Wind Limits on Rain Layers and Diurnal Warm Layers

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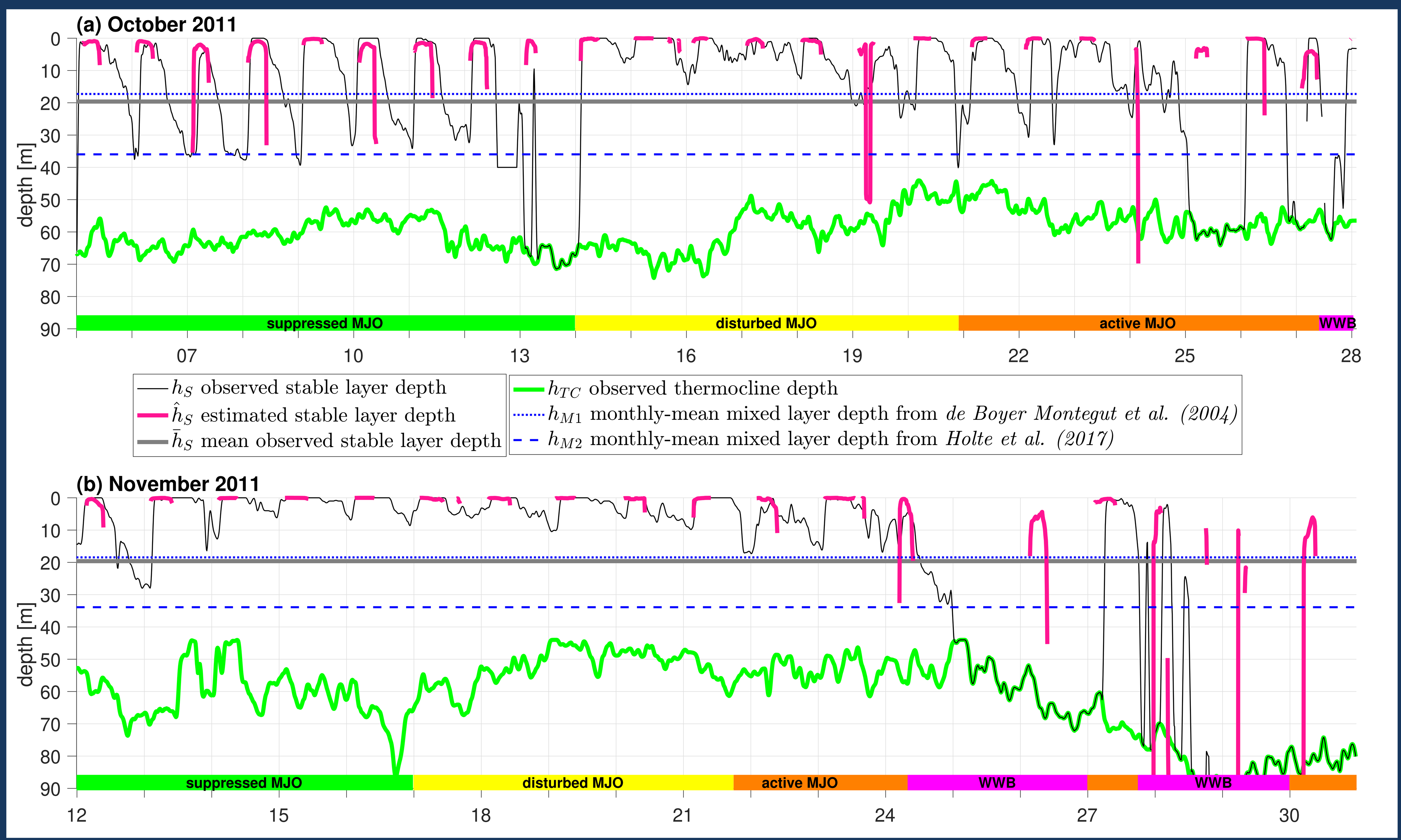
Results

- **Significant density stratification at depths z ≤ 5 m was observed in 38% of a 2-month data set from the central Indian Ocean collected during the DYNAMO experiment (Dynamics of the MJO); Data: 30 warm layers, 32 rain layers, 17 RL-DWL combinations**

- **Wind limit on rain layers: $U_{10} \leq 9.8$ m s^{-1}**
- **Wind limit on warm layers: $U_{10} \leq 7.6$ m s^{-1}** ***limit = 99th percentile values of U_{10}***

- **Rain layers (and their combinations with diurnal warm layers) formed often in disturbed and active MJO periods before westerly wind bursts. Diurnal warm layers were most commonly observed in suppressed and disturbed MJO conditions**
- **Upstream rain input to the ocean was tracked with shipboard radar, and analyzed in addition to shipboard rain gauges.**

- **Estimates of $\widehat{h_S}$ and $\widehat{U_S}$ from U_{10} and B predicted:**
  - **36 out of 44 observed stratification events, 88% success rate**
  - **the wind limits of these events (99th % values of U_{10})**
- **This suggests a means to determine the presence of ocean stable layers at depths of z ≤ 5 m from U_{10} and B.**



The shallowest stable layer depth in the ocean was determined and tracked from DYNAMO in situ T and S observations from Chameleon profiler. From the 10-m adjusted wind speed, U_{10}, and surface buoyancy flux, *B*, we derived estimates of $\widehat{h_S}$, ocean stable layer depth (pink), and compared these estimates to DYNAMO observations (black, green) and monthly mixed layer climatologies (blue).

Estimating ocean stable layer depth from surface buoyancy flux and wind speed

From the ocean friction velocity, u_{*W}^3, and surface buoyancy flux, *B*, we derived estimates of $\widehat{h_S}$, stable layer depth, and $\widehat{U_S}$, the maximum U_{10} for which stratification should persist at h_S for fixed *B*. From Dorrestein (1979), the surface buoyancy flux into the ocean is:

$$B = \frac{g}{\rho_W}\left[S_0\beta(P-E) - \frac{\alpha}{c_p}(Q_{\text{Lat}} + Q_{\text{Sens}} + Q_{\text{Solar}} + Q_{\text{IR}}) + \alpha\Delta TP\right]$$

Rain Freshening and Evaporation Term; Heating Term; Rain Cooling Term

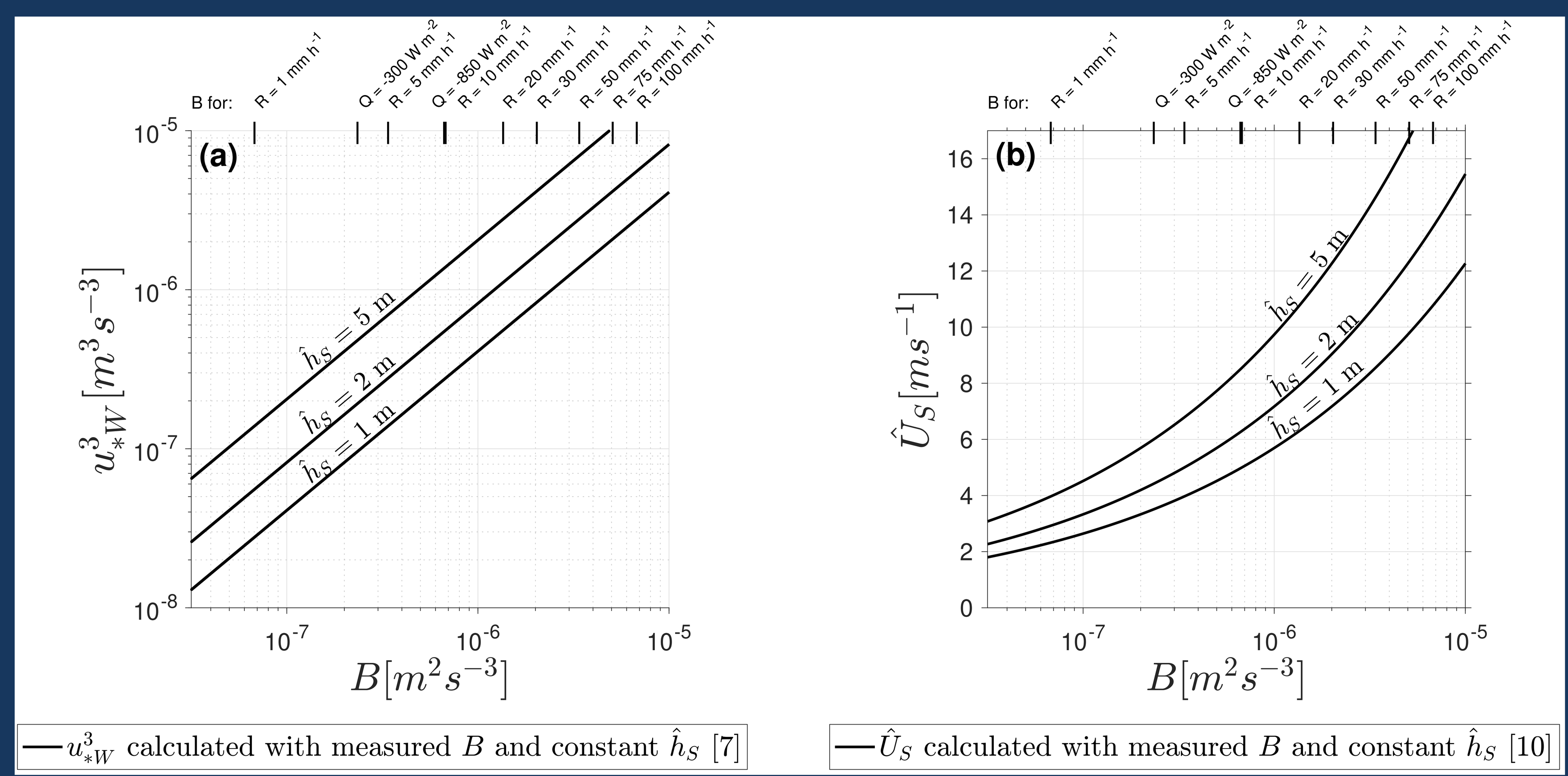
$$\widehat{h}_S = \frac{u_{*W}^3}{\kappa B}$$

$$\widehat{h}_S = C\frac{U_{10}^3}{B}$$

$$\widehat{U}_S = \left(\frac{\widehat{h}_S B}{C}\right)^{1/3} \cong 569\left(\widehat{h}_S B\right)^{1/3}$$

$$C = \frac{1}{\kappa}\left(\frac{C_D\rho_{\text{air}}}{\rho_W}\right)^{3/2} \cong 5.42\times10^{-9}$$

- Rain can produce *B* that is more than an order of magnitude greater than is typically produced by surface heating
- Equivalent *B* is produced by *R* = 10 mm h^{-1} (light to moderate rain rate) and Q_{NET} = -850 W m^{-2} (max daytime heating)
- B produced by rain typically outweighs B destruction by surface cooling.
- The B produced by *R* = 5 mm h^{-1} (light rain rate) is equal and opposite to the B destroyed by cooling of Q_{NET} = 400 W m^{-2} (max surface cooling)
- 1/3 of rain layers formed at night; 4 rain layers lasted from night to day; several rain layers and diurnal warm layers lasted from day to night
- Rain always stabilizes the upper ocean because B produced by rain freshening is 15 × greater than B destroyed by rain cooling.



Calculated (a) u_{*W}^3 and (b) $\widehat{U_S}$ using different stable layer depths of h_S = 1, 2, and 5 m and measured *B*. The value of $\widehat{U_S}$ is the highest value of U_{10} at which wind mixing and *B* should support the existence of a stable layer at depth h_S. Examples of *B* for various net heat fluxes (Q_{Net} abbreviated as *Q* in plot) and rain rates (*R*) are shown for reference.

References:

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