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Volcanic Eruptions in the Deep Sea

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Eruption of molten lapilli, ash,
and sulfur-rich fumes at Hades
vent, West Mata Volcano.

ABSTRACT. Volcanic eruptions are important events in Earth's cycle of magma generation and crustal construction. Over durations of hours to years, eruptions produce new deposits of lava and/or fragmentary ejecta, transfer heat and magmatic volatiles from Earth's interior to the overlying air or seawater, and significantly modify the landscape and perturb local ecosystems. Today and through most of geological history, the greatest number and volume of volcanic eruptions on Earth have occurred in the deep ocean along mid-ocean ridges, near subduction zones, on oceanic plateaus, and on thousands of mid-plate seamounts. However, deep-sea eruptions (> 500 m depth) are much more difficult to detect and observe than subaerial eruptions, so comparatively little is known about them. Great strides have been made in eruption detection, response speed, and observational detail since the first recognition of a deep submarine eruption at a mid-ocean ridge 25 years ago. Studies of ongoing or recent deep submarine eruptions reveal information about their sizes, durations, frequencies, styles, and environmental impacts. Ultimately, magma formation and accumulation in the upper mantle and crust, plus local tectonic stress fields, dictate when, where, and how often submarine eruptions occur, whereas eruption depth, magma composition, conditions of volatile segregation, and tectonic setting determine submarine eruption style.

AN ABUNDANCE OF SUBMARINE VOLCANOES

There are thousands of submarine mountains throughout the deep ocean, and nearly all of them are volcanoes (e.g., Smith and Jordan, 1987; Smith and Cann, 1992). They form long, continuous ridges along seafloor spreading centers, arcuate chains along subduction zones, and both linear chains and widely distributed individual seamounts away from plate boundaries. Submarine volcanism is one of the most important geological processes on the planet, forming a dense, low-lying, igneous crust that floors the vast ocean basins and covers nearly two-thirds of Earth's surface. Studies of active volcanism around the planet provide information that is critical to understanding the range of volcanic conditions in different settings, but they are heavily weighted to the subaerial environment (Figure 1). Deep-sea eruptions are much more difficult to detect and observe than subaerial eruptions, but knowledge about them is increasing rapidly.

This paper provides an overview of active deep-sea volcanism, here taken to mean at water depths greater than approximately 500 m (Mastin and Witter, 2000). Representative case studies describe the general character of eruptions and volcanic histories at different geological settings. Although our focus is on mid-ocean ridge (MOR)

eruptions, we also discuss submarine eruptions in other settings because volcanism is a continuum of conditions and processes across the full range of tectonic settings and eruption depths, and because our sample of MOR eruptions is small and, as yet, lacking direct observation of an ongoing eruption. Photographs of a variety of deep-sea eruption styles and deposits (Figure 2) depict the two general forms that volcanic deposits take: lava flows and pyroclasts (volcanic fragments).

Seafloor eruptions punctuate the tranquility of the deep sea with injections of heat, lava, pyroclasts, and gases into the overlying water, affecting its composition, density structure, and ecology (e.g., Delaney et al., 1998; Kelley et al., 2002). The volcanic and tectonic characteristics of MORs vary with spreading rate and magma supply (e.g., Macdonald et al., 1992; Perfit and Chadwick, 1998; Small, 1998). Along fast-spreading ridges, small (50–1,000 m tall) volcanoes merge into a semicontinuous ridge, producing lava fields several kilometers to either side. Slow-spreading ridges are

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Historically active volcanoes (< 500 y)

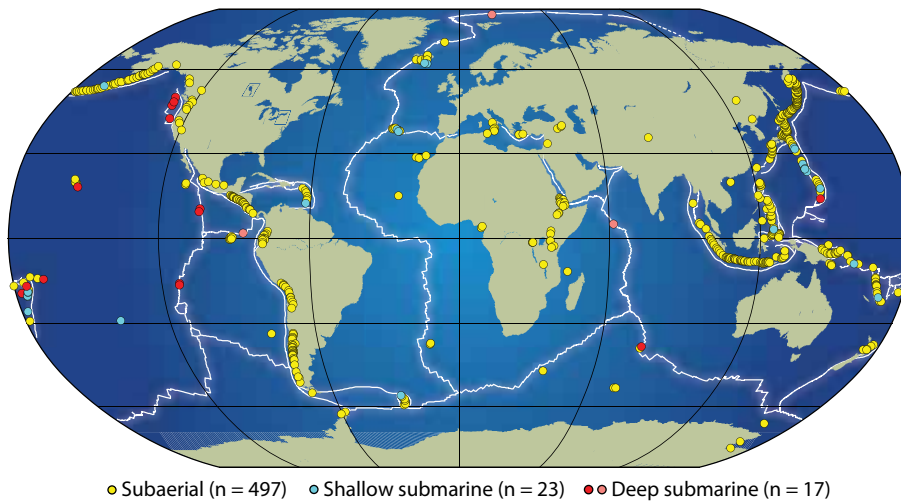


Figure 1. Sites of historical volcanism on land and in the sea over the past 500 years, based on data from the Smithsonian Institution Global Volcanism project and references cited in this paper. Shallow and deep submarine volcanoes are distinguished because the style of volcanism differs greatly at the two. Note the much higher number of sites on land. Only confirmed (red dot) and likely (pink dot) historical deep submarine eruptions are shown. Likely but unconfirmed eruptions are defined as those lacking a dated eruption signal on the seabed and in the water column (i.e., having a seismicity signal and an event plume but no young lava, or having seismicity and visually young lavas but no event plume or high-resolution lava dating).

marked by a 5–15 km wide rift valley in which volcanism occurs on elongate volcanic ridges (Searle et al., 2010). At MORs, tectonic processes modify the zone of active volcanism, making it difficult to construct large volcanic edifices. In contrast, much of Hawaii’s 10,000 m tall Mauna Loa (Earth’s largest active volcano) was built by many thousands of submarine eruptions over about a million years at roughly the same location in the middle of the Pacific Plate. Magmas erupt at subduction zone volcanoes with a distinct style that is largely due to recycled oceanic lithosphere (rocks, sediments, and fluids) in their mantle sources, producing magmas that are compositionally distinct from those at MOR and mid-plate volcanoes (in particular, they are more volatile rich).

Crustal magma bodies that feed submarine volcanic eruptions are the heat sources that drive the majority of hydrothermal activity in the

ocean (e.g., Wolery and Sleep, 1976). Hydrothermal processes create both focused and diffuse fluid flow into the deep ocean, forming edifices of chemical precipitates (e.g., black smoker chimneys) and creating habitats for chemosynthetic animal communities (e.g., Tivey, 1995; Shank et al., 1998; Luther et al., 2001). Chemosynthetic ecosystems are sensitive to hydrothermal fluid compositions, which change with time and distance from the magmatic heat source (Butterfield et al., 1997; Von Damm, 2000, 2004). Volcanic eruptions may directly disrupt these communities by repaving the seafloor with lava or indirectly by reorganizing hydrothermal plumbing. Eruption impacts on local hydrothermal and benthic ecology are greatest for sessile fauna that cannot move from one location to another (Shank et al., 1998) and for animals with limited larval dispersal mechanisms (Kim and Mullineaux, 1998). However,

some organisms flourish even when volcanic events make the environment unstable (Tunnicliffe et al., 2003; Fornari et al., 2004; Embley et al., 2006).

Our understanding of deep submarine volcanism has been greatly aided by mapping and sampling the types and distributions of volcanic products from recent but unknown age eruptions. In particular, seafloor studies at various MOR sites have provided information about eruption styles, sizes, and frequencies over a broad range of dominantly effusive volcanoes (e.g., Bryan and Moore, 1977; Ballard et al., 1979, 1982; Embley and Chadwick, 1994; Sinton et al., 2002, 2010; White et al., 2002; Fornari et al., 2004; Soule et al., 2005, 2009; Stakes et al., 2006; Escartín et al., 2007).

Complementary studies of historical deep-sea eruption deposits are key to understanding short-term evolution of seafloor volcanic landscapes and rapidly changing thermal and subsurface hydrologic conditions in the uppermost crust (e.g., Embley and Chadwick, 1994; Gregg et al., 1996; Embley et al., 2000, 2006; Perfit and Chadwick, 1998; Soule et al., 2007). Ideally, these features are studied before they are modified by post-eruption faulting, volcanic collapse, and background sedimentation, which obscure age contrasts in the landscape. These analyses, along with the need to study temporal aspects of hydrothermal effluent chemistry, the style and location of effluent discharge, and the concomitant changes in marine ecosystems, make submarine eruption detection and rapid response (e.g., see Baker et al., 2012, in this issue), as well as the development of eruption chronologies (e.g., Rubin et al., 1994), integral components of deep submarine volcanology.

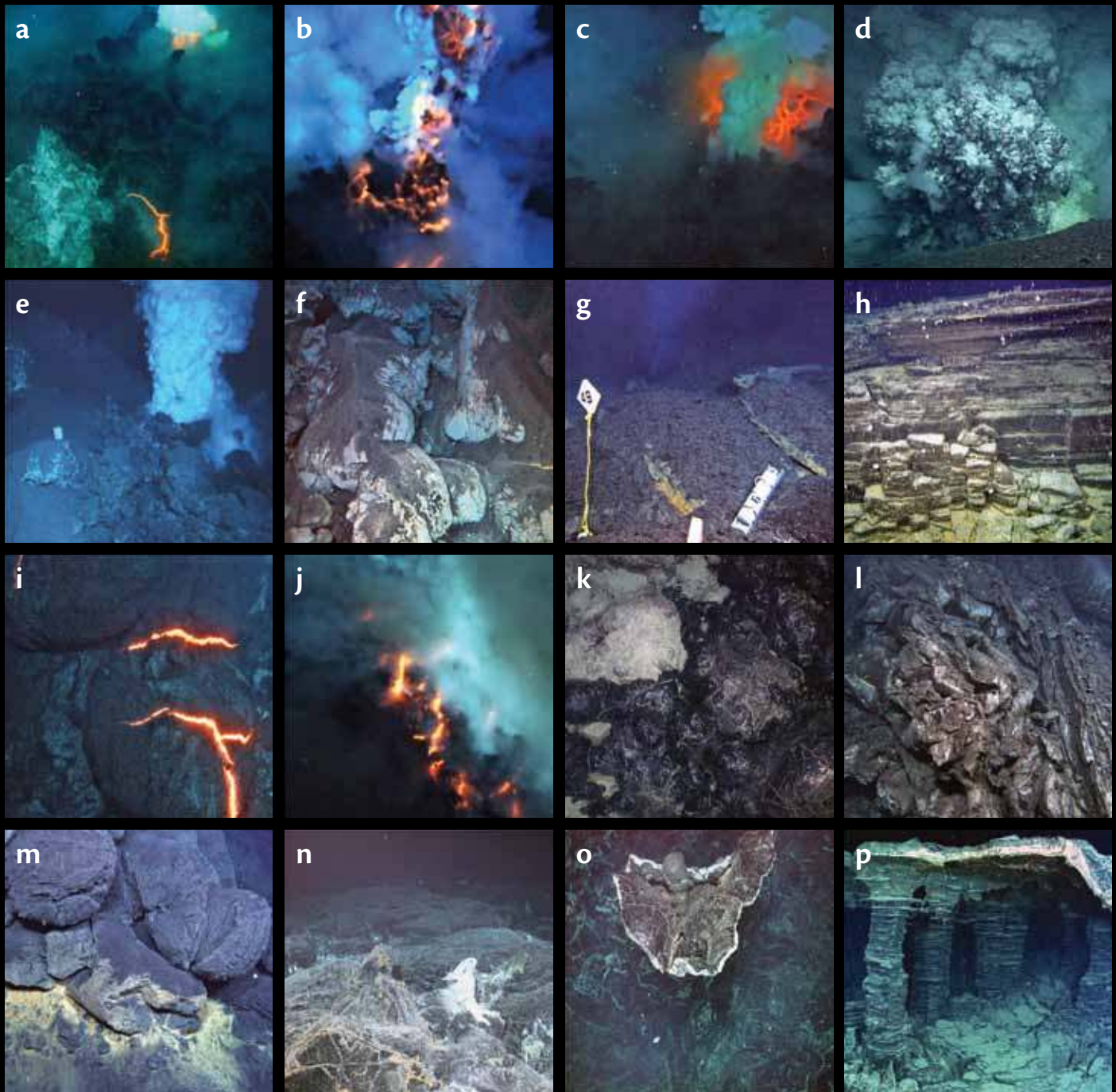


Figure 2. Images of deep-sea eruption styles and products. Panels a–h depict various forms of explosive activity, and panels i–p depict effusive activity. (a) Explosive and effusive activity at Hades vent, West Mata Volcano. (b) Several-meter-high fountain of molten magma, solidifying ejecta, and gas at Hades vent, West Mata Volcano. (c) Strombolian eruption of a meter-wide lava-skinned gas bubble at West Mata Volcano. (d) Ejection of volcanic ash and sulfur at Brimstone vent at Northwest Rota-1 Volcano. (e) Overview of Brimstone vent eruption area (22 cm wide tephra collection bucket for scale). (f) Young volcanic pyroclasts overlying pillow lavas from an eruption at Gakkel Ridge. (g) Coarse-grained pyroclasts cover lavas about 50 m from Hades vent, West Mata Volcano (30 cm long hydrophone for scale). (h) A 1 m high section of consolidated pyroclastic deposit at one of the Vance Seamounts, near the Juan de Fuca Ridge. (i) Active pillow lava lobe inflating on transverse and radial cracks, near Hades Vent, West Mata Volcano. (j) High-effusion-rate sheet flow moving down a steep slope near Hades Vent, West Mata Volcano. (k) Downward-looking view of 2005–2006 pillow lavas overlying older pillows at the East Pacific Rise, 9°50'N. (l) High-effusion-rate sheet flow of the Puipui eruption, Northeast Lau Spreading Center. (m) Months-old pillow lavas at West Mata Volcano colonized by an orange microbial mat. (n) Months-old lobate lavas at the East Pacific Rise, 10°45'N, covered by various microbial mats. (o) Downward-looking view of a remnant piece of lobate crust above the collapsed interior of 2005–2006 lavas at the East Pacific Rise, 9°50'N. (p) Side-on view of 2–3 m tall lava pillars in the interior of the partially collapsed lobate lava flow erupted at Axial Seamount, Juan de Fuca Ridge, in 1998. *Photo sources:* (a–c) Jason dive J2-420, 2009; (d) Jason dive J2-189, 2006; (e) Jason dive J2-408, 2009; (f) *Camper camera sled*, 2007; (g,m) Jason dive J2-418, 2009; (h) *Tiburion dive T1011*, 2006; (i,j) Jason dive J2-414, 2009; (k) *TowCam*, 2006; (l) Jason dive J2-415, 2009; (n) *Alvin dive 3935*, 2003; (o) *TowCam*, 2006; (p) *ROPOS dive R743*, 2003

SIGNIFICANT DEEP SUBMARINE ERUPTION DISCOVERIES

Marine geologists have known that there are volcanic rocks at the seafloor since at least the 1870s, when submarine basalts were recovered from the deep sea by trans-Atlantic cable repair ships (Hall, 1876) and the *Challenger* Expedition (Murray and Renard, 1891). Early MOR photographic surveys and submersible expeditions in the 1960s and 1970s revealed a wide variety of volcanic landforms—some similar to

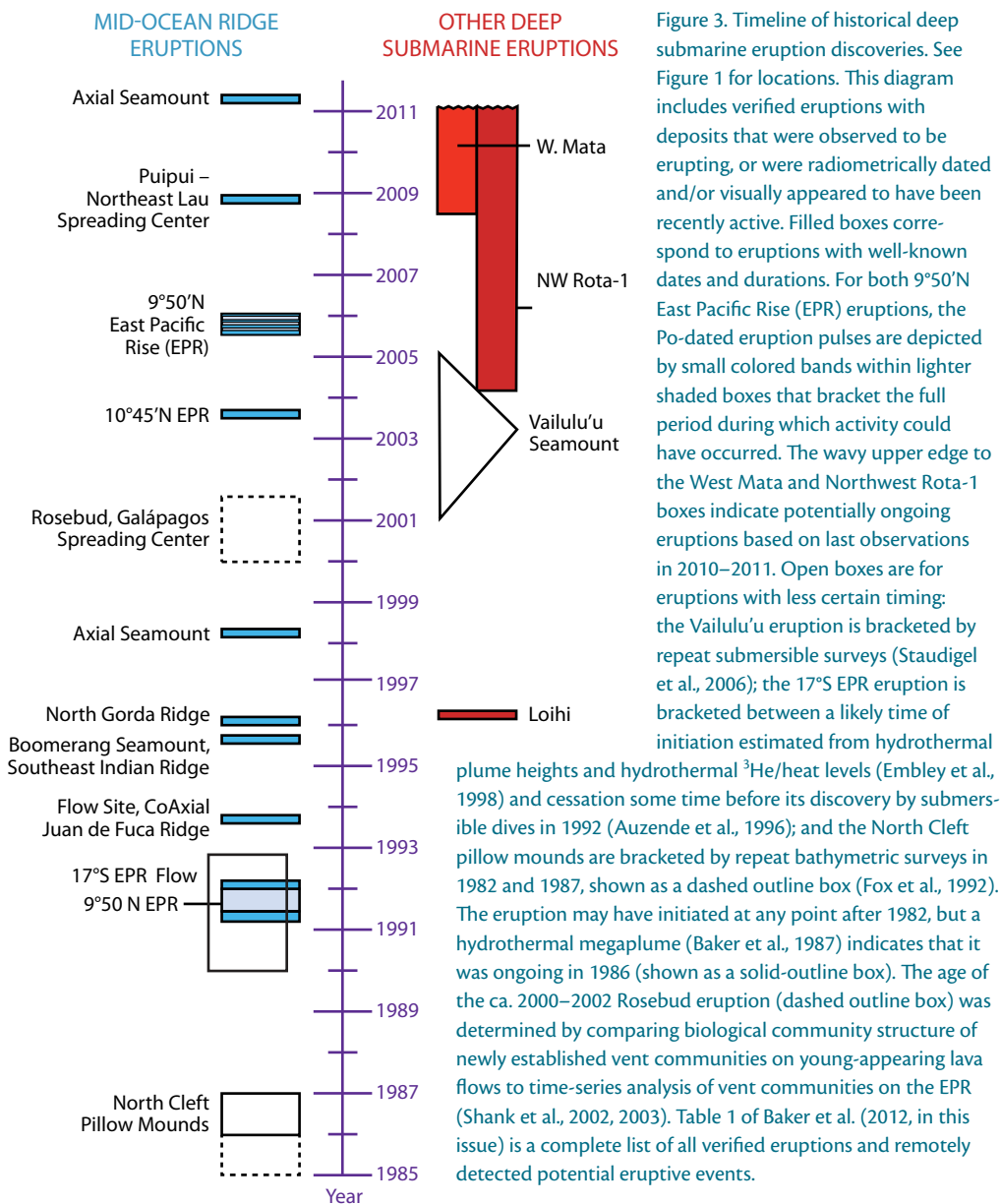
those in subaerial basaltic provinces like Hawaii and Iceland (e.g., Ballard and Van Andel, 1977). The discovery of deep-sea warm springs on the Galápagos Spreading Center (GSC) in 1977 (Corliss et al., 1979) and hot springs on the East Pacific Rise (EPR) in 1979 (Spiess et al., 1980) provided evidence that these MOR volcanoes were underlain by active magmatic systems.

Soon thereafter, increased access to seafloor along MORs and improved observational technologies paved the

way for discoveries of specific eruptions (summarized in Figure 3). A hydrothermal event plume detected along the Cleft segment of the Juan de Fuca Ridge (JdFR) in 1986 (Baker et al., 1987) led to the discovery of the first young deep-sea lava flows of known age: a series of pillow mounds up to 75 m high (Chadwick et al., 1991; Embley et al., 1991; Chadwick and Embley, 1994). Divers in the *Alvin* submersible at 9°50'N EPR made serendipitous observations of the aftermath of a March 1991 deep volcanic

eruption, finding newly dead and charred tubeworms strewn among and under fresh lava along with diffuse hydrothermal vents (“snowblowers”) spewing vast amounts of white, sulfur-rich microbial floc (Haymon et al., 1993). Radiometric dating of lavas demonstrated that the divers had arrived just two to four weeks after the eruption (Rubin et al., 1994). The seafloor volcanic terrain was not jet-black fresh lava as expected, but instead was draped (e.g., Figure 2n) with white/gray gelatinous microbial matter (Haymon et al., 1993) that was largely absent a year later (Shank et al., 1998). Additional visual evidence and radiometric dating indicated that the EPR eruption continued with a second pulse in late 1991 to early 1992 (Rubin et al., 1994). A similar “snowblower” vent and young lavas were discovered at 17°S on the EPR in early 1993 (Auzende et al., 1996; Embley et al., 1998; Sinton et al., 2002).

In 1993, scientists at the National Oceanic and



Atmospheric Administration Pacific Marine Environmental Laboratory (NOAA PMEL) detected an earthquake swarm typical of volcanic eruptions on land on the JdFR CoAxial segment within the first month of real-time acoustic monitoring using the US Navy Sound Surveillance System (SOSUS; Fox et al., 1994). This detection triggered the first dedicated “response” expedition, which observed event plumes, new hydrothermal vents, a still-cooling lava flow (Fox et al., 1995; Embley et al., 1995), and, remarkably, two other previously unknown nearby young lava flows (Embley et al., 2000). This real-time Northeast Pacific hydroacoustic monitoring capability has led to multiple eruption detections and responses over nearly two decades (Perfit and Chadwick, 1998; Dziak et al., 2006, 2011; Baker et al., 2012, in this issue).

In 1996, earthquake swarms were detected on the North Gorda Ridge by SOSUS (Fox and Dziak, 1998) and at the 1,100 m deep summit of Loihi Seamount by US Geological Survey land-based seismometers (The 1996 Loihi Science Team, 1997). Both events were associated with volcanic eruptions (Chadwick et al., 1998; Rubin et al., 1998; Garcia et al., 1998; Clague et al., 2000) and the exhalation of large volumes of hydrothermal and magmatic fluids into the overlying ocean. The Loihi eruption was the first of these new discoveries away from the global MOR. A summit caldera eruption at Axial Seamount on the JdFR in 1998 (Dziak and Fox, 1999; Embley et al., 1999) partially buried a pressure and temperature instrument package at ~ 1,500 m water depth, leading to the first in situ record of the emplacement of a submarine lava flow, which included lava flow inflation and

deflation (Fox et al., 2001; Chadwick, 2003), like many eruptions on land (Hon et al., 1994). In 2005, submersible divers studying the summit caldera of Vailulu'u Seamount, east of American Samoa, observed hydrothermal fluids venting from a new, 300 m high volcanic cone not present on the caldera floor in 2001 (Staudigel et al., 2006).

In 2006, several still-operating ocean bottom seismometers (OBSs) would not release from the seafloor at 9°50'N EPR, and two “response” cruises discovered that newly erupted lavas had trapped them (Tolstoy et al., 2006; Cowen et al., 2007), marking the first time a second eruption had been observed at the same MOR location. It also marked the beginning of very detailed surveys of newly formed deep-sea lava flows using high-resolution bathymetric, side-scan, and photographic data from before and after an eruption (Soule et al., 2007; Fundis et al., 2010) and large-scale radiometric dating of new lavas (Rubin et al., 2008). The eruption occurred in 2005 to 2006, with lava flows extending 19 km along the ridge axis and up to 3 km away from it into much older terrain, with a much greater areal coverage and a volume of roughly five times that of the 1991–1992 eruption.

Another repeat eruption was observed in July 2011 at Axial Seamount in the same area as the 1998 eruption (Caress et al., 2011; Chadwick et al., 2011). The well-instrumented eruption site has provided a wealth of in situ geodetic, seismic, and temperature data, as well as visual observations and samples from three previously scheduled remotely operated vehicle (ROV) and/or autonomous underwater vehicle (AUV) expeditions. The April 2011 eruption was not recognized in real-time SOSUS data due

to recent reductions in system coverage.

The new millennium also brought new seafloor eruption discoveries at deep subduction zone and back-arc settings, beginning in 2004 with ROV observations of pyroclastic eruptions at the 520 m deep summit of Northwest Rota-1 Volcano in the Marianas Arc (Embley et al., 2006; Chadwick et al., 2008; Dearthorff et al., 2011). The volcano appears to have been continuously active from at least February 2003 to March 2010 (Chadwick et al., 2012). In November 2008, two active eruptions were serendipitously detected in the Northeast Lau Basin in the Southwest Pacific by water-column anomalies; they were subsequently verified by May 2009 ROV dives. The mainly effusive “Puipui” eruption on the Northeast Lau Spreading Center (NELSC; e.g., Rubin et al., 2009) produced an event plume containing volcanic fragments (Baker et al., 2011). At West Mata, a nearby back-arc volcano, a second eruption had an intense particle plume and unusually high H₂ concentrations. ROV dives observed dramatic pyroclastic and active effusive pillow eruptions from two ~ 1,200 m deep West Mata summit vents (Resing et al., 2011; Clague et al., 2011). The West Mata discovery allowed for detailed, real-time observations of a deep submarine eruption, including the first imagery of glowing hot lava flowing across the deep seafloor (Figure 2i,j) and the first sample of molten lava from the deep sea (Figure 4). Dramatic video of pillow lavas flowing slowly across the seafloor and of the explosive release of magma and gas into the water column (Figure 2a–c) can be downloaded from the NOAA Vents Program website (<http://www.pmel.noaa.gov/vents/laubasin/laubasin-multimedia.html>).

Acoustic recordings on both seafloor and moored hydrophones provided a long-term record of fluctuations in the intensity of the West Mata eruption (Dziak et al., 2009, 2010) and correlation between eruption sounds and visual observations of different eruptive styles at the vents. High-resolution AUV multibeam mapping indicates that mixed effusive and pyroclastic eruptions have been common at West Mata throughout much of its eruptive history (Clague et al., 2011).

ERUPTION PRODUCTS

The physical character of effusive and explosive eruptions and the deposits they produce are controlled by magma chemistry and physical properties, magma eruption rate, and physical conditions at the volcanic vent. The dominant volcanic product from submarine eruptions is effusive lava flows from vents aligned linearly along an eruptive fissure (common at MORs) or from one or more localized vents (common at submarine arc and intraplate volcanoes). Lava-flow thickness, run out, and surface morphology reflect variations in magma viscosity, effusion rate, local slope, topographic obstructions, and the

sequence of emplacement events during individual eruptions (Gregg and Fink, 1995; Chadwick et al., 1999; Gregg and Smith, 2003; Soule et al., 2005, 2007; Escartín et al., 2007; Fundis et al., 2010). Submarine lava-flow morphologies are as varied as those on land and have been described from near-bottom observations in many locations (Bryan and Moore, 1977; Lonsdale, 1977; Ballard et al., 1979; Francheteau et al., 1979; Bonatti and Harrison, 1988; Fox et al., 1988; Fornari et al., 1998; Embley et al., 1990; Embley and Chadwick, 1994). Lava morphologies are typically classified by the length scale of the quenched-crust units that collectively make up a lava flow. Different types are pillow, lobate, and sheet lavas (Figure 2i–p), although they are more correctly thought of as a continuum controlled by flow and cooling rate.

Although often volumetrically minor at MORs, deposits of lapilli- to ash-sized fragments from explosive submarine activity occur throughout the depth range of observations in the deep sea (Clague et al., 2009). Shallow submarine eruptions, such as the one that formed Surtsey Island off the coast of Iceland in 1963, are primarily driven

by hydrovolcanic steam generation from the interaction of seawater with magma. In the deep ocean, dynamic exsolution of magmatic volatiles primarily drives explosivity, particularly if magmas ascend to shallow crustal levels from the mantle under largely closed-system conditions (Davis and Clague, 2006; Sohn et al., 2008; Clague et al., 2009; Schipper et al., 2010; Helo et al., 2011). Understanding of deep submarine pyroclastic activity (Figure 2a–h) has advanced largely based on observations of active processes at Northwest Rota-1 (e.g., Deardoff et al., 2011) and West Mata (Resing et al., 2011), as well as large pyroclastic deposits at other sites such as Loihi Seamount (Clague et al., 2009), Seamount 6 (Maicher and White, 2001), Gakkel Ridge (Sohn et al., 2008; Barreyre et al., 2011), and the Mid-Atlantic Ridge (Eissen et al., 2003).

The size, volume, thickness, and dispersal of both effusive and pyroclastic volcanic deposits provide key information about the conditions of the eruption(s) that produced them. Today, it is generally straightforward to determine the aerial extent of young submarine deposits by detailed near-bottom photo and multibeam mapping,



Figure 4. Images of the first molten lava sampled in the deep sea by the remotely operated vehicle *Jason* during the 2009 Northeast Lau event response cruise on R/V *TG Thompson*. The *Jason* team skillfully inserted a T-handled rod into a glowing portion of an advancing pillow lobe and pulled out a cantaloupe-sized piece of lava, subsequently shown to be a rare magma type known as boninite (Resing et al., 2011).

but currently there is no technology to directly measure the true thickness of submarine lava over the entire flow without high-resolution observations of the pre-eruption surface. Deposit thickness is traditionally estimated from field relationships (e.g., from flow fronts and collapse margins), but these estimates are difficult to make with just a few direct submersible and ROV observations. Thus, absolute volumes remain uncertain for many submarine eruption units; for example, the 17°S EPR Aldo-Kihi and 2005–2006 9°50'N

EPR eruptions have similar surface area, fissure length, and lava flow morphology yet differ by a factor of six in estimated volume (Sinton et al., 2002; Soule et al., 2007). Differential bathymetric maps of the pre- and post-eruption terrain can also provide volume estimates, but are limited by the resolution and navigational accuracy of both surveys (see Figure 5). For future eruptions, repeat near-bottom high-resolution surveys should provide high-precision thickness estimates at sites with existing high-resolution bathymetry. An additional

uncertainty in determining submarine pyroclastic-erupted magma volume is limited knowledge about internal deposit porosity and fragment vesicularity (i.e., the dense rock equivalent [DRE] volume commonly employed on land). One study suggests a DRE factor of ~ 50% for a deep submarine site on Loihi (Schipper et al., 2011).

ERUPTION DYNAMICS

An eruption may be described by a number of parameters including: eruptive style (e.g., effusive/explosive),

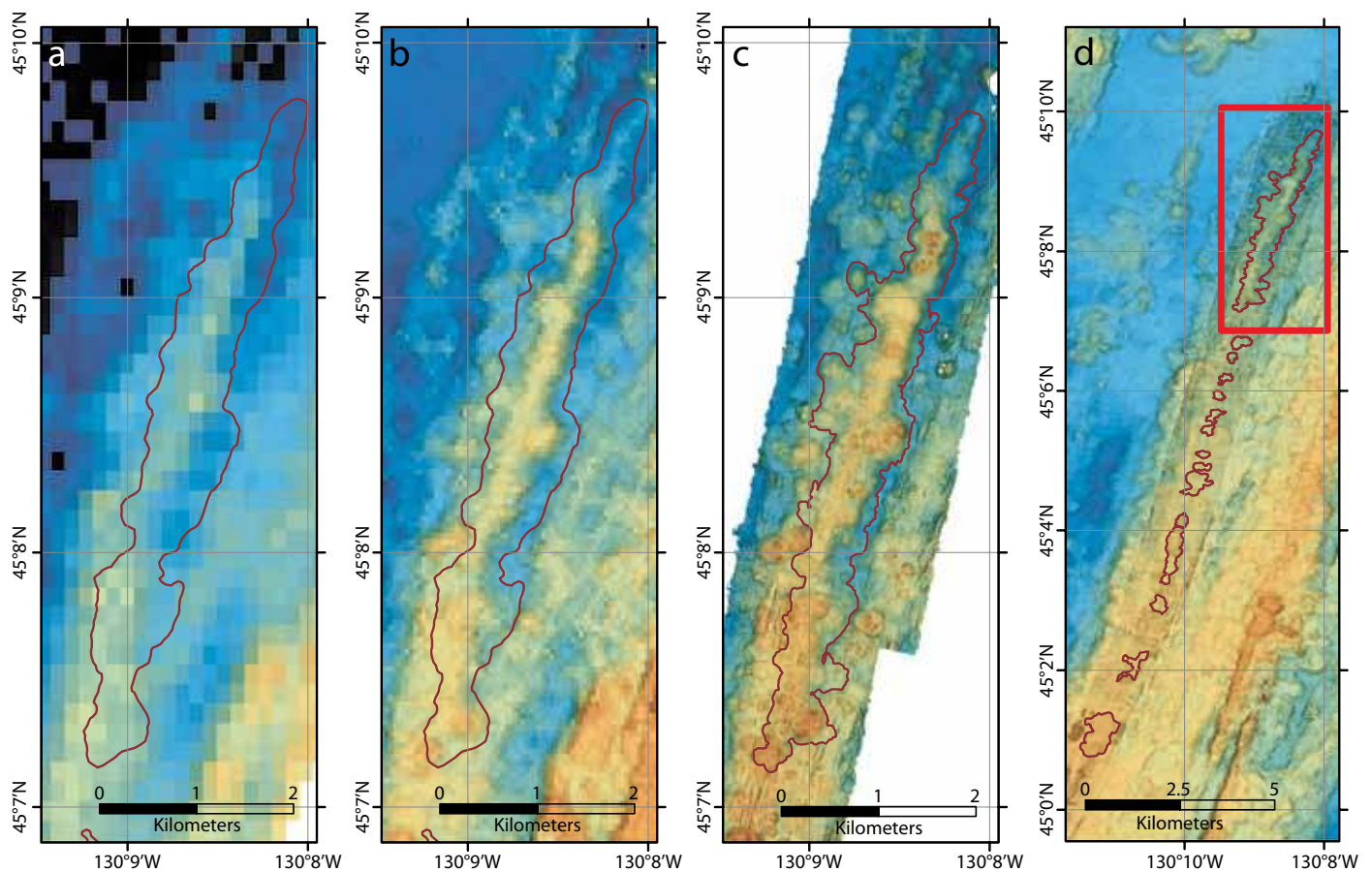


Figure 5. Improvements in resolution of bathymetric mapping and navigation shown in three generations of multibeam sonar with maps of the northern end of the 1986 North Cleft pillow mounds lava flow. (a) 1991 Seabeam classic (12 kHz), Loran-C navigated (Fox et al., 1992), shown at 100 m resolution. (b) 1998 Simrad EM300 (30 kHz), GPS navigated (MBARI Mapping Team, 2001), shown at 30 m resolution. (c) Recent Reson 7125 (200 kHz) multibeam, flown at 75 m altitude with the MBARI Mapping autonomous underwater vehicle (AUV), shown at 1 m resolution (Yeo et al., 2010). (d) Overview of the entire eruption area at the highest available resolution (same data set as in c). The red box shows the region in panels a, b, and c. Color scale ranges for a, b, and c are 2,420–2,260 m, and for d are 2,520–2,110 m depth. Two different outlines of the lava flow (brown) are also shown. In panels a and b, the outline of the northernmost mound (Chadwick et al., 1996) is based on AMS-60 side-scan and bathymetry data, plus camera tow observations; it is a refinement of the original flow outline (Fox et al., 1992; Chadwick and Embley, 1994). In panels c and d, the outline is based on the AUV data in the same panel and MBARI remotely operated vehicles *Tiburón* and *Doc Ricketts* samples and observations (Yeo et al., 2010).

intensity (e.g., mass eruption rate), duration, and volume. Although direct measurements are difficult, the small catalog of historical submarine eruptions (Figures 1 and 3) indicates that these factors vary significantly between tectonic settings. The geological, geochronological, and remote-sensing tools used to infer these parameters have evolved over the years and yet challenges remain in their application to submarine eruptions. The goals are to: (1) visually image the young eruption deposits, ideally in a time series; (2) obtain high-spatial-resolution bathymetry of the new terrain and its contacts with the neighboring seafloor; and (3) determine when various parts of the geological deposits were emplaced, in both relative (i.e., stratigraphic) and absolute (chronologic) frameworks. These observations provide the basis for estimating eruption deposit thickness, area, volume, and morphology as well as the location and extent of hydrothermal activity and biological colonization.

Deep-submergence tools used to achieve the first and second goals have improved dramatically over the years. Current high-frequency (12–30 kHz) surface ship, and near-bottom multi-beam and side-scan sonar systems are capable of tens of meters to submeter resolution, respectively, with ever-improving navigational precision, and a variety of deep-submergence platforms also have high-definition photography and video capabilities. A comparison of the > 50 m resolution repeat surface ship multibeam surveys used in the 1980s to more recent 1 m resolution AUV mapping demonstrates how far mapping capabilities have come in 30 years (Figure 5).

Radiometric dating of lavas in the

context of a detailed geologic framework based on deposit mapping primarily addresses the third goal. ^{210}Po – ^{210}Pb dating provides ultra-high-resolution eruption ages (± 1 – 3 weeks) of volcanic rocks, and has been used to determine emplacement chronologies of multiple submarine eruptions (see also Table 1 of Baker et al., 2012, in this issue). The method works by charting the post-eruption in-growth of initially degassed, volatile ^{210}Po from erupting magma to essentially not-volatile, grandparental ^{210}Pb in lava samples. The best age resolution is obtained when samples are recovered soon (relative to the 138.4 day half-life of ^{210}Po) after eruption. The earliest Po-dating studies used a handful of samples collected during event response or serendipitous eruption discovery cruises to determine if “zero-age” volcanic deposits were present on the seafloor and when they were erupted (e.g., Rubin et al., 1994, 1998; Garcia et al., 1998; Johnson et al., 2000).

There is added power when larger numbers of widely distributed samples are dated from a single eruption deposit. When combined with high-resolution geological and sonar mapping, such data permit reconstruction of the temporal and spatial evolution of a lava flow field to understand basic aspects of eruption dynamics. For example, detailed mapping and Po dating of 15 samples from the 2005–2006 lava flow at 9°50'N EPR demonstrate that this long fissure eruption began in summer 2005, with subsequent smaller eruptive pulses from ever-shortening fissures, culminating in a small eruption in mid to late January 2006 (Rubin et al., 2008, and recent work of author Rubin and colleagues). The Po-dated eruption pulses are correlated in time with seismicity pulses and

hydrothermal fluid exit temperatures recorded by in situ loggers in one chimney. In addition, Po dating of robust sample sets from the 2008–2009 eruption of West Mata Volcano (Resing et al., 2011) and the 2008 Puipui eruption on the NELSC (Baker et al., 2011) provide considerable detail regarding the dynamics of those eruptions (Rubin et al., 2009).

ERUPTION FREQUENCY AND DURATION

The tempo of activity at a volcano, both during and between eruptions, reflects the rate of magma input, the buildup of tectonic stress, and conditions within the crustal magma reservoir. Some volcanoes on land (and probably most MOR volcanoes) operate at a quasi-steady state, whereby eruption volume and repose (the time between eruptions) correlate positively and record a relatively constant magma flux to and from the volcano over significant periods of time (e.g., Wadge, 1982). We currently lack the information to determine these parameters at submarine volcanoes. Multiple researchers have used a steady-state assumption to estimate eruption frequency from magma-supply rate (calculated from spreading rate and assuming constant crustal thickness; Carbotte and Macdonald, 1992; Haymon et al., 1993; Macdonald, 1998; Perfit and Chadwick, 1998; Sinton et al., 2002) and infer the smallest average eruption repose interval (~ 10 yrs) for the fastest spreading rates, which occur on the southern EPR, and an average repose interval of $\sim 1,000$ years or more on the slower-spreading Mid-Atlantic Ridge.

Seismic monitoring, bottom-pressure recorders, lava-flow mapping, and radiometric dating collectively indicate that

each of the historical MOR eruptions observed to date has occurred as one or more short pulses of activity lasting a day to a week, and the longest-lived eruptions (at 9°50'N EPR) have occurred over several short episodes spread over seven to 10 months (i.e., 2005–2006 eruptions). By contrast, both of the eruptions observed at deep submarine subduction zone volcanoes (Northwest Rota-1 and West Mata) have been continuous, multiyear eruptions with low magma flux rates. There are just two such observations thus far, but the style observed contrasts with the punctuated, high-flux MOR eruptions that have decadal repose times.

The building of longer-term, site-specific eruption histories is important for understanding the frequency and size of eruptions, and how they influence the growth of volcanoes and the stability of hydrothermal systems and ecosystems. We lack the necessary observations along most of the global MOR, but at Iceland, which is the only significant portion of the global MOR above sea level, over 1,000 years of historical eruption observations provide important clues. Icelandic eruptions cluster in time followed by decades to multiple centuries of volcanic inactivity (e.g., Gudmundsson, 2000; Sinton et al., 2005). Whether or not such a temporal pattern applies to other spreading centers is difficult to determine with the limited data in hand. Intriguingly though, clustering may be implied by the high apparent eruption frequency at the “Flow” site on the CoAxial segment of the JdFR. The two eruptions it experienced in just 12–13 years indicate much higher frequency than anticipated from spreading-rate-based, steady-state recurrence estimates

(Perfit and Chadwick, 1998).

An active area of research that will help with this effort is the development of high-precision methods for submarine volcanic chronologies that span the past several centuries. A paleomagnetic intensity method (e.g., Bowles et al., 2005, 2006) and a short-lived radioisotope method (e.g., Rubin et al., 2005; Bergmanis et al., 2007; Sims et al., 2008) both show promise. Radiocarbon dating of planktonic and benthic foraminifera collected from lava-flow tops is a promising approach to developing site-specific millennial eruption histories (Clague et al., 2010b).

MAKING AND STORING MAGMA

The composition of magma and how and where it accumulates in the crust before eruption affect eruption character. Nearly all magmas feeding submarine volcanoes are formed from melting of the mantle. Mid-ocean ridge basalts (MORBs) typically display a smaller range in composition (e.g., Rubin and Sinton, 2007) than subduction zone and intraplate volcano magmas (e.g., Stern et al., 2003). In all settings, local and regional variations in source chemistry, magma-formation process, and magma evolution lead to variations in volatile content (which, if high enough, leads to explosive eruptions) and magma viscosity (which is mostly a function of magma temperature, composition, and crystal content). These attributes, along with the shape and location of magma chambers, play a large role in determining eruption style.

Volatile content (e.g., CO₂, SO₂, and H₂O) is typically higher and produces successively more explosive volcanism in intraplate and arc settings. The generally

low volatile content of MORBs and the great ocean depth (> 2,500 m) of most MOR crests limit the amount of explosive activity along ridges. However, some MOR eruptions can have high volatile content (e.g., Helo et al., 2011) that produces small explosive deposits (e.g., Sohn et al., 2008; Clague et al., 2009). Volumetrically larger explosive deposits occasionally occur at near-ridge seamounts (e.g., Maicher and White, 2001; Clague et al., 2009).

Beyond volatiles, variations in magma chamber depth, geometry, size, replenishment rate, and convective heat loss due to hydrothermal activity in the overlying crust affect magma chemistry, temperature, and viscosity, and thus eruption style. Seismic and petrological data suggest that most MOR eruptions are fed from small liquid-rich melt segregations whose size and temporal stability are linked to magma supply (Sinton and Detrick, 1992; Perfit and Chadwick, 1998; Singh et al., 1998). Lenticular magma bodies are persistent in time and are spatially more continuous at faster spreading rates (e.g., Harding et al., 1993; Canales et al., 2006; Carbotte et al., 2012, in this issue), whereas at slow spreading rates, such chambers are rarely detected and are assumed to be ephemeral and laterally restricted (e.g., Singh et al., 2006). The depth of shallowest magma accumulation in the MOR crust and regional average calculated eruption temperature are strongly correlated, such that high magma supply (e.g., higher spreading rate) generally promotes eruption of cooler and more degassed magmas from shallower crustal chambers, and lower supply promotes accumulation of hotter and gassier magma bodies at progressively greater depths (Rubin and Sinton, 2007). This variability

in lateral extent, chemistry, and depth as well as local variations in magma supply and tectonic stress fields appear to affect the style of eruption. Fissure-fed, high-effusion-rate morphologies are more common at fast-spreading MORs (e.g., Sinton et al., 2002; Fornari et al., 2004; Soule et al., 2005, 2007), high-relief constructional pillowed eruptions from point-source vents are typical at slow-spreading MORs (e.g., Smith and Cann, 1992; Searle et al., 2010), and both types occur at intermediate-spreading ridges (e.g., Embley and Chadwick, 1994; Stakes et al., 2006). Less is known about crustal magma bodies at deep submarine arc and intraplate volcanoes, but deposits of both point-source and rift-zone eruptions are known in these settings, so they likely span a range of characteristics that depend on melt supply and maturity of the magmatic system.

The timescales and magnitudes of magma supply to and heat loss from submarine volcanoes affect eruption frequency and character and drive hydrothermal circulation. Studies of chemical variations within and between successive historical eruptions at several MOR sites with moderate to high magma supply imply a relative constancy to magma chamber thermal conditions with, at most, 10° to 20°C of magmatic heat loss over time periods of one to two decades (e.g., 9°50'N EPR, Goss et al., 2010; 17°S EPR, Bergmanis et al., 2007; JdFR CoAxial segment, Embley et al., 2000). Many of these repeat eruptions also preserve magma temperature gradients along the eruptive fissures that indicate short-wavelength spatial thermal variations beneath MORs and point to inefficient mixing of magma along the axis between or during eruptions (e.g., Rubin et al., 2001; Sinton

et al., 2002; Bergmanis et al., 2007; Goss et al., 2010). Furthermore, the preservation of significant compositional variability inherited from the mantle in many MOR magmas and the geometry of their expression in erupted lava flows have begun to provide useful information about how eruptions initiate and progress (e.g., Rubin et al., 1998, 2001; Bergmanis et al., 2007; Goss et al., 2010). The first detected historical eruption on a back-arc spreading ridge (the 2008 Puipui eruption) displays extreme compositional variation in a very small, rapidly emplaced lava unit (Rubin et al., 2009) erupted from short fissure segments spanning only ~ 2 km of ridge crest (Clague et al., 2010a). This variation suggests that the short-lived eruption (Baker et al., 2011) was fed by multiple small, isolated lenses of compositionally unrelated magmas from a complex magma delivery and storage network.

SUMMARY OF DEEP-SEA ERUPTIONS

A growing body of observations from individual eruptions in the deep ocean has significantly improved our understanding of submarine volcanic eruption mechanisms and their relationships to magma generation. At the large scale, differences in eruption style at ridges, arcs, and intraplate volcanoes are primarily related to tectonic setting, magma composition, and volatile content. Short-duration, high-eruption-rate events or clusters of events dominate MORs, whereas explosive, low-mass eruption rates and more continuous eruptions dominate the recently defined deep-sea arc volcanism. However, if arc volcanoes on land are analogs, there is likely a continuum of eruptive activity, ranging from effusive low-magma,

high-gas-throughput eruptions to infrequent, large, explosive, caldera-forming events. Limited data suggest that intraplate volcanoes like Loihi erupt in both MOR and arc-like styles, with a substantial pyroclastic component in some instances. Over time, these differences translate into the type of volcanic edifice that is constructed. MORs produce low-lying elongate volcanoes from the combined effects of fissure eruptions and the constant rafting away of eruption deposits due to plate separation. In the case of fast-spreading ridges, where magma erupts at relatively high frequency from nearly continuous along-ridge melt lenses, the boundaries of individual volcanoes, and individual eruptions, become difficult to define. By contrast, arc and intraplate volcanoes have longer-lived, localized eruptive centers that often produce isolated composite cones with steep sides and unstable slopes.

Decadal-scale studies at several MOR sites with high melt supply reveal that magma chambers are persistent features over eruptive cycles (e.g., Carbotte et al., 2012, in this issue) and that they maintain discernible geochemical gradients within a multi-eruption episode that may span one or several centuries. Whether magma input is steady or time variable, magma chamber pressures increase over time (e.g., Nooner and Chadwick, 2011), triggering increased seismicity rates along the ridge over several years (e.g., Tolstoy et al., 2006). Changes in hydrothermal fluid chemistry and temperature as hydrologic conditions adjust to a vertically migrating thermal boundary layer in response to short-term fluctuations in magma recharge accompany this seismicity (e.g., Fornari et al., 2012, and

Kelley et al., 2012, both in this issue). When sufficient pressure has built in the magma body, diking initiates and may reach the surface, given sufficient overpressure. This diking is probably related to the timing of when event plumes form (e.g., Dziak et al., 2007) by advection of magmatic heat to the seafloor and/or discharge of hydrothermal fluids stored in the crust (e.g., Baker et al., 2011, and 2012, in this issue).


Eruption rate variations, which themselves depend on episodes of magma overpressure and variations in magma chamber geometry/connectivity, will produce a range of lava morphologies. Lava flows will advance for hours or days during each eruption pulse. Eruption pulses may continue if the pressure release was incomplete and/or pressure continues to build in the melt lens by recharge. The amount of magmatic gas within the system and the eruption depth will determine the extent of explosive activity that might accompany the effusive component of the eruption. Widespread diffuse venting of hydrothermal fluids continues through the carapace of a new lava flow well after the eruption wanes. For some eruptions (e.g., CoAxial 1993, North Gorda 1996, Puipui 2008), this type of hydrothermal activity is the only one observed, but for sites with pre-existing focused-flow hot springs, these systems may regain a steady-state condition within several years of the last magmatic event. (e.g., EPR 9°50'N; see Fornari et al., 2012, in this issue).

Researchers are currently moving eruption detection and response studies into a new realm, with new tools and methods, as described in Baker et al. (2012, in this issue). Observations at sites of continuous eruptive activity

and rapid response to serendipitous eruption detection are extraordinary research opportunities that will continue to inform our knowledge of submarine volcanism so long as humans have a presence on the seafloor. There is much yet to learn about volcanism in Earth's most active and prolific volcanic province. It will take a concerted effort by the international scientific community to establish an adequate submarine volcanic monitoring and response effort at the scale necessary to significantly advance our understanding of active eruptions in this largely still hidden world.

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