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Large-scale modes of the tropical atmosphere. Part II: analytical modeling of Kelvin waves using the CAPE closure

Željka Fuchs[1] *and Antun Marki*[2]

[1] Physics Department and Geophysical Research Center, New Mexico Institute of Mining and Technology, Socorro, New Mexico, USA

[2] Department of Geophysics, Faculty of Science, University of Zagreb, Zagreb, Croatia

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The thermal assumption of the model is based on the convective available potential energy (CAPE) closure, i.e. increased CAPE, represented by decreased midlevel potential temperature, results in increased precipitation. The dynamic assumption of the model is that the vertical heating profile has the shape of the first baroclinic mode, while the vertical dependence of modeled fields is calculated, i.e. the model is vertically resolved. The modeled modes are free Kelvin waves and convectively coupled Kelvin waves. It is shown that the CAPE closure is not sufficient to produce the observed destabilization of the Kelvin mode, but that the dynamical properties of the model give the observed phase speeds.

Keywords: CAPE, Kelvin waves

1. Introduction

Fuchs and Marki (2006) in part I of the two-paper sequence gave a short overview of the differences between physics in the tropics and in the middle latitudes, as well as an outline of the large-scale disturbances in the tropics. Those are Madden-Julian oscillation (MJO), Kelvin waves, inertio-gravity waves, equatorial Rossby waves and mixed Rossby-gravity waves. Fuchs and Marki (2006) concentrated on the physics of the Kelvin wave; i.e. presented one way of modeling the convectively coupled Kelvin waves by using the boundary-layer quasi-equilibrium approximation. The model included the wind-induced surface heat exchange (WISHE). It was vertically resolved and it produced the free Kelvin waves and the convectively coupled Kelvin waves of the observed phase speeds (Straub and Kiladis, 2002). The convectively coupled Kelvin waves were unstable only for unrealistically long wavelengths.

The second part of the paper sequence continues with modeling physics of the convectively coupled Kelvin waves. The observations show the instability of the convectively coupled Kelvin waves and its propagating phase speed of 17 ms^{-1}. The goal of this paper is to:

1. Apply different physics assumptions (than in part I), i.e. the convective available potential energy (CAPE) closure to get a better understanding of what causes the instability of the convectively coupled Kelvin wave.

2. Investigate the impact of the dynamics on the modeled phase speeds of the convectively coupled Kelvin waves.

We first give a brief historical outline of large-scale wave modeling and then give the reason for choosing the CAPE closure.

The models that simulate the interactions between large-scale motions and deep convection in the tropics can be characterized as either convergence-driven models or quasi-equilibrium models. The convergence-driven models or wave-CISK (conditional instability of the second kind) models (Charney and Eliassen, 1964; Lindzen 1974) are models in which the convection is driven by low-level mass convergence. They assume large reservoirs of CAPE and tend to produce large wave growth rates on small spatial scales. The quasi-equilibrium models owe their origin to Manabe et al. (1965), followed by Arakawa and Schubert (1974) and tend to be stable unless some additional process occurs. Emanuel (1987) and Neelin et al. (1987) suggested that wind-induced surface heat exchange (WISHE) is the destabilization mechanism. In their linear model with WISHE, Neelin and Yu (1994) used the Betts and Miller (1986) quasi-equilibrium closure with an adjustment time of two hours. The model assumed that the vertically integrated convective heating is proportional to CAPE and inversely proportional to a relaxation time and thus assumed that the precipitation and CAPE are positively correlated. However, using the daily-mean sounding data averaged over the five KWAJEX (Kwajalein Experiment) locations, Sobel et al. (2004) show that the correlation between CAPE and precipitation is weak. By investigating the influence of CAPE on the convectively coupled Kelvin waves, we test the results of Sobel et al.

We now focus on the second goal of the paper by giving an overview of basic dynamical properties of the Kelvin wave and supporting the use of the vertically resolved model (Fuchs and Raymond, 2007)

The free adiabatic Kelvin wave phase speed is $c = (gh)^{1/2} = \Gamma_B^{1/2} / m_z$ where h is the equivalent depth, $m_z = 2\pi / L$ is the vertical wavenumber and $\Gamma_B^{1/2}$ is the Brunt-Väisälä frequency. The phase speed can be adjusted by varying either the vertical wavelength or the dry static stability. For the vertical wavelength of L = 30 km, which corresponds to the first baroclinic mode, the equivalent depth is 232 m and the phase speed is 48 ms^{-1}. The convectively coupled Kelvin waves have the observed phase speed of about 15–17 ms^{-1} which implies an equivalent depth of h = 23 m and a vertical wavelength of 9.4 km. Alternatively, we could adjust the equivalent depth through the static stability. If the Brunt-Väisälä frequency, which represents the dry static stability, is replaced by an effective static stability (Neelin et al., 1987; Neelin and Held, 1987; Emanuel et al., 1994; Neelin and Yu, 1994), it is possible to reduce the phase speed of convectively coupled modes without reducing the vertical scale. It then falls upon the various modelling schemes to define the physics of the effective static stability parameter. For instance, Emanuel et al. (1994) note that in strict quasi-equilibrium (SQE) the vertical structure of the vertical velocity of a simple linear mode is completely determined by the moist adiabatic lapse rate. The vertical profile of the vertical velocity in this case is close to that of the fundamental baroclinic mode. The overall conclusion is that large scale ascent in convective atmosphere is associated with a slight reduction of temperature, i.e. there exists positive effective static stability. This occurs because the large scale upward motion increases convection and therefore reduces the boundary layer entropy through the downdrafts. But the free atmosphere cools, since by their assumption, the atmosphere maintains a moist adiabatic lapse rate keyed to the surface conditions. The resulting effective static stability is typically reduced by an order of magnitude compared to dry static stability.

The recent observations of convectively coupled equatorially trapped waves (Wheeler and Kiladis, 1999; Wheeler et al., 2000; Straub and Kiladis, 2002) tell us that the vertical structure for the propagating modes is not that of the first baroclinic mode. In particular Straub and Kiladis (2002) analyzed a convectively coupled Kelvin wave propagating in the Eastern Pacific ITCZ (intertropical convergence zone) and found that the temperature profile associated with the Kelvin wave is anomalously cold in the lower part of the troposphere and warm in the upper part, within (and sometimes leading) the region of strong convective heating. The vertical heating profile, however, has a vertical wavelength twice the depth of the troposphere. This result is fundamentally in conflict with the quasi-equilibrium picture outlined above.

The new approach of modeling the dynamics of convectively coupled equatorially trapped waves started with Mapes (2000) and was followed by Majda and Shefter (2001) and Majda et al. (2004). In his model Mapes assumes two separate vertical modes: one that represents the stratiform rain regions and the one which corresponds to the convective rain regions. The mode corresponding to stratiform rain has a significantly smaller vertical wavelength than the one corresponding to the convective region. Fuchs and Raymond (2007) presented a simple analytical vertically resolved model in a non-rotating atmosphere that offers an alternative approach that does not assume two vertical modes a priori, nor does it specify the values of equivalent depth and effective static stability. The dynamics of their model is based on only one assumption and that is that the heating has the specified vertical structure of the first baroclinic mode as shown by Straub and Kiladis (2002). The vertical profiles of the vertical velocity perturbation as well as of the other thermodynamic fields are then calculated.

By implementing the CAPE closure on a vertically resolved model (Fuchs and Raymond, 2007), we will be able to see whether the modeled phase speeds of the convectively coupled Kelvin waves depend on various closures or whether they are a sole consequence of the dynamical properties of the system. Thus we will achieve the second goal of this paper.

Combining the results of this model with those from part I will give us a better insight into the physics of convectively coupled Kelvin waves.

2. Model

In part I we applied the boundary-layer quasi-equilibrium approximation and did not need to give the full set of equations. Now we present the linearized governing equations for large scale motions in the nonrotating atmosphere. Then we explain the thermodynamics of the model.

2.1. Governing equations

Our governing equations are based on Boussinesq theory and are given for the meridional velocity, $v = 0$, case. We write our governing equations for moist atmosphere in terms of dry and moist entropy instead of the more conventional variables potential temperature θ and equivalent potential temperature θ_e. As perhaps this transformation is not trivial, we will briefly go through it starting with the hydrostatic equation:

$$\theta \nabla \tilde{\Pi} - g\hat{k} = 0 \tag{1}$$

where the Exner function is given by $\tilde{\Pi} = C_p(p / p_R)^{\kappa}$, where C_p is the specific heat at constant pressure p, p_R is a reference pressure, $\kappa \equiv R / C_p$ where R is the gas constant, g is the acceleration of gravity, and $\hat{k}$ is the vertical unit vector. The potential temperature is defined as $\theta = T_R \exp(s_d / C_p)$ where T_R is a reference temperature and s_d is the dry entropy. After applying the perturbation method and linearization we obtain: $\theta' / \theta_0 = s'_d / C_p$ where the subscript zero implies the base state and prime indicates the perturbation. Noting the use of Boussinesq approximation and the definition of Exner function: $\theta_0 \nabla \tilde{\Pi}' \sim \nabla \theta_0 \tilde{\Pi}' \equiv \nabla \Pi'$ and applying the perturbation method and linearization on (1): $\theta_0 \nabla \tilde{\Pi}' = -g\hat{k}\theta' / \theta_0$, it is straightforward to obtain the linearized hydrostatic equation:

$$\frac{\partial \Pi'}{\partial z} - \frac{g}{C_p} s'_d = 0 \tag{2}$$

Equation (2) corresponds to the hydrostatic equation $\partial\Pi'/\partial z = b'$, where $b' = g\theta' / \theta_0$ used by Fuchs and Raymond (2002), so we define the buoyancy perturbation as $b' \equiv g s'_d / C_p$.

The dry entropy equation is:

$$\frac{ds_d}{dt}=\frac{Q}{T} \tag{3}$$

where Q is the heating source and T is the temperature that becomes T_R in linearization. After perturbation, linearization and use of the buoyancy perturbation definition, we get:

$$\frac{\partial b'}{\partial t}+\frac{g}{C_p}\frac{ds_{d0}}{dz}w'=\frac{g}{C_pT_R}Q' \tag{4}$$

which is the scaled dry entropy equation. The moist entropy is defined approximately as $s = s_d + Lr / T_R$, where L is the latent heat of condensation and r is the mixing ratio. We now define scaled moist entropy and scaled mixing ratio perturbations as $e' \equiv gs' / C_p$, $q' \equiv gLr' / (C_pT_R)$, maintaining the equality $e' = b' + q'$ that easily comes out of the moist entropy definition after applying the perturbation method.

The mixing ratio equation is:

$$\frac{dr_t}{dt} = -P_r \tag{5}$$

where $r_t = r + r_c$, r is the water vapor mixing ratio and r_c is the cloud droplet mixing ratio. P_r is the moisture source term. After the perturbation method and linearization we get the equation for scaled mixing ratio:

$$\frac{\partial q'}{\partial t}+\frac{gL}{C_pT_R}\frac{dr_{t0}}{dz}w'=-\frac{gLP_r'}{T_RC_p} \tag{6}$$

The moist entropy equation is:

$$\frac{ds}{dt}=\frac{Q_R}{T_R}-\frac{\partial F_e}{\partial z} \tag{7}$$

where Q_R is the radiation source and F_e is a perturbation of all small scale eddy fluxes. Again, after the perturbation method and linearization we get the equation for scaled moist entropy:

$$\frac{\partial e'}{\partial t}+\frac{g}{C_p}\frac{ds_0}{dz}w'=\frac{gQ_R'}{T_RC_p}-\frac{g}{C_p}\frac{\partial F_e'}{\partial z} \tag{8}$$

To summarize, our linearized system of governing equations for $v = 0$ with the Boussinesq approximation is:

$$\frac{\partial u}{\partial t}+\frac{\partial \Pi}{\partial x}=0 \tag{9}$$

$$\frac{\partial \Pi}{\partial z}-b=0 \tag{10}$$

$$\frac{\partial u}{\partial x}+\frac{\partial w}{\partial z}=0 \tag{11}$$

$$\frac{\partial b}{\partial t}+\Gamma_B w=S_B \tag{12}$$

$$\frac{\partial q}{\partial t}+\Gamma_Q w=S_Q \tag{13}$$

$$\frac{\partial e}{\partial t}+\Gamma_E w=S_E \tag{14}$$

Note that all the variables are perturbations. The momentum (9) and continuity (11) equations are straightforward and are not discussed in detail. The system of equations (9) – (14) was used in the model of Fuchs and Raymond (2007), but that model implemented moisture closure and therefore had different thermal assumptions than in the presented model.

2.2. Thermal assumptions of the model

We assume the simplest scenario that implements only the CAPE closure so that we can observe the impact of only that factor. We ignore other physical mechanisms such as cloud-radiation interactions and WISHE. Thus we only need to calculate the vertically integrated dry entropy source term S_B that depends on the precipitation rate:

$$B = \int_0^h S_B(z)dz = P \tag{15}$$

where h is the depth of the troposphere. P is the scaled perturbation in the precipitation rate that depends on CAPE:

$$P = -\eta \int_0^h b dz \tag{16}$$

where η^{-1} is the buoyancy relaxation time. The equation (16) tells us that increased CAPE, represented by decreased midlevel potential temperature, results in increased precipitation; this is called the CAPE closure. It is important to note that no assumptions have been made yet about the vertical structure of the fields. We now use the calculated vertical velocity profile from Fuchs and Raymond (2007):

$$w(z) = \frac{Bm_0}{2\Gamma_B(1-\Phi^2)}\left[\sin(m_0 z) + \Phi\exp(-i\frac{\pi}{\Phi})\sin(mz)\right] \tag{17}$$

where $m_0 = \pi / h$, h is the depth of the troposphere, $m = k\Gamma_B{}^{1/2} / \omega$, k is the horizontal wavenumber, ω is the frequency and $\Phi = m_0 / m$. Φ is also defined as Ω / κ. The equation for vertical velocity perturbation consists of two sinusoidal components of different wavelengths. The term $\sin(m_0 z)$ in (17) corresponds to the deep convective component and the term $\exp(-i\pi/\Phi)\sin(mz)$ to the shallow component of the mode.

3. Calculating the dispersion relation

We write the polarization relations for our system of equations (9) – (14):

$$u(z) = \frac{i}{k}\frac{\partial w(z)}{\partial z} \tag{18}$$

$$\Pi(z) = \frac{i\omega}{k^2}\frac{\partial w(z)}{\partial z} \tag{19}$$

$$b(z) = \frac{i}{\omega}[S_B(z) - \Gamma_B w(z)] \tag{20}$$

$$q(z) = \frac{i}{\omega}[S_Q(z) - \Gamma_Q(z)w(z)] \tag{21}$$

$$e(z) = \frac{i}{\omega}[S_E(z) - \Gamma_E(z)w(z)] \tag{22}$$

The next step is to calculate the dispersion relation. The vertical heating profile of the first baroclinic mode structure $S_B(z) = 0.5Bm_0\sin(m_0 z)$ is assumed, but the value of the vertically integrated heating, $B = \int_0^h S_B(z)dz$, is unknown. However, we know the vertical velocity profile. The approach that is taken is the following: we start with the expression for the vertically integrated heating, B:

$$B = \int_0^h S_B(z)dz = P = -\eta\int_0^h b(z)dz \tag{23}$$

It is straightforward to calculate the vertical integral of the scaled entropy perturbation, b, from the polarization relation (20):

$$\int_0^h b(z)dz = \frac{i}{\alpha\kappa\Phi}\left[\int_0^h S_B(z)dz - \Gamma_B\int_0^h w(z)dz\right] \tag{24}$$

We also know that the vertical velocity perturbation is given by (17). Its integral is then:

$$\int_0^h w(z)dz = \frac{B}{\Gamma_B(1-\Phi^2)}\left\{1+\frac{\Phi^2}{2}\exp\left(-i\frac{\pi}{\Phi}\right)\left[1-\cos\left(\frac{\pi}{\Phi}\right)\right]\right\} \qquad (25)$$

The nondimensionalized horizontal wavenumber is $\kappa = k\Gamma_B^{1/2} / \alpha m_0$, the nondimensionalized frequency is $\Omega = \omega / \alpha$ and $\Phi = \Omega / \kappa$ is the nondimensionalized phase speed. α is here just a normalising constant chosen as an inverse of a day.

Combining (23) – (25), the dispersion relation with CAPE closure is:

$$\kappa\Phi^2 + i\gamma\Phi - \kappa + 0.5i\gamma\Phi\exp\left(-i\frac{\pi}{\Phi}\right)\left[1-\cos\left(\frac{\pi}{\Phi}\right)\right] = 0 \qquad (26)$$

where $\gamma = \eta / \alpha$ is a nondimensional CAPE parameter that we will vary.

4. Results

All the modes that come out of the dispersion relation (26) are now plotted in figure 1 for the case when the buoyancy relaxation time, η^{-1}, is twelve hours and thus $\gamma = 2$. The phase speeds and the growth rates of the modes are plotted as a function of the planetary wavenumber l, where $l = 1$ corresponds to a wavelength equal to the circumference of the earth. The upper panel shows the dimensional phase speed of the wave, while the bottom panel shows the growth rate in units of 1 / day. The growth rate is represented by the imaginary part of the frequency because all the variables have $\exp(-iwt)$ dependence in time, so that the positive imaginary part of the frequency gives the exponential growth and negative part, the decay.

The modeled modes are the convectively coupled Kelvin mode and the fast Kelvin modes. The convectively coupled Kelvin mode has the phase speed of 15 – 17 ms^{-1}. The mode is damped as a consequence of diabatic effect, in this case because of the CAPE. The modeled fast Kelvin modes present no surprise: their phase speed is 48 ms^{-1} and they decay at a very slow rate as expected.

Figure 2 shows the convectively coupled Kelvin mode for different choices of CAPE parameter. Even though there are some differences when the CAPE parameter is varied, the convectively coupled Kelvin mode remains stable.

The heating and temperature distributions in horizontal and vertical dimensions for the convectively coupled Kelvin mode are given in figure 3. The strongest heating is where the shading is the lightest and the temperature distribution is given by the contour plots. The temperature perturbations are strong and the positive temperature perturbations are found in the area of the strongest heating, as expected. However, there is no "boomerang shape" in

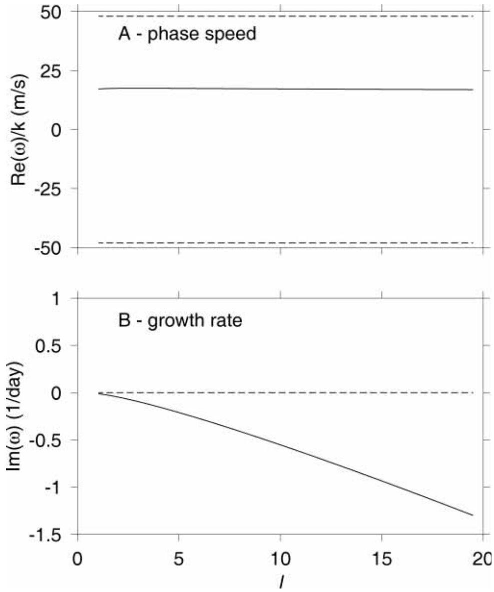

Figure 1. Dimensional dispersion curves as a function of the planetary wavenumber l when η^{-1} = 12 hours. Solid line represents convectively coupled Kelvin mode and dashed lines represent fast Kelvin modes. The upper panel shows the phase speeds while the bottom panel shows the growth rates of the waves.

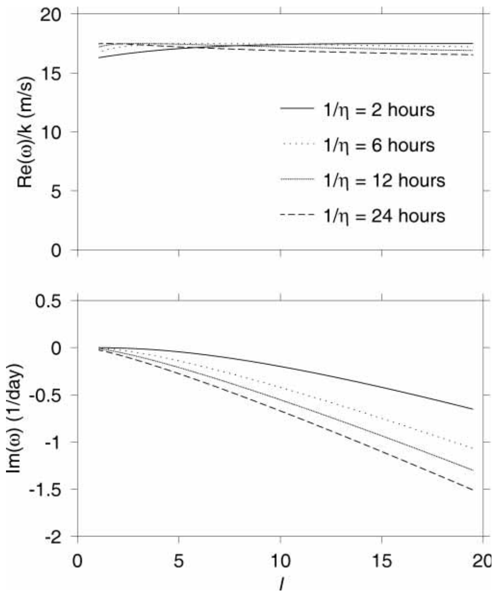

Figure 2. Dimensional dispersion curves for convectively coupled Kelvin mode as a function of the planetary wavenumber l when η^{-1} is varied.

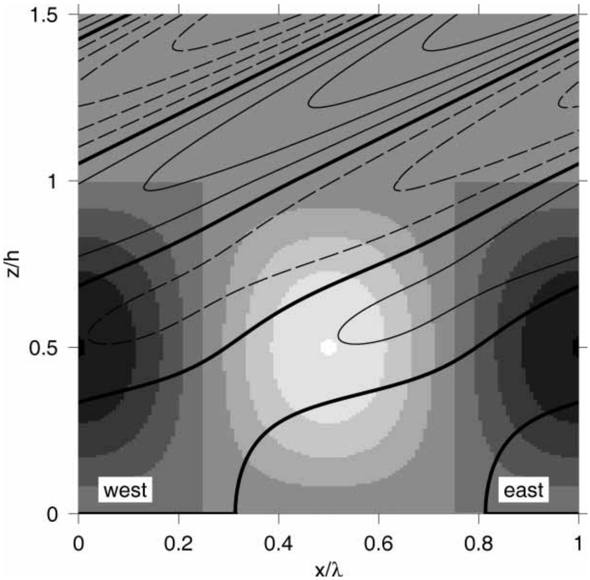

Figure 3. Heating and temperature for the convectively coupled Kelvin mode when planetary wavenumber l = 10 in horizontal and vertical dimensions: x / λ is the phase of the wave and z / h is the scaled height. The heating is maximum where the image is white and minimum where it is black. Temperature is given as a contour plot where the thick solid line corresponds to the zero contour; negative perturbation contours are given as dashed lines and positive as solid. The interval between the contours is arbitrary.

temperature contours suggested by Straub and Kiladis (2002), because the mode is stable.

The results in figure 1 are similar to those in part I (Fuchs and Marki, 2006). The phase speed of the convectively coupled Kelvin waves remains the same and the CAPE closure has little effect on the stability.

5. Conclusions

The convectively coupled Kelvin mode in nonrotating atmosphere maps onto the convectively coupled Kelvin wave in a rotating atmosphere and so we can compare it with the observations, Straub and Kiladis (2002). The modeled convectively coupled Kelvin waves are very robust in the sense that their phase speed is not significantly influenced by physical parameters of the model, i.e. the CAPE closure. Furthermore Fuchs and Marki (2006) showed the same phase speed results for convectively coupled Kelvin waves while using the boundary-layer quasi-equilibrium approximation.

Comparing our model to others is quite a challenge as it is based on completely different assumptions. The quasi-equilibrium models in general assume one vertical wavelength, usually the one of the first baroclinic mode. Emanuel (1987) modeled an unstable WISHE mode that has a phase speed of 30 m/s for long wavelengths and 15 m/s for short wavelengths. At that time those modes were called Kelvin modes or were perceived as being similar to Madden-Julian oscillation (MJO) modes. However, after Straub and Kiladis (2002) it has become apparent that the Kelvin wave has a shorter vertical wavelength than that of the first baroclinic mode, and thus quasi-equilibrium models cannot give the observed convectively coupled Kelvin wave.

In summary, the presented model showed two important things:

1. The damping of the convectively coupled Kelvin waves indicates that the full physics of the interaction between Kelvin waves and convection is not captured by the simple thermodynamic assumptions postulated here. It confirms the results of Sobel et al. (2004) that the correlation between CAPE and precipitation is weak and as such is not responsible for the instability of the Kelvin wave.

2. By imposing a single dynamic assumption of fixed vertical heating profile, the model captures the convectively coupled Kelvin waves with the observed phase speeds. Furthermore, it shows that the phase speed of the convectively coupled Kelvin waves is a consequence of including the right dynamics in the model.

In combining the results of this paper and those of Fuchs and Marki (2006) it is evident that neither the CAPE closure nor the boundary-layer quasi-equilibrium assumption with WISHE presented in part I are causing the observed instability in convectively coupled Kelvin waves. How the instability originates remains a challenge. The dynamic assumption of the models, however, is responsible for modeling the observed phase speed of the convectively coupled Kelvin waves, while robust to CAPE closure and the boundary-layer quasi-equilibrium assumption with WISHE, and as such can be used on a broader range of models.

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SAŽETAK

Dugoperiodički modovi u tropskoj atmosferi Drugi dio: Analitičko modeliranje Kelvinovih valova povezanih s konvekcijom uz konvekcijsku raspoloživu potencijalnu energiju

Željka Fuchs i Antun Marki

Termalna pretpostavka modela je da konvekcijski raspoloživa potencijalna energija utječe na količinu oborine. Konvekcijski raspoloživa potencijalna energija ovisi o potencijalnoj temperaturi u srednjoj troposferi. Uzrok povećane količine oborine je smanjena potencijalna temperatura u srednjoj troposferi. Dinamika modela je razvijena uz pretpostavku da vertikalni profil grijanja ima oblik prvog baroklinog moda dok se u modelu računa ovisnost termodinamičkih polja po vertikali. Modelirani valovi su slobodni Kelvinovi valovi i Kelvinovi valovi povezani s konvekcijom. Iz modela se vidi da konvekcijski raspoloživa potencijalna energija nije dovoljna za opaženu nestabilnost Kelvinovih valova povezanih s konvekcijom, ali dinamika modela daje opaženu faznu brzinu.

Ključne riječi: CAPE, Kelvinovi valovi

Corresponding author's address: Željka Fuchs, Physics Department and Geophysical Research Center, New Mexico Institute of Mining and Technology, Socorro, New Mexico, USA, tel: + 1 505 83 55 407, fax: + 1 505 83 55 707, e-mail: zeljka@kestrel.nmt.edu