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An implosive seismoacoustic source for seismic experiments on the ocean floor

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Abstract

Bottom shots have been used for a number of years in seismic studies on the ocean floor. Most experiments utilized explosives as the energy source, though researchers have recognized the usefulness of collapsing water voids to produce seismoacoustic signals. Implosive sources, however, suffered generally from a lack of control of source depth. We present a new experimental tool, called SEEBOSEIS, to carry out seismic experiments on the seafloor utilizing hollow glass spheres as controlled implosive sources. The source is a 10-inch BENTHOS float with penetrator. Inside the sphere we place a small explosive charge (two detonators) to destabilize the glass wall. The time of detonation is controlled by an external shooting device. Test measurements on the Ninetyeast Ridge, Indian Ocean, show that the implosive sources can be used in seismic refraction experiments to image the subbottom P-wave velocity structure in detail beyond that possible with traditional marine seismic techniques. Additionally, the implosions permit the efficient generation of dispersed Scholte waves, revealing upper crustal S-wave velocities. The frequency band of seismic energy ranges from less than 1 Hz for Scholte modes up to 1000 Hz for diving P-waves. Therefore, broadband recording units with sampling rates $>$2000 Hz are recommended to sample the entire wave field radiated by implosive sources.

Introduction

Despite the fact that seismic energy must always propagate through the uppermost several hundred metres of crust to be detected by either ocean bottom seismographs or sonobuoys, nature appears to conspire against investigation of this otherwise most accessible portion of oceanic crust. The geometrical constraint of a thick water layer masks many arrivals that travel through the upper crust. To isolate rays turning near the seabed a concomitant isolation of other interfering energy paths is required. Therefore, seismic velocity determination of uppermost crust took a major leap forward when both the receiver and the source were placed on or close to the seafloor (e.g., Purdy, 1987). Such experiments are akin to refraction experiments on land with geophones and seismic sources on the land surface.

It has been 60 years since marine seismologists first used bottom shots. These experiments described by Ewing et al. (1946) and most subsequent studies utilize explosive charges as sources (e.g., Davis et al., 1976; Purdy, 1986; Koelsch et al., 1986; Kirk et al., 1991), but heavy weights dropped onto the seafloor have also been used (Davis et al., 1976; Whitmarsh and Lilwall, 1982). In addition, researchers have recognized the usefulness of collapsing water voids at depth to produce seismoacoustic signals. Davis et al. (1976) constructed steel tubes 15 cm in diameter and roughly 4 m long. The tubes were weighted at one end with concrete fill to ensure vertical falling stability and an explosive charge with a delay fuse was

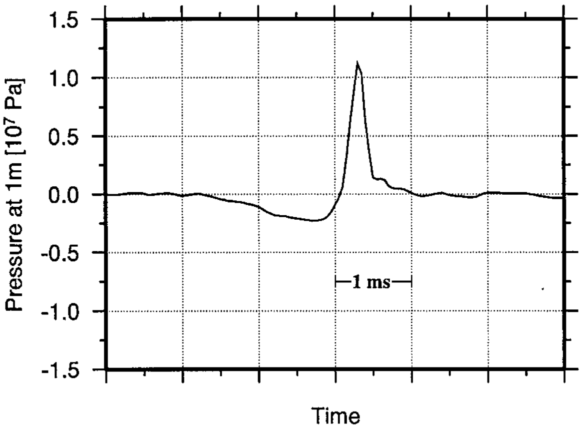

Figure 1. Acoustic signal from an imploding glass sphere at 2660 m depth (Orr and Schoenberg, 1976).

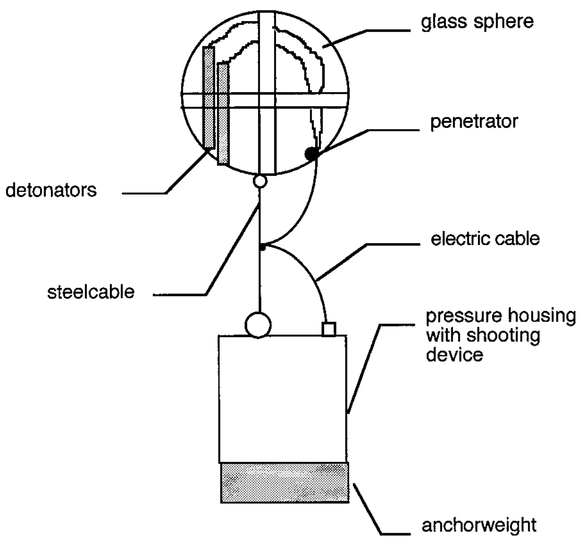

Figure 2. Principal design of the seismoacoustic implosive source for seismic experiments on the ocean floor (not to scale).

placed outside at the top to remove the upper seal upon detonation and allow violent implosion of water into the tube. These sources, however, suffered from the unknown sinking rate and delay time between deployment and detonation. Orr and Schoenberg (1976) examinated the potential of spherical glass floats. In their tests, they dropped preweakened glass spheres through the water column, recorded the implosion and analyzed the sounds emitted. These sources, however, were lowered into the sea until increasing water pressure caused failure. Thus as for the steel tubes the lack of control of depth makes it impossible to use this technique for controlled source on-bottom seismic studies. Nevertheless, due to the impulsive character

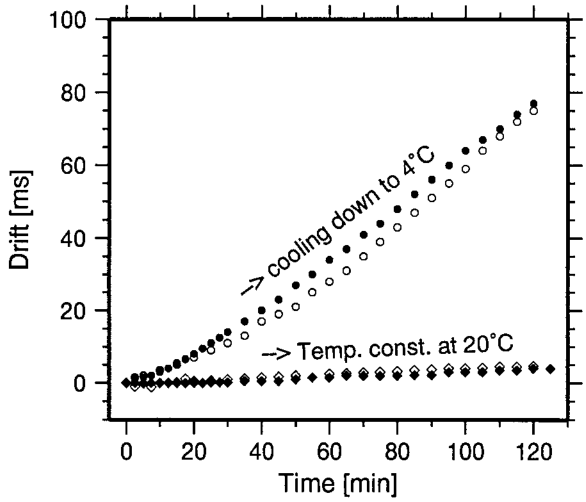

Figure 3. Drift function of the quartz oscillator used to determine the time of ignition. The functions can be used to assess the source time more precisely. 20 °C is taken to be the temperature in the laboratory; 4 °C is assumed to be the temperature at the seabed. Each device was tested twice, providing two sets of measurements (open and solid symbols).

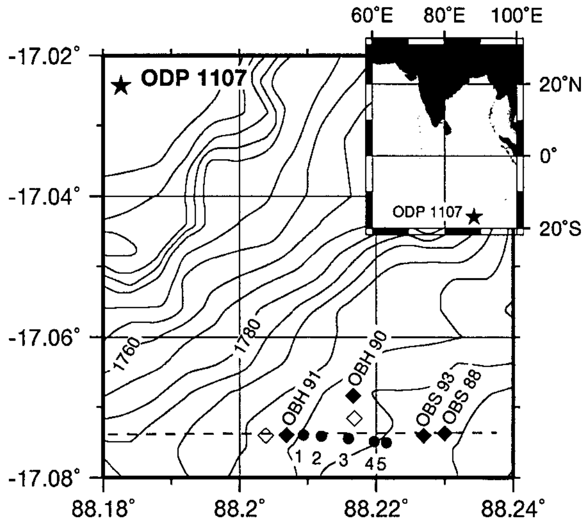

Figure 4. Location map of the field test on Ninetyeast Ridge, Indian Ocean, near ODP site 1107. OBH/OBS stations are squares, solid squares are the stations used in this study; bottom shots are solid circles.

and the high peak pressure of the signal that is radiated upon implosion (Figure 1), glass spheres show promise as a tool in the study of uppermost crustal structure. With more than 10^7 Pa at 1 m the pressure generated by an imploding glass sphere is more than 10 times higher than the peak pressure of a 8-l or 32-l airgun, providing 3.3×10^5 Pa at 1 m and 7.1×10^5 Pa at 1 m, respectively. In contrast to surface shots, bottom shots do not suffer from the divergence of seismic energy while the sound wave is propagating through the water column. Thus, the entire energy radiated by a bottom source reach the seabed. However, with increasing source depth the bubble pulse frequency of seismic sources increases. Therefore, frequency dependent attenuation and scattering might be important for bottom shots.

There have been unpublished reports which show that the behaviour of glass under high pressure prevented reliable initiation of collapse of BENTHOS floats (Sam Raymond, personal communication). Nevertheless, we took on the challenge to develop a new experimental tool, called SEEBOSEIS, to carry out on-bottom seismic experiments utilizing BENTHOS glass spheres as seismoacoustic implosive sources. The use of such sources deployed on the ocean floor will allow us to study the uppermost crustal velocity structure in detail beyond that possible with conventional marine seismic techniques.

The experimental tool

The SEEBOSEIS system consists of two separate components: a broadband ocean-bottom recording system and the implosive sources. Both components are operated and controlled separately. For the experiment described here, we deployed two kinds of ocean bottom recording systems: (i) prototype recorders developed at the University of Hamburg. These instruments are characterized by a dynamic range of 90 dB and are capable of recording frequencies of 1 Hz to 15 kHz; (ii) newly developed Marine Broadband Seismic (MBS) recorders developed by SEND GmbH with support from the GEOMAR Research Centre. These instruments have a dynamic range of >100 dB. Depending on the sensor (OAS E-2PD hydrophone or 4.5 Hz Lennartz geophone), the pre-amplifier and the sampling rate chosen to digitize the wave field, frequencies of $\sim$ 0.1 Hz up to some 1000 Hz can be detected. The hydrophones will have flat responses up to the Nyquist frequency. The geophones, however, may exhibit spurious resonances far above their natural frequencies. We therefore used primarily hydrophone data to draw our conclusions. As instrument housing we used GEOMAR pop-up systems (Flüh and Bialas, 1996) and a Hamburg-OBS (Herber et al., 1981). However, any other ocean bottom instrument could be used instead, though the high bubble pulse frequency of implosions at depth (Orr and Schoenberg, 1976) requires recording systems with sampling rates of 500 to 1000 Hz, or higher. In addition, bottom shots permit the generation of dispersed interface waves (Scholte modes) and these wavetrains may have frequencies lower than 1 Hz (Nolet and Dorman, 1996; Essen et al., 1981, 1998). Therefore, broadband recording units are recommended to sample the entire wave field radiated by the bottom shots. In practice, however, it might be reasonable to use a high frequency system for the hydrophones and a low frequency system for the geophones.

The implosive source itself is a 10-inch (25.4 cm) BENTHOS hollow glass sphere with penetrator (i.e., a hole in the wall to allow an electrical connector to run into the sphere). In the sphere, a small explosive charge (two electrical detonators, about 1 g of explosives) was fixed on the 0.9 cm thick glass wall (Figure 2). When ignited by an external shooting device the detonators destabilize the glass sphere and in turn the ambient water pressure causes the implosion. The attraction of using such spheres is a consequence of the large amount of potential energy available in the ocean when a cavity is formed. The total energy is equal to the pressure times the interior volume so that our 23.6 cm cavity at a depth of 2000 m has a potential energy of about 10^5 J. The work of Orr and Schoenberg (1976) suggest that the efficiency of the conversion of available potential energy to acoustic energy is approximately 15–20%.

The detonation itself is controlled by an electronic shooting device, which has two main components: a compact electronic timer and a capacitor. The timer has to control the shooting time. It is built completely of HCMOS technology for low power operation and high speed performance. The capacitor is used to produce a detonation voltage of more than 100 V DC, necessary to fire the detonators. The electronic device itself is powered by two 9 V batteries, providing about 20 h of operating time. To guarantee an exact start time, the device is synchronized with a reference clock (for example a DCF signal); dip switches allow the time until ignition to be set between 1 and 65 535 s. At the detonation time, a relay switches the voltage from the capacitor to the detonators with a time delay of less than 1 ms. To minimize the time drift of the clock, we used a programmable quartz oscillator with a frequency variation of 5 ppm. Each device, however, will have its own characteristic drift. To determine the drift, we run the devices under temperature conditions of 20 °C and 4 °C (assumed to be the temperature in the laboratory and at the seabed), respectively. Consid-

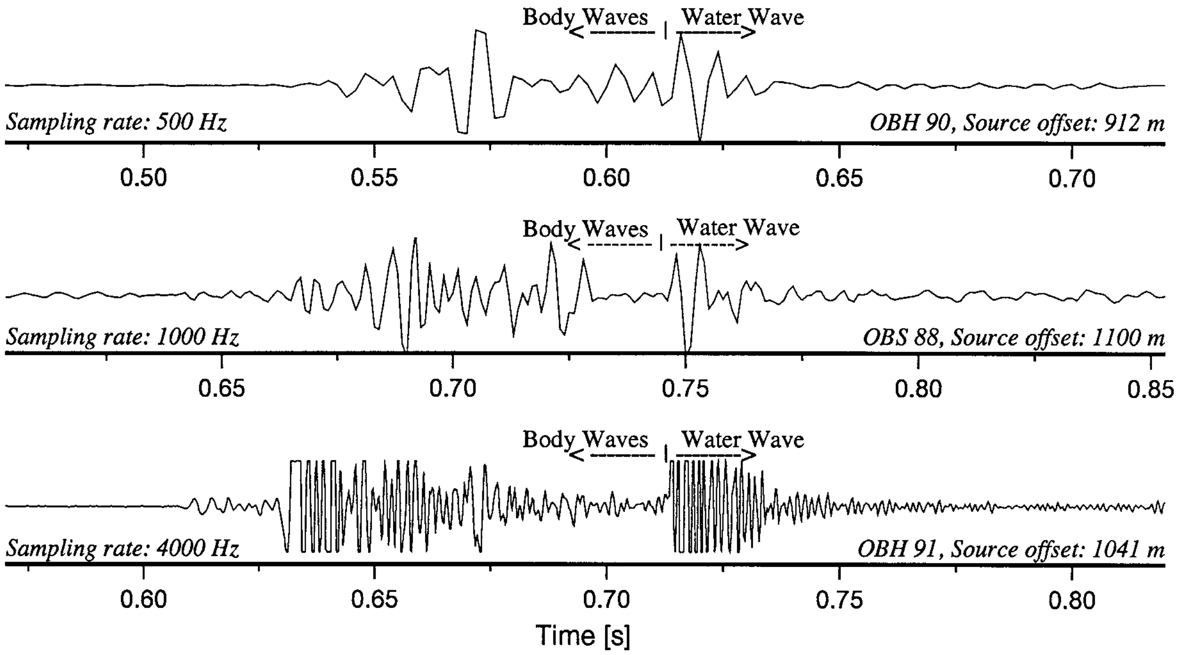

Figure 5. Wavetrains from bottom shots recorded on hydrophones by instruments with different sampling rates. OBS 88 and OBH 90 are high pass filtered at 40 Hz. The recording unit of OBH 91 discriminates frequencies less than 40 Hz.

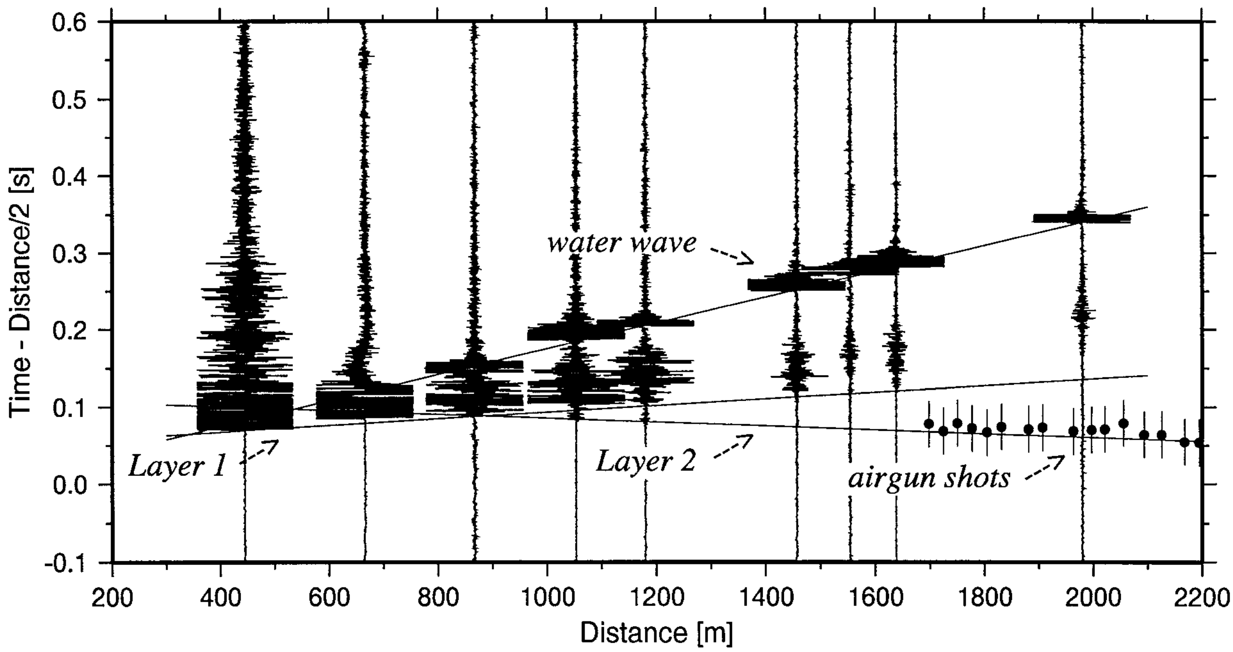

Figure 6. Seismograms (reduced by 2 km s^{-1}) recorded on the hydrophones of OBH 91 and OBS 93. The shot at 1460 m offset was used to calculate the power spectrum shown in Figure 7. In addition, we included water path corrected airgun shots and mark the water wave and body wave arrivals (layer 1 and layer 2) by straight lines. Most profoundly, surface shots did not allow the detection of arrivals at shot-receiver ranges of less than 1700 m. The only way to study these offsets is the deployment of bottom shots. The bottom shots detected at least one layer not imaged by the surface shots.

ering the drift functions (Figure 3), we can determine the shooting time within 10–15 ms. However, for the experiment we deployed several sensors on the ocean floor (Figure 4). Therefore, the position of the source relative to the sensors and the time of detonation can be re-calculated by applying techniques used to determine local earthquakes parameters (cf., Klein, 1978).

Field test

During R/V *Sonne* cruise 131 (Flüh and Reichert, 1998) we carried out test measurements near Ocean Drilling Program (ODP) site 1107, Ninetyeast Ridge (Indian Ocean). The hole was drilled in Spring of 1998 by ODP's drilling vessel *Joides Resolution*. A location about 5 km to the south of the site provided a flat seabed with horizontally layered strata (Flüh and Reichert, 1998). The average water depth was about 1825 m. Two pairs of ocean bottom sensors (i.e., either ocean bottom seismographs [OBS] or ocean bottom hydrophones [OBH]) were placed along a west-east trending line (Figure 4). A 2-km line of five bottom shots was placed between the OBH/OBS pairs. A third pair of OBHs was placed about 600 m off line. Both instruments and bottom shots were deployed by free fall using Global Positioning System (GPS) for droppoint positioning. The instrument locations on the seabed were further constrained using water path arrival times from airgun shots collected while the ship was navigated with GPS. The sound speed model used to calculate the ranges is based on conductivity-temperature-depth (CTD) measurements (Flüh and Reichert, 1998). In addition, we should mention that for safety concerns, we first lowered the glass spheres down to the water before we connected the shooting devices with the detonators inside the spheres.

Four instruments provided data useful for geophysical interpretation. Thus 20 shots were detected, covering shot-receiver offsets of 445 to 2190 m. Shot to receiver ranges were calculated from the arrival times of the direct waves and from the sound speed at the seabed. Examples of bottom shots are shown in Figures 5 and 6. Surprisingly, even at shot-receiver offsets of 900 to 1100 m the recording units are overloaded by both the direct wave traveling along the seabed and diving P-waves. We had expected that only the direct wave would overload the instruments at shot-receiver offsets of less than 1000 m. However,

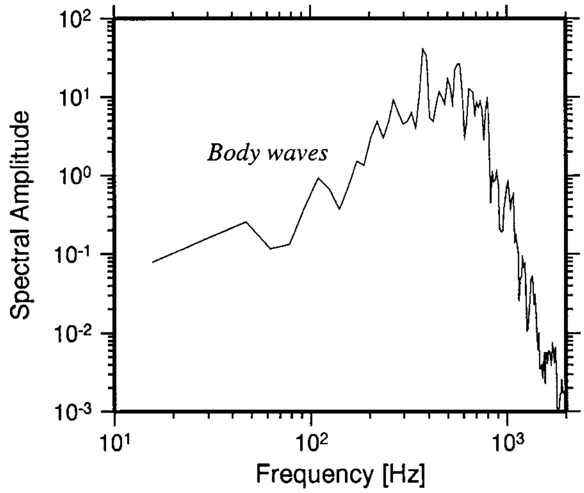

Figure 7. Spectrum of shot no. 4 recorded on the hydrophone channel of OBH 91 (offset: 1460 m; see Figure 6).

even at offsets of 1600 to 2200 m the receivers clipped the signals. This behaviour clearly proves the effectiveness of the implosive sources, because P-waves from airgun shots never overloaded the recording systems. Furthermore, it may indicate that the dynamic range of any available digital instrument might be insufficient to recover true amplitudes from deep water implosions. Nevertheless, an initial assessment of the entire wave field might be possible by using different amplification rates for different channels to sample near and far field separately.

Figure 5 has wavetrains from bottom shots detected by instruments with sampling rates of 500, 1000 and 4000 Hz. Most profoundly, the detected wave field is biased by the sampling rate. Although OBS 90 (sampling rate 500 Hz) has recorded first breaks, there might be aliasing in the data. A comparision of wavetrains recorded by OBS 88 (sampling rate 1000 Hz) and OBH 91 (resampled at 4000 Hz) may suggest that 1000 Hz might be sufficient to sample first breaks. Nevertheless, the image of the P-wave field sampled at 4000 Hz is much clearer. Figure 7 has the power spectrum from shot no. 4 recorded on OBH 91 (source offset 1460 m; Figure 6). Most of the energy carried by body waves is in a frequency band of 200–800 Hz, with little energy at lower or higher frequencies. The uppermost crust of the Ninetyeast Ridge, however, may act as a low-pass filter. We therefore recommend instruments with sampling rates of more than 2000 Hz to detect implosive bottom shots. The water wave, however, may also provide energy at much higher frequencies.

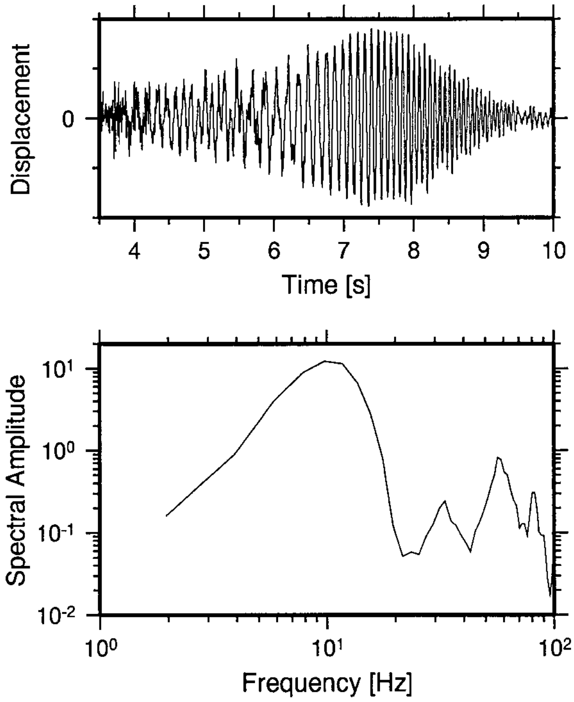

Figure 8. Wavetrain and power spectrum of a dispersed Scholte wave detected on the vertical geophone of OBS 88.

Figure 6 shows the advantage of bottom shots recorded on the seabed compared to conventional surface shots recorded on either ocean bottom seismographs or sonobuoys. Because the strong water wave arrivals interfere with comparatively weak arrivals from energy turning in the shallow crust, the water wave tend to mask those arrivals. Therefore, experiments using surface sources are generally not able to provide direct evidence for the seismic properties of the shallow crust. The experiment presented in this paper was carried out at a water depth of $\sim$ 1825 m. The OBHs recorded diving P-waves from airgun shots at offsets of about 3000 m (Grevemeyer et al., 1999). Corrected for the water path the arrivals correspond to offsets on the seabed of more than 1700 m (Figure 6). The bottom shots, however, covered offsets from about 400 m to 2000 m. As indicated in Figure 6, they provided arrivals from at least one layer not detected by the surface shots. Thus, modelling of surface shots only would be biased. However, the figure also indicates that frequency-dependent attenuation and scattering may control the propagation of seismic energy in surficial deposits. Grevemeyer et al. (1999) found an excellent agreement between the structure sampled by drilling and the structure derived from both bottom sources and surface shots, suggesting that layer 1 is composed of nannofossil ooze while layer 2 provided volcaniclastic material. Rays turning within the pelagic sediment of layer 1 are clearly detectable out to 2000 m offset, while energy propagating within the volcaniclastics is rapidly attenuated. On the other hand, airgun shots provided clear arrivals from layer 2. This behaviour may indicate that attenuation and scattering is stronly dependent on the signal frequency and lithology. Compared to airgun shots, high-frequency implosive sources are inherently affected by attenuation and scattering; airgun shots and implosive bottom shots are characterized by frequencies of 10–60 Hz and 200–800 Hz, respectively. Nevertheless, the deployment of bottom shots is the only way to detect seismic energy traveling through the shallow crust. Thus bottom shots are recommended for all studies interested in the seismic properties of the uppermost crustal rocks.

Near-bottom seismic experiments generally allow a determination of P-wave velocities by modelling the detected P-wave arrivals (e.g., Purdy, 1987; Christeson et al., 1994a; Grevemeyer et al., 1999), while S-wave velocities of unlithified and partially lithified marine sediments have been difficult to measure. However, the ability to locate a seismic source at the seafloor also permits the efficient generation of dispersed Scholte waves that travel along the ocean floor as Rayleigh waves travel along the surface of the Earth (e.g., Essen et al., 1981; 1998; Nolet and Dorman, 1996). Indeed, after the water wave a wave group traveling with a speed of some 200 m s^{-1} is detected. This slow wave is clearly dispersed and provides energy at frequencies of less than 20 Hz (Figure 8). Dispersion analysis (e.g., Essen et al., 1981; 1998) or waveform inversion (Nolet and Dorman, 1996) may yield upper crustal S-wave velocities from the wavetrain.

Determination of true shot locations

The determination of the locations of bottom shots deployed by free falling is generally difficult and tenuous. So far, shot-receiver offsets have been constrained using arrival times and arrival time differences of the direct and surface reflected water waves (e.g., Davis et al., 1976; Kirk et al., 1991), though Hammer et al. (1994) and Wiggins et al. (1996) used transponder navigation to determine the position of bottom shots. This technique, however, is limited to a couple of shots and is comparatively expensive. In recent years the

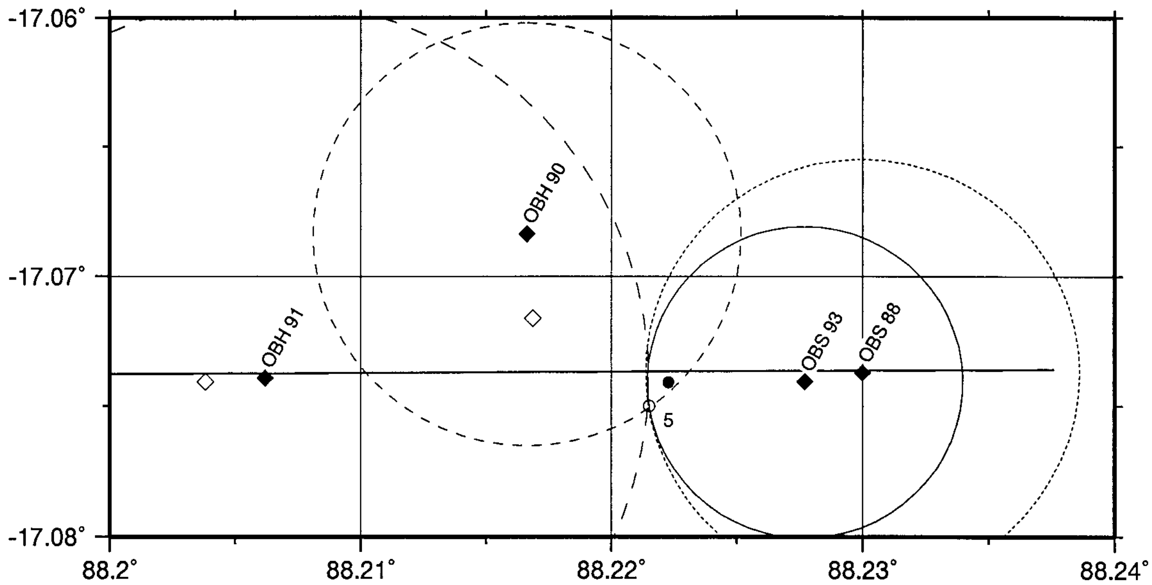

Figure 9. Relocated position (open circle) of bottom shot No. 5. The position was estimated by applying acoustic triangulation.

number of OBSs available for marine seismic experiments increased significantly. Therefore, techniques and computer codes for local earthquake studies can be used to relocate bottom shots fired within an array of ocean bottom seismographs. Using the velocity at the seabed derived from CTD measurement, we were able to estimate the true shot location on the ocean floor. We assumed receiver positions and depths on the seafloor (calculated from the travel time of the direct water wave between airgun shots at the sea surface and the seabed receivers) to be fixed. The Hamburg-OBS, however, was attached to a buoy floating at the sea surface and therefore suffered from bad sea conditions. Figure 9 shows the relocated position of shot No. 5. The histogram in Figure 10 shows the different between the shot locations calculated from the best fit shot location from all instruments and from the relative epicentral distance calculated for each instrument individually. Because the Hamburg-OBS was pulled over the seabed, only pop-up instruments have been used to calculate the final shot location. These recordings provide ranges within 15 m (r.m.s.-error) of the best fit location.

These parameters allow us to study the drift of the free falling shots (Figure 11). The absolute drift on the way down to the ocean floor is less than 180 m. By calculating the longitudinal and latitudinal components, we observed a systematic change in latitudinal direction, while longitudinal drift seems to vary randomly. All five shots were deployed within 30 min.

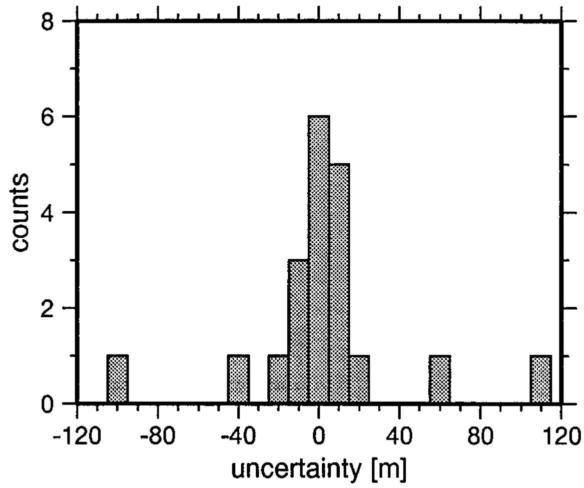

Figure 10. Histogram showing the uncertainty of shot locations. Note, the largest uncertainties of >40 m are from OBS 93. The instrument was attached to a buoy floating at the sea surface. Due to bad sea conditions the instrument was pulled over the seabed. 93% of the pop-up instruments, however, have a r.m.s.-error of 8.6 m.

Therefore, temporary variations of the wind and/or sea state or changes of currents within the water column may affect the location of bottom shots significantly.

Future improvements

Our test measurements with seismoacoustic implosive sources clearly indicate that the sampling of the en-

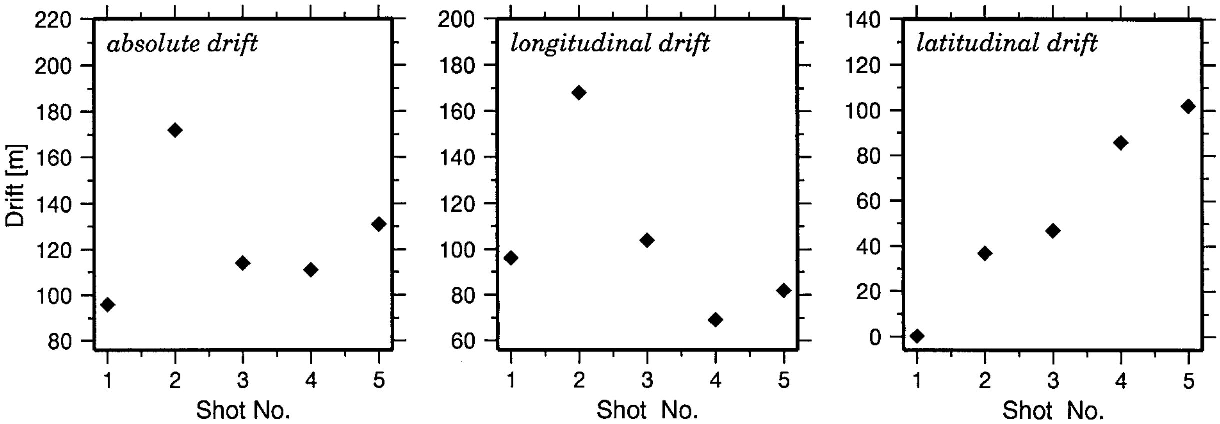

Figure 11. Drift of the bottom shots with respect to the deployment position.

tire wave field radiated by imploding BENTHOS glass spheres is a critical issue. For the identification and separation of different wave groups and for the assessment of the onset time, we need unclipped signals. At small shot-receiver ranges, for example, it is difficult to separate the direct water wave from diving body waves. But also at offset of about 2000 m the direct wave overloads the receivers. We believe that the dynamic range of any available digital recording unit will be insufficient to recover the entire wave field. Therefore, we suggest recording the near field and the water wave of the far field separately from the far field P-waves. We roughly estimate that amplification factors must differ by about 60 dB.

During the experiment described here, we deployed each source separately and very carefully. The drift rates (Figure 11), however, suggested that the sources have suffered from strong passat winds during Spring time in the Central Indian Ocean, though the rates may represent a reasonably worse case. We believe that heavier ground weights may reduce the drift by a shorter stay in the water column.

Conclusions and outlook

Over the last three decades, explosive charges have been used as the energy sources for seismic experiments on the ocean floor. They have been used for both generating dispersed Scholte waves and for seafloor refraction experiments (e.g., Purdy, 1987; Kirk et al., 1991; Nolet and Dorman, 1996). The usage of implosive sources, however, will eliminate safety concerns associated with explosives, i.e., compared to several kilograms of explosive used for previous experiments, we only have to handle the detonators. Furthermore, we are planning to replace the detonators by a device not restricted by any public law. We therefore suggest that glass sphere utilized as seismoacoustic implosive sources can replace explosives in most on- or near-bottom seismic experiments.

The results presented here indicate that the implosive sources radiate sufficient energy to detect P-waves at shot-receiver ranges of more than 2000 m. These arrivals have been turned within the uppermost 500–1000 m of the crust. Traditional marine seismic refraction experiments generally did not allow the study of this portion of crust, i.e., the geometrical constraint of the water layer masks those arrivals. Therefore, the implosive sources will allow us to study the shallow seismic structure in detail beyond that possible with conventional seismic techniques. In addition, the implosions permit the efficient generation of Scholte modes, which are closely related to the S-wave velocity of the immediately subjacent sediment. The detection of P-waves and Scholte modes, however, also allow the study of P- and S-wave attenuation as well (Christeson et al., 1994b; Nolet and Dorman, 1996). The shallow seismic structure and sediment properties from seafloor seismic experiments provide important information for the planning of marine drill sites. Therefore SEEBOSEIS might be an important tool for ODP presite surveys. However, there are other targets for on-bottom seismic studies, like the investigation of upper crustal accretion at seafloor spreading centres (e.g., Christeson et al., 1994a,b), revealing age-dependent trends in upper crustal structure (e.g., Grevemeyer and Weigel, 1997; Grevemeyer et al., 1998) or studying seismic velocities of gas hydrate bearing sediments at continental margins (e.g., Hyndman and Spence, 1992; Singh et al., 1993).

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