A simple model of a balanced boundary layer coupled to a large-scale convective circulation

Bob Beare [1] and Mike Cullen [2]

[1]University of Exeter.

[2]Met Office, Exeter.

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Introduction

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- Reverse approach- start with large-scale balances and how do the physics preserve these?
- Tropical troposphere follows Weak Temperature Gradient (WTG) approximation.
- Significant gradients of temperature within boundary layer (connected to sea-surface temperature). Associated gradients of pressure in balance with the drag.

Climatology (Back and Bretherton 09)

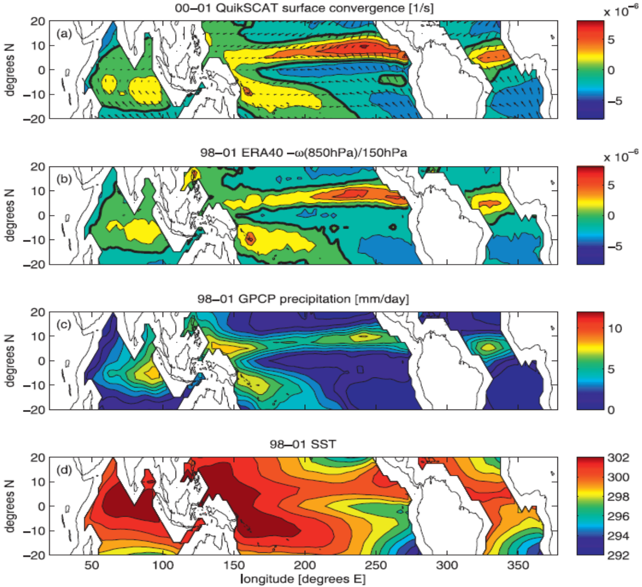
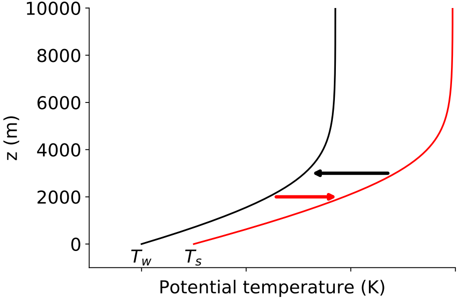


FIG. 1. (a) 2000–01 surface convergence from QuikSCAT with contours of $2 \times 10^{-6}\ s^{-1}$ (heavy contour shows zero convergence). (b) 1998–2001 ERA-40 $-\omega_{850}/(150\ hPa)\ s^{-1}$, representative of mean convergence in the boundary layer, with the same contours as (a). The GPCP 1998–2001 (c) precipitation (contours of 1 mm day^{-1}) and (d) SST (contours of 1 K).

Weak temperature gradient profile

Maintenance of WTG profile

- Convection tries to relax to moist adiabat from the SST (T_s).
- Equal and opposite relaxation back to WTG (T_w).



Components of simple model

1. Maintenance of WTG vertical profile
2. WTG mass balance
3. Boundary-layer balance
4. Moisture balance

Convection layer

$$M_c = \gamma_c \frac{T_s - T_w}{\tau_c} \quad |x| \le L_c/2,$$
$$\frac{P}{L\rho_0 H} = \gamma_q \frac{q_s - q_w}{\tau_c} \quad |x| \le L_c/2,$$

where M_c is the mass flux divided by density, P the precipitation flux, τ_c the relaxation timescale.
Assume WTG and a constant radiatively-driven subsidence velocity (w_s).
Mass balance in the Convection layer is

$$L_x w_s + L_c \langle M_c \rangle = 0$$

where L_x domain length and L_c width of convection. Angle brackets are horizontal average over the convecting region.

Schematic of model

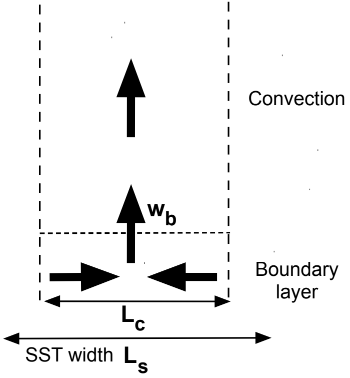


Figure: Schematics of the flows and length scales in the simple model

Mass flux

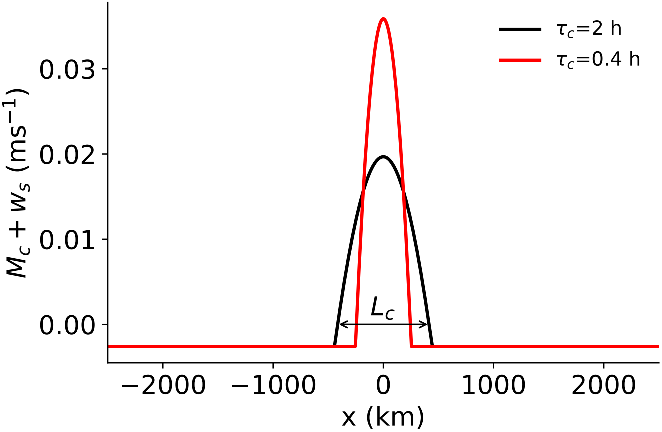


Figure: The sum of subsidence and mass flux for the WTG layer. Shown are profiles for the control ($\tau_c = 2$ h, black) and $\tau_c = 0.4$ h (red). The convective width for $\tau_c = 2$ h is marked by the horizontal arrow.

SST and WTG temperatures

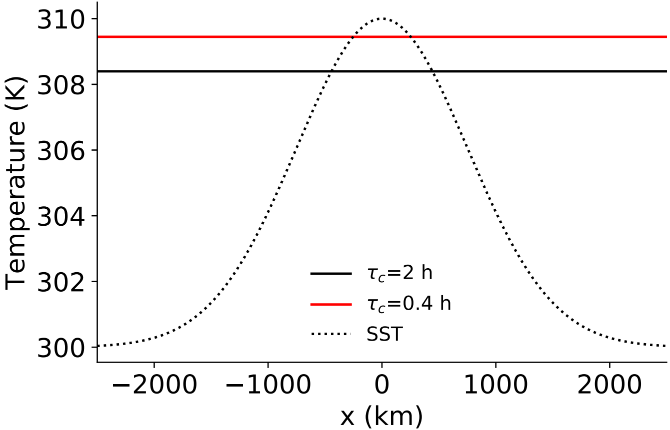


Figure: The SST and WTG temperatures for the control ($\tau_c = 2$ h, black horizontal line) and $\tau_c = 0.4$ h (red horizontal line).

Contraction of convection width

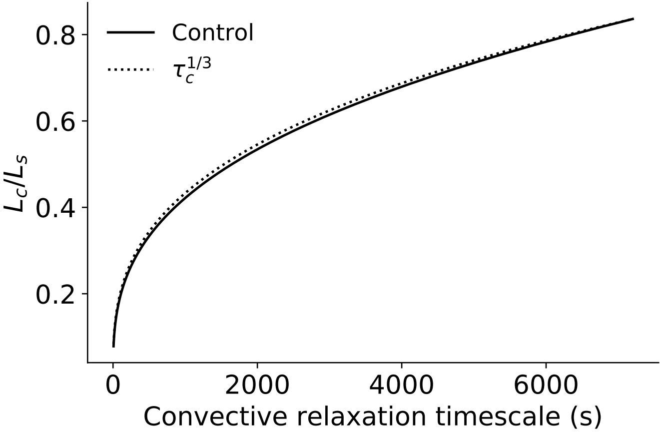


Figure: The convective width (normalised by width of SST) plotted against convective relaxation timescale. Also shown is a $\tau_c^{1/3}$ power law (dotted).

Maximum mass flux

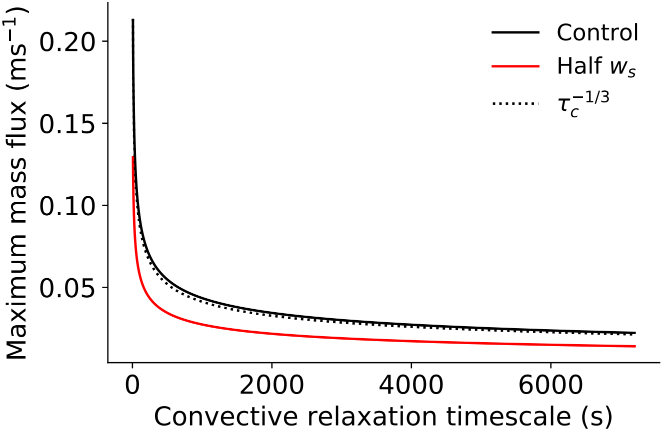


Figure: Maximum mass flux plotted against convective relaxation timescale for the control (black) and half subsidence (red) cases. Also shown is a $\tau_c^{-1/3}$ power law (dotted).

Coupling to boundary layer

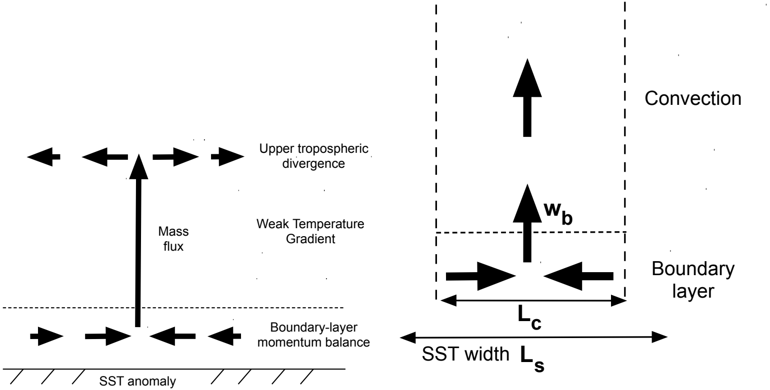


Figure: Schematics of the flows and balances in the simple model

Coupling to boundary layer

In contrast to the WTG convection layer, thermal gradients are significant within the boundary layer, so we need a momentum balance

$$\overbrace{\frac{d\phi_b}{dx}}^{\text{Pressure gradient}} = \overbrace{-\frac{u_b}{\tau_b}}^{\text{Drag}},$$

where u_b is boundary-layer wind, ϕ_b geopotential, τ_b the Rayleigh boundary-layer timescale. Boundary-layer top vertical velocity (w_b) is calculated using continuity and hydrostatic balance is given by

$$w_b = -\frac{du_b}{dx}h, \quad \phi_b = -\frac{h}{2}\frac{g(\theta_b - \theta_0)}{\theta_0}$$

where h is the boundary-layer depth. The boundary layer potential temperature matches the ascent in the convection region.

$$-\frac{\tau_b g h^2}{2\theta_0}\frac{d^2\theta_b}{dx^2} = w_b = M_c + w_s.$$

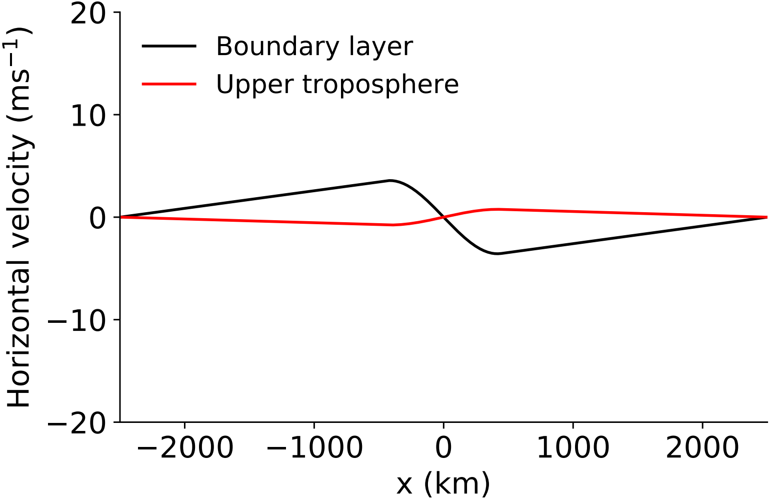


Figure: Horizontal winds for : boundary layer (u_b, black) and upper troposphere (u_u, red) for control case.

Moisture fluxes

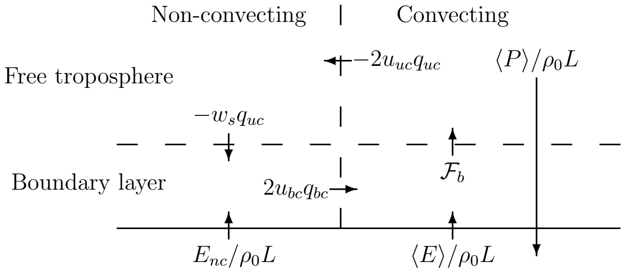


Figure: The moisture fluxes assumed between convecting (right), non-convecting (left), boundary layer and free troposphere regions. All fluxes shown are positive.

Evaporation

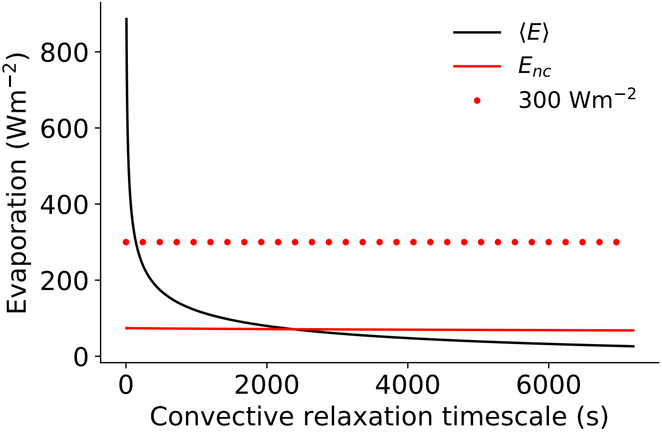
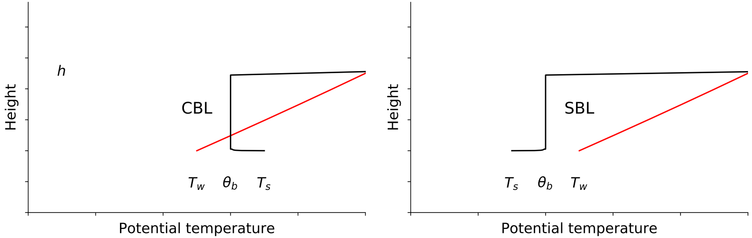


Figure: The evaporation averaged over convecting region ($\langle E \rangle$) and from the non-convecting region (E_{nc}) plotted against convective relaxation timescale. The red dotted line is the 300 Wm^{-2} threshold.



Vertical profiles of potential temperature for the boundary layer (black) with respect to the WTG moist adiabat based on T_w (red) for the: (left) convective boundary layer or (right) stable boundary layer.

Boundary-layer potential temperature

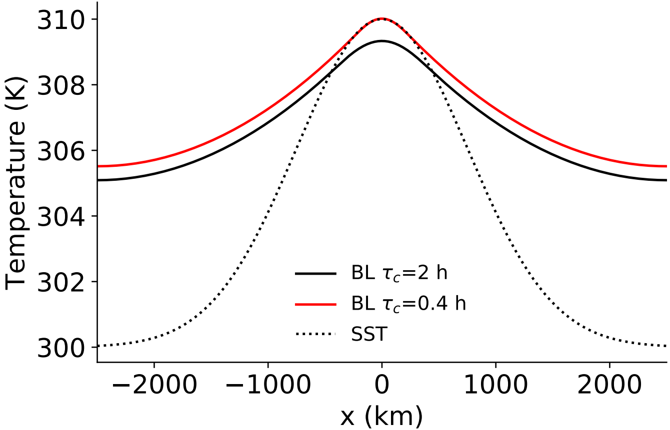


Figure: The distribution of SST (dotted) and boundary-layer potential temperature for $\tau_c = 2$ h (black) and $\tau_c = 0.4$ h (red).

Sensitivity to drag

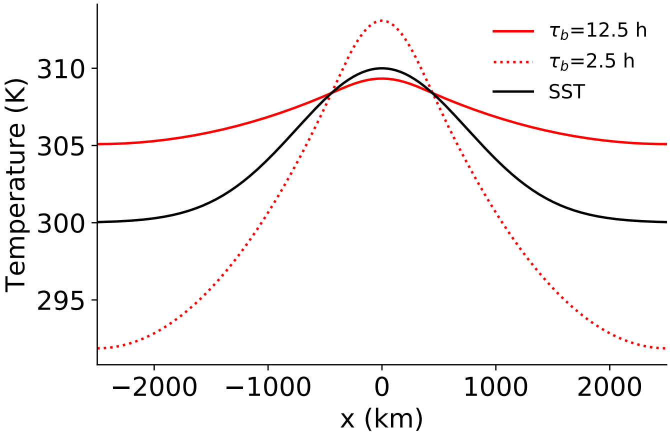


Figure: The sensitivity of the boundary-layer potential temperature to decreasing τ_b (increasing drag) from 12.5 h (red) to 2.5 h (red dotted).

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  - How well does convection scheme maintain WTG?
  - Does the convective width decrease with increased efficiency of convection scheme?
  - How close to Ekman balance is tropical convergence on weekly-monthly timescales?
  - See Susannah Hearn's poster - defining balanced regimes in MetUM.

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  - See Susannah Hearn's poster - defining balanced regimes in MetUM.
- Beare and Cullen (2018), *submitted to JAS*.
- See also Beare and Cullen (2013 PhilTransA, 2016 QJRMS) for balanced diagnostics applied to mid-latitudes.

Cullen 2018, in preparation

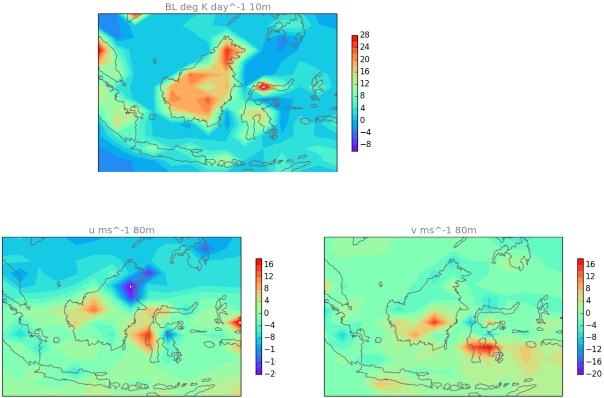

Figure 4. Diagnostics calculated over a region 10°S to 10°N and 100°E to 130°E at 80m height above the surface. (a) Boundary layer heating, units °K day^{-1}: (b) total zonal wind calculated from (19), (c) total meridional wind calculated from (19).

Diagnosed convergence from MetUM over Borneo, based on Ekman balance (semi-geotriptic theory).