3D.3

EXCHANGE COEFFICIENTS AND AIR-SEA INTERACTION IN TROPICAL CYCLONES

K. J. E. Walsh[1], G. B. Brassington[2] and P. Sandery[2]
[1]School of Earth Sciences, University of Melbourne
[2]CAWCR, Bureau of Meteorology, Melbourne

1. INTRODUCTION

Surface fluxes are a fundamental source of energy for tropical cyclones (e.g. Emanuel, 1986). Knowledge of the magnitude and variation of these fluxes is limited by observational constraints, as flux measurements are difficult to achieve at high wind speeds. Nor is it entirely clear what form the exchange coefficients governing the fluxes should take during high winds.

For the drag coefficient, there is considerable evidence that it reaches a maximum at hurricane wind speeds (30-40 ms^{-1}) and either levels out or decreases thereafter (Frank 1985; Emanuel 1995; Powell et al. 2003; Andreas 2004; Donelan et al. 2004; Moon et al. 2004; Makin, 2005; Bye and Jenkins 2006; Jarosz et al. 2007; Moon et al. 2007; French et al. 2007; Drennan et al. 2007).

For the exchange coefficient, there are few measurements at high wind speeds, leaving considerable uncertainty regarding the functional form of this parameter. In addition, the effects of sea spray on the total heat flux become significant at higher wind speeds (Fairall et al. 1994; Andreas and DeCosmo 1999; Wang et al. 2001; Andreas and Emanuel 2001; Emanuel 2003). Here we investigate possible formulations of the exchange coefficient based on observations and theoretical assumptions of the relationship between drag and exchange coefficients.

2. METHODS

The results of numerical simulations of ocean surface cooling are compared with satellite observations of sea surface temperatures (SSTs). Various parameterizations of the drag and exchange coefficients are tested to determine whether the resulting simulated cooling of the sea surface is realistic, in an attempt to constrain the possible values of the coefficients.

The functional form of the assumed drag coefficient C_D is shown in Fig. 1. A linear variation of drag was assumed at low wind speeds. At speeds between 20 and 45 ms^{-1}, C_D is assumed to vary following measurements by Jarosz et al. (2007). For values above 45 ms^{-1}, the drag coefficient is assumed to asymptote to a constant coefficient of $1x10^{-3}$ at winds of 80 ms^{-1}. There is only some theoretical guidance for the choice of this parameter; this value was chosen to be consistent with the results of Fairall et al. (2008). In addition, if the mechanism for the decrease of drag coefficient with wind speed above 40 ms^{-1} is flow separation (Donelan et al. 2004), then it is likely that a limiting value of C_D will be reached at higher wind speeds, similar to the behavior shown in Fig. 1.

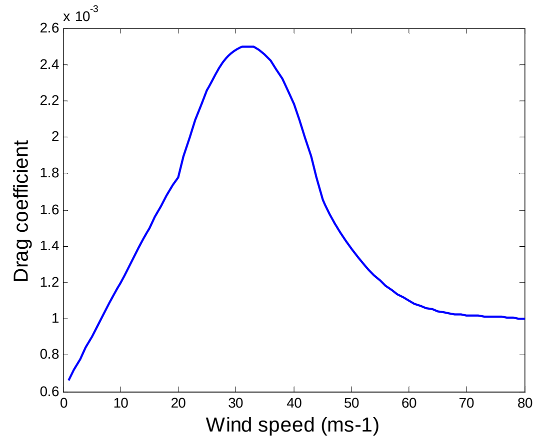


Figure 1. Assumed form of drag coefficient versus 10m wind speed, as described in text.

The relationship between the drag coefficient C_D and the surface flux exchange coefficient C_k is assumed to follow that of the simplified tropical cyclone model of Emanuel (1995) for the high wind regime ($u_{10m} >$ 45 ms^{-1}). For wind speeds less than this, C_k is assumed to be constant at C_k=0.0012, consistent with the available observations at wind speeds up to 30 ms^{-1} (Black et al. 2007).

The resulting ratio C_k/C_D is shown in Fig. 2. There is a small mismatch between values in the different wind regimes. The estimated ratio shown in Fig. 2 is consistent with independent estimates of this quantity derived from laboratory tank results (Fairall et al. 2008). The rise in the ratio for wind speeds above 40 ms^{-1} is attributed by Fairall et al. (2008) to the production of spray and the resulting increase in C_k, combined with a decrease or leveling out of C_D. Note that Emanuel (1995) contends that the ratio would be expected to rise in the high wind regime, as development of a tropical cyclone could only occur if the amount of heat flux energy being supplied to the storm exceeded the amount being removed by drag.

Winds are specified from the wind field model of Holland (1980), with cyclone parameters, including radius of maximum wind, taken from data produced by the Joint Typhoon Warning Center. Initial ocean profiles were derived from the BRAN 2.1 ocean analysis (Schiller et al. 2008; Oke et al. 2008). A number of different representations of surface fluxes were tested,

* *Corresponding author address*: Kevin J.E. Walsh, School of Earth Sciences, University of Melbourne. E-mail: kevin.walsh@unimelb.edu.au

including bulk fluxes alone, and bulk fluxes modified by the spray parameterizations of Fairall et al. (1994) and Andreas and DeCosmo (1999).

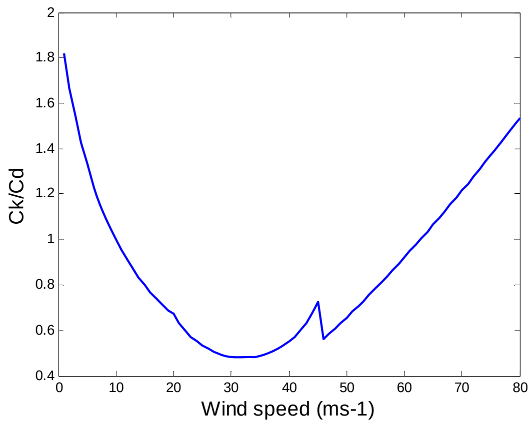


Figure 2. Derived ratio of exchange coefficient to drag coefficient (C_k/C_D) as a function of 10m wind speed.

The ocean model employed is CLAM (Coupled Limited Area Model), a regional configuration of the BLUElink global model (Oke et al. 2008), which is run in uncoupled mode from a reanalysed initial state from BRAN2.1. Features and parameters in the 3-D model are set to be as close as possible to the 1-D model so that differences in results can be mainly attributed to advection and horizontal mixing. The main difference between the two models is that the 3-D model uses 10 m vertical resolution in the upper 200 m, yet resolves depths to greater than 5 km, while the 1-D model has 2 m vertical resolution to a fixed depth of 200 m. Both 3-D and 1-D simulations of the model are performed, in order to examine the relative role of advective processes in the cooling of the sea surface caused by the passage of a tropical cyclone. The mixed layer formulation of the model is based on Chen et al. (1994). This model combines a bulk mixed layer formulation similar to that of Niiler and Kraus (1977) with a variant of the vertical shear-produced mixing model of Price et al. (1986). Thus both wind-driven deepening of the mixed layer and mixing due to vertical current shears are included.

Results were produced for two intense tropical cyclones, both traveling over the Coral Sea east of Australia. Severe Tropical Cyclone Zoe (2002; Figure 3) reached a maximum intensity of 155 knots, while Severe Tropical Cyclone Ingrid (2005; Figure 4) had a maximum intensity of 130 knots in this region. Both storms thus constitute a severe test of any surface flux parameterization.

3. RESULTS

3.1 *Cyclone Zoe*

Figure 5a shows objectively-analysed daily minimum SST, derived from TMI and AMSRE satellite data by Gentemann et al. (2003), for a day and time close to that of the maximum observed cooling. Figure 5b shows results for a 1-D model run using bulk fluxes; here, advection is turned off in the model and all grid points run independently. Figure 6 shows time series comparing the observations to model runs with bulk and the Andreas and DeCosmo (1999) spray flux parameterization and to the BRAN 2.1 analysis, at a point close to the location of maximum observed cooling. The results show that the 1-D model cools in response to surface flux and wind-driven mixing, but recovers too rapidly compared with the observations and the 3-D BRAN 2.1 analysis. Note that the BRAN analysis is a 10 m depth average and it is forced by fluxes from ECMWF analyses, which do not fully resolve tropical cyclone winds.

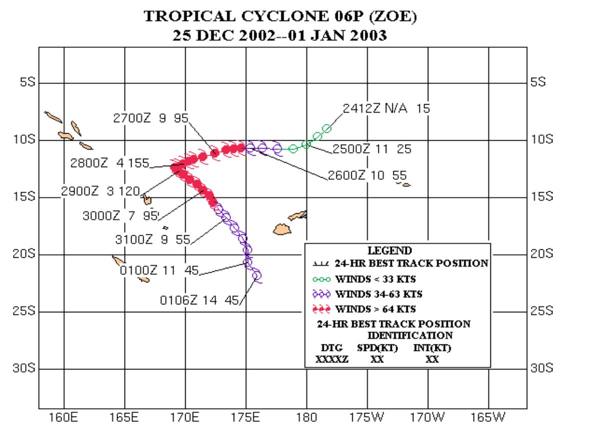


Figure 3. Track and intensity of Severe Tropical Cyclone Zoe. From the Joint Typhoon Warning Center.

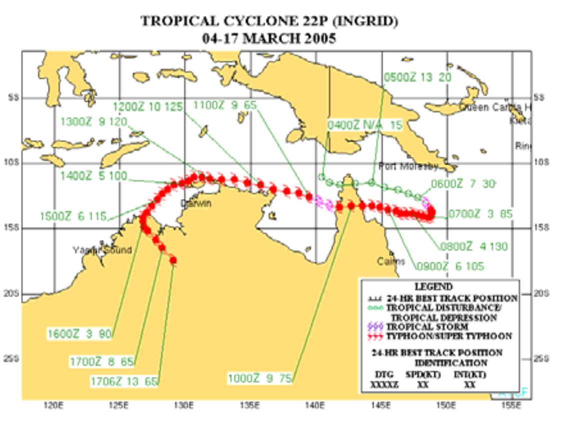
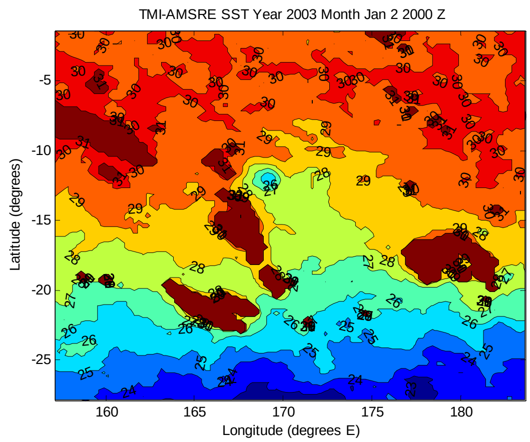


Figure 4. The same as Fig. 3 for Severe Tropical Cyclone Ingrid.

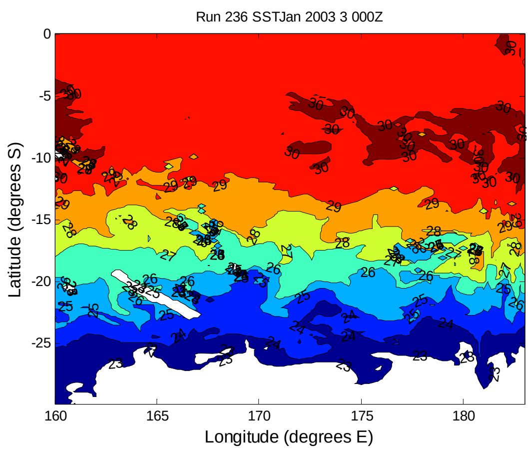
The SST response of the 1-D model is substantially affected by the heat fluxes. This explains why the decrease in SST is so much greater for the run including spray evaporation (Fig. 6, green points) than for the run including only bulk fluxes (blue points), despite the fact that they have the same wind stresses.

This is quite different from the expected 3-D response, as previous work has shown that in hurricanes the vertical current shear dominates the surface cooling processes (e.g. D'Asaro 2003).

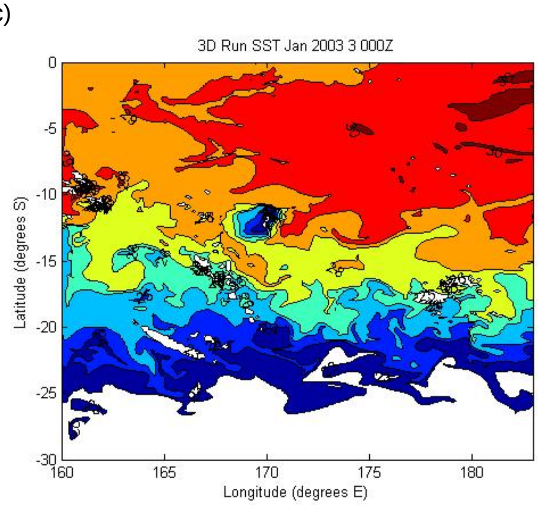
(a)



(b)



(c)



(d)

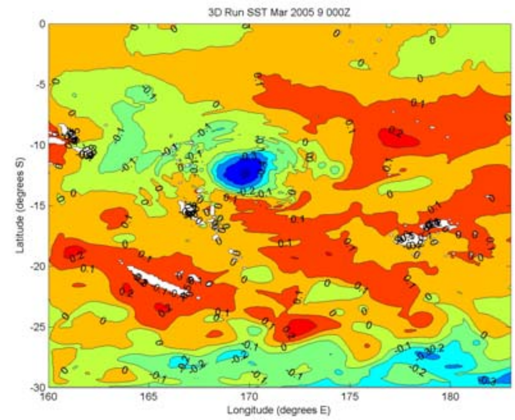


Figure 5. (a) Observed daily minimum SST (at approximately 8 am local time, Jan 2 2000Z) from objectively analysed blended TMI and AMSRE observations; (b) simulated 1-D model results using bulk fluxes; (c) 3-D model simulation of SST corresponding to same time and surface fluxes as 1D simulation; and (d) sea surface height anomaly (m) for the same time as Fig. 7(c).

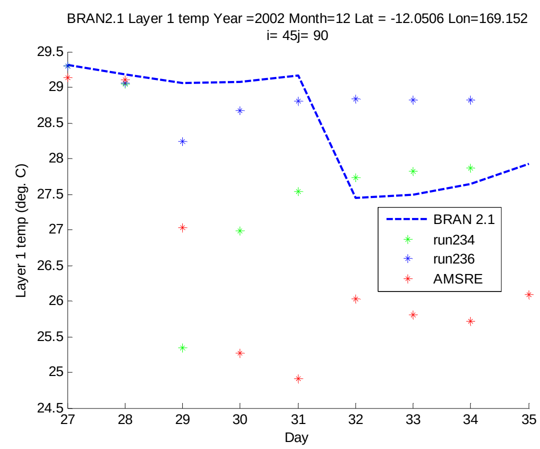


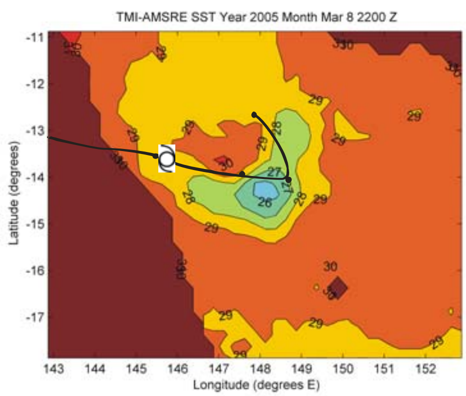
Figure 6. Comparison between SST observations (red), and 1-D model simulations using bulk fluxes (blue), Andreas spray fluxes (green), and BRAN 2.1 analysis (dashed line), for simulation commencing 0000Z Dec 27 2002, at a point close to the location of maximum observed cooling.

Results from the 3-D simulation (Figure 5c) show good agreement with TMI-AMSRE observations in terms of the location of the storm induced cool SST area, but this area is larger in extent and ~2^{o}C cooler than observations on the left side of the storm track, due to resonance excitation (e.g. Shay et al. 1990). SSTs on the right side of the track are in close agreement with observations. Figure 5d shows the corresponding sea level anomaly. The anomalies are in generally good agreement with available observations (not shown).

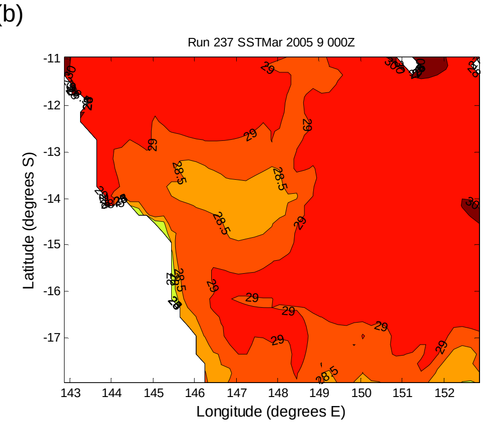
4.2 *Cyclone Ingrid*

Figure 7a shows SST observations for Cyclone Ingrid, while Figure 7b shows 1-D model simulations. In common with the simulations for Zoe, the cooling in the 1-D model is earlier than the cooling in the observations (Fig. 8). In addition, the simulated cooling tends to lie over the cyclone track compared with the observed cooling, which is to the left of the track, as is typical in the Southern Hemisphere.

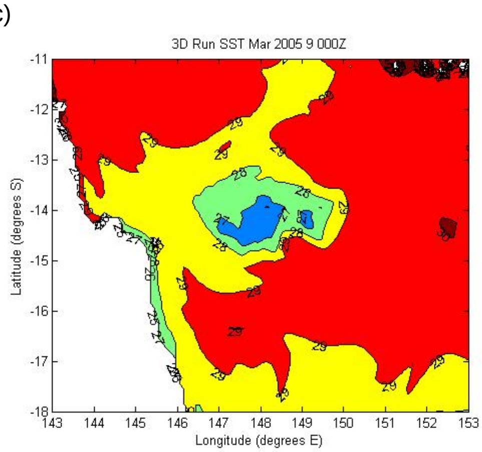
(a)



(b)



(c)



(d)

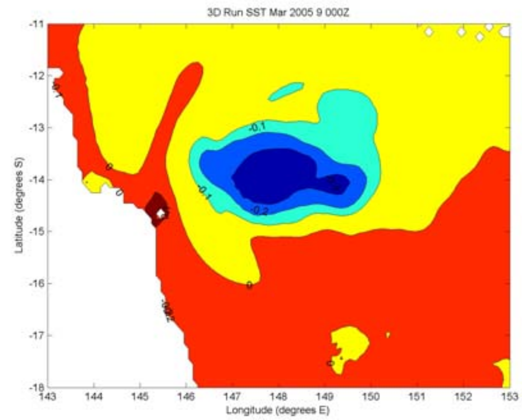


Figure 7. (a) TMI-AMSRE SST at Mar 8 2200Z, with Cyclone Ingrid track indicated along with position at that time; (b) 1-D model simulation of SST at Mar 9 0000Z using bulk fluxes; (c) 3-D model simulation of SST corresponding to same time and surface fluxes as 1-D simulation; and (d) sea surface height anomaly (m) at the same time as Fig. 7(c).

The 3-D simulation shows considerably improved agreement over the 1-D model with respect to TMI-AMSRE observations in terms of the magnitude and location of the storm-induced cool SSTs (Figure 7c). The 3-D simulation has cooler water on the left side of the storm track, due to resonance excitation which cannot be resolved by the 1-D model. The sea surface height anomalies (Fig. 7d) are roughly in agreement with the available observations, although these are less detailed than similar observations for Zoe.

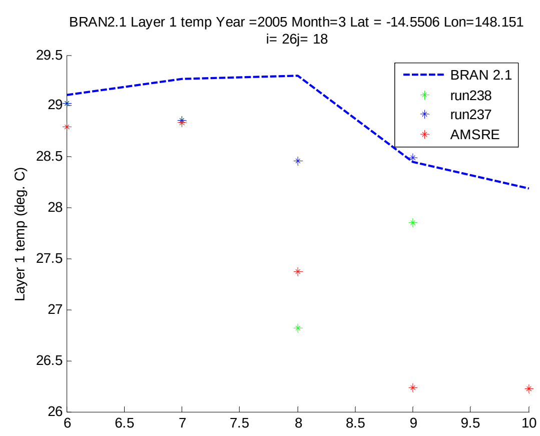


Figure 8. The same as Figure 6 but for a location close to the minimum observed temperature for Cyclone Ingrid, for a simulation commencing at 0000Z March 6 2005.

4. CONCLUSION

It is known that advective processes are exceedingly important in determining the SST anomalies generated by tropical cyclones. The results shown here reinforce this conclusion. Realistic results of surface cooling are achieved using a 3-dimensional model combined with a representation of surface drag appropriate to very high wind speeds. Further work will involve testing other hypothesized representations of the drag coefficient, as well as further simulating the effect of spray evaporation on the SST simulation.

REFERENCES

Andreas, E.L., 2004: Spray stress revisited. *Mon. Wea. Rev*., **34**, 1429-1440.

Andreas, E.L., and J. DeCosmo, 1999: Sea spray production and influence on air–sea heat and moisture fluxes over open ocean. *Air–Sea Exchange: Physics, Chemistry and Dynamics*, Edited by G. L. Geernaert, Kluwer, 327–362.

Andreas, E.L., and K.A. Emanuel, 2001: Effects of spray on tropical cyclone intensity. *J. Atmos. Sci.*, **58**, 3741–3751.

Black, P.G., E. A. D'Asaro, W. M. Drennan, J.R. French, P.P. Niiler, T.B. Sanford, E.J. Terrill, E.J. Walsh, and J.A. Zhang, 2007: Air-sea exchange in hurricanes. *Bull. Amer. Meteorol. Soc.*, **88**, 357-374.

Bye, J.A.T., and A.D. Jenkins, 2006: Drag coefficient reduction at very high wind speeds. *J. Geophys. Res.*, **111**, C03024, doi: 10.1029/2005JC003114.

Chen, D., L. Rothstein, A.J. Busalacchi, 1994: A hybrid vertical mixing scheme and its application to tropical ocean models. *J. Phys. Oceanogr.*, **24**, 2156-2179.

D'Asaro, E.A., 2003: The ocean boundary layer below Hurricane Dennis. *J. Phys. Oceanogr.*, **33**, 561-579.

Donelan, M.A., B.K. Haus, N. Reul, W.J. Plant, M. Stiassnie, H.C. Graber, O.B. Brown, E.S. Saltzman, 2004: On the limiting aerodynamic roughness of the ocean in very strong winds. *Geophys. Res. Letters*, **31**, L18306, doi:10.1029/2004GL019460.

Drennan, W.M., J. Zhang, J. R. French, C. McCormick, P. G. Black, 2007: Turbulent fluxes in the hurricane boundary layer. Part II: Latent heat flux. *J. Atmos. Sci.*, **64**, 1103–1115.

Emanuel, K.A., 1995: Sensitivity of tropical cyclones to surface exchange coefficients and a revised steady-state model incorporating eye dynamics. *J. Atmos. Sci.*, **52**, 3969-3976.

Emanuel, K.A., 2003: A similarity hypothesis for air-sea exchange at extreme wind speeds. *J. Atmos. Sci*., **60**, 1420-1428.

Fairall, C.W., J.D. Kepert, and G.J. Holland, 1994: The effect of sea spray on surface energy transports over the ocean. *Global Atmos. Ocean Syst.*, **2**, 121–142.

Fairall, C. W., W. Asher, M. Banner, and W. Peirson, 2008: Investigation of the physical scaling of sea spray spume droplet production. *J. Geophys., Res.*, in preparation.

Frank, W.M., 1985: Analysis of the inflow layer and air–sea interactions in Hurricane Frederic (1979). NASA-CR-175616, 8 pp. [Available from National Technical Information Service, 5285 Port Royal Rd., Springfield, VA 22151].

French, J.R., W.M. Drennan, J.A. Zhang, and P.G. Black, 2007: Turbulent fluxes in the boundary layer. Part I: Momentum flux. *J. Atmos. Sci.*, **64**, 1089-1102.

Gentemann, C., C.J. Donlon, A. Stuart-Menteth, and F.J. Wentz, 2003: Diurnal signals in satellite sea surface temperature measurements. *Geophys. Res. Lett.*, **30**, 1140-1143.

Holland, G.J., 1980: An analytic model of the wind and pressure profiles in hurricanes. *Mon. Wea. Rev.*, **108**, 1212-1218.

Jarosz, E., D.A. Mitchell, D.W. Wang, and W.J. Teague, 2007: Bottom-up determination of air-sea momentum exchange under a major tropical cyclone. *Science*, **315**, 1707-1709.

Makin, V.K., 2005: A note on the drag of the sea surface at hurricane winds. *Bound. Layer Meteorol.*, **115**, 169-176.

Moon, I.-J., I. Ginis, and T. Hara, 2004: Effect of surface waves on Charnock coefficient under tropical cyclones. *Geophys. Res. Letters*, **31**, L20302, doi:10.1029/2004GL020988.

Moon, I.-L., I. Ginis, T. Hara, and B.Thomas, 2007: A physics-based parameterization of air–sea momentum flux at high wind speeds and its impact on hurricane intensity predictions. *Mon. Wea. Rev.*, **135**, 2869-2878.

Niiler, P.P., and E.B. Kraus, 1977: One-dimensional models of the upper ocean. *Modeling and Prediction of the Upper Layers of the Ocean*. Edited by E.B. Kraus, Pergamon Press, 323 pp.

Oke, P. R., G. B. Brassington, D. A. Griffin and A. Schiller, 2008: The Bluelink Ocean Data Assimilation System (BODAS). *Ocean Modelling*, **20**, 46-70, doi:10.1016/j.ocemod.2007.11.002.

Powell, M.D., P.J. Vickery, and T.A. Reinhold, 2003: Reduced drag coefficient for high wind speeds in tropical cyclones. *Nature*, **422**, 279-283.

Price, J.F., R.A. Weller, R. Pinkel, 1986: Diurnal cycling: Observations and models of the upper ocean response to diurnal heating, cooling and wind mixing. *J. Geophys. Res*., **91**, 8411-8427.

Schiller, A., P. R. Oke, G. B. Brassington, M. Entel, R. Fiedler, D. A. Griffin, J. Mansbridge, 2008: Eddy-resolving ocean circulation in the Asian-Australian region inferred from an ocean reanalysis effort, *Progress in Oceanography*, **76**, 334-365.

Shay, L. K., S. W. Chang and R. L. Elsberry, 1990: Free surface effects on the near inertial ocean current response to a hurricane. *J. Phys. Oceanogr*., **20**, 1405-1424.

Wang, Y., J.D. Kepert, and G.J. Holland, 2001: The effect of sea spray evaporation on tropical cyclone boundary layer structure and intensity. *Mon. Wea. Rev*., **129**, 2481-2500.