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Contribution to the Themed Section: '*Revisiting Sverdrup's Critical Depth Hypothesis*'

## Food for Thought

# Has Sverdrup's critical depth hypothesis been tested? Mixed layers vs. turbulent layers

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Sverdrup (1953. On conditions for the vernal blooming of phytoplankton. Journal du Conseil International pour l'Exploration de la Mer, 18: 287–295) was quite careful in formulating his critical depth hypothesis, specifying a "thoroughly mixed top layer" with mixing "strong enough to distribute the plankton organisms evenly through the layer". With a few notable exceptions, most subsequent tests of the critical depth hypothesis have ignored those assumptions, using estimates of a hydrographically defined mixed-layer depth as a proxy for the actual turbulence-driven movement of the phytoplankton. However, a closer examination of the sources of turbulence and stratification in turbulent layers shows that active turbulence is highly variable over time scales of hours, vertical scales of metres, and horizontal scales of kilometres. Furthermore, the mixed layer as defined by temperature or density gradients is a poor indicator of the depth or intensity of active turbulence. Without time series of coincident, *in situ* measurements of turbulence and phytoplankton rates, it is not possible to properly test Sverdrup's critical depth hypothesis.

**Keywords:** critical depth, diffusivity, dissipation, mixed layer, mixing, stratification, Sverdrup, turbulence.

## Introduction

### Sverdrup's critical depth hypothesis

More than 60 years ago, Sverdrup formulated his critical depth hypothesis (SCD: Sverdrup, 1953) to explain the well-documented North Atlantic spring phytoplankton bloom. The SCD hypothesis is a simple model based on the premise that winter phytoplankton growth is limited by the amount of available light: strong turbulence in the upper layer of the ocean moves phytoplankton rapidly through the water column. These conditions result in each phytoplankter receiving the average amount of light over the depth of the mixed layer, which—since light decays exponentially with depth—is significantly less light than would be available if the phytoplankters were stationary within the euphotic zone. Sverdrup (1953) postulated that there existed a particular depth—the *critical depth*—at which the vertically integrated phytoplankton growth (assumed proportional to the exponentially decreasing local irradiance) equalled the vertically integrated phytoplankton losses (assumed to have no depth dependence). When the mixed layer is shallower than the critical depth, integrated growth outweighs the losses, and a bloom can occur. This was hypothesized to occur through a spring shoaling of the mixed layer, driven by surface heating.

Since its formulation, this hypothesis has served as a basis for understanding bloom dynamics throughout the world's oceans (e.g. Fischer *et al.*, 2014). Over the decades, many studies have attempted to test the SCD hypothesis with mixed results. Using satellite imagery and mixed-layer climatologies, Obata *et al.* (1996), Siegel *et al.* (2002), Brody *et al.* (2013), Brody and Lozier (2014), and Chiswell *et al.* (2013) found evidence supporting the SCD hypothesis (though that support was qualified in some cases), while Behrenfeld (2010), Boss and Behrenfeld (2010), Behrenfeld and Boss (2014), and Behrenfeld *et al.* (2013) rejected the SCD hypothesis based on the observation that net phytoplankton growth was positive and increasing in the middle of winter.

Before proceeding, it is important to clarify some terms. Technically "mixing" refers to the homogenization of gradients of a property. Once those gradients disappear, there is nothing left to mix, thus "mixing" stops, although the water may still be turbulent. Since we are concerned with the movement of phytoplankton through a light gradient, the strength of turbulent motions of the water is the more relevant quantity, and care will be taken to use the term "turbulence" or "dissipation" (see below) to refer to the turbulent motions of the water, rather than "mixing". A "mixed layer", on the other hand, is hydrographically defined as the region in which the temperature or density difference is less than a given amount (see below).

In formulating his model (both conceptual and mathematical), Sverdrup's first explicit assumption was that "there exists a thoroughly mixed top layer of thickness D". His second assumption was, "Within the top layer the turbulence is strong enough to distribute the plankton organisms evenly through the layer". The turbulent displacement of a particle can be expressed mathematically using an eddy diffusivity, which I will denote $K$ (units: $m^2\ s^{-1}$). In the SCD model, Sverdrup made the following assumptions: (i) the diffusivity was strong enough that over a day, every phytoplankter spent an equal amount of time at every depth in the mixed layer, and (ii) this movement resulted in phytoplankton experiencing the average irradiance within the mixed layer over the course of a day. Interestingly, because of his assumptions, neither the magnitude nor the vertical structure of $K$ appears in Sverdrup's (1953) equations.

The central issue to testing the SCD hypothesis is understanding the meaning of a "thoroughly mixed" layer. In this context, "thoroughly mixed" means that every phytoplankter in the turbulent layer receives the same average amount of irradiance over some time interval: a photoperiod or a growth period, for example. To first order, the average time $\tau_L$ it takes for a particle to move a distance $\pm L$ from its present depth with a local diffusivity $K(z)$ is given by

$$\tau_L = \frac{L}{2K(z)}. \qquad (1)$$

To build our intuition, $\tau_L$ can be thought of as a residence time of a particle in a layer $2L$ thick, centred on the particle's present depth. It is clear from (1) that the residence time is inversely proportional to the diffusivity $K(z)$: larger diffusivities (stronger turbulence) lead to shorter residence times. Thus, a vertical gradient in diffusivity—or turbulence—would also result in a vertical gradient in residence time of phytoplankton with depth, thus affecting the amount of light they are exposed to.

In his paper, Sverdrup (1953) clearly states, "...a phytoplankton population may increase independently of the thickness of the mixed layer *if the turbulence is moderate*" (page 290, my italics). Sverdrup was clearly aware of the distinction between a mixed layer (defined hydrographically by a vertical temperature or density difference) and the intensity of turbulence in that mixed layer. This is the basis of Huisman *et al.*'s (1999) formulation of the "critical turbulence" hypothesis: if $K$ is weak enough, turbulent motions will still occur, but the residence time of the phytoplankton in the euphotic zone will be long enough to cause net positive growth. Townsend *et al.* (1992), Ellertsen (1993), and Chiswell (2011), for example, observed the initiation of shallow spring blooms in apparently unstratified water columns, that is, water columns with deep mixed layers. Chiswell (2011) concluded that the upper mixed layer in which the bloom occurred was not turbulent enough to satisfy the SCD criterion of a "thoroughly mixed layer".

Huisman *et al.* (1999) point out that it is not sufficient to simply define a mixed layer: it is also necessary to specify how vigorously that mixed layer is mixing, i.e. to quantify the strength of the turbulence. This point has been echoed and amplified in subsequent studies (e.g. Ebert *et al.*, 2001; Chiswell, 2011; Taylor and Ferrari, 2011; Brody and Lozier, 2014). Sverdrup (1953) avoided the necessity of including the vertical diffusivity, $K$, by explicitly assuming it was large enough to be ignored. But he clearly understood the ramifications of weaker turbulence—he just chose not to include it in his model. Huisman *et al.* (1999) explicitly include a vertically constant diffusivity $K$ in their model. Brody and Lozier (2014) parameterized vertical overturning time scales for the turbulent layer based on large-scale hydrographic properties and atmospheric forcings; again, they assumed that the turbulent motions were vertically uniform. Using higher-order turbulence closure models, Taylor and Ferrari (2011) and Mahadevan *et al.* (2012) are among the only studies to have employed a vertically varying diffusivity in their analyses of the SCD hypothesis. One of the points I make in this synthesis is that it is crucial to know not only the intensity of the turbulence, but also its vertical structure and temporal variability. That is, to test the SCD hypothesis, we must find the part of the water column that is consistent with the assumptions of the SCD hypothesis: "a thoroughly mixed top layer". This requires measurements of the actual turbulence, rather than the hydrographic results of the mixing: the mixed layer.

As I will show below, vertical gradients of turbulence can, and often do, exist even when the turbulence is strong enough to homogenize hydrographic properties such as temperature and density. Thus, if the "thoroughly mixed" layer is defined by uniform temperature or density, as is commonly the case (see below), and this layer contains a vertical gradient in turbulence, Sverdrup's first assumption is not met and the SCD hypothesis is not properly being tested. Intuitively, then, the details of the vertical and temporal structures of the turbulence are fundamental to our definition of "thoroughly mixed", regardless of the vertical distribution of other hydrographic properties like temperature or density. In other words, the presence of a mixed, homogenous, layer in temperature or density does not imply either that turbulence is ongoing or that turbulence is strong enough to move the phytoplankton completely and rapidly (relative to temporal changes in the light field) through the (hydrographically defined) mixed layer, as required by the SCD hypothesis.

As the debate concerning the validity of the SCD hypothesis continues, this paper explores the physics and the spatial and temporal dynamics of one of the foundations of the SCD hypothesis: turbulence within the mixed layer. I will show that the vertical structure and intensity of turbulence depends strongly on the source of energy that drives the turbulence, and the sources of stratification that can suppress it. I will show that restratification—the formation of vertical density gradients through heating, freshwater fluxes or slumping of horizontal density gradients—can occur on time scales of hours and spatial scales of kilometres. Even extremely weak restratification can inhibit turbulence, and this can occur rapidly (hours). Furthermore, I will show that including this short-time scale variability has important consequences for understanding dynamics in the upper ocean. In particular, I hope to demonstrate that a deeper understanding of the spatial and temporal dynamics of turbulent layers may help to reconcile the various studies supporting and rejecting the SCD hypothesis, and to further the investigation of the mechanisms underlying the spring bloom.

## Mixed layer vs. turbulent layer

Sverdrup (1953) wrote about a "mixed layer", though he was clearly referring to a "turbulent layer"—the waters that are kept in motion through turbulence. This distinction has largely been forgotten, and most subsequent tests of the SCD hypothesis have used measures of the mixed layer, not the turbulent layer. Brainerd and Gregg (1995) may have been the first to formally distinguish a mixed layer from a turbulent layer. As they note, "The distinction is significant, because it is often important to match the mixed layer time scale to that of the process being studied ". As we will see below, the time scales of variability of turbulence and biological time scales ($\sim$1 d) are often well matched when considering phytoplankton photosynthesis and growth.

*Mixed layers* are typically defined operationally as the shallowest depth at which a difference in temperature or density, measured from the surface (or 10 m in some cases), reaches a given threshold (see Kara *et al*., 2000; De Boyer Montégut *et al*., 2004; Lorbacher *et al*., 2006; Holte and Talley, 2009, for summaries of the various thresholds used). These thresholds are often instrument-dependent, with larger thresholds for lower-resolution sensors. A typical temperature threshold is 0.2°C less than the surface value, or a density increase of 0.125 kg m$^{-3}$ above the surface value. These criteria give the location of the seasonal pycnocline, and time series of mixed-layer depths will usually produce well-behaved, smooth annual cycles (e.g. Holte and Talley, 2009). The mixed-layer depth in temperate waters deepens from late fall into winter, and then shoals gradually from winter into spring. Mixed-layer depths during winter can reach several hundred metres (or more than 1000 m in regions of strong convection and deep water formation), and are typically a few tens of metres during summer.

As discussed above, the issue of relevance to the SCD hypothesis is not the depth of the mixed layer, but more precisely, the depth (and intensity) of active turbulence—the *turbulent layer*. Turbulence is usually measured in the field in terms of the dissipation rate of turbulent kinetic energy, $\varepsilon$ (units: m$^2$ s$^{-3}$ or W kg$^{-1}$, often just called "dissipation" for short). Large values of $\varepsilon$ indicate strong turbulence: there is a great deal of kinetic energy from turbulence being dissipated at small (mm–cm) scales. Unfortunately, there is no $\varepsilon$-based criterion for operationally identifying the base of the turbulent layer, partly because $\varepsilon$ is very difficult to measure, requiring specialized instrumentation and expertise, and partly because any such criterion would depend on the problem at hand (e.g. phytoplankton or temperature might have different criteria). Brainerd and Gregg (1995) show that a density step as small as 0.0025–0.005 kg m$^{-3}$ would often mark the base of the actively turbulent surface layer—the region of surface-enhanced $\varepsilon$. For most instruments, a density change this small would be considered noise. Indeed, Brainerd and Gregg (1995) pointed out that there were many density steps of this magnitude in their vertical profiles; in the absence of coincident dissipation measurements, it was impossible to say which one would mark the base of the turbulent layer. The strongly turbulent layer was almost always equal to or shallower than the mixed layer; on some occasions, "remnant turbulence" could still be intense below a newly forming mixed layer.

Since it is the turbulence—not the depth of the mixed layer—that is of relevance to the SCD hypothesis, it is worth building some intuition about the dynamics that control it. When mixing occurs in a stably stratified ocean, light water is pushed down and heavy water brought up, both moving against gravity. This vertical mixing actually raises the centre of gravity of the water column, thus increasing its potential energy. The gain in potential energy comes at the expense of kinetic energy, which is dissipated in the process of moving water up and down. Thus, to have vertical turbulence, one has to have a source of kinetic energy that is strong enough to overcome the existing density stratification. Turbulence in the surface boundary layer, then, is a trade-off between the kinetic (and sometimes potential—see the "Convectively driven turbulence" section) energy available to drive the turbulence, and the density stratification that can suppress it.

In the sections to follow, I first present and discuss some sources of energy that drive turbulence, and then sources of stratification that suppress it. These sources are presented not to be comprehensive, but more as vehicles to enhance our intuition concerning the time and space scales of changes in turbulence and restratification. A careful consideration of these time and space scales is essential to proper testing of the SCD hypothesis.

## Sources of turbulence

To mix fluid across a density gradient requires energy. This can be kinetic energy transferred across the ocean's surface by wind, waves, Langmuir circulations, or gravitational instability caused by a cooling of the ocean's surface (convection).

### Wind-driven turbulence

When the wind blows on the ocean's surface, it imparts kinetic energy to the ocean. Whether this kinetic energy goes into accelerating the surface waters horizontally (the Ekman layer) or into turbulence depends largely on the ambient density stratification. If the stratification is weak enough, kinetic energy from the wind will cause mixing of the density gradient. A criterion for determining whether the wind-driven shear will cause mixing in the face of stratification is given by the gradient Richardson number:

$$Ri = \frac{N^2}{(\partial u/\partial z)^2}, \tag{2}$$

where $N$ is the buoyancy frequency,

$$N^2 = \frac{g}{\rho}\frac{\partial \rho}{\partial z}. \tag{3}$$

$u$ is the horizontal velocity, $z$ the vertical coordinate, $g$ the acceleration due to gravity, and $\rho$ the density. The numerator of $Ri$ is a measure of the vertical stratification $\partial\rho/\partial z$, while the denominator quantifies the vertical shear. When $Ri < 0.25$, that is, when the vertical shear squared is greater than four times the buoyancy frequency squared, turbulent mixing is likely to occur. Strong shear or weak stratification (low buoyancy frequency) will allow turbulence and will generate vertical mixing of the density gradient.

The intensity of wind-driven turbulence varies with the strength of the wind. This is usually given through the friction velocity, $u^*$:

$$u^* = \sqrt{\frac{\tau}{\rho}}, \tag{4}$$

where $\tau$ is the surface windstress and $\rho$ the water density. The dissipation rate of turbulent kinetic energy $\varepsilon$ is then found from the "law of the wall" through the scaling

$$\varepsilon = \frac{u^{*3}}{\kappa z}, \tag{5}$$

where $\kappa$ is von Karman's constant (0.4, dimensionless). Examination of (5) shows that the dissipation rate—the intensity of turbulence—is predicted to decay with depth away from the surface as $z^{-1}$ (i.e. 1/depth, Figure 1). This decay with depth is less sharp than, say, the exponential decrease in irradiance with depth, but still leads to a surface-intensified distribution of turbulence.

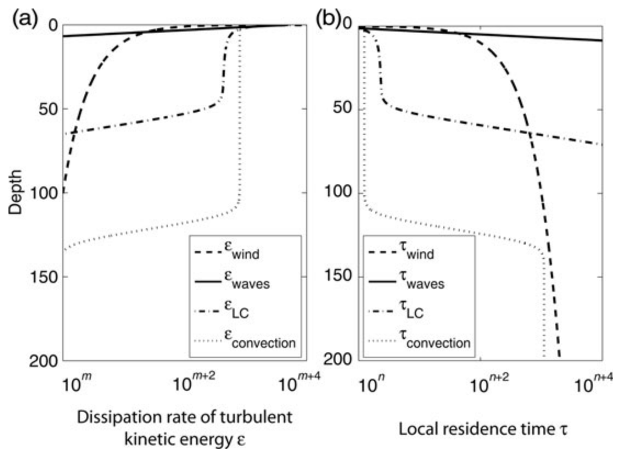


**Figure 1.** (a) Schematic of the vertical structure of the dissipation rate of turbulent kinetic energy, $\varepsilon$, generated by wind, waves, Langmuir circulations, and convective mixing. (b) Schematic of the vertical structure of the local residence time $\tau$ in the various types of turbulent layers in (a). Note that the actual values of dissipation and residence time are strongly dependent on the conditions: the magnitudes of these profiles should not be interpreted literally. The horizontal axes give the order-of-magnitude changes of the variables.

One implication from (4) and (5) is that wind-driven turbulence increases exponentially with the windstress; thus, small changes in windstress will drive proportionately larger changes in dissipation. These changes in dissipation can occur on very short time scales—when the source of energy changes, the dissipation changes within hours or less (e.g. Brainerd and Gregg, 1993a; D'Asaro, 2001).

The eddy diffusivity, $K$, which parameterizes the strength of the mixing in stratified waters is usually calculated from the dissipation and the buoyancy frequency following Osborn (1980):

$$K = \frac{0.2\varepsilon}{N^2}. \tag{6}$$

The "constant" of 0.2 is not technically a constant, but represents the "mixing efficiency": only 20% of the available energy dissipated by turbulence actually drives mixing. Eddy diffusivities for momentum (more properly called an eddy viscosity), density, temperature, salt, and phytoplankton could all be different, though typically the diffusivity for density is used as the diffusivity for plankton. From (6), we can see that the diffusivity will be higher when dissipation ($\varepsilon$) is stronger, and weaker when the density gradient ($N^2$) is stronger. It is clear from this formulation that the turbulence and stratification work against each other to give the resultant rate of mixing of the property. Note that this formulation is not applicable to turbulence driven by convection (Osborn, 1980) or in a mixed layer: the density gradient is zero ($N = 0$) or negative, and the predicted diffusivity is infinite (see the "Convectively driven turbulence" section).

A first-order estimate of the depth of the wind-mixed layer $L_{\rm wind}$ can be found from (Price and Sundermeyer, 1999; Wang and Huang, 2004)

$$L_{\rm wind} = \frac{u^*}{2f}, \tag{7}$$

where $f$ is the Coriolis frequency. Though this estimate does not account for variations in the strength of turbulence through the layer, it is clear that the turbulent layer will be thicker with a stronger wind or at more equatorward latitudes for a given windstress.

The local (with depth) residence time of a particle in a turbulent layer will vary inversely with the dissipation rate, regardless of the source of dissipation. For wind-driven turbulence, the residence time $\tau_{\rm wind}$ increases rapidly with depth (Figure 1), suggesting that phytoplankton not far from the surface will experience relatively long periods of constant irradiance. Thus, windforcing alone will not necessarily lead to a deep "thoroughly mixed layer".

## Wave-driven turbulence

Comparisons of field data with the $z^{-1}$ depth scaling of wind-driven dissipation sometimes agree (e.g. Harcourt and D'Asaro, 2008; D'Asaro, 2014). More often, though, the profiles of $\varepsilon$ show significant enhancement near the surface that cannot be accounted for by the $z^{-1}$ depth dependence of wind-forced turbulence (e.g. Anis and Moum, 1992, 1995; Brainerd and Gregg, 1993a). Part of this surface-enhanced dissipation is due to the actions of the surface wave field.

Turbulence driven by surface waves typically falls off exponentially from the surface (Huang and Qiao, 2010; Huang *et al.*, 2012; Sutherland *et al.*, 2013; Figure 1). The dissipation $\varepsilon$ is proportional to both the friction velocity $u^*$ (and thus the windstress) and the Stokes velocity $u_{S0}$ at the surface:

$$\varepsilon = C u_{S0} u^{*2} e^{-2kz}, \tag{8}$$

where $C$ is an empirical constant and $k$ is the horizontal wave number ($=2\pi$/wavelength) characterizing the surface wave field. The wave-driven turbulence thus depends strongly on the windstress (though less so than the wind-driven turbulence), but it decays more rapidly with depth than wind-driven turbulence, with a depth scale of $1/2k$. This exponential decay with depth often gives a strongly surface-intensified dissipation field (e.g. Anis and Moum, 1995; Huang *et al.*, 2012; McWilliams *et al.*, 2012; Sutherland *et al.*, 2013) with dissipation levels orders of magnitude higher than those due to wind alone (Figure 1). This wave effect on surface mixed layer turbulence is particularly apparent in the open ocean (Huang *et al.*, 2012).

Because of the rapid decrease in wave-driven turbulence with depth, the residence time $\tau_{\rm wave}$ of particles increases very rapidly with depth (Figure 1). Thus, it is unlikely that waves alone could cause strong turbulence over a sufficiently deep layer to keep phytoplankton well mixed through any significant portion of the euphotic zone.

## Langmuir circulations

Research in the last two decades has identified the importance of Langmuir circulations to mixing and turbulence in the upper ocean. D'Asaro and Dairiki (1997) showed that excess dissipation in the surface turbulent layer over that predicted by the law of the wall (i.e. windstress alone) could be accounted for by Langmuir circulations—or more specifically, the Craik–Leibovich vortex force (Craik and Leibovich, 1976), which dominates near the surface during large waves (D'Asaro *et al.*, 2014). Harcourt (2013) showed that the Craik–Leibovich vortex force gave dissipation rates well above the law of the wall scaling [$z^{-1}$, Equation (5)], generating surface-intensified and significantly enhanced dissipation rates in the turbulent surface layer (Figure 1).

McWilliams *et al.* (1997) characterized the importance of Langmuir circulations through a turbulent Langmuir number

$$La_t = \sqrt{\frac{u^*}{u_{S0}}}, \quad (9)$$

which is the ratio of the friction velocity (4) to the Stokes velocity at the surface. Harcourt and D'Asaro (2008) modified this to include the vertical gradient of the Stokes velocity:

$$La_{SL} = \sqrt{\frac{u^*}{\Delta u_{SL}}}. \quad (10)$$

Here, $\Delta u_{SL}$ takes the Stokes drift averaged over the top 25% of the mixed layer and subtracts from it the value near the base of the mixed layer, thus decreasing the sensitivity of (9) to the surface value. This scaling gave excellent representations of the turbulence intensity as a function of friction velocity (e.g. Harcourt and D'Asaro, 2008; D'Asaro, 2014).

The vertical distribution of dissipation driven by Langmuir circulations has mostly been studied through models. Skyllingstad and Denbo (1995), McWilliams *et al.* (1997), and Grant and Belcher (2009) found dissipation rates that decreased with depth at about the same rate as wind-driven turbulence [$z^{-1}$, Equation (5)]. Recently, Harcourt (2013) performed a fairly thorough modelling analysis of the vertical structure of dissipation in Langmuir circulations, and found a surface region in which dissipation decreased as $z^{-1}$ (though more slowly than wind-driven turbulence), lying above a thicker layer of relatively constant dissipation (Figure 1). Dissipation decreased rapidly below the mixed layer. This suggests that Langmuir circulations are quite efficient at moving particles vertically through the mixed layer—more efficient than wind- or wave-driven turbulence, but less efficient than convectively driven turbulence.

Particle residence times in Langmuir turbulence, $\tau_{LC}$, are relatively constant through an upper layer (usually corresponding to the mixed layer), increasing rapidly below (Figure 1). With strong Langmuir circulations, these residence times could be short enough to satisfy the conditions for the SCD hypothesis of a "thoroughly mixed top layer". The intense turbulence driven by the Craik–Leibovich vortex force (of which Langmuir circulations are an example) is often strong enough to inhibit restratification through heating (Kukulka *et al.*, 2013). D'Asaro and Dairiki (1997) showed rapid deepening of the turbulent layer (25 m over 10 h) presumably driven by Langmuir circulations, followed by a similarly rapid restratification after the wind died. D'Asaro (2014) points out that the effects of Langmuir circulations are still poorly understood, and their scalings are not well developed. It is clear, however, that they are an important source of turbulence in the upper ocean, though they probably do not extend as deep as turbulence driven by convection.

## Convectively driven turbulence

When the ocean loses heat through its surface, the top layers of the ocean become colder. Similarly, evaporation can increase the surface salinity; both these processes increase the density of the surface water. If these changes are great enough, the surface water can become denser than the water below it. This is gravitationally unstable: the surface water must sink, and in the process generates convectively driven turbulence. Because of the sinking of dense water to an equilibrium level, convectively driven turbulence is very efficient.

Convectively driven turbulence was first quantified in the ocean surface layer by Shay and Gregg (1984) in a Gulf Stream eddy during winter cooling. They showed intense dissipation throughout the surface turbulent layer, with an abrupt, two to three orders of magnitude decrease in dissipation at the base of the turbulent layer. While the dissipation due to windforcing decreases as $z^{-1}$ (5), convectively driven dissipation tends to be constant throughout the turbulent layer (e.g. Shay and Gregg, 1986; Lombardo and Gregg, 1989; Anis and Moum, 1994) with a magnitude scaled by the buoyancy flux $J_b$ (Lombardo and Gregg, 1989; D'Asaro, 2014; Figure 1). The depth at which the relative contributions of wind- and convection-driven turbulence are equal is given by the Monin–Obukhov length $L_{MO}$ (Shay and Gregg, 1986):

$$L_{MO} = \frac{u^{*3}}{\kappa J_b}. \quad (11)$$

At depths shallower than $L_{MO}$, turbulence is likely dominated by wind; dissipation due to convection is particularly apparent at depths $>L_{MO}$. The strong turbulence caused by convection tends to extend to the base of the mixed layer, and defines the seasonal pycnocline.

During convective mixing, the vertical temperature gradient can be negative, giving a negative buoyancy frequency. In this situation, Equation (6) cannot be used to calculate the diffusivities used to model particle and tracer motions. A better approach is the use of a turbulence closure model such as the K-profile parameterization (KPP—Large *et al.*, 1994) or Mellor and Yamada (1974) to model vertical profiles of diffusivities driven by boundary forcing. The strong, relatively uniform turbulence driven by convection leads to a constant vertical profile of plankton residence times, $\tau_{\text{convection}}$, in the convecting layer (Figure 1). These strongly turbulent layers are the most likely to satisfy the SCD hypothesis, with short residence times of plankton at any given irradiance in the upper layer. As I show below, however, small changes in stratification can rapidly (hours) shut down convective turbulence, leaving a well mixed but quiescent layer in which a bloom could form.

Measurements have shown that the intensity of dissipation, $\varepsilon$, tends to have a lognormal distribution, characterized by a few extremely intense events, with many much weaker occurrences. This lognormal distribution of dissipation leads to skewed distributions of diffusivities (6), with frequent occurrences of weak turbulence, and rare occurrences of intense turbulence events. These intense patches of turbulence can then drive infrequent, large fluxes over short time and space scales. Stevens *et al.* (2011) showed that the measured mean and median diffusivities were very different—the mean was biased towards large values by the rare occurrence of extremely high diffusivities. The biased mean would give a very different picture of the mixing climate for the phytoplankton: rather than experiencing weak turbulence most of the time with a few rare bursts of strong turbulence as suggested by the median diffusivity (the more accurate view), the mean diffusivity would suggest fairly strong turbulence all the time.

## Relevance to the SCD hypothesis

Each of the turbulence-generating mechanisms described above has a different vertical distribution of turbulence intensity or dissipation (Figure 1). From (1), this means that phytoplankton would have different residence times in different parts of the water column, affecting their local net growth and potentially allowing blooms in the presence of a much deeper (hydrographically measured) mixed layer. Wind- and wave-driven turbulence is strongly surface-intensified, giving longer residence times in deeper waters that are potentially still well within the euphotic zone. Convectively driven turbulence tends to move water in the turbulent layer most evenly, and would be the closest to satisfying the assumptions of the SCD hypothesis. Convection is a common source of turbulence during winter at temperate and high latitudes. As we shall see, however, convectively driven turbulence is relatively easy to suppress through extremely small changes in stratification.

## Sources of stratification

Waters in an unstratified ocean would be easy to move via turbulence, as vertical motions would not be inhibited by any density gradient. This is seldom the case, however, as there are many sources of density stratification that are constantly operating. As stratification increases, more kinetic energy is needed to create the same turbulent mixing. Here, I discuss three main sources of stratification in the ocean: heat flux, freshwater flux, and horizontal density gradients.

### Heat flux

The surface buoyancy flux $J_b$ has two main components: the surface sensible heat flux $Q_s$, and the latent heat flux $Q_l$:

$$J_b = \frac{g}{\rho}\left(\frac{\alpha}{c_P}Q_s + \frac{\beta S}{(1-S)H}Q_l\right), \tag{12}$$

where $\alpha$ is the thermal expansion coefficient, $c_P$ the specific heat of seawater, $\beta$ the haline contraction coefficient, $S$ the salinity, and $H$ the latent heat of vaporization (e.g. Brainerd and Gregg, 1993a). When the first term on the right is positive (depending on the orientation of your vertical coordinate), the surface of the ocean will heat, leading to enhanced vertical stratification. The second term on the right determines the change in density through evaporation (and consequent increases in salinity: see the "Freshwater flux" section).

The heat flux through the ocean's surface varies a great deal, both seasonally and daily. In many regions of the ocean, the heat flux will change from positive (net heating) to negative (net cooling) over a few hours during the course of the day, due to the sun's position in the sky. A negative night-time heat flux can drive deep convection, which is subsequently shut down by heat-induced stratification during the day (Figure 2). This daily cycle of heat flux strongly modulates the daily cycles of surface-layer turbulence.

Surface-driven turbulence is very sensitive to heating. Brainerd and Gregg (1993a) and Shay and Gregg (1986), for example, showed that a $<0.2°$C step in temperature or a $<0.005$ kg m$^{-3}$ step in density was sufficient to shut down dissipation below the diurnal turbulent layer (but above the seasonal mixed layer). The dissipation decayed with a time scale of $\sim$2 h (Caldwell *et al.*, 1997), giving decreases in the vertical diffusivity of a factor of 100 over a few hours (Brainerd and Gregg, 1993a; Peters *et al.*, 1994). The short time scales for changes in the intensity and vertical distribution of turbulence have been appreciated for many decades. Mellor and Durbin (1975) found that turbulent mixing decayed with a time scale of 20% of an inertial period ($<$1 d) in their mixed-layer model. Shay and Gregg (1986) observed dissipation to decrease in $<$1 d with changes in surface forcing, while Brainerd and Gregg (1993a) measured a 40$\times$ decrease in $\varepsilon$ over 4 h when convective turbulence was suppressed. D'Asaro and Dairiki (1997) observed restratification in 10 h after a storm passed their coastal study site.

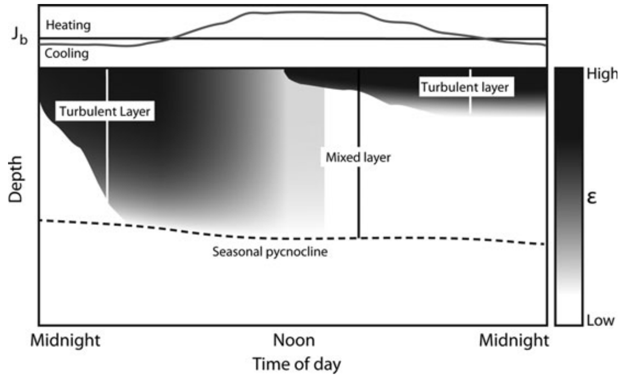


**Figure 2.** Schematic view of diel cycle of dissipation (grey scale, bottom panel) during net heating and net cooling at the surface (grey line, top panel). Cooling at night creates dense water at the surface that sinks, causing rapid convectively driven turbulence down to the seasonal pycnocline. Heating during the day can cause restratification that suppresses the turbulence from the surface, leading to a shallow turbulent layer with a remnant (non-mixing) mixed layer below (dashed line). A surface freshwater flux can similarly restrict the vertical extent of turbulence.

Stramska and Dickey (1993) showed that including the absorption of heat by phytoplankton could cause a 0.2°C increase in temperature relative to a water column without phytoplankton in a model of the North Atlantic during winter–spring transition. This phytoplankton-driven temperature increase could be sufficient to shut down convective turbulence, stabilizing the water column. Indeed, Stramska and Dickey (1994) showed that including the heating due to phytoplankton led to local blooms before large-scale stratification of the water column: the heating decreased the local turbulence, allowing blooms in the sense of the SCD hypothesis.

A few recent studies have compared diel-resolved fluctuations in surface heat flux $J_b$ and windstress $\tau$ to more averaged fluctuations in global models. Kamenkovich (2005) found that the high-frequency forcing led to shallower mixed layers in high-latitude regions in winter. Bernie *et al.* (2007) found that including the diurnal cycle led to trapping of momentum fluxes from the surface windstress near the surface in the time mean, leading to a reduction in vertical mixing and shallower turbulent layers. Kawai and Wada (2007) reviewed the literature and concluded that including diurnal variability is important, and that diurnal variations at high latitude, in particular, have been poorly studied.

### Freshwater flux

Changes in salinity at the ocean's surface can occur through evaporation, rain, or horizontal advection of waters of different salinity. Price (1979) was one of the first to detail the changes in vertical density distributions over the course of a rain event. He showed that the surface salinity decreased by 0.25, which was coincidentally similar to the change seen by Brainerd and Gregg (1997) in another study. The low-salinity signal was quickly (a few hours) mixed throughout the surface turbulent layer, decreasing its average density. This increased the density contrast at the base of the turbulent layer, which inhibited further mixed-layer deepening. The relatively static mixed layer could then accumulate more heat, which further enhanced the density step at the base of the mixed layer.

The inhibition of vertical turbulence by a low-salinity layer was also seen by Peters *et al.* (1994) and Wijesekera and Gregg (1996), who showed that the low-salinity "puddles" created by a rainstorm had a horizontal scale of ~10 km, and a vertical scale of 1 m. The salinity-induced barrier to turbulence was almost instantaneous, and lasted almost a day. This is consistent with observations of Hosegood *et al.* (2006) who found considerable vertical structure above the mixed layer after a rainfall, indicating decreased vertical turbulence due to salinity stratification.

Vertical variations in salinity through the interleaving of different water masses can have a profound effect on vertical turbulence. For example, Christensen and Pringle (2012) showed that a subsurface low-salinity layer in the Gulf of Maine offset the increased density of surface waters due to cooling, causing less turbulence than might be expected through convective heat loss alone. Shay and Gregg (1984) found a similar salt-stabilized temperature inversion in a Gulf Stream ring, which caused a 2–3 order of magnitude drop in dissipation at the base of the strongly turbulent layer. Long *et al.* (2012) found that wind-driven horizontal advection of low-salinity water was sufficient to stabilize the water column, allowing a bloom to form. Similarly, Ji *et al.* (2007, 2008) found that the timing and spatial pattern of the spring bloom in the Gulf of Maine depended on the freshwater flux into the Gulf from the Scotian Shelf: increased freshening caused earlier blooms.

In developing a data-based mixed-layer climatology, De Boyer Montégut *et al.* (2004) found that including salinity effects in the definition of the mixed layer led to shallower estimates of mixed layers during January, February, and March in the North Pacific and western North Atlantic, but deeper estimated mixed layers in the eastern North Atlantic due to salinity compensation of temperature gradients. This study re-emphasizes the point that neglecting salinity effects will affect estimates of mixed-layer depth, which are commonly used in tests of the SCD hypothesis. Furthermore, salinity variations are another factor causing differences between a mixed-layer depth and the actual depth of turbulent mixing.

## Horizontal stratification

With a few notable exceptions (e.g. Taylor and Ferrari, 2011; Mahadevan *et al.*, 2012), most studies examining SCD have assumed a one-dimensional (vertical) trade-off of turbulence and stratification. However, we have known since at least the 1990s (e.g. Brainerd and Gregg, 1993a, b) that the tilting (flattening or slumping) of horizontal density gradients can cause vertical density gradients that will decrease vertical turbulence (Figure 3). The tilting of isopycnals from near vertical to horizontal is known as "restratification": horizontal density gradients become vertical density gradients. Brainerd and Gregg (1993a, b, 1997) showed that horizontal gradients could cause restratification over the course of a few hours, during which the buoyancy frequency $N$ could increase by a factor of 10. Caldwell *et al.* (1997) found that ~40% of the restratification they observed was due to the relaxation of horizontal gradients, rather than local vertical processes. Hosegood *et al.* (2006, 2008) found that most of the restratification they observed in the surface mixed layer was through the tilting of existing horizontal density gradients. Stevens *et al.* (2011) also noted stratification induced by the slumping of horizontal density gradients during a Southern Ocean iron fertilization experiment, while Long *et al.* (2012) found horizontal advection-driven restratification to dominate turbulence, allowing phytoplankton blooms.

In addition to windforcing (Ekman flux), there appear to be two main mechanisms that cause horizontal density gradients to slump in the mixed layer: baroclinic (and other, often smaller-scale) instabilities, and the vertical shear of near-inertial waves. Hosegood *et al.* (2006) found a great deal of horizontal structure in temperature, salinity, and density at scales down to 2 km (the resolution of their vehicle). These structures occur at the "submesoscale" (e.g. Lévy *et al.*, 2012), and are often surface-intensified. A variety of instabilities can act on these horizontal gradients to cause them to slump, restratifying the mixed layer (Figure 3). Hosegood *et al.* (2006) concluded that near-surface baroclinic instabilities were responsible for the observed restratification, and that these dynamics occurred at horizontal scales that are not resolved in most models of ocean physics (but see Taylor and Ferrari, 2011; Mahadevan *et al.*, 2012 for counterexamples).

Tandon and Garrett (1994, 1995) showed that near-inertial waves can cause restratification. These low-frequency waves generated mainly by wind events propagate almost vertically, causing horizontal layers of water to oscillate horizontally relative to each other. This creates a vertical shear that can flatten existing horizontal

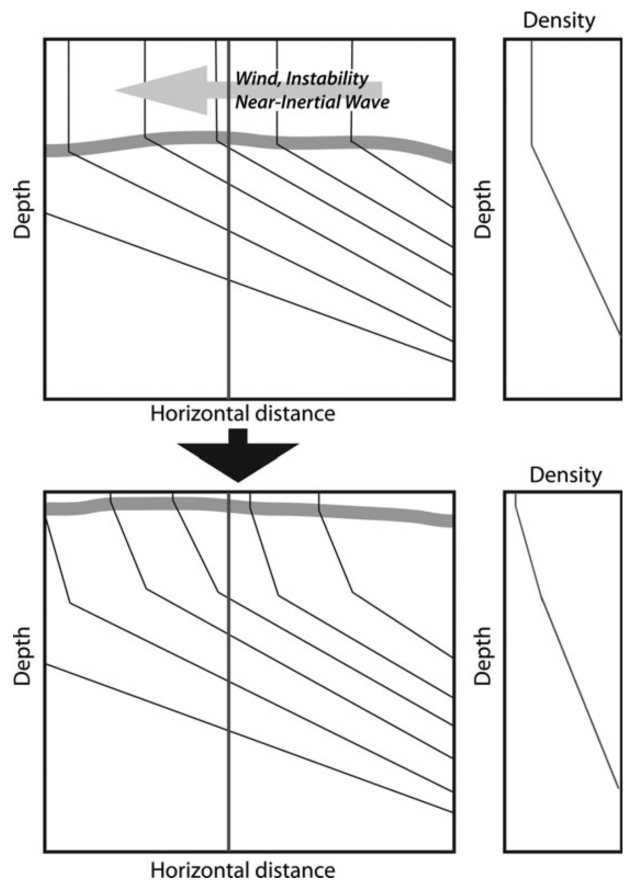


**Figure 3.** Schematic showing how a horizontal density gradient can slump to create a vertical density gradient that suppresses vertical turbulence. (Upper left) A horizontal density gradient in a region with deep turbulent layers is perturbed by a forcing with a vertical shear such as a baroclinic instability or a near-inertial wave. (Upper right) The vertical density profile taken at the vertical grey line in the upper-left panel, showing the deep mixed layer. (Lower left) The horizontal density gradient in the mixed layer has tilted, creating vertical density gradients that suppress turbulence (turbulent layer shown by thick grey line). (Lower right) The vertical density profile taken at the vertical grey line in the lower left panel, showing the shallow mixed layer after slumping.

density gradients, causing restratification (e.g. Hosegood *et al.*, 2008). The waves can also tilt horizontal gradients upwards, decreasing the local vertical density gradients, depending on the phase of the wave and the geostrophic currents.

Vertical shear of horizontal density gradients is thus a rapid and efficient mechanism to increase vertical stratification and damp vertical turbulence. It will occur wherever there are horizontal density gradients, and cannot be accounted for in one-dimensional models.

### Relevance to the SCD hypothesis

The sources of stratification determine how the kinetic energy of turbulence is distributed in time and space. Surprisingly small steps in density can suppress turbulence, leaving a well mixed but quiescent layer below an actively turbulent layer. Changes in density arise from heat flux, freshwater flux, and the slumping of horizontal density gradients, showing that the temporal and spatial structures of turbulence must be studied in a three-dimensional framework. The short time scales (hours) and small horizontal spatial scales (kilometres) of turbulence are likely to drive patchiness of phytoplankton growth on similar scales. This suggests that phytoplankton spring blooms should be patchy, transient, local phenomena, such as seen by Mahadevan *et al.* (2012).

## Discussion

In formulating his critical depth hypothesis, Sverdrup (1953) assumed a "thoroughly mixed top layer" with turbulence "strong enough to distribute the plankton organisms evenly through the layer". Subsequent tests of the SCD hypothesis have generally ignored the turbulence, and have instead used the operationally defined "mixed-layer depth" as a (very inadequate) proxy for turbulence intensity and vertical structure. The smooth seasonal cycles of mixed layers—averaged over weeks to months in time and 10–1000 km in space—used by most researchers remove most of the important aspects of turbulence that are essential to understanding the timing and location of a spring bloom. Bernie *et al.* (2007, 2008) included diel variability in forcing in a global circulation model focusing on atmosphere–ocean coupling in the tropics, and found significant changes in heating due to the diel rectification of the daily mean sea surface temperature. These hourly fluctuations propagated up to cause changes in seasonal and interannual cycles of both the ocean and atmosphere. In their review, Kawai and Wada (2007) underscore the importance of including diel variability in heat fluxes and windstress when exploring larger-scale problems: the short-time scale variability affects the long-term behaviour of the system. This problem is particularly acute at high latitudes where we have a paucity of high-frequency observations (but see Martin *et al.*, 2011; Cetinic *et al.*, 2012; Mahadevan *et al.*, 2012 for notable exceptions). Fundamentally, even if mixed-layer depths reflected the depth of strong turbulence, weekly or monthly estimates of mixed-layer depth should be used with caution and some scepticism when testing the SCD hypothesis.

Time averages of surface mixed-layer dynamics remove a great deal of important structure. One basic feature of turbulent layers is that they do not gradually move upwards in the water column as heating increases or wind decreases. Rather, they reform at the surface, leaving a remnant layer below. Tiny increases in temperature or density ($<0.02^\circ$C or $<0.005$ kg m$^{-3}$) can be sufficient to shut down wind-forced or convectively driven turbulence below. This shut down can occur over a few hours, leaving an unstratified but quiescent remnant mixed layer below a shallow, actively turbulent layer. At a local scale, then (and the scale that is relevant to the phytoplankton), the depth of the turbulent layer is highly variable, deepening when kinetic energy is available, and reforming at the surface when stratifying fluxes overcome the sources of turbulence. This leads to a discontinuity in the turbulent-layer depth: it deepens and then re-forms at the surface, often over the course of a day. This discontinuity has important implications for understanding the turbulent environment experienced by the phytoplankton: it is highly variable in intensity and vertical structure over relatively small horizontal spatial scales. From one hour to the next, the turbulent layer could shoal from 100 to 20 m due to some local heating, a rainstorm, or slumping of horizontal gradients, giving a newly quiescent environment for phytoplankton photosynthesis.

It is becoming increasingly clear from observations and models that vertical turbulence is intimately tied to submesoscale horizontal gradients in density. In their surveys, Hosegood *et al.* (2006) found horizontal density gradients on the scale of 1–2 km (the smallest scales they could resolve); these gradients significantly affected the intensity and distribution of vertical turbulence in the study area. Patchiness of surface salinity due to rain was estimated to have 10 km horizontal scales (Wijesekera and Gregg, 1996), and we would expect significant horizontal density structure at oceanic mesoscales (10–100 km) that would affect vertical turbulence. The horizontal wavelength of near-inertial waves is $<$10 km, similar to many mixed-layer instabilities. These horizontal density gradients and vertical shears should drive variations in turbulent layer depth on the same horizontal scales. It is clear then, that 100–1000 km horizontal averages of mixed-layer depth will smooth over a great deal of variability that is fundamental to understanding the local growth of the phytoplankton.

In this synthesis, I have tried to make the point that turbulence varies locally with time scales of hours and spatial scales of kilometres. One might ask then, how this local view of turbulence and consequent phytoplankton response can be reconciled with the large-scale (many weeks, 1000s of km) view of a spatially and temporally coherent spring bloom. As Chiswell *et al.* (2013) notes, "13 years of data suggest that this spring bloom initiation progresses smoothly, [while] uncomposited images of surface chlorophyll show that at any given time, surface chlorophyll is dominated by seemingly near chaotic processes ". In this view, then, the coherence of the spring bloom arises from averaging away the small-scale variability. The large-scale forcings and conditions are modulated by the local forcings and conditions to determine the local timing and intensity of the bloom: the large-scale forcings indicate when a bloom *might* occur; the local dynamics indicate whether it *will* occur. A local bloom could occur earlier or later than large-scale forcings would indicate, due to local winds, local freshwater flux, a local change in the heat flux (less cloud, for example), and mesoscale or submesoscale changes in the turbulent-layer depth, etc.

An obvious question then is, "Can large-scale averages of properties (e.g. temperature, mixed-layer depth, chlorophyll) and forcings (e.g. heat flux, wind, irradiance) be used to test the SCD hypothesis?" Upper-ocean turbulence is, after all, driven largely by surface forcings, as indicated in the synthesis above. Is it reasonable to use averaged forcings to infer averaged dynamics to compare with averaged data to test what is essentially a local hypothesis? Many of the properties being averaged (turbulence, mixed-layer depth, phytoplankton growth) have very non-linear responses to their forcings. The average of a non-linear variable depends greatly on the time- or space-scale of the averaging. Furthermore, the average of products of non-linear variables, for example, is not the same as the product of the averages, which is what is normally used to test the SCD hypothesis. I do not believe we can definitively answer the question posed here. But I do believe that we should approach such averaged analyses with healthy scepticism until we have better measurements (both remote and *in situ*), and better means of analysing the observations.

In reviewing mixed layers and turbulent layers, I have shown that both the sources of energy that drive vertical turbulence and the sources of stratification that suppress vertical turbulence are extremely patchy in time and space. The key point, however, is that to properly test the SCD hypothesis one must obtain coincident measurements of the turbulence and the phytoplanktonic rates, as it is the turbulent water motions that move the phytoplankton through the vertical light gradient. Drawing conclusions about the movement of phytoplankton in a mixed layer whose depth is determined by vertical profiles of temperature or density is potentially very misleading. Furthermore, using large-scale averages of properties to infer dynamics can easily lead to incorrect conclusions. Sverdrup (1953) was quite explicit about the requirement of a "thoroughly mixed layer"; currently, the only accurate measure of the existence of such a layer is through measurements of turbulence, *in situ*. I would maintain, therefore, that with the possible exception of Mahadevan *et al.* (2012), the SCD hypothesis has yet to be thoroughly tested in the field. Perhaps quantifying the turbulent layer rather than the mixed layer will lead to a deeper understanding of the timing and structure of spring phytoplankton blooms, and a reconciliation of the various tests of the SCD hypothesis.

## Summary

(i) Because of the variety of mechanisms creating turbulence in the surface turbulent layer, the mixed-layer depth (defined by temperature or density) is usually a poor indicator of the depth of turbulence, the vertical structure of turbulence, and the intensity of turbulence.

(ii) Different forcing mechanisms drive very different vertical distributions of turbulence intensity, from surface-intensified, to relatively uniform with depth.

(iii) Forcings that increase stratification can shut down active turbulence in hours, with density steps as small as $<0.005$ kg m$^{-3}$.

(iv) Turbulent layers do not shoal monotonically, but rather tend to re-form at the surface, and then entrain downwards with time.

(v) Fluctuations in turbulence occur over time scales of hours and spatial scales of kilometres.

(vi) It is essential to include the short time scale and small spatial scale variations of turbulence in analyses when testing the SCD hypothesis. The depth of the mixed layer is an insufficient and usually inaccurate measure of the turbulent layer depth.

(vii) Inclusion of the structure and intensity of turbulence may lead to a reconciliation of the various tests of the SCD hypothesis.

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