HYDROACOUSTIC MONITORING OF OCEANIC SPREADING CENTERS

Past, Present, and Future

BY ROBERT P. DZIAK, DELWAYNE R. BOHNENSTIEHL, AND DEBORAH K. SMITH

Ocean bottom hydrophone within the caldera of Axial Seamount, Juan de Fuca Ridge. *Photo courtesy of William Chadwick, Oregon State University/NOAA*

ABSTRACT. Mid-ocean ridge volcanism and extensional faulting are the fundamental processes that lead to the creation and rifting of oceanic crust, yet these events go largely undetected in the deep ocean. Currently, the only means available to observe seafloor-spreading events in real time is via the remote detection of the seismicity generated during faulting or intrusion of magma into brittle oceanic crust. Hydrophones moored in the ocean provide an effective means for detecting these small-magnitude earthquakes, and the use of this technology during the last two decades has facilitated the real-time detection of mid-ocean ridge seafloor eruptions and confirmation of subseafloor microbial ecosystems. As technology evolves and mid-ocean ridge studies move into a new era, we anticipate an expanding network of seismo-acoustic sensors integrated into seafloor fiber-optic cabled observatories, satellite-telemetered surface buoys, and autonomous vehicle platforms.

INTRODUCTION

New seafloor is magmatically emplaced, cooled, and then faulted within the mid-ocean ridge (MOR) system, forming one of the most active and longest belts of seismicity on the planet (Figure 1). Detection of MOR magma intrusion and rifting events is critical to our understanding of the global pace of volcanism and the ephemeral physical and chemical impacts these events have on the ocean and seafloor ecosystems. MOR magmatic and tectonic activity, however, typically has no expression at the sea surface, and therefore our only means of identifying these events is via remote detection of small-magnitude (typically $M < 4$) earthquakes caused by intrusion of magma and faulting of brittle oceanic crust. One of the most effective means of detecting small-magnitude, deep-ocean seismicity is to record an earthquake's hydroacoustic phase, or "*T*-phase," that enters and propagates within the ocean's sound channel.

This article discusses the hydroacoustic monitoring methods used to study mid-ocean ridges from the first discovery of seismically generated acoustic phases in the 1920s up to the present, and speculates upon future applications of these techniques. The spreading rate of an MOR largely determines the character of seismicity it produces. Fast-spreading ridges exhibit significant magmatic activity; however, the earthquakes tend to be small and more difficult to detect regionally (Figure 1). At slow-spreading ridges, fault processes dominate, exemplified by more frequent, large-magnitude earthquakes but sporadic magmatic activity. MOR volcanism can be detected by (1) earthquakes caused from magma intrusion through oceanic crust and/or faulting resulting from the intrusion, (2) volcanic tremor produced by flow of either magma or hydrothermal fluid through oceanic crust, or (3) violent magmatic explosions from degassing. All three volcano-acoustic sources can be useful for remote detection of volcanism on MORs because of their unique signal characteristics and because they are the loudest natural sounds recorded in the ocean.

Because of the combined dependence of ocean sound speed on pressure and temperature, much of the global ocean exhibits a low-velocity region known as the SOund Fixing And Ranging (SOFAR) channel (Figure 2). The channel's axis lies ~ 1,000 m below the sea surface in equatorial regions, becoming progressively shallower and then disappearing near the poles in response to the changing temperature structure of the water column. Seismically generated acoustic energy may become trapped in the SOFAR channel, where it propagates laterally in range (r) via a series of upward- and downward-turning refractions and has little interaction with the seafloor or sea surface. The attenuation due to geometric spreading is cylindrical ($\sim 1/r$) for SOFAR guided waves, making transmission significantly more efficient relative to solid Earth phases that undergo spherical spreading loss ($\sim 1/r^2$) and allowing acoustic phases to propagate thousands of kilometers with little energy loss.

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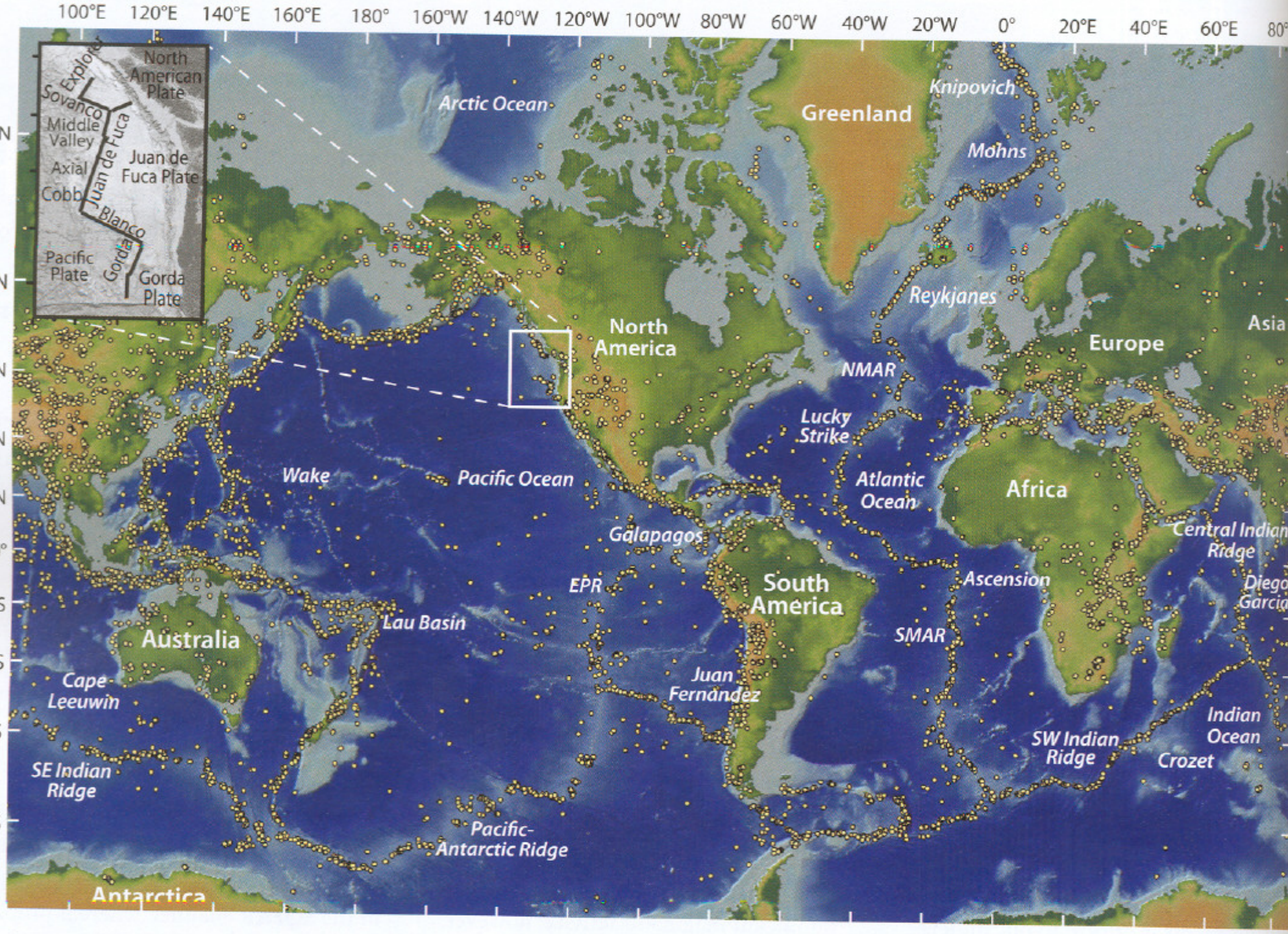

Figure 1. Global map of seismicity from the National Earthquake Information Center catalog, M ≥ 4.5, 1976–2009. Mid-ocean ridges and oceanic transforms are defined by narrow bands of shallow hypocenter earthquakes. Spreading centers and approximate full spreading rates:

- East Pacific Rise (EPR), ~ 110–140 mm yr^{-1}
- Pacific-Antarctic Ridge, 65 mm yr^{-1}
- Galápagos Spreading Center, ~ 45–60 mm yr^{-1}
- Northern Mid-Atlantic Ridge (NMAR), 25 mm yr^{-1}
- Southern Mid-Atlantic Ridge (SMAR), ~ 30 mm yr^{-1}
- Central Indian Ridge, ~ 35 mm yr^{-1}
- Southwest Indian Ridge, ~ 15 mm yr^{-1}
- Southeast Indian Ridge, ~ 70 mm yr^{-1}
- Mohns Ridges, ~ 15–20 mm yr^{-1}
- Reykjanes Ridge, ~ 20 mm yr^{-1}
- Juan de Fuca and Gorda Ridges (inset), ~ 60 mm yr^{-1}

Earthquake data from http://earthquake.usgs.gov/eqcenter.

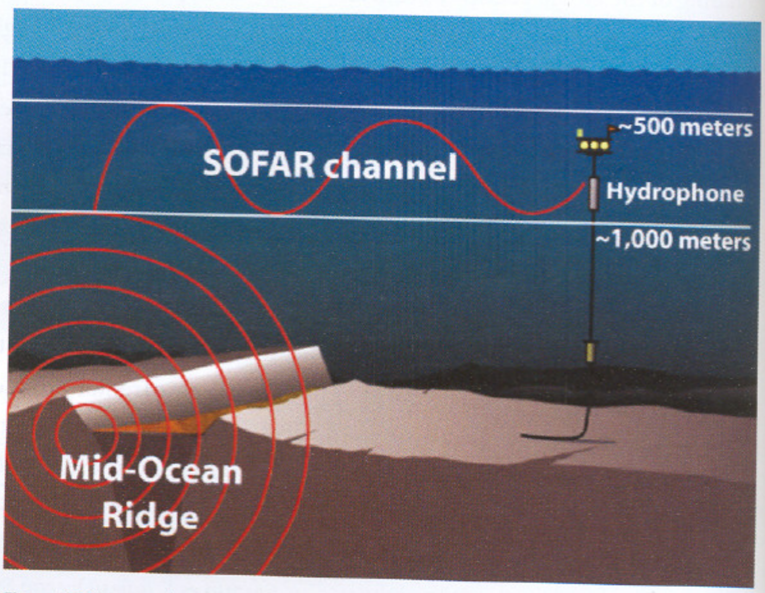

Figure 2. Diagram showing how a hydrophone deployed in the ocean sound channel is used to record acoustic waves from seismic and volcanic activity originating at a mid-ocean ridge. Red circles represent hydroacoustic wave fronts produced by seafloor seismic events that eventually become trapped and laterally propagate along the sound channel. A float keeps the hydrophone moored vertically in the water column. The entire mooring package can be retrieved via acoustic release near the seafloor.

HISTORICAL HYDROACOUSTIC MONITORING (PRE-1990)

The Discovery of *T*-phases

The first published report of a teleseismic *T*-phase wavetrain is attributed to Jagger (1930), who described a high-frequency arrival recorded on Hawai'i Volcano Observatory seismometers within the coda of a large 1927 Alaskan earthquake (Okal, 2008). Several years later, Collins (1936) noted that the seismogram of a Caribbean earthquake featured a third arrival following the primary (P) and secondary (S) phases recorded on short-period channels. Linehan (1940) first coined the term "*T*-phase" when he identified this third or "tertiary" arriving phase in the West Indian region, although he could only speculate as to its source.

The effort to develop antisubmarine warfare techniques during World War II brought considerable progress in hydroacoustics. Ewing et al. (1946) were the first to suggest that underwater sounds recorded during an ocean acoustic experiment were generated by submarine volcanic activity, and they proposed that a network of sound channel hydrophones could be used to monitor seafloor volcanism. Ewing and Worzel (1948) provided the theoretical basis for this phenomenon by describing the basics of long-range propagation in the ocean sound channel. Two years later, Tolstoy and Ewing (1950) provided the first identification of *T*-phases as the water-borne phases of an earthquake source resulting from the conversion of seismic energy to acoustic energy at the seafloor-ocean interface. The first detection of *T*-phases from MOR earthquakes was reported by Bath (1954), who used a land-based seismic network to detect earthquakes from the Mohns and Knipovich Ridges, establishing the existence of long-range acoustic propagation in the Arctic, despite the sound channel being surface limited there.

The Hawai'i *T*-phase Group (1960s to Early 1970s)

The study of *T*-phases and the effectiveness of hydroacoustic methods for volcano monitoring were advanced in the 1960s by use of a Pacific-wide network of sound channel hydrophones, the Air Force's Missile Impact Location System (MILS; Figure 3), whose data were analyzed by researchers at the Hawai'i Institute of Geophysics. Johnson et al. (1963) introduced the concept of down-continental-slope conversion for the generation of *T*-phases by subduction earthquakes. A potential generation mechanism for mid-ocean ridge and abyssal *T*-phases was later introduced by Johnson et al. (1968), who proposed the coupling of acoustic energy into the sound channel via sea surface scattering directly above the earthquake epicenter.

Northrop et al. (1968) and Northrop (1970) presented analysis of MILS hydrophone records of *T*-phases from the Gorda Ridge, a spreading center in the Northeast Pacific Ocean. These studies focused on the discrepancy between seismic and *T*-phase locations of ridge earthquakes (with *T*-phase locations being more accurate). Hammond and Walker (1991) used MILS data to show that the Juan de Fuca Ridge produced 58 *T*-phase events during a two-year period in the mid-1960s, all of which were located either at areas of vigorous hydrothermal venting or volcanic edifices near the spreading axis. Moreover, Walker and Hammond

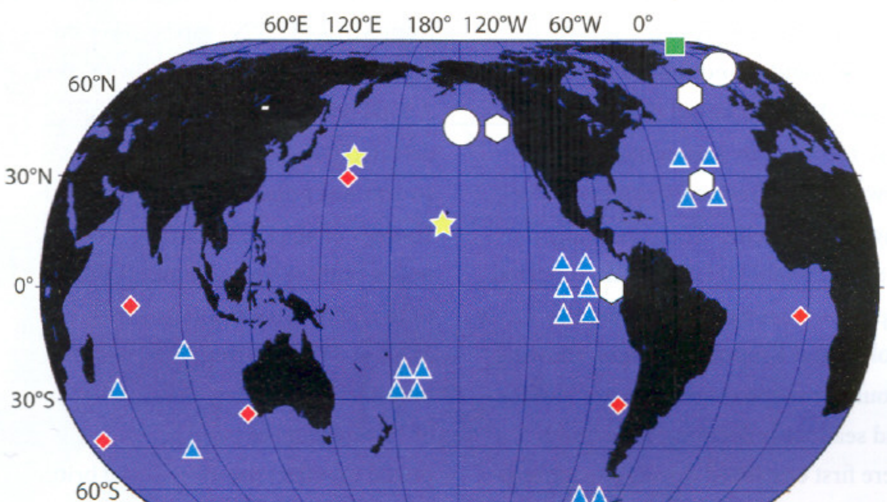

Figure 3. Map of major hydroacoustic arrays and systems discussed in text. Legend at right shows the system names associated with each icon. SOSUS = Sound Surveillance System. MILS = Missile Impact Location System.

(1998) showed that of the 644 *T*-phase Gorda Ridge events detected, nearly all occurred in discreet swarms centered on the ridge axis, and the swarms remained confined to individual ridge segments.

or slip on graben faults. Sonobuoy experiments also were performed along the Mid-Atlantic Ridge (MAR) at 36.5°–37°N (Reid and Macdonald, 1973; Spindel et al., 1974), the Galápagos Rift (MacDonald and Mudie,1974), and the Nansen Ridge in the Arctic Basin (Kristoffersen et al., 1982).

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Global Mid-Ocean Ridge Hydroacoustic Observations (1960s to Early 1990s)

There was increasing awareness that ocean *T*-phases were being recorded at coastal and island seismic stations. Cooke (1967) reported that *T*-phases were recorded by seismometers in New Zealand from earthquakes occurring on the Pacific-Antarctic Ridge. Bath and Shahidi (1971) again reported *T*-phases from earthquakes on the Mohns and Knipovich Ridges, showing acoustic rays propagated by sea surface and seafloor reflections. Sonobuoys were first employed during this time as an inexpensive method to detect mid-ocean ridge seismicity at the sea surface. Reid et al. (1973) used sonobuoys to detect multiple large earthquake swarms along the Guaymas basin rift in the Gulf of California. They attributed these swarms to either magma movement or slip on graben faults.

Brocher (1983) recorded *T*-phases from MAR (31.6°N) earthquakes on seafloor seismometers off Nova Scotia. A total of 16 *T*-phase events were detected over 30 hours, indicating the swarm was possibly of volcanic origin. Toomey et al. (1985) used an array of ocean-bottom hydrophones (OBHs) to record microearthquakes from the MAR at 23°N. During three weeks of monitoring, they detected an average of 15 earthquakes per day, interpreted to reflect the extensional brittle failure of the crust.

In May 1989, an earthquake swarm of 45 teleseismic events (M= 4–5.5) was detected from the Reykjanes Ridge, 500 km southwest of Iceland. The swarm was interpreted as a seafloor eruption (Crane et al., 1997), and an array of sonobuoys was air deployed to record acoustic arrivals and refine earthquake locations. These data showed the earthquakes were centered on a volcanic complex on the Reykjanes rift axis. Subsequent side-scan and submersible surveys showed the earthquake source region was an area of high acoustic backscatter and relatively recent lava flows.

A hydrophone array suspended through the Arctic sea ice was used to record the *T*-phases generated by 32 mid-Arctic Ridge earthquakes (Keenan and Dyer, 1984). The sea surface ice scattered the earthquake acoustic rays, enabling long range *T*-phase propagation despite the absence of a sound channel in the polar ocean. Keenan and Merriam (1991) provided further spectral analysis of Arctic *T*-phases and their long-range propagation characteristics, as well as giving estimates of their acoustic magnitudes. The earthquake locations published in these studies indicate the events were near the Spitzbergen Fracture Zone (transform fault) and not the Gakkel Ridge, which was not discernible from the bathymetry available at the time.

RECENT PAST AND PRESENT DAY HYDROACOUSTIC MONITORING (1990–2010)

The US Navy Sound Surveillance System

Following the end of the Cold War, the US Navy sought environmental applications for many military assets and developed a dual-use program to share technology with civilian researchers. Two projects were initiated to use the Navy's networks of bottom-mounted hydrophones, referred to as the Sound Surveillance System (SOSUS), one in the Atlantic (Nishimura and Conlon, 1994) and one in the Northeast Pacific (Fox and Hammond, 1994; Figure 3). The Atlantic effort focused on cetacean research; however, one notable exception was the use of northern Atlantic SOSUS arrays to detect a submarine eruption on the Mohns Ridge (Blackman et al., 2000). *T*-phase earthquake swarms concentrated at various locations along a 50–70 km length of the ridge, which Blackman et al. (2000) interpreted as representing focused eruptions at a few discrete areas rather than the intrusion and propagation of a magma dike.

In 1991, the National Oceanic and Atmospheric Administration (NOAA) Pacific Marine Environmental Laboratory's *T*-phase monitoring project began the first systematic effort to use hydrophone data to produce a continuous catalog of mid-ocean ridge seismicity by screening SOSUS data for earthquakes from the Juan de Fuca and Gorda Ridges. Availability of SOSUS to the civilian research community enabled real-time detection of *T*-phases from seafloor earthquakes, reducing the detection threshold of seismicity by almost two orders of magnitude (Fox et al., 1994). The excellent array geometry of the SOSUS network relative to Northeast Pacific spreading centers, combined with the existence of a well-defined velocity model of the ocean, yielded significant improvements in the accuracy of derived earthquake locations (Slack et al., 1999).

This much enhanced earthquake detection capability led to the first real-time observations of a seafloor spreading event on the Juan de Fuca Ridge (Fox et al., 1995). Nearly 700 earthquakes were tracked during a 23-day period following the intrusion of a magma dike 60 km along the rift axis of the CoAxial ridge segment (Dziak et al., 1995; Figure 4). The remote detection of this volcanic episode resulted in several in situ multidisciplinary studies that led to the discovery of a hydrothermal plume, a still cooling lava flow,

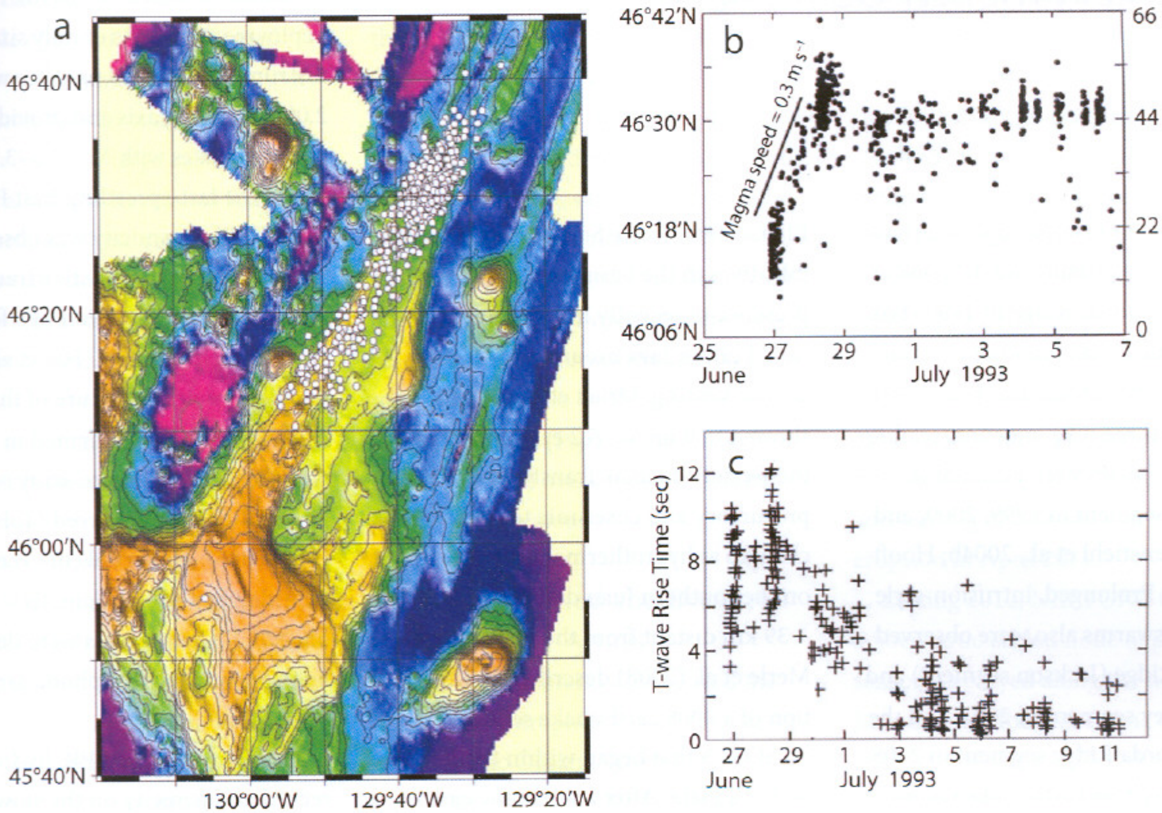

Figure 4. (a) Bathymetric map of the CoAxial segment of the Juan de Fuca Ridge. Circles show earthquake locations recorded during a 23-day period. (b) Along-segment position of earthquake locations vs. time during the swarm. (c) *T*-wave rise time vs. time during the swarm. Events show a decrease in rise time consistent with shoaling of earthquakes as the magma approaches the seafloor, erupting fresh lava along the northern part of the segment. *Reproduced from Dziak et al. (1995)*

and microbial communities living in a subseafloor ecosystem (e.g., Baker et al., 1995; Embley et al., 1995; Holden et al., 1998). Estimates of relative earthquake depths were also obtained using the *T*-phase rise time, defined as the time between the onset of the signal and its amplitude peak (Schreiner et al., 1995).

amounts of hydrothermal fluid during MOR spreading events.

The SOSUS earthquake locations also have been used in several studies of Northeast Pacific transform faults. Dziak et al. (1996) presented evidence of earthquake and volcanic-tremor activity from an extensional basin within the western

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Blanco Transform. Subsequent submersible dives in the basin confirmed the presence of recently formed constructional pillow lava mounds as well as diffuse venting. Dziak et al. (2003) showed that an M_w 6.2 earthquake on the western Blanco Transform caused precursory and coseismic temperature changes at hydrothermal vent sites on the southern Juan de Fuca Ridge, ~ 39 km distant from the main shock. Merle et al. (2008) described the detection of a 2008 earthquake sequence on SOSUS that began within the Juan de Fuca Plate. After a week, the earthquakes moved to the eastern Blanco Transform and ended 30 days later with a seafloor-spreading event on the northern Gorda Ridge.

Two other major eruptive episodes have been documented using SOSUS, one at the northern Gorda Ridge in 1996 (Fox and Dziak, 1998) and another at Axial Volcano in 1998 (Dziak and Fox, 1999). In addition, three dike injection/seafloor spreading episodes were detected at Endeavour Segment in 1999, 2000, and 2005 (Bohnenstiehl et al., 2004b; Hooft et al., 2010). Prolonged, intrusion-style earthquake swarms also were observed at the Gorda Ridge (Jackson segment) and Middle Valley segment in 2001, and the northern Gorda Ridge segment in 2008. Using this *T*-phase earthquake record, Dziak et al. (2007) were able to infer that a rapid intrusion of magma within a rift zone typically leads to seafloor eruptions and expulsion of massive

Autonomous Portable Hydrophones

Early successes using SOSUS facilitated the development of moored autonomous underwater hydrophone systems (AUHs) that could be used to monitor global ridge segments. In this design, the hydrophone sensor and instrument package are suspended within the SOFAR channel using a seafloor tether and foam flotation (Figure 2). These instruments have been deployed successfully along mid-ocean ridge spreading centers in the Atlantic (Smith et al., 2002, 2003; Goslin et al., 2005; Simao, 2010), eastern Pacific (Fox et al., 2001), and Indian (Royer et al., 2009) Oceans, as well as along the back-arc spreading systems of the western Pacific (Dziak et al., 2005) and Antarctic Peninsula (Dziak et al., 2010). A typical deployment consists of only six to seven instruments that can monitor more than 2,000 km along axis and provide a record of earthquakes with M > 2.5–3.

On the fast-spreading East Pacific Rise (EPR), seismicity was observed to concentrate along active transform faults, with the exception of a few swarm sites on the ridge axis (Fox et al., 2001). The likely volcanic nature of these EPR swarms was confirmed in 2006 when an in situ seismic array observed an eruption as it occurred (Tolstoy et al., 2006). Thousands of events were recorded by the seismometers, whereas only 23 *T*-phase events were detected by the regional hydrophone array (Dziak et al., 2009).

In contrast to the EPR, hydrophone-recorded seismicity on the slow-spreading MAR is common along both the rift axis and its transform offsets (Smith et al., 2002). Smith et al. (2003) showed ridge-crest seismicity correlates