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PLANETARY BOUNDARY LAYER FEEDBACK AND ITS ROLE IN THE CLIMATE CHANGE

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1. INTRODUCTION

An energy balance model of the Earth's Climate System (ECS) could be written as (Schwartz, 2007)

$$C\, d\theta_a / dt = S - L \quad (1)$$

where C is the ECS heat capacity (per unit area); θ_a is the ECS temperature, which could be associated with the surface (potential) air temperature (a popular characteristic of climate); and $S-L$ is the balance of incoming and outgoing radiation at the top of the atmosphere. In majority of climate publications, C is taken as given property of the ECS. It is assumed that the system heat capacity is probably known with uncertainty (Hansen et al., 1985) but neither changing significantly with climate nor imposing significant feedbacks on the ECS. Hence the main attention of climatologists is focused on processes and feedbacks directly or indirectly affecting the radiation balance $S-L$.

The formulation (1) could be further specified for an atmospheric column as

$$d\theta_a / dt = Q / C \quad (2)$$

where Q is the heat flux divergence in the column, which includes also the divergence of the horizontal heat fluxes, the heat storage in the soil/ocean and the balance of the latent heat fluxes due to condensation and evaporation. If the vertical turbulent mixing in the planetary boundary layer (PBL) of thickness, h, is faster than the other dynamical or radiation processes, then the vertical temperature gradient, $\partial\theta_a / \partial z$, is height independent inside the PBL and smaller than the gradient in the free atmosphere. In well mixed convective PBL, this quantity is $\partial\theta_a / \partial z$ = 0. Then the temperature evolution will be determined by a trivial equation

$$d\theta_a / dt = Q / C = Q / \left(\rho c_p h\right) \quad (3)$$

where ρ, c_p are air density and the air specific heat at constant pressure.

Integration of Eq. (3) over climatological time scales would give the mean surface air temperature in the ECS. There is however an important difficulty, the global and even more so the regional heat capacity of the ECS is determined by the PBL thickness h. This quantity varies by several orders of magnitude on daily and seasonal time scales depending on the atmospheric stability, measured, in the case of small baroclinicity, with the Brunt-Vaisala frequency $N = \sqrt{g\beta\partial\theta_a / \partial z}$, where g =9.81 [m s^{-2}] is the acceleration due to gravity and β =0.003 [K^{-1}] is the thermal expansion coefficient.

The PBL thickness $h(t)$ is an integral measure of the vertical turbulent mixing in the PBL. This quantity and its evolution depend on a number of control parameters, including Q as well as characteristics of the atmospheric circulation such as the large-scale wind speed U. Early studies did not include $h(t)$ or the turbulent mixing directly, but apply a convective adjustment procedure. The procedure adjusted the vertical temperature gradient in a model to the observed one, i.e. $\partial\theta_a / \partial z > \partial\theta_{obs} / \partial z$, and in this sense implicitly accounted for the turbulent mixing. Manabe and Strikler (1965) were probably the first who recognized the role of the turbulent mixing in the ECS. In their continuously heated radiation-convection model, this role was strongly negative and lowered the surface temperature by 44^{o}K relative to the ECS radiation equilibrium.

This and following studies with the radiation-convective models (e.g. Cunnington and Mitchell, 1990; Moraes et al., 2005) were not very realistic as they studied the conditions of continues heating in the ECS. Contrary, as Eq. (3) suggests, the largest response on variations of Q, which includes changes in the radiation balance due to the accumulation of the greenhouse gases, should be found in the thinnest PBL. The latter ones are typical for the conditions of the net radiation cooling, i.e. in nocturnal and long-lived wintertime high-latitude PBL. Up to my knowledge, Manabe and Wetherald (1975) were first who has attributed the amplified θ_a response in high latitudes on doubling CO2 concentration in simulations with GFDL climate model to the restrictions on the vertical turbulent mixing in a more statically stable polar atmosphere. This effect is quoted as the polar amplification of the global warming. Contrary to the common believe, complex, physically sophisticated climate models are not required to obtain the polar amplification. It can be reproduced in the radiation-convection models (e.g. Moraes et al., 2005) where climate feedbacks usually invoked to explain the polar amplification are absent.

Despite the clear importance of the turbulent mixing for the ECS, there was almost no progress in its understanding during recent decades. An obvious reason is that the theoretical understanding of the stably stratified PBL was inadequate. Only recently, the progress in high-performance computing and remote sensing of the turbulence made it possible to derive asymptotic dependences for h (Zilitinkevich and Esau, 2003; Fedorovich et al., 2004; Ziliinkevich et al., 2007) and relate those with θ_a and external (relative to the PBL) control parameters (Zilitinkevich and Esau, 2005; Esau and Zilitinkevich, 2006; Zilitinkevich and Esau, 2007). The key role of $h(t)$ for the ECS and its sensitivity became obvious (Esau, 2008), although the quantification of this role and a proper correction of the climate models require additional efforts.

So far, we discussed the effect of $h(t)$ on θ_a assuming Q to be an external parameter. In fact, Q affects θ_a in both ways directly as well as indirectly through its impact on h. Thus, a dynamical feedback between h and θ_a is possible. GABLS results (Cuxart et al., 2006; Beare et al., 2006) suggest that h is simulated with a large error in the models. Moreover, it is usually too high in the stably stratified PBL and often in convective PBL too. It may explain the damping of the observed warming signal in the models. The problem is seen for instance as the reduced daily temperature range (Easterling et al., 1997; Vose et al., 2005) and as reduced sensitivity of simulated climate (Rahmstorf et al., 2007; Stroeve et al., 2007; van Oldenborgh et al., 2008).

This type of feedbacks is the subject of the study. In Section 2, the paper repeats the formulation of the PBL-feedback given in Esau (2008). In Section 3, estimations of this feedback efficacy through turbulence-resolving simulations are given. In Section 4, some evidences from the ERA-40 reanalysis are given to support the PBL-feedback hypothesis with the atmospheric data. Summary is given in Section 5. Appendix describes the procedure to obtain the PBL-feedback in the turbulence-resolving simulations.

2. PBL-FEEDBACK FORMULATION

Planetary boundary layer feedback (PBL-feedback) is a response of the surface air temperature, θ_a (for convenience, the potential temperature will be used here), on changes in the vertical turbulent mixing. A system transient sensitivity is defined as the time integrated temperature change $\delta\theta_a$ in response on the perturbation of the forcing δQ. It could be written down using Eq. (3) as

$$\frac{\delta\theta_a}{\delta Q} = \frac{1}{\rho c_p}\int\left(h^{-1} - Qh^{-2}\frac{\delta h}{\delta Q}\right)dt \qquad (4)$$

where integration should be done over a climatologically significant period of time. Here, we will simplify the problem neglecting the role of the large-scale circulation and cloudiness in the PBL development. In such conditions, the mean surface air temperature θ_a and its change $\delta\theta_a$ are defined by the asymmetry of the PBL diurnal cycle.

In the climate science, the inverse quantity

$$\lambda = C/\tau = \left(\delta\theta_a / \delta Q\right)^{-1}, \qquad (5)$$

that is so called feedback parameter (Hansen et al. 1984; Bony et al., 2006), characterizes the climate sensitivity. Here, τ is the sensitivity time scale. One can define a feedback gain as amplification of the black body (Plank) response on the temperature change. The gain reads $g = \lambda/\lambda_0$ where λ_0 is the Plank feedback. Different estimations give λ_0 from 2.1 W m^{-2} K^{-1} for the mid-latitude winter to 3.4 W m^{-2} K^{-1} for tropics (Huang and Ramaswamy, 2007) and up to 3.8 W m^{-2} K^{-1} in the Stefan-Boltzmann law by equating the outgoing long wave radiation to the fourth power of temperature and assuming the earth's emission temperature of 255 K. Here, λ_0 = 3.2 W m^{-2} K^{-1} (Bony et al., 2006) will be used to allow for the comparison. The feedback gain is convenient for the comparison of the feedback strengths. The traditionally quoted g in the IPCC models (Bony et al., 2006) are 0.563 for the water vapour feedback; -0.263 for the lapse rate feedback; 0.216 for the cloud feedback; and 0.081 for the surface ice-albedo feedback. Thus, the ice-albedo feedback appears to be the strongest feedback in the Earth's climate system. It is necessarily to note that the IPCC values have somewhat different meaning from the values of the PBL-feedback in this paper. The IPCC values have been obtained from models where different feedback mechanisms interact and therefore g could differ from their values for the non-interacting feedbacks without interactions.

Although Eq. (4) looks simple its consequences are non-trivial. The major problem is the evolution of $h(t)$ itself (Medeiros et al., 2005). The PBL thickness is a nonlinear function of several parameters, which will be considered later. It is well defined quantity only in turbulence-resolving models and probably in the ocean mixed layer (Lorbacher et al., 2006). Its diagnosis in observations (Siebert et al., 2000) and climate models (Cuxart et al., 2006) meets considerable difficulties. Moreover, λ may change sign, at least locally, as Eq. (2) admits. It occurs when $h = Q\partial h/\partial Q$.

Now consider two distinct case for which the analytical asymptotic equations for h were given in Zilitinkevich (1991), Fedorovich et al. (1994), Zilitinkevich et al. (2007). These cases have been considered in more details in Esau (2008). In the convective PBL,

$$h = \left(c_{CBL}(\rho c_p)^{-1}Qt\right)^{1/2}N^{-1} \quad (6)$$

where c_{CBL} =1.67 (Fedorovich et al., 2004). Assuming sinusoidal heating with period 2 $\tau_{heating}$ and $h(t=0) >> 0$ (to avoid singularity of integrations) and using Eq. (6), PBL-feedback estimation becomes

$$\frac{\delta\theta_a}{\delta Q} = \frac{1}{\rho c_p}\frac{\tau_{heating}}{\max h}, \quad (7)$$

Where $\max h$ is the PBL thickness at time of the maximum heating. Figure 1 shows the analytical PBL-feedback in the convective PBL. As one can see, the convective PBL-feedback, studied intensively so far, is weak. So it is not surprising that the interest to its analysis has faded away by the last decade. Taking $\tau_{heating}$ = 12 hours and typical $\max h$ of (500 m; 1500 m; 5000 m) corresponding to shallow, mid-latitude and deep tropical convection without significant release of the latent heat, the feedback gain g will be (4.4; 14.7; 45.0), which is much larger than the gains quoted in Bony et al. (2006).

It should not be confused however with the impact of the convection parameterizations in the climate models on the model results. Firstly, $\max h$ could be greatly erroneous, usually smaller, and hence the PBL-feedback is stronger. Secondly, the PBL-feedback is interacting with other feedbacks in the model, which may greatly enhance the overall model response.

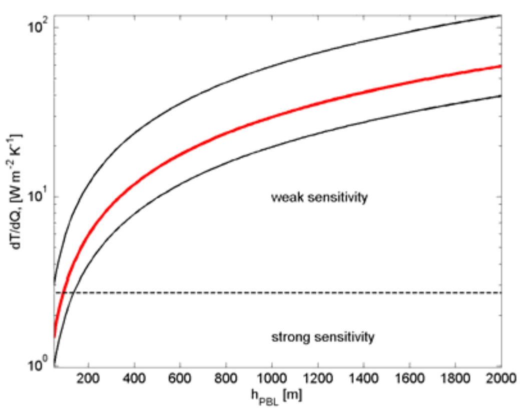


Figure 1. The analytical representation of the PBL-feedback after Eq. (7). The red line is for $\tau_{heating}$ = 12 hours; the upper black line – for 18 hours; the lower black line – for 6 hours.

In the stably stratified PBL,

$$\begin{aligned} h &= \frac{u_*}{f^{1/2}}\left(\frac{f}{C_R^2} + \frac{N}{C_{CN}^2} + \frac{\beta}{C_{NS}^2\rho c_p u_*^2}Q\right)^{-1/2} \\ &= A(B+CQ)^{-1/2} \end{aligned} \quad (8)$$

$$A = \frac{u_*}{f^{1/2}},\ B = \frac{f}{C_R^2} + \frac{N}{C_{CN}^2},$$

$$C = \frac{\beta}{C_{NS}^2\rho c_p u_*^2}.$$

Here, f [s^{-1}] is the absolute value of the Coriolis parameter; u_* [m s^{-2}] is the friction velocity. Constants C_R = 0.65, C_{CN} = 1.36 and C_{NS} = 0.51 were empirically fitted to the large-eddy simulations (Zilitinkevich et al., 2007). Assuming a constant cooling over time $\tau_{cooling}$ and no changes in the atmospheric stratification N , Eq. (8) results in the following PBL-feedback estimation

$$\frac{\delta\theta_a}{\delta Q} = \frac{1}{2\rho c_p}\left(3h^{-1} - \frac{B}{A^2}h\right)\tau_{cooling} \quad (9)$$

This equation is intriguing. It suggests a possibility for sign change for the PBL-feedback parameter in stably stratified PBL. It happens when.

$$h = h^0 = A\sqrt{3/B} \quad (10)$$

Taken f = 1.2 10^{-4} s^{-1}, u_* = 2.5 10^{-2} m s^{-1}, N ~ 10^{-2} s^{-1} in long-lived stable PBL, N ~ 0 s^{-1} in nocturnal stable PBL. Then the expected critical PBL thickness will be 55 m and 235 m correspondingly. These numbers are frequently observed under the assumed conditions. If the parameter values are hold, the climate feedback in shallower PBL $h < h^0$ will be positive and in deeper PBL $h < h^0$ will be negative. The positive feedback denotes here temperature increase in response on reduction of the negative heat flux divergence. The negative feedback denotes paradoxical temperature decrease. Whether the negative feedback is observed on the climate time scales needs further investigations.

3. ESTIMATIONS OF THE PBL-FEEDBACK EFFICACY THROUGH TURBULENCE-RESOLVING SIMULATIONS

A turbulence-resolving simulation technique is a suitable tool to obtain efficacy of the PBL-feedback in the absence of other feedbacks. This work has been done using the DATABASE64 data base of several tens of independent model runs produced by the large-eddy simulation code LESNIC (Esau and Zilitinkevich, 2006).

The PBL-feedback in the stably stratified PBL is presented in Figure 2. Observe that in these runs, B/A^2 is not kept constant. It results in a slight scatter of the direct λ estimations from the DATABASE64 (black dots) around its analytical estimation after Eq. (9) (curves).

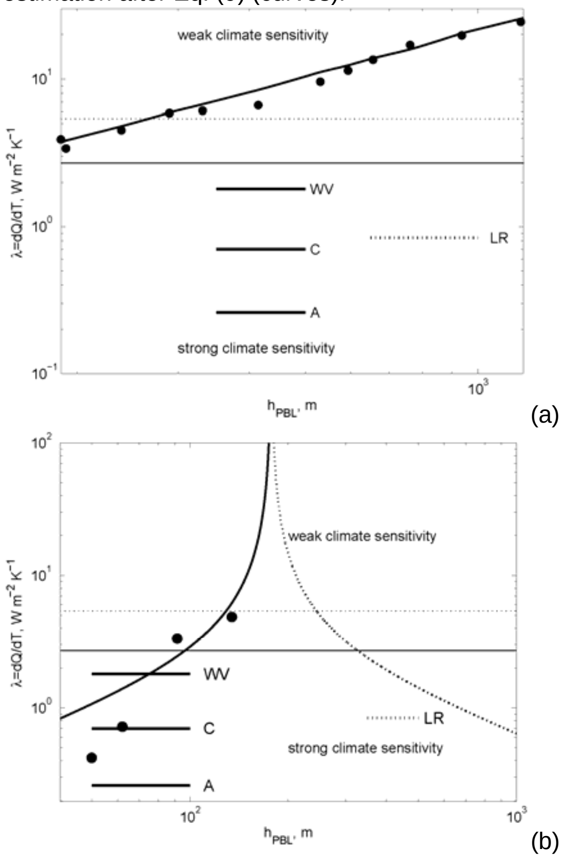


Figure 2. Falsification of the analytically obtained PBL feedback parameter in the nocturnal PBL (a) and the long-lived stably-stratified PBL (b) against DATABASE64. The bold solid curve represents the PBL-feedback from Eq. (9) with h and B/A^2 parameter taken from DATABASE64 runs and piecewise (extra-) interpolated to cover the interval of h variability. The dotted curve is extrapolation of the feedback in the parameter interval of the negative feedbacks. Bold dots represent the PBL feedback parameter directly computed from DATABASE64. The negative PBL-feedback has not been captured in DATABASE64 runs with the selected value of B/A^2. Thin horizontal line represents the commonly quoted total climate sensitivity of 2.7 W m^{-2} K^{-1}(solid line) and 50% of that value (dashed line). Values of other feedback parameters from Bony et al. (2006) are given for comparisons for: A – the ice-albedo feedback; C – the cloud feedback; WV – the water vapor feedback; and LR – the lapse rate feedback.

The agreement between the analytical theory and the nocturnal runs in DATABASE64 is remarkable. The cruelest assumption in derivations of Eq. (9), namely, the assumption of time independence of h, seems to exhibit very little impact on the accuracy of the PBL-feedback estimations.

Bony et al. (2006) work can be used to compare the PBL-feedback strength with the strength of the other feedbacks. In Figure 2, the feedback parameters are given for the ice-albedo, cloud, water vapor and lapse rate feedbacks. It is obvious that the nocturnal PBL-feedback is generally weak. Long clear sky night and low wind are required to make the SBL-feedback equally strong to the total of the other feedbacks. Long-lived PBL-feedback is considerably stronger as the free atmosphere lapse rate imposes limitations on h. The lapse rate and PBL-feedbacks should be interacting so that they should be considered in a coupled radiation-convection model where the convection module should be based on turbulence-resolving model.

The turbulence-resolving simulations allow for more detailed research of the PBL-feedback composition. Particular effects of the changes in the mean atmospheric lapse rate (the stratification of the free atmosphere N) and the nocturnal heat flux divergence Q (this quantity is very sensitive to cloudiness) are investigated.

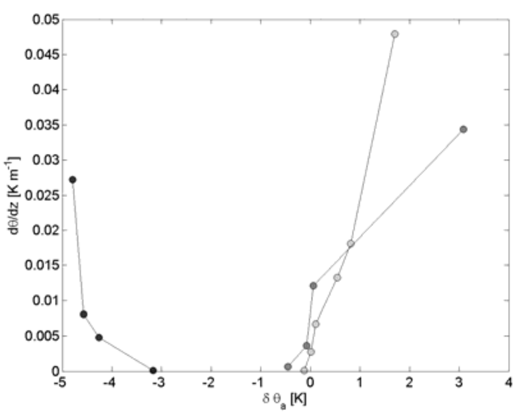


Figure 3. The asymptotic temperature response in the long-lived stably stratified PBL cooled at the fixed rates Q = 2 Wm^{-2}; 4 Wm^{-2}; and 8 Wm^{-2} from light to dark dots correspondingly.

Figure 3 shows the PBL-feedback as a function of N. Whereas the response on stability increase in a strongly cooled PBL (the darkest dots on the left of the plot) is in line with expectations, the response in a weakly cooled PBL is contra-intuitive. The simulations revealed that the weakly cooled PBL shows actual increase of θ_a in response on increasing stability of the free atmosphere, and thus according to Eq. (8), in the response on the reduction of h. This is explicit demonstration of the possibility to the stably stratified PBL-feedback to change sign under realizable conditions. The reason for such an increase of θ_a is clearly identified in Figure 4 where the normalized profiles of the sensible heat flux are given for the runs cooled by 8 W m^{-2} (dotted line) and 2 W m^{-2} (bold line). It is clearly seen that in the latter case the downward heat flux from the free atmosphere is comparable or even larger than the surface cooling. Thus, strengthening of the capping inversion, which is expectable effect of the climate change in the Arctic, may reverse the already weak surface cooling.

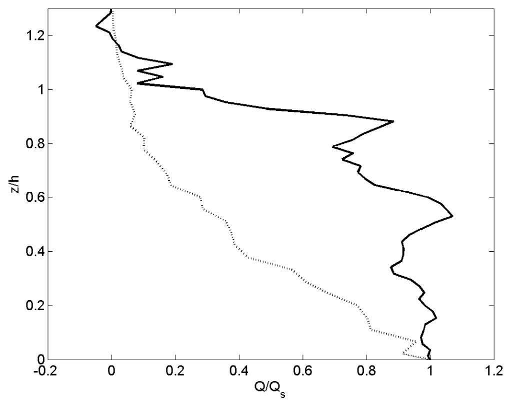


Figure 4. Normalized vertical profiles of the sensible heat flux given for the runs cooled by 8 W m^{-2} (dotted line) and 2 W m^{-2} (bold line) from Fig. 3.

Unlike the effect of the mean temperature lapse rate in the atmosphere, the effect of variations in the nocturnal (N = 0) cooling is not surprising. Figure 5 shows that the clear-sky nights, where the large heat flux divergence and therefore the strong cooling are observed, should be more sensitive (small values of the feedback parameter λ) to the heat flux perturbations δQ . It is interesting however that even a moderate wind (U = 5 m s^{-1} as it is in these runs) in clear-sky nights makes the nocturnal PBL-feedback relatively weak.

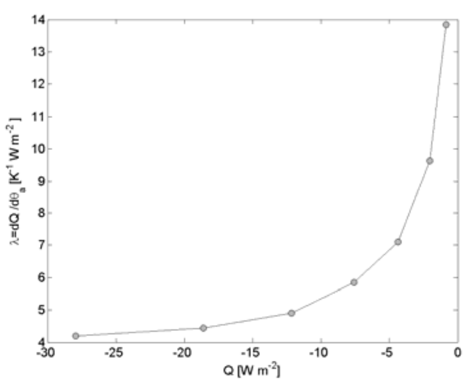


Figure 5. PBL-feedback parameter in clear-sky nocturnal (N = 0) PBL as function of the surface sensible heat flux divergence obtained from DATABASE64 with the geostrophic wind speed U = 5 m s^{-1}.

4. PBL-FEEDBACK IN ERA-40 REANALYSIS DATA

ERA-40 reanalysis (Uppala et al., 1999) provide the necessary data to estimate the PBL-feedback in the ECS. It can be calculated after Eq. (4) but in this case the unknown sensitivity $\partial h / \partial Q$ needs to be obtained. The PBL-feedback can be also calculated after Eqs. (7) and (9) utilizing the good agreement between those simplified analytical equations and the turbulence-resolving simulations. In this case however, the succession of heating and cooling periods needs to be obtained to complete the integration. This work is in progress. Therefore, here, only preliminary results and indirect support to the PBL-feedback are presented.

Figure 6 presents zonal variations of the surface air temperature trend (1979-2001) in the ERA-40 and the CRU (Climate Research Unit, Jones and Moberg, 2003) data along with the inverse zonally averaged PBL thickness in the ERA-40 and the CHAMP (CHAllenging Minisatellite Payload radio occultation, von Engeln et al, 2005) satellite data.

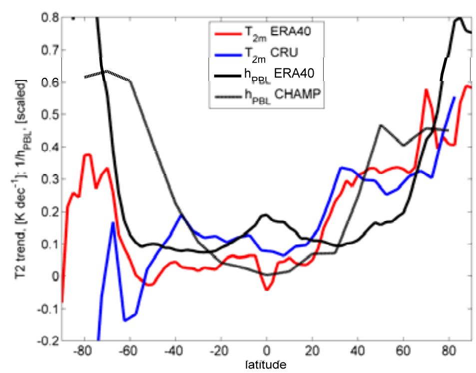
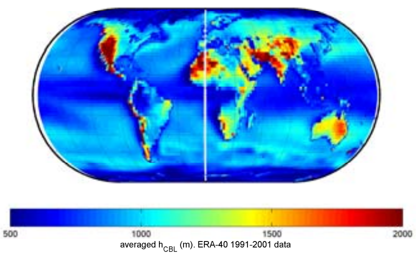


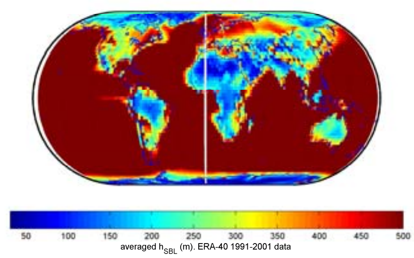
Figure 6. Zonal variations of the surface air temperature at 2m trends (1979-2001) [K dec^{-1}] in ERA-40 (red), CRU (blue) data along with zonal variations of the inverse averaged PBL thickness (in normalized units) from ERA-40 (solid black curve) and the CHAMP (dotted curve) satellite data sets.

Although the co-variation of those quantities is not perfect, the general tendency of the temperature trend to be larger in the areas of shallower PBL is clearly observed. Especially interesting is sharp increase of both the trend and h^{-1} (from the CHAMP data) in the latitude band 25N-35N. This is the band of tropical deserts where air subsidence in the anticyclones makes the PBL shallow.

The averaged PBL-thickness in the ERA-40 data is given in Figure 7 for convective and stably stratified conditions (observe the difference in the color scales). One drawback of the ERA-40 algorithm is that it erroneously determines the thickness of the PBL with convective clouds, which gives a false impression as if the PBL in the equatorial and convergence zones is shallower than the PBL in the tropical deserts. As expected, h over the ocean is fairly constant due to the large heat capacity of the underlying surface.



(a)



(b)

Figure 7. The mean PBL thickness h in the ERA-40 data for convective (a) and stably stratified (b) PBL.

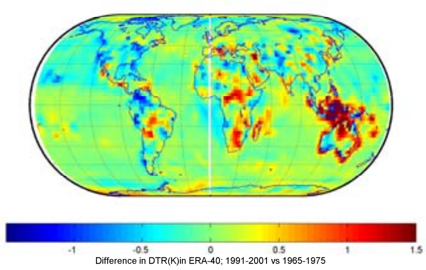
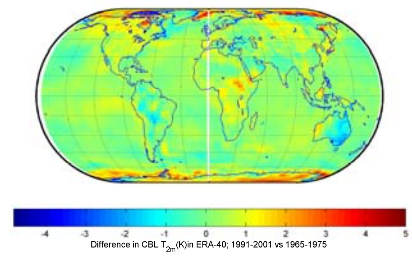
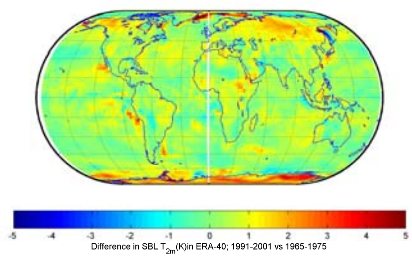


Figure 8. The difference in the diurnal temperature range between 1991-2001 and 1965-1975 periods in the ERA-40 data.

From Eqs. (4) and (9), one expects that θ_a increases significantly in the areas of thin PBL. This is not quite the case as Figure 7 and 9 reveal. One notorious disagreement with this expectation is large increase of θ_a in Western Siberia in the area of obvious storm tracks. In these areas the negative PBL-feedback probably dominates. Diurnal temperature range (DTR) changes (Figure 8) are also not in particular agreement with the theoretical expectations. Moreover, they disagree with Easterling et al. (1997) and Vose et al. (2005) analysis. This analysis has been carried out without selection of clear-sky cases, so that cloudiness may affect the result significantly. It seems that in areas of generally clear sky, like the tropical deserts and Eastern Siberia, the DTR has decreased as predicted.



(a)



(b)

Figure 9. The difference in the surface air temperature between 1991-2001 and 1965-1975 periods in the ERA-40 data for convective (a) and stably stratified (b) PBL.

5. SUMMARY

In this study, the PBL-feedback in the Earth's Climate System has been formulated. The effects and efficacy of the feedback were first obtained analytically through application of the asymptotic equations for the PBL thickness. The analytical work has been supported with analysis of the turbulence-resolving simulations from the DATABASE64. The research is now focused on quantification of the PBL-feedback in the ECS data. One of possible source of data is the ERA-40 reanalysis. Its processing meets some difficulties, notoriously, the low quality of the PBL thickness and fluxes (e.g. Serreze et al., 2007) determination. The work with ERA-40 continues.

APPENDIX

The PBL-feedback from the turbulence-resolving simulations in the DATABASE64 was computed as the follows. Each run in the DATABASE64 was conducted for 16 hours under prescribed control parameters. The first 12 hours of data, averaged over 1 hour intervals, were sampled where the first sample defines the initial conditions and the last sample defines the steady-state conditions of the run. The temperature change was determined as

$$d\theta_a / dt \sim \delta\theta_a / \delta t = \\ = \left(\theta_a\big|_{t=12} - \theta_a\big|_{t=1}\right)/(12 \cdot 3600) \qquad \text{(A.1)}$$

The PBL thickness was determined as the height where the momentum flux becomes less than 5% of its surface value. The PBL thickness is always taken from the last sample. Each run in the DATABASE64 has prescribed U, N and Q. To estimate the feedback parameter, different runs, each with own combination of the parameters, must be utilized, e.g.

$$d\theta_a / dQ \sim \delta\theta_a / \delta Q = \\ = \frac{\left[\underbrace{\theta_a\big|_{t=12} - \theta_a\big|_{t=1}}_{run\quad i}\right] - \left[\underbrace{\theta_a\big|_{t=12} - \theta_a\big|_{t=1}}_{run\quad j}\right]}{\left[\underbrace{Q}_{run\quad i}\right] - \left[\underbrace{Q}_{run\quad j}\right]} \qquad \text{(A.2)}$$

where $i \neq j$.

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