughout the whole range from ground level to more than 8 km height. It is also of interest to notice that in Fig. 2 the 90^{th} percentile values and, hence, the values of the overall third quartile turn out to be appreciably lower (by about 20%) than those of the vertical profile relative to the maximum values of *f* for clear-sky conditions, throughout the whole height range from sea-level to 9 km. In fact, the values of the overall third quartile mostly ranged between 60% and 75% in the lower part of the troposphere below 4 km altitude, and gradually decreased with height from about 70% to a few percents throughout the upper part. Thus, the choice of values of *f* below the third quartile values guarantees that these data do not refer to saturation or supersaturation conditions in cloudy atmospheres, even in cases where the largest dry bias errors described by Miloshevic et al. (2001) should have been made. The dry bias errors mentioned by Wang et al. (2002) cannot be easily detected: since they do not alter in practice the initial clear-sky values of *f* given by the sonde before reaching the cloud, the data affected by these errors can be assumed as representative of cloudless conditions of the atmosphere. It should also be taken into account that the presence of extended cirriform cloud layers may cause wet bias errors due to freezing processes that take place inside the capacitive film of the HUMICAP sensor. Since ice can remain inside the sensor also in cloudless conditions, until it is completely sublimated during the ascent and the equilibrium conditions between sensor and external air are reinstated, the sensor continues to provide values of *f* close to saturation conditions also after the sonde has left the cloud. However, the choice of threshold values, as given at the various levels by the 90^{th} percentile curve should account for all these possible occurrences. The results shown in Fig. 2 clearly indicate that the values of the third quartile of the overall data-set can be confidently used as threshold values of *f* at all significant levels, suggesting that values lower than those of the third quartile can be correctly assumed as representative of cloudless atmospheric conditions.

Therefore, we re-examined the overall data-set of radiosounding measurements and discarded all the HUMICAP-A measurements yielding values of f higher than the threshold values defined in Fig. 2 by the third quartile vertical profile for the overall data-set. We then calculated the values of parameters p and T using a detailed grid in altitude, consisting of the following levels:

(1) the 189 levels from the ground-level to 12 km height, as defined above;

(2) 52 levels taken in steps of 250 m from 12 to 25 km; and

(3) other additional levels, taken in regular steps of 500 m, throughout the upper altitude range from 25 km to the top-level reached by the radiosonde.

In addition, we calculated the values of moisture parameters T_d, e, $E(T)$, and q at all the above-fixed levels from the ground to 12 km, where the reliability of relative humidity measurements had been carefully ascertained through the above procedure. More precisely, the values of total air pressure p were calculated at each level by log-linear interpolation in height between the radiosounding data corrected above for the lag-effects, according to the exponential form of the well-known hydrostatic equation. The corresponding values of air temperature T and dew-point T_d were calculated by linear interpolation in height between the above-corrected values of these two physical quantities, realistically assuming that the vertical gradient of temperature assumes stable values at all the intermediate altitudes. For all the pairs of T and T_d found at the above fixed levels, we then calculated: (i) the values of water vapour partial pressure e, as given by eq. (3), where T_d was set in place of T; (ii) the values of relative humidity f in terms of ratios between e and $E(T)$, the latter being calculated according to eq. (3); and (iii) the values of absolute humidity q as a function of parameters e and T, according to eq. (4).

The launch base of the radiosonde balloons at Terra Nova Bay is situated at an altitude of 55 m a.m.s.l.. Since we are attempting to define a set of mean atmospheric models covering the tropospheric height range from sea-level to a stratospheric altitude of 32 km for air pressure and temperature, and from sea-level to the tropopause height for moisture parameters, we decided to determine the values of the various meteorological parameters from the ground-station level to the sea-level, taking into account the information on the thermodynamic characteristics of the surface layer provided by the meteorological data occasionally measured by a baro-thermo-hygrograph placed at the Terra Nova Bay base, at an altitude of a few meters above sea-level. From the comparison between the measurements of parameters T and T_d taken at sea-level and the simultaneous measurements of the same two quantities defined from the radiosoundings at the 75 m level, we calculated the average differences between the measurements performed at the two levels for the various ten-day periods from late October to mid-February, determining the mean ten-day values of vertical gradients of T and T_d within the surface layer of 75 m depth. The mean ten-day values were used to determine the vertical profiles of parameters T and T_d and, consequently, of e, f and q from the 75 m level to the sea-level, for all the radiosounding measurements taken at the Terra Nova Bay station. Thus, the complete data-set obtained for each radiosounding consisted of:

(1) values of parameters p and T at (i) 41 levels taken in steps of 25 m from the sea-level to 1 km altitude, (ii) 80 levels from 1 to 5 km, in steps of 50 m, (iii) 70 levels from 5 to 12 km, in steps of 100m, (iv) 52 levels from 12 to 25 km, in steps of 250 m, and (v) other additional levels from 25 km up to the top-level of the radiosonde, which only rarely exceeded 32 km height; and

(2) values of parameters T_d, e, f and q taken at (i) 41 levels, in steps of 25 m from the sea-level to 1 km altitude, (ii) 80 levels from 1 to 5 km, and (iii) 70 levels from 5 to 12 km, in all cases where the HUMICAP-A sensor was capable to provide realistic measurements of relative humidity up to 12 km.

5.- DETERMINATION OF THE ATMOSPHERIC MODELS OVER TEN-DAY PERIODS

The vertical profiles of p, T, T_d, e, f and q obtained in the previous section from the overall data-set of 1330 radiosoundings were divided into twelve ten-day sub-sets, each sub-set corresponding to one of the ten-, eleven- or twelve-day periods, fixed from October 20 to February 20 as shown in Table 1. Each of the sub-sets includes data taken in various years, the first and second sub-sets in 4 years only, the third to the sixth sets in 6 or 7 years, and the seventh to the twelfth sets in 8÷10 years. About half of the radiosoundings were taken at 00:00 GMT, i.e. when the solar elevation angle was close to its maximum daily value, varying between 25° and 39° throughout the period from mid-October to mid-February. The other radiosoundings were taken at around 12:00 GMT, for lower values of the solar elevation angle, ranging from negative values in October and mid-February to no more than 8° throughout the rest of the measurement period. Examining a large set of radiosounding measurements taken at different Antarctic stations at both 00:00 GMT and 12:00 GMT, Connolley and King (1993) verified that the differences among the measurements of the moisture parameters taken at different hours are not significant “in the sense that the differences between the means are less than 1 standard error”. Similarly, through the careful analysis of a five-year set of radiosounding measurements taken at the Terra Nova Bay station, Vitale and Tomasi (1994) determined the mean vertical profiles of temperature and absolute humidity for the six fifteen-day periods from mid-November to mid-February, examining separately the radiosounding data taken at different GMT hours. The temperature difference ΔT between 00:00 and 12:00 GMT values was found to vary mostly between –1.5 and +0.8 °K and, hence, to be appreciably smaller than the corresponding standard deviations. This indicates that only limited variations in the thermal features of the troposphere are usually produced by the solar heating of the ground. Moreover, Vitale and Tomasi (1994) found that the relative difference in absolute humidity at the various tropospheric levels varies mostly between –25% and +5%, with a median value of –8%, substantially confirming the remarks of Connolley and King (1993). However, before assuming that diurnal and nocturnal temperature and humidity parameters do not differ greatly throughout the troposphere and, hence, that it is meaningful to analyse the measurements performed at different hours all together, we carried out a preliminary analysis of the present twelve-year data-set by calculating the average differences between 00:00 and 12:00 GMT mean values of such thermodynamic parameters. The values of ΔT calculated within the lower tropospheric layer of 3 km depth was found to vary between –0.5 and +0.6 °K during the period from the end of October to mid-February, confirming that only relatively weak warming effects in the low troposphere can be ascribed to changes in the incoming solar radiation due to variations in the sun zenith angles. Correspondingly, the differences in absolute humidity observed within the 0÷3 km layer were evaluated to range between +0.08 and –0.03 g/m^3, i.e. between approximately +10% and –3%.

In view of these results, we decided to analyse the 00:00 and 12:00 GMT data all together, obtaining data-sets consisting of more numerous cases for the various ten-day periods. As can be seen in Table 1, the ten-day sub-sets consist of more than 100 radiosoundings, except for the first ten-day period (23 cases), the second (65 cases), the third (97 cases) and the last one (63 cases). The division of the entire data-set into twelve ten-day sets allowed us to define the mean temperature and moisture features of the atmosphere in the various ten-day periods and to describe the main evolutionary features of air temperature in the troposphere and stratosphere, as well as those of moisture parameters in the troposphere. For each ten-day set, we calculated the average values of parameters p and T at all the 257 levels fixed above from sea-level to 32 km height, together with the standard deviation σ_p of air pressure p and standard deviation σ_T of air temperature T. We then calculated (i) the average values of T_d at all the levels fixed above, from the sea-level to the tropopause region, together with the corresponding standard deviations σ_d, and (ii) the corresponding mean values of humidity parameters e, f and q at all the fixed levels, together with their standard deviations σ_e, σ_f and σ_q, respectively.

Two examples of the results obtained are shown in Fig. 3, which presents the vertical profiles of air temperature T and absolute humidity q determined for the first and the fourth ten-day periods defined in Table 1, together with their standard deviations calculated at all the fixed levels. The vertical profile of temperature for the first ten-day period exhibits considerably lower mean values than those defined in the other period at all the tropospheric levels, to an extent which turns out to be larger than the standard deviation σ_T, varying between 1.4 and 3.3 °K in the 0÷10 km height range. In the lower stratosphere, parameter T describes a wide minimum, reaching very low values of nearly 205 °K at altitudes between 12 and 14 km. The temperature begins to increase considerably at altitudes above the 15 km level, assuming values comparable to those of the fourth ten-day period at heights around 23 km. In the first ten-day period, σ_T varied between 1.1 and 8.0 °K in the 10÷20 km height range, and between 2.9 and 9.1 °K in the 20÷30 km height range, while in the fourth period it varied between 2.9 and 4.8 °K in the 0÷10 km range, between 4.7 and 8.2 °K from 10 to 20 km, and between 3.1 and 4.6 °K from 20 to 30 km. The differences between the average values of T determined for the two ten-day periods in Fig. 3 were evaluated to vary (i) between 7.1 and 10.5 °K from the sea-level to 10 km altitude, with the sum of the corresponding standard deviations ranging between 5.2 and 7.7 °K; (ii) between 7.2 and 18.9 °K in the altitude range from 10 to 20 km, with values of the overall standard deviation varying between 6.2 and 13.2 °K; and (iii) between −11.5 and 11.3 °K throughout the upper region, with values of the overall standard deviation ranging between 5.0 and 13.3 °K. The results allow us to evaluate the intensity and extension of the marked warming of the troposphere and low stratosphere occurring from the end of October to the end of November, in the region between 10 and 24 km altitudes, with an average temperature increase of about 7 °K at middle tropospheric levels and more than 10 °K within the 12÷21 km height interval. Thus, Fig. 3 gives evidence of the marked variations occurring in the thermal conditions of the atmosphere during the period from late October to the end of November, at all tropospheric and stratospheric levels.

The vertical profiles of absolute humidity obtained in the troposphere following the above procedure are shown in the left part of Fig. 3, for the same two ten-day periods, with values of σ_q which result to be mostly less than 2 10^{-2} g/m^3 in the lower troposphere and less than 10^{-4} g/m^3 in the upper troposphere. The comparison shows that the average values of q double in practice at all the tropospheric levels from the end of October to the end of November. In order to give a comprehensive picture of all the above variations, Tables 2, 3, 4 and 5 present the average values of air pressure p, air temperature T, dew-point T_d, and absolute humidity q at 13 fixed levels from the sea-level to 10 km height, and for air pressure and temperature only at another 14 fixed levels from 11 to 32 km height.

6.- EVOLUTIONARY FEATURES OF THE TEMPERATURE AND ABSOLUTE HUMIDITY VERTICAL PROFILES

Examination of the ten-day average values of temperature, dew-point and absolute humidity given in Tables 2 ÷ 5 offers an opportunity to evaluate the mean changes in the thermodynamic conditions of the troposphere and low stratosphere from the last days of October to mid-February. The comparison made in Fig. 3 between the mean vertical curves of parameter T provides the measure of the important variations caused by the warming processes occurring during the first month of the observation period at the various tropospheric levels and at stratospheric levels up to more than 20 km height. These can be reasonably explained in terms of strong dynamic exchanges of heat between troposphere and stratosphere and intense heat transport from mid- to high-latitude regions. Fig. 3 also shows a considerable increase in absolute humidity occurring from late October to the end of November throughout the whole troposphere.

In order to evidence the marked temperature variations measured at various atmospheric levels during the radiosounding period of about four months, Fig. 4 presents the mean vertical profiles of air temperature T obtained for the twelve ten-day periods. The parameter was found to increase considerably during the first month, at all the tropospheric levels, assuming more stable values in December and January, and then decreasing slowly in February. To give an idea of such variations, air temperature at sea-level increases from 257.9 to 272.5 °K from late October to late December, while there is a simultaneous increase of T from 237.9 to 246.3 °K at the 4 km level, and from 213.0 to 223.9 °K at 8 km. During January and February, parameter T was found to decrease slowly, reaching the value of 267.3 °K at sea-level and

244.0 °K at 4 km height. At altitudes ranging from less than 10 km to about 24 km, a large increase in T was observed from the end of October to early December, followed by a slow decrease in the subsequent ten-day periods: parameter T increases from 204.9 to 232.3 °K at the 15 km level, during the period from late October to early January, and then decreases towards the mean value of 230.2 °K in mid-February. At stratospheric levels higher than 25 km, T was observed to decrease gradually from October to February, assuming a value of 257.7 °K at the 30 km altitude, in late October, and lower values in the subsequent periods, down to 236.8 °K in mid-February.

The evolutionary time-patterns of air temperature present negative values of the average vertical gradient γ calculated within the tropospheric region from sea-level to 8 km height: this quantity was found to (i) be equal to –5.6 °K/km in the first ten-day period, (ii) assume gradually higher absolute values over the subsequent two months, (iii) reach the value of –6.3 °K/km in the sixth ten-day period, (iv) decrease to –6.1 °K/km in the seventh ten-day period, as can be seen in Table 6, and (v) assume gradually less marked negative values during January, until reaching –5.4 °K/km in mid-February. A more detailed analysis of the mean values of parameter γ calculated within layers of 1 km depth from 0 to 8 km altitude indicates that parameter γ reveals the most marked features within the first layer, mainly varying between –8 and –6 °K/km, less marked values in the second, third and fourth layers, mainly ranging between –6 and –5 °K/km, and appreciably lower values in the upper layers, mainly ranging between –6 and –4 °K/km from 6 to 8 km.

As mentioned above, air temperature describes a wide minimum at all the levels between 8 and 15 km, presenting the lowest value T_M at gradually decreasing altitudes from the first to the seventh ten-day period. Simultaneously, a strong warming was observed at the upper stratospheric levels, up to about 24 km height. Conversely, a marked temperature decrease turned out to occur at the upper levels, mostly during the first two ten-day periods. As shown in Table 6, parameter T_M assumes the value of 204.7 °K at level z_M = 13.7 km in the first ten-day period, and then continues to increase up to 224.1 °K in the last period, while z_M lowers gradually to less than 8 km in mid-February. At stratospheric levels from 8 to 20 km, parameter T varies only slightly in January and February, showing a decreasing trend of the average vertical gradient calculated within the 10÷20 km altitude interval, with values gradually decreasing from 1.1 to 0.4 °K/km.

The mean vertical profiles of absolute humidity q determined in the troposphere for the twelve ten-day periods are shown in Fig. 5. The results indicate that q decreases as a function of height in an approximately exponential fashion throughout the first 10 kilometres, characterised by values of the mean exponential coefficient ranging between 0.45 and 0.60 km^{-1} within the first tropospheric kilometres and between 0.50 and 1.00 km^{-1} in the rest of the troposphere. The values of q found in the first ten-day period do not appreciably differ from those determined in the second period at all the tropospheric altitudes. However, they turn out to increase considerably in the third ten-day period, by 50÷80% in the lower part of the troposphere and by 80÷100% in the upper part. During the four subsequent ten-day periods, absolute humidity was found to increase slightly, in general by overall percentages lower than 30% at all the tropospheric levels. The tropospheric vertical profiles of q obtained in the last five ten-day periods exhibit very similar features, characterised by relative variations that are (i) positive and equal to a few percents in early January, at all the tropospheric levels, (ii) generally negative in mid-January, and (iii) of variable sign during the last month. Correspondingly, precipitable water w was evaluated to almost double during the first three ten-day periods, increasing from 0.12 to 0.23 g cm^{-2}, and to increase slowly from late November to early January, assuming values gradually passing from 0.27 to 0.36 g cm^{-2}, as a result of the more stable moisture conditions of the troposphere. In January and February, i.e. during the last five ten-day periods, parameter w was found to decrease gradually from 0.35 to 0.26 g cm^{-2}, presenting the time-fluctuations of the absolute humidity field described above.

7.- MIXING RATIO VERTICAL PROFILES IN THE STRATOSPHERE

One of the principal aims of the present study was to obtain atmospheric models suitable for calculating the effects produced by radiative transfer processes occurring in the atmosphere, also at stratospheric heights. Therefore, it was very important for us to define realistically the humidity parameters within the stratosphere. Considering that the humidity measurements provided by the HUMICAP-A sensor at stratospheric levels are not reliable, due to the numerous errors and limited accuracy of instrumentation, we decided to analyse the mixing ratio measurements available in the literature, carried out in the Antarctic stratosphere by means of more sophisticated techniques than the one utilized in the radiosoundings.

To evaluate realistically the mean evolutionary time-patterns of moisture parameters in the stratosphere during the period from mid-October to mid-February, we determined the average values of water vapour mixing ratio *R* in the lower stratosphere from independent measurements taken at different high-latitude sites in the Southern Hemisphere, as follows:

(1) the vertical profiles of water vapour mixing ratio *R* obtained by Rosen et al. (1991) on October 20 and 21, 1990, over the South Pole region by means of balloon-borne observations, presenting values of *R* increasing from 2 to 8 ppmv at heights varying between 15 and 30 km;

(2) the measurements of *R* obtained at 70 °S latitude by Rind et al. (1993) in March 1987 from water vapour observations taken during the SAGE II experiment, providing values of *R* varying between 4 and 10 ppmv in the 10÷25 km height range;

(3) the vertical profiles of *R*, obtained at 77 °S latitude by Harries et al. (1996) in October 1992, during the HALOE experiment, with an infrared solar limb occultation radiometer, presenting values of *R* varying between 2.3 and 5.6 ppmv throughout the altitude range from 15 to 31 km, with standard deviations varying between 0.5 and 1 ppmv;

(4) the water vapour measurements obtained by the Microwave Limb Sounder on the UARS satellite and examined by Lahoz et al. (1996), giving cross-sections of *R* as a function of height and latitude, also within the latitude range from 70 °S to 80 °S: in the present study, we considered the measurements taken at 75 °S and 80 °S on October 30 and November 28, 1992, yielding values of *R* ranging between 4.6 and 5.6 ppmv in the 24÷32 km height interval;

(5) the vertical profiles of the zonally averaged water vapour mixing ratio *R* derived by Chiou et al. (1997) from the observations performed during the SAGE II Experiment aboard the Earth Radiation Budget Satellite (ERBS) from January 1986 to May 1991: we considered the average values of *R* determined in the 80 °S-60 °S latitude zone over the 3-month period including September, October and November and the one including December, January and February, with 1-km vertical resolution in the altitude range from 10.5 to 32.5 km; and

(6) the vertical profiles of the monthly mean values of *R* derived by us at 75 °S latitude in the height range from 10 to 26 km, for the five months from October to February, from the meridional cross sections of the monthly zonal mean water vapour mixing ratio *R* determined by Randel et al. (2001) through the analysis of the Halogen Occultation Experiment (HALOE) satellite observations spanning the period from 1991 to 2000, plus Microwave Limb Sounder (MLS) climatology.

Examination of the above measurements suggests the following remarks: (i) within the height range from 16 to 22 km, the measurements of *R* at latitudes lower than 75 °S are considerably higher than those obtained at the higher southern latitudes and, hence, were discarded as they were not representative for the Terra Nova Bay area; (ii) the measurements of *R* carried out by Rosen et al. (1991) at the South Pole agree very closely with those performed by Harries et al (1996) at 77 °S, throughout the whole height range from 15 to 21 km; (iii) appreciable but not large differences exist within the 22÷26 km height range between the measurements performed by Rosen et al. (1991) at 90 °S, those of Harries et al. (1996) at 77 °S and those of Lahoz et al. (1996) at 85 °S; and (iv) a substantial agreement characterises the values of *R* measured by Rosen et al. (1991), Harries et al. (1996) and Lahoz et al. (1996) at high latitudes within the 26÷32 km height range, while the values of *R* found at lower latitudes turn out to be only slightly smaller at the same altitudes.

All the above values of *R* are shown in Fig. 6, divided into four periods: (i) the last ten-day period of October; (ii) November and December; (iii) January; and (iv) February and March. The comparison in Fig. 6 gives rise to the following considerations:

(a) the measurements taken in October assume the lowest values at levels varying between 12 and 18 km, describing a pronounced minimum, while they yield higher values of *R* than those measured in the subsequent periods at levels above 20 km;

(b) the values of *R* obtained in November and December exhibit more stable features in the lower part of the stratosphere, without defining a marked minimum as in the previous case, and then slowly increase at upper levels to become stable above the 20 km altitude and close to those of the first ten-day period;

(c) in the subsequent period from January to March, the mixing ratio turns out to change only slightly with respect to the vertical distribution features of the previous two months.

Thus, we determined the mean vertical profile of *R* for each of the twelve ten-day periods by linear interpolation in time between the vertical curves of *R* defined in Fig. 6. This procedure was followed at all the 83 significant levels from 11 to 32 km height, where the values of air pressure *p* and air temperature *T* were determined through the procedure described in Section 4. Using the mean vertical profiles of total air pressure *p*, determined above and given in Tables 2 ÷ 5 for all the twelve ten-day periods, together with the above vertical profiles of *R*, we calculated the values of water vapour partial pressure *e* at all the 83 fixed levels from 11 to 32 km height and for each of the twelve ten-day periods, by means of the approximated formula

$$e = (29/18)\, R\, p\,, \qquad (5)$$

adopted by Connolley and King (1993) to define the moisture conditions at various levels. Each vertical profile of *e* was then associated to the corresponding mean vertical profile of air temperature T determined in Section 4 to calculate at all the above-fixed levels the corresponding values of T_d, f and q, following the procedure based on eqs. (3) and (4), and taking into account the corresponding values of T given by the radiosoundings. This procedure was used to determine the vertical profiles of parameters e, T_d, f and q for all the 1330 selected radiosoundings, from 11 to 32 km height. Tables 2÷5 present the mean values of T_d and q at fourteen significant stratospheric levels (italics). The vertical profiles of absolute humidity q are shown in Fig. 7, presenting in its left part those relative to the first six ten-day periods and the other six vertical profiles on the right. The comparison highlights the marked increase in moisture conditions during late October and early November at all the stratospheric levels, as well as the further increase throughout the subsequent three ten-day periods at altitudes lower than 20 km. Only small and negligible variations in the moisture conditions appear to take place at all the stratospheric levels after mid-December.

8.- CONCLUSIONS

The mean vertical profiles of air temperature determined for twelve ten-day periods from late October to mid-February show that parameter T increases at all the tropospheric levels, significantly modifying the vertical distribution curves of temperature in the tropopause region. The average increase of T turns out to be greater than 14 °K at the sea-level, more than 8 °K at 4 km and nearly 11 °K at 8 km altitude, indicating that a more intense warming takes place in that period within the first 2 kilometres of the atmosphere and at heights above 6 km. During January, the tropospheric temperature decreases slowly with time, by 1÷2 °K only at all levels, and more rapidly during the first twenty days of February. In fact, the average variation in T during the last twenty days is about –3 °K at all levels in the ground-layer of 1 km depth and of –0.8 °K in the upper troposphere, with intermediate values in the middle region.

The average gradient of tropospheric temperature becomes steeper from the end of October to mid-December, passing from –5.6 to –6.3 °K/km. It assumes quite constant values in January, ranging between –6.1 and –5.9 °K/km, and becomes more moderate in February, reaching a mean value of –5.4 °K/km in the last ten-day period. Correspondingly, the moisture conditions of the troposphere increase considerably from early November to mid-December, by 2÷3 times within the first 5 kilometres and by more than 3 times in the upper troposphere. Subsequently, *q* was found to decrease slowly in January and February, more sharply within the ground-layer and at levels close to the tropopause than in the middle troposphere, by percentages varying between 40% and 80% at the various levels. As a result of these changes in the humidity conditions of the troposphere, precipitable water w turns out to (i) increase by three times from late October to early December, passing from 0.12 to 0.36 g cm^{-2}, (ii) assume values fluctuating between 0.30 and 0.36 g cm^{-2} throughout the rest of December and the first half of January, and (iii) decrease from 0.35 to 0.26 g cm^{-2} in the remaining period. Taking into account that the vertical profiles of *q* were defined by removing from the whole set of radiosounding data the relative humidity measurements exceeding the values of the 90^{th} percentile profile (relative to the clear-sky data-set), the tropospheric vertical profiles of *q* found in the present study appear to be suitable for representing correctly the moisture conditions typical of a cloudless atmosphere.

The present results also confirm that the low stratosphere undergoes a strong warming from the end of October to mid-December, subsequently presenting more stable temperature profiles from mid-December to mid-January, and gradually cooler conditions in the next period, with negative rates that slowly increase with height. Correspondingly, the absolute humidity tends to increase considerably in November and more slightly in early December, to such an extent that the water vapour content of the low stratosphere increases weakly during the first weeks of December, reaching almost stable values in January and early February.

Therefore, the set of mean vertical profiles of *T* and *q*, as determined for the twelve ten-day periods and shown in Figs. 4, 5 and 7, can be usefully employed in radiative transfer calculations within the Antarctic atmosphere from sea-level to more than 30 km height, in the presence of clear-sky conditions, at latitudes close to 75 °S, as in the case of the Terra Nova Bay (74° 42' S) area. These vertical profiles can be also used for calculations of the solar radiation absorption due to water vapour. Moreover, the vertical profiles of parameters *p*, *T* and *q* can be appropriately used for calculations of the Rayleigh scattering coefficient per unit volume at the various altitudes, so as to take into account the effects produced by pressure, temperature and moisture conditions on the air density and scattering properties of air molecules. In addition, the mean variations in the radiation terms that contribute to define the local energy balance of the Antarctic atmosphere can be appropriately evaluated by using the data given in Tables 2÷5, which describe the average evolutionary features of temperature and moisture conditions from the end of October to mid-February.

The above remarks indicate that January is the month characterized by the most stable temperature and humidity conditions. The vertical distribution curves of these meteorological parameters were defined by Sissenwine (1969) in the “Arctic (75 °N) atmosphere, July” model (see also COESA Working Group, 1966), giving air pressure, temperature and relative humidity as a function of height. Considering that July is the warmest month of the year in the Arctic regions, we decided to analyze the present data-set to determine a mean atmospheric model representing the average thermodynamic conditions of the troposphere and the low stratosphere at latitudes close to 75 °S, during the warmest month of the year in Antarctica. Thus, we grouped together all the mean vertical profiles of p, T, T_d, f, e and q, relative to the 8^{th}, 9^{th} and 10^{th} ten-day periods. We obtained an overall data-set consisting of 454 radiosounding measurements and containing as many vertical profiles of parameters p, T and q, each of them providing the values of the said physical quantities at 257 of the above fixed heights. The average values were then calculated at all the levels, determining the mean vertical profiles of the above meteorological quantities, together with their standard deviations.

Fig. 8 shows on the left the mean vertical profile of air temperature T with the corresponding standard deviations given in steps of 1 km height, from the sea-level to 32 km altitude. The standard deviations turn out to range between 2 and 4 °K in the troposphere, assuming higher values from 4 to 14 km height and decreasing values at upper levels, these large values being due to the less stable thermal conditions characterizing the tropopause region. The right part of Fig. 8 presents the mean vertical profile of absolute humidity q, characterized by very small standard deviations at all levels. The profile exhibits an evident deflection in the tropopause region and a slower decrease in the upper region below 20 km height.

These results allow us to give form to an atmospheric model that can be considered to be representative of the Antarctic atmosphere conditions occurring at around 75 °S latitude during the warmest month of the year. Therefore, the set of vertical profiles shown in Fig. 8 and the vertical distribution curves correspondingly found for air pressure p, dew-point T_d, relative humidity f and water vapour partial pressure e can be used to represent the “Antarctic (75 °S) atmosphere, January”, as we have called it in analogy to the one defined by Sissenwine (1969). This atmospheric model can be used for carrying out calculations of the most important radiative processes occurring in the Antarctic atmosphere, such as water vapour absorption of both short-wave and long-wave radiation, emission of thermal radiation and Rayleigh scattering.

ACKNOWLEDGMENTS

The authors gratefully acknowledge L. M. Miloshevich for his valuable comments and suggestions. This research has been supported by the Programma Nazionale di Ricerche in Antartide (PNRA) and has been developed in the frame of Sub-project 6.5 “Cloud and aerosol particle influences on the radiative balance of the Antarctic atmosphere” and Sub-project 2.6 “Meteo-Climatological Antarctic Observatory”. Supplementary support is currently provided by the Sub-project 2003/6.7 “Characterization of aerosol induced climatic effects in polar regions: an assimilation and analysis of multispectral sun-photometer data from the POLAR-AOD network”.

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FIGURE LEGENDS

Fig. 1.- Two examples of comparison between the vertical profiles of air temperature T (dashed curves) and dew-point T_d (solid curves) obtained from the original radiosounding data recorded at the Terra Nova Bay station on November 24, 1990 (left) and January 31, 1991 (right). The corresponding vertical profiles of T (dotted curves) and T_d (dotted and dashed curves) obtained by correcting the original radiosounding data for the lag-effects of the THERMOCAP and HUMICAP sensors, by using the present procedures.

Fig. 2.- Vertical profiles of the three quartiles (dashed curves), the 90th percentile (solid triangles) and the maximum values (open circles) obtained from the relative frequency histograms of relative humidity f defined at 189 fixed levels from the ground-level (55 m a.m.s.l.) to the 12 km height, as determined for the data-set consisting of 84 radiosoundings performed for clear-sky conditions. These vertical profiles are compared with those of the three quartiles obtained for the overall set of 1330 radiosoundings (solid curves).

Fig. 3.- Left: comparison between the mean vertical profiles of air temperature T obtained for the first ten-day period from October 20 to 31 (solid circles) and for the fourth ten-day period from November 21 to 30 (open squares). The horizontal bars give the values of the standard deviation σ_T at all the 257 fixed levels. Right: as on the left, for the vertical profiles of absolute humidity q obtained for the same two ten-day periods at tropospheric heights only.

Fig. 4. - Mean vertical profiles of air temperature T, as obtained from the radiosounding data-sets relative to the six ten-day periods from October 20 to December 20 (left) and to the six ten-day periods from December 21 to February 20 (right).

Fig. 5. - Mean vertical profiles of absolute humidity q in the troposphere, obtained from the radiosounding data-sets relative to the six ten-day periods from October 20 to December 20 (left) and to the six ten-day periods from December 21 to February 20 (right).

Fig. 6 – Vertical profiles of the logarithm of water vapour mixing ratio R, obtained at various high-latitude sites of the Southern Hemisphere from radiometric measurements taken from

satellites and balloons: open circles refer to measurements performed in October (Rosen et al., 1991; Harries et al., 1996; Lahoz et al., 1996; Chiou et al., 1997), solid triangles to November and December (Lahoz et al., 1996; Chiou et al., 1997; Randel et al., 2001), open squares to January (Chiou et al., 1997; Randel et al., 2001), and solid diamonds to February and March (Chiou et al., 1997; Rind et al., 1993).

Fig. 7.- Mean vertical profiles of absolute humidity q in the lower stratosphere from 11 to 32 km height, obtained from the mean vertical profiles of air pressure p and air temperature T given in Tables 2 ÷ 5 (see also Fig. 4) and the water vapour mixing ratio data obtained through linear interpolation in time between the vertical profiles shown in Fig. 6.

Fig. 8.- Mean vertical profiles of air temperature T (left) and absolute humidity q (right), where solid symbols refer to the average HUMICAP-A data and open symbols to the data derived from satellite and balloon-borne observations), obtained with their standard deviations (horizontal bars) from the data-set consisting of the three ten-day data-sets of January.

TABLE 1.- Division of the 1330 radiosounding measurements into the twelve sub-sets relative to the ten-day periods during the twelve campaigns performed at the Terra Nova Bay station from 1987 to 1998.

Ten-day Period	Measurement campaign												
	1986/ 1987	1987/ 1988	1988/ 1989	1989/ 1990	1990/ 1991	1991/ 1992	1992/ 1993	1993/ 1994	1994/ 1995	1995/ 1996	1996/ 1997	1997/ 1998	Total
1st (Oct. 20-31)	–	–	–	–	–	–	–	3	7	7	6	–	23
2nd (Nov. 1-10)	–	–	–	–	–	–	–	15	16	16	15	–	62
3rd (Nov. 11-20)	–	–	–	5	–	–	–	16	18	19	20	16	94
4th (Nov. 21-30)	–	–	–	17	13	–	–	18	21	20	20	21	130
5th (Dec. 1-10)	–	–	–	17	16	–	–	18	17	16	16	1	101
6th (Dec. 11-20)	–	–	–	18	20	–	7	19	20	20	5	–	109
7th (Dec. 21-31)	–	–	5	19	23	–	10	24	24	23	18	–	146
8th (Jan. 1-10)	–	4	14	14	15	–	7	13	18	17	12	6	120
9th (Jan. 11-20)	–	4	12	18	20	12	–	19	20	13	18	17	153
10th (Jan. 21-31)	2	–	23	20	24	–	–	24	24	21	24	19	181
11th (Feb. 1-10)	7	15	17	16	18	–	–	15	17	15	6	9	135
12th (Feb. 11-20)	2	4	11	5	13	–	–	9	20	–	–	12	76
Total	11	27	82	149	162	12	24	193	222	187	160	101	1330

TABLE 2.- Twelve-year mean values of air pressure *p*, air temperature *T*, dew-point T_d and absolute humidity *q* at 27 significant levels, obtained from the data-sets relative to the three ten-day periods from October 21 to November 20.

Height (km)	1st ten-day period (October 20 - 31)				2nd ten-day period (November 1 – 10)				3rd ten-day period (November 11 – 20)			
	p (hPa)	*T* (°K)	T_d (°K)	*q* (g m^{-3})	*p* (hPa)	*T* (°K)	T_d (°K)	*q* (g m^{-3})	*p* (hPa)	*T* (°K)	T_d (°K)	*q* (g m^{-3})
0	985.8	257.9	245.7	$5.45\ 10^{-1}$	985.0	262.0	249.0	$7.21\ 10^{-1}$	986.2	267.2	254.8	$1.18\ 10^{0}$
0.5	922.3	255.0	243.3	$4.38\ 10^{-1}$	922.5	259.2	245.3	$5.19\ 10^{-1}$	924.8	263.8	251.3	$8.75\ 10^{-1}$
1	862.2	252.0	240.7	$3.48\ 10^{-1}$	863.3	256.0	243.4	$4.42\ 10^{-1}$	866.5	260.2	248.9	$7.22\ 10^{-1}$
1.5	805.4	249.1	240.4	$3.40\ 10^{-1}$	807.3	252.8	241.3	$3.66\ 10^{-1}$	811.2	256.8	246.4	$5.82\ 10^{-1}$
2	751.8	246.4	238.3	$2.82\ 10^{-1}$	754.3	250.0	239.0	$2.97\ 10^{-1}$	762.1	253.9	244.1	$4.78\ 10^{-1}$
3	653.8	242.7	233.4	$1.74\ 10^{-1}$	657.1	244.8	234.1	$1.85\ 10^{-1}$	665.3	248.6	239.2	$3.04\ 10^{-1}$
4	567.3	237.9	226.4	$8.33\ 10^{-2}$	570.6	239.1	227.2	$9.09\ 10^{-2}$	579.2	243.6	232.4	$1.56\ 10^{-1}$
5	490.6	231.9	220.7	$4.49\ 10^{-2}$	493.6	233.0	221.2	$4.70\ 10^{-2}$	502.6	237.4	225.9	$7.89\ 10^{-2}$
6	422.5	225.2	214.3	$2.10\ 10^{-2}$	425.4	226.9	214.1	$2.04\ 10^{-2}$	434.4	230.7	217.6	$3.08\ 10^{-2}$
7	362.2	218.5	208.2	$9.77\ 10^{-3}$	365.3	220.8	207.0	$8.28\ 10^{-3}$	373.9	224.2	211.2	$1.43\ 10^{-2}$
8	309.1	213.0	201.1	$3.70\ 10^{-3}$	312.3	215.9	200.9	$3.54\ 10^{-3}$	320.6	218.7	204.1	$5.52\ 10^{-3}$
9	263.0	209.9	194.9	$1.46\ 10^{-3}$	266.4	213.1	195.7	$1.63\ 10^{-3}$	273.9	215.1	198.1	$2.35\ 10^{-3}$
10	223.8	208.7	186.5	$3.66\ 10^{-4}$	226.9	211.9	192.5	$9.93\ 10^{-4}$	233.6	213.2	194.4	$1.34\ 10^{-3}$
11	189.9	207.4	*194.1*	*$1.31\ 10^{-3}$*	193.2	211.2	*194.0*	*$1.26\ 10^{-3}$*	199.0	213.3	*193.9*	*$1.23\ 10^{-3}$*
12	161.2	206.2	*191.1*	*$8.01\ 10^{-4}$*	164.5	210.9	*191.5*	*$8.41\ 10^{-4}$*	169.6	213.7	*192.0*	*$8.99\ 10^{-4}$*
13	136.5	205.4	*189.2*	*$5.93\ 10^{-4}$*	139.9	211.0	*189.9*	*$6.49\ 10^{-4}$*	144.6	214.3	*190.6*	*$7.19\ 10^{-4}$*
14	115.5	204.8	*187.9*	*$4.72\ 10^{-4}$*	119.2	212.1	*188.7*	*$5.23\ 10^{-4}$*	123.4	215.9	*189.4*	*$5.85\ 10^{-4}$*
15	97.8	204.9	*186.9*	*$4.02\ 10^{-4}$*	101.6	214.8	*187.8*	*$4.47\ 10^{-4}$*	105.4	218.3	*188.6*	*$5.06\ 10^{-4}$*
16	82.9	206.5	*186.1*	*$3.42\ 10^{-4}$*	86.9	218.1	*187.0*	*$3.84\ 10^{-4}$*	90.1	221.4	*187.9*	*$4.37\ 10^{-4}$*
18	59.5	212.8	*185.4*	*$2.96\ 10^{-4}$*	63.7	225.5	*186.1*	*$3.18\ 10^{-4}$*	66.5	227.5	*186.7*	*$3.48\ 10^{-4}$*
20	43.4	222.9	*185.4*	*$2.80\ 10^{-4}$*	47.5	232.8	*186.0*	*$3.00\ 10^{-4}$*	49.4	234.0	*186.4*	*$3.17\ 10^{-4}$*
22	32.2	234.8	*185.1*	*$2.55\ 10^{-4}$*	35.5	237.8	*185.7*	*$2.79\ 10^{-4}$*	37.0	238.9	*186.0*	*$2.91\ 10^{-4}$*
24	24.1	242.3	*183.5*	*$1.83\ 10^{-4}$*	27.0	241.3	*184.0*	*$2.03\ 10^{-4}$*	28.0	242.1	*184.1*	*$2.06\ 10^{-4}$*
26	18.3	248.6	*182.0*	*$1.37\ 10^{-4}$*	20.9	244.5	*182.7*	*$1.57\ 10^{-4}$*	21.5	243.9	*182.8*	*$1.60\ 10^{-4}$*
28	13.9	255.7	*180.6*	*$1.03\ 10^{-4}$*	16.1	245.5	*181.3*	*$1.21\ 10^{-4}$*	16.6	244.7	*181.4*	*$1.23\ 10^{-4}$*
30	10.7	257.7	*179.3*	*$7.85\ 10^{-5}$*	12.5	247.1	*180.0*	*$9.45\ 10^{-5}$*	12.8	242.8	*180.1*	*$9.69\ 10^{-5}$*
32	-	-	-	-	9.5	242.7	*178.8*	*$7.60\ 10^{-5}$*	9.8	245.0	*178.9*	*$7.75\ 10^{-5}$*

TABLE 3.- Twelve-year mean values of air pressure p, air temperature T, dew-point T_d and absolute humidity q at 27 significant levels, obtained from the data-sets relative to the three ten-day periods from November 21 to December 20.

Height (km)	4^{th} ten-day period (November 21 – 30)				5^{th} ten-day period (December 1 – 10)				6^{th} ten-day period (December 11 – 20)			
	p (hPa)	T (°K)	T_d (°K)	q (g m^{-3})	p (hPa)	T (°K)	T_d (°K)	q (g m^{-3})	p (hPa)	T (°K)	T_d (°K)	q (g m^{-3})
0	990.1	268.4	257.5	$1.46\ 10^{0}$	988.2	269.7	261.8	$2.05\ 10^{0}$	987.4	271.6	260.6	$1.86\ 10^{0}$
0.5	928.7	265.2	254.0	$1.11\ 10^{0}$	927.2	266.1	258.0	$1.52\ 10^{0}$	926.9	268.0	256.9	$1.40\ 10^{0}$
1	870.5	261.8	251.4	$8.91\ 10^{-1}$	869.3	262.3	254.2	$1.12\ 10^{0}$	869.4	264.3	254.0	$1.11\ 10^{0}$
1.5	815.3	258.6	248.4	$6.92\ 10^{-1}$	814.2	258.9	251.2	$8.82\ 10^{-1}$	814.6	260.6	251.0	$8.69\ 10^{-1}$
2	762.9	255.6	245.4	$5.31\ 10^{-1}$	761.9	255.6	248.7	$7.12\ 10^{-1}$	762.7	257.2	248.6	$7.13\ 10^{-1}$
3	666.6	250.7	238.8	$2.89\ 10^{-1}$	665.8	250.7	241.8	$3.87\ 10^{-1}$	666.7	250.7	242.4	$4.11\ 10^{-1}$
4	580.9	245.5	232.9	$1.63\ 10^{-1}$	580.2	245.6	235.4	$2.10\ 10^{-1}$	581.0	245.2	235.2	$2.08\ 10^{-1}$
5	504.5	239.2	225.9	$7.89\ 10^{-2}$	504.1	239.6	229.3	$1.13\ 10^{-1}$	504.7	239.7	229.3	$1.14\ 10^{-1}$
6	436.5	232.6	219.8	$3.98\ 10^{-2}$	436.3	232.9	223.8	$6.30\ 10^{-2}$	436.8	233.4	223.1	$5.89\ 10^{-2}$
7	376.2	226.3	212.3	$1.63\ 10^{-2}$	375.9	225.9	217.6	$3.14\ 10^{-2}$	376.6	227.0	217.5	$3.12\ 10^{-2}$
8	323.0	221.0	205.5	$6.66\ 10^{-3}$	322.6	220.2	209.6	$1.15\ 10^{-2}$	323.3	221.3	208.7	$1.02\ 10^{-2}$
9	276.5	217.4	198.9	$2.62\ 10^{-3}$	275.9	217.3	201.3	$3.62\ 10^{-3}$	276.8	218.5	202.0	$4.01\ 10^{-3}$
10	236.2	215.7	194.4	$1.33\ 10^{-3}$	235.8	217.2	198.6	$2.45\ 10^{-3}$	236.8	219.6	199.3	$2.66\ 10^{-3}$
11	201.6	216.2	*194.7*	$1.39\ 10^{-3}$	201.6	218.7	*196.5*	$1.80\ 10^{-3}$	202.8	221.4	*196.5*	$1.79\ 10^{-3}$
12	172.2	217.0	*192.2*	$9.23\ 10^{-4}$	172.5	220.3	*192.0*	$8.74\ 10^{-4}$	173.9	222.5	*192.0*	$8.72\ 10^{-4}$
13	147.2	218.0	*191.0*	$7.47\ 10^{-4}$	147.8	221.8	*190.5*	$6.76\ 10^{-4}$	149.3	224.1	*190.5*	$6.76\ 10^{-4}$
14	125.9	219.7	*189.9*	$6.22\ 10^{-4}$	126.9	224.4	*189.6*	$5.78\ 10^{-4}$	128.3	226.3	*189.7*	$5.80\ 10^{-4}$
15	107.8	221.8	*189.2*	$5.51\ 10^{-4}$	109.0	227.1	*189.1*	$5.27\ 10^{-4}$	110.4	229.3	*189.2*	$5.29\ 10^{-4}$
16	92.5	224.7	*188.7*	$4.93\ 10^{-4}$	93.9	229.8	*188.9*	$4.99\ 10^{-4}$	95.3	232.2	*189.0*	$5.01\ 10^{-4}$
18	68.5	230.6	*187.5*	$3.95\ 10^{-4}$	69.9	234.3	*188.3*	$4.40\ 10^{-4}$	71.1	236.2	*188.4*	$4.44\ 10^{-4}$
20	51.1	235.2	*186.8*	$3.43\ 10^{-4}$	52.5	238.1	*187.3*	$3.68\ 10^{-4}$	53.4	238.7	*187.4*	$3.73\ 10^{-4}$
22	38.4	238.8	*186.1*	$2.97\ 10^{-4}$	39.6	240.2	*186.1*	$2.94\ 10^{-4}$	40.1	240.2	*186.2*	$2.98\ 10^{-4}$
24	29.0	242.0	*184.3*	$2.12\ 10^{-4}$	29.9	242.0	*184.6*	$2.25\ 10^{-4}$	30.3	241.3	*184.7*	$2.28\ 10^{-4}$
26	22.1	244.5	*182.9*	$1.63\ 10^{-4}$	22.6	243.4	*183.1*	$1.70\ 10^{-4}$	22.9	242.8	*183.1*	$1.72\ 10^{-4}$
28	16.8	245.6	*181.4*	$1.24\ 10^{-4}$	17.1	245.3	*181.7*	$1.30\ 10^{-4}$	17.3	244.5	*181.7*	$1.31\ 10^{-4}$
30	12.7	247.0	*180.0*	$9.46\ 10^{-5}$	13.0	246.2	*180.3*	$1.00\ 10^{-4}$	13.1	246.3	*180.4*	$1.01\ 10^{-4}$
32	9.8	245.7	*179.0*	$7.84\ 10^{-5}$	10.1	249.2	*179.2*	$7.95\ 10^{-5}$	10.1	248.4	*179.2*	$7.99\ 10^{-5}$

TABLE 4.- Twelve-year mean values of air pressure p, air temperature T, dew-point T_d and absolute humidity q at 27 significant levels, obtained from the data-sets relative to the three ten-day periods from December 21 to January 20.

Height (km)	7th ten-day period (December 21 – 31)				8th ten-day period (January 1 – 10)				9th ten-day period (January 11 - 20)			
	p (hPa)	T (°K)	T_d (°K)	q (g m^{-3})	p (hPa)	T (°K)	T_d (°K)	q (g m^{-3})	P (hPa)	T (°K)	T_d (°K)	q (g m^{-3})
0	987.9	272.5	258.4	1.56 10^{0}	989.5	272.4	261.8	2.05 10^{0}	989.7	272.2	260.6	1.86 10^{0}
0.5	927.4	268.5	255.1	1.20 10^{0}	929.0	268.3	258.0	1.52 10^{0}	929.1	268.0	256.9	1.40 10^{0}
1	869.9	264.8	252.4	9.71 10^{-1}	871.4	264.5	254.2	1.12 10^{0}	871.3	263.9	254.0	1.11 10^{0}
1.5	815.3	261.2	250.0	7.95 10^{-1}	816.3	261.0	251.2	8.82 10^{-1}	816.4	260.2	251.0	8.69 10^{-1}
2	763.4	257.5	247.2	6.31 10^{-1}	764.3	257.5	248.7	7.12 10^{-1}	764.2	256.7	248.6	7.13 10^{-1}
3	667.5	251.4	240.4	3.39 10^{-1}	668.3	251.6	241.8	3.87 10^{-1}	668.0	250.7	242.4	4.11 10^{-1}
4	581.9	246.3	234.0	1.84 10^{-1}	582.6	246.3	235.4	2.10 10^{-1}	582.0	245.0	235.2	2.08 10^{-1}
5	505.9	240.7	228.2	1.02 10^{-1}	506.4	240.6	229.3	1.13 10^{-1}	505.5	239.4	229.3	1.14 10^{-1}
6	438.2	234.6	222.8	5.67 10^{-2}	438.6	234.3	223.8	6.30 10^{-2}	437.5	233.3	223.1	5.89 10^{-2}
7	378.1	228.5	216.6	2.77 10^{-2}	378.3	227.9	217.6	3.14 10^{-2}	377.2	227.2	217.5	3.12 10^{-2}
8	325.1	223.9	208.0	9.33 10^{-3}	325.2	223.4	209.6	1.15 10^{-2}	324.1	223.7	208.7	1.02 10^{-2}
9	279.2	223.2	201.1	3.52 10^{-3}	279.4	223.4	201.3	3.62 10^{-3}	278.4	225.3	202.0	4.01 10^{-3}
10	239.3	224.3	200.1	3.02 10^{-3}	240.4	225.6	198.6	2.45 10^{-3}	239.9	227.9	199.3	2.66 10^{-3}
11	205.6	226.2	196.6	1.77 10^{-3}	206.8	228.1	196.6	1.77 10^{-3}	206.7	229.8	196.6	1.75 10^{-3}
12	176.9	227.3	192.1	8.68 10^{-4}	178.2	229.6	192.2	8.66 10^{-4}	178.2	230.6	192.2	8.62 10^{-4}
13	152.3	228.6	190.6	6.76 10^{-4}	153.6	230.6	190.7	6.76 10^{-4}	153.7	231.2	190.7	6.74 10^{-4}
14	131.2	230.1	189.8	5.83 10^{-4}	132.5	231.3	189.9	5.86 10^{-4}	132.6	231.4	189.9	5.86 10^{-4}
15	113.2	231.8	189.3	5.36 10^{-4}	114.4	232.3	189.4	5.41 10^{-4}	114.4	232.0	189.4	5.42 10^{-4}
16	97.7	233.1	189.1	5.12 10^{-4}	98.8	233.4	189.2	5.17 10^{-4}	98.8	232.6	189.2	5.19 10^{-4}
18	73.0	235.1	188.5	4.58 10^{-4}	73.8	234.7	188.6	4.64 10^{-4}	73.7	233.9	188.6	4.65 10^{-4}
20	54.7	236.5	187.5	3.86 10^{-4}	55.2	236.1	187.6	3.90 10^{-4}	55.1	235.2	187.6	3.91 10^{-4}
22	40.9	238.2	186.3	3.07 10^{-4}	41.4	237.1	186.3	3.12 10^{-4}	41.3	236.0	186.3	3.12 10^{-4}
24	30.8	240.0	184.8	2.33 10^{-4}	31.1	238.0	184.8	2.38 10^{-4}	30.9	236.9	184.8	2.38 10^{-4}
26	23.2	241.4	183.2	1.76 10^{-4}	23.4	239.5	183.3	1.79 10^{-4}	23.2	238.3	183.2	1.79 10^{-4}
28	17.5	243.3	181.8	1.33 10^{-4}	17.6	242.2	181.8	1.35 10^{-4}	17.5	240.6	181.8	1.35 10^{-4}
30	13.2	245.7	180.4	1.02 10^{-4}	13.3	245.2	180.4	1.03 10^{-4}	13.2	244.0	180.4	1.03 10^{-4}
32	10.0	252.8	179.1	7.74 10^{-5}	10.2	248.7	179.2	8.10 10^{-5}	10.0	248.6	179.1	7.91 10^{-5}

TABLE 5.- Twelve-year mean values of air pressure *p*, air temperature *T*, dew-point T_d and absolute humidity *q* at 27 significant levels, obtained from the data-sets relative to the three ten-day periods from January 21 to February 20.

Height (km)	10th ten-day period (January 20 - 31)				11th ten-day period (February 1 - 10)				12th ten-day period (February 11 – 20)			
	p (hPa)	*T* (°K)	T_d (°K)	*q* (g m^{-3})	*p* (hPa)	*T* (°K)	T_d (°K)	*q* (g m^{-3})	*p* (hPa)	*T* (°K)	T_d (°K)	*q* (g m^{-3})
0	989.2	271.0	258.4	1.56 10^{0}	990.4	269.7	257.9	1.51 10^{0}	987.0	267.3	255.5	1.24 10^{0}
0.5	928.4	267.0	255.1	1.20 10^{0}	929.3	265.8	255.0	1.20 10^{0}	925.5	263.5	252.8	1.00 10^{0}
1	870.5	263.0	252.4	9.71 10^{-1}	871.1	262.2	251.8	9.22 10^{-1}	867.1	260.2	249.9	7.88 10^{-1}
1.5	815.5	259.4	250.0	7.95 10^{-1}	815.9	258.9	248.8	7.18 10^{-1}	811.7	257.0	247.8	6.59 10^{-1}
2	763.2	255.9	247.2	6.31 10^{-1}	763.6	256.0	246.2	5.71 10^{-1}	759.2	254.2	245.5	5.41 10^{-1}
3	666.9	249.8	240.4	3.39 10^{-1}	667.4	250.9	240.9	3.54 10^{-1}	662.9	249.2	238.8	2.92 10^{-1}
4	580.9	244.8	234.0	1.84 10^{-1}	581.7	246.4	235.9	2.22 10^{-1}	577.2	244.0	233.7	1.77 10^{-1}
5	504.5	239.1	228.2	1.02 10^{-1}	505.7	241.0	230.1	1.23 10^{-1}	501.0	238.3	228.3	1.03 10^{-1}
6	436.6	233.0	222.8	5.67 10^{-2}	438.3	235.1	224.6	6.91 10^{-2}	433.4	232.2	223.0	5.84 10^{-2}
7	376.3	227.1	216.6	2.77 10^{-2}	378.3	229.0	219.1	3.74 10^{-2}	373.4	226.6	215.5	2.44 10^{-2}
8	323.3	223.5	208.0	9.33 10^{-3}	325.3	224.5	209.9	1.19 10^{-2}	321.1	224.2	206.3	7.40 10^{-3}
9	277.8	225.3	201.1	3.52 10^{-3}	279.8	224.2	202.7	4.46 10^{-3}	276.6	225.4	204.2	5.51 10^{-3}
10	239.5	228.2	200.1	3.02 10^{-3}	241.7	225.8	199.5	2.78 10^{-3}	238.6	227.4	204.4	5.61 10^{-3}
11	206.3	229.8	*196.6*	*1.75 10^{-3}*	208.0	228.1	*196.1*	*1.63 10^{-3}*	205.5	229.6	*195.5*	*1.47 10^{-3}*
12	177.9	230.6	*192.2*	*8.61 10^{-4}*	179.2	229.3	*192.1*	*8.49 10^{-4}*	177.2	230.0	*191.8*	*8.17 10^{-4}*
13	153.4	231.0	*190.7*	*6.74 10^{-4}*	154.4	229.6	*190.8*	*6.91 10^{-4}*	152.7	230.1	*190.8*	*6.91 10^{-4}*
14	132.4	231.2	*189.9*	*5.86 10^{-4}*	133.0	229.8	*189.9*	*5.98 10^{-4}*	131.6	230.2	*189.9*	*5.96 10^{-4}*
15	114.2	231.9	*189.4*	*5.41 10^{-4}*	114.7	230.0	*189.4*	*5.48 10^{-4}*	113.4	230.2	*189.4*	*5.42 10^{-4}*
16	98.6	232.3	*189.2*	*5.19 10^{-4}*	98.9	230.6	*189.0*	*5.11 10^{-4}*	97.8	230.4	*188.8*	*4.95 10^{-4}*
18	73.6	233.6	*188.6*	*4.65 10^{-4}*	73.6	231.5	*188.4*	*4.56 10^{-4}*	72.7	230.9	*188.2*	*4.40 10^{-4}*
20	55.0	234.6	*187.6*	*3.91 10^{-4}*	54.8	232.3	*187.6*	*3.94 10^{-4}*	54.0	231.0	*187.5*	*3.90 10^{-4}*
22	41.1	235.0	*186.3*	*3.12 10^{-4}*	40.9	233.0	*186.3*	*3.13 10^{-4}*	40.1	231.4	*186.1*	*3.10 10^{-4}*
24	30.8	235.5	*184.8*	*2.38 10^{-4}*	30.5	233.4	*184.7*	*2.38 10^{-4}*	29.8	231.7	*184.6*	*2.35$10^{-4}$*
26	23.1	236.9	*183.2*	*1.78 10^{-4}*	22.8	234.4	*183.1*	*1.78 10^{-4}*	22.2	232.5	*183.0*	*1.75 10^{-4}*
28	17.4	239.8	*181.7*	*1.34 10^{-4}*	17.1	236.8	*181.6*	*1.34 10^{-4}*	16.5	234.6	*181.5*	*1.30 10^{-4}*
30	13.1	242.5	*180.3*	*1.02 10^{-4}*	12.8	239.7	*180.2*	*1.01 10^{-4}*	12.3	236.8	*180.0*	*9.92 10^{-5}*
32	9.8	242.3	*179.0*	*7.93 10^{-5}*	9.7	244.3	*178.9*	*7.78 10^{-5}*	-	-	-	-

TABLE 6.- Values of temperature minimum height z_M, temperature minimum T_M at the tropopause level, average vertical gradient γ of the air temperature in the troposphere, and precipitable water w.

Ten-day period	z_M (km)	T_M (°K)	γ (°K/km)	w (g cm^{-2})
1st period (October 20 to 31)	13.7	204.7	-5.6	0.12
2nd period (November 1 to 10)	12.5	210.8	-5.8	0.14
3rd period (November 11 to 20)	10.2	213.1	-6.1	0.23
4th period (November 21 to 30)	10.3	215.6	-5.9	0.27
5th period (December 1 to 10)	9.7	217.0	-6.2	0.36
6th period (December 11 to 20)	9.2	218.4	-6.3	0.35
7th period (December 21 to 31)	8.7	222.8	-6.1	0.30
8th period (January 1 to 10)	8.5	222.8	-6.1	0.36
9th period (January 11 to 20)	8.1	223.6	-6.1	0.35
10th period (January 21 to 31)	8.1	223.5	-5.9	0.30
11th period (February 1 to 10)	8.5	224.0	-5.7	0.30
12th period (February 11 to 20)	7.9	224.1	-5.4	0.26