A turbulence closure scheme in the wave boundary layer and its application in a coupled model

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Introduction

The wind stress in current models is estimated by

$$\boldsymbol{\tau}_{\mathrm{a}}(0) = \rho_a C_d |S_{10}| \boldsymbol{S}_{10}$$

- ρ_a is the air density
- $\boldsymbol{S}_{10}$ is the mean wind at 10 m
- $C_d = \frac{\kappa^2}{ln^2(10/z_0)}$ is the drag coefficient

Over the ocean, the roughness length is estimated

$$z_0 = \alpha \frac{u_*^2}{g} + \frac{0.11 \times \nu_a}{max(u_*, 0.05)},$$

α is the Charnock coefficient (depend on wave states), ν_a is the air kinematic viscosity.

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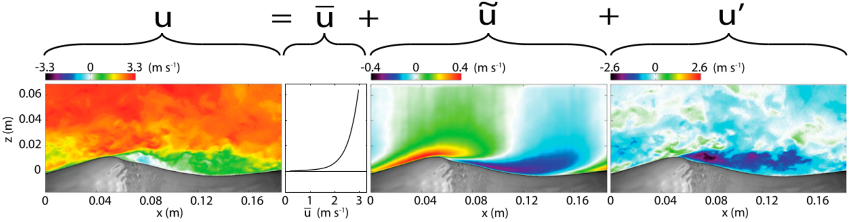
Motivation

Two limitations of current parameterization ($\boldsymbol{\tau}_{\mathrm{a}}(0) = \rho_a C_d |S_{10}| \boldsymbol{S}_{10}$):

1. The wind stress always aligns with wind direction
2. The momentum flux is always from atmosphere to ocean/waves (downward)

Based on eddy-covariance method, $\boldsymbol{\tau}_{\mathrm{a}} = -\overline{u'w'} - \overline{v'w'}$.
In the wave boundary layer,



$$\boldsymbol{\tau}_{\mathrm{a}} = -\overline{u'w'} - \overline{v'w'} - \overline{\tilde{u}\tilde{w}} - \overline{\tilde{v}\tilde{w}}$$

(Buckley and Veron, 2016)

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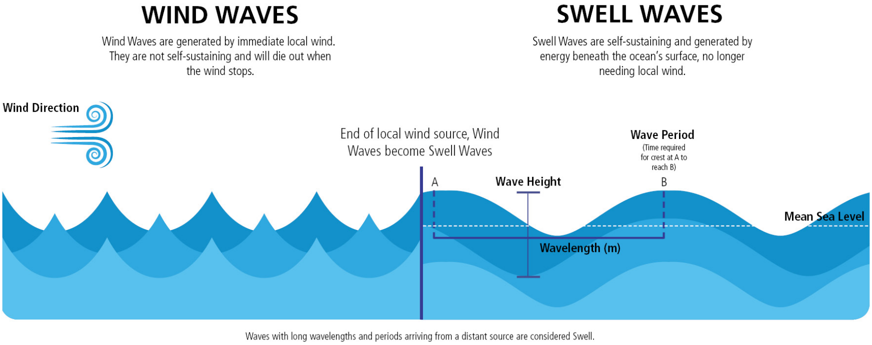
Motivation


WIND WAVES
Wind Waves are generated by immediate local wind. They are not self-sustaining and will die out when the wind stops.
SWELL WAVES
Swell Waves are self-sustaining and generated by energy beneath the ocean's surface, no longer needing local wind.
Wind Direction
End of local wind source, Wind Waves become Swell Waves
Wave Period
(Time required for crest at A to reach B)
A
Wave Height
B
Mean Sea Level
Wavelength (m)
Waves with long wavelengths and periods arriving from a distant source are considered Swell.

How can we paramterize the stress in WBL?

The wind stress in the wave boundary layer:

$$\boldsymbol{\tau}_{\mathrm{a}}(z) = \boldsymbol{\tau}_{\mathrm{t}}(z) + \boldsymbol{\tau}_{\mathrm{v}}(z) + \boldsymbol{\tau}_{\mathrm{ww}}(z) + \boldsymbol{\tau}_{\mathrm{sw}}(z),$$

- $\boldsymbol{\tau}_{\mathrm{t}}$ shear-induced turbulent stress
- $\boldsymbol{\tau}_{\mathrm{v}}$ viscous stress
- $\boldsymbol{\tau}_{\mathrm{ww}}$ wind wave-induced stress (downward)
- $\boldsymbol{\tau}_{\mathrm{sw}}$ swell wave-induced stress (upward)

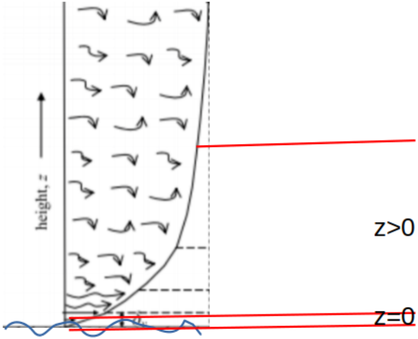
Note:

1. Wave-induced stresses can alter from the wind direction.
2. Swell-induced stress is upward (from waves to atmosphere)
3. Wave-induced stress can extend to a certain height.

How can we paramterize the stress in WBL?

The wind stress in the WBL:

$$\boldsymbol{\tau}_{\mathrm{a}}(z) = \boldsymbol{\tau}_{\mathrm{t}}(z) + \boldsymbol{\tau}_{\mathrm{v}}(z) + \boldsymbol{\tau}_{\mathrm{ww}}(z) + \boldsymbol{\tau}_{\mathrm{sw}}(z),$$



At air-sea interface ($z \approx 0$), the shear-induced stress approches to 0:, then

$$\boldsymbol{\tau}_{\mathrm{a}}(0) = \boldsymbol{\tau}_{\mathrm{v}}(0) + \boldsymbol{\tau}_{\mathrm{ww}}(0) + \boldsymbol{\tau}_{\mathrm{sw}}(0)$$

In the wave boundary layer (z $>$ 0): $\boldsymbol{\tau}_{ww}(z)$ decay faster and $\boldsymbol{\tau}_{\mathrm{v}}(z)$ approach 0, then

$$\boldsymbol{\tau}_{\mathrm{a}}(z) = \boldsymbol{\tau}_{\mathrm{t}}(z) + \boldsymbol{\tau}_{\mathrm{sw}}(z)$$

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How can we paramterize the stress in WBL?

The wind stress terms are estimated by:

$$\boldsymbol{\tau}_{\mathrm{v}}(0) = \rho_a C_v |S_{10}| \boldsymbol{S}_{10}$$

$$\boldsymbol{\tau}_{\mathrm{ww}}(0) = \rho_w g \int_0^{2\pi} \int_0^{\infty} \frac{\mathbf{k}}{\omega} S_{in} d\omega d\theta,$$

$$\boldsymbol{\tau}_{\mathrm{sw}}(0) = \rho_w g \int_0^{2\pi} \int_0^{\infty} \frac{\mathbf{k}}{\omega} S_{out} d\omega d\theta,$$

$$\boldsymbol{\tau}_{t}(z) = -\rho_a K_m \frac{dS}{dz},$$

$$\boldsymbol{\tau}_{\mathrm{sw}}(z) = \boldsymbol{\tau}_{\mathrm{sw}}(0) e^{-Ak_p z}$$

Note:

1. $\boldsymbol{\tau}_{\mathrm{v}}(0)$, $\boldsymbol{\tau}_{\mathrm{t}}(z)$ align with the mean wind direction.
2. The direction of $\boldsymbol{\tau}_{\mathrm{ww}}(z)$ and $\boldsymbol{\tau}_{\mathrm{sw}}(z)$ is determined by 2D wave spectrum.

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How can we paramterize the stress in WBL?

In current models, the wind speed at 10 m,

$$S_{log}(z) = \frac{u_*}{\kappa}\left[ln\frac{z}{z_0} - \psi_m(z/L)\right].$$

After considering the wave-induced stress:

$$U_{10} = S_{log}(10)cos(\theta_{wind}) - S_{sw}(10)cos(\theta_{sw})$$

$$V_{10} = S_{log}(10)sin(\theta_{wind}) - S_{sw}(10)sin(\theta_{sw}),$$

in which θ_{wind} and θ_{sw} are the wind direction and the direction of the swell-induced stress. and

$$\boldsymbol{S}_{sw}(z) = \frac{\boldsymbol{\tau}_{sw}(0)}{\rho_a \kappa u_*}\int_{\infty}^{z} e^{-Ak_p z}/z dz. \quad (1)$$

WRF-SCM simulations

Simulations using the single-column version of the Weather Research and Forecasting Model (WRF-SCM)

- MYNN-2.5 turbulence closure scheme
- 100 vertical layers in the bottom 6 km of the atmosphere.
- 5 m in the lowest model level and 16 layers are in the lowest 200 m
- The geostrophic wind is $U_g = 5 \ ms^{-1}$ and $V_g = 0 \ ms^{-1}$.
- The initial conditions of the wind and temperature are from the Large-Eddy simulation (ZN1) in Nilsson et al., 2012.
- The friction velocity is 0.11 ms^{-1} in all simulations.

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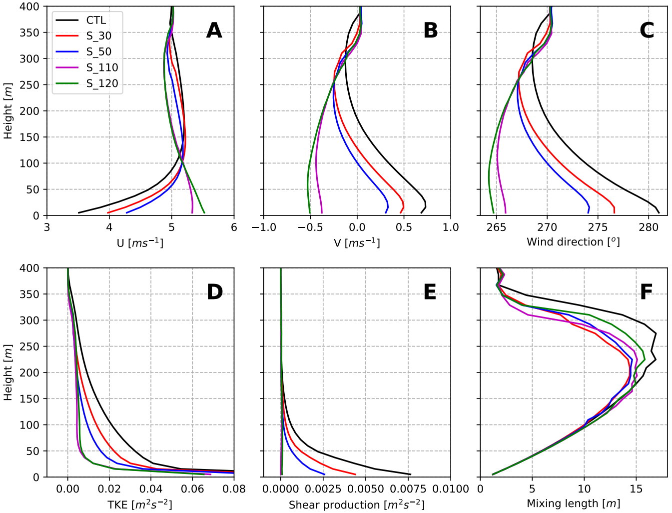
Impact of τ_{sw}:



- $\tau_{sw}(0)$ aligns with wind.
- xx in name *S_xx* represents $\tau_{sw}(0)/\tau_l$
  τ_l is the downward momentum flux
- Neutral condition

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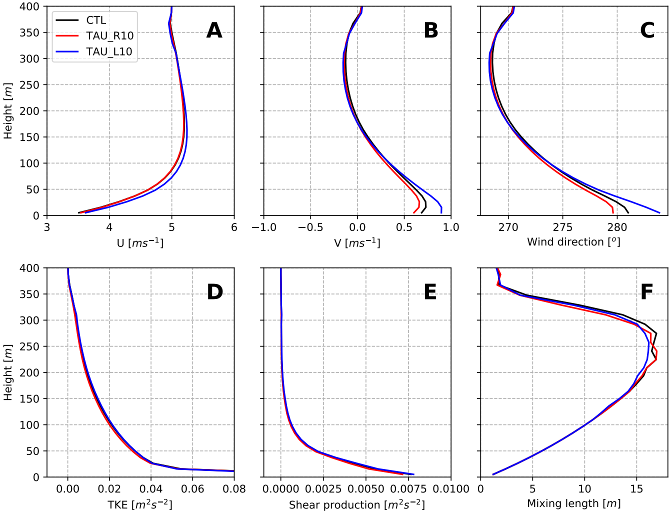
Impact of the direction difference between τ_l and S_{10}:



- $\tau_{sw}(0) = 0$
- The direction difference between τ_l and S_{10} is 10^o on the right and left in R_10 and L_10, respectively
- Neutral condition

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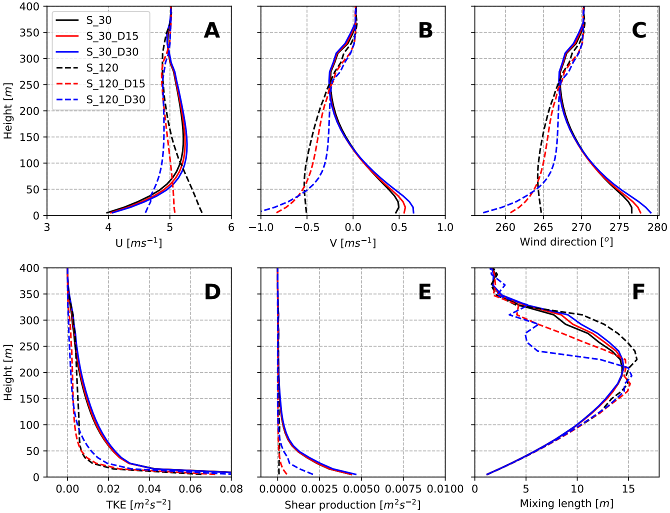
Impact of the direction difference between τ_{sw} and S_{10}:



xx and yy in the name *S_xx_Dyy*

- xx represent τ_{sw}/τ_l
- yy represent the direction difference between τ_{sw} and τ_l

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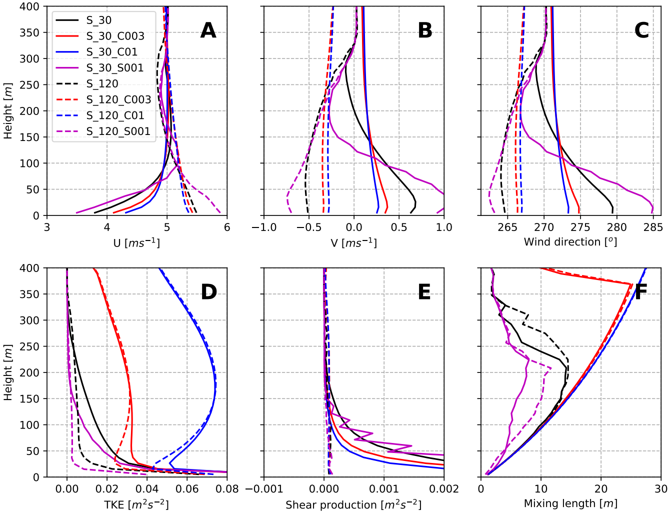
Impact of the atmospheric stability:



xx and yy in the name *S_xx_Cyy* or *S_xx_Syy*

- xx represent τ_{sw}/τ_l
- yy represent the heat flux in convective (C) or stable (S) conditions

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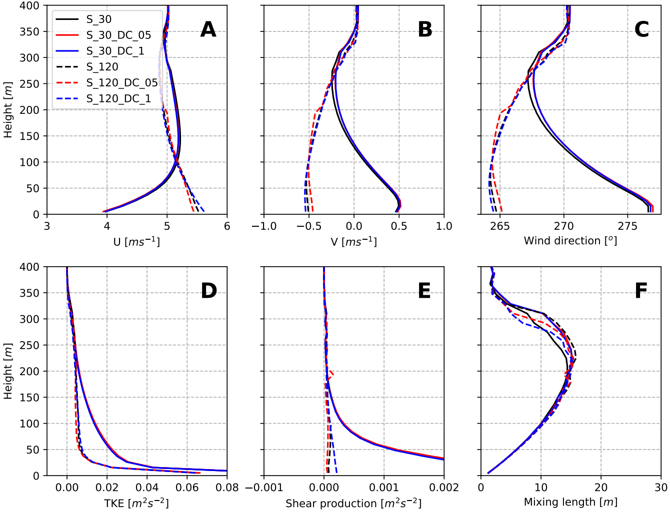
Impact of the decay coefficient of swell-induced upward momentum flux:



xx and yy in the name *S_xx_DCyy*

- xx represent τ_{sw}/τ_l
- yy represent the decay coefficient A in $\tau_{\mathrm{sw}}(z) = \tau_{\mathrm{sw}}(0)e^{-Ak_pz}$

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Real case simulations

One-month long simulations (Jan., 2017) are done over the ocean off California using an atmosphere-wave coupled model (UU-CM).

Control experiment:

$$\boldsymbol{\tau}_a(0) = \rho_a C_d |\boldsymbol{S}_{10}| \boldsymbol{S}_{10},$$

$$z_0 = \alpha \frac{u_*^2}{g} + \frac{0.11 \times \nu_a}{max(u_*, 0.05)},$$

$$\alpha = \frac{0.0095}{\sqrt{1 - \tau_{ww}(0)/\tau_l}}.$$

$$S_{log}(z) = \frac{u_*}{\kappa}\left[ln\frac{z}{z_0} - \psi_m(z/L)\right].$$

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Real case simulations

One-month long simulations (Jan., 2017) are done over the ocean off California using an atmosphere-wave coupled model (UU-CM).

Coupled experiment:

- At air-sea interface ($z \approx 0$)

$$\boldsymbol{\tau}_{\mathrm{a}}(0) = \boldsymbol{\tau}_{\mathrm{v}}(0) + \boldsymbol{\tau}_{\mathrm{ww}}(0) + \boldsymbol{\tau}_{\mathrm{sw}}(0)$$

- In the wave boundary layer ($z > 0$):

$$\boldsymbol{\tau}_{\mathrm{a}}(z) = \boldsymbol{\tau}_{\mathrm{t}}(z) + \boldsymbol{\tau}_{\mathrm{sw}}(z)$$

- The wind speed at 10 m:

$$U_{10} = S_{log}(10)cos(\theta_{wind}) - S_{sw}(10)cos(\theta_{sw})$$

$$V_{10} = S_{log}(10)sin(\theta_{wind}) - S_{sw}(10)sin(\theta_{sw}),$$

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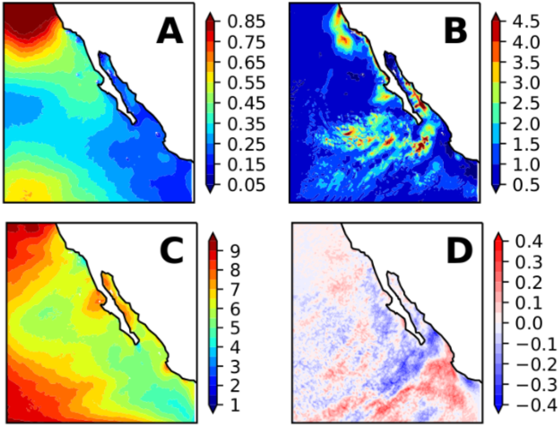
Real case simulations



(A): the mean τ_{sw} $[\times 10^{-3} Nm^{-2}]$, (B) the ratio of the mean τ_{sw} and τ_l [%]
(C) the mean S_{10} $[ms^{-1}]$, (D) the difference of S_{10} between FUL and CTL $[ms^{-1}]$

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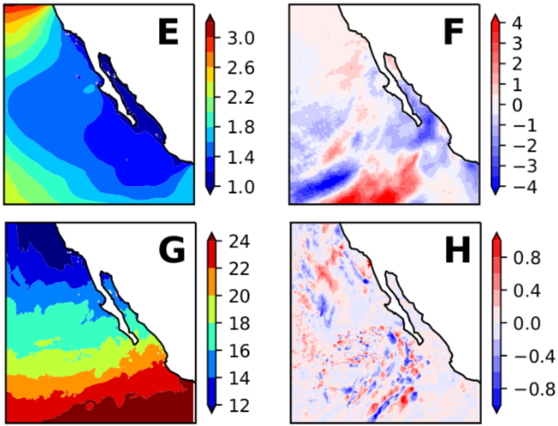
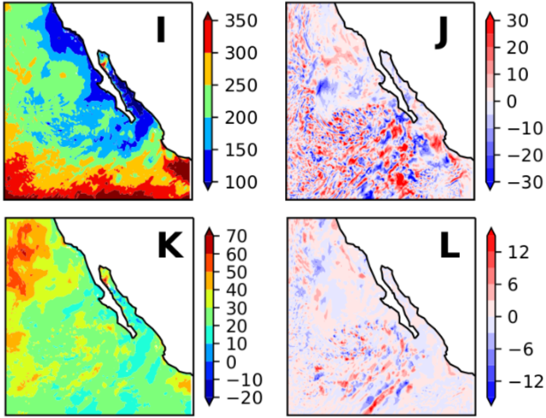
Real case simulations



(E) the mean H_s [m], (F) the difference of H_s between FUL and CTL [cm]
(G) the mean T_2 [°C], (H) the mean difference of T_2 between FUL and CTL [°C]

Real case simulations



(I) the mean latent heat flux [Wm^{-2}], (J) the difference between FUL and CTL [Wm^{-2}]
(K) the mean sensible heat flux [Wm^{-2}], (L) the difference between FUL and CTL [Wm^{-2}].

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Conclusions

- The swell-induced upward momentum flux increases the surface wind and changes the wind direction
- The misalignment between the upward momentum flux and wind has a more significant impact on the wind profile than that from the downward momentum flux
- The impact of swell-induced upward momentum flux decreases with atmospheric convection.
- The surface wind can be altered up to 5% by ocean surface gravity waves over the ocean off California.