Abrupt monsoon transitions as seen in paleo-records can be explained by moisture-advection feedback

Long version of Comment on “Near-linear response of mean monsoon strength to a broad range of radiative forcings” by W.R. Boos and T. Storelvmo

A. Levermann[1,2,3,*], V. Petoukhov[1], J. Schewe[1], H.J. Schellnhuber[1,4]

[1] Potsdam Institute for Climate Impact Research, Potsdam, Germany

[2] Columbia University, New York, USA

[3] Potsdam University, Potsdam, Germany

[4] Santa Fe Institute, Santa Fe, New Mexico, USA

* correspondence should be addressed to: anders.levermann@pik-potsdam.de

Abrupt monsoon changes exist in paleo-records

Paleo-climatic records show evidence of abrupt and strong monsoon shifts during the last two glacial cycles (Burns et al. 2003; Wang 2005; Wang et al. 2008), the last deglaciation (Stager et al. 2011) and the Holocene (Gupta et al. 2003; Hong et al. 2003; Wang 2005; Berkelhammer et al. 2013; Dixit et al. 2014) in India, the Bay of Bengal, and East Asia. While some of these shifts have been linked either to variations in solar insolation or to climatic changes in other parts of the planet, they are often much more abrupt and/or larger than expected for a linear response to external forcing, thus suggesting that monsoon systems may be capable of non-linear transitions. Such transitions were first reproduced in a conceptual model by (Zickfeld et al. 2005) and later for the more specific case of the Wang et al. 2008 data by (Schewe et al. 2012). Levermann et al. (2009) carved out the physical mechanism for such transition in a very simple conceptual model stating in the abstract "Though details of monsoon circulations are complicated, observations reveal a defining moisture-advection feedback that dominates the seasonal heat balance and might act as an internal amplifier, leading to abrupt changes in response to relatively weak external perturbations. Here we present a minimal conceptual model capturing this positive feedback."

Both conceptual models agree when operating in physically reasonable regime

Boos & Storelvmo (2015) based their article on the statement that the introduction of adiabatic cooling into the conceptual monsoon model of Levermann et al. (2009) eliminates the abrupt transitions. That is not true as can be seen in their figure 1a. Boos & Storelvmo obtain the same abrupt transition if the adiabatic cooling is not compensating for the latent heat release. Thus their argument is based on the assumption that most of the energy from latent heat release is consumed by adiabatic cooling. While we are obviously aware of the existence of the physical process of adiabatic cooling (and show below that it is implicitly accounted for in our model), it is not a valid assumption that most of the latent heat release is consumed by this process. To the contrary, the monsoon circulation over the continent is predominantly sustained by the release of latent heat and subsequent warming of the atmospheric column over land as stated in a number of classic studies as for example (Webster et al. 1998).

For that reason we deliberately addressed in our minimalistic model (and we clearly declared this in the article) the simplest, but fundamentally important cases with predominantly advective character of the low-atmosphere circulations in the main monsoon regions. These situations are illustrated by Figure S2 from our article, where we showed maps of the low-troposphere winds over these regions, with dominant contribution from the advective component. It is further shown in our Figure 2 which compares the latent heat release to other energy components for different regions.

Adiabatic cooling in the Levermann et al. (2009) model

A very specific incorporation of adiabatic cooling into our conceptual model let Boos and Storelvmo (2015) to their central result of a vanishing of the threshold in monsoon rainfall. The specific representation they choose is based on an approximation of the second horizontal derivative of the temperature by a linear function of its first horizontal derivative (horizontal velocity), dividing the velocity scale by the horizontal length scale (their equations (1)-(3) and (S1) and (S2)). This approximation is very crude. Here we show that our model implicitly incorporates adiabatic cooling without eliminating the possibility of abrupt monsoon transitions as seen in paleo-records:

Levermann et al. 2009 paper is based on the classical equations and conventional assumptions. In that article we start with the equation for the specific entropy (see, e.g., (Lorenz 1967, page 13)) in z-coordinate:

$$\frac{ds}{dt} = \frac{Q_m}{T}, \qquad (1)$$

where $s = c_p \ln \vartheta$ is the specific entropy of the air, c_p and ϑ are the specific heat at constant pressure and the potential temperature of the air, t is time, Q_m is the net heating rate per unit mass, and T is kinetic temperature. Multiplying the left hand side of equation (1) by the air density ρ and using the continuity equation yields:

$$\rho \frac{ds}{dt} = \frac{\partial \rho s}{\partial t} + \nabla \cdot (\rho s \vec{V}), \qquad (2)$$

where $\partial / \partial t$ is the partial derivative with respect to time, ∇ is the three-dimensional gradient vector, $\vec{V}$ is the three-dimensional vector of atmospheric velocity, and $\cdot$ is a scalar multiplication sign. Integrating equation (2) with respect to z vertically from the earth's surface to the tropopause and horizontally over the monsoon land region, assuming quasi-stationarity of the process and imposing zero boundary conditions on the vertical velocity w at the surface and at the tropopause, z_{tr}, using equation (2) one obtains

$$c_p \int_{\Sigma} \left(\int_0^{z_{tr}} \nabla_H \cdot \rho \vartheta \vec{V}_H dz \right) d\sigma = \int_{\Sigma} \left(\int_0^{z_{tr}} \frac{\vartheta}{T} Q_{V,P} dz \right) d\sigma + \int_{\Sigma} \left(\int_0^{z_{tr}} \frac{\vartheta}{T} Q_{V,R} dz \right) d\sigma, \qquad (3)$$

where ∇_H is the horizontal gradient vector, $\vec{V}_H$ is the vector of the horizontal velocity, Σ is the total area of the monsoon land region, while $Q_{V,P}$ and $Q_{V,R}$ are, respectively the net heating rate per unit volume due to the condensation and radiation. In our simple monsoon model we neglected the sensible heat flux at the land surface. The condensation heating rate is positive throughout the entire troposphere. The radiation heating rate is non-positive throughout the entire troposphere in the monsoon land regions (McFarlane et al. 2007). The factor ϑ / T is positive in the troposphere. Thereby one can bring this factor outside the integrals in the right side of equation (3), representing it by constant parameters S_P , and S_R , respectively, in the first and second terms in the right hand side of equation (3). We can denote S_P, and S_R as the integral static stability parameters of the monsoon land system, associated with the condensation and radiation processes. This way, the deviation of ϑ / T from 1 is the indicator of the contribution from the vertical motions to the atmosphere heat balance: in the adiabatic (i.e., with $T = \vartheta$) atmosphere the vertical motions could not contribute to the heat balance. Equating condensation to precipitation rate over the monsoon land region and prescribing as the parameter the ratio $\varepsilon = H / L$, where H and L are respectively, the vertical extent of the lower branch of the monsoon inflow to the land and the horizontal scale (distance between the coast line and the remote boundary of the monsoon region, see Levermann et al. 2009), we get the equation whose general structure is identical to Eq. (1) from that paper, namely

$$LPS_P - \varepsilon c_p W \Delta T + RS_R = 0, \quad (4)$$

where L is the latent heat of condensation, P and R are the averaged over the entire monsoon land region precipitation and radiation rates per unit square, and

$$W = \alpha \Delta T \ , \quad (5)$$

where α is a constant parameter and $\Delta T = T_L - T_O > 0$ is atmospheric temperature difference between land and ocean.

When deriving equation (4) we use a conventional representation of the potential temperature $\vartheta \approx T + \gamma_A z$ within the troposphere, where γ_A is the constant adiabatic lapse rate, so that the horizontal derivatives of ϑ and T match up in that case.

A conservation law for the water vapor mass places a condition that an overall influx of the water vapor into the system should be equal to zero, for the steady states. This means, in our case, that in the absence of the land evaporation and under the condition that the vertical velocity is zeroed at the surface and at the tropopause, an overall (integrated over the entire side surface of the monsoon land region) income of water vapor at the lateral boundaries of the system should be balanced by the outcome of water vapor due to precipitation rate integrated over the entire lower surface of the monsoon land region. Going this way we get the equation for the atmospheric humidity that is identical to the equation (3) from Levermann et al. 2009:

$$\varepsilon W \rho (q_O - q_L) - P = 0, \qquad (6)$$

where q_O and q_L are the specific humidity over ocean and land, respectively.

We further represent P as follows (see Levermann et al. 2009):

$$P = \beta q_L, \qquad (7)$$

where β is constant parameter. Finally, combining equations (4) - (7) one can get

$$W^3 + \frac{\beta}{\varepsilon\rho} W^2 - \frac{\alpha}{\varepsilon c_p}(LS_P \beta q_O + RS_R) W - \frac{\alpha\beta}{\varepsilon^2 \rho c_p} RS_R = 0, \qquad (8)$$

The general structure of this equation is identical to that of the governing equation in Levermann et al. 2009.

As a consequence, the threshold behavior is not eliminated by adiabatic cooling unless it consumes practically all of the energy of the latent heat release.

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