Deep-Sea Research I 94 (2014) 107–123

Contents lists available at ScienceDirect

Deep-Sea Research I

journal homepage: www.elsevier.com/locate/dsri

Influence of surface forcing on near-surface and mixing layer turbulence in the tropical Indian Ocean

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ARTICLE INFO

Article history:
Received 11 November 2013
Received in revised form
14 July 2014
Accepted 19 August 2014
Available online 6 September 2014

Keywords:
Dissipation of turbulent kinetic energy
Near-surface turbulence
Remnant layer
Dissipation decay time
Air-Sea Interaction Profiler (ASIP)
Diurnal cycle
Wind-driven mixing
Convectively-driven mixing

ABSTRACT

An autonomous upwardly-moving microstructure profiler was used to collect measurements of the rate of dissipation of turbulent kinetic energy (ε) in the tropical Indian Ocean during a single diurnal cycle, from about 50 m depth to the sea surface. This dataset is one of only a few to resolve upper ocean ε over a diurnal cycle from below the active mixing layer up to the air–sea interface. Wind speed was weak with an average value of $\sim$5 m s^{-1} and the wave field was swell-dominated. Within the wind and wave affected surface layer (WWSL), ε values were on the order of 10^{-7}–10^{-6} W kg^{-1} at a depth of 0.75 m and when averaged, were almost a factor of two above classical law of the wall theory, possibly indicative of an additional source of energy from the wave field. Below this depth, ε values were closer to wall layer scaling, suggesting that the work of the Reynolds stress on the wind-induced vertical shear was the major source of turbulence within this layer. No evidence of persistent elevated near-surface ε characteristic of wave-breaking conditions was found. Profiles collected during night-time displayed relatively constant ε values at depths between the WWSL and the base of the mixing layer, characteristic of mixing by convective overturning. Within the remnant layer, depth-averaged values of ε started decaying exponentially with an e-folding time of 47 min, about 30 min after the reversal of the total surface net heat flux from oceanic loss to gain.

1. Introduction

The flux of energy, momentum, and heat across the air–sea interface sets the vertical structure of the ocean by mixing the water column on scales from micrometres to hundreds of metres (Lazier, 1980; Peirson and Banner, 2004). The principal action[2] of the wind stress drives net downward transfer of momentum across the air–sea interface. This drives turbulence via the generation of a mean current shear (and resulting Kelvin–Helmholtz instabilities), and additionally through wave-field related processes such as the generation of the Stokes drift vertical shear and wave breaking (at large enough wind speeds). These mechanisms help convert mean kinetic energy (MKE) and wave kinetic energy (WKE) into turbulent kinetic energy (TKE) which partly acts to mix the upper ocean (conversion of TKE to potential energy), and is partly dissipated by viscous forces into heat. Buoyancy fluxes, on the other hand, can either lead to convectively-driven turbulence when the ocean loses buoyancy (via heat loss and/or an increase in salt concentration via freshwater evaporation), or it can promote stable stratification which suppresses turbulence during oceanic buoyancy gain (via heat gain or a decrease in salt concentration via freshwater input).

The upper ocean boundary layer (OBL) is that layer which is directly affected by surface forcing. Understanding of the dynamics of turbulence within the OBL relies, in part, on the ability to adequately observe its spatial and temporal evolution as it is being forced by the wind and waves together with heat and mass fluxes. This requires measurements that extend in space from just below the mixing layer right up to the ocean surface, and in time so as to capture transient wind events and systematic diurnal changes in the surface buoyancy fluxes. The mixing layer can be defined as the depth-zone in which turbulence is directly and actively driven by surface forcing (Brainerd and Gregg, 1995; Stevens et al., 2011). Field campaigns measuring turbulence and temperature microstructure using vertically rising microstructure profilers offer the possibility of observing the OBL on these spatial and temporal scales.

Microstructure profilers have been used in the past to observe the OBL over one or more diurnal cycles (*e.g.*, Gregg et al., 1985; Shay and Gregg, 1986; Lombardo and Gregg, 1989, Brainerd and

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[1] Now at Scripps Institution of Oceanography.
[2] In conditions where swell waves travel in the same direction but more quickly than the overlying wind, there may be a net transfer of momentum upwards from the waves to the air (Donelan and Dobson, 2001).

http://dx.doi.org/10.1016/j.dsr.2014.08.009
0967-0637/

Gregg 1993a,b; Caldwell et al., 1997). Studies like these have helped form the basis for much of our knowledge about diurnal variability of turbulence in the upper ocean. However, several previous studies had to discard the turbulence measurements within the upper 5–10 m due to contamination from the ship wake, and by using microstructure profilers in a descending mode (*e.g.*, Oakey and Elliott, 1982; Lombardo and Gregg, 1989; Brainerd and Gregg, 1993a). It is becoming increasingly clear from *in situ* measurements that wave breaking and wave-turbulence interaction also play significant roles in determining the magnitude and distribution of turbulence at depths from the free surface to O (10 m) (*e.g.*, Agrawal et al., 1992; Gemmrich and Farmer, 2004; Gemmrich, 2010). Vertically rising microstructure profilers offer the possibility to obtain snapshots of oceanic processes within this highly turbulent near-surface region that cannot be resolved by descending profilers.

The Air-Sea Interaction Profiler (ASIP) used in this study is described in Ward and Fristedt (2008), and the latest version of the profiler is described in Ward et al. (2014). It is a fully autonomous vertically-rising oceanographic microstructure profiling instrument capable of measuring right up to the air–sea interface in the oceanic mixed layer. Here we report on measurements of the rate of dissipation of turbulent kinetic energy, ε, from 95 profiles in the upper 45 m of the Indian Ocean layer during a single diurnal cycle acquired by ASIP. This dataset is one of only a few datasets that have resolved the upper ocean at depths from below the active mixing layer to the air–sea interface. The evolution of turbulence in the upper 45 m of the water column during a single 13-h period is examined and the turbulence within the mixing layer is scaled in terms of shear production by the wind and surface buoyancy fluxes. The decay timescale of convectively generated turbulence in the remnant layer (defined below) is also quantified.

The layout of the paper is as follows: Section 2 presents a conceptual view of the OBL, Section 3 describes the study area, data collection and processing, the results are presented in Section 4 which is followed by a discussion in Section 5 and the concluding remarks are given in Section 6.

2. Conceptual view of the OBL

The general evolution of ε within the upper ocean mixing layer over a diurnal cycle is fairly well understood and is depicted schematically in Fig. 1. The energetics within the mixing layer can be quantified with the turbulent kinetic energy (TKE) equation. Following Belcher et al. (2012) and assuming horizontally homogeneous flow, the z-component of the TKE equation may be written as follows:

$$\underbrace{\frac{Dq^2}{Dt}}_{(a)} = -\underbrace{\overline{u'w'}\frac{\partial \overline{U}}{\partial z}}_{(b)} - \underbrace{\overline{u'w'}\frac{\partial \overline{U_s}}{\partial z}}_{(c)} + \underbrace{\overline{w'b'}}_{(d)} - \underbrace{\frac{\partial}{\partial z}\left[\overline{w'\left(q^2+\frac{p'}{\rho_o}\right)}\right]}_{(e)} - \underbrace{\varepsilon}_{(f)} \qquad (1)$$

where overbars represent averaged quantities, primed quantities represent fluctuations from mean quantities, q^2 is the TKE, ρ_o is the density of seawater, p is the pressure, $\overline{U}$ is the mean horizontal current, $\overline{U_s}$ is the mean Stokes drift and u' and w' are the horizontal and vertical velocity fluctuations respectively. The LHS of Eq. (1) quantifies the rate of change of TKE in response to the forcing terms on the RHS. Terms (b) and (c) represent the Reynolds stress ($u'w'$) working against the mean current shear ($\partial\overline{U}/\partial z$) and the mean Stokes drift shear ($\partial\overline{U_s}/\partial z$) respectively, with the latter often denoted as Langmuir turbulence (Belcher et al., 2012). These two terms are always a source of TKE as they convert mean kinetic energy (MKE) and wave kinetic energy (WKE) to TKE. Term (d) is the work done by buoyancy fluctuations ($b' = -g\rho'/\rho_o$) against the background stratification and this can be a source or sink of TKE depending on the density gradient. Term (e) represents the vertical flux of TKE via turbulence–turbulence interactions and the last term represents the conversion of TKE to heat *via* viscous forces and is termed the rate of dissipation of TKE. However, our understanding of the detailed physical processes affecting the specific magnitude and vertical distribution of ε within the layer, where wind and wave forcing (from both breaking and non-breaking waves) are expected to dominate the production of turbulent kinetic energy, is not yet complete (Belcher et al., 2012).

2.1. General description of upper ocean evolution during a diurnal cycle

During night-time cooling the ocean loses heat to the atmosphere (defined here as a positive heat flux in the direction from the ocean to the atmosphere, (*e.g.*, Lombardo and Gregg, 1989)) and potential energy is converted to turbulent kinetic energy [term (d) in Eq. (1) is positive resulting in an increase of TKE]. This drives convective overturns that result in the deepening of the diurnal thermocline entraining more quiescent water upward from below. The diurnal thermocline typically reaches its maximum depth at the transition between night-time cooling and daytime heating, when the surface heat flux reverses sign (Fig. 1a).

With the reversal of the heat flux from cooling to warming, the upper ocean may begin to re-stratify. At this point the convective overturns are no longer actively forced and begin to decay. The layer of decaying turbulence that is capped by the wind and wave affected surface layer is depicted in Fig. 1a and is termed as the remnant layer. Within the remnant layer, the main balance in Eq. (1) is between term (a) and term (f), and the rate of dissipation of turbulent kinetic energy has been found to decay exponentially in time (Brainerd and Gregg, 1993a). Following the night–day transition and accompanying change in heat flux direction, the rate of re-stratification is a function of the surface insolation and the downward turbulent heat flux driven by wind and wave-driven mixing (*e.g.*, Brainerd and Gregg, 1993a,b). The re-stratification process can form a shallow (order several metres), relatively warm and well-defined surface mixing layer. The base of this layer may be marked by a sharp increase in the buoyancy frequency that can effectively isolate the water column below from the effects of surface forcing above.

At the transition from day to night-time the surface heat flux changes from warming to cooling which marks a return to convectively driven overturning. The diurnal thermocline deepens in response to the magnitude of these overturns and the overlying wind stress, but it is also sensitive to the stratification and levels of turbulence of the remnant layer, into which the diurnal thermocline is deepening (*e.g.*, Brainerd and Gregg, 1993a,b). The maximum depth of the diurnal thermocline, or the active mixing layer, can vary on timescales of days depending on the diurnal forcing. The average effect of the daily variability in mixing layer depth over timescales of weeks and months acts to set the depth of the seasonal thermocline (Lombardo and Gregg, 1989).

Langmuir circulation cells can be superposed upon these characteristic features of the ocean boundary layer (OBL) throughout the diurnal cycle (Leibovich, 1983). This circulation pattern can be generally described as a series of counter-rotating vortices with axes that are parallel, or nearly parallel, to the direction of the wind (Langmuir, 1938). The Craik–Lebovich theory of Langmuir circulation generation proposes that interactions of the Stokes drift with the vertical vorticity generated by cross-wind perturbations in the along-wind surface current drive a convergent surface current that points into the vertical plane of the maximum along-wind surface current, giving rise to an associated and co-located