A dual-tracer approach to estimate upwelling velocity in coastal Southern California

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A R T I C L E   I N F O

Article history:
Received 19 December 2014
Received in revised form 18 March 2015
Accepted 9 April 2015
Available online 28 April 2015
Editor: J. Lynch-Stieglitz

Keywords:
upwelling
particle export
radioactive tracers

A B S T R A C T

The distribution of the cosmogenic radionuclide $^{7}$Be ($t_{1/2} = 53$ d) in the surface ocean has previously been used to estimate upwelling velocity in the open ocean. However, the loss of $^{7}$Be to particle export has limited this approach in high particle density environments like the continental margins. In this study, we combine a mass balance of $^{7}$Be with a $^{234}$Th budget in the surface ocean to constrain the loss of $^{7}$Be to particle sinking at the San Pedro Ocean Time-series (SPOT) in the inner Southern California Bight during spring 2013. Upwelling velocities (all in m d$^{-1}$) determined from the $^{7}$Be mass balance were observed to increase from 0.5 ± 0.6 in January to 2.5 ± 1.3 in May, then decrease to 1.2 ± 0.5 in June. These results agree within uncertainty with upwelling velocities derived from the monthly Bakun Upwelling Index, which ranged from 0.1 to 2.8 m d$^{-1}$, supporting the pressure-field-based approach. Evidence from a heat budget and the nutrient distribution over the course of the study supports that the upwelling signal at SPOT (20 km offshore) is not transported from coastal upwelling near shore, but instead is dominantly a local signal, likely driven by wind-stress curl.

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1. Introduction

The inventory and distribution of the cosmogenic radionuclide, $^{7}$Be ($t_{1/2} = 53$ d), in the surface ocean has been used as a tracer of circulation, including upwelling velocity into the upper thermocline (Kadko and Johns, 2011; Haskell II et al., 2015). Because of its typically small magnitude, upwelling velocity has been notoriously difficult to measure and most chemical tracers (i.e. $^{14}$C, $\delta^{13}$C, Apparent Oxygen Utilization (AOU), $\delta^{3}$He, and pCO$_{2}$) previously used to estimate its magnitude have long lifetimes in the water column (Broecker and Peng, 1982; Broecker et al., 1978; Quay et al., 1983; Wanninkhof et al., 1995; Klein and Rhein, 2004). The mean lifetime of $^{7}$Be in the water column is on the order of weeks to months, thus it is able to capture the dynamic range of upwelling over the timescale of biological production. Previous applications of this approach have been limited to open ocean regimes because the loss of $^{7}$Be adsorbed onto particles must be better constrained in higher particle flux environments (Haskell II et al., 2015).

Coastal upwelling zones are some of the most productive regions in the world’s oceans, accounting for 30–50% of global marine primary productivity (Jahnke, 1996). Future climate forcing may affect upwelling patterns and water column stratification (Levitus et al., 2001; Barnett et al., 2001). In order to better understand the control that upwelling of deep-water nutrients has on marine productivity over the continental margins and how these ecosystems may respond to future climate change, constraints on upwelling velocities on the timescale of bloom evolution are needed in these highly productive surface ocean regimes.

The Southern California Bight (SCB) is located between Point Conception to the North and the US/Mexico border to the South (Fig. 1). From winter to summer, the surface currents increase in magnitude from the northwest as a portion of the California Current (CC) splits off into the inner SCB. The combined effect of the Coriolis acceleration and wind stress drive net offshore transport of surface water, causing both near-shore coastal upwelling and wind stress curl-driven upwelling (as defined by Rykaczewski and Cheekley, 2008; Jacox and Edwards, 2011). This effect is strongest during spring and early summer. During the late summer and early fall, prevailing wind directions shift, and the Southern California Countercurrent (SCC) begins to flow from the Southeast against the CC. During fall and early winter months, the SCC dominates the surface current flow, upwelling velocity is minimal and the surface ocean returns to a more eddy diffusion-dominated transport system. This seasonal upwelling pattern is reflected in the Bakun Upwelling Index, an estimate of offshore Ekman transport driven by geostrophic wind stress, based on measurements of atmospheric pressure fields (Bakun, 1973; Schwing et al., 1996). However, the Bakun Upwelling Index may reflect large-scale off-shore
mass transport, rather than local upwelling dynamics, since it is derived from pressure gradients over large distances (~300 km).

In this study, we use a dual-tracer approach to estimate upwelling velocity in the upper thermocline in San Pedro Basin from January to June of 2013. Measurements of $^7$Be inventory, primarily sensitive to loss by upwelling, are combined with those of $^{234}$Th, primarily sensitive to particle export, to apply these tracers in a high particle flux environment. These measurements are part of a larger effort to characterize the biological response to upwelling, the Upwelling Regime In-Situ Ecosystem Efficiency (Up.R.I.S.E.E.) study, located at the San Pedro Ocean time-series (SPOT) station (33°33′N, 118°24′W).

2. Material and methods

2.1. Beryllium-7

Two depths were sampled for $^7$Be at the SPOT hydrostation (Fig. 1), approximately every two weeks from January to June 2013. Samples from the mixed layer (40 L) and 10 m below the mixed layer (80 L), as determined from temperature profiles during CTD descent, were taken with a CTD equipped with twelve, 12 L Niskin bottles. In this study, we define the base of the mixed layer as the depth where the temperature differs from the surface value by 0.5°C. Water from each depth was transferred into 10 L centrifuge tubes and taken back to the lab. Each centrifugation was spiked with a known amount of $^8$Be used as a yield tracer and equilibrated for ~24 h, then acidified with 12.5 mL of 10 N HCl to pH of <2 and bubbled with N2 for at least one hour to drive off all inorganic carbon. FeC2 (50 mg) was added to each centrifugation and allowed to mix thorough for one hour before 12.5 mL of NH4OH was added to raise the pH and co-precipitate $^7$Be and Fe(OH)$_3$. Each centrifugation was allowed to settle for 2 days, ~8 L of solution was decanted off the top, and then the remaining solution containing the precipitate from all centrifugations for a single depth was combined into a single centrifuge tube and centrifuged down to fit into a counting tube.

Each sample was counted on an Ortec low-background intrinsic Germanium (Ge) gamma detector (well-type, 100cc active volume). $^7$Be has a readily identifiable gamma peak at 478 keV. Detection efficiency was determined by measuring gammas emitted by standards of known activity. The effect for sample geometry was found by counting a solution of known gamma activity in sample tubes filled to different heights. The uncertainty of extraction efficiency and the detector efficiency was in all cases smaller than the statistical counting error and the uncertainty in the blank. Recovery yield of $^8$Be was measured following gamma counting by first re-dissolving the Fe(OH)$_3$ precipitate in a 10% nitric acid matrix, equilibrating with BIO-RAD AG 50W-X12 cation-exchange resin to remove Fe (von Blankenburg et al., 1996), and then analyzing the Be eluate on a Microwave Plasma Optical Emission Spectrometer (MP-OES) to measure recovered $^8$Be. This recovery yield was used in all $^8$Be calculations to correct for methodological efficiency (Mean: 0.88; Standard Deviation: 0.13). All water column $^7$Be activities are reported in Table 1.

The activity of $^7$Be in rain ($A_{Be}$) was measured on rain collected by bucket on the rooftop of one of the tallest buildings (~6 stories tall; separated from other tall buildings) on the USC campus intermittently throughout the study period. The same Fe(OH)$_3$ co-precipitation and isotopic dilution procedures used to process seawater samples were performed on the rainwater, and they were counted on the same gamma detectors (average activity = 201 dpm L$^{-1}$; Table S1).

2.2. Thorium-234

Vertical profiles from the surface to 200 m were collected for thorium via Niskin/CTD on every cruise. Ten liters were sampled at eight to ten depths, chosen based on fluorescence measured during the CTD’s descent. The analysis of each sample follows the method published in Haskell II et al. (2013). An isotope dilution method using a $^{226}$Th spike of known activity and co-precipitation with MnO$_2$ was used to measure $^{234}$Th (Rutgers van der Loeff and Moore, 1999; Mean Yield: 0.83, Standard Deviation: 0.15). The $^{234}$Th deficit relative to its parent isotope, $^{238}$U, was calculated by trapezoidal integration of the profile. $^{238}$U activity was calculated as a function of salinity ($^{238}$U activity (dpm L$^{-1}$) = 0.07097 * Salinity; Chen et al., 1986).

2.3. Sediment traps

During 4 cruises, one surface-tethered drifting sediment trap of the Particle-Interceptor-Traps (PITs) design was deployed at 100 m with a total surface area of 0.0851 m$^2$ (Knauer et al., 1979). The trap had 12 collection tubes with an aspect ratio of 6.4 and 1 cm x 1 cm baffles fitted into the opening of the tubes. Funnels with centrifuge tubes were attached into the base of each trap tube. The centrifuge tube contained a brine solution of NaCl (in excess of sea water by 5 ppt) that was poisoned with 3% formaldehyde and buffered with disodium tetraborate. Deployments lasted ~24 h. After recovery, each of the twelve centrifuge tubes were put through a series of centrifuging/descanting/diluting cycles to remove all salts, then combined onto Whatman Nuclepore polycarbonate membrane (0.4 µm) filters and dried at room temperature. All material from each trap was folded and put in a single counting tube and counted on the same counters as the other $^7$Be and $^{234}$Th samples.

2.4. Nutrients

Dissolved nutrient samples were collected via Niskin bottles at 12 depths from the surface to 400 m and filtered through 0.8/0.2 µm syringe filters into two Nalgene bottles for each depth. One was used for silica and phosphate analyses done colorimetrically at USC with a Hitachi UV/vis-spectrophotometer (Parsons et al., 1984). Nitrate samples were collected in acid-washed bottles and frozen at ~20°C, until analysis. Nitrate concentrations were determined by converting nitrate to N$_2$O (Sigman et al., 2001) and quantifying the amount of N$_2$O as integrated sample areas on an Isotope Ratio Mass Spectrometer (IRMS) in the D. Sigman lab at Princeton University. Prior to analyses, nitrite was chemically removed from samples (Grager and Sigman, 2009).
2.5. Bakun Index upwelling velocities

The Bakun Upwelling Index (Bakun, 1973) outputs reported in this study are the monthly indices provided by the Pacific Fisheries Environmental Laboratory (PFEL) through their NOAA Live Access Server website (NOAA, 2013a). Monthly indices describe the offshore Ekman transport of surface waters with positive values being offshore and negative values representing onshore transport, reported in m$^3$ s$^{-1}$ 100 m coastline$^{-1}$. Assuming that water transported offshore is replaced by deeper upwelled waters near the coast, the volume of water transported offshore equals the volume of water upwelled to replace it. The monthly indices use a global 3° spherical mesh that is interpolated from the monthly-averages of six-hourly 1° pressures. To calculate the pressure-gradient forced geostrophic wind stress, the first derivative of surface pressure at the point of interest is estimated by taking the difference in pressure between grid points (3° apart) on either side of the point of interest in both the N/S and E/W directions, then dividing by the 6° angular mesh length for each direction, giving pressure gradients per degree in both zonal and meridional directions (Schwing et al., 1996). Coincidentally, the closest grid point to our study area is also ~1° from the coast (centered at 33°N, 119°W, ~10 km from our study site). Assuming the Bakun Index at this location is a calculation of vertical transport (in m$^3$ s$^{-1}$ 100 m coastline$^{-1}$ degree$^{-1}$), at a location 1° from the coast (~100 km), we use the indices to calculate an average upwelling velocity per 10$^3$ m$^2$ (100 m coastline ≈ 100 km) of sea surface in a region bounded by 36°N, 30°N, 122°W and the coastline. Smaller scale spatial heterogeneity cannot be detected using this approach, since it represents an average found throughout the region, although it is likely to be patchy.

3. Calculation

3.1. Be box model

Following the formulation in Haskell II et al. (2015), we used a simple mass balance of $^7$Be for the surface mixed layer of depth $H$ with uniform concentration ($C_H$). Taking $z$ as positive downward, the change in concentration of $^7$Be over time (per m$^2$ of ocean) in this box should be given by:

$$\frac{\partial C_H}{\partial t} = F_a - J - \lambda C_H + w_H C_H - P_{Be}$$  (1a)

where $J$ represents vertical flux through the bottom of the box by advection and diffusion:

$$J = w_H C_H - K_z \frac{\partial C}{\partial H}$$  (1b)

Other terms include $F_a =$ atmospheric input flux, $K_z =$ eddy diffusivity, $\frac{\partial C}{\partial z} =$ concentration gradient with depth, $\lambda C_H =$ radioactive decay within the surface box (where $\lambda$ is the decay constant for $^7$Be), $P_{Be} =$ particle export of $^7$Be, and $w_H C_H =$ horizontal export as upwelled water exits the mixed layer in horizontal flow. In practice, the $^7$Be activity measured in the mixed layer is assumed equal to that at the base of the mixed layer. Therefore, the concentration of $^7$Be in the water transported across this boundary ($C_H$) into the surface box is the same as the water inside the box ($C_H$). Because depth as taken as positive downward ($z \geq 0$), upwelling $w$ is ≤ 0, although the negative sign will be dropped in tables where it is given as upwelling velocity. Note that the effective horizontal export of the tracer from the mixed layer must have the equivalent velocity and the opposite sign of the advective input from upwelling if the surface layer is a constant thickness and well-mixed.

If $w$ and $K_z$ are assumed constant below the mixed layer, $^7$Be should decrease exponentially below the mixed layer with a depth attenuation coefficient of $\alpha$. This value is determined by curve fit of two points, which introduces a considerable amount of uncertainty into our calculations, and relies on the assumption that $^7$Be exhibits an exponential profile beneath the mixed layer. However, we are confident that this is a reasonable assumption since previous studies in many locations have shown an exponential distribution below the mixed layer (Silker, 1972; Kadko, 2009). The uncertainty in the exponential fit to determine $\alpha$ is included in our uncertainty estimates described below. As shown in detail in Appendix A, combining Eqs. (1a) and (1b), assuming steady-state and integrating over the entire water column yields:

$$w_H = \frac{1}{C_H} (J + \lambda H C_H - F_a + P_{Be}) = \frac{\lambda}{\alpha} + \lambda H = \frac{F_a + P_{Be}}{C_H}$$  (2)

The terms in parentheses represent the difference between the inventory of $^7$Be activity in the upper ocean and the input from the atmosphere ($F_a$). If the inventory is less than the input, $w < 0$, implying upwelling. In this model, we assume that all atmospherically derived $^7$Be is either soluble or adsorbed to the sinking fraction of particles. Any horizontal export of $^7$Be on suspended particles is considered negligible in the budget.

The uncertainty in the $^7$Be inventory was estimated using a Monte Carlo approach, which determined the inventory for each sampling date with 10,000 simulations of randomly selected values for each data point that were assumed to be normally distributed about the measurement, based on the analytical uncertainty for the sample. The upper and lower bounds for inventory were then calculated as one standard deviation from the best estimate, determined as the percentile below which 16% and 84% of the simulated inventories fell. Three runs of 10,000 simulations each were

Table 1

<table>
<thead>
<tr>
<th>Cruise ID</th>
<th>Date</th>
<th>MLD (m)</th>
<th>$^7$Be activity ML (dpm m$^{-3}$)</th>
<th>$^7$Be act. 10 m BML (dpm m$^{-3}$)</th>
<th>$^7$Be inventory* (dpm m$^{-2}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SPOT 42</td>
<td>1/16/13</td>
<td>15</td>
<td>43</td>
<td>28</td>
<td>36</td>
</tr>
<tr>
<td>Up1</td>
<td>2/14/13</td>
<td>35</td>
<td>125</td>
<td>24</td>
<td>18</td>
</tr>
<tr>
<td>Up2</td>
<td>2/20/13</td>
<td>28</td>
<td>67</td>
<td>17</td>
<td>15</td>
</tr>
<tr>
<td>UP3</td>
<td>3/14/13</td>
<td>35</td>
<td>71</td>
<td>16</td>
<td>6</td>
</tr>
<tr>
<td>Up4</td>
<td>4/3/13</td>
<td>10</td>
<td>89</td>
<td>16</td>
<td>74</td>
</tr>
<tr>
<td>Up5</td>
<td>4/25/13</td>
<td>13</td>
<td>75</td>
<td>33</td>
<td>54</td>
</tr>
<tr>
<td>Up6</td>
<td>5/10/13</td>
<td>12</td>
<td>96</td>
<td>36</td>
<td>45</td>
</tr>
<tr>
<td>Up7</td>
<td>5/23/13</td>
<td>10</td>
<td>52</td>
<td>28</td>
<td>7</td>
</tr>
<tr>
<td>Up8</td>
<td>6/20/13</td>
<td>12</td>
<td>54</td>
<td>18</td>
<td>21</td>
</tr>
</tbody>
</table>

* The water column inventory of $^7$Be calculated from mixed layer activity and integrating under an exponential curve fit below the mixed layer. Uncertainty is reported as one standard deviation calculated using a Monte Carlo simulation of water column inventories using normally distributed possible combinations of ML and BML values. The uncertainty is skewed toward higher values when the difference between mixed layer and sub-mixed layer measurements is small, due to the high sensitivity of an exponential curve fit of two values when their uncertainties overlap.
done with statistical ranges that varied by less than 2%, and the averages of these ranges are reported in Table 1. However, the distribution of the calculated inventories is non-Gaussian and skewed toward higher values, resulting in an imbalance in uncertainties for most sampling dates. In these simulations, we used a maximum bin value of 10,000 dpm m⁻² for all sampling periods, except March 14 (the largest best estimate inventory), where we used 20,000 dpm m⁻². This value is greater than the largest inventory possible, with the stipulation that the deeper activity could not exceed the value in the mixed layer for each sample. The imbalance in the distribution of uncertainty is the result of the exponential fit and overlapping of uncertainties in sampling points on a few occasions, which causes the inventory to approach infinity when the deep activity is greater than in the mixed layer. All parameters in the ⁷Be balance are shown in Tables 1, 2 and 3.

3.2. Atmospheric ⁷Be input

The input flux of ⁷Be by atmospheric fallout is dominated by wet precipitation globally (~80%; Kadko and Prospero, 2011), however a portion of the input flux is contributed from dry deposition (~10–38%; Fogh et al., 1999; McNearney and Baskaran, 2003). For this study, we use Eq. (3a) to define the total depositional flux of ⁷Be (Fₐ) over the previous 24 days, which contributes to the water column ⁷Be inventory at the time of sampling:

\[ Fₐ = F_{wet} + F_{dry} \]  

with the subscripts 'wet' and 'dry' referring to the effective wet and dry input fluxes defined in Eqs. (3b) and (3c):

\[ F_{wet} = \frac{0}{-24} \sum R_t \cdot A_{Be} \]  

The wet ⁷Be input flux by each rainfall event that contributed to the water column inventory at the time of sampling (F_wet) was estimated by dividing the product of the ⁷Be activity in the rain (A_{Be}) and a regional average of rainfall estimates for that event (R_t) by the time between the rainfall event and the sampling date (t; days). This term reflects the time-averaged input flux of rain into the water column from any given rainfall event over the period of time between the event and the sampling day. The total effective ⁷Be input by wet deposition at the time of sampling (F_wet) was then calculated by summing the average flux contributed by each event within 24 days prior to the sampling date. The A_{Be} measured closest to the day of each rainfall event was used in Eq. (1b) in the first half of the season and the average of all A_{Be} measured was used for each event in the latter half of the season, due to a lack of rain samples collected during that period.

As a boundary condition, we only include rainfall events within 24 days of each sampling date in our input function. This is based on a calculation of the residence time of ⁷Be in the upper water column of the inner SCB, calculated by adding the loss rates of ⁷Be due to radioactive decay (1.3% d⁻¹; λ_{Be} = 0.013 d⁻¹), horizontal export due to upwelling (5% d⁻¹ at a velocity of 1 m d⁻¹ from wind-speed estimates; Schwab et al., 1996), and export on particles (2% d⁻¹; see Tables 1 and 2). This equals 12 days on average. Therefore, ~90% of the ⁷Be inventory is completely flushed from the region in 24 days. This is comparable to published residence times of surface water in the inner SCB (Hickey, 1992). We
acknowledge that this is an oversimplification of more complex dynamics but we believe it represents the behavior of this system reasonably well.

For the rainfall average ($R_i$), we used rainfall estimates at Santa Monica Municipal Airport, Los Angeles International Airport, Long Beach Airport, Santa Barbara Municipal Airport, Oxnard Airport, Oceanside Municipal Airport, San Nicolas Island and the Avalon Airport on Catalina Island (NCDC, 2013; mean for each event was 0.7 cm and the mean standard deviation between sites was ~43%). These do not differ significantly from rates at Monterey and San Diego. For the April 25 sampling, we used only the rainfall measured in Santa Barbara, since the surface currents were determined to be coming predominantly from the north for 3 weeks prior to the sampling date using the JPL Regional Ocean Models (ROMS) output (NASA, 2013). We do not consider rainfall from further north than Santa Barbara, as the surface 7Be concentration in the California Current has been shown to be fairly uniform between 20° N and 40° N (Silker, 1972).

To estimate input flux due to dry deposition, we adopted Eq. (3c), which estimates the reloading of 7Be associated with aerosols in the atmosphere following each rain event, similar to the equation presented by Caliliet et al. (2001). We assume that this value is equal to the dry depositional flux into the water column ($F_{dry}$):

$$F_{dry} = (r_{max}) \times (1 - e^{-k_r t})\tag{3c}$$

with $k_r$ being the reload rate, $t$ representing time (days) between the last rainfall event and sampling date, and $r_{max}$ being a maximum dry deposition rate, which is a step function that reflects seasonal variability in the density of aerosol particles, as shown by satellite aerosol optical depth from the wet winter season to the dry summer season (NASA, 2014), and the reported 7Be concentration in marine air masses (Bourcier et al., 2011; Al-Azmi et al., 2001). We use an $r_{max}$ of 50 dpmm$^{-2}$ d$^{-1}$ in January and February, 100 dpmm$^{-2}$ d$^{-1}$ in March, April, and early May, 120 dpmm$^{-2}$ d$^{-1}$ in late May and June (the maximum value being half of the atmospheric production rate of 240 dpmm$^{-2}$ d$^{-1}$; Nagai et al., 2000). The rate constant of inventory growth ($k_o$) should be equal to 1 divided by the residence time of 7Be in the atmosphere, which has been estimated previously by Yu and Lee (2002) to be between 3 to 12 days, Rosner et al. (1996) to be 6 days and by Papastefanou and Bondietti (1991) to be between 7 to 9 days. 2013 was an anomalously dry year in Southern California (~3× lower rainfall than the previous 30 yr mean). In light of low wet deposition of 7Be, we estimate the residence time of 7Be on aerosols in this region to be 10 days ($k_o$ value of 0.10 d$^{-1}$), slightly longer than the average of reported empirically derived values. During two periods of heavy rainfall within a week of our sampling dates (wet deposition >2× the seasonal average), we assume dry deposition was completely washed out by precipitation. The average dry deposition rate over the course of this study was 52 dpmm$^{-2}$ d$^{-1}$, 21% of the total 7Be input flux, which agrees well with previously published dry/total fluxes (22%; Tanaka and Turekian, 1995; 23% Pham et al., 2013). Table S2 in the supplementary material summarizes dry deposition rates from 13 different locations in northern mid-latitudes, which average 51 dpmm$^{-2}$ d$^{-1}$. Values for $F_{wet}$, $F_{dry}$ and $F_o$ for each sampling date used in the calculation of upwelling velocities are reported in Table 2 and the input fluxes for the entire season are shown in Fig. 2.

3.3. Combined Be/Th model

In order to constrain particle export of 7Be ($P_{be}$), we calculated 234Th export from the upper 100 m, by integrating the 234Th/238U disequilibrium (Coale and Bruland, 1985; Buesseler et al., 1992; Bacon et al., 1996; Charette and Moran, 1999; Benitez-Nelson et al., 2001). A previous study in this region determined that large particle sinking events near shore can cause mixed layer 234Th/238U disequilibrium which can be transported horizontally to contribute to the measured 234Th/238U disequilibrium at SPOT (Haskell et al., unpublished). In this study, we measured the gradient in surface mixed layer activity from the L.A. Harbor to SPOT, assumed a travel time of one week to estimate ingrowth of 234Th during transport (mean speed ~5 cm s$^{-1}$), and calculated the inherited 234Th/238U disequilibrium. This value was on average, <10% of the 234Th/238U disequilibrium at SPOT. Though relatively small, this effect was removed from the integrated vertical 234Th/238U disequilibrium for each sampling date to improve the accuracy of the integration. Assuming steady-state and that particle export and radioactive decay are the only removal processes in a parcel of seawater in which upwelling introduces water from below the box and exports the surface 234Th value horizontally, the particulate removal of 234Th can be represented by:

$$P_{Th} = \lambda_{Th} \int_0^{100} (U - Th)dz + w_{HF}(Th_{deep} - Th_{surf})$$

$$= D_{Th} + w_{HF}(Th_{deep} - Th_{surf})$$

where $\lambda_{Th}$ is the decay constant of 234Th, $U$ is the activity of 238U (dpmm$^{-3}$; the soluble parent nuclide of 234Th), Th is the activity of 234Th (where subscript ‘deep’ signifies below the zone of deficiency, and ‘surf’ signifies within the zone of deficiency), and $D_{Th}$ represents the depth-integrated thorium deficiency (Table 3). The 234Th/238U deficit in the upper ocean is also dependent on the upwelling velocity, as water near secular equilibrium for 234Th/238U (Th$_{deep}$) is introduced into the surface layer, decreasing the measured 234Th deficit ($D_{Th}$). The last term in Eq. (4) accounts for this. In this derivation, we ignore diffusive transport of Th since its residence time in the surface is likely short due to its strong dependence on the particle export term and because advection is likely the dominant transport mechanism. We then solved Eq. (4) for $w_{HF}$ and set it equal to Eq. (2) to solve for the true particle flux of 234Th (Appendix B):

$$P_{Th} = \frac{\frac{D_{Th}}{Th_{deep} - Th_{surf}} - \lambda_{Th} \frac{1}{(U - Th) + \frac{F_0}{C_{HF}}}}{\frac{1}{(Th_{deep} - Th_{surf})} + \frac{(Be/Th)_{part}}{C_{HF}}}$$

If the 7Be/234Th ratio on the particles exported from the upper thermocline is known, and all other terms are measured, $P_{Th}$ and $P_{be}$ can be calculated, allowing Eq. (2) to be used to find upwelling. The (Be/Th)$_{part}$ ratio was measured on material caught in sediment traps set at 100 m deployed four times from January
to June 2013. In this study, we use the average $^{7}\text{Be}^{234}\text{Th}$ activity ratio of all trap deployments (0.028; Table 3). $^{234}\text{Th}$ export and upwelling velocities calculated using this method are shown in Table 4. The total uncertainty in both $P_{\text{Th}}$ and $w_{H}$ reported in Table 4 is calculated by propagating the uncertainties in the $^{7}\text{Be}$ inventory calculated using the Monte Carlo simulation, in the total input of $^{7}\text{Be}$, and in the depth-integrated thorium deficiency calculation. For this estimate, we use the uncertainty in the lower limit of the $^{7}\text{Be}$ inventory as if it represents both upper and lower limits because this is likely more realistic if our assumption that $^{7}\text{Be}$ activity decays exponentially with depth is true (i.e. the deeper activity can never be greater than the shallower activity). The uncertainty in the upper bound is inflated due to overlapping uncertainty in the shallow and deep $^{7}\text{Be}$ activities, causing the inventory determined by the exponential curve fit to approach infinity in some cases, which is unrealistic.

3.5. Numerical model

Finally, we used a non-steady state numerical model (NM) to calculate upwelling velocities, avoiding the approximation of a time-averaged input flux of $^{7}\text{Be}$ and non-steady state term, which introduced significant uncertainty in the NSS model. However, some uncertainty remains, as the precise timing of changes in upwelling, mixed layer depth, and particle flux is unknown. Using the NM approach, the daily change in $^{7}\text{Be}$ inventory is calculated using daily inputs and an average upwelling velocity for the period since the previous sampling. The system is initialized by assuming that $^{7}\text{Be}$ on 12/11/12 was equal to that measured the following month, with mixed layer depth and the scale depth for $^{7}\text{Be}$ penetration in the water column linearly interpolated between sampling periods. The removal rate for $^{7}\text{Be}$ was linked to the $^{234}\text{Th}$ balance by specifying the ratio of removal rates of each isotope ($k_{\text{Be}}/k_{\text{Th}}$) = 0.18 and varying $w_{H}$ to find a value that matched the $D_{\text{Th}}$ (Appendix C). One $^{7}\text{Be}$ profile was ignored (3/14/13) due to its large uncertainty. The choice of $k_{\text{Be}}/k_{\text{Th}}$ was done slightly differently than in the previous model, as removal of both isotopes was set to follow first order kinetics, with the ratio constrained by the median value of $^{7}\text{Be}/^{234}\text{Th}$ on sinking particles divided by that dissolved in surface waters. The measured $^{7}\text{Be}$ inventories and the calculated inventories using the NM approach are shown in Fig. 3.

4. Results and discussion

4.1. Upwelling in the SCB

The upwelling velocities calculated using both the SS and NSS approaches are shown for each cruise in Fig. 4b. The steady state formulation values are shown with gray dashed lines and non-steady state values are shown with thick black bars. The width of these bars represents the approximate timescale over which each value integrates (~2 weeks) and extends backwards in time from the sampling date. Also shown in Fig. 4b are the monthly average upwelling velocities calculated from the Bakun Upwelling Index output at 33°N, 119°W (Bakun, 1973; NOAA, 2013a). Shown in Fig. 4a is sea-surface temperature (SST) measured 48 times daily on NOAA buoy 46222, ~10 km inshore from SPOT, during the same time period as our sampling (NOAA, 2013b). Dramatic decreases in SST (inside red circles in Fig. 4a) coincide with two periods of increasing upwelling velocity. This is presumably due to short pulses of cold, deep-water upwelling in the region. The overall trend in SST does not follow the upwelling trend because of seasonal warming into the summer months.

The largest disagreement between the SS and NSS formulations (Feb. 28) coincided with SST event labeled, ‘1,’ and reflects the large decrease in $^{7}\text{Be}$ inventory from Feb. 14 to Feb. 28 (Table 1;
Figs. 3 and 4). This event lasted for ~3.5 days, 5 days prior to sampling and the most recent rainfall event occurred 9 days prior to sampling. Since our input is time-averaged and the change in inventory so large, the NSS formulation resulted in an upwelling velocity of 3.2 ± 1.5 m d⁻¹ (Fig. 4b, in parentheses), which likely reflects the vertical velocity during this event, but not the mean over the previous two weeks since this water mass was likely out of equilibrium with the surrounding waters. To get a more realistic estimate of the mean upwelling since Feb. 14, we used a weighted average of the SS and NSS velocities for this sampling, assuming the NSS upwelling velocity represented the ~3.5 days (25%) of low SST and the SS velocity represented the remaining ~10.5 days (75%) of the 2 weeks. The resulting \( w_H \) is then 1.9 ± 0.9 m d⁻¹, which we report with the NSS upwelling velocity (Table 4; Fig. 4b). The NM upwelling velocity also agreed rather well for this period.

In general, the NM approach produced upwelling velocities that were slightly lower than, but either within or nearly within the uncertainty of, the NSS approach for all sampling periods with the exception of March and early April, where they were very low (Fig. 4b). This was an unexpected result, but we suspect that the measured inventories for these sampling periods may have been overestimated due to the small difference in \(^7\text{Be}\) activity between the two points used in the depth-integration. Note the high uncertainty for the measured inventories in March and April (Fig. 3), and that the calculated (NM) inventory was much lower than the measured inventory on March 14. There is also a memory effect when using a numerical approach. If the inventory was overestimated in March, then the April calculation may also be affected from the first overestimation. The systematically lower NM upwelling velocities compared to the NSS are likely due to a combination of the differences in: 1) determining the \(^7\text{Be}\) input (stepped through each day vs. time-averaging), and 2) the assumptions made during derivation of each approach (Appendices A and C).

There is a general increase in the NSS \(^7\text{Be}\)-based upwelling velocity between 0.6 ± 0.3 to 2.5 ± 1.3 m d⁻¹ from January to May, then a decrease back to 1.2 ± 0.5 m d⁻¹ in June. The Bakun Index-derived upwelling velocities increase throughout the spring and agree with our estimates to within uncertainty, with the exception of the last month of our study. The Bakun Index upwelling velocity continues to rise to 2.7 m d⁻¹ in June, but all three \(^7\text{Be}\)-derived upwelling velocities are less than half that value. Considering the large spatial scale over which the pressure field estimate applies, the discrepancy in June might be attributed to spatial heterogeneity, which may affect the \(^7\text{Be}\) balance more so than the Bakun Upwelling Index if upwelling is patchy. NSS \(^7\text{Be}\) upwelling velocity in January was 0.9 ± 0.6 m d⁻¹, approximately 3 times greater than the Bakun Index. It is generally believed there is little upwelling in winter in the SCB, but the NSS velocity may be an overestimate because without a previous sampling, we were unable to include a NSS term in the balance, which may have lowered this value. Therefore, the upwelling velocity determined using both the NSS and SS approaches are reported as equivalent for this sampling. Using the NM approach, which includes a month ‘spin-up’ before the first sampling, the upwelling velocity was approximately half the NSS value (0.5 ± 0.6 m d⁻¹), likely closer to the true upwelling velocity. The uncertainty for the \(^7\text{Be}\)-based upwelling velocities (shown for the NSS approach in Fig. 4b) generally increases towards the end of the study because of increasing vertical velocity and decreasing rainfall frequency. This combined effect results in lower \(^7\text{Be}\) inventory, and thus a higher counting uncertainty that propagates into the uncertainty in upwelling velocity.

4.2. Sensitivity analysis

Upwelling velocities calculated using each approach have significant uncertainties. A sensitivity analysis was performed to determine the relative contributions of each variable to the overall uncertainty in the NSS upwelling velocity by varying each variable by 1σ and reporting the effect in upwelling velocity for two sampling dates, May 10 and Feb. 14, which represent periods of high and low upwelling, respectively. The results of this test are presented in Table 5. Four variables were tested: the \(^7\text{Be:}^{234}\text{Th}\) ratio on sinking material, the depth-integrated \(^{234}\text{Th}\) deficiency (\(D_{\text{Th}}\)), the \(^7\text{Be}\) input flux (\(F_{\text{in}}\)) and the \(^7\text{Be}\) inventory (\(C_{\text{inv}}\)). For both sampling periods, upwelling was most sensitive to the \(^7\text{Be}\) input flux.
Table 5
Model sensitivity to variables contributing to the upwelling velocity calculations on May 10 and February 14.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Estimated σ</th>
<th>May $\omega_H$ best estimate</th>
<th>May $\omega_H$ +1σ</th>
<th>May $\omega_H$ −1σ</th>
<th>Feb. $\omega_H$ best estimate</th>
<th>Feb. $\omega_H$ +1σ</th>
<th>Feb. $\omega_H$ −1σ</th>
</tr>
</thead>
<tbody>
<tr>
<td>$^7\text{Be}^{234}$Th</td>
<td>50%</td>
<td>2.5</td>
<td>2.9</td>
<td>2.2</td>
<td>0.6</td>
<td>0.8</td>
<td>0.5</td>
</tr>
<tr>
<td>$^{234}$Th deficiency ($D_{\text{Th}}$)</td>
<td>30%</td>
<td>2.5</td>
<td>2.6</td>
<td>2.5</td>
<td>0.6</td>
<td>0.7</td>
<td>0.6</td>
</tr>
<tr>
<td>$^7\text{Be}$ input ($F_a$)</td>
<td>35%</td>
<td>2.5</td>
<td>1.4</td>
<td>3.6</td>
<td>0.6</td>
<td>0.0</td>
<td>1.3</td>
</tr>
<tr>
<td>$^7\text{Be}$ inventory ($C_{\text{inv}}$)</td>
<td>40%</td>
<td>2.5</td>
<td>2.6</td>
<td>2.4</td>
<td>0.6</td>
<td>0.8</td>
<td>0.4</td>
</tr>
</tbody>
</table>

B) NM

| $^{234}$Th deficiency ($D_{\text{Th}}$) | 30% | 2.0 | 2.0 | 1.9 | 0.3 | 0.3 | 0.3 |
| $^7\text{Be}$ input | 30% | 1.9 | 1.4 | 2.5 | 0.3 | −0.1 | 0.6 |
| $^7\text{Be}$ inventory ($C_{\text{inv}}$) | 30% | 1.9 | 2.8 | 1.5 | 0.3 | 0.7 | 0.0 |

estimate: a 1σ change in the $^7\text{Be}$ input flux estimate causes a 44% and 102% change in upwelling velocity on May 10 and Feb. 14, respectively. The other three variables rank in the following order for May 10: $^7\text{Be}$/$^{234}$Th (16%), $^7\text{Be}$ inventory (5%), and $D_{\text{Th}}$ (2%). For Feb. 14, a 1σ variation caused the following percentage change in upwelling velocity: $^7\text{Be}$ inventory (34%), $^7\text{Be}$/$^{234}$Th (32%), and the $D_{\text{Th}}$ (5%). It is important to note that the contribution of the $^7\text{Be}$ input flux to the uncertainty in the upwelling velocity on Feb. 14 (102%) may seem large, but because of the small magnitude of upwelling velocity on Feb. 14, this only corresponds to an uncertainty of $\sim 0.6 \text{ m} \text{ d}^{-1}$.

The NM-derived upwelling velocities were most sensitive to uncertainties in the $^7\text{Be}$ input and the $^7\text{Be}$ inventory. Increasing input or decreasing inventory by 30% increased calculated upwelling by 0.0 to 0.7 m d$^{-1}$ for each period, an average of 0.4 m d$^{-1}$, with a corresponding reduction for the opposite change in each of these parameters. Uncertainties in particle scavenging (Th inventory or $k_{\text{Be}}$/k$_{\text{Th}}$) contribute less than 0.1 m d$^{-1}$ uncertainty. Assuming the uncertainties in these parameters are not correlated, error propagation indicates the systematic uncertainties for the ensemble of values should be about 0.6 m d$^{-1}$. Because of model memory effects, there could be additional point-to-point variability that is introduced by random variation in the data set, possibly due to patchy upwelling. Regardless, the ability of the NM approach to match the NSS inventory every sampling date except one (Fig. 3) suggests that even though individual days may have large uncertainties, the mean upwelling velocity for the season should be reasonable because the uncertainty on the magnitude of the entire NM ensemble should be small.

4.3. Hydrography

Fig. 5a shows CTD depth profiles of salinity, temperature and density at SPOT throughout the study. Strong thermal stratification in the near-surface begins around late-March and continues into the summer, driving the density stratification during the second half of the sampling season. The density gradient below 50 m remains fairly similar from January to June, suggesting that upwelling waters typically come from a relatively shallow depth. This may be evidence that wind-stress curl driven upwelling, as defined by Rykaczewski and Checkley (2008), is the primary upwelling source, rather than coastal upwelling. Further north along the coast of the western North America, the CC is strong near the coast, where there is also a steep shelf break. This combination drives strong near-shore Bottom Boundary Layer (BBL) transport of a large volume of upwelled waters up the shelf, which results in shoaling deep isopycnals that breach the surface layer near the shoreline (Jacox and Edwards, 2011). However, in the SCB, the portion of the CC that continues near-shore is not as strong and is episodic, thus upwelling that occurs in the SCB is likely due to interior water column advection driven by wind-stress curl (Chelton, 1982). We present evidence that this is case in the following two sections. The NOAA buoy temperature record shown in Fig. 4a may be evidence that there are periods when short pulses of BBL transport may breach the surface layer intermittently. However, this appears to be sporadic and it is unclear whether the temperature deviation was caused by BBL or interior vertical transport.

4.4. Heat budget

To test whether there is a substantial influence of BBL-upwelled water that breaches the surface layer over the shelf, and then advects horizontally to SPOT, we constructed a simple one-dimensional heat budget using temperature profiles taken via CTD at SPOT and at a station sampled over the shelf near the Paslo Verdes peninsula (near 500 m depth; marked in Fig. 1 with the letters, 'PV'). PV was chosen because it is believed to be a site where episodic pulses of upwelling have occurred in the past, based on JPL ROMS sea-surface temperature (SST) (NASA, 2013). Both stations were sampled three times on Feb. 28, Apr. 3, and May 10, which spanned a range of upwelling velocities from 0.6 to 2.5 m d$^{-1}$ (temperature profiles are given in the supplementary material (Fig. S1); the difference in mean surface and deep temperature at each station was ∼4 °C). Assuming the net heat input (solar insulation, back radiation, evaporative loss and sensible heat transfer) is equal at both locations and that vertical diffusion is small, the change in integrated heat over time should then be proportional to the relative difference between the average upwelling velocity at each station over this time period. For example, if there were regular upwelling pulses that contributed significantly to the vertical advective flux over the shelf at PV, but not at SPOT, then the change in depth integrated heat over time during the period of warming between winter and summer (Fig 4a) would be lower at PV than at SPOT. It is important to note that this is a crude calculation, not accounting for many factors (e.g. variability in meridional currents). Therefore, we cannot calculate the magnitude of upwelling velocity in this manner; we can only roughly estimate the relative magnitude of upwelling at each station.

The change in heat content over time at PV was ∼2× that at SPOT, suggesting the amount of upwelled water at PV is likely less than that at SPOT. Assuming that this station is representative of the entire inner shelf, the magnitude of BBL upwelling up the shelf slope is likely less than the magnitude of wind-stress curl-driven interior water column upwelling. Furthermore, if upwelling was not generally concentrated in a BBL near the shelf, then it follows that the signal measured at SPOT is likely not greatly influenced by water upwelled over the shelf and transported horizontally in the surface layer.

4.5. Nutrient distribution

There is also evidence in nutrient distribution for the trends in upwelling velocities reported in this study. Fig. 5b shows
phosphate, nitrate, and silica concentrations in the upper 250 m of the water column at SPOT during the study period, collected on each day. \( ^{7}\)Be samples were collected. The upper 75 m is outlined with a red square to highlight where the nutrient gradient is strongest. From the beginning of March until mid-May, during the period of increasing upwelling velocity, the nutrient gradients not only shoal by \(\sim 20\) m, but also become stronger. Then after the peak in upwelling velocity, when upwelling relaxes, so does the strength of the nutrient gradients. We argue this is an effect of vertical advection, and not dictated by the thermal gradient. This interpretation seems likely because the strength of the thermal gradient increases dramatically from the end of March through June due to very strong surface warming, but the strength of the nutrient gradient relaxes over this same time period. Under the assumption of steady state, the vertical flux of each nutrient is constant and equals the sum of the flux due to vertical eddy diffusion and advection (Eq. (1b)). If this is the case, and the change in upwelling velocity is faster than the change in eddy diffusivity over large length scales (tens of meters), then an increase in upwelling velocity will increase the strength of the nutrient gradient, whereas a decrease in upwelling velocity (or an increase in eddy diffusivity) will decrease the strength of the gradient. If upwelling were not occurring beneath our sampling location, but instead along the coastline and advected to our sampling site, the changes in the nutrient gradients would likely not correspond with changes in upwelling velocity. We interpret this as further evidence that upwelling is local beneath the SPOT hydrostation and not a signal that is advected from a zone over the inner shelf.

The Bakun Upwelling Index does not decrease going into June, but instead continues to increase, contrary to the behavior indicated by \(^{7}\)Be. We attribute this contrast to a difference in observational scales. The Bakun Index probably reflects a larger-scale signal for the Southern California Bight than our \(^{7}\)Be-based upwelling velocities, which may reflect localized patchiness in upwelling. Evidence for this comes from our sampling on February 28. Both the CTD data shown in Fig. 5a, as well as the nutrients in Fig. 5b, show a clear destabilizing event shortly before the sampling date, which introduced cold, dense, nutrient-rich water into the mixed layer. The isopycnals that shoaled to depths around 10 m during this event are typically found around the base of the euphotic zone at \(\sim 40-50\) m. We interpret this to be either a local upwelling pulse or possibly an eddy that moved through the region a few days before our occupation of SPOT. The \(^{7}\)Be inventory, and thus \(^{7}\)Be upwelling velocity, apparently responded to this local event, although the monthly Bakun Index shows no deviation from the fairly consistent increase from January to June.

5. Conclusion

In this study, we use a mass balance for \(^{7}\)Be in the surface ocean, similar to the approaches of Kadko and Johns (2011) and Haskell II et al. (2015), to estimate upwelling velocity in Southern California during the seasonal upwelling cycle. We demonstrate that by using a balance of \(^{234}\)Th in the surface ocean to estimate \(^{7}\)Be loss to particle export, it is possible to use this approach to estimate upwelling velocity in a high particle density environment, which was not possible before. Although an extensive measurement suite is required, this approach is necessary to use \(^{7}\)Be as an upwelling tracer in coastal environments. The main advantage of this approach is that it estimates upwelling velocity over the timescale of weeks, the same timescale over which biological blooms occur. Between January and June 2013, upwelling...
velocity increased, peaking around late May, then decreased to about half the peak value in June. The monthly Bakun Upwelling Index at 33°N, 119°W, a pressure field-based approach to estimate upwelling, showed generally the same seasonal trend. It agreed with our upwelling estimates within uncertainty, for every sampling period except one in June. SST did not clearly reflect this seasonal trend in upwelling because of increased insolation from winter to summer. However, SST did drop rapidly during two periods of high upwelling velocity, presumably from pulses of deep water reaching the surface layer. Although these pulses may occur over the inner shelf periodically during the upwelling season, the overall signal of upwelling velocity at the SPOT hydrosat station appears to be dominated from below, suggesting wind stress curl-driven upwelling. There is evidence for this in both the regional heat budget and local nutrient distribution: 1) Heat budgets for the SPOT hydrosat and station PV, located over the inner shelf, show that the increase in depth-integrated heat content at PV from winter to summer was greater than that at SPOT throughout a period of increasing upwelling velocities, suggesting upwelling is not focused preferentially near the shelf. 2) Nutrient gradients in the upper 75 m of the water column shoaled and strengthened during increasing upwelling velocities, and then became weaker as upwelling velocities decreased while the temperature and density gradients increased. Overall, the 7Be-based upwelling estimates agree well with the Bakun Upwelling Index, however we believe that the 7Be method is able to resolve local upwelling dynamics better than the Bakun Upwelling Index, a great advantage in heterogeneous near-shore environments.

Acknowledgements

We would like to thank Troy Gunderson for CTD training and SPOT management, Baron Barrera and Diana Molina for sample collection and processing; Jotaus Baranoss for silica and phosphate analysis, Corinne Calhoun for nitrate analysis, and all other participants in the UpRISEE cruises who helped with sample collection. None of this work could have been possible without the helpful assistance of the captain and crew of the R/V Yellowfin. We would also like to thank the USC Wrigley Institute of Environmental Science for the use of their CTD. Special thanks to Dave Kadko for helpful discussions on this topic. Thanks to Will Berelson for use of his sediment trap array. We are also grateful to two anonymous reviewers for their helpful suggestions. Financial support for this project came from a graduate research grant from the International Association for Geochemistry to William Haskell, and grants 1260296 to Maria Prokopenko and 1260692 to Douglas Hammond from the Chemical Oceanography program of the National Science Foundation.

Appendix A. Mathematical formulation of 7Be upwelling model equations

At steady state, the balance described by Eq. (1) becomes:

\[ 0 = F_a + K_z \left( \frac{\partial C}{\partial z} \right)_{H} - \lambda C_H H - P_{Be} \]  

(A.1)

Assuming that radioactive decay is the only important reaction for 7Be, the change in concentration below the mixed layer box over time is given by:

\[ \frac{\partial C}{\partial t} = \frac{\partial}{\partial z} K_z \left( \frac{\partial C}{\partial z} \right) - \lambda C - \frac{\partial C}{\partial z} \]  

(A.2a)

This equation can be solved at steady state, with the boundary condition that the 7Be goes to zero at infinite depth, and assumptions that \( K_z \) and \( w \) are independent of depth:

\[ C(z) = C_H e^{-\alpha(z-H)} \]  

(A.2b)

where \( z \) is water depth and the depth attenuation coefficient (\( \alpha \)) is given by:

\[ \alpha = \frac{-w}{2K_z} + \frac{1}{2} \left\{ \left( \frac{w}{K_z} \right)^2 + \frac{3\lambda}{K_z} \right\}^{1/2} \]  

(A.2c)

At steady state, the downward flux of 7Be from the mixed layer (\( J \)) must be balanced by radioactive decay, and can be estimated by integrating 7Be activity to infinite depth:

\[ J = \int_{-\infty}^{H} \lambda C(z) dz = \frac{\lambda}{\alpha} C_H \]  

(A.3)

Combining Eq. (A.3) with Eq. (A.1) at steady state yields

\[ w_H = \frac{1}{C_H} (J + \lambda H C_H - F_a + P_{Be}) = \frac{\lambda}{\alpha} + \lambda H - \frac{F_a + P_{Be}}{C_H} \]  

(A.4)

Appendix B. Mathematical formulation of combined 7Be/234Th model

The rate of change over time of 234Th activity in a unit of seawater is given by:

\[ \frac{\partial \text{Th}}{\partial t} = \lambda_{\text{Th}} U - \lambda_{\text{Th}} \text{Th} - P_{\text{Th}} + V \]  

(B.1)

where \( \text{Th} \) is the activity of total 234Th (dpm/m^2); \( dpm = \text{disintegrations per minute} \), \( U \) is the activity of 234U (the soluble parent nuclide of 234Th), \( \lambda_{\text{Th}} = \text{the decay constant of 234Th (0.0288 day}^{-1}) \), \( V \) is the net input (or loss) by advective and diffusive transport (dpm/m^2 d), and \( P_{\text{Th}} \) is removal rate by settling particles (dpm/m^3 d). Assuming steady-state, integrating with depth, and assuming that the \( V \) term in Eq. (B.1) dominated by upwelling of water from below the box and exports the surface Th value horizontally, the particulate removal of 234Th can be represented by Eq. (4). We then solved Eq. (4) for \( w_H \):

\[ w_H = \frac{P_{\text{Th}} - D_{\text{Th}}}{Th_{\text{deep}} - Th_{\text{surf}}} \]  

(B.2)

and set it equal to Eq. (A.4):

\[ \frac{\lambda_{\text{Be}}}{\alpha} + \lambda_{\text{Be}} H - \frac{F_a + P_{\text{Be}}}{C_H} = \frac{P_{\text{Th}} - D_{\text{Th}}}{Th_{\text{deep}} - Th_{\text{surf}}} \]  

(B.3)

We then solved for the particle flux of Th (\( P_{\text{Th}} \)):

\[ P_{\text{Th}} = \frac{D_{\text{Th}}}{Th_{\text{deep}} - Th_{\text{surf}}} - \lambda_{\text{Be}} (\frac{1}{\alpha} + H) + \frac{F_a}{C_H} \]  

(B.4)

Using this equation, we are able to estimate the true particle export of 234Th and 7Be if we can estimate the 7Be/234Th ratio on the particles exported from the upper thermocline. Once this ratio is established and Eq. (B.4) is solved for \( P_{\text{Th}} \), we are able to use this value to solve Eq. (A.4) for upwelling velocity. The 7Be/234Th ratio was measured on material caught in sediment traps set at 100 m deployed four times from January to June 2013. In this study, we use the average 7Be/234Th ratio of all trap deployments (0.028).

Given that the ratio of 7Be/234Th in the water column is reported here as ~0.05 on average (Tables 1 and 3), and the published partition coefficient of Be is approximately an order of magnitude lower than Th (~10^5 vs. ~10^6 Lkg^{-1}; Nyffeler et al., 1984; Hennyman and Santachi, 1988), it would follow that the expected ratio on particles would be ~0.005, ~5x lower than our measured
mean value on sinking particles. Two of the measured ratios averaged to 0.005 (Up-5 and Up-7; Table 3) and two averaged to 0.05 (Up-3 and Up-8). Interestingly, the two large values were measured both early and late in the season, during periods of moderate upwelling velocity and moderate rainfall, whereas the two low values were measured during periods of high rainfall within 5 days of sampling and just weeks preceding and following the peak in upwelling velocity. Likely, the fluctuations in this ratio are due to variability in rainfall, particle export, composition of the particles and vertical and horizontal advection in this highly dynamic system. Following the method described in Section 4.2, using the measured lower limit for the $^{7}$Be/$^{234}$Th ratio (0.005) results in a ~36% change in the estimated upwelling velocity for Feb. 14 and ~25% change for the peak upwelling velocity on May 10. Both of which are within the reported uncertainty.

Appendix C. Mathematical formulation of time variable numerical model

The $^{7}$Be inventory ($J$; dpm m$^{-2}$) is affected by atmospheric input ($F_a$; dpm m$^{-2}$ d$^{-1}$), decay, loss due to upwelling, and particle loss to the mixed layer. Assuming $^{7}$Be concentrations ($C_H$) is constant for depths 0 m to the base of the mixed layer ($H$) and decreases exponentially from $H$ to infinity ($C = C_H \exp(-\alpha z)$; $\alpha$ = depth attenuation coefficient and $z$ = depth), the change in inventory over time can be represented by the following equations:

$$\frac{\partial J}{\partial t} = b \frac{\partial (gC_H)}{\partial t} = F_a - \lambda b C_H - w_H C_H - k_{Be} b C_H$$

$$(C.1a)$$

$$\frac{\partial I}{\partial t} = F_a - g b C_H$$

$$(C.1b)$$

where $b = H + \frac{1}{\alpha}$

and

$$g = \lambda + \frac{w_H}{\alpha} + k_{Be}$$

$$(C.1c)$$

where $\lambda$ is the $^{7}$Be decay constant (d$^{-1}$), $w_H$ is the upwelling velocity (m d$^{-1}$), $k_{Be}$ equals the rate constant for $^{7}$Be particle removal (d$^{-1}$).

Here, this equation is integrated over time, using daily time steps to establish $g$ for the interval between successive field observations. Once $g$ is established for $^{7}$Be, this factor is used with the $^{234}$Th balance to obtain both $w_H$ and $k_{Be}$. For $^{234}$Th, a similar equation can be written:

$$\frac{\partial I}{\partial t} = D_{Th} + w_{Th}(T_{deep} - T_{surf}) - H'k_{Th}Th$$

$$(C.2)$$

where $D_{Th}$ represents mean $^{234}$Th in the zone of deficiency, $T_{deep}$ is $^{234}$Th activity at the base of the integrated depth, $T_{surf}$ is $^{234}$Th in the upper layer defined for Be $H'$ is the depth used to calculate the $^{234}$Th deficiency ($D_{Th}$), and $k_{Th}$ is the rate constant for $^{234}$Th removal by particulates. The equations for $^{234}$Th and $^{7}$Be are linked by a defined relationship of $k_{Be}/k_{Th} = 0.18$ based on observations in sediment traps, allowing the $^{234}$Th and $^{7}$Be balances (based on $g$) to be solved for $w_H$ and the rate constants $k_{Be}$ and $k_{Th}$. This works because $^{7}$Be is primarily sensitive to upwelling, while $^{234}$Th is primarily sensitive to particle removal. The NM simulation indicates that the average $^{7}$Be removed over the course of the study period is 50% by upwelling, 40% by decay, and 10% by particle export, although these fractions vary from one interval to the next.

Appendix D. Supplementary material

Supplementary material related to this article can be found online at http://dx.doi.org/10.1016/j.epsl.2015.04.015.

References


