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MIDDLE ATMOSPHERE TEMPERATURE AND DYNAMICS AS REVEALED FROM D-REGION OBSERVATIONS

A. D. Danilov

Institute of Applied Geophysics
Glebovskaya 20-b
Moscow 107258, USSR

The concept of so-called meteorological control of the ionospheric D-region is presently undergoing development (see, e.g., TAUBENHEIM, 1983; DANILOV and TAUBENHEIM, 1983). According to this concept the electron concentration in this region is governed not only by solar and geomagnetic parameters (W, χ, Ap, etc.) but strongly depends on the temperature and dynamical regime of the mesosphere and stratosphere. The aim of this paper is to consider how the above connection between D-region and meteorological parameters can be used to obtain some information about middle atmosphere temperature and dynamics. For this purpose it is worthwhile summarizing briefly the essential points of the meteorological control concept.

The best illustration of the meteorological control presents the well-known phenomenon of Winter Anomaly in radio wave absorption (WA). There are two components of the WA. Average absorption, L, in the SW-band in winter is higher than that in summer ("normal" WA component). On some winter days ("anomalous" days of WA) L is much higher than on previous and following days. Since the increase of L is due to the enhancement of electron concentration, [e], at altitudes 75-85 km (FRIEDRICH et al., 1979; OFFERMANN, 1979), WA provides a good example of [e] variations in the D region, which are not directly connected with any changes in solar zenith angle χ, solar or geomagnetic activity.

Using a data bank of rocket [e] measurements, compiled by DANILOV and LEDOMSKAYA (1983a), it was pointed out by DANILOV et al. (1982), that there is one more difference in [e] behavior in the upper D region between summer and winter. If we look to the electron concentration at fixed altitude in summer there is nearly no day-to-day scattering of the data and the variation with χ is well pronounced. In winter, however, even for undisturbed days (non-anomaly days) there is strong day-to-day scattering of [e]. This is illustrated in Figure 1, where rocket data for 80 km height taken from the above-mentioned data bank, only for quiet geomagnetic conditions (Kp<4) and middle geomagnetic latitudes, are presented.

To explain the effects of the meteorological control mentioned above one should examine possible mechanisms of neutral atmosphere influence on the electron concentration. The principal approach to the problem is rather simple. The electron concentration in the upper D region is governed by the photochemical equilibrium equation.

$$q = [e]^2 \alpha_{eff} \qquad (1)$$

where q is the ionization rate and α_{eff} the effective recombination coefficient. Thus changes in [e] can be due to variations of either q, or α_{eff} or both. To reveal the mechanisms of meteorological control it is necessary to look for the dependence of q and α_{eff} on the meteorological situation in the strato-mesosphere.

It is widely known (MITRA and ROWE, 1972; MITRA, 1974; DANILOV, 1975) that the effective recombination coefficient depends on the positive ion composition:

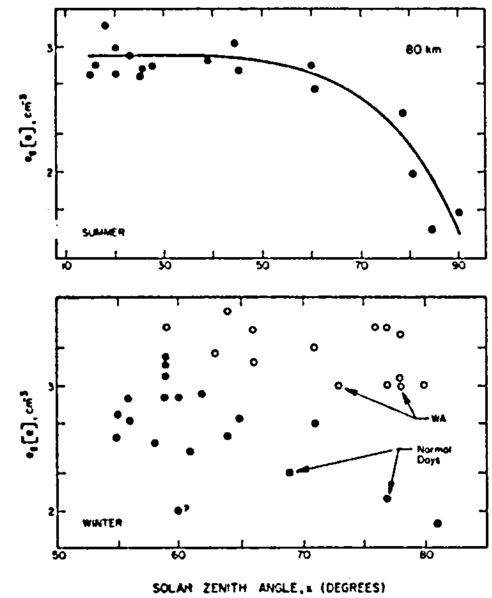

Figure 1. Electron concentration versus solar zenith angle χ at 80 km (DANILOV et al., 1982) for quiet conditions (Kp < 4) and middle geomagnetic latitudes (Λ = 30-52°).

$$\alpha_{eff} = \frac{\alpha^*(NO^+, O_2^+) + f^+ \alpha^*(clust^+)}{1 + f^+} \qquad (2)$$

f^+ being an ion composition parameter, $f^+ = [clust^+]/[NO^+ + O_2^+]$, and $\alpha^*(NO^+, O_2^+)$, $\alpha^*(clust^+)$ the dissociative recombination coefficients for normal and clustered positive ions, respectively. Though the constants α^* themselves depend on neutral temperature as T^{-n}, with n = 0.5 to 1, the major effect on the meteorological control through α_{eff} is due to f^+ variations.

It was shown by DANILOV and SIMONOV (1981, 1982), that f^+ has a well pronounced seasonal variation: at fixed altitude f^+ is higher in summer than in winter, and that this change of f^+ accounts for a variation of the effective recombination coefficient, the latter being under quiet conditions in summer 1.5 to 2 times higher than in winter. This seasonal effect in α_{eff} explains the existence of the normal component of the WA (average [e] in quiet conditions being about a factor of 1.5 higher in winter than in summer).

The reason of f^+ seasonal variation is in the strong temperature dependence of the formation rate of clustered ions (DANILOV and TAUBENHEIM, 1983; SMIRNOVA et al., 1983). Reactions, leading to cluster ions, first of all the reaction

$$NO^+ + N_2 + N_2 \rightarrow NO^+ \cdot N_2 + N_2 \qquad (3)$$

have an inverse temperature dependence, whereas most of the reverse reactions destroying clusters have a direct dependence on temperature. As a result, the net effectiveness of cluster formation from NO^+ ions $B(NO^+)$ has a strong inverse temperature dependence. Figure 2 from SMIRNOVA et al. (1983) shows the life time, $\tau(NO^+) = 1/B(NO^+)$, as a function of temperature according to ARNOLD and KRANKOWSKY's (1980) ion composition data and to calculations based on REID'S (1977) theoretical scheme. One can see, that the experimental and the theoretical data fit each other and lead to a very strong temperature dependence of the type of $B(NO^+) \propto T^{-14}$.

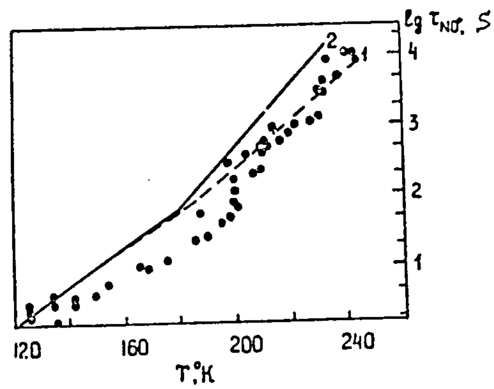

Figure 2. Life-time of NO^+ ions against clustering ($\tau(NO^+) = 1/B(NO^+)$) versus neutral temperature. Points:data from in situ ion composition measurements (ARNOLD et al., 1980), solid curve: calculations, based on theoretical scheme of ion transformation with $(H_2O) = 10^{-6}(M)$, dashed curve:same with $(H_2O) = 5 \times 10^{-6}(M)$ (Smirnova et al., 1983).

It is pertinent to stress here that such a strong temperature sensitivity of cluster formation efficiency is a vital fact for the existence of a strong temperature control of the electron concentration.

Recently, the annual variations of the efficiency B of cluster ion formation were calculated using a theoretical scheme and temperatures from CIRA model (SMIRNOVA et al., 1983). The results are shown in Figure 3 (curves) together with values of B estimated from experimental data of ion composition. As it is seen from Figure 3, theoretical calculations based on CIRA model well reproduce the observed features of the annual behavior of B. Summer B values are higher than winter ones, the amplitude of summer-to-winter variations being larger at high latitudes.

The arrows in Figure 3 show variations of B if T is changed by ±20° from the CIRA values. It is seen that these relatively small changes in T lead to rather strong effects in B. Further estimates show that these B variations both in summer and winter produce nearly equal changes of f^+. But in summer the variations in f^+ virtually do not influence α_{eff} (and so [e]) because $f^+ >> 1$ so that $\alpha_{eff} = \alpha^*(clust^+)$. In winter (when f^+ is of the order of 1) the effective recombination coefficient is very sensitive to f^+ variations so that changes in f^+ induced by temperature variations would lead to [e] variations of the order of factor 2 at middle latitudes and factor 3 in the auroral region.

Thus, day-to-day variations of the electron concentration in winter mentioned above can be accounted for in terms of day-to-day temperature variations in the mesopause region, which are quite able to reach 15-20°. An essential point should be stressed here. The difference in the ion composition between winter and summer conditions leads to quite a different reaction of [e] to small variations of the neutral temperature. So these variations would make nearly no effect in summer, but produce considerable day-to-day variations in winter. During WA events usually a temperature increase is observed which is responsible for the observed decrease of f^+ during WA (see below).

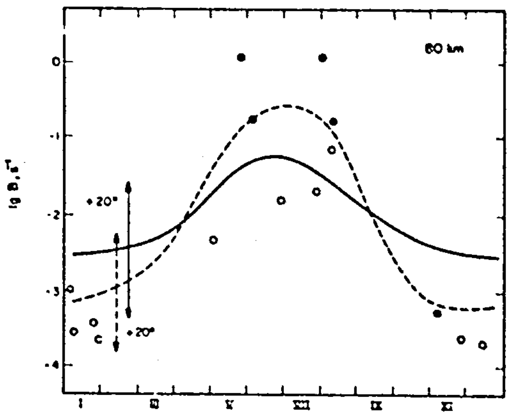

Figure 3. Annual variation of the effectiveness B of clustered ion formation at 80 km (Smirnova et al., 1983). Dots: values of B, calculated from simultaneous data on electron concentration and ion composition. Solid and dashed lines: annual variation of B at middle (≈ 40°) and high (≈ 70°) latitudes, respectively, calculated with detailed model of clustered ions formation using the temperature from CIRA model. Arrows show the limits of B if ±20° temperature deviations from CIRA values are used in the calculations.

The role of the humidity variations on B values also has been considered by SMIRNOVA et al. (1983). The curves in Figure 3 are calculated for $[H_2O] = 10^{-6}[M]$. If the H_2O mixing ratio is enhanced to 5×10^{-6}, the corresponding enhancement of B is higher in winter (about factor 4) and lower in summer (factor 1.5-2).

Resuming, the above calculations show that the experimentally observed seasonal variations of the ion composition (which, as shown above, leads to the systematic summer-to-winter difference in the electron concentration) can be accounted for by seasonal changes of the mesospheric temperature. The reasonable assumption about the existence of day-to-day temperature variations of ±20° explains the scatter of the [e] data in winter. During WA events the observed temperature enhancement can well account for the detected decrease of f^+ and thus contribute to the [e] increase. Atmospheric temperature is therefore an important parameter strongly influencing the upper D-region ionization-recombination cycle through the channel:

$$T \rightarrow B \rightarrow f^+ \rightarrow \alpha_{eff} \rightarrow [e] \rightarrow L$$

It is widely known that the principal ion production process in the upper D-region in absence of solar flares and corpuscular intrusion is photoionisation of nitric oxide molecules by solar Lyman-α emission. Since nitric oxide is not practically produced in the midlatitude D-region, one may expect that the influence of the chemical processes and the temperature on its concentration should be rather weak. But [NO] should be very sensitive to changes in strato-mesospheric dynamics because it is transport processes that govern the nitric oxide distribution in the D-region. The NO influence on the D-region behavior is strongly confirmed by the fact that during WA events nitric oxide concentration at 70-90 km has been found to be much higher than during normal days (BERAN and BANGERT, 1981).

Calculations, based on rocket experiments in which [e] and the ion composition have been measured simultaneously (DANILOV et al., 1982), also show that in order to account for strong increase of [e] during WA events one has to assume (in addition to the decrease of α_{eff} due to decrease of f^+) an increase of [NO] up to a factor of 10. The same estimates, on the other hand, show no pronounced difference between average q values for quiet summer and winter days, which means, that for quiet conditions there is no regular seasonal difference in [NO].

The increase of nitric oxide concentration on some winter days giving rise to WA events is the strongest known manifestation of meteorological influence on the D region. Knowing the mechanisms of the influence, we would be able to get information about the dynamics of the mesosphere and lower thermosphere from D-region observations.

There are two main processes which are able to supply NO to the midlatitude D-region: downward transport from the E region due to eddy diffusion and mean motion, and equatorward transport of NO molecules from the high latitude D-region, where those molecules are formed because of corpuscular intrusion. Estimates show (LAUTER et al., 1976; DANILOV and LEDOMSKAYA, 1982), however, that in order to get a sufficient effect on midlatitude [NO] through the second process one has to have rather strong latitudinal gradient of [NO] of about a factor 10 between mid- and high latitudes, which does not agree with satellite observations. Nevertheless, some correlation between [e] in the D-region and equatorward horizontal circulation has been found (GELLER et al., 1976; HESS and GELLER, 1978).

Thus, the main source of NO in the midlatitude D-region is downward transport from the E-region, where NO molecules are produced in the ion reactions. Concentrations of nitric oxide at 70-90 km depend strongly on the value and the vertical profile of the eddy diffusion coefficient, K_t. Figure 4 shows [NO] distributions calculated by DANILOV and LEDOMSKAYA (1983b) with various assumptions about the K_t profile. It is seen from Figure 4 that by varying K_t in reasonable limits of 10^5-$2\times10^7 cm^2s^{-1}$ one can get any nitric oxide values in the interval 10^6-$10^7 cm^{-3}$, which means that one can account for any experimentally observed NO variations including increase of [NO] by a factor 3-5 in WA conditions. Not accounted for are only very high [NO] values of the order of 10^8-$10^9 cm^{-3}$ reported by some authors during strong WA events. To explain these values in terms of increased eddy diffusion one has to assume very high K_t values above $2\times10^7 cm^2s^{-1}$ which seem to be unrealistic. Further estimates show that enhanced mean motions are effective in transporting [NO] molecules downward to the D-region, when turbulence is weak (K_t low), but cannot provide essential input to NO downward flux when K_t is high.

Therefore, turbulence provides the major dynamical mechanisms of meteorological control, influencing nitric oxide distribution, thereby changing the ion production rate q. Thus the eddy diffusion coefficient K_t might be in principle considered as one of the "governing" meteorological parameters for the D-region.

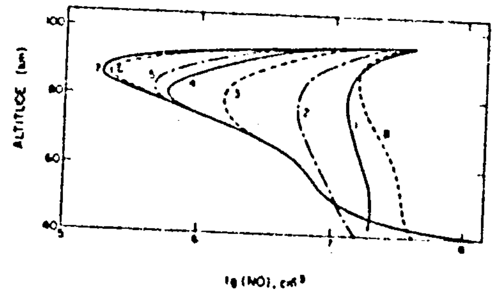

Figure 4. Calculated nitric oxide profiles (DANILOV and LEDOMSKAYA. 1983), using various values and vertical profiles for eddy diffusion coefficient K_t. Curves labeled 1,2,3,4,6 are exponential profiles with K_t(80 km) = 2×10^5, 1.2×10^5, $1.8 \times 10^7 cm^2 s^{-1}$, respectively. In these calculations a constant value of $0.5 cm\ s^{-1}$, respectively. Curves 5,7,8 are profiles with maximum at 90 km and K_t(80 km) = 2×10^5, 1.2×10^5, $1.8 \times 10^7 cm^2 s^{-1}$, respectively. In these calculations a constant value of $0.5 cm\ s^{-1}$ was accepted for the velocity of the mean downward motion.

Unfortunately our knowledge about the turbulence regime in the middle atmosphere, its nature, sources (e.g., wave dissipation, wind shears), characteristic scales, seasonal variations etc., is rather scarce. Valuable information can be found from mass-spectrometer measurements of the turbopause level h_t (DANILOV et al., 1979, 1980). In particular it was found that h_t, supposed to be inversely proportional to K_t, is decreasing with the increase of the neutral temperature at 120 km (DANILOV et al., 1979). This result initially obtained for high latitude measurements, has recently been confirmed for mid-latitude flights. It has been interpreted by KALGIN and POKHUNKOV (1981) as confirmation of the theoretical statement that the value of K_t near the turbopause should be inversely proportional to the temperature gradient between the mesopause and the turbopause:

$$K_t \approx (dt/dh)^{-1}_{80-120\ km.}$$

If so, we can try to connect temperature variations observed in WA events with the suggested increase of turbulence. In fact, temperatures at the mesopause level during WA days are 20-50 K higher than during quiet days (e.g., OFFERMANN, 1979; OFFERMANN et al., 1979), but there is no increase of T at turbopause heights, sometimes even some decrease is detected. That means that on WA days dT/dh at 80-120 km is much lower than on normal days, which can account for higher K_t values and consequently for higher [NO] in the D-region. This conception is schematically illustrated in Figure 5.

Now, the very important question is for the causes of the above increase of the temperature, and whether this increase is the reason for the enhancement of turbulence as described above, or both temperature and K_t variations are due to the same initial source. This question is still open, but there is a strong temptation to connect both facts with the dissipation of the wave energy in the upper mesosphere. Unfortunately we still know very little about the mechanisms of wave propagation through the middle atmosphere, about the conditions of its energy dissipation etc., so at the moment it is difficult even to identify these waves - internal gravity, or planetary waves. But it seems worth mentioning (see e.g., DICKINSON, 1968; GELLER, 1981) that the penetration of planetary

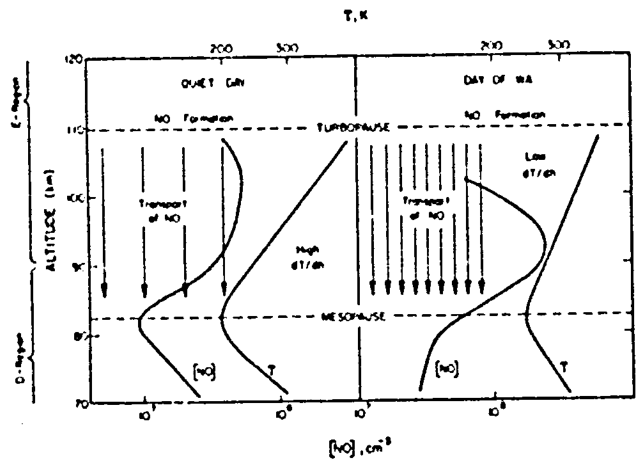

Figure 5. Schematic representation of the processes of NO transport on normal and WA days.

waves through the stratosphere is possible only during periods of west-to-east zonal circulation, which take place only in winter. If that is so and if dissipation of the planetary waves causes the above effects in T and K, then we have at least an explanation why the Winter Anomaly is observed during winter only.

Thus, all the above can be reformulated in terms of our initial task to get information about the middle atmosphere from D-region observations. The regular seasonal variations of electron concentration in the upper D-region give a good indication on the amplitude of summer-to-winter variation of neutral temperature. More detailed information, including water vapor concentration, can be obtained if both [e] and ion composition are measured together. Day-to-day variations in [e] (or L as the more widely available parameter) may provide information about day-to-day changes in T. Days of anomalous winter absorption show that there is an increase of temperature at mesopause heights and an intensification of turbulence at 80-120 km. as well as an increase of nitric oxide concentration in the mesosphere, and presumably an enhancement of wave processes. At the modern stage of our D-region studies the above relations can be used to get more information about the character of temperature variations and the nature and variations of the turbulence and wave processes.

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