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During solstices a plasma pressure difference exists between geomagnetic conjugated points located in the topside ionosphere. This difference is responsible for an interhemispheric plasma flow. Here we investigate the possibility of this interhemispheric plasma flow to drag along neutral particles and to influence the neutral winds.

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OBSERVAÇÕES E NOTAS

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INFLUENCE OF AN INTERHEMISPHERIC PLASMA FLOW ON THE NEUTRAL ATMOSPHERE ION-DRAG EFFECTS

C.J. Zamlutti (Instituto de Pesquisas Espaciais INPE - C.P. 515, 12201 - São José dos Campos, SP - Brasil)

During solstices a plasma pressure difference exists between geomagnetic conjugate points located in the topside ionosphere. This difference is responsible for an interhemispheric plasma flow. Here we investigate the possibility of this interhemispheric plasma flow to drag along neutral particles and to influence the neutral winds. (Ion-drag effects, transports process, ionospheric-atmospheric interaction.

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INTRODUCTION

Interhemispheric transport of energy by a plasma flow has been considered both on theoretical grounds (e.g. Schunk, 1975; Raitt et al., 1977; Schunk and Watkins, 1979; Young et al., 1980) and in its practical aspects (Bailey and Moffett, 1979). One may distinguich two states in the processa: a transient state, for which equilibrium along the vertical at geomagnetic conjugate locations has not yet been reached by ionized particles, and a steady state situation, for which such equilibrium was reached.

To consider the ion-drag effect of the plasma flow on the neutral atmosphere, we assume that the steady state situation was established. Under such circunstances, along the path, the ionized particles drift by the action of partial pressure gradient against the gravitational and frictional forces (Banks and Holzer, 1965).

In this paper we want to determine to what extent the ion-drag produced by the interhemispheric plasma flow affects neutral winds.

THE PHYSICAL MECHANISM

During solstices the daytime summer hemisphere receives about 20% more solar radiation than the winter hemisphere. The resulting energy stored in the summer hemisphere, however, ends up being smaller than that stored in the winter hemisphere. This peculiar situation is produced by the active dynamics of the upper atmosphere, as follows. The photoelectrons, produced in larger scale in the summer hemisphere, travel along the geomagnetic field lines so that the summer to winter flow exceed the one in the reverse sense. As a result of the collision between photoelectrons and neutral particles, primary electrons are produced. These electrons are high energy electrons (nonthermal electrons) and, colliding again with neutral particles, can produce more ionization (secondary electrons), which may eventually be nonthermal. The ionization at the winter hemsiphere is, therefore, constituted mostly by lower order electrons (secondary thermal electrons). The reverse winter to summer) flow contains a lot of primary and secondary nonthermal electrons. The ionization at the summer hemisphere is then constituted by higher order electrons (secondary, tertiary). Therefore, not only the ionization is larger in the winter hemisphere, but also the neutral atmosphere temperatures are larger, because of the effective electron neutral collisional energy transfer process. The ionized particles temperatures, however, are larger in the summer hesmisphere due to the active electron thermalization mechanism. This transient has been studied by Torr et al. (1980), who considered the back and forth flow of both nonthermal and thermal electrons between opposite hemispheres.

Under steady state condition, the plasma pressure difference between opposite hemispheres yields a meridional flow of plasma along the geomagnetic field lines. This subject has been the matter of a series of papers by Bailey and co-workers (Bailey et al., 1978, 1982; Bailey and Moffett, 1979). Additionally, an horizontal meridional wind component produced by seasonal asymetries appears.

Seasonal winds were considered by Volland and Mayr (1972a,b), using the method of perturbation of the hydrodynamic equations and by Blum and Harris (1975), who solved the equations of motion employing atmospheric static models.

Seasonal winds affect the flow of plasma through air-drag (King and Kohl, 1965). Their relevance to the plasma flow through air-drag was treated theoretical by Banks and Holzer (1965) and experimentally by Salah and Holt (1974), Roble et al. (1974, 1977) and Bittencourt and Sahai (1978). Conversely, the flow of plasma carries the neutral particles through a small ion-drag (King and Kohl, 1965). The ion-drag net effect on a global scale, which is the matter of the present work, has been incorporated in computer modelling of the neutral atmosphere dynamics (e.g. Dickinson et al., 1975, 1977), but so far nort determined separately using measured ionospheric and atmospheric parameters.

At low latitudes, the protonospheric link does not significantly alter the flow of plasma and there plasma pressure difference between conjugate ionospheres is quickly brought into balance (Bailey et al., 1982). At middle latitude, the photonosphere acts as a reservoir and its large plasma content significantly reduces the flow of plasma.

A SIMPLE MATHEMATICAL FORMULATION

We assume first that the geomagnetic and geographic axis are coincident. We further consider that it is possible to define an interhemispheric dynamical equilibrium situation, for which no plasma transport between hemisphers occurs and which is roughly reached during equinox.

Under nonequilibrium conditions (solstices), a geomagnetic field aligned plasma velocity, $\underline{\delta v}$, is established between hemispheres to compensate the pressure gradient estabilished between them. This velocity is given by:

$$\underline{\delta v} = \delta \underline{v}_d + \delta \underline{u}_s, \qquad (1)$$

where $\underline{v}_d$ the plasma diffusion velocity along the geomagnetic field line and $\underline{u}_s$ is the neutral particle velocity in the same direction. The operator δ represents the differences between the actual value of the parameter and its reference val ue for the equilibrium situation. Neglecting electric fields, thermal diffusion and assuming the validity of chemical equilibrium, the diffusion velocity is given by:

$$\underline{v}_d = -D\,[\nabla_s\, p/p - \sin I/H_p] \qquad (2)$$

(Banks and Holzer, 1965). Here, $D = k(T_i + T_e)(m_i\, \nu_{in})^{-1}$ is the diffusion coefficient, $p = Nk(T_i + T_e)$ is the plasma pressure, $H_p = k(T_i + T_e)(m_i\, g)^{-1}$ is the plasma scale height, I is the dip angle and ∇_s is the gradient operator along the field lines. The others are the standard symbols for electron number density (N), eletron temperature (T_e), ions temperature (T_i), ion mass (m_i), ion-neutral collision frequency (ν_{in}), Boltzman constant (k) and acceleration of gravity (g).

The velocity in (1) produces and ion-drag force which tends to carry the neutral particles from one hemisphere to the other hemisphere. A viscosity reaction tendes to impede the motion. The equilibrium is reached when:

$$\partial^2 \delta \underline{u}_s/\delta z^2 = -\delta[(\rho/\mu)\ \nu_{ni}\,\underline{v}_d], \qquad (3)$$

where μ is the viscosity coefficient, ρ is the neutral particles density, ν_{ni} is the neutral-ions collision frequency and z is the altitude.

The neutral-ion drag force is related by Newton's Third law to the ion-neutral drag forces resulting:

$$\rho\nu_{ni} = N\, m_i\, \nu_{in} \qquad (4)$$

Inserting (2) and (4) into (3) we obtain:

$$(\partial^2 \delta \underline{u}_s/\partial z^2) = -\mu^{-1}(\sin I)\ \delta[\nabla_z\, p + N\, m_i\, g] \qquad (5)$$

for the nearly vertical range of the magnetic field line.

At altitudes where sin I decreases significantly, a modification is expected from fluid to individual particles behavior for the ions (Banks and Holzer, 1965). The individual ion-neutrals collisions are very low because of the rarefied air density. Under these conditions no ion drag contributes to modify the wind velocity.

RESULTS

A consistent set of most of the necessary parameters, to solve Equation 5, is provided by the incoherent scatter technique. Since there are not two incoherent scatter radars at geomagnetic conjugate locations a compromise solution must be considered to this problem. The one chosen in this work was to assume that the summer (winter) ionosphere parameters at a given location are identical to the summer (winter) ionospheric parameters at its geomagnetic conjugate location. Therefore, the data collected at one single site during one entire year can, in principle, provide the necessary elements to simulate this summer-winter contrast existing between geomagnetic conjugated locations.

In order to estimate the reliability of the proposed theory, we carried out the integration of Expression 5 using numerical methods (Simpson rule). The necessary ionospheric parameters were obtained from the data published by Evans (1975). For the neutral atmosphere we used the Jacchia (1970) model with the same exosphere temperatures of Roble et al. (1977). As boundary conditions we assumed that the average velocity, $\delta \overline{u_s}$, as well as its first and second derivatives all vanish at 200 km. The ramaining parameters were the viscosity and ion mass. For the ion mass we considered an ion composition varying linearly from pure atomic oxygen ions at the F_2 layer peak to pure helium at 650 km during daytime and to pure atomic hydrogen ions at 700 km during nighttime (Rishbeth and Garriott, 1969). The viscosity depends on the neutral atmosphere composition and is essentially that of atomic oxygen. Our results are marked with x in figures 1 and 2.

Another alternative method to derive the wind velocity induced by ion drag is the use of Expression 1. This method was employed by Roble et al. (1974, 1977), who used the component of plasma drift along the geomagnetic field line, measured with the incoherent scatter technique. Their diffusion velocity was computed using a slightly different expression than our Expression 2. Moreover, they included the thermal diffusion to compute the diffusion velocity. The results obtained with this approach are presented in Figures 1-4.

The results exhibit a satisfactory agreement considering the fact that the experimental accuracy of wind measurements is of the order of ±10m/s (Salah and Holt, 1974), which yields and accuracy of ±20m/s for the difference $\delta \overline{u_s}$. Our results, depending on the computations of the diffusion velocity, introduce an erros of ±2m/s at 300 km (Salah and Holt, 1974). Since we used numerical integration which is an error reducing method, the uncertainties tend not to grow with altitude because of the increasing number of points used in the integration. The shortcommings of our approach are the boundary conditions imposed at 200 km, which only hold for daytime. According to our estimates for the whole range of altitudes, the neglecting of thermal diffusion does not introduce more than 20% error in the daytime results and is of no importance for the nighttime computations.

Another restriction of our technique is that Equation 3 does not consider the time variation of wind speed required during sunrise and sunset. This, however, is a complementary term which can easily be introduced.

DISCUSSION

The poposed approach, to compute the net global effect of ion drag on the neutral atmosphere, has proved satisfactory as far as the order of magnitude is concerned. However, it has a severe intrinsic limitation, namely, the assumption of steady state dynamics. This may be tolerated for daytime and nighttime (noon and midnight), but is not true during sunrise and sunset.

Regarding the influence of electric fields, the reader is referred to the work by Salah and Holt (1974). Since both the experimental and computed values neglected their contribution, we will not consider them further. If a detailed account of them is required, we certainly need a more complete set of measurements.

The influence of the other simplifications undertaken in this work were commented in the preceding section because of their relevance to matching computed with experimental results.

Comparing the obtained results with the winds derived from Roble et al. (1977), we see that ion drag effects on the wind velocities range fron roughly 20% at 300 km to as much as 80% at protonospheric altitudes.

CONCLUSION

A method was presented to compute the net global effect of ion-drag on the neutral atmosphere during solstices. The first tests performed with it were satisfactory. The obtained results exhibit a clear tendency for a net transport of mass between the summer and the winter hemisphere above 300 km with reversal of direction from day to night. The extent of ion-drag ranges from 20%, at 300 km, to as much as 80% at protonospheric altitudes.

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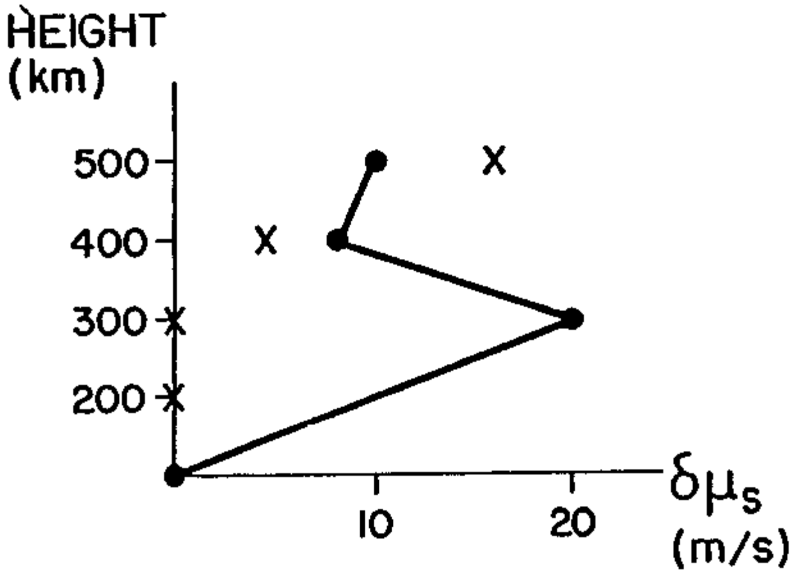
FIGURE CAPTIONS

Figure 1 - Nighttime wind velocities produced by summer to winter differences in ion drag above Millstone Hill. Positive values corresponds to summer to winter direction. Computed values are marked with x.

Figure 2 - Same as figure 1 for daytime wind velocities. Computed values are marked with x.

Figure 3 - Same as figure 1 for sunrise wind velocities.

Figure 4 - Same as figure 1 for sunset wind velocities.



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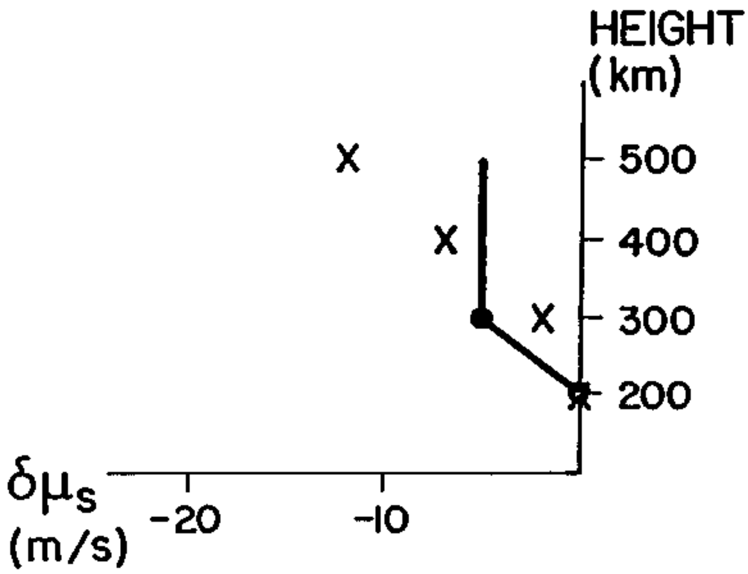

Figure 2

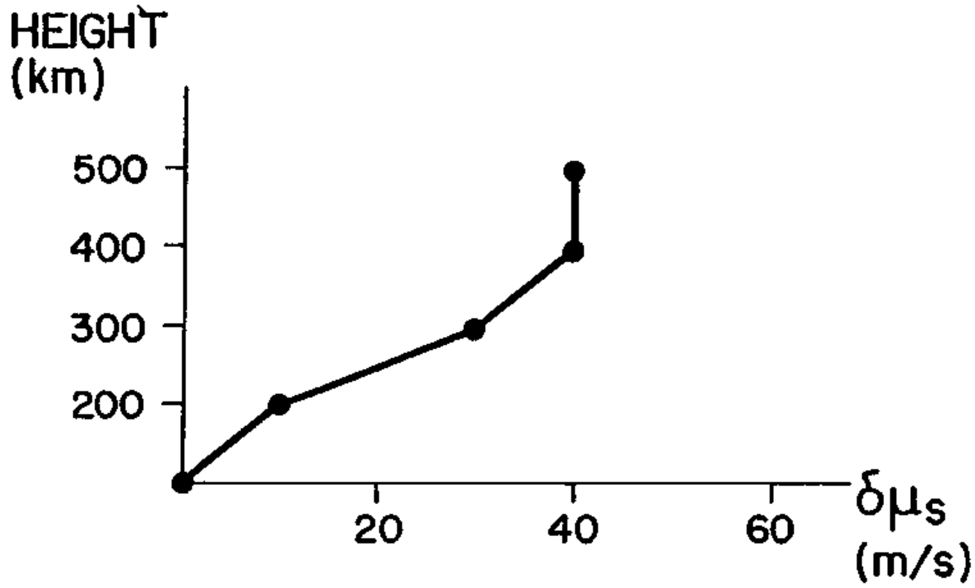

Figure 3

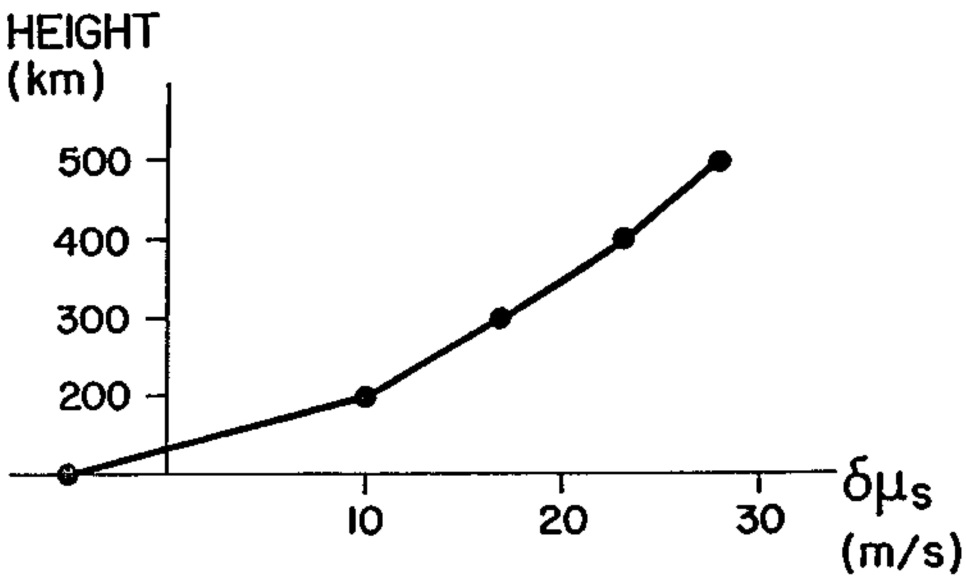

Figure 4