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Atmospheric ozone. A short review

E. Antonello[1]

Istituto Nazionale di Astrofisica – Osservatorio Astronomico di Brera, Milano-Merate, Italy

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Abstract. The review includes: *a)* a brief description of the lower atmosphere and its complex dynamics, and the main features of stratospheric ozone distribution; *b)* a comparison of total and effective ozone and a discussion of the variations of ozone at the different time scales; *c)* a brief description of solar activity and volcanism effects, of the ozone hole and of some atmospheric effects on UV radiation, with a conclusion regarding the possible relation with pandemics.

Key words. Troposphere and stratosphere, atmosphere dynamics, ozone hole, solar activity, UV radiation, pandemics

1. Introduction

After realizing that ozone was responsible for the absorption of solar UV spectrum in terrestrial atmosphere, in the last century an effort was made to derive its spatial and temporal properties (Dobson 1968). Starting from the seventies, a significant part of the specialized literature was devoted to the problem of the ozone depletion and ozone hole over Antarctica. In the last decades many observations were performed by several satellites and ground-based networks, and projects are going on for the generation of multi-decadal time series of harmonised ozone data [1].

In the present work we attempt to give a short review of the properties of ozone which could be of interest for the researchers of INAF involved in the project of solar UV radiation and its effects on viruses. The review is based mainly on the online textbook published by NASA (2000), with updates from more recent papers. We will start with a brief description of troposphere and stratosphere and the atmospheric dynamics, then the main features of ozone production, destruction and distribution will be discussed. We will supply some information concerning the observational methods and the difference between total ozone and effective ozone. The temporal variability of ozone at the different time scales and latitudes will be discussed with some detail. We will recall the impact of solar activity and volcanism, and the problem of the ozone hole. Finally we will mention some atmospheric effect on UV radiation.

2. Troposphere and stratosphere

In the atmosphere, the regions of interest for our review are the troposphere and the stratosphere (Fig. 1). The troposphere is the lowest region where the temperature decreases with altitude. Its height depend on the latitude and varies with the season, the average value be-

[1] e.g.: https://climate.esa.int/en/projects/ozone https://ozonewatch.gsfc.nasa.gov/ https://community.wmo.int/activity-areas/gaw

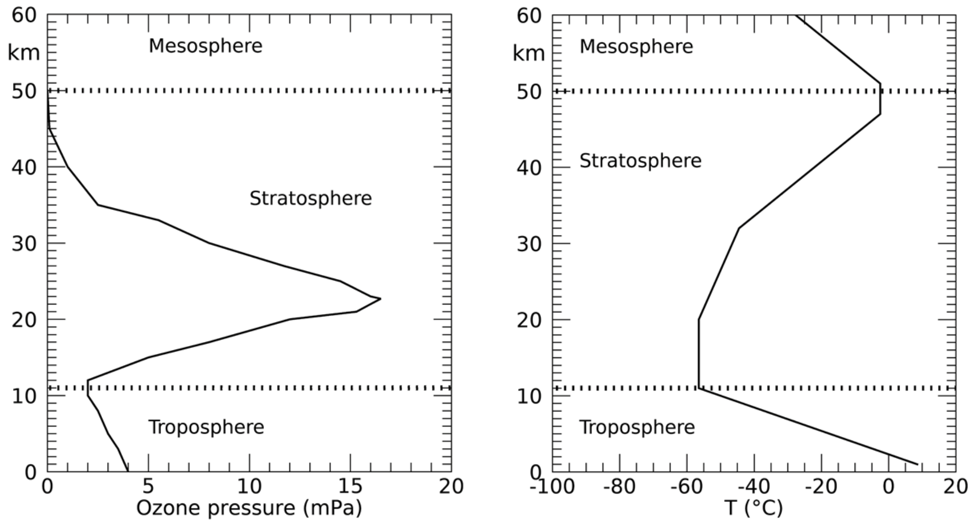

Fig. 1. *Left panel:* vertical ozone profile for mid-latitudes; the ozone concentrations are very small, only a few molecules O_3 per million molecules of air. *Right panel:* temperature of the standard atmosphere as a function of altitude; both the pressure and the air density decrease with the altitude.

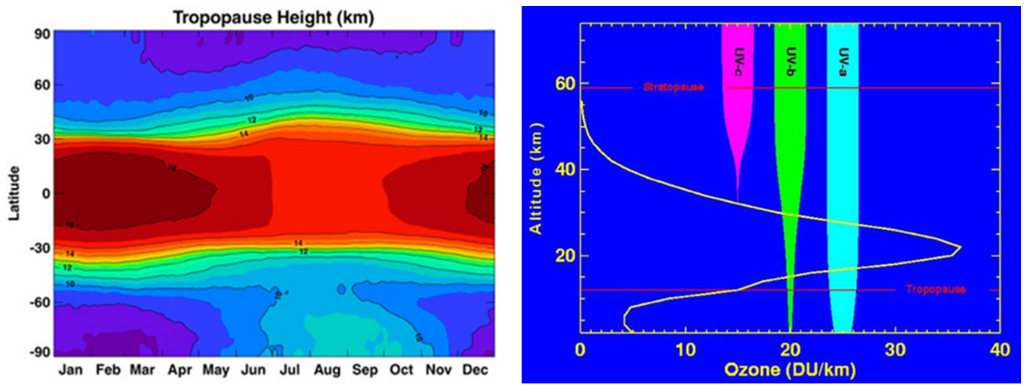

Fig. 2. *Left panel:* tropopause height as a function of latitude and time; the height decreases from about 17 km at the equator to less than 8 km at the poles (NASA 2000). *Right panel:* solar radiation and ozone; the absorbed radiation is UV-C (200-280 nm) and most of UV-B (280-320 nm), while most part of UV-A (320-400 nm) reaches the ground (NASA 2000).

ing about 12 km (the upper border is termed tropopause (Fig. 2, left panel).

In the stratosphere, the region between about 12 km and 50 km, the temperature increases with altitude; most of the ozone (90%) is located there, and the ozone concentrations are greatest between 15 and 30 km (Fig. 1).

A characteristics of the troposphere related to its temperature gradient is the convective instability, and this play a major role for most of the observed weather. On the contrary, the

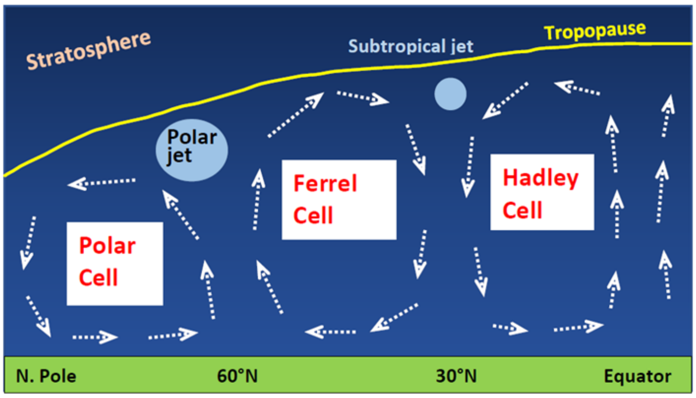

Fig. 3. A schematic representation of the three large-scale circulation cells and the jets in the troposphere of the northern hemisphere. The Coriolis effect cause poleward-moving mass to deviate to the East, and equatorward-moving mass to deviate to the West, therefore the winds at the base of the Polar, Ferrel and Hadley cell are the polar easterlies, the westerlies and the trade winds (easterlies), respectively.

stratosphere is convectively stable and vertically stratified, so the vertical motions are often small, while the horizontal motions can be significant [2].

Though it is only a minor constituent of air even in the stratosphere, ozone is the main responsible for its warming and different temperature gradient, since it absorbs solar UV radiation and reemits it in all directions as thermal longwave radiation (see the *Chapman cycle* in Section 4). The absorbed radiation is UV-C (200-280 nm) and most of UV-B (280-320 nm), while most part of UV-A (320-400 nm) reaches the ground (Fig. 2, right panel).

Essentially, it is the ozone layer that is responsible for the existence of the stratosphere. Therefore, the future trend in this part of the atmosphere will depend not only on the ozone depleting substances, such as chlorine compounds (Langematz 2019), but also on the greenhouse gases, since they appear to heat and expand the troposphere and to cool and shrink the stratosphere (Pisoft et al. 2021).

[2] In the atmosphere, both the pressure and the density of air decrease with altitude; of course, horizontal changes of temperature, pressure and air density are related to the weather system.

3. Atmosphere dynamics

The complex motion in the atmosphere, that affects the temporal and spatial distribution of ozone, is driven by the differential heating due to the Sun and by the rotation of the Earth (Coriolis force). Temperature changes are directly linked to the speed and direction of the winds, and changes in wind speed in the horizontal and vertical direction give rise to the instabilities of the weather system. The Coriolis effect cause poleward-moving mass to deviate to the East, and equatorward-moving mass to deviate to the West.

3.1. Winds, jet stream, polar vortex

In the troposphere, the atmospheric heating produces the three large-scale circulation cells in both the hemispheres, Polar (above 60° of latitude), Ferrel (between about 60° and 30°) and Hadley (between 30° and the equator) cell (Fig. 3). At about 60° the air rises to the tropopause, moves poleward and deviates toward the East. When it reaches the polar areas, it descends and is driven away toward the 60th parallel deviating westwards (polar easterlies). At the equator the air rises to the tropopause and moves toward the 30th parallel; then it descends and, moving toward the equator, it deviates toward West: these are the trade winds

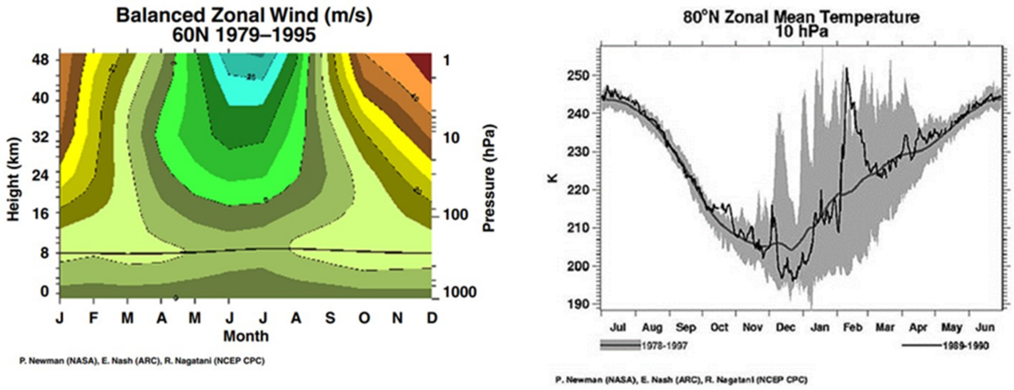

Fig. 4. *Left panel:* zonal wind in the stratosphere as a function of altitude for northern latitudes (60° N), from January to December (NASA 2000). At these latitudes, differential heating in winter is much greater than in summer. This results in sharper temperature gradients and hence stronger winter zonal winds, about 45 m/s (in November-January), than summer winds, about 20 m/s (in June-July), above 40 km of height. *Right panel:* Stratospheric Mean Temperature at 80°N at 10hPa or 10 mb. Note the sudden stratospheric warming in February 1990 (NASA 2000).

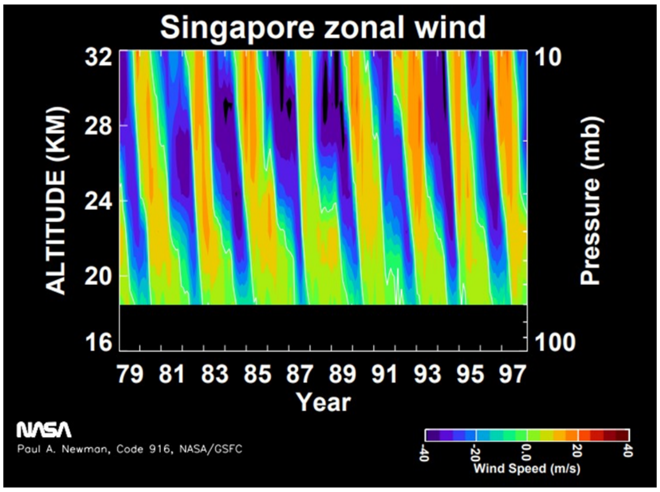

Fig. 5. Zonal winds for tropical latitudes (Singapore data) showing the Quasi-Biennial Oscillation effect in the stratosphere (NASA 2000).

or easterlies. At about 60° of latitude part of the rising air moves toward lower latitudes, it descends at the 30th parallel and then moves poleward deviating toward East: these are the westerlies. While the trade winds are permanent, the westerlies are stronger in winter and weaker in summer, since they are linked to the tropospheric polar vortex (Waugh et al. 2017), and they are particularly strong in the southern hemisphere, where, differently from the northern one, there are large tracts of open ocean. Jet streams are located near the tropopause and are westerly winds; the strongest ones are: *a)* the polar jets located at an altitude of 9–12 km at the interface between Polar and Ferrel circulation cells, and *b)* the somewhat weaker subtropical jets at 10–16 km near the boundary of the Ferrel and Hadley circulation cells.

The path of the jet typically has a meandering shape, and these meanders propagate eastward at lower speeds than that of the actual wind within the flow; they are known as Rossby waves (planetary waves).

In the stratosphere, the zonal wind is the component of the wind field blowing parallel to lines of latitude, while the meridional wind is the component in the North-South direction. The seasonal variation of zonal winds is significant in those places (and altitudes) where the temperature change during the course of the year is large, e.g. at the latitude of 60°, in comparison to the lower variability at the equator (Fig. 4, left panel). In the middle to upper stratosphere, during winter, the polar night jet sets up along the zone of greatest temperature change along the polar night terminator (north of the Arctic Circle, or south of the Antarctic Circle). The region poleward of the polar night jet is known as the stratospheric polar vortex. It is a region of air isolated (by the polar night jet) from the rest of the stratosphere, and where extremely cold temperatures develop. The polar vortex breaks down in spring and vanishes in summer, and it is during spring that a large decrease of ozone occurs in the polar region (ozone hole in Antarctica). One should note that the tropospheric polar vortex is much larger than the stratospheric one (and the two are not directly connected; Waugh et al. (2017)); it is poleward of the polar jet stream, and, differently from the stratospheric polar vortex, it is permanent, but it is stronger in winter and weaker in summer.

During winter, the cold stratosphere can rapidly warm up in a very short time period, usually by 5 – 10 °C, and more rarely up to 30 °C, in only a few days; total ozone rises substantially over these same days (Fig. 4, right panel). The warmings are the result of the rapid displacement of the polar vortex from a roughly symmetric circulation about the pole to a circulation that is offset from the pole. Midwinter warmings are fairly common in the northern hemisphere, while in the southern hemisphere they are quite rare (due to the different topography).

3.2. Waves and QBO

The atmosphere exhibits many wave-like motions with a variety of space and time scales ranging from slow moving planetary scale waves to much faster and smaller gravity waves. Waves are responsible for asymmetries in the polar vortex, stratospheric sudden warmings, mixing of polar vortex air with mid-latitude air, the forcing of the Quasi Biennial Oscillation and the control of the mid-latitude mean meridional circulation.

Planetary scale, stationary Rossby waves are forced by topography, i.e. the large-scale features like the Himalaya-Tibet complex, by land-ocean heating contrasts, and by instabilities arising from horizontal and/or vertical gradients in the temperature and wind distributions. After being generated in the troposphere, planetary waves propagate into the stratosphere, growing in size as they move upward. Because of the hemispheric asymmetries (such as the much greater land area and the more extensive mountain ranges in the northern hemisphere), the wave energy is significantly larger in the northern stratosphere. These large scale waves have the largest influence in the winter hemisphere stratosphere outside of the equatorial region.

The direction of the winds in the tropical stratosphere is observed to reverse from easterly to westerly and back to easterly again approximately every 26 to 28 months: it is the Quasi Biennial Oscillation (QBO). The phase of this oscillation depends on the altitude: the winds blowing in one direction descend in altitude in time and are replaced by winds blowing in opposite direction (Fig. 5) [3]. Convective activity and other forcings in the tropics generate a variety of atmospheric waves, some of which propagate vertically from the troposphere and dissipate in the stratosphere; the effect is an alternating wind acceleration or deceleration depending on the easterly or westerly wind in the lower stratosphere. While it is clear that similar waves are responsible for the QBO, their ultimate source and characteristics remain

[3] The QBO winds are fairly regular, however a breakdown was observed in early 2016 (Osprey et al. 2016), (Lapillonne et al. 2020).

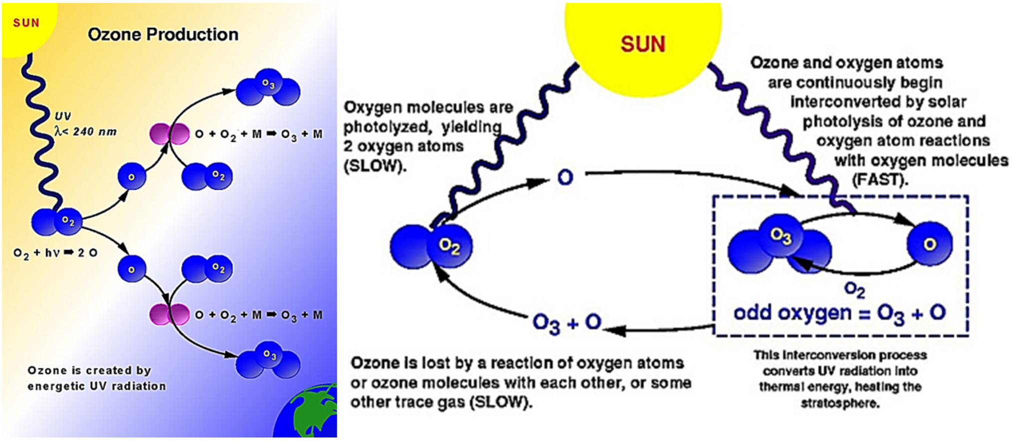

Fig. 6. *Left panel:* ozone photochemical production (NASA 2000). *Right panel:*the Chapman ozone life cycle (NASA 2000).

elusive. The vertically propagating Kelvin and mixed Rossby-gravity waves were firstly proposed as the source, and then gravity waves alone have been suggested; see e.g. Giorgetta et al. (2002). Although QBO is mainly a tropical phenomenon, its effects are felt well beyond the tropics; however, there is no well accepted explanation for how the equatorial QBO is transmitted to extratropical latitudes.

4. Ozone formation, destruction and distribution

The distribution of ozone is determined by the photochemical processes of creation (production) and destruction (loss), and by the transport into or out of the region. At the tropical latitudes close to equator, in the upper stratosphere photochemical processes control most of the ozone budget, since the lifetime of an ozone molecule is very short compared to the length of transport processes. The transport still plays an indirect role there, because of its impact on the temperature structure and hence on the ozone amounts. It is in the lower stratosphere, where there is much less UV light and ozone lifetimes are much longer, that transport is fundamental for determining the ozone distribution (Fig. 6).

a) *Chapman cycle*. An ozone molecule's life begins when an intense UV radiation (wavelength less than 240 nm) breaks apart an oxygen molecule; the two atoms react with other two oxygen molecules to form two ozone molecules. The rate at which ozone is formed is slow. The ozone molecule spend most of its life absorbing UV radiation. The UV photon breaks the ozone molecule into an oxygen molecule and an oxygen atom, then there is the recombination of the atom with another oxygen molecule to reform ozone, and in this process thermal energy is emitted. In other words, ozone and oxygen atoms are continuously interconverted by solar photolysis of ozone and oxygen atom reactions with oxygen molecules; this interconversion process converts UV radiation into thermal energy, heating the stratosphere.

b) *Ozone destruction*. The loss of ozone is a natural process resulting from normal levels of gases such as methane, nitrous oxide, methyl bromide and methyl chloride. The gases, however, are emitted also by human activities; methane, for example, is emitted by natural wetlands and by rice paddies and biomass burning. In the case of CFCs, the energetic UV radiation above most of the ozone layer breaks down the molecule, freeing the chlorine. Under the proper conditions, this chlorine has the potential to destroy large amounts of ozone (over its lifetime in the stratosphere, a Cl atom can destroy about 100.000 ozone molecules); eventually, the Cl atoms react with methane to form HCl. c) *Tropospheric ozone*. We just recall that tropospheric ozone is a pollutant found in high concentrations in smog; it is produced also by biomass burning (jungle, savannas, and farm land). It is formed by the interaction of sunlight with hydrocarbons and nitrogen oxides. In urban areas, high ozone levels usually occur during warm summer months.[4] There is an exchange of ozone between troposphere and stratosphere which is discussed in the subsection *Stratospheric-Tropospheric Exchange*.

4.1. Brewer-Dobson circulation

Ozone production occurs mainly at tropical latitudes near the equator, but most ozone is found outside the tropics, and there is relatively few ozone at equatorial latitudes (Fig. 7, left panel). A very slow circulation, known as Brewer-Dobson circulation, moves the ozone from the tropics into the middle and polar latitudes. The simple model consists of three part. The first part is rising tropical motion from the troposphere into the stratosphere; it is very slow since the time needed to lift an air parcel from the tropical tropopause (near 16 km) to 20 km is about 4 - 5 months. The second part is poleward transport in the stratosphere. The third part is descending motion in both the stratospheric middle and polar latitudes. Poleward of about 30°N and 30°S, the circulation becomes downward as well as poleward, and it tends to increase ozone concentrations in the lower stratosphere of the middle and high latitudes. The cooling of air is accompanied by a sinking motion, and it is this sinking motion that establishes the meridional overturning (for mass continuity reasons) from equator to pole in the winter hemisphere. The mechanism behind the Brewer-Dobson circulation is complex, and it results from wave motions in the extratropical stratosphere, that is from the Rossby standing planetary waves. They either remain stationary or move slowly westward, and they eventually propagate vertically into the stratosphere. So while the Brewer-Dobson circulation cell is created due to mass continuity requirements, its existence is due to the breaking of planetary waves into the winter hemisphere polar stratosphere. That is, the Brewer-Dobson cell is a winter time circulation, and it is almost nonexistent in the summer hemisphere.

The midwinter Brewer-Dobson circulation cells in the southern and northern hemispheres are quite different, owing to the hemispheric differences in planetary wave forcing coming out of the troposphere, related to the different topography. The southern hemisphere has significantly less land than the northern hemisphere and is almost entirely ocean from 55°S to the Antarctic continent. The northern stratosphere, therefore, has more frequent and intense planetary wave activity (and stronger Brewer-Dobson circulation) during the northern winter than the southern hemisphere stratosphere during the southern winter. The weak winter wave activity in the southern hemisphere means that the Antarctic polar vortex is much more isolated than its Arctic counterpart, and as a result, temperatures in the Antarctic polar vortex get extremely cold.

4.2. Ozone distribution

The concentrations of ozone change with time and latitude, owing to the Brewer-Dobson circulation (Fig. 7, right panel).

At northern mean latitudes (45°), the ozone tends to increase from around November from a minimum value of about 250 DU (Dobson Unit), and to decrease from around April from a maximum value of about 450 DU; the global average of total ozone is 300 DU [5].

At southern mean latitudes (45°) the trend change occurs in March (about 240 DU) and October (about 400 DU), respectively; the values are lower than those of northern hemisphere.

[4] Breathing ozone is lethal at dosage levels of a few molecules per million air molecules.

[5] The Dobson Unit is defined as the thickness, in units of 0.01 mm, of the layer of pure ozone which would be formed by the total column amount at standard conditions for temperature and pressure. Hence 300 DU correspond to 3 mm.

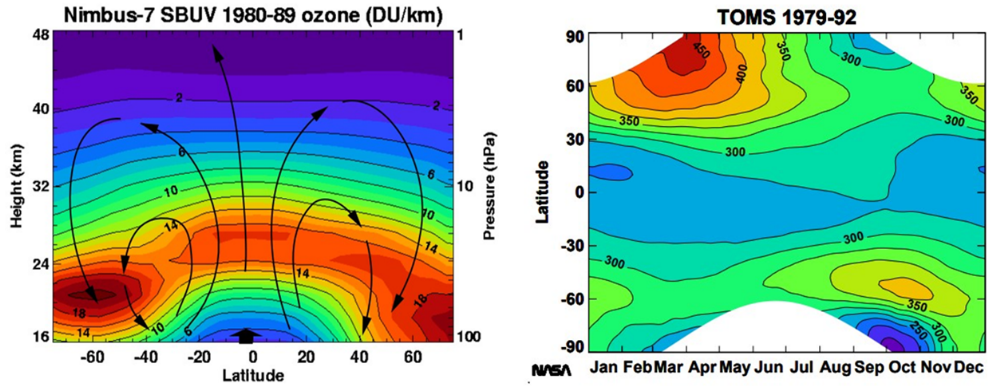

Fig. 7. *Left panel:* Brewer-Dobson circulation. Average number density of ozone plotted versus latitude and height; high levels of ozone (Dobson Units per kilometer) are found at high latitudes (above 60° S and 60° N) between 16 and 24 km; the black arrows show the Brewer-Dobson circulation which varies by season and by hemisphere (NASA 2000). *Right panel:* total ozone versus time and latitude. Average total column ozone as measured by the Nimbus-7 Total Ozone Mapping Spectrometer (TOMS); the highest levels of ozone (Dobson Units) are located in the polar regions at the beginning of the spring season (NASA 2000).

The ozone concentration is small near the equator, and it shows a double minimum and double maximum along the year (range about 230 - 260 DU).

In the Arctic region the ozone get the maximum stratospheric value by the end of March. In the Antarctica ozone does not change very much from January to July; it decreases in August-September (below 150 DU, ozone hole) and then it increases with a peak in November.

4.3. Stratosphere-Troposphere Exchange

The Stratosphere-Troposphere Exchange refers to the transport of material across the tropopause, and it has direct implications on the distribution of atmospheric ozone, in particular the decrease of lower stratospheric ozone and increase of tropospheric ozone. The transport of anthropogenic gases such as CFCs from the troposphere into the stratosphere affects the chemical balance in both regions, and provides the catalysts necessary for stratospheric ozone destruction. For time scales greater than several months, the mass flux through the tropopause is ultimately driven by the Brewer-Dobson circulation.

High pressure areas in the troposphere, that is anticyclones or blocking high, have the effect to lower the column ozone. The anticyclonic flow brings lower latitude air poleward, and the warmer temperatures associated with a blocking high cause an increase in the vertical scale of the troposphere, resulting in a lower column ozone density.

Cut-off lows are upper level cyclones which become separated from the main flow of the upper tropospheric jet stream, and they isolate air with characteristics of its polar source region (higher ozone content). Cut-off lows are related to large scale cumulus convection, and the tropospheric air can be transported across the tropopause, creating a vertically mixed region of stratospheric and tropospheric air with increased ozone content. As already stated by Dobson (1968), it is well known that there is more ozone in cyclonic (low pressure) conditions than in anticyclonic (high pressure) conditions.

Models predict for the present century a significant increase of tropospheric ozone, transported from the stratosphere across the tropopause, mainly as an effect of greenhouse gases (Langematz 2019).

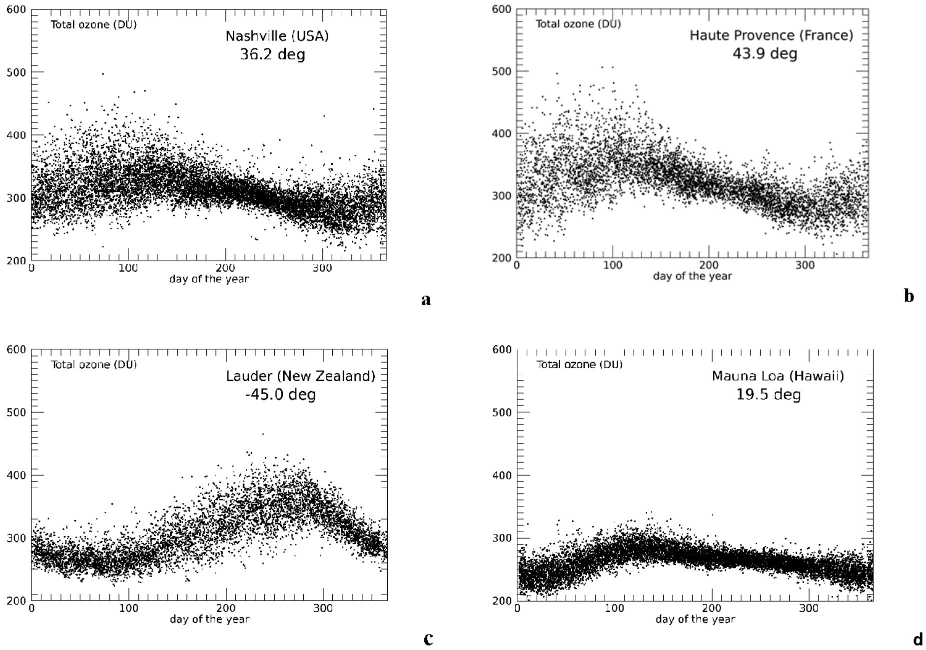

Fig. 8. Total ozone measurements at different places; data from NOAA (2021). *a*: Nashville (USA), direct Sun observations 1962-2017; *b*: Haute Provence (France), direct Sun observations 1983-2019; *c:* Lauder (New Zealand), direct Sun observations 1987-2019; *d:* Mauna Loa (Hawaii, USA), direct Sun observations 1963-2021. One should note the larger dispersion of data during the period from autumn to spring locally.

5. Total ozone and effective ozone

Many ground stations regularly perform total ozone measurements in the network of the Global Atmosphere Watch Programme (GAW). The observations are predominantly taken with Dobson and Brewer ozone spectrophotometers; the Brewer network is continuously increasing due to the installation of Brewers at newly established stations and because of replacement of Dobsons by Brewers at existing stations (Vanicek 2006). Some examples taken from the Global Monitoring Laboratory, NOAA (2021), are shown in Fig. 8 and Fig. 9.

The Dobson and Brewer ozone spectrophotometers measure total column ozone in the atmosphere by observations of direct Sun spectral irradiances of solar radiation at selected wavelengths in the UV part of the spectrum with strong and weak absorption by ozone. The total ozone values are derived by differential spectroscopy techniques; the calculation takes into account the logarithms of extraterrestrial and ground spectral irradiances measured by the instruments, the linear combinations of ozone absorption and Rayleigh molecular scattering coefficients, the relative optical air masses of the ozone layer and the entire atmosphere, and the air pressure. The linear combinations eliminate influences of the atmospheric aerosol on the observations (Vanicek 2006). In the case of observations from scattered zenith sky radiation (e.g. owing to clouds on the Sun), the total ozone is determined by means of zenith polynomials, which are empirical functions and depend on the instrument and location (Vanicek 2006). For a detailed de-

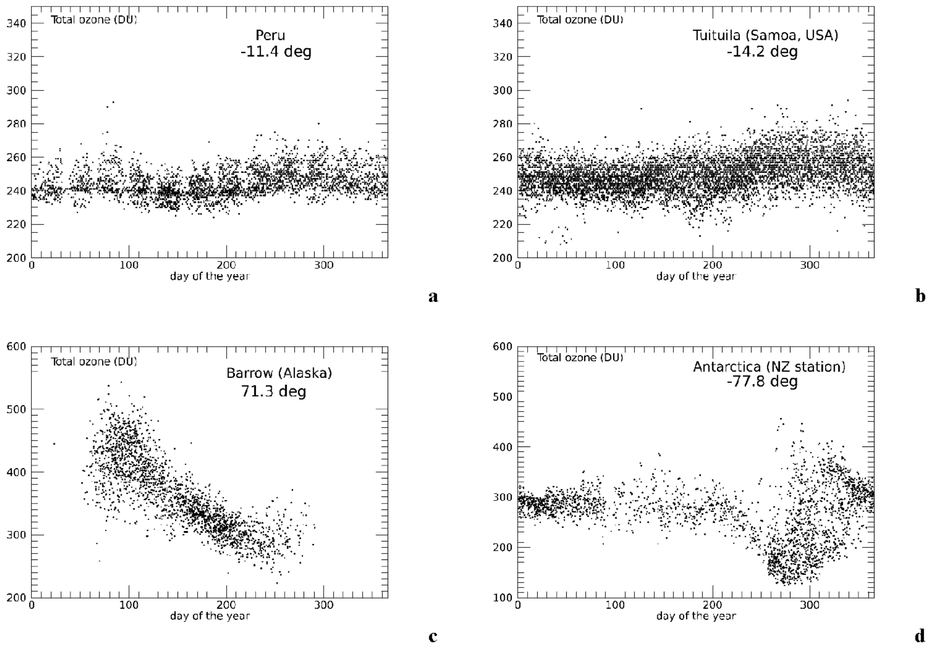

Fig. 9. Total ozone measurements at different places; data from NOAA (2021). *a:* Peru, Marcapomacocha Station, direct Sun observations 2000-2019, note the very low amplitude oscillation (with two maxima) and low ozone values; *b:* Tuituila Is. (Samoa, USA), direct Sun observations 1976-2021, note the low ozone values; *c:* Barrow Atm. Baseline Obs. (Alaska), direct Sun observations 1972-2021, note the largest values in March-April; *d:* Antarctica, Arrival Heights New Zealand Station, direct Sun or direct Moon (during winter) observations 1988-2018, note the decrease in August and the ozone hole (with very small values) in September-October.

scription see e.g. the operations handbook by Komhyr (1980).

Total ozone observations performed by space instruments such as TOMS (Total Ozone Mapping Spectrometer) on board the Earth Probe (NASA) use the backscatter ultraviolet techniques. That is, they observe solar radiation that has penetrated to the Earth's lower atmosphere and is then scattered by air molecules and clouds back through the stratosphere. A description of this and other observational techniques can be found in Chapter 7 of NASA (2000).

The effective ozone is the product of the total ozone (at the zenith) and the airmass that depends on the solar zenith angle (Nozawa et al. 2007); it is considered by researchers working on the biological effects of solar UV radiation since the possible inactivation or killing efficiency of sunlight depends of course on the position of the Sun (Furusawa et al. 1990). The available data of total ozone in the literature can be used to estimate the effective ozone taking into account the latitude of the site and the date and time (see the example in Fig. 10); an example of the killing effect of solar UVB radiation is shown in Fig. 11.

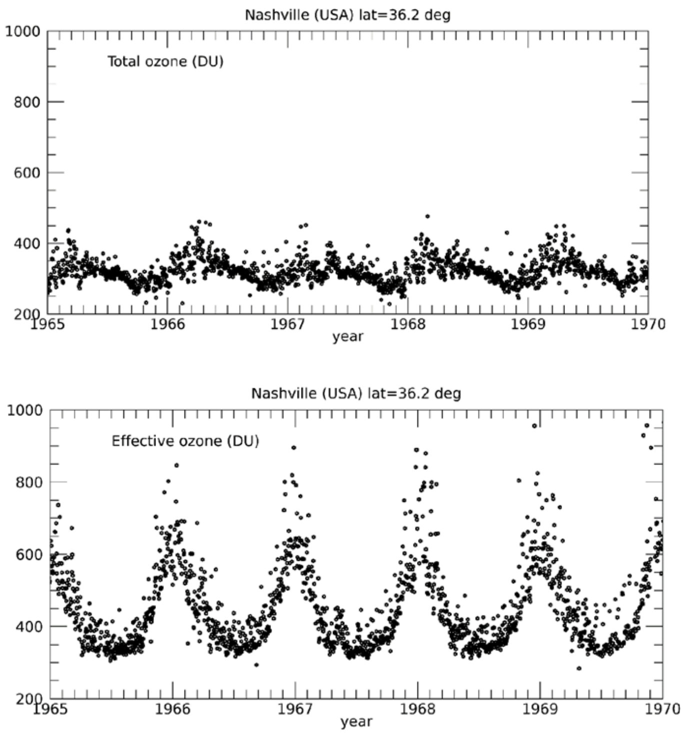

Fig. 10. Total ozone (upper panel) and effective ozone (lower panel) at Nashville (USA) from 1965 to 1970; data from NOAA (2021). Direct Sun measurements were taken usually at noon, and partly two or three hours before or after the noon.

6. Temporal behavior

The variations that are significant for the total column measurements of ozone occur in the lower stratosphere; however, we will mention also those occurring in the upper stratosphere, even though they involve only the few percent of ozone present there.

a) *Short term variability*. In the upper stratosphere, variations in ozone occur with the daily rising and setting of the Sun. When sunlight is present, production and loss are balanced, and when the Sun sets both production and loss are turned off. After sunset, there is occasionally a conversion of an oxygen atom to ozone via a three-body reaction with another molecule, so the ozone concentration increases slightly; there is reconversion to atomic oxygen at the dawn. However, all this has little measurable effect as regards the total ozone. Much larger total column ozone variations occur in the lower stratosphere, corresponding to the variations of the weather system as mentioned in the subsection *Stratospheric Tropospheric Exchange*. The observations at the stations located at mean latitudes indicate large variations during the period from autumn to late spring, that is from about September to May in the northern hemisphere (ozone can change even by 150 DU in few days), and from March to October in the southern hemisphere. There is less variability during summer maybe due to the prevalence of anticyclonic condi-

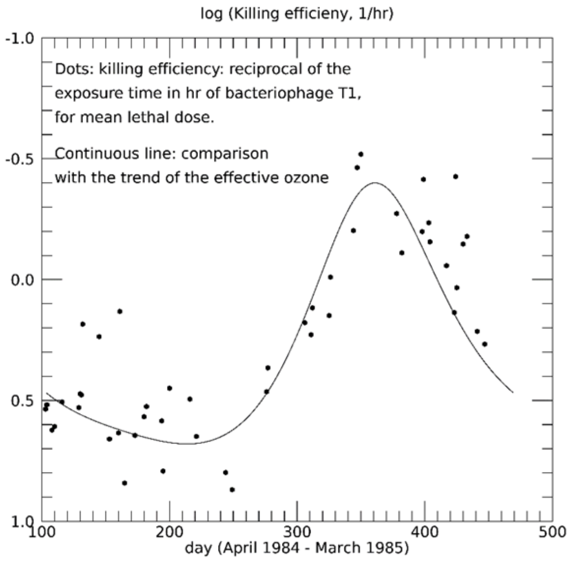

Fig. 11. Example of the effect of solar UVB radiation on viruses; *dots*: killing efficiency of bacteriofage T1 measured by Furusawa et al. (1990) in experiments carried out throughout one year at noon (approximately from 11 a.m. to 1 p.m.) in Isehara; *continuous line*: qualitative comparison with the trend of effective ozone for a latitude of 36°.

tions [6]. Small variations in the upper stratosphere are linked to the 27-day rotation period of the Sun, but more significant variations are related to solar proton events.

b) *Seasonal variability*. In the upper stratosphere the photochemical processes of ozone are sensitive to the temperature, and there is an anticorrelation between temperature and ozone concentration, depending on the effect of the Sun and the planetary waves; as already stated, this has little effect on the total ozone. Much more important are the variations in the lower stratosphere; they depend on the Brewer-Dobson circulation, which is active essentially during winter. At mean latitudes in northern hemisphere, ozone increases from about November to April, and decreases from about April to November, while in the southern hemisphere it increases from about March to October and decreases from about October to March.

c) *Interannual variability*. The year-to-year variations in the upper stratosphere are mainly related to the 11-year solar cycle and to the possible volcanic eruptions. The estimated amplitude of the solar cycle effect is about 5 DU for all the latitudes. In the lower stratosphere, the variations are mainly due to the QBO and to the El Niño Southern Oscillation (ENSO). The ozone QBO contribution usually is largest in the tropics, where the amplitude is about 12 DU from maximum to minimum.

d) *Long-term variability and trends*. In the last sixty years the most important variation in the stratospheric ozone has been its harmful depletion due the CFCs; it was followed by its (partial) recovery. There are doubts about the future trend, given the unexpected change of CFC emissions in recent years. There is of course a rich literature on the subject, see e.g. Zerefos et al. (2012), Fountoulakis et al.

[6] For the relation between stratosphere and climate extreme events see e.g. Domeisen & Butler (2020)

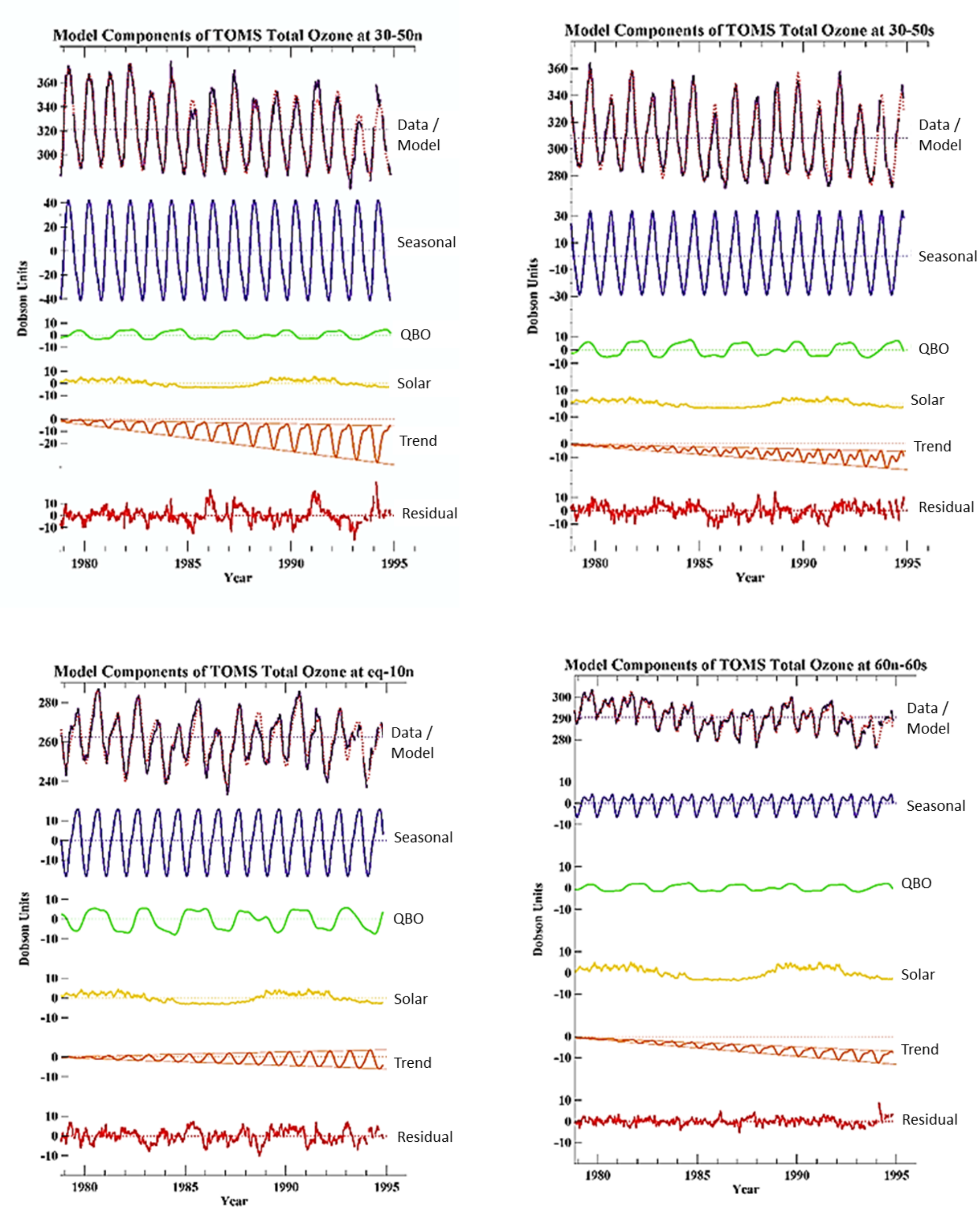

Fig. 12. Results published by NASA (2000) of the statistical regression analysis of TOMS data (monthly average time series of total ozone) performed by adopting a model with four components: seasonal cycle, QBO, 11-yr solar cycle and a long-term trend, and with a final residual. One should note the asymmetry between the northern (*upper left panel*: latitude 30°-50° N) and southern (*upper right panel*: latitude 30°-50° S) hemisphere, in terms of ozone amount and phase of seasonal oscillation. It explains the oscillations in the *lower right panel* with evident annual double maxima and minima of *Data/Model* and *Seasonal* of the difference between 60°N and 60°S. The general decreasing trend in the four panels is related to the CFC effects. QBO is stronger near equator (*lower left panel*, Equator - 10°N).

(2018), Fang et al. (2019), Solomon et al. (2020) and references therein [7]. The long-term variability of solar activity should have some influence on the upper stratosphere.

Fig. 12 shows some examples of the observed temporal variability derived from data analysis.

7. Solar activity

The modulations in the solar UV output that depend on the sunspot cycle of 11 years have a direct effect on ozone photochemistry, and it is possible to detect the effect in the upper stratosphere as interannual variability. However, the solar cycle of the total column ozone is not well quantified, and some of the variability of about 5 DU may be due to other factors, particularly aerosol contamination. During solar maxima, it is possible to detect in the upper stratosphere the modulation related to the 27-day rotation period. One cannot exclude very long-term variability, due for example to the Gleissberg cycle of about 90 years.

A solar proton event (SPE) can cause ozone depletion at polar latitudes; see e.g. Jackman et al. (2001). According to Denton et al. (2018), stratospheric ozone measurements from sites that are within the polar vortex show a decrease in ozone partial pressure, following SPEs, that is of about 5–10% and persists for about one month. No decrease in stratospheric ozone is detected following SPEs in late summer or autumn. Denton et al. (2018) conclude that the polar vortex is an essential factor for causing stratospheric ozone depletion following SPEs.

8. Aerosols and volcanoes

Sulfate aerosols are typically composed of a solution of sulfuric acid and water, at least in the middle latitudes where the temperatures are warm enough to maintain the particles in a liquid state. The sulfuric acid comes from carbonyl sulfide (COS) and sulfur dioxide (SO_2) carried into the stratosphere via tropical lifting by the Brewer-Dobson circulation, or by direct injection of SO_2 into the stratosphere from very explosive volcanic eruptions, such as El Chichon in 1982 or Mount Pinatubo in 1991. Most sulfate aerosols are carried out of the stratosphere by Brewer-Dobson circulation descent in the higher latitudes.

SO_2 is the most important of the gases emitted from explosive volcanos as far as the stratosphere is concerned. SO_2 is oxidized in the stratosphere to sulfuric acid (H_2SO_4) which coalesces into small particles or aerosols, and this takes about one month after the injection. These particles are generally much smaller than those originally emitted by the volcano. They act more like gases, and turbulent motions are sufficient to keep them in the same air mass with any remaining gas from the volcano. These aerosol particles can enhance the formation of polar stratospheric clouds, key players in the formation of the Antarctic ozone hole.

The effect of Pinatubo (the largest eruption in the past ninety years) on the total ozone was a decrease of some percent, with the largest values for northern latitudes. According to Angell (1997), ozone decreased by 8% in Europe. However, there are significant differences among the results obtained by the various authors; see Angell (1997) for a comparison.

9. Ozone hole

During August and September periods, ozone decreases quite rapidly over Antarctica region. While there have been severe losses of ozone over the Arctic, usually there is not a similar 'hole' over the North Pole. Although very rare, an ozone hole occurred effectively in the Arctic in March 2020 (Witze 2020).The losses are directly caused by chlorine and bromine catalytic reactions, that are accelerated on the surfaces of cloud particles in the stratosphere. The CFCs that destroy ozone are distributed throughout the atmosphere. However, the holes form as a result of exceptionally cold temperatures which occur during the winter season over the polar regions. These cold temperatures isolate air in the polar regions from ozone rich air at lower latitudes and allow formation of

[7] See also the papers on *Renewed emissions of ozone depleting substances* (26 July 2021) in https://www.nature.com/collections/ggcbbedahg/

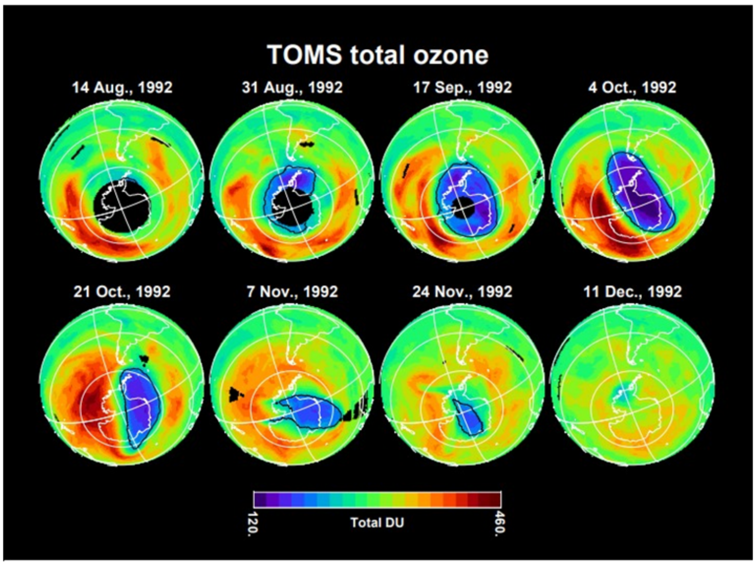

Fig. 13. TOMS Southern Hemisphere total ozone images for August-December 1992. Note the extent of the hole on 4 Oct. 1992 to South America (NASA 2000).

polar stratospheric clouds that enhance ozone destructive processes.

The effects of the hole are not limited to Antarctica because of the change of its shape from the approximately symmetrical one centered on the South Pole (Fig. 13). A typical pattern is the elongation of the ozone hole that slowly rotates eastward, and it can cover the tip of South America (Nozawa et al. 2007); its harmful effects at Punta Arenas were remarked for example by Abarca & Casiccia (2002). After the breakdown of the polar vortex, the hole breaks up, and parcels of ozone depleted air mixed with mid latitude air move northwards, occasionally causing a reduction in ozone values [8].

10. Atmosphere and UV radiation

In this section we recall some other atmospheric effect on UV radiation. Calbó et al. (2005) have reviewed the empirical studies of cloud effects, and found that the ratio between measured UV radiation in a cloudy sky and calculated radiation for a cloudless sky range from 0.3 to 0.7, depending both on cloud type and characteristics. However, it is known that hazy/cirrus sky can have an enhancing UV effect, up to 8% according to Sabburg & Wong (2000), or maybe more (Calbó et al. 2005). Enhancements are explained by reflections in cloud surfaces and increased forward scattering in some types of clouds; a review of cloud enhancement effects in the UV band is given by Parisi et al. (2004).

Many works have been devoted to the analysis of the atmospheric effects during several years on UV-B and total solar radiation, by considering ozone, water vapour, clouds, and aerosols; just as an example, we recall the studies by El-Nouby Adam (2014) in Egypt, Lee et al. (2019) in Korea, Eerme et al. (2015) in Estonia [9]. Fountoulakis et al. (2018) compared the results obtained in Canada, Europe and Japan, and found that in northern hemi-

[8] Such as in southern Australia; Bureau of Meteorology, http://www.bom.gov.au/uv/faq.shtml

[9] At summer solstice, Eerme et al. (2015) observed that the shortwave threshold in the UV-B range was 294 nm in normal column ozone and atmospheric transparency conditions; in early morning and late evening the shortwave threshold dropped to approximately 310 nm.

sphere, the long-term changes in the UV-B radiation vary greatly over different locations, and the main drivers are changes in aerosols and total ozone. At higher latitudes, part of the changes may be attributed to the surface reflectivity and clouds; however, since the connection between the changes in UV-B irradiance and those in the different factors is not clear, it is obviously essential to reduce the uncertainties in the measurements and to improve the understanding of the interactions between solar UV radiation and the related geophysical variables. Wild et al. (2021) have discussed the 'global dimming and brightening', that is, the solar radiation decrease and increase at the Earth surface in the past decades; as regards central Europe, aerosol pollutants appear to play a crucial role, since the variations occurred also in cloud-free clear sky conditions.

The "amplification factor" is the percentage increase in the biologically active UV irradiance that would result from 1% decrease in the column amount of atmospheric ozone (Durzan & Smertenko 2005). In the case of the Pinatubo eruption, Zerefos et al. (2012) noted that there was an enhancement of the amplification factor of UV-B as a result of enhanced scattering processes caused by the volcanic aerosols, and some years later there was an increase of total ozone and *increasing* trend in UV irradiance, a "paradox" caused by the decline of the aerosol optical depth.

11. UV radiation and pandemics

Ultraviolet radiation in sunlight is the primary virucidal agent in the environment (Sagripanti & Lytle 2007), hence a better understanding of this ability of UV radiation could reveal a link between solar UV and seasonal influenza. Seasonal cyclicity is a ubiquitous feature of acute infectious diseases and may be a ubiquitous feature of human infectious diseases in general (Martinez 2018); each acute infectious disease has its own seasonal window of occurrence, which may vary among geographic locations and differ from other diseases within the same location. However, one may say that the seasonality of influenza is a phenomenon that has eluded explanation throughout history.

Several papers have claimed to find a significant association between pandemics and physical effects such as solar activity or volcanism. Towers (2017) in particular has performerd a careful statistical assessment of the purported association between sunspot activity and the timing of influenza pandemics. In all cases, he found no statistically significant evidence of any association. Towers (2017) underlines that the faults found in the past analyses are common pitfalls, and that inattention to analysis reproducibility and robustness assessment are common problems in the sciences ("unfortunately not noted often enough in review"). Therefore, an accurate statistical analysis is mandatory in all cases when studying the physical or atmospheric (e.g. temperature and humidity) effects on pandemics.

According to Sagripanti & Lytle (2020), SARS-CoV-2 should be inactivated relatively fast during summer in many populous cities of the world, indicating that sunlight should have a role in the occurrence, spread rate and duration of coronavirus pandemics; this view is supported by the model developed by Nicastro et al. (2021). Therefore, it is important to improve our understanding of the possible association between UV radiation and pandemics by performing focused experiments and measurements such as those carried on by Italian researchers.

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