~~Temperature~~Equilibrium temperature distribution and Hadley circulation in an axisymmetric model.

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Abstract

The impact of the equilibrium temperature distribution, θ_E, on the Hadley circulation simulated by an axisymmetric model is studied. The ~~temperature~~θ_E distributions that drive the model are modulated here by two parameters, *n* and *k*, the former controlling the horizontal broadness and the latter ~~defining change in~~controlling the vertical ~~lapse rate.~~stratification of θ_E. In the present study, ~~the changes~~variations of the ~~temperature~~θ_E distribution mimic changes of the energy input of the atmospheric system leaving as an almost invariant the equator-poles θ_E difference. Both equinoctial and time-dependent Hadley circulations are simulated and results compared. The results give evidence that concentrated ~~temperature~~θ_E distributions enhance the meridional circulation and jet wind speed intensities even with a lower energy input. The meridional circulation and the subtropical jet stream widths are controlled by the broadness of horizontal ~~temperature~~θ_E rather than the vertical ~~lapse rate~~ ~~k~~stratification, which is important only when the temperature distribution is concentrated at

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the equator. The jet stream position does not show any dependence with n and k, except when the ~~temperature~~θ_E distribution is very wide and in such a case the jet is located at the mid-latitude and the model temperature clamps to forcing θ_E. Using n=2 and k=1 we have the formulation of the potential temperature adopted in classical literature. A comparison with other works is performed and our results show that the model running in different configurations (equinoctial, solstitial and time- dependent) yields results similar to one another.

1 Introduction

The earth's atmosphere is driven by differential heating of the earth's surface. At the equator, where the heating is larger than that at other latitudes, air rises and diverges poleward in the upper troposphere, descending more or less at 30° latitude (subtropics). This circulation is known as Hadley cell. Because of the earth's rotation, this circulation produces two subtropical jets~~,~~ at about 30° north and 30° south. A poleward shift (Fu and Lin, 2011) and an enhanced wind speed of these jets (Strong and Davis, 2007) are associated with a possible Hadley cell widening and strengthening, which has been observed in the last decades (Fu et al., 2006; Hu and Fu, 2007; Seidel et al., 2008; Johanson and Fu, 2009; Nguyen et al., 2013).

There are a few studies suggesting possible causes of these phenomena. One of the theories postulates global warming as a possible cause of Hadley cell widening (Lu et al., 2009). However, the atmosphere is a complex system containing many subsystems interacting with one another and the global warming might not be the only cause that is suggested to explain the widening. Ozone depletion (Lu et al., 2009; Polvani et al., 2011), SST warming (Chen et al., 2013; Staten et al., 2011) and aerosol (Allen et al., 2012) have also been invoked to explain the Hadley cell widening.

Climate models vary to some extent in their response and the relationship between global warming and Hadley cell is not straightforward. For instance, Lu et al. (2007) found a smaller widening than the observed one. Gitelman et al. (1997) showed that the meridional temperature gradient decreases with increasing global mean temperature and the same result can be found in recent modeling studies (Schaller et al., 2013).

Much of our understanding on the Hadley cell comes from theories using simple models (Schneider 1977, Schneider and Lindzen 1977 and Held and Hou 1980, hereafter HH80) and

such a simple model will be adopted here in order to understand how temperature distributions can change the Hadley circulation. How much temperature change impacts the real Hadley circulation is not clear yet, perhaps because of discrepancies between observations, reanalysis (~~Weliser~~Waliser et al., 1999) and climate model outputs, although these differences are becoming less marked because of newer observational datasets or correction of the older ones (Sherwood 2008, Titchner et al., 2008, Santer et al., 2008). Hence, it is critical to understand the possible mechanisms behind the cell expansion starting from a simple model.

The objective of this study is to analyze the sensitivity of a model of the symmetric circulation to the radiative-convective equilibrium temperature distribution. Our point of departure is the symmetric model used by Cessi (1998), which is a bidimensional model considering atmosphere as a thin spherical shell. This model will be briefly described in Sect. 2. The model describes mainly a tropical atmosphere, hence it does not allow for eddies. Although eddies may play a central role in controlling the strength and width of the Hadley cell (e.g. Kim and Lee, 2001; Walker and Schneider, 2006), a symmetric circulation, driven by latitudinal differential heating, can exist even without eddies and it is a robust feature of the atmospheric system (Dima and Wallace, 2003). The temperature distributions used in this study represent some paradigms of tropical atmospheres. Among the possible causes that can change temperature distributions there are El Niño, global warming and change of solar activity. We will show, in Sect. 3, that the energy input is not as important as the ~~temperature~~forcing distribution. Our results are consistent with those obtained both by Hou and Lindzen (1992) (hereafter HL92), and recently by Tandon et al. (2013) who performed experiments similar to those described here. The conclusions will be drawn in Sect. 4.

2 The model

The model used in this study is a bidimensional model of the axis-symmetric atmospheric circulation described in Cessi (1998). The horizontal ~~variable~~coordinate is defined as $y = asin\phi$ from which we have

$$c(y) = cos\phi = \sqrt{(1 - y^2/a^2)} \quad (1)$$

where a is the radius of a planet having a rotation rate Ω, the height of atmosphere is prescribed to be H.

The model is similar to the Held and Hou model (HH80), but ~~the difference is that, the model~~it prescribes a horizontal diffusion ν_H other than the vertical diffusion ν_v. The prognostic variables are the angular momentum M, defined as $M = \Omega ac^2 + uc$ where u represents the zonal velocity; the zonal vorticity ψ_{zz} with the meridional stream function ψ defined by

$$\begin{array}{l} \partial_y\psi \equiv w; \quad \partial_y\psi \equiv w; \\ \partial_z\psi \equiv -cv \quad \partial_z\psi \equiv -cv \end{array} \quad (2)$$

and the potential temperature θ that is forced towards a radiative-convective equilibrium temperature θ_E. Starting from the dimensional equations of the angular momentum, zonal vorticity and potential temperature, we will obtain a set of dimensionless equations. The new equations are non-dimensionalized using a scaling that follows Schneider and Lindzen (1977), but the zonal velocity u is scaled with Ωa. A detailed description can be found in Cessi (1998).

The non-dimensional model equations are:

$$M_t = \frac{1}{R}\left\{M_{zz} + \mu\left[c^4(c^{-2}M)_y\right]_y\right\} - J(\psi, M) \quad (3a)$$

$$\psi_{zzt} = \frac{1}{(R^2E^2)} yc^{-2}(M^2)_z - \frac{1}{c^{-2}} J(\psi, c^{-2}\psi_{zz}) + \frac{1}{(RE^2c^{-2})}\theta_y + \frac{1}{(Rc^{-2})}\left[c^{-2}\psi_{zzzz} + \mu\psi_{zzyy}\right] \quad (3b)$$

$$\theta_t = \frac{1}{R}\left\{\theta_{zz} + \mu\left[c^2\theta_y\right]_y + \alpha[\theta_E(y,z) - \theta]\right\} - J(\psi,\theta) \quad (3c)$$

The term $J(A,B) = A_yB_z - A_zB_y$ is the Jacobian.

The thermal Rossby number *R*; the Ekman number *E*, the ratio of the horizontal to the vertical viscosity μ and the parameter α are defined as

$$R \equiv gH\,\Delta_H/(\Omega^2a^2)\,;\, E \equiv \nu_V/(\Omega H^2)\,;\, \mu \equiv (H^2/a^2)\,\nu_H/\nu_V\,;\, \alpha \equiv H^2/(\tau\nu_V) \quad (4)$$

The term α is the ratio of the viscous timescale across the depth of the model atmosphere to the relaxation time τ toward the radiative-convective equilibrium.

The boundary conditions for the set of Eq. 3 are:

$$\begin{aligned} &M_z = \gamma(M - c^2),\ \ \psi_{zz} = \gamma\psi_z\ ; \\ &\psi = \theta_z = 0\ at\ z = 0; \\ &M_z = \psi_{zz} = \psi = \theta_z = 0\ at\ z = 1. \end{aligned} \quad (5)$$

Where $\gamma = \frac{CH}{\nu_V}$ is the ratio of the spin-down time due to the drag to the viscous timescale, the bottom drag relaxes the angular momentum *M* to the local planetary value Ωac^2 through a drag coefficient *C*.

The model flow started from an isothermal state at rest and is maintained by a Newton heating function where the heating rate is proportional to the difference between the model potential temperature and a specified radiative-convective equilibrium temperature distribution, which follows the HH80 one:

$$\theta_E = \frac{4}{3} - y^2 + \frac{\Delta_V}{\Delta_H}\left(z - \frac{1}{2}\right). \quad (6)$$

~~We will assume that Eq. 6 is the radiative convective equilibrium temperature distribution of the control experiment; we define a general form of the Eq. 6 as~~

$$\theta_E = \frac{4}{3} - |y|^n + \frac{\Delta_V}{\Delta_H}\left(z^k - \frac{1}{2}\right). \quad (7)$$

Equation 6 is used extensively in dry axisymmetric models (e.g. HH80, Farrell, 1990, Cessi 1998) and it is related to the thermal forcing term of the equation system. A statically stable state as a vertical profile of θ_E is also assumed by Eq. 6. HH80 suggested that the impact of latent heat released by water vapor condensation can be incorporated in dry axisymmetric models by modifying the meridional distribution of θ_E. HL92 followed the HH80 argument and altered the concentration of θ_E under the constraint of equal energy input. The resulting θ_E distributions used by HL92 were peaked distribution on and off the equator resulting in a stronger Hadley circulation with respect the circulation obtained applying Eq. 6. Tandon et al. (2013) used narrow and wide thermal forcing to mimic El Niño or global warming effect on a tropical circulation in a Global Circulation Model. On the opposite side, in fact, we can suppose that if a warmer climate, especially in the tropical regions, happens a very weak gradient of the equilibrium temperature θ_E will be more extent in latitude, expanding consequently the tropical region. This is already occurred in the past, especially in the mid Cretaceous and Eocene when the tropics extended up to 60°. This is the so called equable climate (e.g. Greenwood and Wing, 1995) where the horizontal temperature gradient was weaker than the present one. During those geological ages the temperature was generally higher everywhere, but summing up a constant to the temperature does not change the response of this kind of models. The equator-pole temperature gradient was smaller than the present situation, whereas we prescribe equator-pole θ_E gradient at the surface constant. As we shall show afterwards this is necessary to demonstrate that it is the tropical temperature

gradient rather than the equator-pole one the driver of the Hadley circulation. Thus, in order to study systematically these different conditions we adopt the strategy to build forcing functions dependent on a parameter that controls the θ_E gradient in the tropical regions. Since, with different horizontal distributions of θ_E we can figure out that even the vertical distribution could be affected by some physical mechanisms that make the atmosphere more or less stable than the stratification described by the z component of Eq. 6. The changes of meridional extension and vertical stability can be obtained by changing the exponents of y and z in Eq. 6 transforming Eq. 6 in the following equation:

$$\theta_E = \frac{4}{3} - |y|^n + \frac{\Delta_V}{\Delta_H}\left(z^k - \frac{1}{2}\right). \quad (7)$$

~~The values n and k control the horizontal homogenization of the temperature and the lapse rate respectively. When n=2 and k=1 Eq. (7) becomes the reference equilibrium temperature given in Eq. 6. In many other works (e.g. Schneider 1977, HH80, Caballero et al. 2008), the considered atmosphere is essentially dry; however the distribution of temperature of a dry atmosphere can reflect an action of the water vapor condensation (HL92). Tandon et al. (2013) used narrow and wide thermal forcing to mimic El Niño or global warming effect on a tropical circulation in a Global Circulation Model.~~

~~Starting from Eq. 7 a set of experiments were performed changing n and k in such a way to have a set of numerical results. In order to isolate the contribution of the temperature distribution on the solution of Eq. 3, a set of parameters will be used:~~

$a = 6.4 \times 10^6\ m$ ~~$\Omega = 2\pi/(8.64 \times 10^4)\ s^{-1}$~~

$\Delta_H = 1/3 \qquad \Delta_V = 1/8$

$g = 9.8\ ms^{-2} \qquad C = 0.005\ ms^{-1}$

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$H = 8 \times 10^3 m \qquad \tau = 20\ days$

$\nu_V = 5\ m^2 s^{-1} \qquad \nu_H = 1.86\ m^2 s^{-1}$ ~~(8)~~

The values *n* and *k* control the horizontal distribution of θ_E and its stratification respectively. Small values of *n* are associated with concentrated θ_E distributions. Increasing *n* means increasing broadness of the θ_E distribution. A larger value of *k* increases the vertical stability, especially at upper levels. Thus it comes quite natural to explore the response of Hadley circulation by changing ~~The parameters in Eq. 8 are the same as those used by Cessi (1998).~~

~~The meridional and vertical gradients are controlled by~~ the parameters *n* and *k*, ~~where~~which control the distribution of θ_E, in closest ranges of 2 and 1 respectively, they vary from 0.5 to 3 with a 0.5 step, in such a way that we have a set of 36 simulations. In fact, when *n*=2 and *k*=1 Eq. (7) becomes the reference equilibrium temperature given in Eq. 6 and the experiments performed with n=2 and k=1 will be considered as the reference experiments.

The average ~~temperature~~θ_E along the latitudes and heights are shown in Fig. 1. Heating functions with *n* value equal to 0.5 should not be regarded as unreal, but merely as a simple way to represent a specific state of the atmosphere. The same assertion is valid for all other parameters. As *n* increases the average temperature increases as well, but the meridional gradient decreases~~. High *n* values represent situations with a model atmosphere temperature homogenized along the meridional direction (Fig. 1a).~~ in the tropical regions.

~~With~~ the prescribed ~~temperature~~θ_E as ~~stated in~~specified by Eq. 7, the ~~temperature~~θ_E values at the boundaries and its equator-pole difference temperature remain invariant with respect to *n* ~~and~~, for a specific *k* value. The ~~mean temperature~~energy input is not constant here, which differs from HL92, which analyzed the influence of concentration heating perturbing the forcing function θ_E *(y,z)* in such a way that ~~its average~~θ_E averaged over the

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domain remained constant. It is easily visible in Fig. 1b. Higher *n* values, keeping *k* invariant, have higher ~~mean temperatures~~averaged θ_E at all levels. The same is true for *k*, with higher *k* values, for *n* constant~~, the mean temperature for~~; θ_E at each level is always higher than that with lower k values. The pole-equator ~~temperature~~θ_E difference at upper and lower vertical boundaries are the same for all the experiments~~, but~~ having the same *k*, the ~~meridional (~~vertical~~) temperature gradient~~ averaged θ_E changes as a function of *k*, for *n*, ~~(*k*)~~constant. Whether global warming makes the ~~earth~~equilibrium temperature distribution narrower or wider is beyond the aim of the paper. One can expect that global warming broadens the temperature distribution, but at the same time it could have an impact above all on the ~~SST~~ sea surface temperature (SST) bringing more water in the upper atmosphere which changes the vertical distribution of the temperature in the ~~intertropical convergence zone.~~inter-tropical convergence zone (ITCZ). It is supposed that the oceans force the atmosphere, so we have to allow for the possibility that increasing SST can change the forcing distribution. Increasing uniformly SST might could a poleward expansion as showed by Chen et al. (2013) with an aquaplanet model, but in that case the mechanism was supposed to be related mainly to mid-latitude eddies rather than a tropical forcing. Since other causes can change the temperature distribution of a planet such as changes in the solar activity for instance, we will focus on the temperature distribution regardless of its cause.

~~Although in~~In this model the atmosphere is dry as in many other studies (e.g. Schneider 1977, HH80, Caballero et al. ~~,~~2008), changing the ~~temperature~~θ_E distribution allows for a change in the static stability. Looking at the ~~mean temperature~~average θ_E along the vertical direction, low values of *k* are related to low values of static stability, especially in higher level of the model atmosphere.

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The Brunt–Väisälä frequency ~~$N^2 = g\frac{\theta_z}{\theta}$~~, when the atmosphere reaches the equilibrium will be

$$N^2 = \frac{(gk\Delta_V/\Delta_H z^{(k-1)})}{[4/3 - y^n + \Delta_V/\Delta_H(z^k - 1/2)]}. \quad (\text{~~9~~8})$$

It is clear from Eq. ~~9~~8 that the Brunt–Väisälä frequency does not depend on *n* at the poles and equator. On the contrary, it depends on *k*; large values of *k* imply a more stable atmosphere in the upper levels, especially at poles, making the model atmosphere more similar to the real one, simulating in some respects a sort of tropopause. Moreover, this is equivalent to creating a physical sponge layer in the upper levels of the model that will have some effects on the vertical position of stream function maximum.

Starting from Eq. 7 a set of experiments were performed changing *n* and *k* in such a way to have a set of numerical results. In order to isolate the contribution of the θ_E distribution on the solution of Eq. 3, a set of parameters will be used:

$a = 6.4 \times 10^6\ m$ $\Omega = 2\pi/(8.64 \times 10^4)\ s^{-1}$

$\Delta_H = 1/3$ $\Delta_V = 1/8$

$g = 9.8\ ms^{-2}$ $C = 0.005\ ms^{-1}$

$H = 8 \times 10^3 m$ $\tau = 20\ days$

$\nu_V = 5\ m^2 s^{-1}$ $\nu_H = 1.86\ m^2 s^{-1}$ (9)

The parameters in Eq. 9 are the same as those used by Cessi (1998).

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3 Numerical Results

This section is divided into three subsections, the first showing the results of the model applying the equinoctial condition, when the sun is assumed to be over the equator. The solution is steady as already shown for instance in Cessi (1998). The second subsection will show the results of the model having a ~~temperature~~θ_E distribution described by Eq. 7 but moving following a seasonal cycle. The case n=2 and k=1 is discussed in the third subsection in comparison with previous studies.

3.1 Equinoctial conditions

The axially symmetric circulation is forced by axially symmetric heating as in HH80 and many others and as prescribed by Eq. (7). The model started from an isothermal state and it was run for 300 days, even though it reached its equilibrium approximately after 100 days, in order to be sure that the model does not have instabilities in the long run.

The absolute value of the maximum stream function intensity at the equilibrium conditions for the 36 experiments is shown in Fig. 2. When n=0.5, with k constant, the circulation is always the strongest. The stream function intensity is inversely proportional to n (Fig. 2a). With n=0.5 the experiment resembles the one described in HL92 where they concentrated the latitudinal extent of heating and this led to a more intense circulation. However, they imposed the forcing function $\theta_E(x, y)$ in such a way that its average over the domain remained the same as in the control experiment, i.e. without changing the energy input. They found that concentration of the heating through a redistribution of heat within the Hadley cell led to a more intense circulation without altering its meridional extent. Instead, here, it is evident from Fig. 1 that the experiment with n=0.5 has an energy input lower than the other cases. Nevertheless, the Hadley circulation is always more intense than the other

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cases and contrary to higher n value experiments, the circulation is confined close to the equator. ~~-~~Thus the results of HL92 are extended to a more general case with a lower energy input. It is worth ~~noting~~noticing the constraint of an equal pole-equator gradient of mean ~~temperature~~θ_E is assumed here ~~which is different~~differently from HL92 (Fig. 1a).

The dependence on k is not as straightforward as the one on n, instead. The stream function reaches the highest value for n=0.5 and k=3. With a high n values the Hadley cell stream function intensity is lower and the dependence on k loses its importance. In other words, in our model, the symmetric circulation strength is modulated by k only when the equilibrium temperature distribution is concentrated to the equator.

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Figure 2b shows the maximum zonal wind speed as function of n and k, it is inversely proportional to n, the dependence on k is not as clear as the one on n and when n=3 it almost vanishes in accordance with the behavior of the maximum stream function. These results are in agreement with HL92, who found a stronger zonal wind when the ~~temperature~~forcing was concentrated at the equator.

Some studies define the border of a Hadley cell as that by the zero line of the 500 hPa stream function (e.g. Frierson et al., 2007). ~~Since~~Since in this kind of model the zero stream function is at the poles, it is problematic to define an edge of the Hadley cell based on the zero stream function. Moreover, the circulation intensity changes greatly in our experiments, so it is problematic to define ~~a width~~an edge of the Hadley cell based on ~~the~~an absolute value of the circulation itself. ~~Moreover the stream function goes to zero in the model only at the poles.~~ Hence, we will define the position of the cell equal to the position of the maximum value of stream function, in this way we will study a possible poleward shift of the cell as a function of the two parameters n and k. The ~~width~~edge of the cell ~~will~~might be defined ~~more or less~~ by

values of isolines ~~having 1/4 of~~that are relative with respect to the maximum value, for example 1/4 of the stream function. ~~It is worth noting that~~For the sake of clarity this definition is an operational one and does not ~~resemble~~follow the definition used for example by Dima and ~~Wallance~~Wallace (2003) or Frierson et al. (2007).

The latitude of the maximum stream function value shows a general dependence on *n* and *k*. It increases with *n* and decreases with *k*. However, as shown in Fig. 3a, this dependence is not straightforward or linear, although we have a few exceptions, for instance when $k=n=0.5$. Hence in general when *n* increases, and the total energy input is larger, the stream function is weaker ~~but poleward. This~~and the Hadley cell moves poleward. Although this result is in agreement with other model outcomes (Lu et al., 2008; Gastineau et al., 2008; and Tandon et al., 2013), it is in contrast with the recent observations where a slight strengthening and widening of the Hadley ~~Circulation~~circulation for the past three decades was observed by Liu et al. (2012) and ~~a poleward expansion was also found by~~ Hu and Fu (2007). However, Liu et al. (2012) showed that if the observations start from 1870, the Hadley cell ~~becomes~~has become more narrow and stronger.

The height of the maximum stream function value is confined for almost all the simulations under 2200 m and the general rule is that when *n* increases, the height of maximum lowers, however a few experiments, those with $k=0.5$ and $n=0.5$, 1 and 1.5, have the maximum value between 4300 and 5600 m exhibiting an increase in the height with *n* (Fig. 3b).

In general, the location of the maximum zonal wind speed does not show any evident relationship with the parameters *n* and *k*. It is always confined between 26° and 29° off the equator; however when $n=3$, there is an abrupt transition to about 48°, independently from the

k value. In Table 1, we show the latitude of the maximum wind speed when *k=1* for different *n* values.

~~It is worth calculating the prediction of the cell edge following the HH80 assumptions. HH80 showed that the Hadley cell has a finite width with the edge ending at a specific latitude φ_H and calculated a scaling for this latitude. Starting with the vertically integrated hydrostatic equation and a balanced zonal wind, HH80 obtained a formulation for θ in the limit inviscid:~~

$$\bar{\theta}(\phi) = \bar{\theta}(0) - \bar{\theta}_0 \frac{\omega^2 a^2 sin^4(\phi)}{2gH cos^2(\phi)} \quad (10)$$

~~Assuming continuity of the potential temperature $\bar{\theta}(\phi_H) = \bar{\theta}_E(\phi_H)$ and conservation of vertically averaged potential temperature $\int_0^{\phi_H} \bar{\theta} cos(\phi) d\phi = \int_0^{\phi_H} \bar{\theta}_E cos(\phi) d\phi$, they found φ_H as a function of the Rossby number R. The hypothesis of continuity and conservation of potential temperature is equivalent to solving those two equations by means of a geometric "equal area" construction.~~

~~For the general case described by Eq. 7, we calculated a similar relationship between φ_H and the Rossby number, assuming the same hypothesis of HH80 described previously.~~

$$R = \left[\frac{(n+1)}{2n}\right] \frac{\left[\left(y_H^5/(1-y) + y_H + 1/3 y_H^3 - 1/2 ln((1+y_H)/(1-y_H))\right)\right]}{y_H^{(n+1)}} \quad (11)$$

~~The *n* value goes from 0.5 to 3. Equation 11 should be compared with Eq. 17 of HH80. We represent the solutions of Eq.11 in Fig. 4. Evidently for R=0.121 (that is the Rossby number used here), there is no agreement when *n*=3 between analytic solution that predicts a smaller φ_H and the numerical one, which differs from all the other solutions that are quite close to one another.~~

~~This behavior could be related to the diffusivity. Figure 5 shows the analytic and numerical vertically averaged potential temperature for n=k=3. It is evident that not only $\bar{\theta}$ is not conserved when n=3 and there is not redistribution of energy, hence the assumption made by HH80 about the continuity and conservation of potential energy and that leads to Eq. 11 never takes place in the numerical model for n=3. However, when the vertical diffusion is very small or k has low values $\bar{\theta}$ approaches to $\bar{\theta}_E$ (not shown) but the jet streams have their maxima at the mid latitudes.~~

~~Figure 6 shows the stream function and the zonal wind speed for the experiments n=k=0.5 (Fig. 6a) and n=k=3 (Fig. 6b~~The difference between θ_E and θ, once the model reaches the equilibrium, is quite interesting. Figure 4 shows meridional distributions of θ_E and θ for n=3 and k=0.5, 1 and 3. In Fig. 4a, θ_E is underθ, when k=1 we find θ_E is over θ in a region around the equator (Fig. 4b), with θ_E crossing θ at about 47°, finding again the equal area condition suggested by HH80 and that explains even the jet location, whereas in Fig.4c, with k=3, we can see how θ_E is over θ. Despite these differences in the distributions of θ_E and θ the model produces with these different k values almost the same solution, in terms of circulation strength and jet location. For other values of n the situation is similar, but with more peaked distributions the differences are not so remarkable.

We can understand these findings in the light of Cessi (1998) results obtained by expanding the variables M, θ and ψ in power series of R. The term R^2 in nonlinear expansion part, the meridional advection depends on the differences between θ_E and θ, on the cube of the meridional temperature gradient, and linearly on the imposed stratification deducing that for unstable stratifications, this term would appear as a negative diffusivity term. This seems to be the case, in our simulation when k=0.5. The thermal energy obtained in the model is larger

than the imposed temperature (Fig.4a). Although the stratification imposed by Eq. 7 is stable, i.e. $\frac{\partial \theta_E}{\partial z} > 0$, the second derivative is negative when k=0.5 reducing the stability at upper levels, so it can be thought as a way to simulate the effect of the latent heat released by water vapor condensation. In any case the model acts to bring the vertical temperature gradient in a more stable configuration and a Hadley circulation is in any case reproduced demonstrating the robustness of the model.

With n getting larger, the θ_E distribution becomes flatter in the tropical region and θ clamps to θ_E. In general, we expect that a vigorous circulation occurs in a fast rotating planet unless the thermal gradient becomes small in the tropics. In such a case the angular momentum homogenization is equivalent to a weaking of the rotation (Cessi, 1998). If the circulation is proportional to the cube of the meridional temperature gradient, it is quite evident that when such a gradient has high values the circulation is vigorously driven by this term, whereas when it approaches to zero it is the term $\theta_E - \theta$ dominates. ~~). The parameter n controls the Hadley cell and jet stream widths. The experiment with n=k=0.5 has Hadley cells and jet streams quite narrow. As far as the vertical position of the maximum value of the stream function is concerned, the experiments with k=0.5, 1 and 1.5 exhibit particular behavior with respect to the other experiments. The stream function has its maximum at upper levels. It is likely that such a combination of the parameters favors air to move to higher levels with respect to experiments with higher k values.~~

HH80 found that the edge of the Hadley cell was at the mid-latitudes when the planetary rotation was lower than that of the earth. Since this phenomenon is here observed for a wider ~~temperature~~forcing distribution, this common result may be attributed to a low efficiency in the process of homogenization of momentum and temperature.

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In order to explain equable climates like those supposed to be occurred in Cretaceous and Eocene, Farrell (1990) formulated an axisymmetric model starting from the Held and Hou model and a forcing with n=2 and k=1 where the temperature gradients became flat because of a dissipation term. For high values of n the θ distributions are similar to those obtained by our forcing conditions. In some respects, flattening of forcing distributions is equivalent to have the same dissipation term in the Farrell (1990) model.

Figure 5 shows the stream function and the zonal wind speed for the experiments n=k=0.5 (Fig. 5a) and n=k=3 (Fig. 5b). The parameter n controls the Hadley cell and jet stream widths. The results show that such with n=k=0.5 the Hadley cell and jet streams are quite narrow. As far as the vertical position of the maximum value of the stream function is concerned, the experiments with k=0.5, 1 and 1.5 exhibit particular behavior with respect to the other experiments. The stream function has its maximum at upper levels. This is related to the different stratification imposed by the parameter k. Stratification with low values of k favor air to move to higher levels with respect to experiments with higher k values.

3.2 Time-dependent simulations

Since heating depends on solar irradiation, it is of interest to analyze the solutions obtained by the annually periodic thermal forcing and to compare it with the steady solutions described previously in this paper. Starting from Eq. (7), we can formulate ~~a~~an equilibrium temperature distribution having the maximum heating off the equator at latitude ~~-~~y_0~~÷~~:

$$\theta_E = \frac{4}{3} - |y - y_0|^n + \frac{\Delta_V}{\Delta_H}\left(z^k - \frac{1}{2}\right). \quad (\text{~~12~~}10)$$

where y_0 in Eq. (9) is dependent on time according to

$$y_0(t) = sin\left(\frac{\varphi_0 \pi}{180}\right) \cdot sin\left(\frac{2\pi t}{360 days}\right)$$

(~~13~~11)

where φ_0 is the maximum latitude off the equator where heating is maximum. Equations ~~12~~11 and ~~13~~12 are the same used by Fang and Tung (1999) with the choice of maximum extension of φ_0 consistent with the choice of Lindzen and Hou (1988), i.e. $\varphi_0 = 6$~~°.~~°. A prescribed equilibrium temperature varying seasonally makes the simulations more realistic. As described previously, here we will focus on the average and maximum values, in absolute terms, of the stream function and zonal speed obtained during 360 days of simulations. The averaged values are obtained in these cases by averaging the outputs obtained every 30 days, starting from the minimum corresponding to the summer Hadley cell in the boreal hemisphere.

The annual averages of the time-dependent and equinoctial circulations shows that maximum stream ~~function s~~functions and zonal wind speeds behave quite similarly ~~in the annual averages of the time dependent and equinoctial circulations. Nevertheless the time-dependent solutions never attain the symmetric circulation obtained by averaging the single snapshots~~ (Fig. ~~7). Even the real earth circulation never reaches the mean~~6), nevertheless the instantaneous Hadley circulation~~, even because of eddies, but symmetric Hadley cells are visible in~~ almost never resembles the ~~average~~modeled circulation (Fang and Tung, 1999) as well as the real one (Dima and Wallace, 2003).

The maximum stream function is obtained here ~~by the~~when k=n=0.5 (Fig. ~~7a~~6a). In general, for k=0.5, we have stronger circulations and winds. It has confirmed the tendency to a weaker stream function and wind speed when n increases. However, the circulation strength expressed as averaged value is weaker ~~in~~than the time-dependent solution, when n is low and

k is high, otherwise it is stronger, but it is never twice as strong as that of the equinoctial solution as found by Fang and Tung (1999). When n=2 and k=1 it ~~appears more~~is consistent with the results of Walker and Schneider (2005) as discussed in the Subsect. 3.3. The maximum zonal wind speed shows a behavior slightly different from the stream function intensity; there is a clear dependence on n and k. For example, there is not an analog maximum when n=0.5 and k=3 found in the steady solution and thus for other k values where the stream function has a relative maximum. With high k value the static stability is high at upper levels and the maximum of circulation remains confined to lower levels prevents air upwelling at the high levels. Thus, the transfer of momentum to high level is less effective with respect to the case k=0.5 where it is favored, instead.

The meridional position and the height of the stream function maximum shows that there is no clear dependency on n and k (Fig. ~~8~~7). The difference between the time-dependent simulations and the average of the ~~non time dependent simulations~~steady solutions is quite interesting. ~~-~~ It is to be noticed that the latitude of the stream function maximum in the time-dependent solution is in the range of 12.5° and 16° (Fig. ~~8a~~7a), whereas in the equinoctial solutions the correspondent latitude is within a ~~-~~larger range. It is probable that this more narrow interval is due to the averaging operation. The maximum stream function is located at higher levels, between 4500 and 6000 for n less than 2.5. Otherwise the maximum is positioned under 2500 m except when n=3 and k=0.5 (Fig. ~~8b~~7b).

More than the steady solution, it is evident that the height of the maximum stream function is lower when k=3. In the steady solution this phenomenon is not that evident. When k=3, the vertical gradient of ~~the potential temperature~~ θ_E is higher in upper levels and it prevents, evidently more than the equinoctial solution, air from moving higher leaving

circulation occurring at lower levels. The case k=3 is equivalent to imposing a "natural" sponge layer at the top of the model. Thus it does not come as a surprise that the maximum stream function is lower than those observed in simulations with other k values. This result is analogous to that of Walker and Schneider (2005) that removed the maximum stream function at higher levels found by Lindzen and Hou (1988) adopting a numerical sponge layer at the top of the model. A comparison with previous works of the simulations with n=2 and k=1 will be discussed in the Subsect. 3.3.

The position of the jet stream is almost similar to the one observed in the steady solution. It is confined between 28° and 30°, with latitude of averaged jet remaining almost at the same place ~~or moving equatorward with n~~, except when n=3 the jets are located at about 44° confirming the abrupt transition of the jet stream position when n=3 already found for the equinoctial experiment. Fu and Lin (2011) suggest that the jets moved poleward of about 1° per decade in the last several years but Strong and Davis (2007) observed that Northern hemisphere subtropical jet shifted poleward over the east Pacific, while an equatorward shift of the subtropical jet was found over the Atlantic basin. Excluding the case n=3, all the other subtropical jets in the different experiments have the position of the maximum very close to one another and the shifting range is very limited. ~~However, when we use the jet latitude to define the edge of the Hadley cell, there is no significant shift but when n=3. This appears to be in contrast with the Held and Hou model.~~Thus, when a vigorous circulation occurs the jet location must be located at about 30°, whereas reducing too much the tropical gradient the process of homogenization becomes weaker like in a slow rotating planet and this is confirmed in the time-dependent solution. Both Tandon et al. (2013) and Kang and Polvani (2011) found a discrepancy in this area with the jets that do not follow the Hadley cell edge. In an axisymmetric model, defining the Hadley edge as a function of the stream function and

connecting it to the jet location is problematic because of lacking of a zero value of the stream function.

Figure ~~9~~8 shows the annually averaged circulation for the same cases as shown in Fig. ~~6~~5, which is obtained by annually averaged heating. It is impressive how the steady and time-dependent solutions resemble each other. As in Fang and Tung (1999) the annual mean meridional circulation has the same extent, but differently from them the strength of the annual mean circulation of the time-dependent solution is almost the same of the steady solution.

When the heating center is off the equator the intensity of the winter cell is stronger, whereas the cell of the summer hemisphere is weak and sometimes almost absent. Figures ~~10~~9 and ~~11~~10 show the maxima of the stream function and zonal wind speed at the winter solstitial as a function of *n* and *k*. The maximum stream function as a function of *n* and *k* has the same configuration of the steady solution. Here, as expected the intensity of the meridional circulation (Fig. ~~10a~~9a) is twice as strong as that of the steady solution. The zonal wind has a different configuration instead, the maximum zonal wind speed is obtained when *n*=1 (Fig.~~11~~10).

We can inspect a couple of simulations when the stream function reaches its maximum in the boreal hemisphere. Figure ~~12~~11 shows the stream function and the zonal wind speed when *n*=2 and *k*=0.5 (Fig. ~~12a~~11a, b) and *n*=2 and *k*=3 (Fig. ~~12 b~~11 c, d). When *k*=0.5 (upper panels) the boreal (winter) circulation is much stronger when *k*=0.5, with the austral (summer) circulation almost absent. The vertical extent is larger and the maximum is located at higher levels. The summer and winter jets are both more intense than their counterparts for *k*=3. The tropical easterly winds are in this case stronger than those for *k*=3 (13.8 ms^{-1} vs 11.4 ms^{-1}) and

the easterly region is also wider. When k=3, it is noted that the ~~boreal~~winter cell is located closer to the equator than the ~~austral~~summer cell (not easily visible in the figure when k=0.5).

3.3 A discussion on the case n=2 k=1

When n=2 and k=1, corresponding to the classic case discussed in many studies, we found that the time-dependent solution is only slightly stronger than the steady solution. Lindzen and Hou (1988) proposed a study of the Hadley circulation in which the maximum heating was 6° off the equator. In their non-time-dependent model, the solution showed an average circulation much stronger by a factor 15 for $\phi_0 = 6°$with respect to the equinoctial solution. ~~–~~Lindzen and Hou (1988) suggested that this exceptional strength was due to a nonlinear amplification of the annually averaged response to seasonally varying heating, although Dima and Wallace (2003) in a study on the seasonality of the Hadley circulation did not observe any nonlinear amplification.

With the parameters used for equinoctial and time-dependent simulations we performed an experiment like that of Lindzen and Hou (1988), with $\phi_0 = 6°$ that will be referred to as solstitial experiment. We found that the winter circulation is stronger by a factor three with respect to the steady solution obtained with the equinoctial heating consistent with the result of the axisymmetric model in Walker and Schneider (2005). However, the average circulation~~,~~ obtained by averaging two solstitial experiments, with ~~-~~$\phi_0 = 6°$ and ~~-~~$\phi_0 = -6$~~°,~~° respectively is only 1.5 times stronger than the steady solution with $\phi_0 = 0°$ and it has a maximum in the upper levels of the model domain as in Lindzen and Hou (1988). We suggest that this maximum is due to a numerical effect caused by averaging the single solstitial experiments rather than a spurious effect caused by the rigid lid as suggested in Walker and Schneider (2005), even though a sponge layer actually lowers ~~this~~the maximum stream function height

and we can see the effects of a stronger vertical gradient in the upper levels especially in the time-dependent solution (cf Fig. 3 and Fig. ~~9).~~ 7). Single solstitial experiments did not show a maximum in upper levels and so the equinoctial and time-dependent experiments (Fig. ~~13~~12). Consequently the only operation performed to produce Fig. ~~13c~~12c, which exhibits the upper levels maxima was to average the two solstitial experiments, which causes the maximum at upper levels.

Finally, we notice that comparing a time-dependent solution with $\varphi_0 = 6°$ with the equivalent steady solution having the heating off the equator is not properly correct, since for the time-dependent model φ_0 represents only the maximum extension of heating, hence a more correct comparison between time and no time-dependent solutions should be performed with the time-dependent solution having$\varphi_0 = 3°$. In such a case, the average solution is only slightly weaker than the Hadley circulation driven by annually averaged heating or by a time-dependent heating which does not show any maximum in the upper levels. Thus, the results of equinoctial, time-dependent and solstitial ($\varphi_0 = 3°$) experiments are mutually consistent.

4 Conclusions

The ~~temperature distribution~~forcing of an Earth-like planet can change for several reasons. For instance, a change of ~~the temperature~~forcing distribution can be caused by different factors such as global warming or long-term variation of solar activity.

Under the assumption of an equal equator-pole difference at the surface we used an axisymmetric model to study the sensitivity of the tropical atmosphere to different ~~temperature~~θ_E distributions modulated by two parameters, *n* that controls the broadness of the distribution and *k* that modulates how the ~~temperature~~ $\theta_E$is distributed vertically. Equinoctial and time-dependent solutions were simulated and compared. Moreover for the case *n*=2 and

k=1, corresponding to the classical distribution used in literature, a few solstitial experiments were also run. When n=2 and k=1, the annually averaged circulation of equinoctial, time-dependent and solstitial experiments are quite close to one another, consistent with the results of Walker and Schneider (~~2006~~2005). However, the results differ from those of Lindzen and Hou (1988) and Fang and Tung (1999). As in all those works the maximum of the stream function of the solstitial experiment ~~appears to be~~is at upper levels, but it seems to be related to a spurious effect of the averaging operation rather than a spurious effect due to the rigid lid.

The results provide evidence that concentrated equilibrium temperature distributions enhance the meridional circulation and jet wind speed intensities, confirming findings of Lindzen and Hou (1988) even though these authors imposed the same energy input. However, in the present study the concentrated distribution at the equator has lower energy input.

The width of the Hadley cell is proportional to n, but when the cell width increases its intensity decreases. ~~The~~Since the equator-pole gradient is the same for all the experiments~~;~~ ~~hence~~ with the same k; it ~~is~~evident that the gradient ~~equator-subtropic that~~ in the tropical region controls the circulation strength. ~~Even k, hence the lapse rate,~~The term k controlling the imposed stratification has influence on the actual temperature distribution that can differ remarkably from θ_E distribution.

Vertical stratification is important in determining the position and intensity of the Hadley cell and jet when n is low, whereas k loses its importance when the ~~temperature~~θ_E distribution is wider. This latter result is consistent with results of Tandon et al. (2013) who found that the Hadley cell expansion and jet shift had relatively little sensitivity to the change ~~in~~of the lapse rate. Consequently, the subtropical jet stream intensities are controlled by the broadness of horizontal temperature rather than the ~~vertical lapse rate~~stratification, with higher values of

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the jet when the thermal forcing is concentrated to the equator. However, results show that the jet stream position does not show any dependence with n and k, except when the ~~temperature~~ θ_E distribution is the widest (n=3); in such a case an abrupt change occurs and the maximum of the zonal wind jet is located at mid-latitudes. The analytic study of this model performed by Cessi (1998) suggest that when the meridional gradient becomes too small the process of homogenization of temperature and momentum occurs slowly and the circulation behaves as that of a slow rotating planet.

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Table 1. Latitudes (in degrees) of the maximum wind speed for the equinoctial and time-dependent solutions when k=*1* as a function of the parameter *n*.

n	0.5	1	1.5	2	2.5	3
Equinoctial	27.4	28.7	27.4	26.1	28.7	47.7
Time dependent	28.7	28.7	28.7	27.4	27.4	44.4

Figure Captions

Figure 1. Meridional (a) and vertical (b) ~~mean~~average of non-dimensional equilibrium temperature as a function of n with k=1(panels a) and ~~b)~~k with n=0.5, 1 and ~~k~~1.5 (panel b~~)~~).

Figure 2. Maximum non-dimensional stream function (a) and zonal wind speed [ms^{-1}] (b) as function of parameters n and k for the steady solution.

Figure 3. Latitude [degree] (a) and Height [m] (b) of maximum non-dimensional stream function.

Figure 4. ~~Relationship between Rossby number and poleward boundary of the Hadley cell as given by Eq. 10, $y_H = \sin\varphi_H$.~~Vertically averaged the θ (blue line) and θ_E (red line) for the simulations with n=3 and k=0.5 (a), k=1 (b) and k=3 (c).

Figure 5. ~~Vertically averaged potential temperature for the simulation with n=k=3.~~

~~Figure 6.~~ Non-dimensional stream function (~~a and c~~contours) and zonal wind speed ~~(b and d~~[ms^{-1}] (colors) for the steady cases n=0.5, k=0.5 (~~upper panels~~a) and n=3, k=3~~, k=3 (lower panels~~ (b).

Figure ~~7~~6. Maximum of annually averaged non dimensional stream function (a) and zonal wind speed [ms^{-1}] (b) as function of parameters n and k for the time-dependent simulations.

Figure ~~8~~7. Latitude [degree] (a) and Height [m] (b) of maximum annually averaged non-dimensional stream function for the time-dependent solution.

Figure ~~9~~8. Annually averaged non-dimensional stream function (~~a and c~~contours) and zonal wind speed ~~(b and d~~[ms^{-1}] (colors) for the steady cases n=0.5, k =0.5 (~~upper panels~~a) and n=3, k=3 (~~lower panels).~~b)

Figure 9.~~Figure 10.~~ Maximum of non-dimensional stream function (a) and zonal wind speed [ms^{-1}] (b) -as function of parameters n and k for the time-dependent simulations.

~~Figure 11.~~Figure 10. Latitude [degree] (a) and Height [m] (b) of maximum non-dimensional stream function for the time-dependent solution.

Figure ~~12. Winter~~11. Boreal winter circulation, non-dimensional stream function (a and c) and zonal wind speed [ms^{-1}] (b and d) ~~when~~for the time-dependent simulation with n=2, k=0.5

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(upper panels) and n=2, k=3 (lower panels~~) for the time dependent simulation~~). Dashed lines indicate negative values.

Figure ~~13~~12. Non-dimensional annually averaged stream function (~~a,c,e~~contours) and zonal wind speed ~~(b,d,f~~[ms^{-1}] (colors) when n=2 and k=1 for the steady (a), time-dependent (b) and with the maximum heating 6° off the equator (c).

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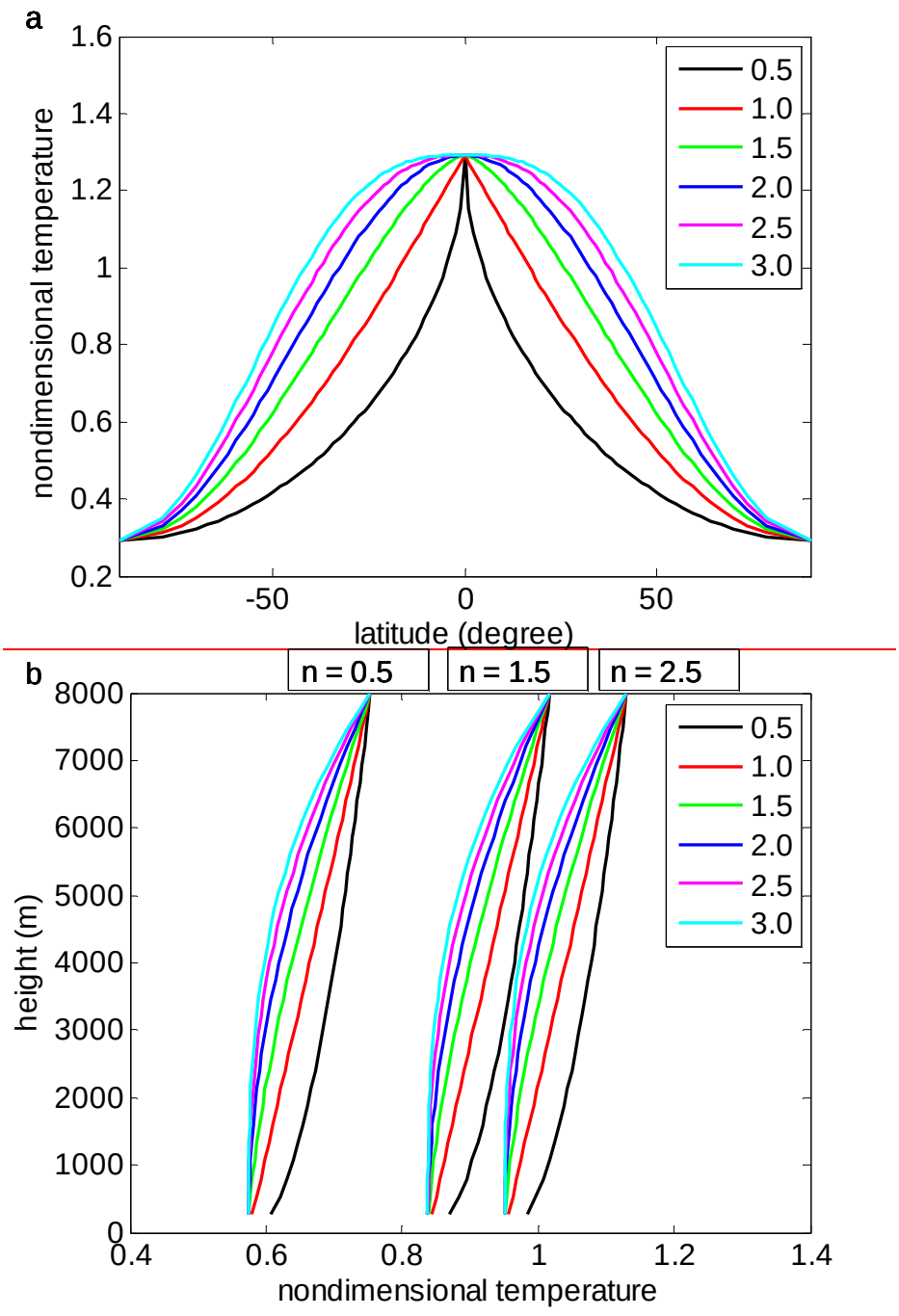
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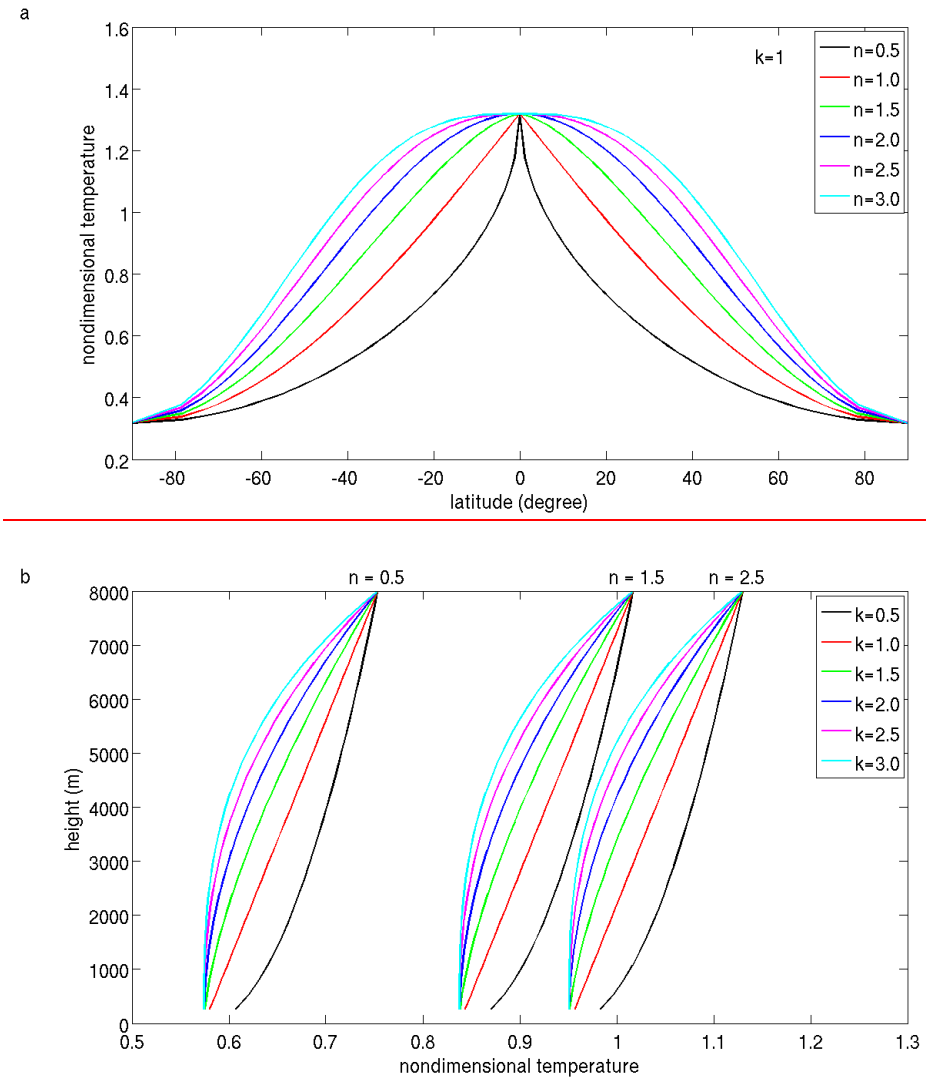
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Figure 1.

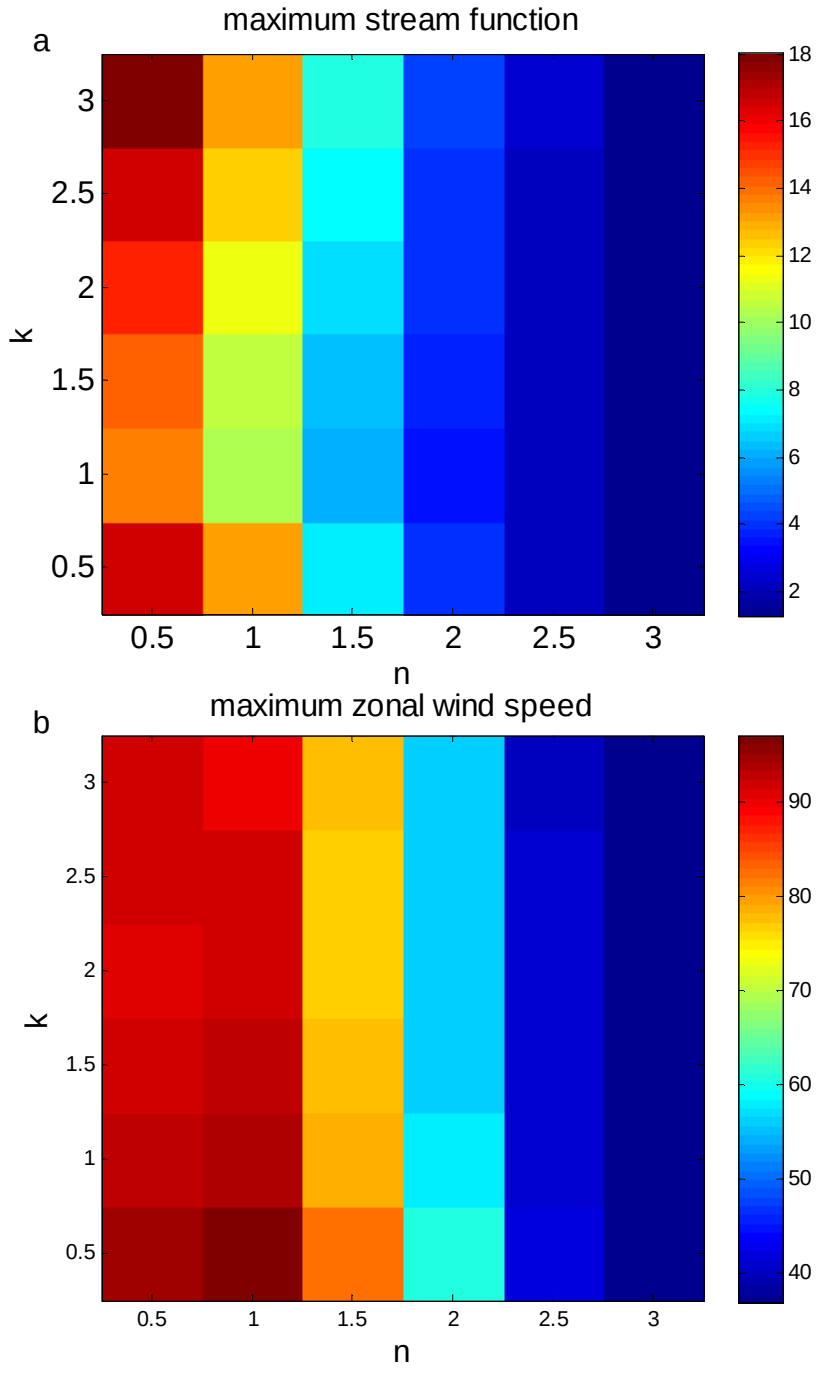


Figure 2.

Formattato: Allineato al centro

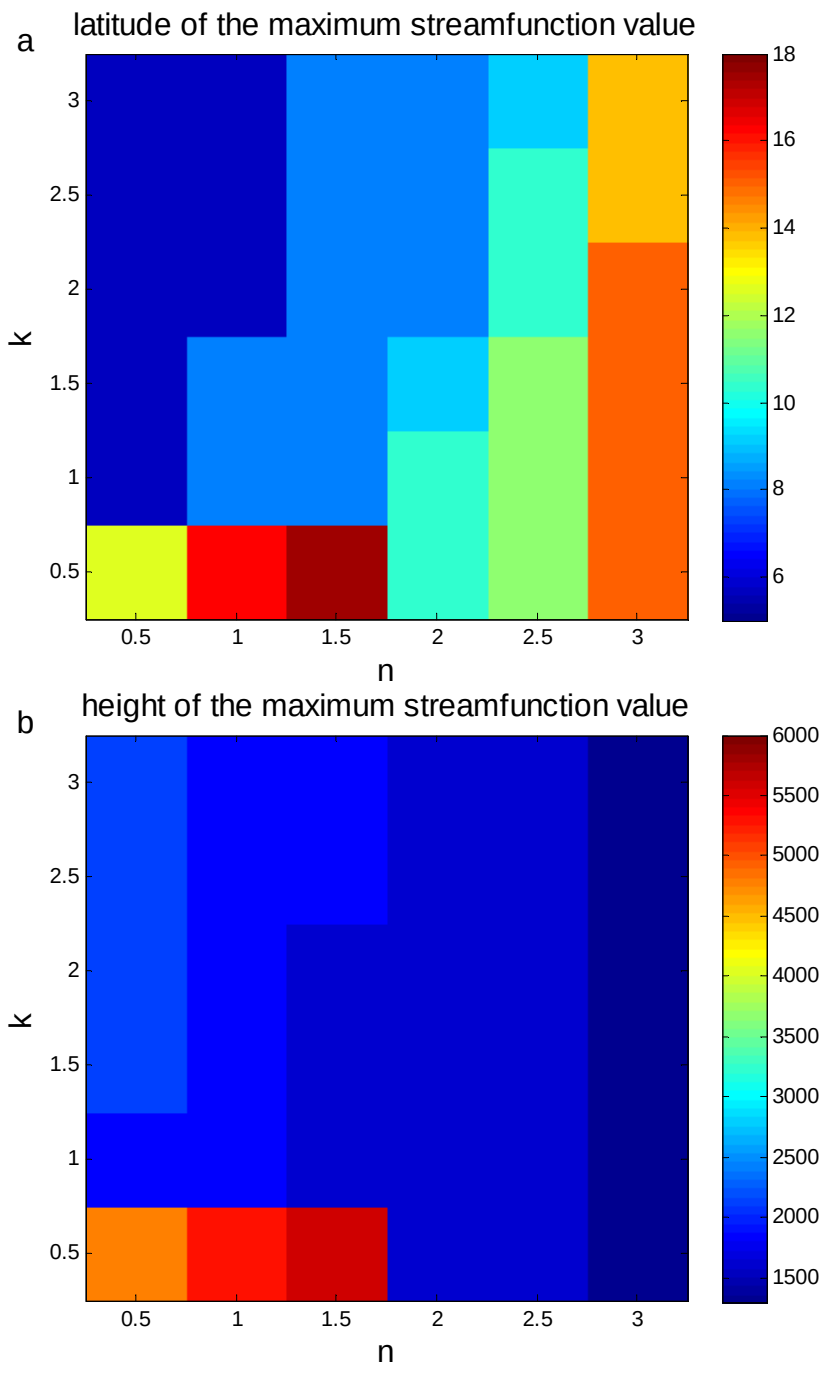
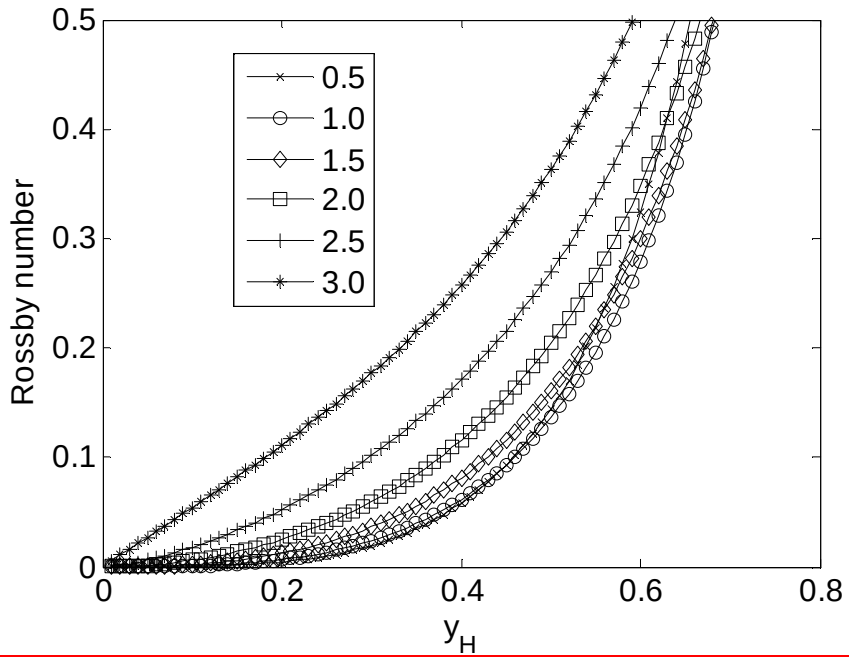
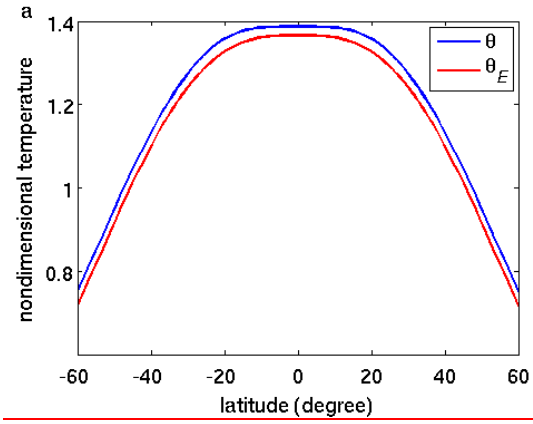


Figure 3.

Formattato: Allineato al centro


0.5
1.0
1.5
2.0
2.5
3.0
Rossby number
0
0.1
0.2
0.3
0.4
0.5
0
0.2
0.4
0.6
0.8
y_H

1


a
θ
θ_E
nondimensional temperature
0.8
1
1.2
1.4
-60
-40
-20
0
20
40
60
latitude (degree)

2

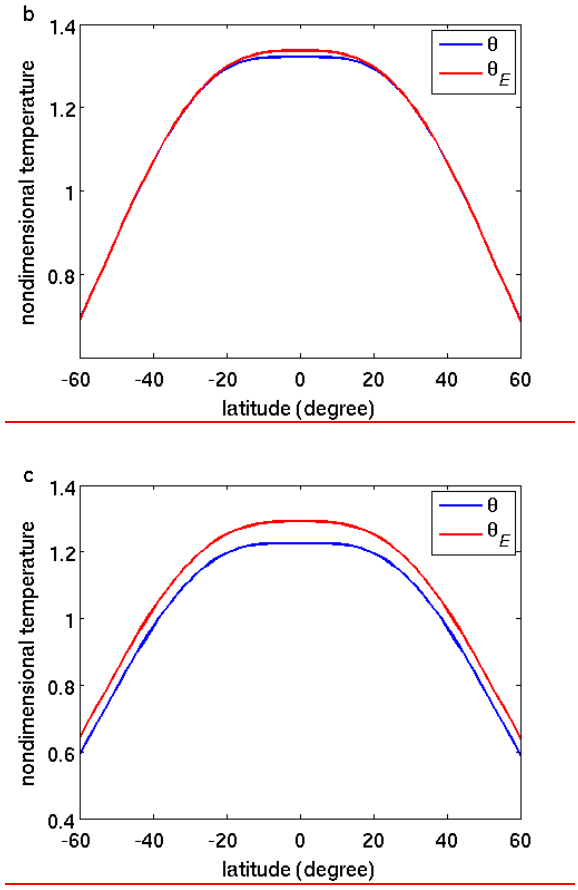
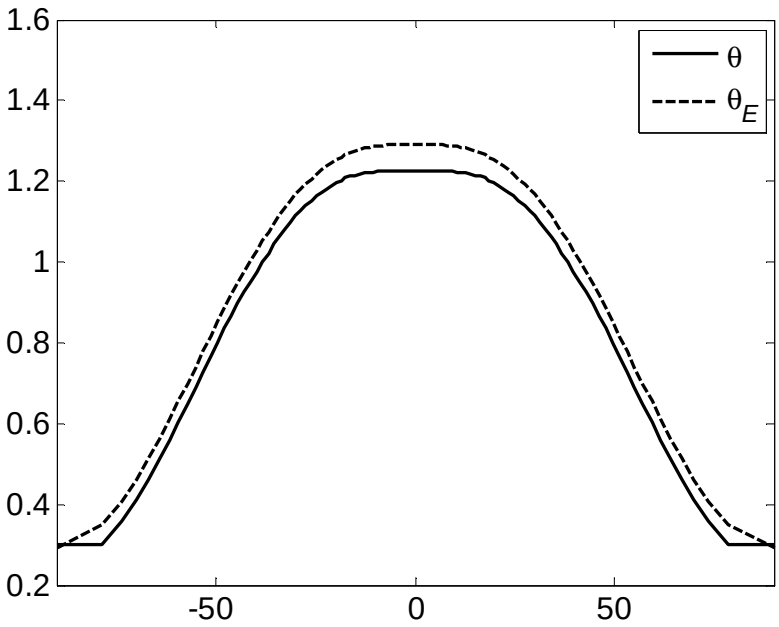


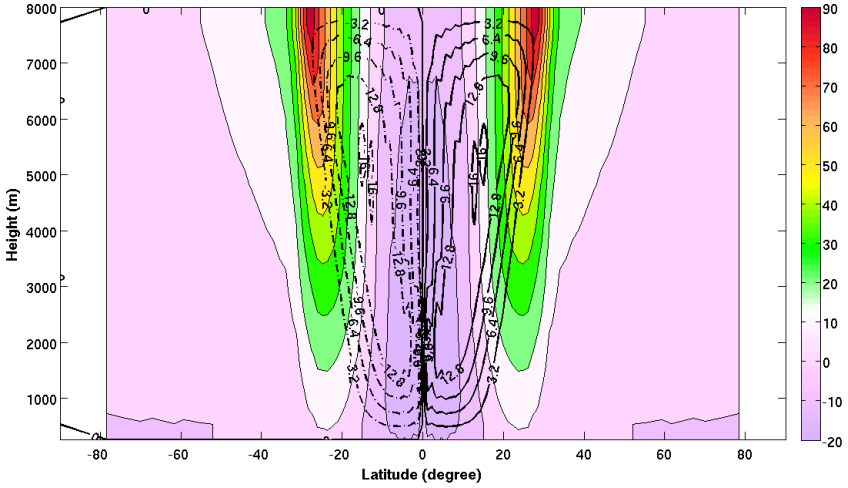
Figure -4.

Formattato: Allineato a sinistra


1.6
1.4
1.2
1
0.8
0.6
0.4
0.2
-50
0
50
θ
θE

1

a


Height (m)
Latitude (degree)
8000
7000
6000
5000
4000
3000
2000
1000
-80
-60
-40
-20
0
20
40
60
80
90
80
70
60
50
40
30
20
10
0
-10
-20

2

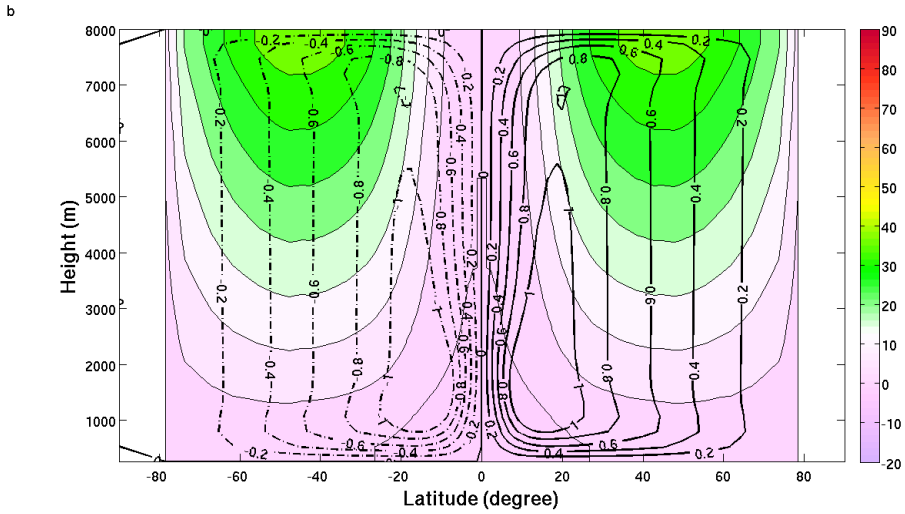
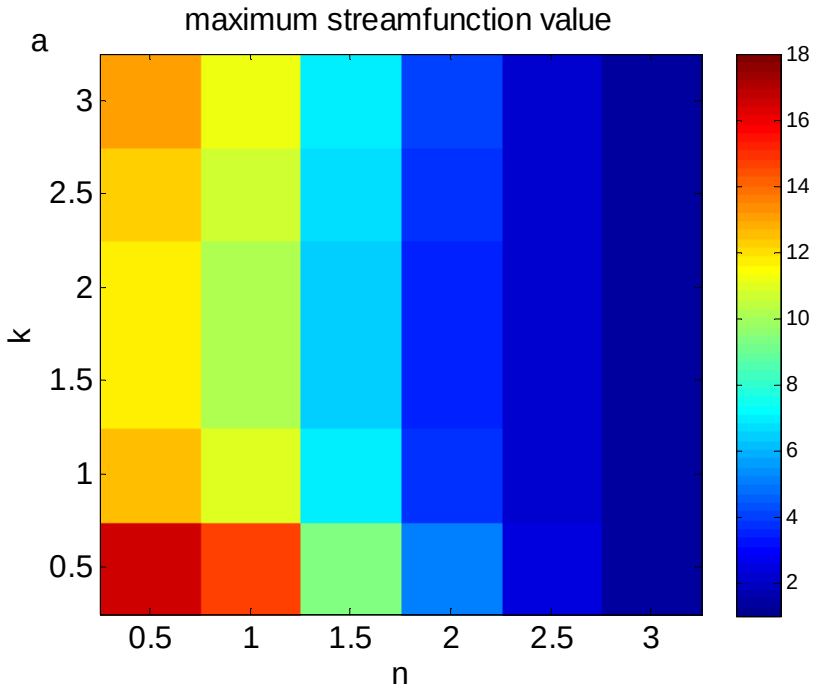


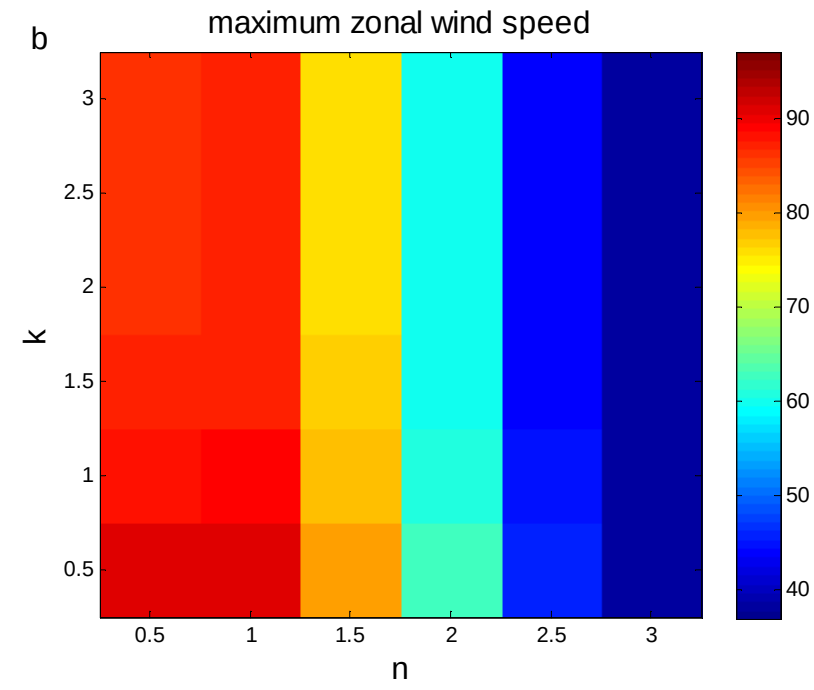
Figure 5.

Formattato: Normale

Formattato: Tipo di carattere: Times New Roman, 12 pt, Inglese (Stati Uniti)



1



2

Formattato: Inglese (Stati Uniti)

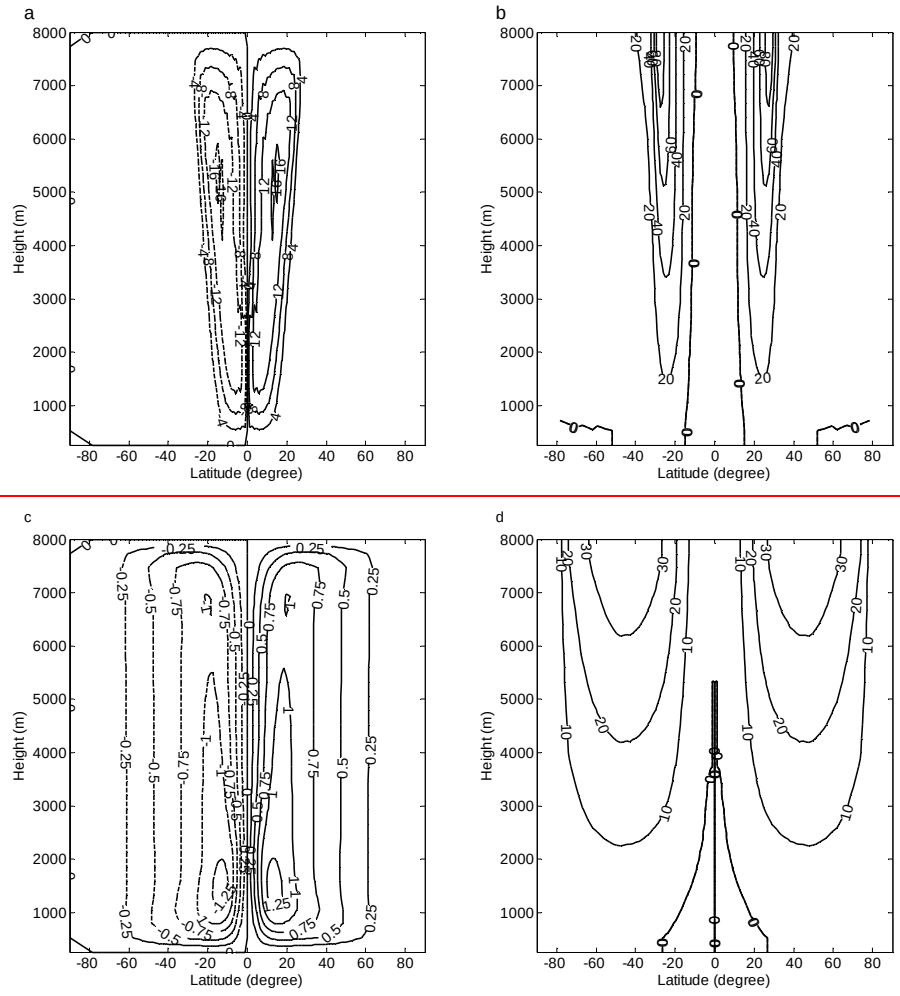
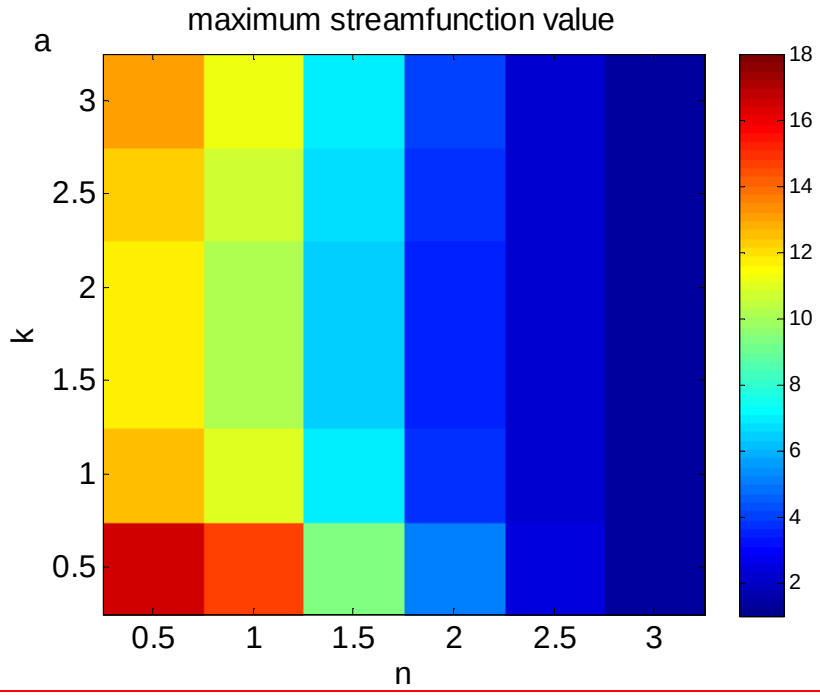
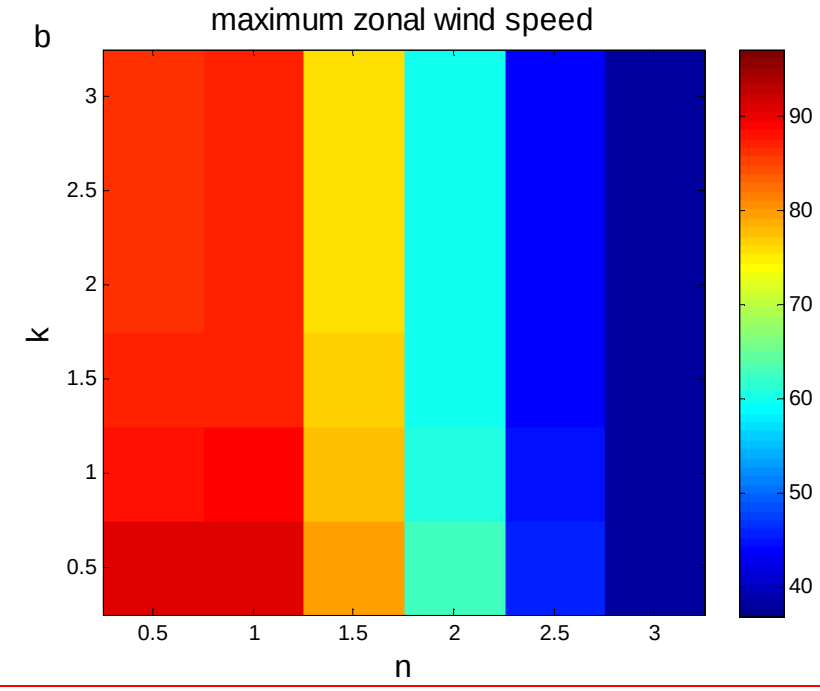

a
b
c
d
Height (m)
Latitude (degree)
8000
7000
6000
5000
4000
3000
2000
1000
-80 -60 -40 -20 0 20 40 60 80

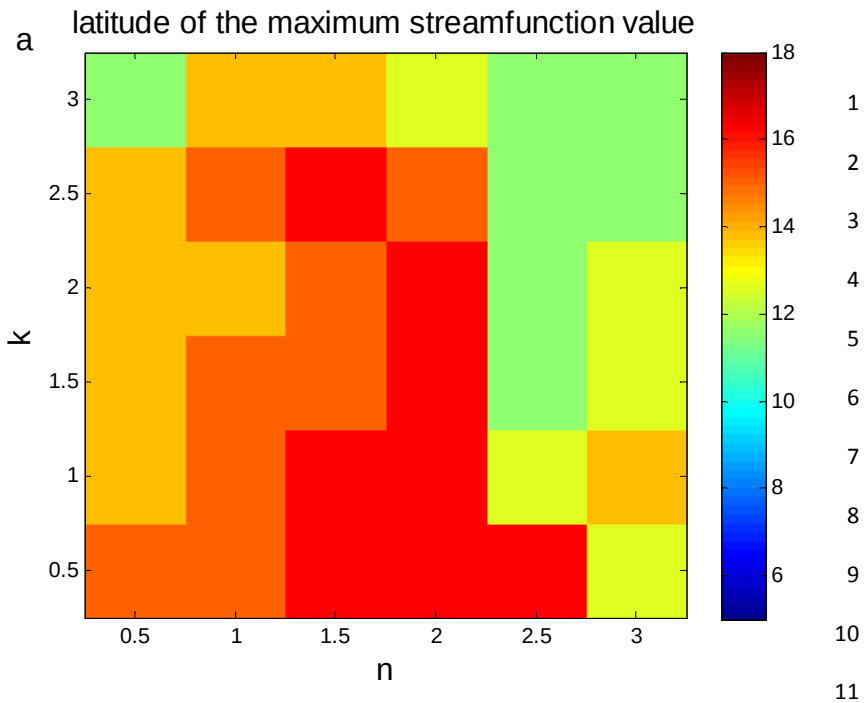
Figure 6.

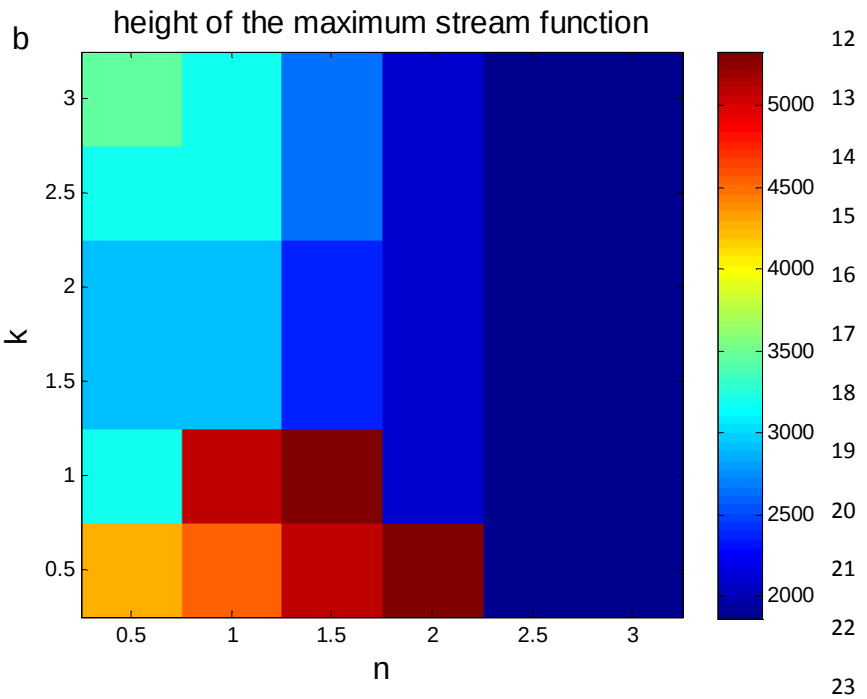

maximum streamfunction value
a
k
n
0.5
1
1.5
2
2.5
3
18
16
14
12
10
8
6
4
2

1


maximum zonal wind speed
b
k
n
0.5
1
1.5
2
2.5
3
90
80
70
60
50
40

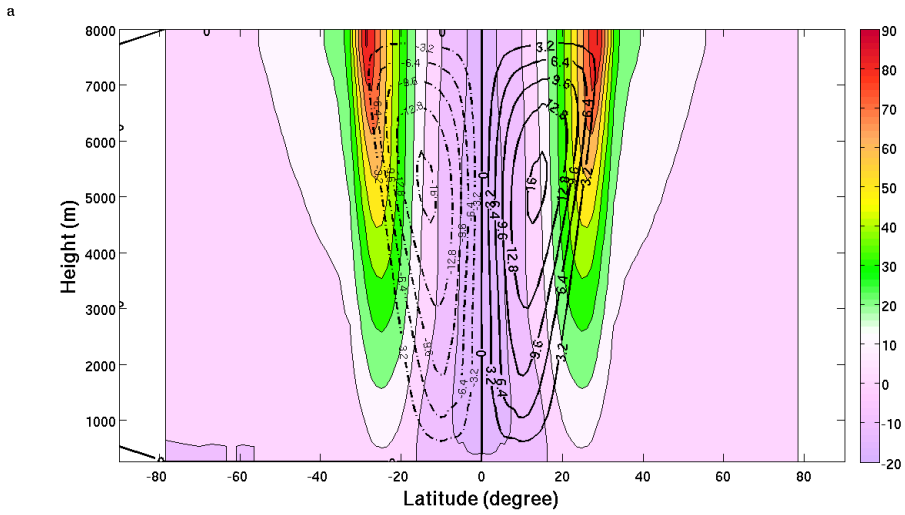
2

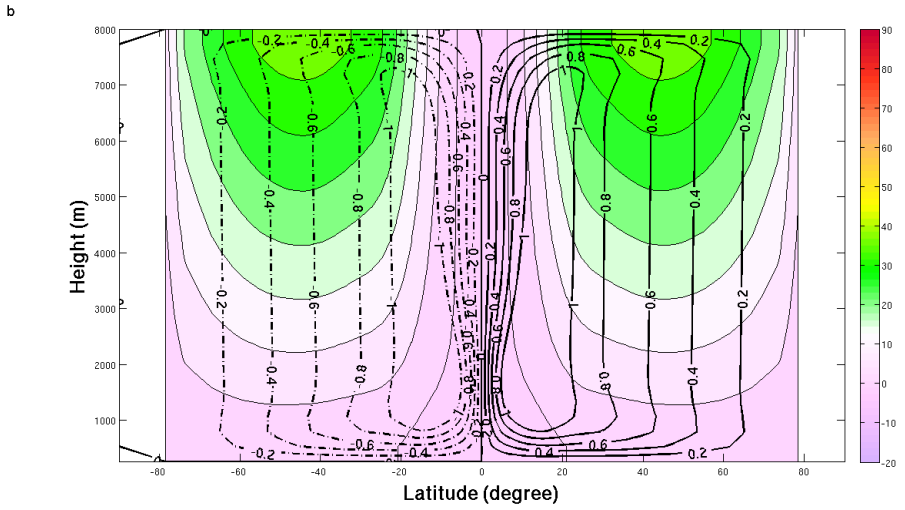




Formattato: Giustificato

Figure 7.





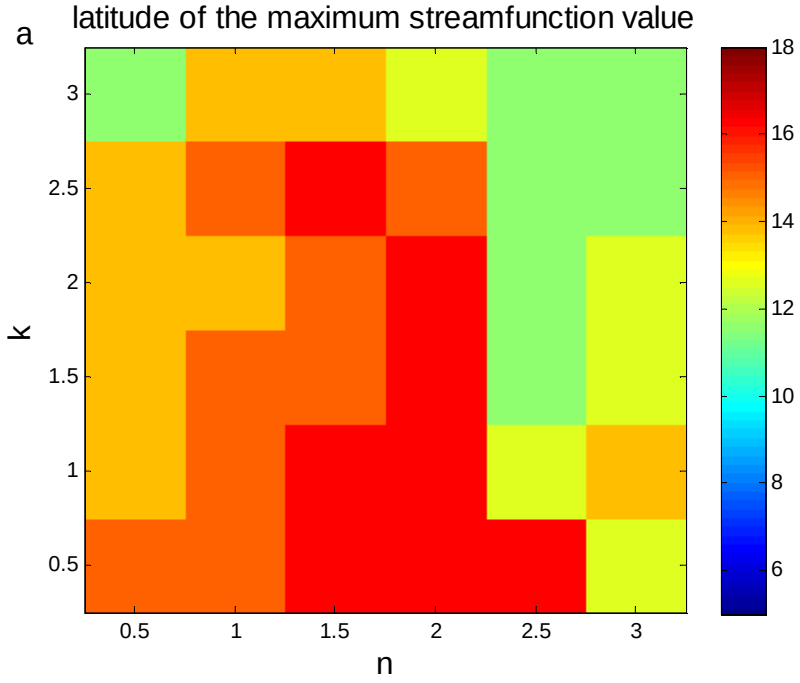
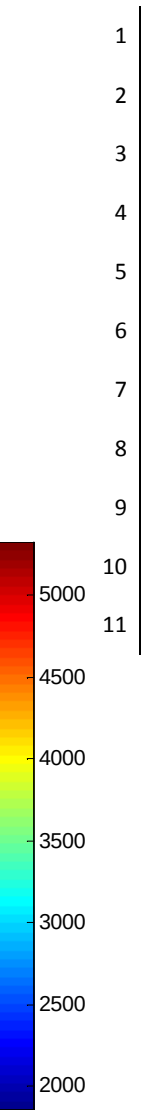
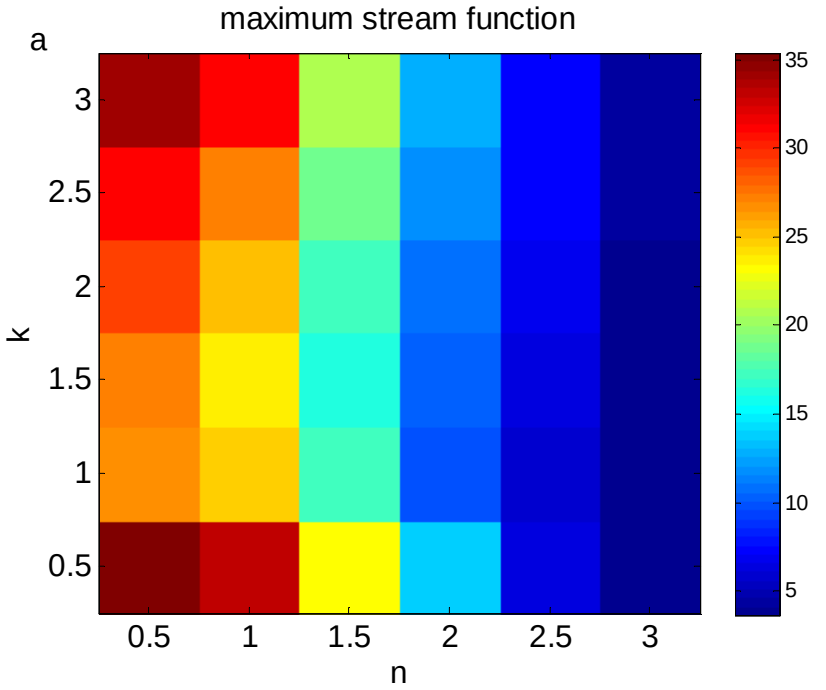
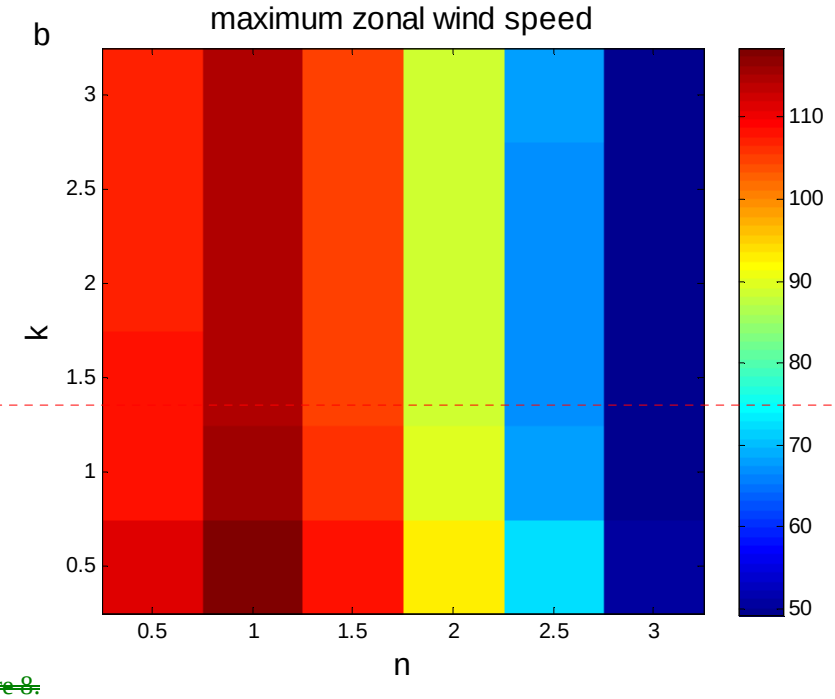


Figure 8.

Formattato: Tedesco (Germania)


1
2
3
4
5
6
7
8
9
10
11
5000
4500
4000
3500
3000
2500
2000





~~Figure 8.~~

Formattato: Tedesco (Germania)

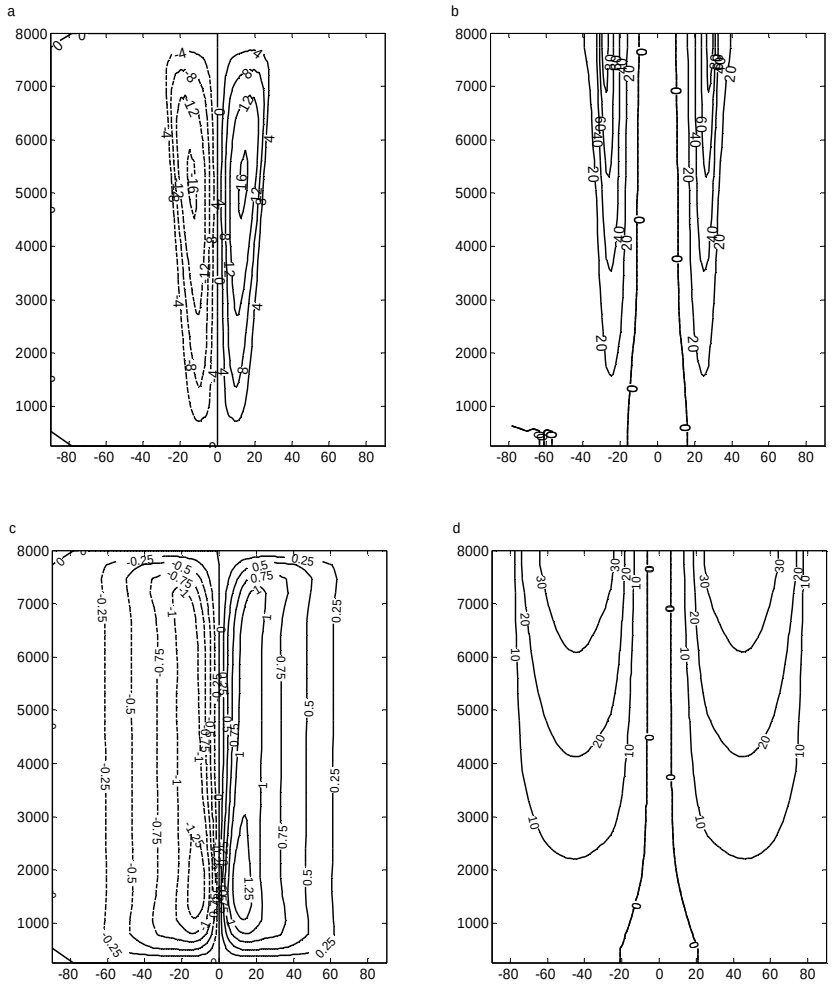
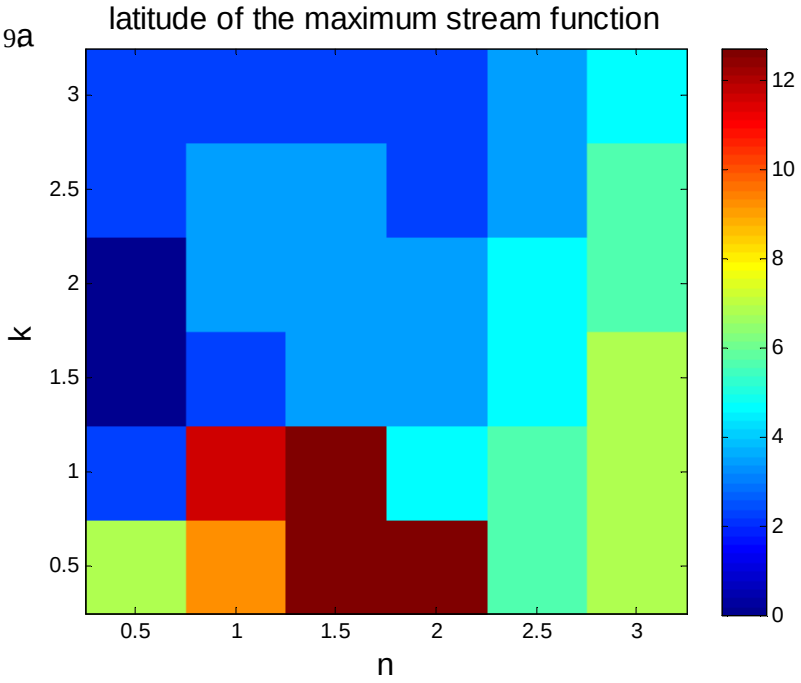
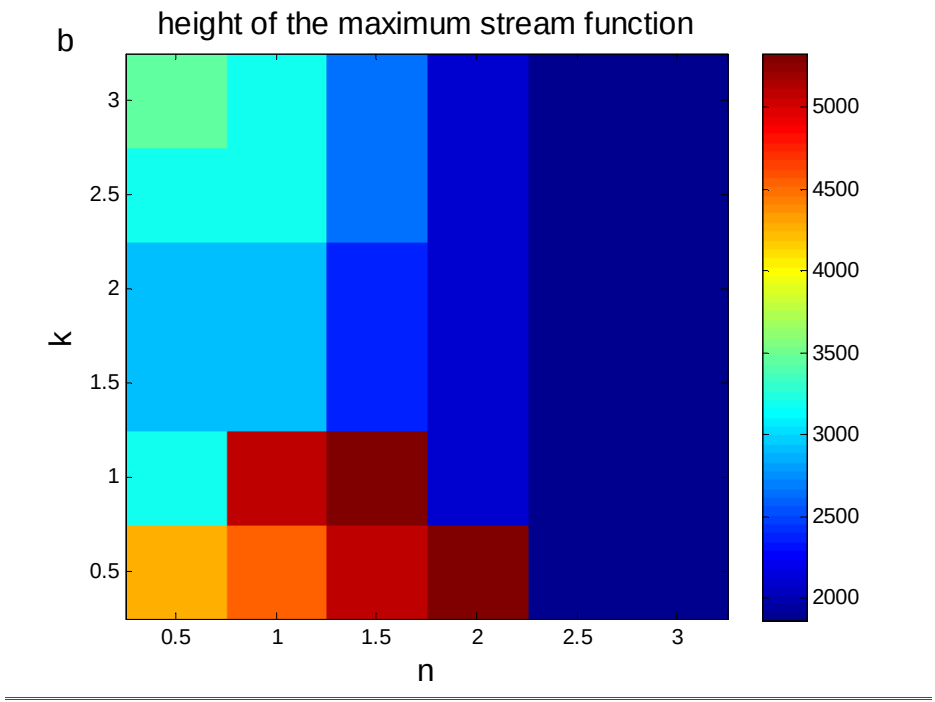


Figure 9a




b
height of the maximum stream function
k
n
3
2.5
2
1.5
1
0.5
0.5
1
1.5
2
2.5
3
5000
4500
4000
3500
3000
2500
2000

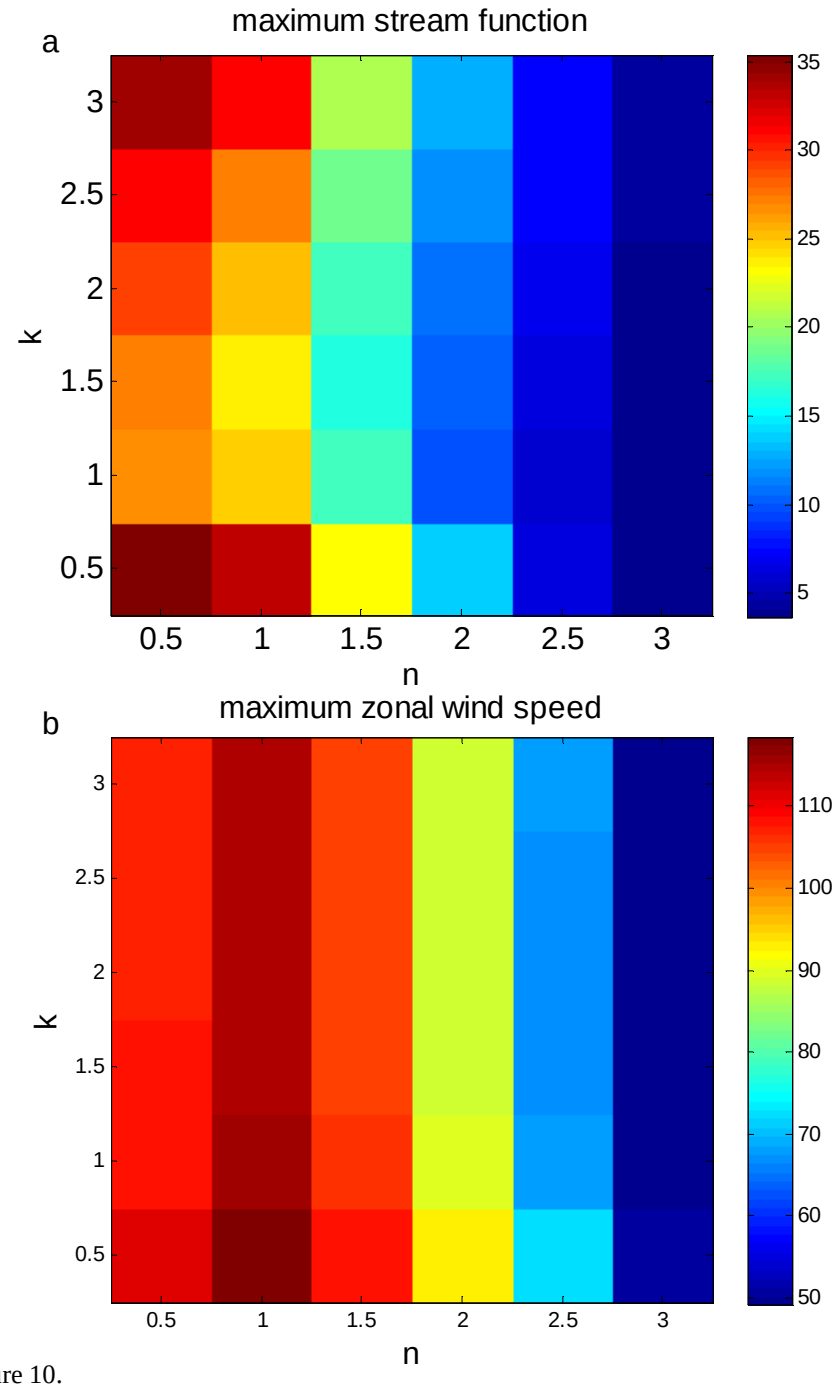
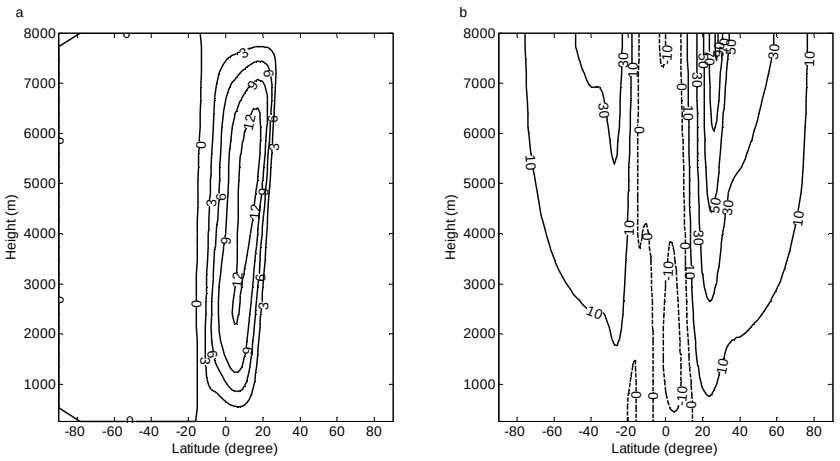
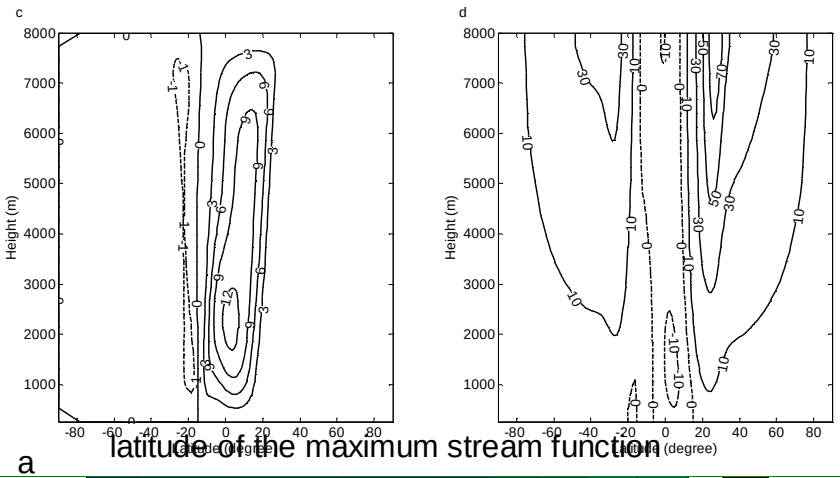


Figure 10.

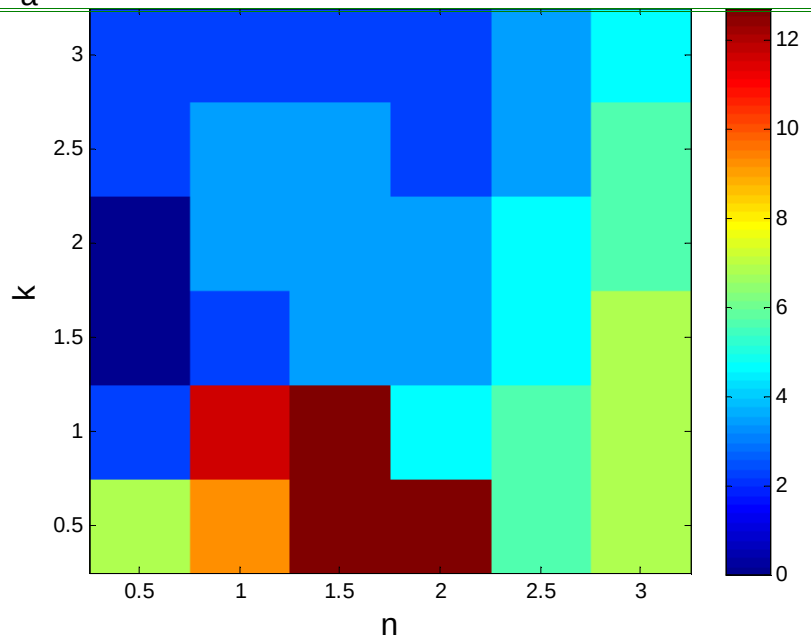
Formattato: Allineato a sinistra





latitude of the maximum stream function

a



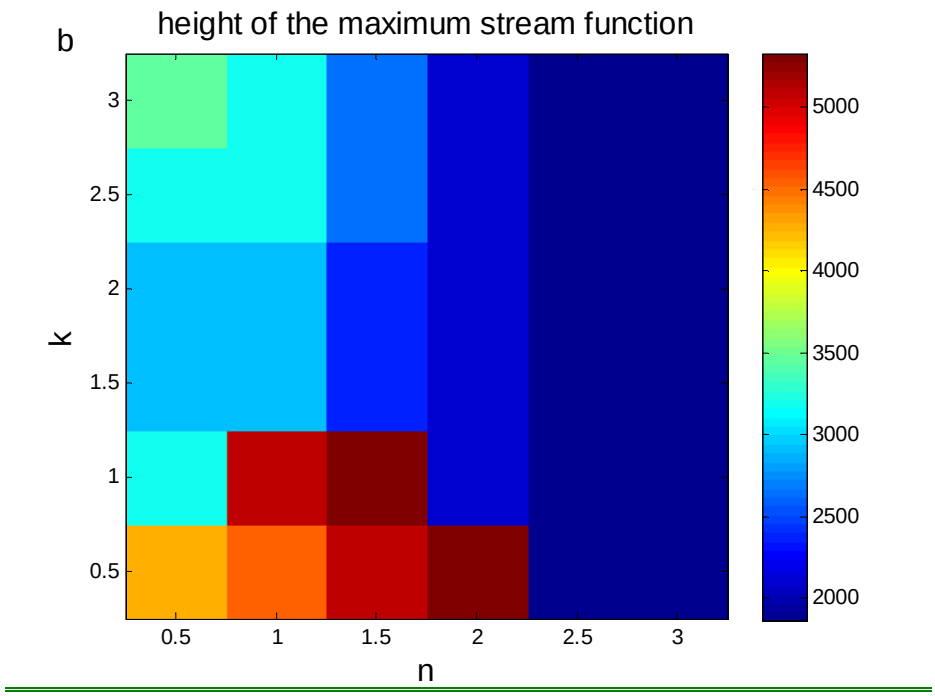
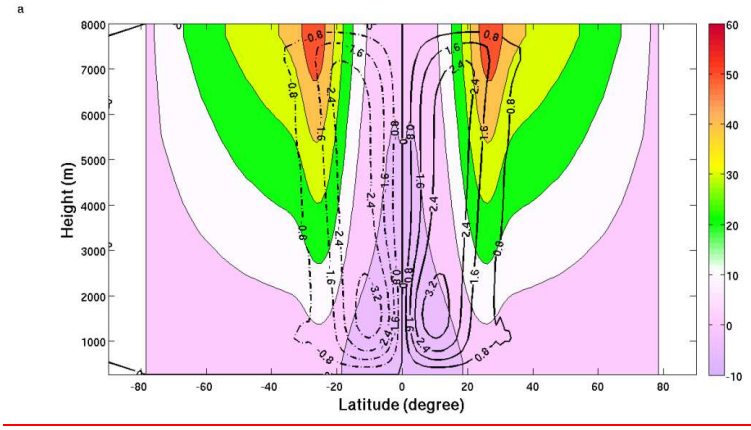
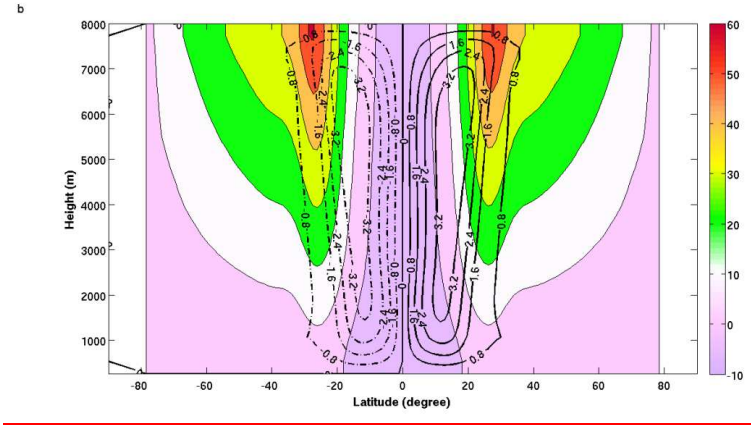

b
height of the maximum stream function
k
n
3
2.5
2
1.5
1
0.5
0.5
1
1.5
2
2.5
3
5000
4500
4000
3500
3000
2500
2000

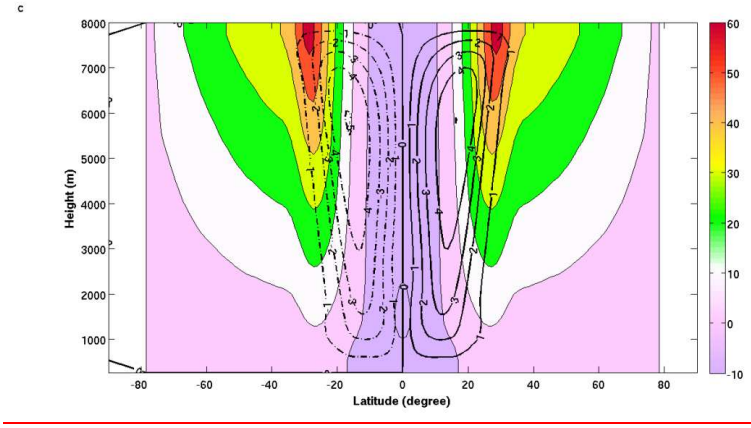
Figure 11.


a
Height (m)
Latitude (degree)

1


b
Height (m)
Latitude (degree)

2


c
Height (m)
Latitude (degree)

3

4

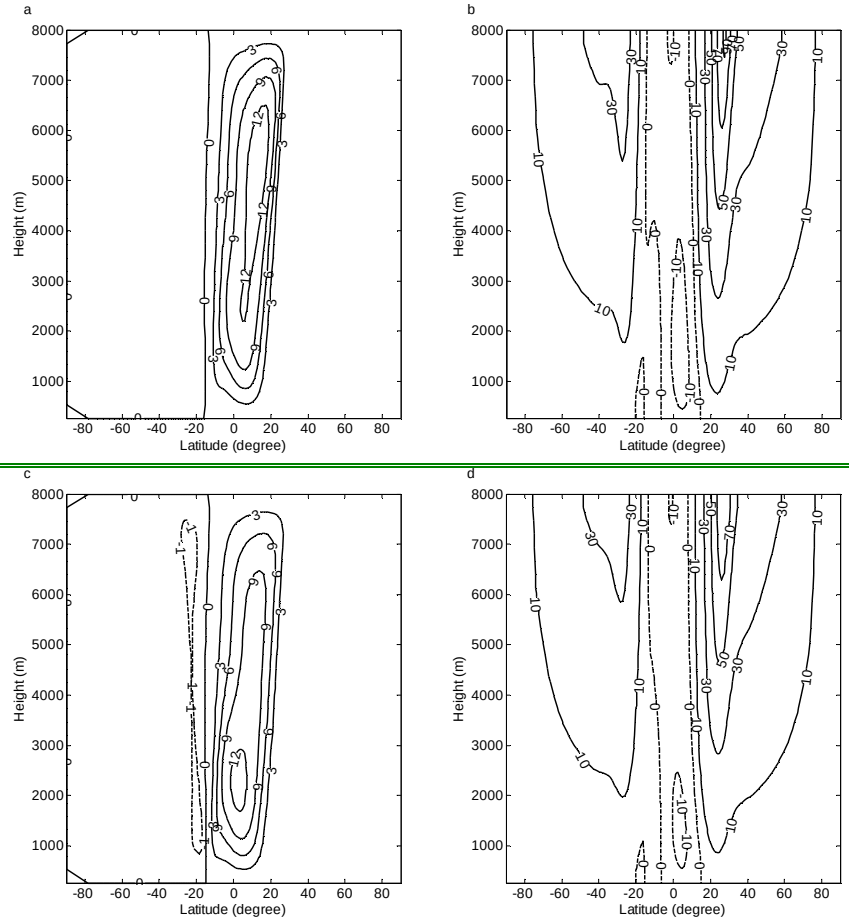
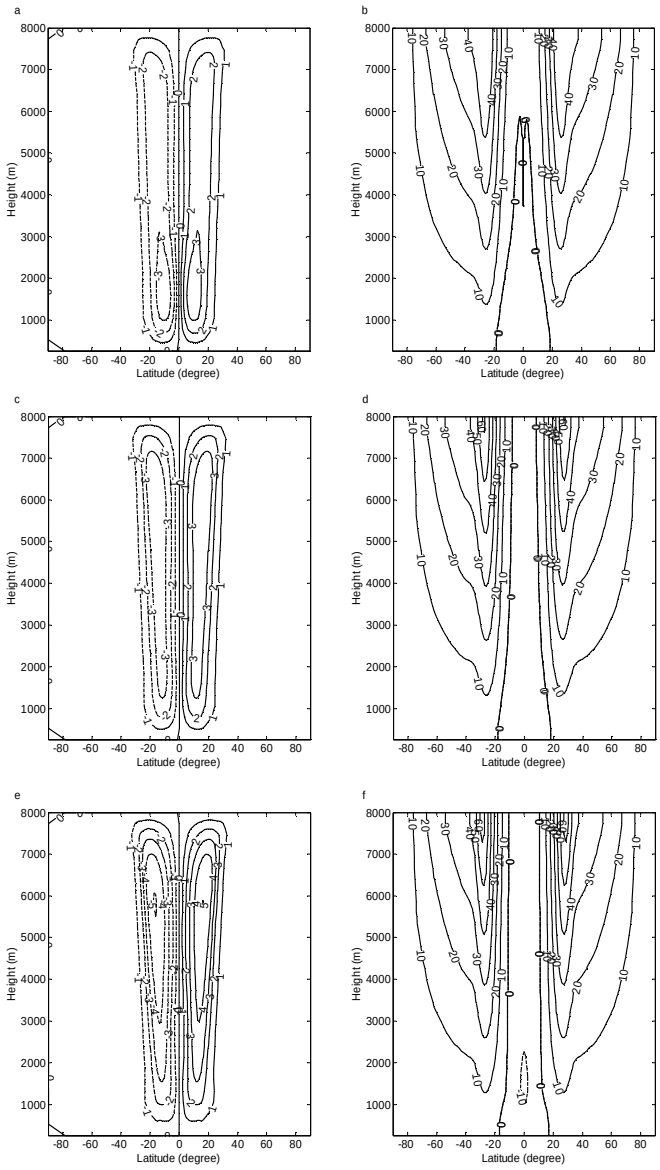

a
b
c
d
Height (m)
Latitude (degree)
8000
7000
6000
5000
4000
3000
2000
1000
-80 -60 -40 -20 0 20 40 60 80

Figure 12.



~~Figure 13.~~

Formattato: Default Style