Method of estimation of the temperature seasonal variations in the lower thermosphere from the ionospheric data

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Abstract. A comparison of the latest versions of the empirical models of the atmosphere (MSIS) and ionosphere (IRI) in the 100–115 km altitude range for the conditions of moderate solar activity and middle latitudes is performed. Relations are derived allowing us to estimate the temperature seasonal variations in the lower thermosphere on the basis of the data on the electron concentration n_e and/or the critical frequencies of the ionospheric E layer f_oE. It is demonstrated that the character of the annual variation of the temperature T and molecular oxygen concentration $[O_2]$ in the MSIS model contradicts the conclusions of the IRI model on a constancy during the year of the E-layer maximum height h_mE and on the semiannual variations of the $[NO^+]/[O_2^+]$ ratio for the above-mentioned conditions.

1. Introduction

Givishvili and Leshchenko [2001] were the first to try to reveal seasonal features of the temperature long-term trends at the height of the E-layer maximum h_mE on the basis of the electron concentration and/or the critical frequency f_oE measurements. However, the annual variations of the molecular oxygen concentration $[O_2]$, the principal ionized constituent at E-layer heights, have been taken into account by *Givishvili and Leshchenko* [2001] only indirectly, using the concentration ratio of the atomic oxygen O and molecular nitrogen N_2. The aim of this study is to consider more accurately factors determining the seasonal behavior of the electron production and loss rates in order to be able to analyze the fine structure of the seasonal variations of the long-term trends in the lower thermosphere temperature derived from n_e or f_oE measurements.

2. Initial Equations

The critical frequency f_oE is related to n_e by

$$f_oE = 8.98^{-3}(n_e)^{0.5} \text{ MHz} \qquad (1)$$

The equilibrium electron concentration n_e in the E region only slightly depends on diffusion processes, thermospheric winds, and electric fields. Therefore the formulae is correct there:

$$n_e = (q_m/\alpha_D)^{0.5} \text{ cm}^{-3} \qquad (2)$$

where q_m is the ionization rate in the layer maximum and α_D is the effective recombination coefficient. The values of q_m and α_D both depend on T. This fact makes it possible to determine reliably values of T at the height of the E-layer maximum h_mE using the measurements of f_oE if the q_m and α_D dependencies on T are known.

The ionization of the E region is provided mainly by the solar radiation in the $\lambda = 97.7$ nm and (Lyman-β) 102.6 nm lines interacting with oxygen molecules. This two channels provide formation of 75–85% of all charged particles. An additional source of ion formation at heights of 100–120 km is related to the solar X rays interacting with oxygen molecules and also nitrogen molecules and oxygen atoms. Thus about 80–90% of the ionization rate q depend on the O_2 content in the lower thermosphere. Therefore q is determined by the
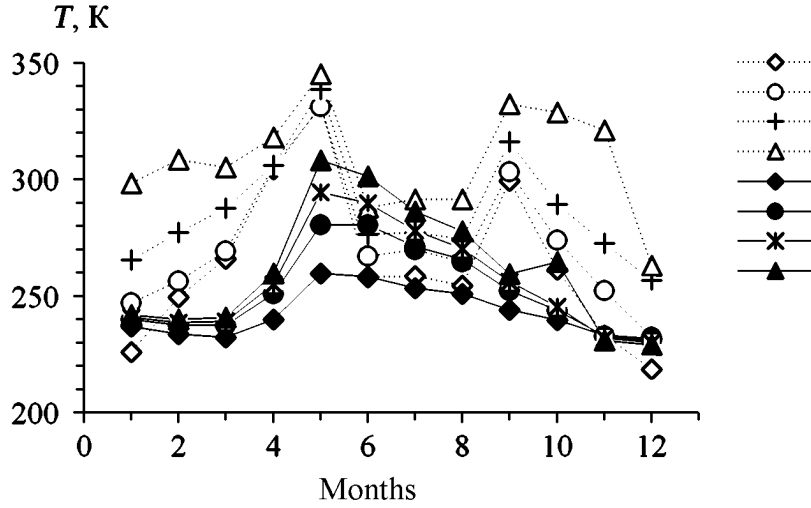
Figure 1. Annual variations of the temperature at a height of 110 km: $T(M)$ corresponds to the MSIS model (solid symbols); $T(12)$ corresponds to calculations by formula (12) for variable values of Θ from the IRI model (open symbols): Alma-Ata (diamonds), Slough (circles), Sverdlovsk (crosses), and Yakutsk (triangles).

Chapman equation [*Chapman,* 1931]:

$$q(O_2^+) = [O_2] \sum \sigma_i J_\lambda \exp\{-\sigma_\lambda e^{-1} \sec \chi\} \qquad (3)$$

where σ_i and σ_λ are the cross sections of ionization and absorption for an O_2 molecule, respectively, χ is the solar zenith angle, J_λ is the radiation flux at the wavelengths 102.6, 97.7, and 1–8 nm. The photoionization cross-sections are 10^{-18}, 2.5×10^{-18}, and $(0.2 - 3.6) \times 10^{-18}$ for 102.6 nm, 97.7 nm, and 1–8 nm, respectively. The absorption cross-sections are 1.5×10^{-18}, 4.0×10^{-18}, and $(0.2 - 0.9) \times 10^{-18}$ for the same wavelengths, respectively [*Samson and Cairns,* 1964; *Watanabe and Hinteregger,* 1962]. All σ_λ and σ_i are constant and do not influence seasonal and long-term changes of the ionization rate. Therefore, to simplify the process of $q(O_2)$ calculations, we accept below generalized values $\sigma_i = 1.82 \times 10^{-18}$ cm^2 and $\sigma_\lambda = 2.34 \times 10^{-18}$ cm^2 taking into account the weighted input of the above-indicated solar radiation ranges into the ionization rate.

In the layer maximum where the $dq(O_2^+)/dh = 0$ condition [*Chapman,* 1931] is fulfilled we have

$$q_m(O_2^+) = \frac{[O_2]\sigma_i J_\lambda}{e} \text{ cm}^{-3}\text{ s}^{-1} \qquad (4)$$

Equation (4) is not convenient for calculations of T because of two reasons. First, one has to have all the information on absolute values and seasonal variations of $[O_2]$. However different authors [*Alcayde et al.,* 1974; *Mayr et al.,* 1976; *Scialmon,* 1974] give significantly different information. Second, to know the seasonal variations of the height where the $dq(O_2^+)/dh = 0$ condition is fulfilled, that is, to which height the derived temperature corresponds, one has to perform additional calculations to reconstruct the $q(h)$ vertical profile using formula (3).

So we rewrite formula (3) in the form

$$q_m(O_2^+) = \frac{\sigma_i J_\lambda \cos \chi}{\sigma_\lambda e H(O_2)} \qquad (5)$$

Here $H(O_2) = kT/mg$ is the scale height for O_2, k is the Boltzmann constant, m is the molecular weight of O_2, and g is the gravity acceleration. Substituting numerical values of k, m, g, σ_1, and σ_2 into formula (5), we obtain

$$q_m(O_2^+) = \frac{1.087 \times 10^6 \cos \chi J_\lambda}{T} \text{ cm}^{-3}\text{ s}^{-1} \qquad (6)$$

Now there is no need to know the absolute values of $[O_2]$. According to *Ivanov-Kholodny and Firsov* [1974] the flux of the radiation ionizing O_2 is

$$J_\lambda = (0.56 + 1.475 \times 10^{-2}F_{10.7}) \times 10^{10} \text{ photon}^{-2}\text{ cm}^{-1}$$

where $F_{10.7}$ is the solar activity index.

The loss of free electrons in the daytime quiet E region occurs in their dissociative recombination with NO^+ and O_2^+ ions with the rate constants [*Mehr and Biondi,* 1969]:

$$\alpha(O_2^+) = 2.2 \times 10^{-7}\left(\frac{300}{T}\right)^{0.5} \text{ cm}^3\text{ s}^{-1} \qquad (7)$$

and

$$\alpha(NO^+) = 4.1 \times 10^{-7}\left(\frac{300}{T}\right)^{0.5} \text{ cm}^3\text{ s}^{-1} \qquad (8)$$

Since at the considered heights $n_e = [NO^+] + [O_2^+]$ and $\alpha(NO^+) \simeq 2\alpha(O_2^+)$, the total loss rate is determined by

$$\alpha_D = \alpha(O_2^+) + \alpha(NO^+) \approx \Theta\,\alpha(O_2^+) \qquad (9)$$

where

$$\Theta = \frac{1 + 2[NO^+]/[O_2^+]}{1 + [NO^+]/[O_2^+]}$$

Finally, we have

$$\alpha_D = \frac{3.8 \times 10^{-6}\Theta}{T^{0.5}} \qquad (10)$$

Two ways to calculate the temperature from the f_oE data are considered. In the first, formula (4) is taken into account, the second is based on formula (6). Substituting these two formulae and formulae (1) and (10) into equation (2), we have

$$T = \frac{7.636 \times 10^{21}(f_oE)^8\Theta^2}{[O_2]^2(0.56 + 1.475 \times 10^{-2}F_{10.7})^2} \text{ K} \qquad (11)$$

and

$$T = \frac{3.45 \times 10^6 \cos^2\chi(0.56 + 1.475 \times 10^{-2}F_{10.7})^2}{\Theta^2(f_oE)^8} \text{ K} \qquad (12)$$

3. Analysis of the Base Equations

According to the International Reference Ionosphere (IRI) model [*Bilitza*, 1997], the height of the E-layer maximum h_mE is constant over the year and equal to 110 km. Therefore we took the concentrations of NO^+ and O_2^+ ions (parameter Θ) and the electron concentration $n_e(f_oE)$ for the altitude of 110 km needed to analyze formula (11). The calculations using formula (11) were performed for 9 vertical ionospheric sounding stations shown in Table 1, conditions of local noon (1200 LT) and moderate solar activity ($F_{10.7} = 120$). The period from September 1971 to August 1972 corresponds to this solar activity level. The data on the molecular oxygen concentration were taken from the MSIS (Mass-Spectrometer and Incoherent Scatter) model [*Hedin*, 1991]. Figure 1 shows the temperature values $T(M)$ from this model for the coordinates of 4 out of 9 stations indicated above.

Calculating the temperature $T(11)$ by formula (11) for h_mE lead to an absurd conclusion that the summer values of $T(11)$ exceed the winter values by a factor of 10 or even a hundred (see Figure 2). At the same time they indicate that the h_mE height can not stay fixed during the year: in summer the E-layer maximum should decent to the region of higher values of $[O_2]$.

Table 1. Coordinates of the Stations for Which the Temperatures Were Calculated

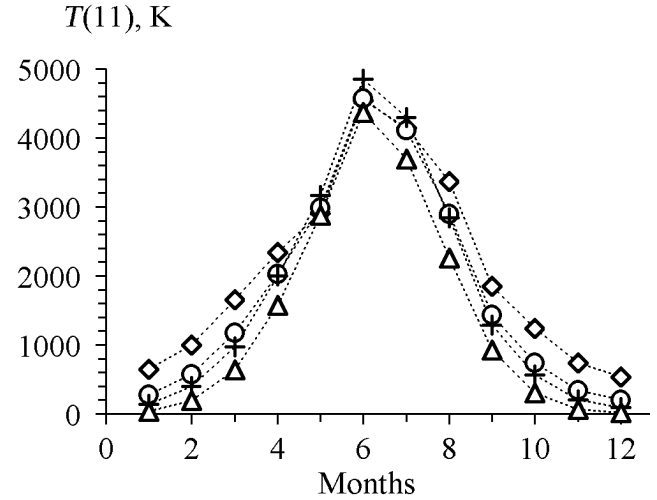|	Station	Coordinates	
1	Alma-Ata	42.3°N,	76.9°E
2	Durbes	50.1°N,	4.6°E
3	Slough	51.5°N,	0.6°E
4	Juliusru	54.6°N,	13.4°E
5	Moscow	55.5°N,	37.3°E
6	Sverdlovsk	56.5°N,	58.6°E
7	Tomsk	56.0°N,	84.0°E
8	Leningrad	60.0°N,	30.2°E
9	Yakutsk	62.0°N,	129.7°E

$T(11)$, K

Figure 2. The annual variations of the temperature $T(11)$ calculated by formula (11) for the same four stations as in Figure 1 (symbols same as in Figure 1).

To specify the character of the seasonal variations of q_m and h_mE and also to estimate the influence of the seasonal variations of the Θ parameter on the temperature evaluation, we use formula (12). Following IRI we presume that $h_mE = 110$ km all over the year. The results of the calculations for the four stations indicated above are shown in Figure 1 as $T(12)$ values. One can see that in this case the deviations between $T(12)$ and $T(M)$ are relatively small in summer but are considerable in equinox periods. The important thing is that a semiannual component which is absent in the annual variations of $T(M)$ is visually seen in the seasonal behavior of $T(12)$.

This semiannual component of the $T(12)$ seasonal behavior is due to the seasonal variations of the Θ parameter (see Figure 3). In the IRI model these variations repeat the annual behavior of the NO^+ concentration. Two features should be mentioned in this behavior: considerable semiannual variations of $[NO^+]$ and some excess of the winter values of $[NO^+]$ over the summer values. However, the data of the observations on board the Orbiting Geophysical Observatory (OGO 6) show that at 105 km the concentration of nitric oxide (the main photochemical source of the NO^+ formation at middle latitudes) changes smoothly from the minimum values of about 10^7 cm^{-3} in winter to the maximum values of about 2.5×10^7 cm^{-3} in summer [*Gravens and Stewart*, 1978]. At the same time, the seasonal behavior of the molecular oxygen concentration controlling the concentration of O_2^+ ions at heights of the E region is also rather monotonous. This concentration decreases from winter months to summer. Thus, there is no semiannual component in the annual behavior of the $[NO^+]/[O_2^+]$ ratio and this ratio should be higher in summer than in winter. Therefore it is not clear what causes the seasonal variations of Θ presented in the IRI model. There is only one source of data on the E-layer ion composition: the data of rocket mass-spectrometer mea-
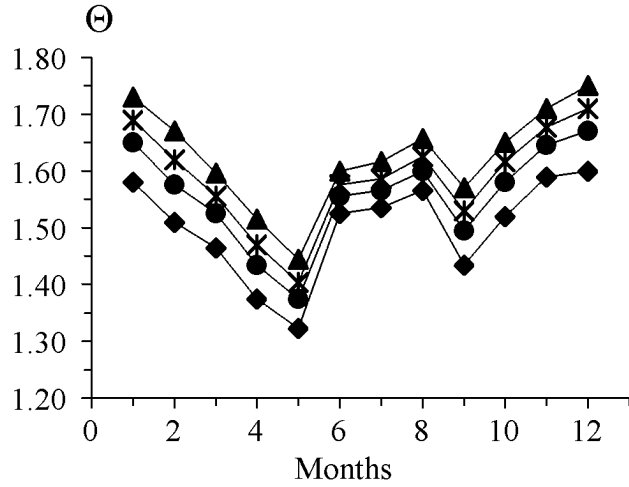
Figure 3. The seasonal variations of the Θ parameter from the IRI model for the same stations (symbols same as in Figure 1).
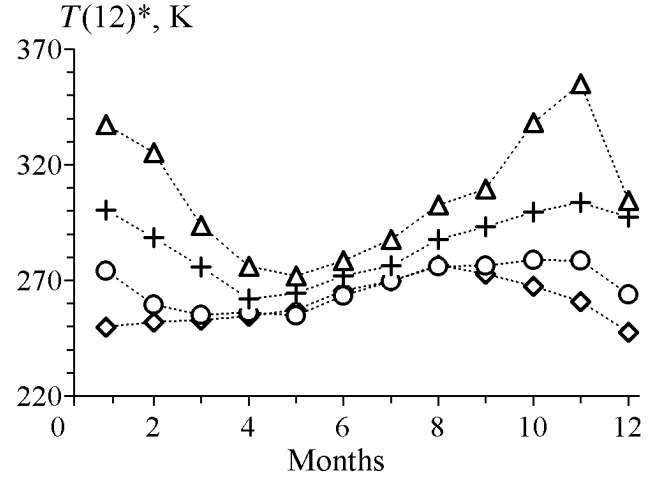

Figure 4. The annual variations of the temperature $T(12)^*$ calculated with constant ion composition over the year and without taking into account of the $[O_2]$ seasonal variations for the same stations (symbols same as in Figure 1).

surements. Evidently, there are not enough of these data to derive statistically significant conclusions on the seasonal variability of the $[NO^+]/[O_2^+]$ ratio.

Since the problem of existence and character of the Θ seasonal variations needs a specification, we assumed that these variations are absent. We accepted this hypothesis as a first approximation and calculated new values of $T(12)^*$ under condition that $\Theta = \Theta_c$, where Θ_c is an annual mean value of Θ at each station considered. The results of the calculations shown in Figure 4 also do not correspond to the structure of the temperature annual variations in the E-layer maximum corresponding to the MSIS model: the obtained values of $T(12)^*$ are maximum in fall-spring months but not in summer. This result demonstrates that formula (12) should contain a function taking into account the seasonal variations of $[O_2]$.

Givishvili and Leshchenko [2001] introduced into formula (12) a parameter γ which took into account the $[O_2]$ seasonal variations according to the data of the rocket measurements of the $[O]/[N_2]$ ratio at a height of 130 km [*Antonova and Katyushina,* 1976]. In other words, T was calculated using

$$T = \frac{3.45 \times 10^6 \cos^2 \chi (0.56 + 1.475 \times 10^{-2} F_{10.7})^2 \gamma^2}{\Theta^2 (f_o E)^8} \ \ \text{K}$$
(13)

The evaluation of γ from the data on the $[O]/[N_2]$ ratio made it possible to estimate the annual T variations at $h_m E$ rather approximately. More explicit description of the relative seasonal variations of $[O_2]$ at height of the E-layer maximum provides the MSIS model, if one determines the γ parameter in another way: as a ratio $\gamma = [O_2]_i/[O_2]_J$, where J stands for January and $i = 1-12$ corresponds to the month number. Thus γ is characterized by the relative changes of $[O_2]$ at about 110 km relative the values in January.

Formula (13) was also analyzed for the conditions of the local noon (1200 LT) and moderate solar activity ($F_{10.7} = 120$) for the coordinates of the same stations. The χ values

corresponded to the 15th day of each month. The data on $f_o E$ were taken from the IRI model. Since according to this model the $h_m E$ height at noon is constant during the year and equal to 110 km, the annual variations of γ should correspond to the annual behavior of $[O_2]$ at this height. The results of calculations are presented in Figure 5 as values $\Delta T = T(13) - T(M)$.

One can easily see that the discrepancy between $T(13)$ and the model values of the temperature $T(M)$ is in this case maximum in summer months when the discrepancy reaches 100–200 K. In winter months the difference between $T(13)$ and $T(M)$ is small. This indicates to the fact that in winter $h_m E$ is actually close to 110 km. In summer months it should decent down to the altitudes where $[O_2]$ is considerably higher than at 110 km, as has been suggested ear-
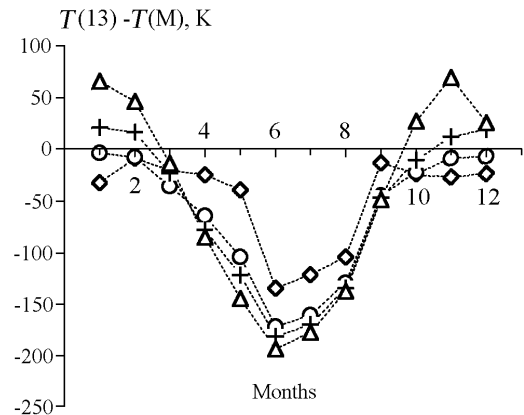

Figure 5. The calculated seasonal trends of the temperature $\Delta T = T(13) - T(M)$. Seasonal variations of $[O_2]$ at a height of 110 km and the constancy of $h_m E$ over the year were taken into account (symbols same as in Figure 1).

lier. This conclusion agrees with empirical models of the midlatitude ionosphere [*Fatkullin et al.*, 1981; *Givishvili and Fligel'*, 1971; *Robinson*, 1960]. According to these models the noon values of h_mE vary during the year from the maximum winter values equal to 108–112 km to the minimum summer values of about 104–108 km. The incoherent scatter measurements conducted in Kharkov (50°N) in 1978–1983 also indicate to a lowering of h_mE from winter to summer [*Ivanov-Kholodny et al.*, 1998]. To check the conclusion on variability of the h_mE and to estimate its changes during the year, we calculated $\gamma(M)$

$$\gamma(M) = \frac{T(M)^{0.5}\Theta(f_oE)^4}{1.86 \times 10^3 \cos\chi(0.56 + 1.475 \times 10^{-2}F_{10.7})} \quad (14)$$

The results of the calculations are shown in Figure 6. They make it possible to see what concentrations of $[O_2]$ at 110 km correspond to the temperature $T(M)$ from the MSIS model, f_oE from the IRI model, and the assumption that $\Theta = $ const. Further, using the ratio $[O_2]_i/[O_2]_J$ from the MSIS model, the heights of the E-layer maximum are found and denoted as $h_m(M)$ (see Figure 7). One can see that the $h_m(M)$ heights vary over the year in the way close to that found in the models [*Fatkullin et al.*, 1981; *Givishvili and Fligel'*, 1971; *Robinson*, 1960]. (It should be noted that according to *Ivanov-Kholodny et al.* [1998] the amplitude of the annual variation of h_mE at a latitude of 50°under $F_{10.7} = 120$ reaches about 7 km. According to our evaluations for the same latitude and moderate solar activity level the seasonal variations of h_mE do not exceed ∼3 km. This discrepancy is rather large, but it does not influence the general conclusion on the variability of $h_m(M)$ over the year.)

Then, using the same ratios $[O_2]_i/[O_2]_J$ from MSIS the values of $T_m(M)$ characterizing the temperature at the $h_m(M)$ height varying during the year are found. Substituting into equation (14) the values $T_m(M)$ instead of $T(M)$, we obtain the values of $\gamma_m(M)$ corresponding to the variable
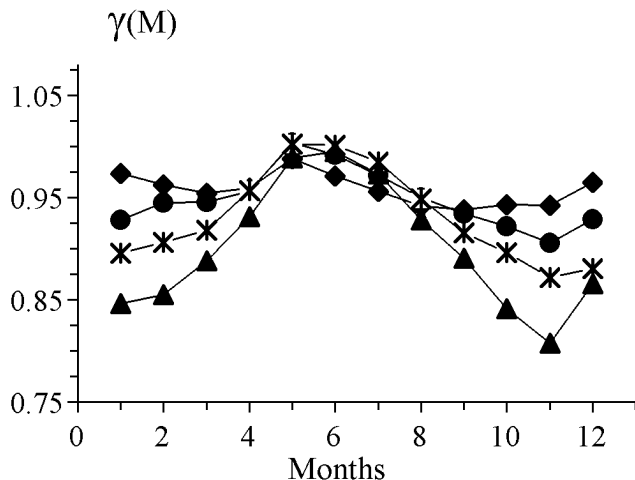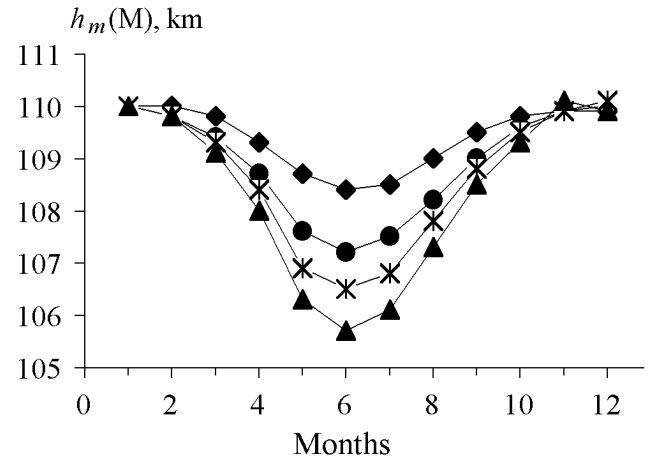

Figure 7. Variations of the $h_m(M)$ height over the year corresponding to the seasonal variations of $[O_2]$ from the MSIS model and f_oE from the IRI model (symbols same as in Figure 1).

$h_m(M)$ height (see Figure 8). Figure 8 shows also the values of γ used by *Givishvili and Leshchenko* [2001]. One can see that the latter are overestimated as compared with the values determined in this paper but by not more than by 10–15% depending on season and latitude. This fact does not influence the results of the above-mentioned paper which deals with the seasonal features of the long-term temperature trends, because the annual behavior of γ in that paper on the whole does not contradict the MSIS model. Nevertheless, to use the vertical sounding data not only to derive the long-term temperature trends but also to evaluate its absolute values it is desirable to use formula (14).

Thus, we have now all the necessary information for calculation of the temperature at the heights of the E-layer
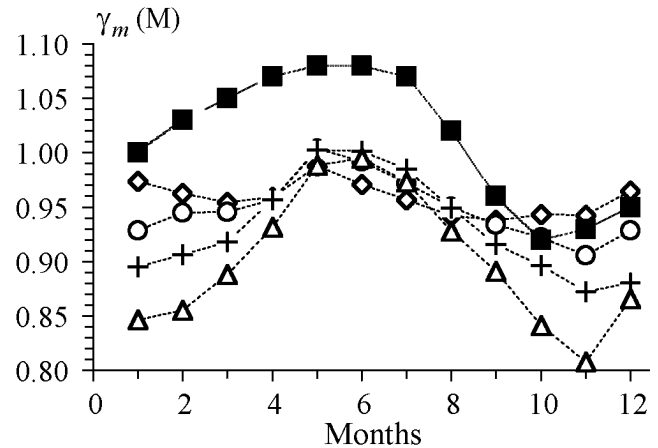

Figure 6. Variations of $\gamma(M)$ manifesting variations of $[O_2]$ at the variable height h_mE for $\Theta = $ const and the temperature at 110 km (symbols same as in Figure 1).

Figure 8. Variations of $\gamma_m(M)$ manifesting the annual variations of $[O_2]$ and T at the variable height $h_m(M)$ (symbols same as in Figure 1). The solid squares show the values of γ accepted by *Givishvili and Leshchenko* [2001].

maximum on the basis of the data on f_oE, including the data on the changes of both this very height (the $h_m(\mathrm{M})$ parameter) and the O_2 concentration at this height, i.e., the $\gamma_m(\mathrm{M})$ parameter.

As for the IRI model, it is worth mentioning that the reliability of the data on the electron concentration (frequency f_oE) in it arises no doubts. Evidently, the same is not true for the seasonal variations of both the maximum height and ion composition of the E layer. The calculations performed show that the h_mE height should vary over the year. However, the assumption on the absence of annual variations of the $[\mathrm{NO^+}]$ and $[\mathrm{O_2^+}]$ generally speaking may not correspond to the reality. Nevertheless, some corrections to the accepted here algorithms of temperature calculations would be reasonable to insert only after statistically reliable data on the parameter Θ seasonal behavior are obtained.

4. Conclusions

The analysis presented above makes in possible to make the following conclusions:

1. The method to evaluate the lower thermosphere temperature from the ionospheric vertical sounding data is specified. A parameter γ is introduced which takes into account the seasonal variations of the O_2 concentration at altitudes of the E-layer maximum. This parameter makes it possible to reproduce the annual variation of the lower thermosphere temperature at middle latitudes ($\varphi = 42° - 62°\mathrm{N}$) on the basis of the critical frequency f_oE measurements.

2. The assumption on the constancy of the E-layer maximum height made in the IRI model, does not fit the reality. There is ground to believe that $h_mE = 110$ km only in winter. During the rest of the year it is slightly lower. The amplitude of the seasonal variations of h_mE depends on latitude, increasing from 1.6 to 4.3 km in the latitudinal band $42°$–$62°\mathrm{N}$.

3. The data on the seasonal variations of the ion composition at E-layer heights presented in the IRI are doubtful. To specify the character of the annual behavior of the $[\mathrm{NO^+}]/[\mathrm{O_2^+}]$ ratio, additional experimental studies are needed.

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