# Physical and biological processes underlying the sudden surface appearance of a red tide in the nearshore

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#### Abstract

The sudden appearance at the surface of an alongshore-parallel band of red tide near Huntington Beach, California, is described in high spatial and temporal resolution using novel instrumentation including a global positioning system–tracked jet-ski. The scale of the surface chlorophyll *a* (Chl *a*) band was small (~ 200 m cross-shore) and ephemeral (3 h) compared with the subsurface extent of the red tide (~ 2 km, > 7 d). The red tide was dominated by the regionally common dinoflagellate *Lingulodinium polyedrum* (F. Stein) and had developed as a subsurface Chl *a* layer during the 7 d prior to the surface appearance. A few hours before the surface appearance, a subsurface patch of elevated Chl *a* (Chl *a* > 30  $\mu$ g L<sup>-1</sup>) was observed in 13-m total depth in the trough of a shoreward-propagating internal wave, consistent with dinoflagellate vertical swimming interacting with the internal wave-driven convergence. Internal wave-breaking–induced vertical mixing in ~ 8-m water depth vertically spread the Chl *a* patch to the surface Chl *a* band were prevented from entering the surf-zone by a density barrier of warm water adjacent to the beach. These high-resolution observations emphasize the role of nearshore physical dynamics in controlling the duration and intensity of red tide exposure to coastal habitats.

Intense phytoplankton blooms ('red tides') and harmful algal blooms (HABs) in the nearshore cause losses of millions of dollars annually to aquaculture (Alonso-Rodríguez and Páez-Osuna 2003), threaten marine mammal and human health (Anderson 1997), and inhibit beach recreation (Backer et al. 2003). An understanding of the mechanisms controlling the intensity and timing of red tides is particularly critical in the nearshore (depth < 20 m) and surf-zone (depth < 3 m) where the often densely populated benthic habitats are vulnerable to red-tide exposure. HABs often begin as thin subsurface layers ('cryptic blooms'; McManus et al. 2008), invisible to satellites and visual inspection. Furthermore, their surface expression is often patchy in time and space, not reflecting the subsurface patterns, making effective monitoring a challenge (Anderson 1997).

Interpretation of surface or near-surface data must be coupled with an understanding of the physical mechanisms that control the vertical and horizontal distributions of chlorophyll a (Chl a) in the water column. Over time scales < 1 d, changes in phytoplankton biomass due to growth are typically small: with a net growth rate  $\mu \sim 0.29 \text{ d}^{-1}$ , the red-tide-forming dinoflagellate Lingulodinium polyedrum (Sullivan and Swift 2003) would show an increase of only  $\sim$  $1.3 \times$  over 1 d. On the other hand, Chl *a* variability driven by advection-for example, tidal currents and stirring (Cloern and Dufford 2005), internal tides (Kamykowski 1974), or internal wave-induced circulation in combination with swimming or sinking of cells (Lennert-Cody and Franks 1999, 2002)—has been shown to cause increases > $2\times$  over time scales of hours or less. Mechanisms that concentrate or advect red tides over short space and timescales may carry severe consequences. For example,

mussels exposed to high concentrations of HAB toxins may surpass quarantine limits within only 1 h, whereas prolonged exposure to lower concentrations may not be harmful (Bricelj et al. 1990).

The study described here presents high spatially and temporally resolved observations of a nearshore red tide and explores the physical dynamics underlying a sudden surface appearance of a dense red-tide band. The surface expression of the red tide was visible from the beach as a brown streak 500 m from shore and extending > 1 km alongshore (Fig. 1). The observations were collected during the Huntington Beach experiment (HB06) in autumn 2006. Within the 1-month HB06 study period, an intense subsurface layer of the dinoflagellate L. polyedrum developed over the 7 d before its peak on 12 October. On 12 October, the red-tide layer was mapped repeatedly using several complementary techniques. The measurements revealed that in the hours preceding the midday appearance of the surface Chl *a* band, an intense subsurface Chl *a* patch formed within a shoreward-propagating supertidal (period < semidiurnal) internal wave trough. The location and intensity of the patch was consistent with concentration increases driven by depth-keeping swimming of the dinoflagellates in the internal wave-induced convergence. The surface alongshore-parallel Chl *a* band did not form directly through the interaction of the internal wave convergence and vertical swimming (see Kamykowski 1974; Lennert-Cody and Franks 1999, 2002). Instead, upon arrival in shallow water (depth < 13 m), internal wave breaking spread the intense subsurface Chl a patch vertically, forming the visible band of surface Chl a just seaward of the surf-zone. After the surface band appearance. Chl a was inhibited from entering the surf-zone by a warm-water plume transported from a nearby marsh by wave-driven alongshore currents.

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Fig. 1. Un-retouched photograph of a brownish alongshoreparallel band of red tide (arrow)  $\sim 500$  m offshore, taken from the beach in early afternoon on 12 October. The small breaking surface wave in the foreground is near the beach, and the 5-m-long CTD + Chl *a* cast boat is offshore.

## Methods

The nearshore HB06 field experiment was conducted at Huntington Beach, California, from 15 September to 17 October 2006, within the framework of the larger scale Southern California Coastal Ocean Observing System and United States Geological Survey programs. Here data are presented from 05 to 16 October with a focus on 12 October when the red tide appeared at the surface. Observations spanned 0.5 km alongshore (y) and 4 km offshore (x) to 25-m depth (Fig. 2a,b). The mean (tidally averaged) water depth is denoted as H, and the vertical coordinate z is positive upward, with z = 0 m at the tidally averaged surface.

Jet-ski surface maps—Five quasi-synoptic maps of nearsurface Chl *a*, turbidity, and temperature (*T*) were constructed from data acquired at 45-min intervals between 12:00 h and 15:00 h on 12 October with a flow-through sampling system mounted on a global positioning system tracked jet-ski (Clark et al. 2009). The jet-ski was driven on 1-km-long cross-shore transects between the shoreline and  $\sim$  13-m depth with eight transects distributed over 600 m alongshore (dashed box in Fig. 2a). Similar jet-ski data were acquired on 05 and 09 October but lacked concurrent Conductivity, Temperature, Depth and Chl *a* (CTD + Chl *a*) transects and so are not discussed.

CTD + Chl a casts—Transects of CTD (Seabird 19) and flow-through Chl a fluorometer (WET Labs WETStar) profiles (x, z) spanning ~ 6 km cross-shore (from H = 40 m to H = 4 m) were conducted between 09:00 h and 11:30 h once per day on 05, 07, 08, 12, and 16 October. In addition, on 12 October, a series of five short cross-shore transects spanning ~ 1 km (H = 13 m to H = 4 m) were conducted between 12:00 h and 15:00 h, concurrent with jet-ski surface mapping. Profile data were averaged into 1-m vertical bins. The jet-ski-derived surface transects (x, y) and CTD + Chl *a*-derived cross-shore transects (x, z) were objectively mapped with a skill threshold of 0.95 and a signal-to-noise ratio of 0.9 (Bretherton et al. 1976).

Moorings-Moorings were deployed on a 4-km-long cross-shore transect at mean (tidally averaged) water depths H = 8 m, 15 m, 18 m, and 27 m (moorings M8, M15, M18, M27 in Fig. 2a,b) and instrumented with 4-6 Star-Oddi (T only) or Sea Bird MicroCAT (S and T) instruments sampling at 3-min intervals between 05 to 16 October (Fig. 2c). A wirewalker, a wave-driven vertically profiling platform (Rainville and Pinkel 2001), was deployed at H = 13 m (M13 in Fig. 2a,b), and instrumented with a CTD (Seabird 49, sample rate 16 Hz) and Chl a fluorometer (WET Labs ECO Triplet, sample rate 16 Hz). Vertical CTD + Chl a wirewalker profiles at M13 were completed approximately every 2 min; data were averaged every 0.1 m in z and interpolated onto regular 4-min intervals. The M13 wirewalker was intermittently operational, but with > 90% data coverage between 10 and 16 October (Fig. 2c). Cross-shore (u) and alongshore (v)currents measured with Acoustic Doppler Current Profilers (ADCPs) deployed at the base of moorings M13 (600 kHz), M15 (600 kHz), and M18 (600 kHz) were averaged to 6min intervals with 1-m vertical bins. A bottom-mounted Nortek Aquadopp at M8 (2000 kHz) sampled every 2.5 min with 0.5-m vertical bins. These current meters were consistently operational from 05 to 16 October (Fig. 2c).

Surf-zone frames-Two fixed frames with Acoustic Doppler Velocimeters, pressure, and temperature sensors (sampled at 8 Hz, 8 Hz, and 0.2 Hz, respectively) were deployed between 05 and 16 October on a cross-shore transect spanning 160 m from near the shoreline to 4-m water depth. One frame was placed just seaward of the surfzone (M4 at H = 4 m) and one within the surf-zone (M1.5 at H = 1.5, see inset in Fig. 2b). Instruments at M1.5 were nonfunctional during exposure at low tide (Fig. 2b). WET Labs ECO Triplet fluorometers measuring Chl a (sampled at 0.25 Hz) were mounted on M4 and M1.5 facing  $30^{\circ}$  from downward, nominally 0.5 m above the seafloor. These fluorometers were repeatedly deployed for 72-h periods with  $\sim$  12-h turnaround time (Fig. 2c). A correction for the data bias induced by surf-zone turbidity was applied to the ECO Triplets (Omand et al. 2009).

*Water samples*—During most CTD + Chl *a* casts, water samples (for Chl *a* analysis and phytoplankton cell counts and identification) were collected near the bottom with a messenger-tripped Niskin bottle mounted 1 m above the CTD, and at the surface by hand. Water samples were also collected near the surface every 20 min on 12 October from 9:00 h to 15:00 h near M1.5 and M4 (Fig. 2c). Samples for taxonomic identification were preserved by transferring 100 mL of raw seawater to a glass Wheaton bottle and combining with 10 mL buffered 37% formaldehyde. The preserved samples were analyzed with the Utermöhl settling method (Utermöhl 1958) and inspected at 16× magnifica-



Fig. 2. Schematic and timeline of HB06 instrumentation: (a) plan view of bathymetry contours vs. the alongshore (y) and cross-shore (x) coordinates, and (b) cross-shore transect of bathymetry (dark shading). The vertical coordinate is z, with z = 0 m at the tidally averaged surface and positive upward. Moored temperature strings (circles, M8–M27) were located at total depths H = 8 m, 15 m, 18 m, and 24 m. A vertically profiling CTD + Chl a wirewalker (M13) was located at H = 13 m. Bottom-mounted ADCPs (stars) were located near moorings M8, M13, M15, and M18. Fixed frames within the surf-zone (M1.5) at H = 1.5 m and seaward of the surf-zone (M4) at H = 4 m measured Chl a (triangles), T (circles), wave height, and currents (*see* inset in [b]). A global positioning system–tracked jet-ski measured surface T and Chl a within the black dashed box in (a) and small boat (CTD + Chl a) completed transects during daylight hours from H = 5–60 m (~ 8 km offshore). The Talbert Marsh outlet is located ~ 1 km south of the instrument transect. (c) Gantt chart indicating velocity (gray), T (black), and Chl a (green) data coverage between 05 October and 16 October. Stars indicate CTD + Chl a transect days and circles indicate M4 and M1.5 bottle samples for phytoplankton identification.

tion using a light microscope. All visible cells (> 5  $\mu$ m) were enumerated and identified to a genus or species level when possible.

Nonphotochemical quenching correction—Nonphotochemical quenching (NPQ) reduces Chl *a* fluorescence in high-light conditions (Kiefer 1973; Falkowski and Raven 1997; Muller et al. 2001). NPQ corrections, derived from continuous profiles of Chl *a* at M13 (Fig. 2a,b) and surface irradiance (measured with a Davis Vantage Pro Plus cosine pyranometer), were applied to all in situ Chl *a* measurements (*see* Web Appendix, <www.aslo.org/lo/toc/vol\_56/ issue\_3/0787a.pdf>). The NPQ correction was generally < 10% at the 17°C isotherm at M13, but was occasionally as large as 50% at the surface at noon. Similar results were obtained with and without the NPQ correction.

## Results

Subsurface red-tide observations—The subsurface Chl a distributions could be separated into two relatively distinct features: an offshore deep Chl a maximum (DCM) formed mainly by diatoms, and an inshore subsurface layer formed by dinoflagellates (Table 1). Chl a within the DCM

Table 1. Phytoplanktonic genera present on 12 October within the nearshore (H < 20 m) and offshore (H > 20 m) regions. The total number of cells counted within each region is *n*. Dinoflagellates are identified in bold type.

Genus	% nearshore $(n = 11,949)$	% offshore $(n = 11,296)$		
Lingulodinium	71.6	4.9		
Chaetoceros	6.4	34.8		
Asterionella	6.3	32.9		
Prorocentrum	4.6	0.9		
Psuedonitzschia	4.3	7.1		
Skeletonema	0.6	7.5		
Other	6.1	11.9		

increased from a maximum of 6  $\mu$ g L<sup>-1</sup>on 05 October to 10  $\mu$ g L<sup>-1</sup> on 12 October (Fig. 3). The DCM lay between ~ 13°C and 16°C, and spanned  $\geq$  6500 m cross-shore. Concurrent CTD + Chl *a* glider surveys conducted ~ 10– 30 km offshore also showed an increase of mean subsurface Chl *a* between 05 and 12 October, but with much lower maximum concentrations (1.5–2.5  $\mu$ g L<sup>-1</sup>; Todd et al. 2009).

The *L. polyedrum* red tide initially developed in a nearshore subsurface layer spanning 2 km cross-shore. The region of this layer with Chl  $a > 10 \ \mu g \ L^{-1}$  between the 16°C and 17°C isotherms will be referred to as the 'Chl *a* layer.' Nearshore Chl *a* was generally low on 05 October (maximum 4.5  $\ \mu g \ L^{-1}$ ; Fig. 3a) increasing to 13  $\ \mu g \ L^{-1}$  by 08 and 09 October (Fig. 3b,c), and peaking at ~ 34  $\ \mu g \ L^{-1}$  on 12 October. By 16 October, the Chl *a* layer concentrations decreased to 12  $\ \mu g \ L^{-1}$  (Fig. 3e).

Although alongshore advection of Chl *a* gradients may have contributed to the observed increase in concentration in the Chl *a* layer between 05 and 12 October, a net growth rate of only  $\sim 0.23 \, d^{-1}$  would also account for these changes. This net growth rate is quite plausible for *L. polyedrum* (Sullivan and Swift 2003) and would lead to a 10-fold increase in Chl *a* over 1 week, consistent with observations (Fig. 3). Over 1 d such a growth rate would lead to only a 1.3-fold increase.

Conditions on 12 October, the focus of this study, were typical of an autumn day at Huntington Beach: winds were weak, blowing slightly offshore in the morning with a northwest 3 m s<sup>-1</sup> breeze in the afternoon. The surface water temperature was ~ 17.5°C, dropping to 14°C at the base of the thermocline at ~ 20 m (Fig. 3d). The phytoplankton taxonomic composition was strikingly different between nearshore and offshore (Table 1). The nearshore was dominated by a near-monoculture of dinoflagellates (> 70% Lingulodinium polyedrum) whereas offshore waters contained a greater diversity of genera, dominated by diatoms from the genera Asterionella and Chaetoceros. Although less prevalent than on 12 October, L. polyedrum cells exceeded 50% in the nearshore on 08, 09, and 16 October.

Diurnal and semi-diurnal frequency fluctuations dominated the current and T variability at moorings M27 to M08 during HB06, similar to prior studies at this location



Fig. 3. Cross-shore sections (x, z) of Chl a ( $\mu$ g L<sup>-1</sup>, colors) and T (°C, black contours) obtained between 09:00 h and 11:30 h on 05 (a), 08 (b), 09 (c), 12 (d) and 16 (e) October. The red circle is M13 surface location and the white triangles indicate CTD + Chl a cast locations. The white line traces the 4- $\mu$ g L<sup>-1</sup> Chl a contour. (d) The black box indicates the nearshore red-tide focus area on 12 October.



Fig. 4. Time series (spanning 54 h) at M13 (H = 13 m) of (a) tidal excursion from mean sea level, (b) vertical profiles of 15-min averaged cross-shore currents u at M13 (positive = onshore), and (c) 15-min averaged vertical profiles of Chl a ( $\mu$ g L<sup>-1</sup>, colors) and T (°C, black contours) at M13, (d) is an enlarged view of the boxed region in (c) with hourly averaged cross-shore currents (u) represented by arrows for  $|u| \ge 0.01$  m s<sup>-1</sup> and by black dots where |u| < 0.01 m s<sup>-1</sup>. The 15min averages from (b) are now shown as arrows here because they crowded the plot and blocked the Chl a colors. The white curves trace the 4- $\mu$ g L<sup>-1</sup> Chl a contour. Time zero is 00:00 h 12 October.

(Boehm et al. 2002; Noble et al. 2009). On 11–12 October, the tidal range was  $\pm$  0.9 m (Fig. 4a). Cross-shore currents at M13 contained both diurnal and semidiurnal baroclinic variability (Fig. 4b), and the Chl *a* layer repeatedly shoaled from near-bottom to near-surface and back down (Fig. 4c). The Chl *a* layer followed the 16.0–17.0°C isotherms (black contours in Fig. 4c) throughout the diurnal cycle, except for a few hours after the 17.5°C isotherm descended, leaving increased near-surface (z > -5 m) Chl *a* concentrations ( $\sim 5 \ \mu g \ L^{-1}$ ) above the 17.5°C isotherm (e.g., at t = -32 h, -8 h, and 14 h in Fig. 4c).

At M13 between 10:00 h and 12:00 h on 12 October, a Chl *a* patch (maximum 34  $\mu$ g L<sup>-1</sup>) was observed between 5 m and 7 m below the surface, at the 17.0°C isotherm (Fig. 4d). The onshore currents ( $u \sim 0.05$  m s<sup>-1</sup> between 10:00 h and 12:00 h) between -7 < z < -5 m advected the Chl *a* patch shoreward. Based on the onshore *u* over the 2-h M13 patch duration (Fig. 4d), the high concentration subsurface Chl *a* patch was estimated to be 200 m wide across-shore. This region of very concentrated Chl *a* patch.



Fig. 5. Sequential (top to bottom) maps of nearshore Chl *a* (left) and *T* (right) on 12 October at 12:00 (a,b), 12:45 (c,d), 13:30 (e,f), 14:15 (g,h) and 15:00 h (i,j) from jet-ski-based horizontal (*x*, *y*) measurements and CTD + Chl *a*-based cross-shore transect (*x*, *z*) measurements. Arrows represent the directions of the near-surface alongshore (at M1.5, M8, and M13) and nearsurface (~ 2 m below mean sea level) and near-bottom (~ 2 m above the seafloor) cross-shore currents (at M8 and M13 only). The jet-ski required 30 min to complete the eight cross-shore transects used for each map, and the centered-time (12:00–15:00 h) is shown to the right of each panel. Each CTD + Chl *a* transect consisted of five or six profiles. Black lines are temperature contours. The white line traces the 4- $\mu$ g L<sup>-1</sup> Chl *a* contour.

Surface red-tide observations—Five CTD + Chl *a* crossshore transects and near-synoptic jet-ski surveys of surface Chl *a* and *T* were conducted on 12 October at 45-min intervals between 12:00 h and 15:00 h inshore of M13 (Fig. 2c). At 12:00 h the Chl *a* patch (up to 39  $\mu$ g L<sup>-1</sup>) was located ~ 5 m below the surface (Fig. 5a). Surface Chl *a* and *T* were spatially homogeneous at 3  $\mu$ g L<sup>-1</sup> and 17.7°C, respectively (Fig. 5a,b). Hourly averaged cross-shore currents were ~ 0.03 m s<sup>-1</sup> onshore near the surface and offshore near the seafloor. The surf-zone alongshore current was 0.15 m s<sup>-1</sup> northward. By 12:45 h, a small (~ 10-m cross- and alongshore) region of ~ 4  $\mu$ g L<sup>-1</sup> Chl *a* appeared at the surface with slightly cooler, 17.5°C water (Fig. 5c,d). The Chl *a* at the surface that formed the alongshore parallel streak will be referred to as the 'Chl *a* band.'

At 13:30 h (Fig. 5e,f), the Chl *a* band had enlarged and intensified to ~ 5  $\mu$ g L<sup>-1</sup> near x = 400 m in 17.5°C water, spanning 100 m across-shore and 250 m alongshore. Near the surf-zone (x < 150 m), the water had warmed to 18°C. The Chl *a* patch was centered on the 17°C isotherm directly below the Chl *a* band, and had increased in thickness. At 14:15 h, the Chl *a* band had increased to 7  $\mu$ g L<sup>-1</sup>, with a cross-shore span of about 200 m (Fig. 5g) and visually estimated alongshore length ~ 1 km.

The near–surf-zone region continued to warm, and the colder (< 17°C) surface water at x > 500 m receded offshore (Fig. 5h). The Chl *a* patch continued to spread vertically into warmer 17.5–18°C water. Later at 15:00 h, the surface Chl *a* band had advected slightly offshore with the < 17.5°C water (Fig. 5i,j).

#### Discussion

A dense, nearshore dinoflagellate bloom developed over  $\sim$  7 d in a subsurface Chl *a* layer spanning 2 km in the cross-shore. The bloom was invisible from the surface until 12 October when a small-scale (a few hundred meters wide) elevated Chl *a* band surfaced for a few hours  $\sim 500$  m from the beach. The brief, small cross-shore scale of the Chl a band contrasts with the persistence and larger extent of the subsurface Chl a layer. The timing and scale of the Chl a band was determined by the combination of several physical-biological processes: (1) shoreward propagation of an internal wave trough, (2) local formation of a dense Chl a patch within the trough, (3) internal wave breaking in H < 13 m, (4) wave-breaking-induced vertical and crossisotherm spreading of the Chl a patch, and (5) a warmwater density barrier to Chl a entry in the surf-zone. Evidence for these mechanisms, and alternate hypotheses underlying each stage, are discussed in the following subsections.

Stage 1: Shoreward propagation of a supertidal internal wave—Estimation of the shoreward propagation speed of the supertidal internal wave (sIW) allows prediction of the arrival time of the sIW in very shallow water where breaking and vertical mixing of the Chl *a* patch occurred. On 12 October between 00:00 h (midnight) and 03:00 h, a mid-water column sIW of depression with an amplitude of ~ 5 m was observed at M27 at 05:00 h (star; Fig. 6a). The sIW trough subsequently passed M18 and M15, reaching M13 at 11:00 h (black stars; Fig. 6b–d). There was no evidence of the sIW trough at M8 (H = 7.8 m; Fig. 6e) implying that breaking had obscured the shape of the trough.

The wave phase speed was calculated as  $C_p = \Delta x / \Delta t$ , where  $\Delta x$  is the cross-shore separation between adjacent moorings, and  $\Delta t$  is the sIW travel time (e.g., time lag between arrivals of the sIW trough, black stars; Fig. 6). The sIW travel time  $\Delta t$  is estimated at the lag for maximum correlation ( $r^2 > 0.63$ , p < 0.001; Table 2) between midwater column isotherm depth anomalies at adjacent moorings (Lerczak 2000) over a 15-h period (from 00:00 h to 15:00 h) that includes the sIW trough (Fig. 6a). Assuming a normally incident sIW, the observed crossshore phase velocity  $C_p = \Delta x / \Delta t$  (where  $\Delta x$  is the crossshore separation between adjacent moorings) decreased from 0.16 m s<sup>-1</sup> between M27 and M18 to 0.12 m s<sup>-1</sup> between M15 and M13 (Table 2). Refraction of the sIW toward normal incidence precludes large angles of incidence. Obliquity of  $30^{\circ}$  from shore-normal causes only a 15% underestimate in  $C_p$ .

The observed cross-shore  $C_p$  variation is similar to the theoretical dependence on water depth of  $C_p$  for linear, long (wavelength >> water depth) internal waves (IW). With constant buoyancy (N) and wave ( $\omega$ ) frequencies, the theoretical phase speed ( $C_p$ ) for a mode 1 cross-shore–propagating internal wave (Gill 1982)

$$C_p = \frac{H}{\pi} \sqrt{N^2 - \omega^2} \tag{1}$$

depends linearly on the depth *H*. The theoretical phase speed  $C_p$  was estimated from Eq. 1 using the vertically averaged *N*, calculated from dT/dz adjusted to density. A nominal wave period of 2 h was chosen based on the observed passage time of the sIW depression (*see* Fig. 6) to estimate  $\omega$ . The buoyancy frequency  $N (\sim 0.02 \text{ s}^{-1})$  was large compared with  $\omega$ , so the theoretical  $C_p$  was insensitive to the choice of  $\omega$  ( $C_p$  varied by < 1% for wave periods between 2 h and 12 h). Theoretical  $C_p$  are similar for a twolayer IW (Sverdrup 1942):

$$C_p = \sqrt{2g\left(\frac{\rho - \rho'}{\rho + \rho'}\right)\frac{hh'}{h + h'}} \tag{2}$$

where  $\rho$  ( $\rho'$ ) and *h* (*h'*) are the upper (lower) layer density and thickness, respectively. The upper layer was defined as the time-averaged (over 2 h preceding the passage of the IW) mean sIW isotherm depth (*see* bold isotherm contours in Fig. 6) and  $\rho$  (calculated from *T*) was vertically averaged across each layer.

The observed  $C_p = \Delta x / \Delta t$  varied from ~ 0.12–0.17 m s<sup>-1</sup> and were similar to theoretical  $C_p$  (Table 2) showing that the coherent propagation over ~ 3 km between M27 and M13 is consistent with simple estimates for IWs. The observed  $C_p$  decrease with decreasing depth, has been observed previously for sIW on the Southern California shelf (Winant 1974; Winant and Olsen 1976; Johnson et al. 2001).

Stage 2: Formation of the intense Chl a patch within an internal wave trough-Motile dinoflagellates such as Lingulodinium polyedrum can maintain water column position despite negatively buoyant cell bodies (Cox 1980). The combination of swimming and convergent advection may lead to accumulation of phytoplankton at fronts (Franks 1992) or above internal wave troughs (Franks 1997; Lennert-Cody and Franks 1999, 2002). In studies conducted  $\sim$  150 km south of Huntington Beach, California, Lennert-Cody and Franks showed that L. *polvedrum* advection by IWs in combination with depthkeeping swimming behavior locally enhanced Chl a above wave troughs. The Lennert-Cody and Franks model suggests that the degree of Chl a enhancement is proportional to the separation between the isotherm and the surface  $\Delta z$ , for a given swimming proficiency. Similarly, systematically elevated Chl a within the sIW-associated isotherm trough was observed at M13, qualitatively consistent with IW-induced convergence and phytoplankton depth-keeping swimming behavior.



Fig. 6. Contours of T (°C) vs. time and z at (a) M27, (b) M18, (c) M15, (d) M13, (e) M8. Black stars represent an internal wave trough propagating onshore. The bold contour is a representative mid-water isotherm. The white star is the trough location at M8 predicted by linear theory. In (d) M13, Chl a ( $\mu$ g L<sup>-1</sup>) is colored. Time zero corresponds to 00:00 h 12 October.

As described in the previous section, the sIW propagated coherently onshore, maintaining shape, with a phase speed consistent with linear theory. At M13 on 12 October, Chl *a* in the sIW trough (black star; Fig. 6d), was maximum (31  $\mu$ g L<sup>-1</sup>) and nearly triple the (vertical) mean

concentration (11.5  $\mu$ g L<sup>-1</sup>) bordering the 17.0°C isotherm between 06:00 h and 17:00 h. During this time period, Chl *a* at M13 (vertically averaged between the 17.0°C isotherm and the surface; solid black curve in Fig. 6d) and the isotherm depth  $\Delta z$  were significantly positively correlated

Table 2. Separation distance ( $\Delta x$ ), time lag maximizing  $r^2$  ( $\Delta t$ ), and correlation coefficient ( $r^2$ ) of the sIW depth contours at the midpoint (Midpt.) between adjacent moorings. The observed phase velocity  $C_p$  (obs.)= $\Delta x/\Delta t$ . At each mooring location, theoretical  $C_p$  ( $m \ s^{-1}$ ) from Eq. (1) and (2) and  $u_{max}$  ( $m \ s^{-1}$ , positive = onshore) above the wave trough. All correlations are significant at the 99% level.

Location	$\Delta x$ (m)	$\Delta t$ (h)	$r^2$	$C_p$ (obs.)	$C_p(1)$	$C_p(2)$	$u_{max}$
M27				_	0.170	0.174	0.08
Midpt. (M27–M18)	1643	2.9	0.75	0.157			
M18					0.129	0.139	0.07
Midpt. (M18–M15)	715	1.6	0.63	0.120			
M15					0.116	0.116	0.06
Midpt. (M15–M13)	549	1.3	0.79	0.117			
M13	—				0.100	0.080	0.06

(Fig. 7). The vertically averaged Chl a was maximum where the isotherms were the deepest. In contrast, T, vertically averaged between the 17.0°C isotherm and the surface, was uncorrelated with  $\Delta z$ . The linear relationship between the vertically averaged Chl a and  $\Delta z$ , and the absence of a  $T - \Delta z$ relationship, do not depend on the choice of isotherm within the range 16.5°C to 17.2°C, associated with the sIW. This along-isotherm Chl a variation is not due to NPQ (see Web Appendix) because the maximum NPQ correction for Chl *a* on the 17.0°C isotherm was 9%, far smaller than the observed variability. The time window of 06:00 h to 17:00 h for the analysis was chosen such that the depth-averaged Twas quasi-stationary (with a time-mean of  $16.9 \pm 0.3^{\circ}$ C) as required by Lennert-Cody and Franks. A weaker Chl a patch (maximum = 22.3  $\mu$ g L<sup>-1</sup>) was also observed near 00:00 h centered over a 16°C isotherm trough (Fig. 4c). Because this occurred at night, it could not have been caused by NPQ, supporting the hypothesis that the daytime patch was also largely formed by the Lennert-Cody and Franks mechanism.

The significant correlation ( $r^2 = 0.57$ , p < 0.01), observed between the vertically averaged Chl *a* and depth of the 17°C isotherm  $\Delta z$  is consistent with the predictions and observations of LCF: Chl *a* (*L. polyedrum*-dominated for both Lennert-Cody and Franks and at HB06) was enhanced in wave troughs where isotherms are strained away from the surface. However, the mean Chl  $a-\Delta z$ relationship is qualitative because the amount of Chl *a* enhancement depends on the phytoplankton swimming proficiency (unknown) and direction (assumed) relative to rate of the vertical movement of the isotherms.

The intensity of the Chl *a* patch formed through the Lennert-Cody and Franks depth-keeping mechanism depends on the swimming proficiency, the ratio of phytoplankton swimming speed to internal wave vertical currents. Laboratory instantaneous swimming speeds of *L. polyedrum* range between 100  $\mu$ m s<sup>-1</sup> and 400  $\mu$ m s<sup>-1</sup> (Buskey 1997; Lewis and Hallet 1997; Mayali et al. 2008). Vertical velocities (*w*) between 10:00 h and 15:00 h estimated from vertical displacements of the 17.0°C isotherm were about 300–1500  $\mu$ m s<sup>-1</sup> (e.g., 2–9 cm min<sup>-1</sup>, *see* bold isotherm displacements; Fig. 6d). The partial overlap between *w* and the range of *L. polyedrum* swimming speeds suggests that the dinoflagellates could have partially or completely countered the sIW–generated vertical velocities as required by Lennert-Cody and Franks.

In addition to forming in the wave trough, it is possible that the Chl *a* patch was advected from offshore (H > 20 m) to M13 within the sIW trough. Nonlinear internal waves of depression transport particles horizontally above the wave trough in the direction of propagation (Lamb 1997). However, the sIW phase speed (0.12–0.17 m s<sup>-1</sup>) was large compared with the mean Eulerian velocity above the wave trough at each mooring  $(0.07 \text{ m s}^{-1} \text{ to } 0.05 \text{ m s}^{-1} \text{ at } M27 \text{ to})$ M13, respectively), so particles may have been transported a short distance but would quickly be left behind the wave trough. Thus, although the Chl a high concentration region formed by the Lennert-Cody and Franks mechanism may have moved onshore with the sIW trough, the organisms making up this Chl a patch were not advected across shore any appreciable distance (Franks 1997). These observations emphasize the potentially important role of swimming in generating Chl a patchiness, and show clearly that Chl a is not a passive tracer in this region of large amplitude, shoreward-propagating, internal waves.

Stage 3: Internal wave breaking—The sIW trough containing the intense Chl *a* patch was evident at M13 (H = 13 m, black star; Fig. 6d), but not at M8 (H = 8 m, white



Fig. 7. Vertically averaged Chl *a* above the 17.0°C isotherm vs. isotherm depth ( $\Delta z$ ) at 4-min intervals between 06:00 h and 17:00 h (number of data points, n = 165) on 12 October 2006). The linear fit (dashed line) has slope 1.29 ± 0.03 µg L<sup>-1</sup>m<sup>-1</sup>, and  $r^2 = 0.57$ , p < 0.01.

star; Fig. 6e) implying that sIW breaking had occurred. Two theoretical criteria for IW breaking were considered. At M13, the sIW amplitude  $a_o$  was ~ 5 m, based on the vertical displacement (crest to trough) of the 17°C isotherm (thick black contour; Fig. 6d). Moving into shallow water, the group velocity of a shoreward-propagating internal wave decreases, and  $a_o$  increases to conserve wave energy flux, until the wave breaks (Gill 1982). For an idealized two-layer system, breaking occurs when the undisturbed lower layer thickness  $h_{low}$  is  $< h_{bp}$  (Helfrich 1992) defined as

$$h_{bp} = \frac{a_o}{(1.6\frac{\lambda}{I} + 0.6)}$$
(3)

where  $\lambda$  is the wavelength and *L* is the distance to the shelf break. L = 8000 m at Huntington Beach, where H = 60 m depth. For sIWs the ratio  $\lambda : L$  is small, and  $h_{\rm bp}$  depends only weakly on  $\lambda$ . Using  $\lambda \sim 500$  m (estimated from a wave period  $\sim 2$  h) an idealized sIW with  $a_o = 5$  m would begin breaking when  $h_{\rm low} < h_{\rm bp} \sim 7$  m. Taking the lower layer thickness  $h_{\rm low}$ as the mean elevation above the seafloor of the 17.0°C isotherm where the sIW was most evident (*see* bold contour Fig. 6d) for 24 h prior to the sIW passage,  $h_{\rm low}$  was  $\sim 9$  m at M13, and according to Eq. 3, the sIW was not breaking. At M8 (H = 8 m),  $h_{\rm low}$  was always less than  $h_{\rm bp}$  (Fig. 6d,e), supporting the hypothesis that wave breaking had occurred.

For shoaling solitary internal waves of depression, breaking occurs when the wave-induced near-surface horizontal currents above the wave trough  $(u_{\text{max}})$  are equal to  $C_p$  (Lamb 2002). At each mooring,  $u_{\text{max}}$  was estimated as the maximum cross-shore current (of the 15-min averaged currents; e.g., Fig. 4b) in the box defined by the top of the wave trough to 3 m below the surface, and over the timescale of the wave (roughly 1–2 h). The ratios  $u_{\text{max}}$ :  $C_p$ from Eq. 2 between M27 and M13 vary from 0.4 to 0.8 (Table 2). Thus the Lamb (2002) breaking criterion  $u_{\text{max}}: C_p = 1$  suggests that the wave would not have broken by M13. This criterion was not investigated at M8 because the sIW was not evident and so an estimate of  $u_{\text{max}}$ was not possible. The observations and theory, thus, support the hypothesis that the sIW broke between M13 and M8 as it propagated onshore.

Stage 4: Cross-isotherm and vertical spreading of the intense subsurface Chl a patch—The shoreward-propagating IW breaking (anticipated to reach H < 13 m at  $\sim 12:30$ h), may have mixed the water column vigorously enough for the subsurface Chl a patch to reach the surface, where it was visually apparent (Fig. 1) and appeared in the jet-ski maps (Fig. 5c,e) as a 200-m-wide, surface Chl a band onshore of M13. The surface Chl a band appeared around 12:45 h on an isotherm warmer ( $T = 17.5^{\circ}$ C) than the isotherms associated with the subsurface Chl a patch (16– 17°C; Fig. 5c) suggesting that cross-isotherm mixing had also occurred. Mixing induced by breaking sIW was illustrated by examining the cross-isotherm and vertical spreading of the subsurface Chl *a* patch. The temporal evolution of the T-z and Chl a-T relationships is quantified using objective maps of Chl a and T from 12:00 h to 15:00 h, within the 750-m-wide region between H = 5-13 m. This region bounds the Chl *a* patch (Fig. 5), includes the estimated time and location of sIW breaking (white star in Fig. 6e), and excludes the warm shallow water region in the surf-zone (x < 200 m, H < 5 m; Fig. 5). To minimize the effects of variability due to cross-shore advection on our analyses, CTD + Chl *a* cast-derived Chl *a* and *T* for each nearshore survey on 12 October were cross-shore–averaged (denoted <Chl a(z)> and <T(z)>), into 1-m-thick vertical bins.

The depth-average of <Chl a(z)> remained roughly constant (within 15%) between 12:00 h and 15:00 h, consistent with small depth-integrated cross-shore Chl *a* fluxes, and small net phytoplankton growth and mortality over these 3 h. The depth-averaged <T(z)> warmed slightly, from 16.9°C to 17.1°C. For each survey <T(z)> were approximately linear with depth (not shown). The vertical temperature gradient d < T(z) > /dz decreased monotonically with time, and the corresponding buoyancy frequency (N) decreased from 0.018 s<sup>-1</sup> at 12:00 h to 0.016 s<sup>-1</sup> at 15:00 h, suggesting that vertical mixing dominated any stratifying effects of solar warming that would tend to increase N over the afternoon.

In addition to a reduction in *N*, cross-isotherm Chl a(z) mixing also was evident in the evolution of the cross-shore– averaged <Chl a(z)>–<T(z)> relationship (Fig. 8a). For all surveys, <Chl a(z)> was maximum near T = 16.9°C. At 12:00 h, <Chl a(z)> was concentrated over a narrow range of T(z), and throughout the afternoon, <Chl a(z)> spread over a larger range of T(z) (e.g., compare the dashed gray with the solid black curve in Fig. 8a). Although the CTD + Chl asurveys exclude the top 1 m of the water column, the increased <Chl a(z)> at <T(z)> > 17.3°C after 12:00 h was in qualitative agreement with the surface appearance of the Chl a band associated with 17.5°C water at the surface (e.g., Fig. 5e–j).

To quantify the vertical spreading of the subsurface <Chl a(z)> layer, which includes cross-isotherm mixing, the <Chl a(z)> center of mass ( $<z>_{Chla}$ ) was calculated (Csanady 1973):

$$\langle z \rangle_{\text{Chl}a} = \frac{\int_{-H}^{0} z \langle \text{Chl}a(z) \rangle dz}{\int_{-H}^{0} \langle \text{Chl}a(z) \rangle dz}$$
(4)

The center of mass deepened linearly from -7.1 m to -7.8 m (Fig. 8b) over 3 h at an average rate of 60  $\mu$ m s<sup>-1</sup>. The deepening of the Chl *a* layer peak is qualitatively evident in the CTD + Chl *a* sections (Fig. 5a,c,e,g,i). The Chl *a* layer squared half-width ( $\sigma^2$ <sub>Chla</sub>; Csanady 1973),

$$\sigma_{\text{Chl}a}^2 = \frac{\int_{-H}^{0} (z - \langle z \rangle_{\text{Chl}a})^2 \langle \text{Chl}a(z) \rangle dz}{\int_{-H}^{0} \langle \text{Chl}a(z) \rangle dz}$$
(5)

increased roughly linearly with time until 14:15 h (solid line; Fig. 8c), consistent with the presence of breaking sIW–generated turbulence. The layer width  $(2\sigma)$  grew from 4.2 m at 12:00 h to 4.7 m at 14:15 h, reflecting the qualitative Chl *a* vertical spreading seen in the CTD + Chl *a* sections (Fig. 5a,c,e,g,i). The slowing of the Chl *a* vertical spreading by 15:00 h may be attributed to subsiding of the breaking sIW–generated mixing.



Fig. 8. (a) Chl *a* vs. *T* (at 12:00 h, 13:30 h, 14:15 h, and 15:00 h) based upon the CTD + Chl *a* surveys (between H = 5-13 m depth; *see* Fig. 4) horizontally averaged along 1-m isobars. Timeseries of (b) depth  $\langle z \rangle_{Chla}$  of the center of mass (COM) of cross-shore averaged  $\langle Chl a (z) \rangle$  and, (c) squared half-width  $\sigma^2_{Chla}$  (Eq. 5) of the Chl *a*(*z*) layer from 11:30 h to 15:30 h on 12 October. The red arrow illustrates the trend of increasing spread of Chl *a* across *T* and *z* with time.

Although phytoplankton swimming is implicated in Chl *a* patch formation above the sIW trough at M13 (Fig. 7), swimming was unlikely to have accounted for the vertical spreading of the Chl *a* patch in H < 13 m. Chl *a* vertical spreading rate owing to random phytoplankton swimming (akin to a random walk) can be scaled by  $1/2 \times \tau \cdot \hat{u}_s^2$  where  $\hat{u}_s$  is a typical swimming speed, and  $\tau$  the de-correlation time of swimming direction. This is effectively a diffusivity, which can be compared with a vertical diffusivity  $\kappa_{zz}$  calculated from changes in the squared Chl *a* layer width  $\sigma^2$  (Csanady 1973):

$$\kappa_{zz} = \frac{1}{2} \frac{\partial \sigma_{\text{Chla}}^2}{\partial t} \tag{6}$$

The best-fit linear slope of  $\sigma^2_{\text{Chla}}$  between 12:00 h and 14:15 h (dashed line; Fig. 8c) yields  $\kappa_{zz} = 0.5 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ . Assuming  $\hat{u}_s \sim 200 \ \mu\text{m} \text{ s}^{-1}$ ,  $\tau = 40 \ \text{min}$  is required to reproduce the observed  $\kappa_{zz}$ . This  $\tau$  is implausibly large, and would induce ballistic dispersion (e.g.,  $\sigma^2_{\text{Chla}} \sim t^2$ , Csanady 1973), instead of the observed linear (Brownian) relationship (Fig. 8c). The timing of the surface Chl *a* band appearance is consistent with the expected arrival time of the sIW trough in shallow water (~ 12:30 h). The evidence supports our hypothesis that the breaking of the sIW in this region drove the vertical mixing that led to the vertical spreading of the Chl *a* patch to the surface.

Stage 5: A density barrier to Chl a entry into the surfzone—During the jet-ski-CTD observation period, the surface Chl *a* band did not enter the surf-zone region (0 < x < 150 m) where the near-surface water was warmer than the near-surface offshore waters (*see* Fig. 5). Here, Chl *a* and *T* observations from very shallow water (H < 5 m) are presented, and a potential mechanism for the low surf-zone Chl *a* is discussed.

Between 10:00 h and 12:00 h, near-surface T was uniform at roughly 17°C from within the surf-zone (M1.5) to M13 (Fig. 9a). Coincident with the M13 subsurface Chl a peak (Fig. 3d), Chl a at M4 (seaward of the surf-zone) was also elevated (between 5  $\mu$ g L<sup>-1</sup> and 10  $\mu$ g L<sup>-1</sup>), with the maximum Chl  $a \sim 25\%$  of that at M13 (Fig. 4d, solid black line; Fig. 9b). The M4 dinoflagellate (mainly L. polyedrum) cell concentrations also reached a maximum of  $7.3 \times 10^4$  cells L<sup>-1</sup> near 11:30 h (solid black line; Fig. 9c), while diatom abundance remained low  $(O[10^3]$  cells L<sup>-1</sup>, dashed black line; Fig. 9c). The qualitative agreement between dinoflagellate cell concentration and Chl a at M4 indicates that the Chl a variation was due to variation in total dinoflagellates. Within the surf-zone, Chl a was lower at about 3–4  $\mu$ g L<sup>-1</sup> (Fig. 9b), consistent with the jet-ski–observed surface Chl a band values in that region at 12:00 h (Fig. 5a). In the surf-zone, dinoflagellate and diatom concentrations were low, usually  $< 10^4$  cells  $L^{-1}$ , and relatively constant throughout the sample period (Fig. 9c).

At 11:30 h, the M4 Chl *a* began dropping and by 12:30 h reached the level at M1.5 (3  $\mu$ g L<sup>-1</sup>). M1.5 and M4 Chl *a* remained constant throughout the afternoon as *T* at both locations rose (Fig. 9a,b). During midday, the near-surface



Fig. 9. Time series from 09:30 h to 18:00 h on 12 October of (a) T (1-min averages), (b) Chl a (1-min averages), and (c) dinoflagellate (solid) and diatom (dashed) concentration from M4 (H = 4 m) and M1.5 (H = 2 m), and in (a) 20-min average T, 2 m below the mean sea surface at M13 (H = 13 m; *see* legend). The CTD + jet-ski sampling period (12:00–15:00 h; Fig. 5) is indicated by the gray bar between vertical dashed lines.

T at M13 also increased slightly (thin curve in Fig. 9a). By 15:00 h, a cross-shore temperature gradient (with  $\Delta T \sim 1^{\circ}$ C) had developed between the warm surf-zone water at M1.5 and the near-surface water M13, consistent with surface T maps (Fig. 5j). A strong afternoon T gradient ( $\Delta T \sim 0.5^{\circ}$ C) was also maintained between M1.5 and M4 over only 80 m in the cross-shore.

The rapid and large T increase at M1.5 over 4 h cannot be explained by solar heating of the water column, and was much larger than the daily heating cycle normally observed near-surface at M13. The warm  $\geq 18^{\circ}$ C surf-zone water present at M1.5 and M4 after 13:00 h likely originated from the Talbert Marsh, located 1 km South (+y) of the fixed instrument array (see Fig. 2a). This warm water created a density barrier to Chl *a* entry into the surf-zone from offshore. Before 12:00 h, the flood tide (Fig. 4a) raised the marsh water level. Over the morning and early afternoon, the shallow marsh water ( $H \sim 2$  m) was strongly solar heated relative to near-surface waters at M13. Based on the marsh area (10<sup>5</sup> m<sup>3</sup>; Grant et al. 2001), tidal prism (2.35 × 10<sup>5</sup> m<sup>3</sup>; Jeong et al. 2008) and the solar insolation, the marsh water would have heated ~ 0.8°C. With the ebb tide at 12:00 h, the warmed marsh water began draining into the surf-zone. The 0.15 m s<sup>-1</sup> northward surf-zone alongshore current (*see* Fig. 5), driven by obliquely incident breaking surface gravity waves (Longuet-Higgins 1970), transported this water to the study region. The predicted increase of ~ 0.8°C is roughly consistent with the near-surface *T* gradient



Fig. 10. Schematic of nearshore (H < 20 m; see black box on Fig. 3d) summarizing the stages of the surface Chl *a* band formation. (a) An intense Chl *a* formed subsurface within 7 d preceding the surface appearance. (b) Chl *a* in an sIW trough was locally concentrated by swimming and sIW strain as the wave propagated onshore, (c) sIW breaking in shallow water caused vertical mixing and outcropping of the 17.5°C isotherm and upper portion of the Chl *a* patch. On ebb tide, warm water carried by the surf-zone alongshore current from the Talbert marsh, blocked Chl *a* entry into the surf-zone.

that developed between M1.5 and M13, and coincided with the reduction of Chl a at M4 (Fig. 9a,b). If the warm water had not been present, the intense Chl a patch may have entered the surf-zone.

Synthesis—The surface Chl a appeared on 12 October through local vertical mixing of a subsurface red-tide layer

(dominated by Lingulodinium polyedrum) that had developed in the study region over the 7 d prior (Fig. 10a). The km-long shore-parallel surface Chl a band (maximum Chl a = 7  $\mu$ g L<sup>-1</sup>) persisted for ~ 3 h at depths H = 6-8 m with a cross-shore width of 200 m. At M13, an elevated Chl a patch was observed in the trough of a supertidal internal wave (sIW) that propagated coherently across the mooring array (Fig. 10b). The elevated Chl a in the sIW trough is consistent with patch formation by phytoplankton swimming interacting with a IW-induced convergent flow. Based on the estimated phase speed, the sIW arrived in shallow water (< 13 m) around 12:30 h, where wave breaking occurred. Turbulence from the breaking sIW led to vertical mixing and the appearance of the subsurface Chl *a* patch at the surface. The wave arrival time (12:30 h), the approximate patch width (200 m), and depth of anticipated wave breaking  $(H \sim 7 \text{ m})$  were all consistent with the appearance and dimensions of the surface Chl a band (Fig. 5). Finally, warm water exiting the Talbert marsh created a density gradient that inhibited the Chl a patch from entering the surf-zone (Fig. 10c).

The high spatially and temporally resolved observations elucidated the dynamics linking a 2-km-wide subsurface red-tide layer to the brief surface appearance of the red tide in a narrow (200 m), along-shore parallel Chl a band. The spatial and temporal scales of surface Chl a were roughly an order of magnitude smaller and shorter than those of the subsurface Chl a layer. The change in scale from the subsurface to the surface highlight the challenges of redtide monitoring. However, an analysis of the physical mechanisms underlying the surface Chl *a* band appearance provided a means of connecting it to the subsurface Chl a field and offered insight into some drivers of short timescale, small spatial-scale nearshore Chl a variability. The highly resolved sampling across a number of different platforms-unique to this field study-was necessary to capture this broad range of nearshore Chl a patchiness.

The mechanisms underlying the surface Chl a band formation demonstrate that the drivers of Chl a distributions in the nearshore are complex. The interaction between motile phytoplankton and propagating internal waves, breaking internal waves, and surface-wave-driven surfzone currents may create striking spatial and temporal changes in Chl a patchiness and community structure over O(100) meters and hourly timescales. Similar dynamics may be anticipated in a broad range of nearshore systems where motile phytoplankton are present, where the stratification is sufficient to support internal waves, and where depth-limited internal wave breaking occurs. These mechanisms that rapidly concentrate, advect, and mix nearshore phytoplankton change the intensity and duration of shoreline exposure to red-tide organisms, and may have large consequences for nearshore ecology.

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