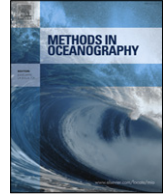




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Full length article

Real time visualization of thermohaline finestructure using Seismic Offset Groups

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ABSTRACT

Seismic oceanography is based on the passage of a regularly repeating acoustic impulsive source and an acquisition streamer along the surface of the ocean, and on summing together all signals reflected from temperature and salinity interfaces in the ocean (where there are acoustic impedance contrasts). Due to the inherent redundancy of the method, random noise is attenuated, while signal is preserved; however, if the original signal-to-noise ratio is large enough, one need not use data from the entire streamer to create a 2D profile. A processing scheme is here devised to obtain consecutive images, known as stacks, of the structure of the water column. The scheme, named Seismic Offset Groups (SOG), consists in splitting the data from the whole streamer at a given geographical position into data produced by different streamer subsets. The method is illustrated by partitioning data from a 5-km long streamer into 7 offset groups separated by 3.5 min in time, thereby imaging the same seafloor-referenced location over a period of 21 min. As the streamer passes over a fixed geographical point, motions within the water column are observed. Each stack, created with a subset of the complete streamer, can therefore be considered an image of the water

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column at a particular time step (animation frame). In this way each image shows a different thermohaline fabric and the animation allows us to visualize internal ocean motions.

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1. Introduction

Visualization of otherwise unseen physical phenomena such as thermohaline finestructure is an important step to a fuller understanding of these phenomena. Broadly, as a tool of remote sensing, visualization may come in the form of a one-time spatial illumination of the body in question (for example, in the case of the specular reflection of light, a photo). Likewise, in the case of the specular reflection of sound waves, we are able to create a stacked seismic profile providing us with information about the Earth or ocean that had not been previously observed. In the context of multi-channel seismic reflection profiling (henceforth, MCS), due to the large scale of the imaging target and its obscurity in the sub-surface, to create an image it is necessary to acquire an abundance of reflected sound wave data and to combine these data using the robust techniques of seismic data processing and analysis (Yilmaz, 2001).

Seismic reflection investigations have been common for delineating the gross structure of the Earth's crust, mantle and core for several decades—first as two-dimensional profiles, later as three-dimensional volumes. More recently, the repeated acquisition of seismic data over the same location in time-steps of several months or years has added the fourth dimension of time to the interpretation of such subjects as petroleum reservoir monitoring (Lumley, 2001), monitoring of CO₂ sequestration (Chadwick et al., 2005) and the active migration of faults (Cheng, 2008).

Holbrook et al. (2003) made the first modern analyses of ocean thermohaline finestructure using the MCS method and found that the acoustic impedance boundaries, which give rise to seismic reflectivity, were the result of temperature and salinity fluctuations, or thermohaline finestructure. These structures have been known to occur for a long time through conventional oceanographic methods (Stommel and Federov, 1967) but were not visualized at high horizontal resolution until the MCS method, as it provides approximately 10 m horizontal resolution images which are to be compared with typical ≥ 1 km resolution of CTD (Conductivity–Temperature–Depth) casts. More recent physical oceanographic studies, using MCS as a tool, have imaged thermohaline finestructure and confirmed that vertical temperature and salinity gradients are responsible for the observed acoustic impedance contrasts. Sallarès et al. (2009) identified the relative contributions of temperature and salinity to reflectivity and found temperature to be by far the dominant physical property which gives rise to acoustic impedance contrasts, and therefore reflectivity. Other seismic oceanography surveys (see Ruddick et al., 2009 for a concise summary of seismic oceanography) have imaged the Kuroshio Current near Japan (Tsuji et al., 2005; Nakamura et al., 2006), the Norwegian Sea (Nandi et al., 2004; Páramo and Holbrook, 2005), the Southern Ocean (Sheen et al., 2009), the Mediterranean Undercurrent and Meddies (Biescas et al., 2008; Buffett et al., 2009) and the Caribbean Sea (Fer et al., 2010). While developing unique components of seismic reflection profiling, all these studies had in common the fact that they represented two spatial dimensions – in depth and horizontally along the direction of acquisition – by creating a quasi snapshot (Vsemirnova et al., 2009) of the ocean. For ocean processes taking place at time scales on the order of the duration of seismic acquisition, this means that all MCS images, except perhaps those in the most quiescent of regions, are ‘smeared’ in time. This problem is exacerbated in the case of the lengthy acquisition of true 3D seismic volumes.

To recap, the dynamic nature of the ocean presents a challenge to MCS for two reasons. First, the ageostrophic motion of ocean currents, and therefore the thermohaline finestructure, is variable on time scales comparable to the data acquisition time, i.e. on the order of minutes or hours (Géli et al., 2009; Klaeschen et al., 2009). This implies that the application of the method to measure movements in the ocean on such short time scales needs to be carefully designed. Second, data acquisition over a precise seafloor-referenced location is very difficult, requiring specialized and often expensive acquisition schemes. The acquisition streamer is almost always subject to some ‘feathering’

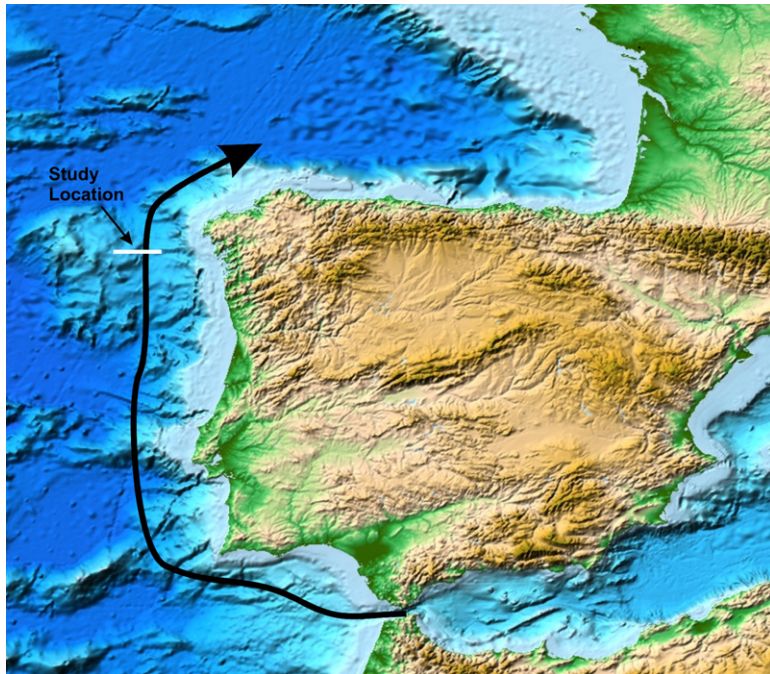


Fig. 1. Location of seismic section (small white line). Black line shows the idealized path of the Mediterranean Undercurrent. Source: Bathymetry map from ETOPO1 Global Relief Model (<http://www.ngdc.noaa.gov/mgg/global/global.html>).

(displacement off-line due to near surface currents) and the ship's azimuth does not always align with the azimuth of the streamer.

Despite these constraints, some efforts have already been started. Blacic and Holbrook (2009) performed the first 3D study of seismic oceanography using two parallel acquisition swaths along the same profile and estimated the orientation of internal wave crests. Klaeschen et al. (2009), during one particular repeat survey of profiles with diametrically opposed azimuths, observed apparent wavelength changes of reflectors that were explained as caused by a Doppler-like effect. By knowing the ship's velocity vector and having a reliable independent model of in situ sound speed, from the simultaneous and co-located deployment of eXpendable BathyThermograph (XBT) probes, they were able to determine that the movement of reflectors was related to real thermohaline motions, i.e. they were not an artifact of acquisition or digital signal processing.

In this study we take a completely different and novel approach to visualize ocean dynamics by taking advantage of inherent seismic acquisition redundancy, to introduce a processing scheme which separately processes seismic data from different grouped offsets (offset is the source to receiver distance) along the same profile. Normally in MCS, if possible, all data from the seismic acquisition streamer is used to create a sub-surface image. This allows maximum statistics and therefore optimal signal-to-noise ratio. However, if the signal quality is sufficient, only a subset of the full streamer is needed to represent a region of the sub-surface in two spatial dimensions, with each successive offset group passing over any given geographical point at a later time. This technique, hereafter named Seismic Offset Group (SOG) processing, allows the creation of several animation frames, thus visualizing in real time, the dynamics of a small region of a changing ocean in (Fig. 1).

To the best of our knowledge, a similar approximation was used by the Marine Geodynamics group at GEOMAR in 2008 to cross-check the results from an automatic movement detection algorithm (D. Klaeschen, personal communication, 7 July 2012). The SOG technique was first presented publicly and applied to real data by Carbonell et al. (2010) and Buffett (2011). Most recently, the approach has been used to investigate non-linear internal wave motion in the South China Sea (VonLanken

et al., 2012). Here we provide the first peer-reviewed detailed description and application of the SOG technique to real data and verify its performance with synthetic data. We present clear images of the movement of thermohaline finestructure which, as will be shown, cannot be explained as an artifact of incorrect processing or acquisition. More comprehensive future studies, using specifically designed SOG analysis and likely in conjunction with simultaneous and co-located thermohaline and velocity in situ measurements, shall be able to associate the finestructure movements to the relatively fast readjustment of pressure anomalies (in comparison with the inertial period ($T/(2 \sin \theta)$), where $T = 24$ h is the period of the Earth's rotation and θ is latitude), including all sorts of transient waves. In this study we do not have the necessary data to make such a conclusive interpretation in the context of physical oceanography so we present the technique as a tool to visualize the motion of thermohaline finestructure.

2. The Seismic Offset Group processing scheme

The SOG methodology used in this experiment differs from most MCS where one image is created from all the streamer's offsets. We use groups of the data offsets to generate several temporally spaced images of the same acoustic impedance interfaces as the streamer passes over them. The potential number of images depends on the streamer length, the vessel speed, the data quality (signal-to-noise ratio), and the shot interval repetition rate. This is because the so-called seismic 'stacks' are the result of summing together all seismic traces within a given offset range. The inherent redundancy of this method means there are many traces that represent the same depth-points but generated from waves arriving from different angles (Fig. 2). The summing together (stacking) of these common-depth-point (or common mid-point, CMP) traces improves the signal-to-noise ratio by constructively boosting signal content and destructively cancelling random noise. The longer the offset range, the better the statistics available for noise cancellation.

Considering the motion of the ship relative to the seafloor we re-sort the seismic data, firstly by CMP and secondly by the appropriate offset range. Next we minimize other possible factors that could account for artificial reflector movements. This consists of two steps. (1) Normalization of frequency bands related to the loss of high frequencies as a function of offset. Each successive offset range contains a progressively lower dominant frequency due to the frequency filtering effect of the ocean as a function of larger incident angles (thus, longer raypaths). Therefore, we apply a low-pass filter to all stacks to minimize this effect by omitting higher frequencies that would only be present in near offset groups. (2) Amplitude is also reduced as a function of offset. We apply a differential geometrical spreading correction to each independently processed offset group to minimize this effect. However, since water is a low viscosity acoustic medium (not elastic), there is minimal amplitude degradation. We then attenuate the direct wave using an eigenvector filter, perform velocity analysis (a technique of determining the sound speed that is associated with the best-fit hyperbola to CMP data, thus optimizing stack coherency (Sheriff, 1991)) and apply a normal moveout correction using the carefully selected stacking velocities. Finally, we apply a stretch mute, stack each group independently and apply a post-stack time migration (Yilmaz, 2001).

3. Test case

To test the SOG method we use MCS data that were acquired in August 1993 as part of a tectonic study to image the Iberian–Atlantic Margin (IAM). These data also showed strong reflections within the water column and resulted in studies by Biescas et al. (2008), Krahmann et al. (2008), Buffett et al. (2009) and Pinheiro et al. (2010), among others. The acquisition system included a Bolt air gun array with a total volume of 0.123 m^3 . Nominal source depth was 10 m and nominal streamer depth was 15 m. The shot point interval was 75 m with peak energy in the 20–50 Hz range. Shot records consisted of 192 channels spreading over 5 km at a receiver interval of 25 m. The CMP spacing was 12.5 m.

Data for this analysis were acquired along line IAM-11, off the western Iberian margin (Fig. 1). This profile was used because of the long streamer, high quality data set and due to the reported decay in laminar finestructure and increased disturbance of isopycnals downstream of the Mediterranean Undercurrent (Buffett et al., 2009), which indicates mixing and therefore high thermohaline

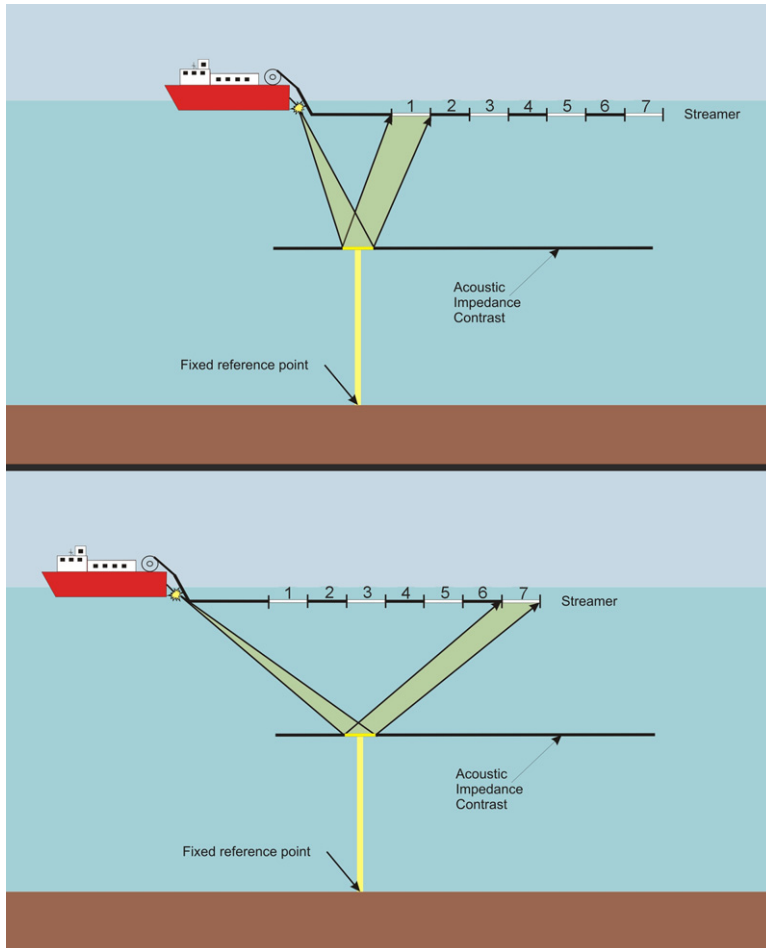


Fig. 2. Schematic diagram showing the first and last streamer offset group imaging the same oceanic structure over a fixed terrestrial reference point.

finestructure variability. We processed the data following the procedure explained in the previous section. In particular, we chose 500 m offsets because they allowed us to create seven images separated by 3.5 min in time while not compromising signal integrity, therefore imaging the same geographic location over a total period of 21 min. Finally, in this application we normalized the frequency bands by using a low-pass filter of 5/10–50/60 Hz.

4. Results

We present a series of seven images of the same water column recorded at different times showing the movement of thermohaline finestructure. The image of the full seismic section, corresponding to the time when the first image (or frame) was taken, is shown in Fig. 3 (a) and its associated animation is shown in Fig. 3 (b). In this figure we may identify the position of four boxes, 100 m deep and 2 km long, to be used to examine the temporal movements of the finestructure. Three of these boxes are located within the water column but the fourth one is found inside the sediment layers. In Fig. 4 we show a close-up of the temporal variability (seven frames at time steps of approximately 3.5 min) of the

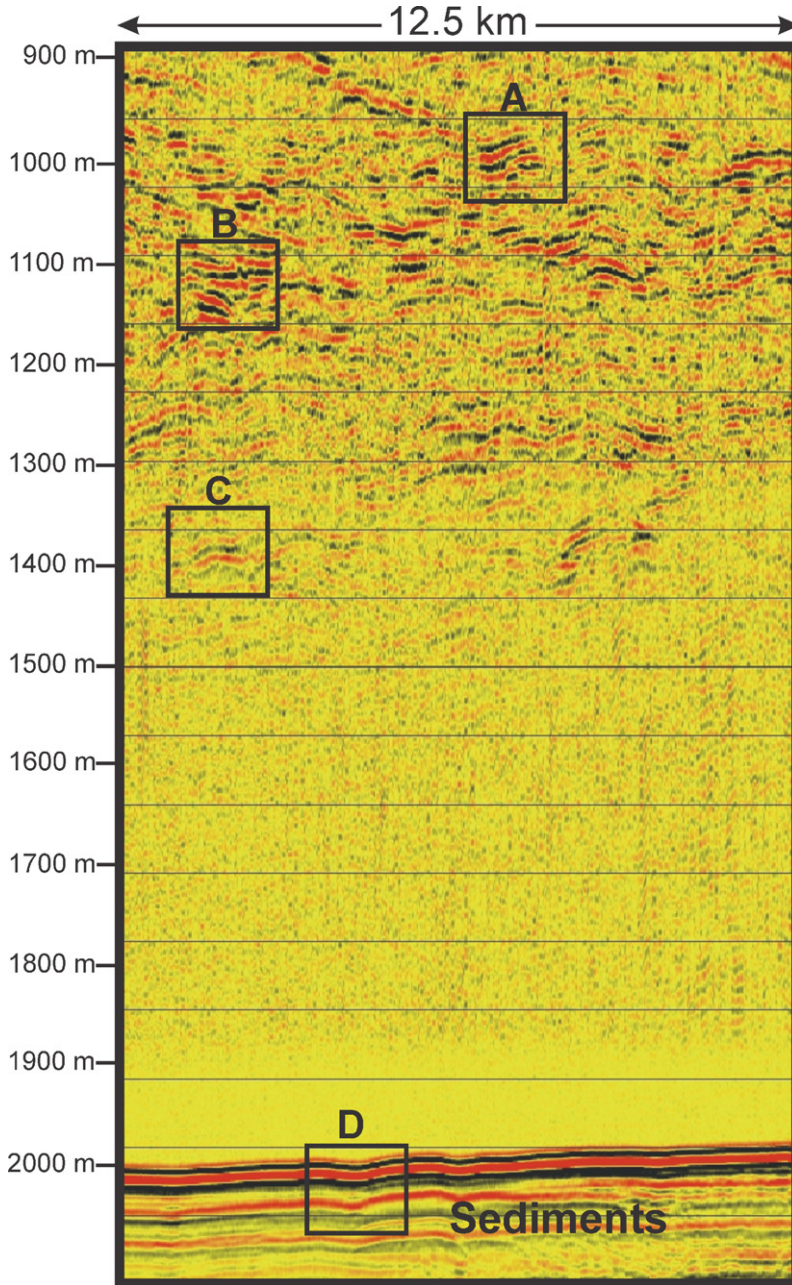


Fig 3. (a) Seismic stack of complete section (from 900 m to the seafloor) at Frame 1, illustrating the position of all boxes later used for zoomed animations. Each box size is 2 km in the horizontal and 100 m in the vertical. (b) Corresponding animation. See Fig. 3(b) in the online version at <http://dx.doi.org/10.1016/j.mio.2012.07.003>.

water column reflectors (Fig. 4(b); Animations A–C) and the seafloor/sediment reflections (Fig. 4(b); Animation D), together with an equivalent simulation from synthetic seismic data, noise-free and necessarily fixed in time (Fig. 4(b); Animation E).

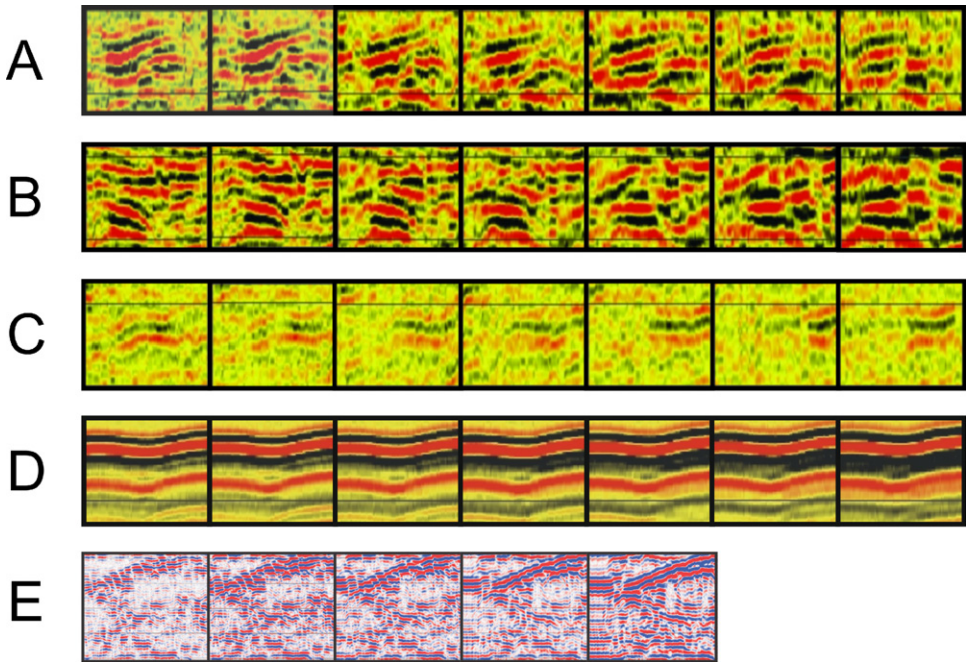


Fig 4. (a) Zoom boxes of different seismic stacks for the same geographical location (Fig. 1) created with different sections of the recording streamer (see Fig. 3 for box locations, boxes A–C are found within the water column, Box D is found within the sediments; Box E is synthetic data). The time step between animation frames, based on known ship velocity, is approximately 3.5 min. The horizontal and vertical scales of each frame are 2 km and 100 m, respectively. Streamer offset ranges are: Frame 1: 500–1000 m; Frame 2: 1000–1500 m; Frame 3: 1500–2000 m; Frame 4: 2000–2500 m; Frame 5: 2500–3000 m; Frame 6: 3000–3500 m; Frame 7: 3500–4000 m. (b) Corresponding animation boxes A–E. See Fig. 4(b) in the online version at <http://dx.doi.org/10.1016/j.mio.2012.07.003>.

An important caveat to be noted is the apparent movement of the reflectors in the Earth's crust and in the synthetic data but, as will be demonstrated, this is only an artifact of the frequency filtering bias toward longer wavelengths as a function of offset (therefore, longer travel paths through the water column). Considerable effort was made to normalize frequency bandwidths across each animation frame by band-limiting each offset stack to the minimum frequencies observed in the most distant stack. Nonetheless, it was deemed not possible to fully equalize all frequencies across each band because the travel of acoustic waves through the water column differentially filters frequencies such that, even within an identical bandwidth, there will inevitably be a different spectrum of frequencies caused by the unique physical filtering of the water column along a given raypath. This effect can be seen in all boxes, i.e. in the thermohaline finestructure, in the reflections from sediments and in the synthetic data. However, while there is a progressive vertical thickening of all seismic traces due to this effect, the reflectivity obtained from the thermohaline finestructure also exhibits smooth temporal motions taking place both in the vertical and horizontal directions which cannot be attributed to this cause. Another artifact visible only in the synthetic data, caused by the modeling software, is the padding of traces to the left leading edge of each subsequent stack resulting in a uniform shift from left to right. The point of argument, however, is in the complete lack of *intra-reflector* variability (which we define as the frequency and amplitude-independent vertical or horizontal displacement of a reflector which is not accounted for by processing artifacts) in the synthetic data as compared with the thermohaline data. In other words, the synthetic reflectors move uniformly like a single, rigid group due to trace padding and show a steady thickening of reflectors with each successive frame due to the frequency filtering effect. Definitively, the synthetic data do not show variability in the *shape* of the reflection package, a feature uniquely characteristic of the thermohaline reflectivity.

Let us briefly describe some of the details of the results in Fig. 4. In Box A (Fig. 4(b); Animation A) we notice a packet of reflectors with positive (red) and negative (black) polarities. From Frames 1 to 7 (0 to 21 min) the packet of reflectors steepens slightly on the right side and then forms an undulating shape; reflectors are observed to successively disappear and restore, with reversals in polarity at places. In Box B (Fig. 4(b); Animation B) there is an undulating shape in the upper right of the first two frames; it becomes flattened in Frame 3, reappears in Frame 4, is disrupted in Frame 5 and is somewhat flattened in the last two frames. Box C (Fig. 4(b); Animation C) shows a weak undulated reflectivity package which becomes flattened toward the last frame; a discontinuity develops but the reflector is restored in the final frame. In Box D (seafloor sediments—Fig. 4(b); Animation D) there is no observable vertical or horizontal displacement of the reflectors. What can be observed is only the thickening of reflectors due to the aforementioned caveat of frequency filtering with increased travel distance.

5. Verification against synthetic seismic data

The purpose of this test was to verify the processing scheme using a synthetic data set with zero fluctuation in the shape of the reflectivity pattern. The synthetic data were created using an explicit time-domain solver (called *SeisSol*) based upon ADER-DG, a discontinuous Galerkin method (Käser and Dumbser, 2006). Reflection coefficients for synthetic seismic data were generated from sound speed profiles derived from CTD data collected during the Geophysical Oceanography project (Hobbs and the GO team, 2007). Given that sound speed is by far the dominant factor influencing reflection coefficient (Sallarès et al., 2009), this provides a static representation of thermohaline finestructure at the location of the CTD casts. The solutions to generate the sections were obtained, up to 4th order accuracy in space and time, on an unstructured triangular mesh of about 20,000 elements. The resolution of the model was 12.5 m in both the horizontal and vertical directions. The velocity model had a 10 m resolution in the vertical dimension based on the model of Papenberg et al. (2010). It was interpolated horizontally to 10 m using a bilinear interpolation. The simulations were carried out on 10 processors of a Linux computing cluster. The synthetic data shot records were then processed using the same processing flow as the real data but, because of the limited virtual offset range generated in the modeling code compared to the real data set, only 5 frames could be created using the same offset group size.

The effect of frequency loss with ‘offset’ can be seen in the synthetic data, where there are no real motions (Fig. 4(b); Animation E). Notice how the synthetic data grow progressively in wavelength but show no intra-reflector variability like the real data. This progressive wavelength increase is seen as a thickening of reflectors in the synthetic data from frame to frame and is a normal result of the representation of frequency filtering as a function of offset in the physical model. The essential feature to observe in the synthetic data is neither the left-to-right apparent motions nor the thickening of reflectors but the fact that there is no intra-reflector variability like in the real thermohaline data. This result substantiates our claim that the application of the SOG technique to real seismic data shows real time movements of the thermohaline finestructure.

6. Discussion

The SOG method consists in the separation of seismic-streamer data into different offset groups to create unique stacked sections which are temporally spaced. Each subsequent section shows changes in the reflectivity pattern of thermohaline finestructure. We have used two different tests to verify that these changes are not an artifact of seismic data processing.

The first test is done with synthetic data. As expected, these data showed no changes in the reflectivity pattern other than the reduction of frequency content consistent with the way high frequencies become attenuated as a function of longer travel paths, i.e. due to the offset or distance between the source and receiver. Frames 1 to 5 in Fig. 4(b) (Animation E) show an apparent steady ‘motion’ from left to right but this is caused by the automatic addition, by the processing software, of duplicate traces to the leading edge of the model and are therefore not considered in our interpretation. However, intra-reflector changes seen in the real seismic oceanography data are not present in the synthetic data. This supports the notion that when a relatively static model of the ocean

based on in situ (CTD) measurements is created, one cannot estimate ocean dynamics to the same precision as with the SOG method. The second test is carried out by simultaneously comparing the seafloor and thermohaline reflectivity of the real data. Both data sets display progressive thickening of reflectors as a function of offset because the higher the frequency the greater it is attenuated by the medium. However, those boxes displaying the ocean reflectors exhibit time variability clearly not present within the seafloor (Fig. 4(b); Animation D). These two independent tests confirm that the fluctuations observed within the water column are likely real time displacements of isopycnals such as those caused by internal waves. Considering a characteristic velocity scale of $1\text{--}2\text{ ms}^{-1}$ (Garrett and Munk, 1979; Gill, 1982) we may estimate that at 3.5 min time intervals there would be fluctuations on the order of 200–400 m in the direction of wave propagation (obviously less if the wave phase velocity does not propagate in the same direction as the seismic stack); these movements should be easily observable on seismic data with a resolution of approximately 10 m.

Further justification for the potential of the SOG method comes from looking at a small part of the seismic section (250 m) along a single arbitrary reflector (Figs. 5 and 6), showing changes in the seismic amplitude and vertical location as a function of time step.

Fig. 5 also shows a contour map of the rise and fall of amplitudes for the positive samples (peaks) in each trace of the chosen reflector. If we take the vertical amplitude displacement of the reflector as indicative of true vertical isopycnal displacement we can estimate its oscillation distance. At a distance of 250 m we see a maximum of 4.5 ms displacements (one-way-travel-time). This displacement occurs from Frame 2 to 3, an interval of 3.5 min. Using an average sound speed of 1510 m/s, derived from the “normal move-out velocity” correction during seismic processing (Yilmaz, 2001), we estimate the displacements to be approximately 7 m ($1510\text{ m/s} \times 0.0045\text{ s}$), likely related to the presence of internal waves.

A more detailed analysis of the amplitudes shows that there is no preferential direction of change of amplitude as a function of offset group (i.e. animation frame) (Fig. 6). The displacement in two-way-time (TWT – the time for sound to reach a reflecting interface and back – a proxy for depth) changes by a few milliseconds. Considering there is little preferential change in amplitude with offset, it is plausible that the observed fluctuations are due to actual reflector displacements, i.e. isopycnal movements associated with traveling disturbances, not an amplitude-versus-offset (AVO) effect such as that used in the study by Páramo and Holbrook (2005).

7. Concluding remarks

By devising a processing scheme to make use of a long streamer as it passes over a fixed geodetic point, we have been able to generate sets of stacked seismic data from different offset groups, what we call Seismic Offset Group (SOG) processing. Because of the ubiquitous existence of traveling perturbations within the stratified ocean, each offset group would display geometrical differences in finestructure. The results of our analysis indicate that the observed thermohaline fluctuations are on the spatial scale of an oscillating internal wave field, although at this time we have no independent or simultaneous observational data that can corroborate this assertion.

A verification of the potential of the SOG method comes from two different tests. First, the seafloor reflectors in the same seismic section do not exhibit the same behavior as those in the ocean and, second, an identical processing scheme applied to synthetic data displays no noticeable intra-reflector motions. All three data sets (thermohaline, seafloor and synthetic) show a lowering of frequency content with offset. In the processing this was partially compensated by band-limiting the data to the minimum possible for the farthest offset group but there is still room for variability in frequency content within a given band-width. It remains a challenge to completely remove this effect as it masks somewhat the true movement of the dynamic ocean.

Unfortunately, given the limited data set it is difficult to put constraints on the velocity of the observed movements using this method alone. There are two main unknowns. First, the motion of the ship at the surface leads to a Doppler-like effect (Blacic and Holbrook, 2009). The observed wavelength is given by

$$\lambda_{\text{mod}} = \frac{v_{\text{ship}} \lambda}{v_{\text{ship}} - v}, \quad (7.1)$$

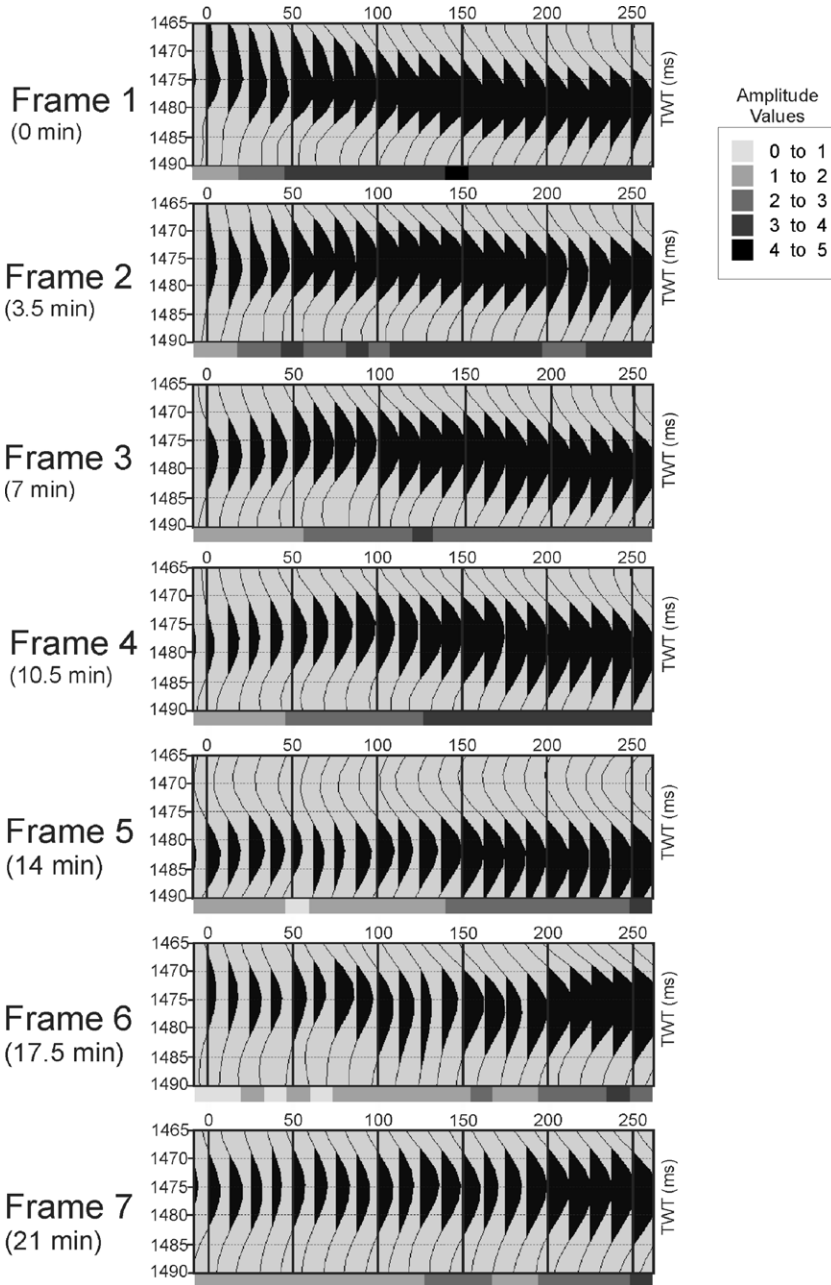


Fig 5. (a) Zoomed region of a single, arbitrary reflector (isopycnal surface) showing its displacement over seven time steps (21 min). Vertical axis is two-way-time (time for sound to travel to an acoustic impedance boundary and back) in milliseconds. Horizontal axis is distance in meters along reflector. Black vertical lines indicate position of peak amplitudes used for analysis shown in Fig. 6. Horizontal shaded scale at bottom of each frame shows the corresponding variation in amplitude of traces along the reflector. (b) Corresponding animation. See Fig. 5(b) in the online version at <http://dx.doi.org/10.1016/j.mio.2012.07.003>.

where v is velocity and λ is wavelength (Vsemirnova et al., 2009). And, second, the velocity vector of the internal wave field does not likely point in the plane of the seismic profile, i.e. it is most probably

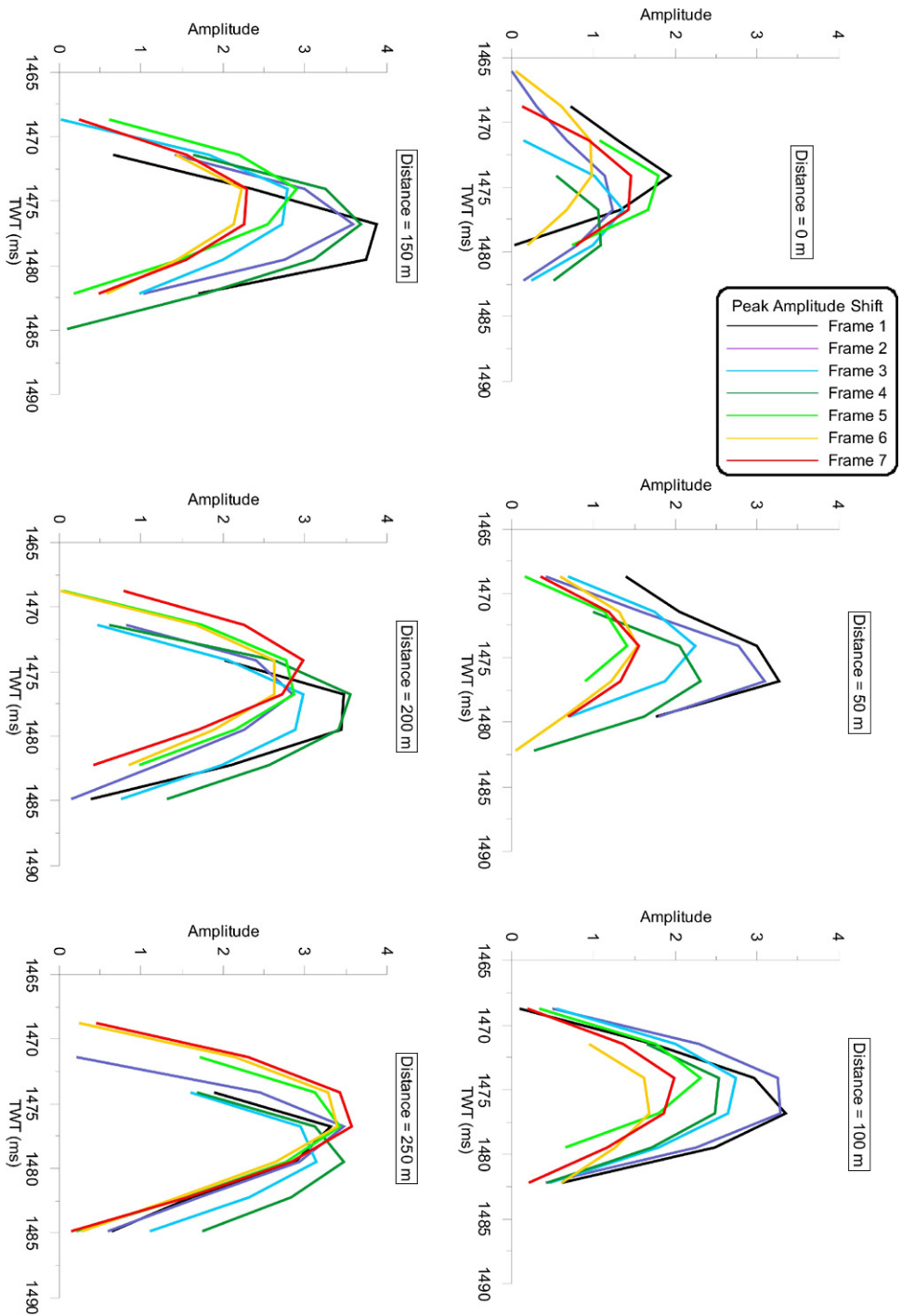


Fig. 6. Variation of position and amplitude as a function of time step (frame) of an arbitrary reflector (Fig. 5) at six intra-reflector locations (0, 50, 100, 150, 200 and 250 m). Compared to reflectors in the synthetic data and the seafloor, Frames 1–7 show distinctive fluctuations in the magnitude and vertical location of the seismic wavelet.

to be angled with respect to the ship's track. Despite these limitations, we have shown that the SOG method applied to real data shows fluctuations in thermohaline finestructure at depths between 900–1500 m (Fig. 4(b); Animations A–C) while the same fluctuations are not observed in either the seafloor (Fig. 4(b); Animation D) or the synthetic data (Fig. 4(b); Animation E).

Another, more practical, limitation of the SOG method is the necessary trade-off between maximizing CMP fold and maximizing the number of stacks (frames) necessary to temporally image ocean dynamics. The data used in this study were of rather good signal quality and were gathered with a relatively long streamer (5 km). Only 4 km of this streamer were used because, during seismic processing, normal move-out (NMO) corrections differentially stretch traces as a function of offset, requiring the distorted shallower events imaged by the last 1 km to be muted. It is for this reason that only the water column between 900 m and the seafloor is shown in Fig. 3. A longer streamer would allow more stacks to be created without reducing the offset range needed to optimize the signal-to-noise ratio; however the NMO stretch effect (Yilmaz, 2001) would remain a problem. This study, nonetheless, shows proof-of-concept.

Independent validation of the SOG method could come about in several ways. One way would be to acquire coincident and simultaneous in situ measurements of local internal waves (by using an acoustic Doppler current profiler and XBT probes, for example) while acquiring the seismic data with a sufficiently long streamer. A second, more technically complex alternative would be through joint measurements from an anchored/moored streamer of a given length together with repeated passes of a conventional airgun source, similar in geometry to a roll-along land seismic survey (Sheriff, 1991). Ideally, this would be done in a region of small or moderate surface currents but with significant undercurrents, high temperature/salinity contrasts and low ship traffic, such as the area imaged herein (Fig. 1). This area was chosen because of the high quality data set (Iberian–Atlantic Margin data) and previously observed changes in finestructure downstream of the Mediterranean Undercurrent (Buffett et al., 2009). In areas where there are more statically stable isopycnals, for example due to double diffusive processes, observing motions on the time scale of a passing streamer would be increasingly difficult.

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Appendix. Supplementary data

Supplementary material related to this article can be found online at <http://dx.doi.org/10.1016/j.mio.2012.07.003>.

References

- Biescas, B., Sallarès, V., Pelegrí, J.L., Machín, F., Carbonell, R., Buffett, G., Dañobeitia, J.J., Calahorrano, A., 2008. Imaging meddy finestructure using multichannel seismic reflection data. *Geophys. Res. Lett.* 35, L11609.
- Blacic, T.M., Holbrook, W.S., 2009. First images and orientation of fine structure from a 3-D seismic oceanography data set. *Ocean Sci.* 6, 431–439.
- Buffett, G.G., 2011. Seismic oceanography: a new tool to characterize physical oceanographic structures and processes. Ph.D. Thesis. <http://www.tesisenred.net/handle/10803/1939>.
- Buffett, G.G., Biescas, B., Pelegrí, J.L., Machín, F., Sallarès, V., Carbonell, R., Klaeschen, D., Hobbs, R.W., 2009. Seismic reflection along the path of the Mediterranean Undercurrent. *Cont. Shelf Res.* 29, 1848–1860. <http://dx.doi.org/10.1016/j.csr.2009.05.017>.
- Carbonell, R., Buffett, G.G., Sallarès, V., Biescas, B., Hobbs, R.W., 2010. Seismic visualization of thermohaline finestructure. (Abstract ID: PO41B-05). Ocean Sciences Meeting, Portland, OR, USA.

- Chadwick, R.A., Arts, R., Eiken, O., 2005. 4D seismic imaging of a CO₂ plume. In: Doré, A.G., Vining, B.A. (Eds.), *Petroleum Geology: North-West Europe and Global Perspectives*, Proceedings of the 6th Petroleum Geology Conference. Petroleum Geology Conferences Ltd., pp. 1385–1399.
- Cheng, X., 2008. Time-lapse imaging of fault properties at seismogenic depth using repeating earthquakes, active sources and seismic ambient noise. Ph.D. Thesis. ISBN: 9781109217353.
- Fer, I., Nandi, P., Holbrook, W.S., Schmitt, R.W., Páramo, P., 2010. Seismic imaging of a thermohaline staircase in the western tropical North Atlantic. *Ocean Sci.* 6, 621–631. <http://dx.doi.org/10.5194/os-6-621-2010>.
- Garrett, C., Munk, W., 1979. Internal waves in the ocean. *Annu. Rev. Fluid Mech.* 11, 339–369.
- Géli, L., Cosquer, E., Hobbs, R.W., Klaeschen, D., Papenberg, C., Thomas, Y., Menesguen, C., Hua, B.L., 2009. High resolution seismic imaging of the ocean structure using a small volume airgun source array in the Gulf of Cadiz. *Geophys. Res. Lett.* 36, L00D09. <http://dx.doi.org/10.1029/2009GL040820>.
- Gill, A., 1982. *Atmosphere–Ocean Dynamics*. Academic Press, p. 662.
- Hobbs, R.W., the GO team, , 2007. GO–Geophysical Oceanography: a new tool to understand the thermal structure and dynamics of oceans. D318 Cruise Rep., Durham Univ., Durham, UK. Available at: <http://www.dur.ac.uk/eu.go/cruise/report.html>.
- Holbrook, W.S., Páramo, P., Pearce, S., Schmitt, R.W., 2003. Thermohaline fine structure in an oceanographic front from seismic reflection profiling. *Science* 301, 821–824.
- Käser, M., Dumbser, M., 2006. An arbitrary high order discontinuous Galerkin method for elastic waves on unstructured meshes I: the two-dimensional isotropic case with external source terms. *Geophys. J. Int.* 166 (2), 855–877. <http://dx.doi.org/10.1111/j.1365-246X.2006.03051.x>.
- Klaeschen, D., Hobbs, R.W., Krahmann, G., Papenberg, C., Vsemirnova, E., 2009. Estimating movement of reflectors in the water column using Seismic Oceanography. *Geophys. Res. Lett.* 36, L00D03. <http://dx.doi.org/10.1029/2009GL038973>.
- Krahmann, G., Brandt, P., Klaeschen, D., Reston, T., 2008. Mid-depth internal wave energy off the Iberian Peninsula estimated from seismic reflection data. *J. Geophys. Res.* 113, C12016. <http://dx.doi.org/10.1029/2007JC004678>.
- Lumley, D., 2001. Time-lapse seismic reservoir monitoring. *Geophysics* 66 (1), 50–53.
- Nakamura, Y., Noguchi, T., Tsuji, T., Itoh, S., Niino, H., Matsuoka, T., 2006. Simultaneous seismic reflection and physical oceanographic observations of oceanic fine structure in the Kuroshio extension front. *Geophys. Res. Lett.* 33, L23605. <http://dx.doi.org/10.1029/2006GL027437>.
- Nandi, P., Holbrook, W.S., Páramo, P., Pearce, S., Schmitt, R.W., 2004. Seismic reflection imaging of water mass boundaries in the Norwegian Sea. *Geophys. Res. Lett.* 31, L23311. <http://dx.doi.org/10.1029/2004GL021325>.
- Papenberg, C., Klaeschen, D., Krahmann, G., Hobbs, R.W., 2010. Ocean temperature and salinity inverted from combined hydrographic and seismic data. *Geophys. Res. Lett.* 37, L04601. <http://dx.doi.org/10.1029/2009GL042115>.
- Páramo, P., Holbrook, W.S., 2005. Temperature contrasts in the water column inferred from amplitude-versus-offset analysis of acoustic reflections. *Geophys. Res. Lett.* 32, L24611. <http://dx.doi.org/10.1029/2005GL024533>.
- Pinheiro, L., Song, H., Ruddick, B., Dubert, J., Ambar, I., Mustafa, K., Bezerra, R., 2010. Detailed 2-D imaging of the Mediterranean outflow and meddies off w Iberia from multichannel seismic data. *J. Mar. Syst.* 79, 89–100.
- Ruddick, B., Song, H., Dong, C., Pinheiro, L., 2009. Water column seismic images as maps of temperature gradient. *Oceanography* 22 (1), 184–197.
- Sallarès, V., Biescas, B., Buffett, G., Carbonell, R., Dañobeitia, J.J., Pelegrí, J.L., 2009. Relative contribution of temperature and salinity to ocean acoustic reflectivity. *Geophys. Res. Lett.* 36, L00D06. <http://dx.doi.org/10.1029/2009GL040187>.
- Sheen, K.L., White, N.J., Hobbs, R.W., 2009. Estimating mixing rates from seismic images of oceanic structure. *Geophys. Res. Lett.* 36, L00D04. <http://dx.doi.org/10.1029/2009GL040106>.
- Sheriff, R.E., 1991. *Encyclopedic Dictionary of Exploration Geophysics*, third ed. Society of Exploration Geophysics, ISBN: 1-56080-018-6.
- Stommel, H., Federov, K.N., 1967. Small scale structure in temperature and salinity near Timor and Mindanao. *Tellus* 19, 306–325.
- Tsuji, T., Noguchi, T., Niino, H., Matsuoka, T., Nakamura, Y., Tokuyama, H., Kuramoto, S., Bangs, N., 2005. Two-dimensional mapping of fine structures in the Kuroshio Current using seismic reflection data. *Geophys. Res. Lett.* 32, L14609. <http://dx.doi.org/10.1029/2005GL023095>.
- VonLanken, A., Holbrook, W.S., Simmons, H., Fortin, W.F., Eakin, D., St Laurent, L., 2012. Investigation of nonlinear internal wave (NLIW) motion in the South China Sea. (Abstract ID:12496). Ocean Sciences Meeting. Salt Lake City, UT, USA.
- Vsemirnova, E., Hobbs, R.W., Serra, N., Klaeschen, D., Quentel, E., 2009. Estimating internal wave spectra using constrained models of the dynamic ocean. *Geophys. Res. Lett.* 36, L00d07. <http://dx.doi.org/10.1029/2009GL039598>.
- Yilmaz, O., 2001. *Seismic Data Analysis: Processing, Inversion, and Interpretation of Seismic Data*, Vol. II, second ed. In: Invest. Geophys., vol. 10. Soc. Explor. Geophys., Tulsa, Okla., p. 2027.