

## Lava flows from a mid-1980s submarine eruption on the Cleft segment, Juan de Fuca Ridge

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**Abstract.** A series of lava flows with a total volume of  $0.05 \text{ km}^3$  were erupted in the mid-1980s along 17 km of the northern Cleft segment of the Juan de Fuca Ridge. Observations from camera tows and submersible dives show that the new flows are all similar in appearance and consist entirely of pillow lava with a mixture of smooth and striated surface textures, suggesting a relatively uniform eruption rate approaching  $1 \text{ m}^3/\text{s}$  at point source vents. The flows vary in size from small patches to large steep-sided ridges and were probably erupted from a dike intruded along the ridge axis because they are aligned along a linear fissure/graben system. Observations at north Cleft show that the physical appearance of new flows changes more rapidly than previously realized and that earlier qualitative dating of young lavas based on sediment cover and glassy surface texture were probably overestimates by an order of magnitude. Sediment accumulation on the lavas is quite variable and locally surprisingly substantial, mainly due to hydrothermal deposits that formed while the lava flows were cooling. Biological vent communities photographed on the new flows in 1989 show that vent animals can colonize new vent sites rapidly but that warm water was still venting only in a few places. Nonvent animals are much slower to colonize the new flows and rates of colonization observed at north Cleft may be useful for making improved age estimates of young (<10 years) lava flows elsewhere. The north Cleft eruption represents about 2% of the estimated average annual volcanic output along the global mid-ocean ridge, implying that many other submarine eruptions are occurring undetected.

### Introduction

A volcanic eruption occurred on the northern Cleft segment of the Juan de Fuca Ridge during the mid-1980s [Chadwick *et al.*, 1991; Embley *et al.*, 1991] and produced a series of isolated mounds of pillow lava along a distance of 17 km, between  $45^{\circ}00'$  and  $45^{\circ}10'N$  (Figure 1). The distribution of new lavas at north Cleft can be readily mapped, because of their fresh appearance and their age contrast with the underlying older lavas (Plates 1a, 1b, and 1c). The timing of this eruption is constrained by repeated Sea Beam bathymetric surveys that document significant seafloor depth changes up to 45 m [Fox *et al.*, 1992], located exclusively where the fresh glassy lava flows have been mapped by bottom photography and submersible observations (Figure 1). The before-and-after surveys are not the same for all the lava flows because some Sea Beam surveys were small and did not cover the whole area; one flow appeared between 1983 and 1987 surveys, but the others are less constrained in time (Table 1). Consequently, we cannot know for sure if the separate lava flows were erupted simultaneously. Nevertheless, throughout the following, we interpret that all of the pillow lava mounds were

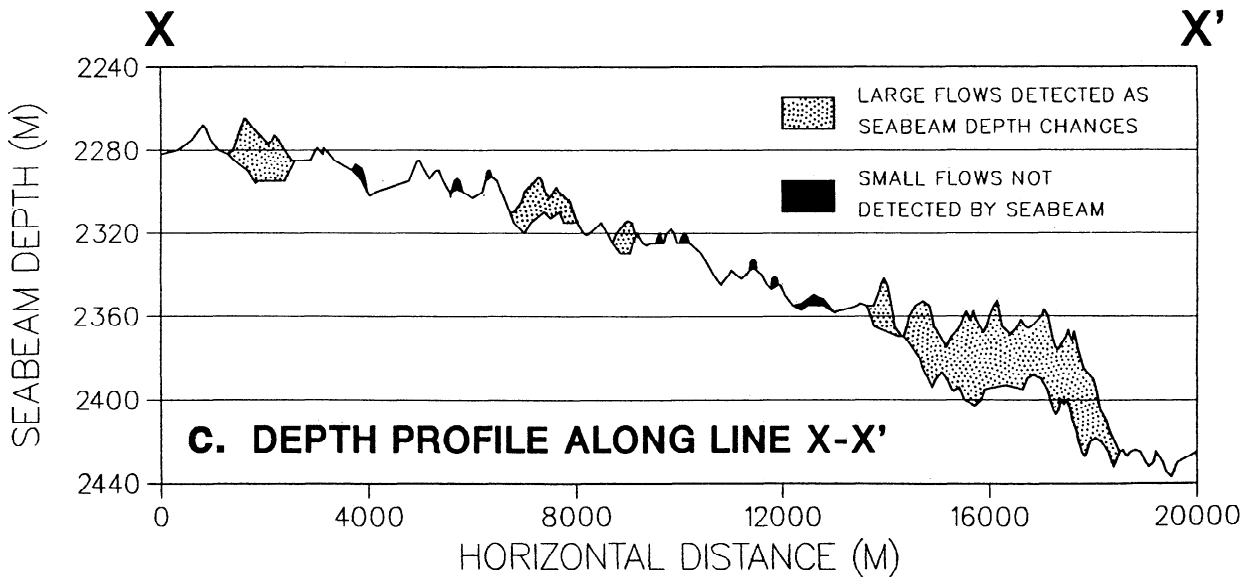
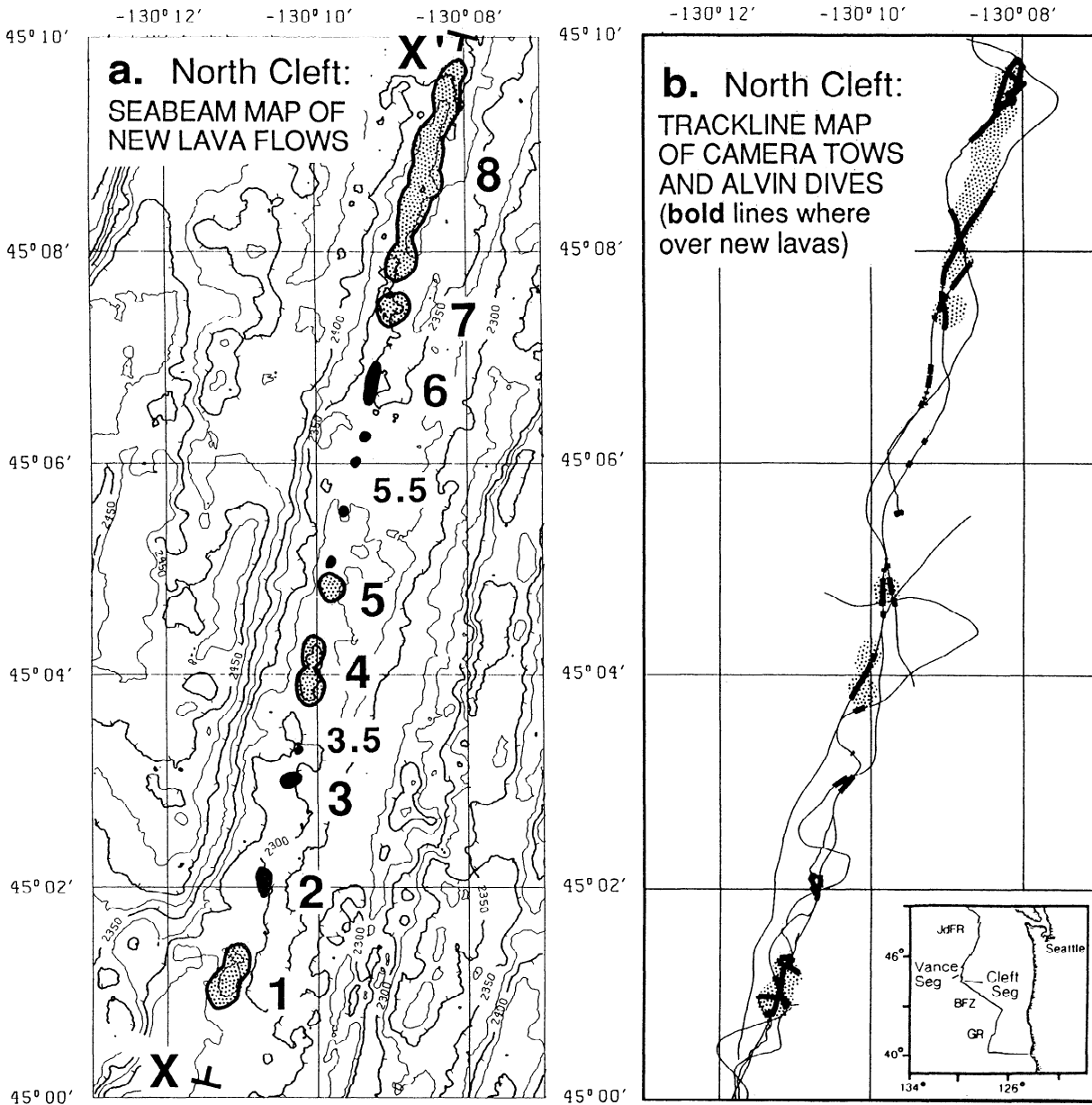
erupted together during one brief volcanic episode, because the flows are aligned along a single fracture system and because they are all similar in morphology, glassy surface texture, and degree of sediment accumulation [Chadwick *et al.*, 1991]. In addition, the chemical composition of separate flows is very similar and is distinct from surrounding older lavas [Smith *et al.*, this issue]. The erupted lavas cover a combined area of  $2\text{--}3 \text{ km}^2$  and have a volume of  $0.05 \text{ km}^3$  (Table 1).

These new pillow mounds are distinguished from a broad ( $2 \text{ km}^2$ ) sheet flow that has been mapped to the south of the mounds between  $44^{\circ}55'$  and  $44^{\circ}59'N$  [Embley and Chadwick, this issue] and is also very young. However, the sheet flow is apparently slightly older than the pillow mounds, based on repeated sidescan sonar and Sea Beam surveys, but probably only by a few years, because no significant differences between the two lavas are observed in terms of their glassy texture, degree of sediment accumulation [Embley *et al.*, 1991; Embley and Chadwick, this issue] or the abundances of sessile and sedentary animals that have colonized the flows [Milligan and Tunnicliffe, this issue]. This indicates that there have been at least two very recent eruptions at north Cleft.

This paper presents geologic observations of the new pillow flows, primarily from five camera tows and three Alvin submersible dives made in 1989–1991, and draws conclusions about the eruption that produced them. The appearance and morphology of the flows are illustrated with bottom photographs, detailed maps and cross sections, and sidescan sonar

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imagery. In many mid-ocean ridge (MOR) studies, relative age assignments of the seafloor are made based on the physical appearance of lavas, including the degree of glassy surface texture and sediment cover [Ballard *et al.*, 1981; Macdonald *et al.*, 1988]. However, due to the difficulty of dating young basalt, very few absolute ages have been obtained to “calibrate” these subjective estimates. The north Cleft lavas are one of the first “zero-age” benchmarks to which other young submarine flows can be compared. The recent eruption at north Cleft presents a rare opportunity to study the character of recent submarine lava flows of known age and the nature of a single eruptive event on the MOR.

## Description of the Lava Flows

### Lava Texture and Flow Morphology

The term “pillow lava” refers to the form of lava flows that consist of numerous rounded or cylindrical flow lobes or cooling units that resemble pillows in cross section. The lava flows that were erupted at north Cleft in the 1980s are composed entirely of pillow lavas, and we refer to these as “pillow mounds” because of their shape. The lavas display a variety of forms and surface textures (Plates 1 and 2) that can be interpreted in terms of eruption and flow processes.

The new pillow mounds consist of a mixture of two distinct surface textures: (1) large elongate pillows with striated surfaces, and (2) smaller smooth-surfaced pillow lobes. The elongate pillows are typically 1–2 m in diameter and display corrugations parallel to their long axes (Plate 1b). The pillows with smooth surfaces are almost always smaller (<1 m in diameter) than the striated pillows, and resemble subaerial pahoehoe toes (Plate 1c). These smooth pillows are similar to the smooth forms described as “pillow buds,” “pillow fingers,” or “elephant trunks,” by Ballard and Moore [1977] from the Mid-Atlantic Ridge but are larger and much more common at north Cleft. The two flow textures are observed in roughly equal proportions and are usually intermingled (Plate 1d), indicating that they are not produced by separate flow events but rather were formed contemporaneously.

Observations of active underwater lava flows in Hawaii, when subaerially erupted flows enter the ocean, have aided in the interpretation of the processes that affect pillow lava morphology [Moore *et al.*, 1973; Moore, 1975; Tribble, 1991]. The form and surface texture of a pillow are dependent on the rate of lava supply and the thickness of surface crust during growth [Ballard and Moore, 1977; Walker, 1992]. Slow growth allows a relatively thick brittle crust to form on a

pillow, which then must be fractured along a spreading crack in order for the pillow to grow further; new crust formed along such a spreading crack is striated perpendicular to the crack. Faster growth allows a pillow to expand rapidly within a thin plastic skin that can stretch without fracturing, and its surface remains smooth in appearance. Smooth-surfaced pillows tend to be smaller because they can only grow to a certain size before the crust becomes too thick and strong to continue stretching plastically [Moore, 1975]. At this point, pillow growth either ceases or continues but is subsequently accommodated by brittle fracture and spreading (Plate 1e). The surface of the smooth pillows is much more glassy and light reflective than the corrugated surfaces of the elongate pillows, because of their different mechanisms of crust formation and because the smaller, shorter-lived smooth pillows cool more quickly.

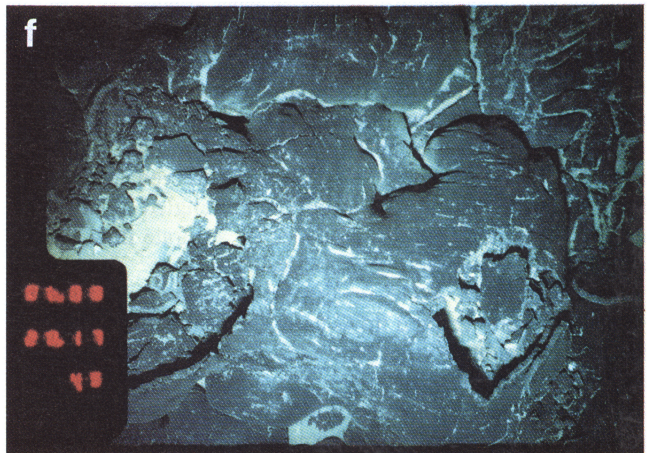
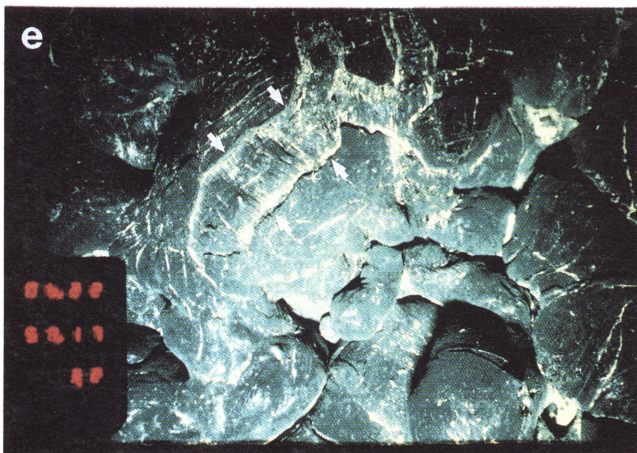
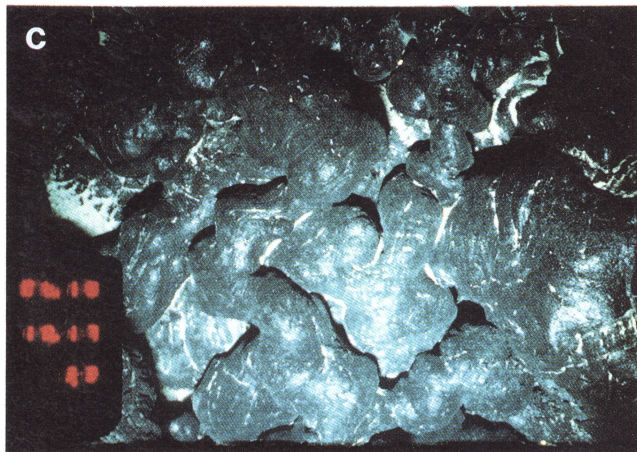
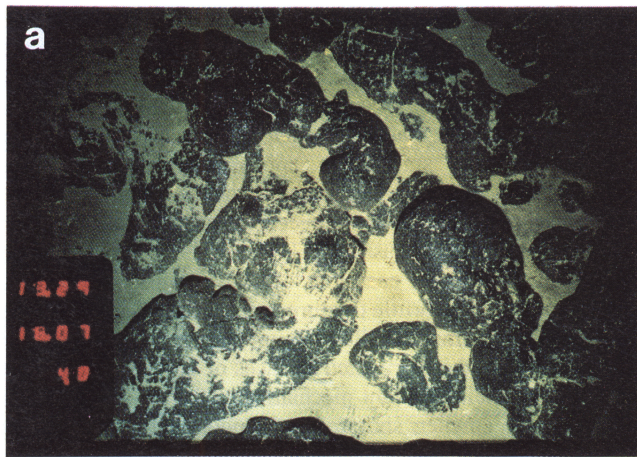
High points on the pillow mounds mark the sites of inferred eruptive vents. Smooth-surfaced pillows tend to be dominant near the high points, where they often display more fluid forms, such as broad, flattened lobes up to several meters across or minor collapses due to lava drain-out (Plate 1f), suggesting a locally high effusion rate. Elongate striated pillows radiate outward from high points and tend to be dominant on the steepest flanks of the mounds. The striated pillows served as distributary tubes that fed fluid lava from an eruptive vent down to the flow front, whereas the smooth pillows appear to be the local output of the distributary system. The smooth lobes commonly have flowed out of the high-standing striated pillow tubes and tend to fill in low areas between them (Plate 1g). The larger diameter of the elongate pillows allowed them to be long-lived feeding channels and the flowing lava within them supplied heat that retarded solidification. It appears the pillow mounds were built up by magma rising up through the mound, erupting out the top, and flowing down the sides. The primary paths of lava distribution from the vents were large pillow tubes, while smooth pillow forms represent local, rapid, and brief “leakage” of lava from the primary distributary system.

The new pillow flows erupted at north Cleft are aligned along a trend oriented 015°, and extend over a distance of 17 km (Figure 1). All the new lavas are similar in appearance, with the same mix of striated and smooth textured pillow lavas. This suggests that the extrusion rate and eruption style were relatively uniform along the entire line of volcanic vents, since submarine lava morphology is primarily a function of eruption rate [Griffiths and Fink, 1993]. The flows vary in size from small isolated patches to larger steep-sided mounds (Table 1 and Figures 2–5), the largest of which is a ridge 4.2 km long, 0.5 km wide, and 45 m thick (mound 8 in Figure 1).

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**Figure 1.** (a) Sea Beam bathymetric map of the northern Cleft segment of the Juan de Fuca Ridge (25-m contours), showing the locations of the recently erupted pillow mounds (numbered for identification in text and Table 1). Larger lava flows (stippled) were detected as depth changes by repeated Sea Beam surveys [from Fox *et al.*, 1992]; smaller flows (solid) are mapped by camera tows (because camera coverage is sparse, there may be other small unmapped flows along the axis of the eruptive fissure). “Mound 5.5” represents four separate small lava patches between mounds 5 and 6. (b) Map of track lines of 1989–1991 camera tows and Alvin dives, tracks are bold where they cross fresh lavas. Stippled areas are Sea Beam depth anomalies from Figure 1a. Inset map shows location of Cleft segment (GR, Gorda Ridge; BFZ, Blanco Fracture Zone; JdFR, Juan de Fuca Ridge). (c) Depth profile along line of lava flows, between X and X' in Figure 1a (40 times vertical exaggeration). Thickness of larger flows (stippled) from before-and-after Sea Beam surveys; smaller flows (solid) are given an arbitrary thickness of 5 m. Note that all maps presented in this paper are in Global Positioning System (GPS) coordinates, but that maps in previous related papers [Chadwick *et al.*, 1991; Embley *et al.*, 1991; Fox *et al.*, 1992] were in LORAN-C coordinates.







**Table 1.** Dimensions, Areas, and Volumes of the North Cleft Pillow Mounds

Mound	Estimates Based on Camera Tows				Estimates Based on Sea Beam Data*			
	Length, m	Width, m	Height, m	Area, m <sup>2</sup>	Before-and-After Sea Beam?, years	Height, m	Area, m <sup>2</sup>	Volume, m <sup>3</sup>
1	1100	500	30	4.3 × 10 <sup>5</sup>	yes, 1983–1987	25	3.3 × 10 <sup>5</sup>	5.2 × 10 <sup>6</sup>
2	400	200	5	6.3 × 10 <sup>4</sup>	no	—	—	—
3	450	200	10	7.1 × 10 <sup>4</sup>	no	—	—	—
3.5	50	50	2	2.0 × 10 <sup>3</sup>	no	—	—	—
4	1150	300	25	2.7 × 10 <sup>5</sup>	yes, 1981–1990	20	2.1 × 10 <sup>5</sup>	2.4 × 10 <sup>6</sup>
5	650	300	20	1.5 × 10 <sup>5</sup>	yes, 1981–1990	15	8.2 × 10 <sup>4</sup>	8.4 × 10 <sup>5</sup>
5.5†	100	100	5	7.9 × 10 <sup>3</sup>	no	—	—	—
6	750	150	5	8.8 × 10 <sup>4</sup>	no	—	—	—
7	600	300	>20	1.4 × 10 <sup>5</sup>	yes, 1981–1987	30	2.0 × 10 <sup>5</sup>	3.3 × 10 <sup>6</sup>
8	4200	500	60‡	1.6 × 10 <sup>6</sup>	yes, 1981–1987	45	1.6 × 10 <sup>6</sup>	3.9 × 10 <sup>7</sup>
Totals				2.8 × 10 <sup>6</sup>			2.4 × 10 <sup>6</sup>	5.1 × 10 <sup>7</sup>

\* Dates are given for before-and-after Sea Beam surveys that detected a bathymetric change at the location of each mound. Mounds that were not detected by the later Sea Beam surveys are consistently smaller than 10–15 m in height and  $1 \times 10^5$  m<sup>2</sup> in area, which is apparently the effective resolution limit of the quantitative comparison technique [Fox *et al.*, 1992]. All lava flows were in place in 1989 when photographed by camera tows. The height of each mound measured by Sea Beam is the maximum value of bathymetric change, but is 5–15 m less than the apparent height measured by camera tow microbathymetry, because of the spatial averaging of the Sea Beam system. The areas and volumes calculated by the Sea Beam comparison technique are minimums, because they do not include the smaller flows or the thin flow margins of larger flows.

† Mound 5.5 actually represents four separate small patches of new lava between mounds 5 and 6 (Figure 1). Because the camera coverage is sparse, there are probably other small unmapped flows along the axis of the eruptive fissure.

‡ Apparent height from *Alvin* dive 2430.

The areas of significant depth change between the before-and-after Sea Beam surveys show the overall shape and outline of the larger mounds, in relation to camera tows and *Alvin* dives (Table 1 and Figures 2–5); the smaller mounds were apparently below the detection threshold of Sea Beam [Fox *et al.*, 1992].

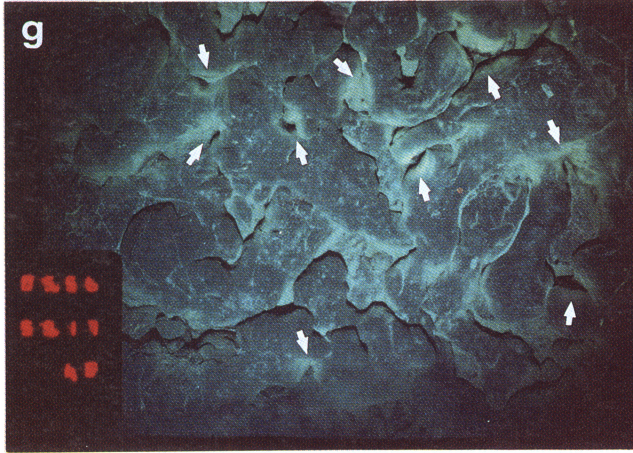
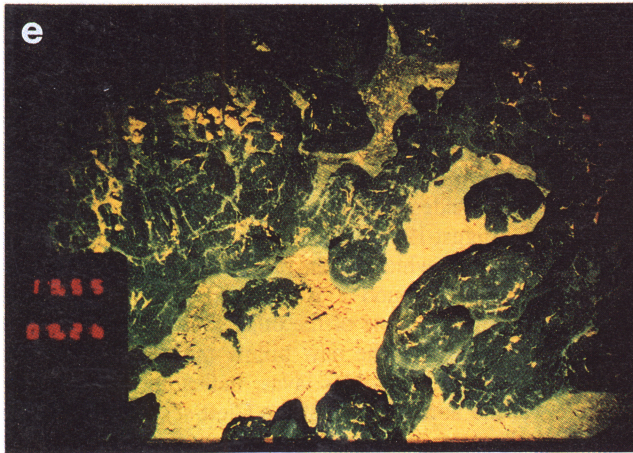
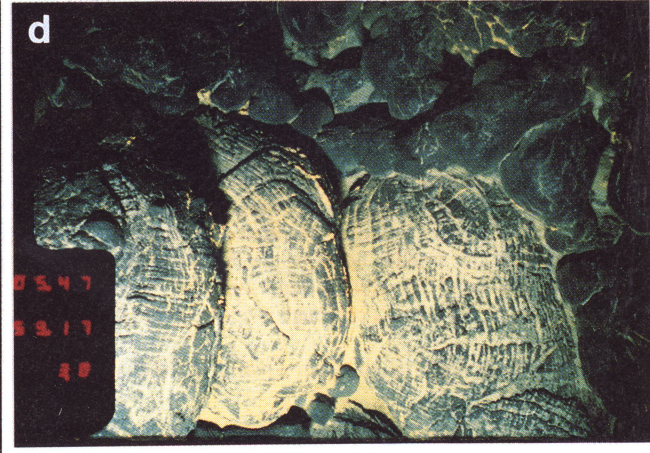
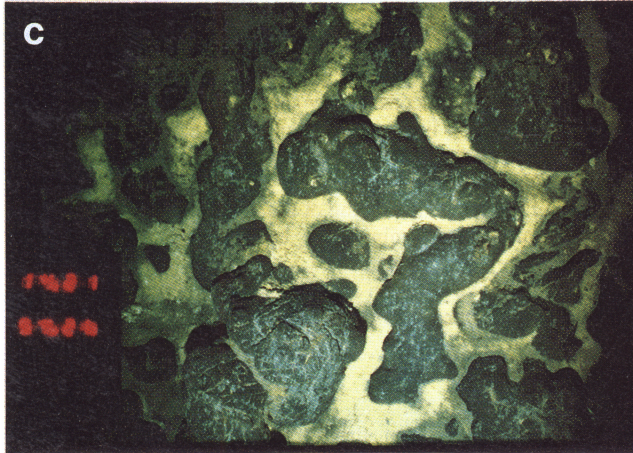
The margins of the flows are typically thin and only one pillow thick (Plate 1h), except in places where they flowed up against preexisting fault scarps, in contrast to the steep (5–20 m) flow fronts reported on some other pillow flows [Ballard and Moore, 1977]. In cross sections perpendicular to the ridge axis, the larger mounds have smooth haystack-shaped profiles suggesting each was built up during a single episode of extrusion (Figures 2c, 3b, 3c, 4c and 4e). Profiles of the mounds parallel to the ridge axis may have multiple culminations along their length (Figures 1c, 2b, 3d, 3e, 4b, 4d, 5b, and 5c). These high points mark the sites of late stage eruptive vents, which are characterized by radiating pillow tubes but no distinct vent structures, such as craters or open eruptive fissures, or “pillow walls” (such as described by Fornari *et al.*

[1978]). A lack of vent structures on pillow flows was also noted at the Mid-Atlantic Ridge [Ballard and Moore, 1977; Ballard and van Andel, 1977], suggesting that lava tends to freeze in place when the eruption of a pillow flow ceases (because of the relatively low extrusion rate), rather than undergoing extensive withdrawal and collapse commonly observed on more fluid sheet flows [Ballard *et al.*, 1979; Francheteau *et al.*, 1979; Embley and Chadwick, this issue]. A SeaMARC I sidescan sonar survey of the Cleft segment, collected in August 1987, shows that mound 1 is a smooth and unfractured feature which covers preexisting structures (Figure 6). Unfortunately, mound 1 is at the northern limit of the sidescan survey, and none of the other mounds were imaged.

The morphology of the new pillow mounds is similar to that of the surrounding older lavas, indicating that previous eruptions at the northern end of the Cleft segment have been similar in character and dominantly pillow flows. Sheet flows are more common on the rest of the segment to the south [Embley and Chadwick, this issue], however, which is consistent with observations elsewhere of a higher ratio of pillow

**Plate 1.** (opposite) Photographs showing various lava morphologies of new pillow mounds and older lava at north Cleft (see text for discussion). Each description is followed by the number of the mound photographed (M number; see Figure 1), the number of the camera tow (C number) or *Alvin* dive (A number), the time, and the horizontal dimension of the photograph. Camera tow views are vertical; *Alvin* views are oblique. The locations of photographs of mounds 1 and 4 are shown in Figures 2 and 4. (a) Older seafloor for comparison with the new lavas (n/a, C89-3; 13:29:16; 4.5 m). (b) Elongate striated pillows (M4, C89-14; 11:43:19; 3.3 m). (c) Smooth-surfaced pillows (M1, C89-11; 06:18:10; 3.1 m). (d) Mixed striated and smooth pillows (M1, A2262; 10:50; ~3 m). (e) Smooth pillow with late stage spreading crack, shown by arrows (M1, C89-11; 06:22:57; 3.0 m). (f) Smooth pillows with broad fluid forms and minor collapse, common near vent areas (M1, C89-11; 06:08:20; 4.2 m). (g) Smooth pillows where they have clearly flowed out of larger striated pillows (M1, A2432; 10:30:18; ~3 m). (h) Southwest contact of mound 1 showing thin flow front over lighter older lava in foreground (M1, A2262; 12:43; ~3 m).







lava to sheet flows near the distal ends of ridge segments [Francheteau and Ballard, 1983; Bonatti and Harrison, 1988].

### Character of the Eruption

Many characteristics of the pillow mounds suggest that they were erupted from a dike that intruded along the north Cleft neovolcanic zone but that only reached the surface intermittently along its length. The individual flows are very similar in their petrologic composition [Smith *et al.*, this issue], gross morphology, and detailed appearance. The mounds are aligned along the northern extension of a hydrothermally active fissure system that bisects the neovolcanic zone, and which is also the apparent source for the recent sheet flow to the south [Embley *et al.*, 1991; Embley and Chadwick, this issue]. Individual mounds are elongate and appear to have been fed from linear eruptive vents, parallel to this trend (Figure 1). It is commonly observed during dike-fed fissure eruptions on land that soon after an eruption begins along a "curtain of fire," most of the fissure system shuts down, the flow of magma localizes, and lava is discharged from a few discrete vents for the remainder of the eruption [Richter *et al.*, 1970; Swanson *et al.*, 1979; Delaney and Pollard, 1982; Wolfe *et al.*, 1987; Bruce and Huppert, 1989]. The morphology of the mounds suggests that this same process happened at north Cleft; a brief early stage of eruption along much of the rift (indicated by the many small flows), followed by more sustained but localized eruption that produced the larger mounds. The fact that 80% of the total volume of lava extruded makes up the northernmost lava flow (Table 1) indicates that this end of the fissure system was by far the longest-lived vent area during the eruption. This may indicate that the northern end of the line of pillow mounds had more accumulated tectonic tension to relieve than the southern end, possibly because the dike that fed the earlier sheet flow eruption just south of the pillow mounds [Embley and Chadwick, this issue] had relieved some of the strain there. The multiple culminations along the length of the larger flows show that even on the scale of a single mound, the eruption localized at discrete vents in its later stages.

The young lavas are unbroken by fissures, whereas faults and fissures are common in the older lavas below. The two southernmost mounds (mounds 1 and 2) were examined in detail by Alvin and both were found to overlie grabens. The graben south of mound 1 is about 100 m wide and 5 m deep, has open cracks at the base of both bounding faults, and has sparse low-temperature hydrothermal vents and tube worm communities along its floor (Figures 2 and 6). The other

graben south of mound 2 formed on an east facing slope and is 50 m wide and 2–5 m deep, with a 2- to 5-m-high horst in the middle (Figure 7). New lava covers the floor of the graben where it emerges from the southern edge of mound 2 (Figure 7). The highest points on mounds 1 and 2 (the inferred eruptive vents) are located directly over the along-strike extension of these grabens (Figures 2, 3, 6, and 7). These relations suggest that the grabens may have formed during the north Cleft dike intrusion and then were partially buried by lava when the dike reached the surface to erupt. In subaerial volcanic rift zones, grabens often form above an intruding dike, because the dike creates two zones of tension near the surface, parallel to the dike axis but offset to either side [Pollard *et al.*, 1983; Rubin and Pollard, 1988; Rubin, 1992]. Graben formation is a manifestation of the horizontal extension that likely occurred across the Cleft neovolcanic zone during this seafloor spreading event (perhaps up to several meters, based on subaerial analogs).

No earthquakes larger than the  $m_b$  4.0 detection threshold occurred between 1981 and 1987 within 1° (~110 km) of the north Cleft eruption site (R.W. Dziak, personal communication, 1991). This is not surprising since the Juan de Fuca Ridge has historically been seismically very quiet and since microearthquakes associated with dike intrusion in volcanic rift zones on land rarely exceed magnitude 4.0 [Brandsdottir and Einarsson, 1979; Einarsson and Brandsdottir, 1980; Klein *et al.*, 1987]. This shows that significant eruptions can occur along the MOR system undetected by land-based sensors.

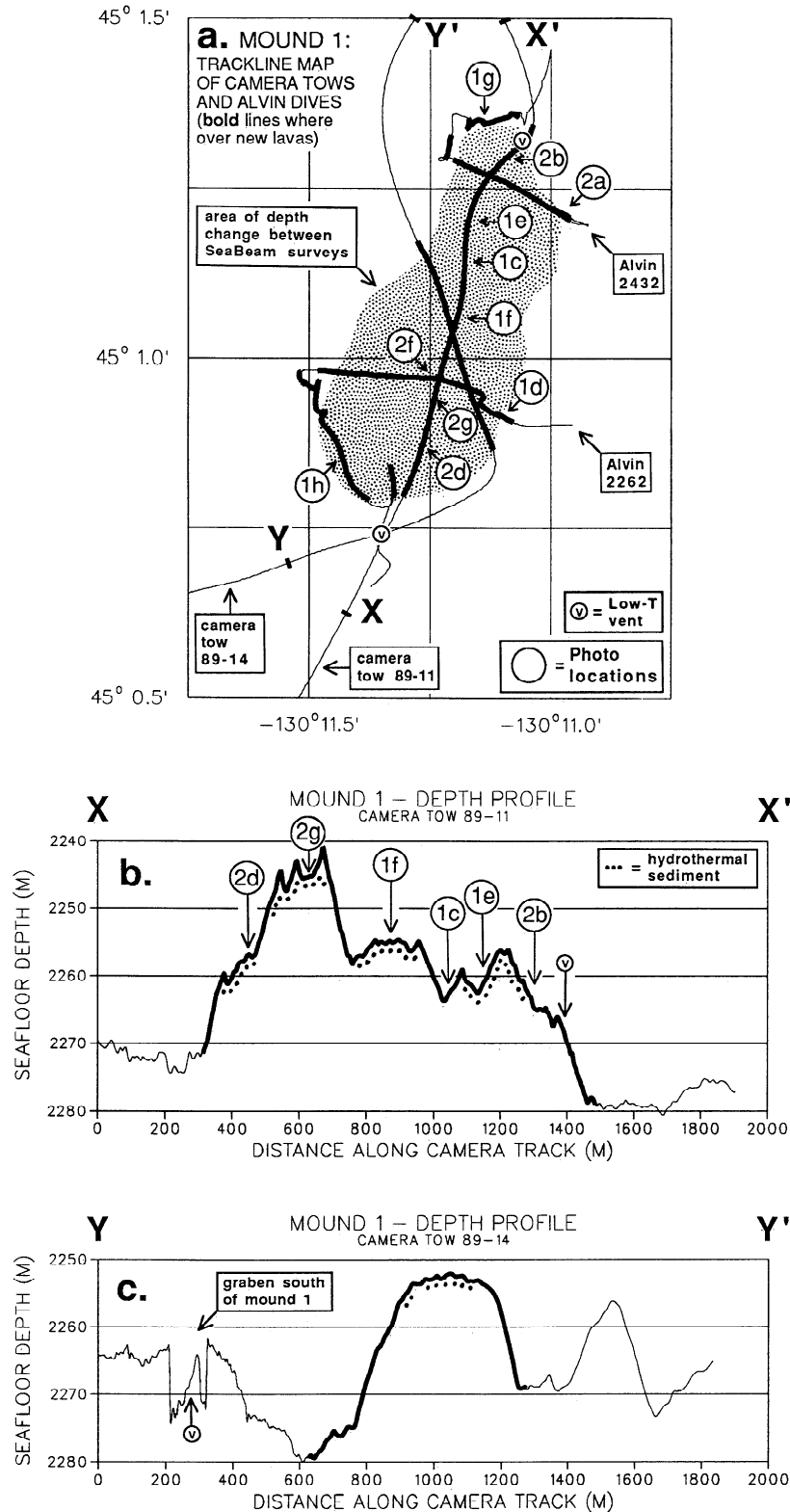
Monitoring data show that dikes in subaerial volcanic rift zones propagate primarily horizontally, away from shallow crustal magma reservoirs [Rubin and Pollard, 1987; Sigurdsson, 1987], and we assume this is also true along the MOR. Without monitoring data, it is impossible to know exactly from where the north Cleft dike intruded. However, we speculate that the dike intruded from south to north, possibly from the sheet flow area (where high-temperature hydrothermal vents suggest an underlying magma body [Embley and Chadwick, this issue]), simply because the pillow mound eruption occurred at the very northern end of the Cleft segment. This direction also follows the topographic gradient along the ridge axis; the depth of the seafloor gradually increases by 140 m from south (2280 m) to north (2420 m) along the line of pillow mounds (Figure 1c).

### Sediment Accumulation on the New Lavas

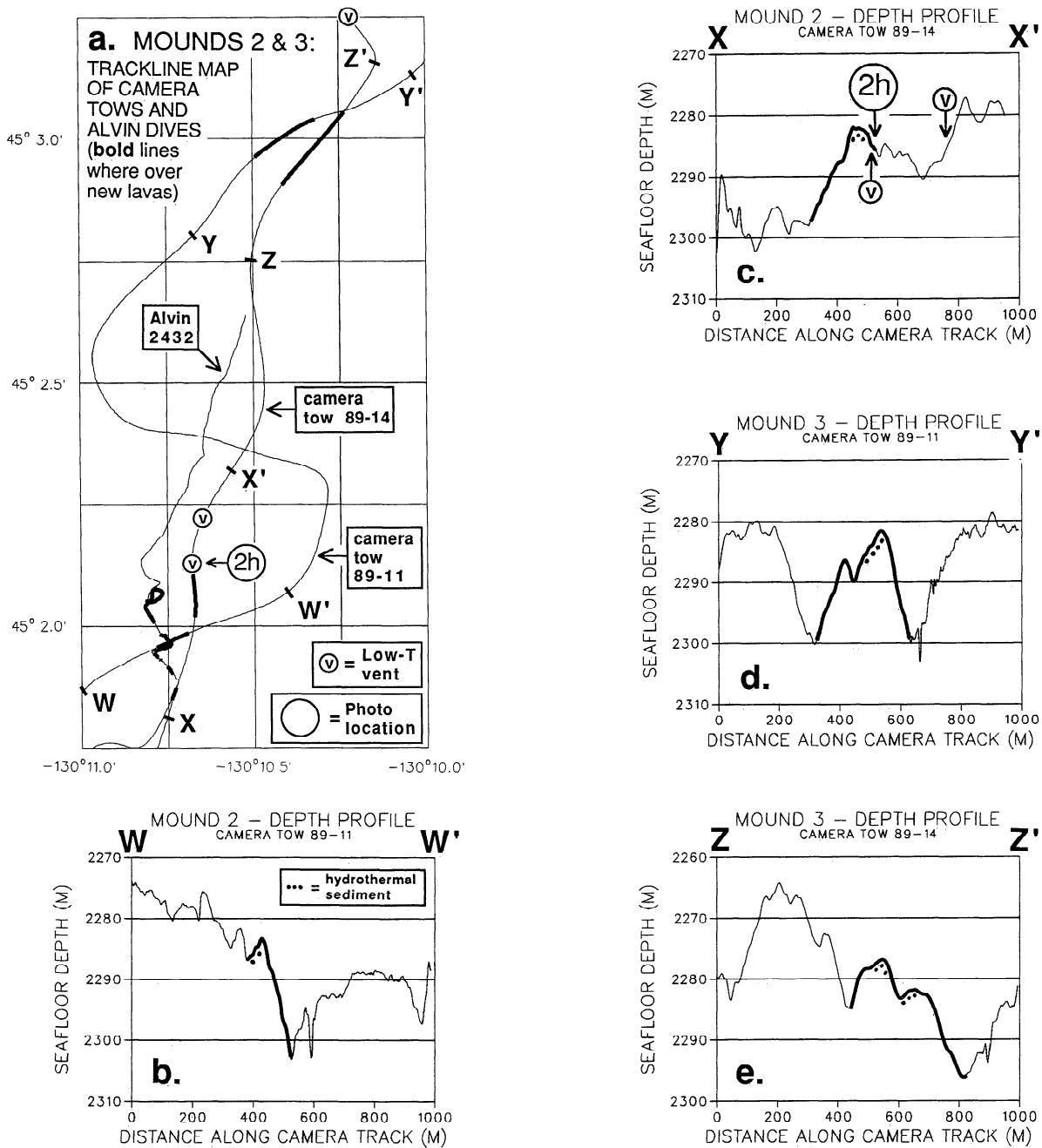
The amount of sediment that has accumulated on the new lavas since the eruption is quite variable. In some places, there is little or no sediment (Plate 2a), but in most areas a light

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**Plate 2.** (opposite) Photographs showing variation in sediment accumulation and biological colonization on the new lavas (see text for discussion). Each description followed by photo data as in Plate 1. The locations of photographs of mounds 1 and 2 are shown in Figures 2 and 3. (a) Nearly pristine lava with little sediment and glassy luster (M1, A2432; 09:50; ~2 m). (b) Light patchy sediment cover (M1, C89-11; 06:31:00; 3.2 m). (c) Heavy accumulation of hydrothermal sediment in interconnected pockets between pillows (M8, C91-8; 14:21:54; ~5 m). (d) Difference in sediment accumulation due only to difference in lava surface texture, because corrugated pillows trap sediment more readily (and appear "older") than smooth pillows (M1, C89-11; 05:47:53; 2.8 m). Note that Fox *et al.* [1992] mistakenly stated that this photograph showed a flow contact, but the photo location is well within mound 1 lavas (see Figure 2). (e) Thick accumulation of clumpy yellow hydrothermal sediment (M8, C91-8; 13:55:09; ~3 m). (f) Light dusting of fine yellow hydrothermal sediment (M1, A2262; 11:16; ~2 m). (g) Halos of fine hydrothermal sediment formed around small orifices and the edges of flow lobes (arrows), where warm water apparently vented diffusely from below (M1, C89-11; 05:56:53; 5.6 m). (h) Biological vent community already colonized on new lavas in 1989 (M2, C89-14; 09:08:29; 4.2 m).



**Figure 2.** (a) Map of camera tow and Alvin dive tracks over mound 1, bold where over new lavas. The area of depth change between Sea Beam surveys (stippled) is from Fox *et al.* [1992]. Circled numbers are locations of photographs in Plates 1 and 2. Circled Vs show locations of low-temperature hydrothermal vents with biological communities. (b) Depth profile along camera track over mound 1 between X and X' in Figure 2a, parallel to inferred eruptive fissure. (c) Depth profile of mound 1 between Y and Y' in Figure 2a, perpendicular to eruptive fissure. Dotted areas beneath profiles show areas with hydrothermal sediment, mainly along the crest of mound. Cross sections have 20 times vertical exaggeration.

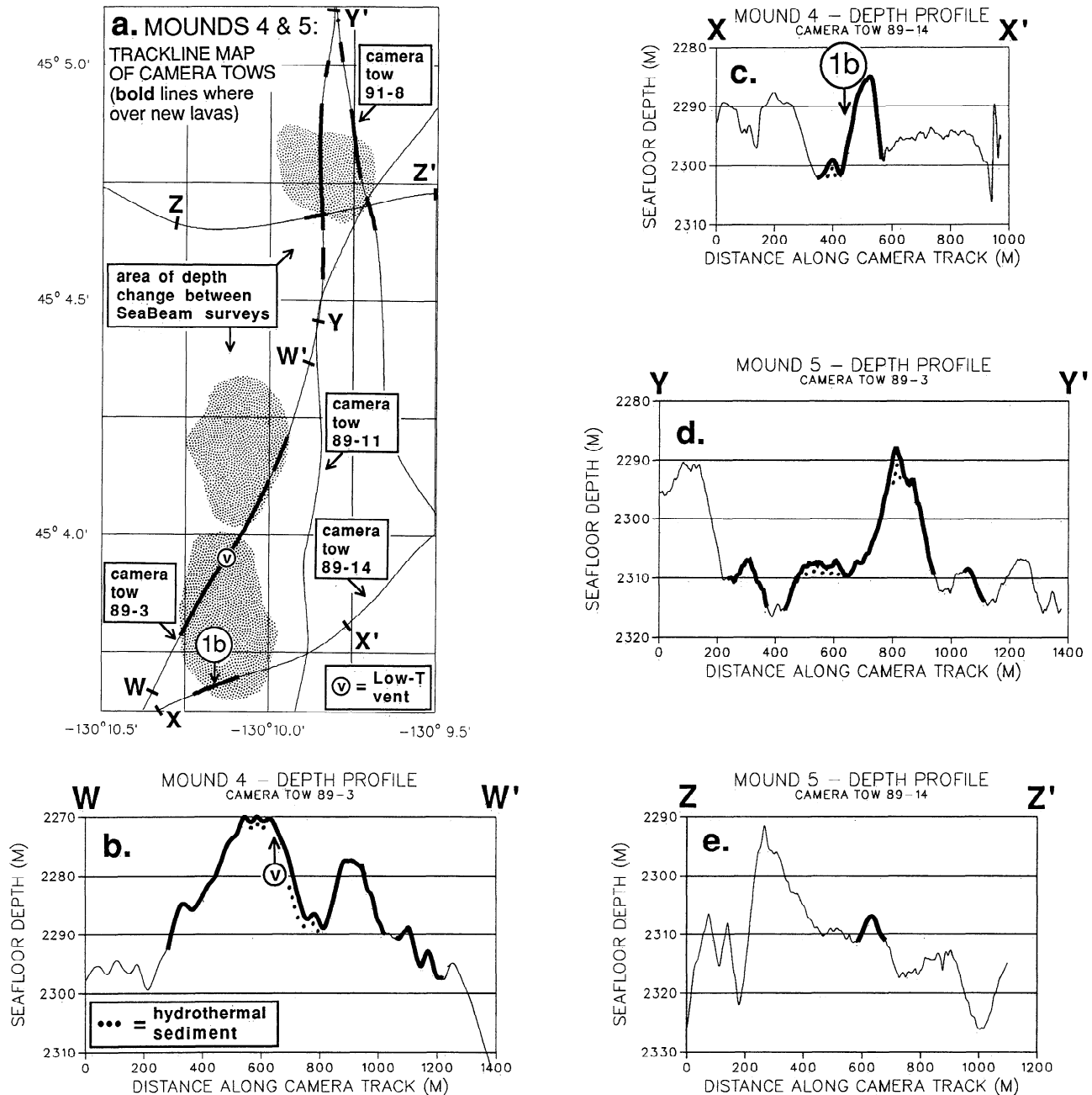


**Figure 3.** (a) Map of camera tow and Alvin dive tracks over mounds 2 and 3, bold where over new lavas. Note these mounds were too small to be detected by Sea Beam. All depth profiles along camera tracks with 20 times vertical exaggeration. (b) Mound 2 between W and W' in Figure 3a; (c) mound 2 between X and X' in Figure 3a; (d) mound 3 between Y and Y' in Figure 3a; (e) mound 3 between Z and Z' in Figure 3a. All symbols as in Figure 2.

dusting is already deposited on the lavas. Locally it is surprisingly abundant (Plate 2b), even collecting in interconnected pockets between the new pillows, perhaps up to 5–10 cm thick (Plate 2c). The long-term sedimentation rate in the vicinity of the southern Juan de Fuca Ridge is about 1–2 cm/1000 years, as measured in deep-sea cores [Carson, 1971]. Short-term sedimentation rates are difficult to quantify, but sediment trap data [Baker et al., 1985, E. Baker, personal communication, 1992] suggest rates of about the same order of magnitude ( $50 \text{ mg m}^{-2} \text{ d}^{-1} = 0.01 \text{ mm/yr}$ , assuming a density of  $1.5 \text{ g/cm}^3$ ). Once on the seafloor, however, the sediment is

redistributed by bottom currents and lava microtopography influences where sediment is swept free or is concentrated. One reason for the variability of sediment cover on the new lavas is that the surfaces of striated pillows trap sediment more readily than the smooth-surfaced pillows (Plate 2d). However, the main reason for this variability is that locally derived hydrothermal sediments, in addition to the background pelagic sediment, have been deposited on the lavas in many places; 21% of camera tow photographs over the new lava flows (355 of 1692) include hydrothermal sediments.

There are two distinct kinds of hydrothermal deposits



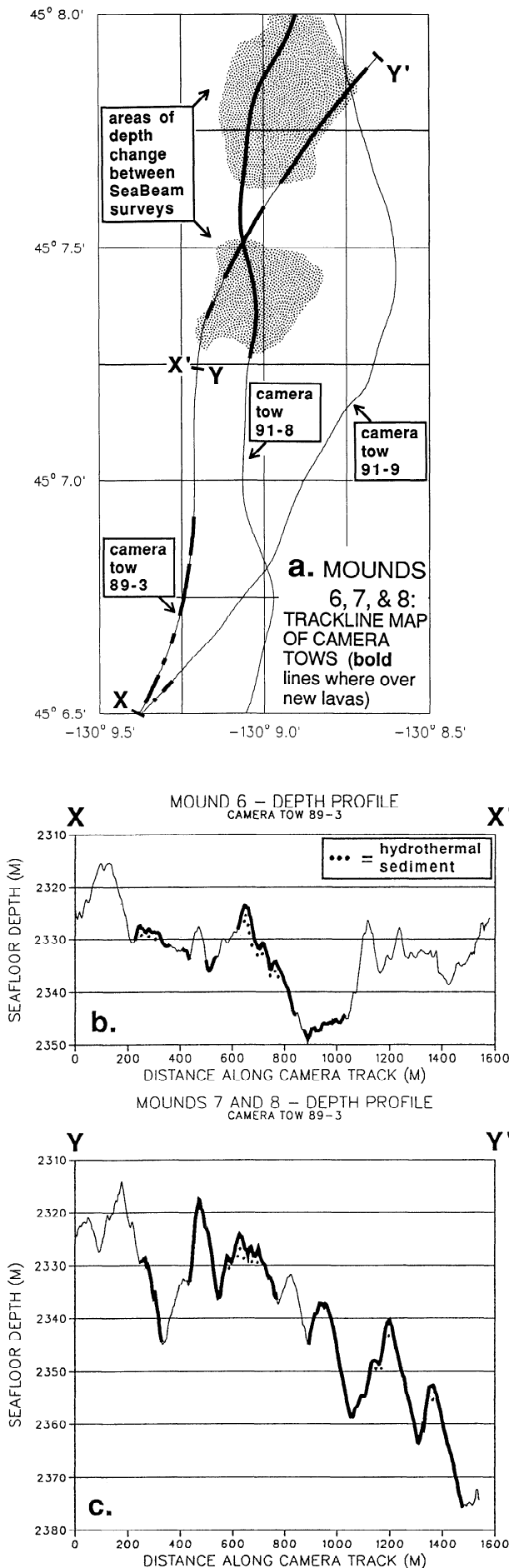
**Figure 4.** (a) Map of camera tows over mounds 4 and 5, bold where over new lavas. Areas of depth change between Sea Beam surveys are stippled. All depth profiles along camera tracks with 20 times vertical exaggeration. (b) Mound 4 between W and W' in Figure 4a; (c) mound 4 between X and X' in Figure 4a; (d) mound 5 between Y and Y' in Figure 4a; (e) mound 5 between Z and Z' in Figure 4a. All symbols as in Figure 2.

observed on the new lavas: (1) clumps of a yellow gel-like iron-rich silica phase that apparently formed along small cracks on individual pillows as they cooled and that later tends to fall off and accumulate between pillows (Plate 2e), and (2) a fine yellow dust that was apparently deposited where warm water vented diffusely out of the flows (Plate 2f). We interpret that both of these deposits formed from short-lived hydrothermal venting as the hot, newly erupted lava flows (and the underlying dike) cooled. The pillow flows are probably quite permeable, because as pillow lava accumulates, a network of open cavities remains between lobes. Heated interstitial water then flowed upward through the mound, interacting chemically with the hot rock, and vented out the top. This interpretation

is supported by several observations: (1) the fine dusty deposits tend to form halos around small orifices created by the edges of adjacent pillow lobes (Plate 2g), consistent with warm water rising from below, (2) the hydrothermal sediments are localized and concentrated near the crests of the mounds (Figures 2–5), which would be expected if water was heated within the mounds and rising through them, and (3) the larger mounds have larger accumulations of hydrothermal sediment, probably because they had more heat to dissipate. Similar deposits were noted by *Lichtman et al.* [1984] on young pillow lavas at the East Pacific Rise (EPR) near 21°N.

Two samples of hydrothermal sediment were obtained on *Alvin* dive 2430 and were analyzed by XRF and a scanning





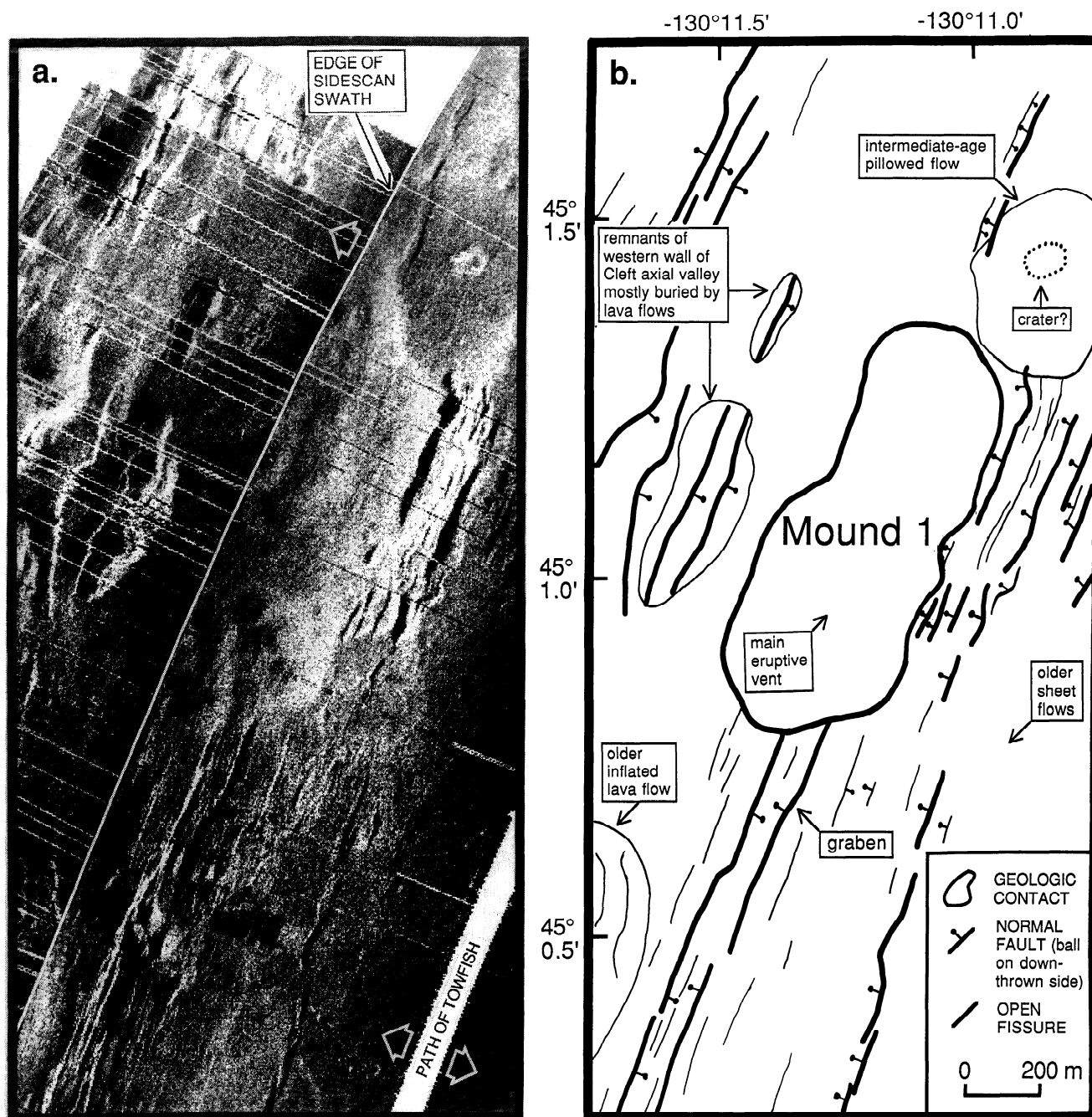
electron microscope. The samples showed a clear hydrothermal character with high concentrations of Fe, S, Cu, Zn, and Ba. Large crystals of chalcopyrite, sphalerite, and barite were identified, suggesting a local rather than a distant source for the hydrothermal sediment (R. Feely, personal communication, 1992). There is no evidence for focused, high-temperature venting on or between the pillow mounds. By 1989, the diffuse venting on the pillow mounds appeared to have stopped, except in a few places where small biological vent communities were observed.

**Biological Colonization on the New Lavas**

The north Cleft eruption provided patches of seafloor that were a “clean slate” on which to observe the process and rates of biological colonization. Small hydrothermal vent communities, including tube worms, limpets, scale worms, and ciliates (V. Tunnicliffe, personal communication, 1992), were photographed in 1989 on three of the southern four pillow mounds (Plate 2h and Figures 2, 3, and 4). Since the southernmost mound was erupted sometime between 1983 and 1987 [Fox *et al.*, 1992], this means that the vents were colonized within 2–6 years. The distribution of vent animals on the lavas is similar to that of the fine dusty hydrothermal sediment described above; they are concentrated at the margins of pillow lobes where warm water is apparently rising from below (Plates 2g and 2h). These few vent communities indicate that although hydrothermal venting had been common soon after the eruption (indicated by the hydrothermal sediment), after several years there were only a few locations where warm water was still exiting the new flows. This also shows that vent animals can colonize new vent areas quickly [Lutz *et al.*, 1992; Milligan and Tunnicliffe, this issue]. Isolated biological vent communities were also photographed on older lavas along the fissure system between the southern four pillow mounds (Figures 2–4), but it is unknown if these vents predated or postdated the eruption. No vent communities have been observed north of 45°04'N.

In contrast, the colonization of nonvent animals on new lavas is slower and therefore provides additional valuable information for attempts to make age assignments to young lava flows of unknown age. Milligan and Tunnicliffe [this issue] have quantified the abundances of sedentary (slow moving) and sessile (permanently attached) macrofauna on both the old and new lavas at north Cleft from towed camera photographs. The densities of sedentary nonvent animals on the new lavas in 1989 were an order of magnitude less than on the surrounding older seafloor. These sedentary animals were apparently actively colonizing the new flows, because they were seen in noticeably increased numbers in 1990 and 1991, especially near the edges of the flows. Sessile nonvent fauna colonize new flows even more slowly because they only spread by dispersing larvae (V. Tunnicliffe, personal commu-

**Figure 5.** (a) Map of camera tows over mounds 6, 7, and 8, bold where over new lavas. Areas of depth change between Sea Beam surveys are stippled (note mound 6 was not detected, and only southernmost part of mound 8, which continues north to ~45°10'N, is shown). All depth profiles along camera tracks with 20 times vertical exaggeration. (b) Mound 6 between X and X' in Figure 5a; (c) mounds 7 and 8 between Y and Y' in Figure 5a. All symbols as in Figure 2. Depth profiles are not available for 1991 camera tows because of depth sensor failure.



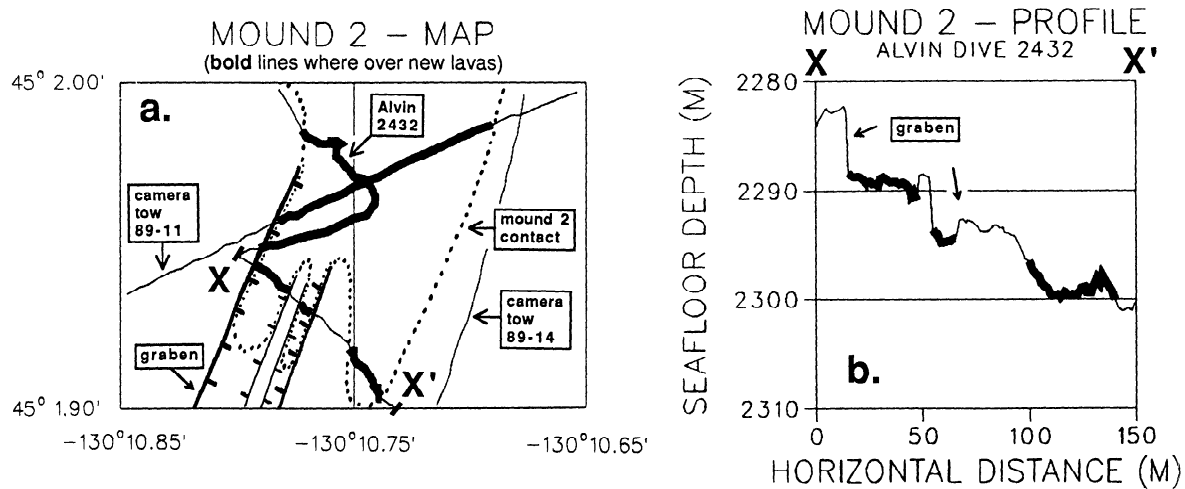
**Figure 6.** Maps of Mound 1 lava flow and surrounding structures. (a) SeaMARC I sidescan sonar mosaic constructed from two, overlapping, 2-km-wide swaths. White arrows indicate the direction of insonification from the towfish; light shades represent high backscatter and dark shades represent low backscatter or acoustic shadow. (b) Geological interpretation.

nication, 1992). Indeed, only a few individual nonvent sessile organisms were observed on the new lavas in 1989 and 1991 photos, the latter date being 4–8 years after the eruption [Milligan and Tunnicliffe, this issue]. These observations may be useful for identifying submarine lava flows in other areas that are less than 10 years old. It should be noted for future studies that many of these organisms are small (several centimeters across) and can only be detected reliably in photographs taken within about 5 m of the seafloor.

### Discussion

It is hard to identify lava flows produced during single eruptions on the seafloor, since relative ages and flow contacts are often ambiguous. The north Cleft eruption provides new

information on the character of a single eruptive event on a medium spreading rate ridge. The observations above suggest that the eruption was fed by a dike that intruded at least 17 km along the ridge axis, probably from south to north, and reached the surface discontinuously, implying that the top of the dike is not far below the surface (<1 km) between the separated eruptive vents. The interpretation that the eruption produced lavas from a single feeder, but at vents that were separated by up to 1 km (also common on subaerial volcanic rift zones), has implications for the interpretation of geochemical variability along ridge crests deduced from recovered rock samples; a single eruption can produce isolated lava flows along distances of tens of kilometers, which may be separated by older lavas of potentially different compositions. Indeed,



**Figure 7.** (a) Map of camera tow and *Alvin* dive tracks (bold where over new lavas) showing location of graben at the southern edge of mound 2 (tick marks on downthrown side of faults). Dotted line is edge of mound 2 lavas. (b) Depth profile along *Alvin* track between X and X' in Figure 7a, with 5 times vertical exaggeration. Graben has a horst in the middle and mound 2 lava covering its floor.

*Smith et al.* [this issue] show that the new pillow mounds have compositions that are significantly different than the surrounding lavas.

The eruption rate was apparently relatively uniform along the line of vents for the pillow mounds, but the duration of eruption varied markedly from vent to vent. The eruption rate was low enough that pillow lava formed exclusively and yet high enough that many pillows with smooth surface texture and fluid shapes were formed. Laboratory experiments involving the extrusion of molten wax into cool water have been used recently to quantitatively investigate the relationship between submarine lava flow morphology and eruption rate [*Fink and Griffiths*, 1990; *Griffiths and Fink*, 1993]. The wax is considered a good analog because, as it is cooled, its surface solidifies and deforms over a molten interior, similar to an underwater lava flow. The wax experiments produce a predictable sequence of flow morphologies (from "pillows," to "rifts," to "folds," to "levees") as a function of extrusion rate [*Fink and Griffiths*, 1990]. Extrapolation of these results to submarine basalts suggests that pillows should form when the extrusion rate from a point source is less than  $1 \text{ m}^3/\text{s}$  (or less than  $3 \text{ m}^2/\text{s}$  per unit length from a line source), assuming a lava viscosity of  $10^3 \text{ P}$  [*Griffiths and Fink*, 1993]. For comparison, the average eruption rate from the recently active Kupaianaha lava shield on Kilauea volcano, Hawaii was  $3.5 \text{ m}^3/\text{s}$  (J. Kauahikaua, personal communication, 1992). Since it is the relative rates of crustal thickening and lateral spreading that control the flow morphology, another way to quantify this result is as follows: in order to form pillow morphology, freshly exposed lava must travel less than 2 m before its surface cools to the solidification temperature, which takes  $\sim 0.1 \text{ s}$  [*Fink and Griffiths*, 1990]. The mixed morphology of the Cleft pillow lavas and the abundance of smooth lava forms suggests that they were erupted at a volume flux near the upper end of the theoretical limits for pillow morphology, perhaps about  $1 \text{ m}^3/\text{s}$  at point source vents.

The entire eruptive episode probably lasted between days to weeks, based on the fact that the volume of lava erupted,  $0.05 \text{ km}^3$ , is about the same as from the 1977 eruption at Kilauea volcano, Hawaii [*Moore et al.*, 1980] and about half that of the 1984 rift eruption at Krafla, Iceland [*Tryggvason*, 1986],

each of which lasted 2–3 weeks. Based on subaerial analogs such as these, we speculate that the dike intrusion that fed the eruption was accompanied by horizontal extension across the neovolcanic zone (perhaps a meter or more) and was probably accompanied by the formation of grabens near eruptive vents. This extension may have been the trigger for the release of megaplumes, huge sudden bursts of hydrothermal fluids, that were discovered in 1986 and 1987 over the southern Juan de Fuca Ridge [*Baker et al.*, 1989; *Cann and Strens*, 1989]. Thus megaplumes may be indirect indicators of seafloor volcanic events [*Chadwick et al.*, 1991; *Embley et al.*, 1991]. The 1980s intrusion/eruption may also have been the fundamental cause of other changes observed in the vent fluid chemistry of the north Cleft hydrothermal system [*Baker and Lupton*, 1990; *Butterfield and Massoth*, this issue].

Many obviously young lavas have been observed along the MOR, but until recently [*Rubin and Macdougall*, 1991; *Goldstein et al.*, 1992], it has been difficult to estimate their absolute ages, which are of interest because they help identify which portions of the ridge system have been recently active and perhaps how often. At north Cleft, we can document the physical appearance of lava flows that are 2–6 years old, and the observed rates of sedimentation and biological colonization have implications for improved qualitative dating of young lavas elsewhere. Perhaps the most important lesson from north Cleft is that newly erupted lava ages in appearance much more rapidly than previously thought, and therefore previous age estimates of young submarine lavas [*van Andel and Ballard*, 1979; *Ballard et al.*, 1981, 1982; *Lichtman et al.*, 1984; *Macdonald et al.*, 1988] are probably at least an order of magnitude too old (Table 2). Previous estimates of lava ages were crude and usually based on the degree of vitreous luster and sediment cover observed in bottom photographs. Such estimates were generally conservative upper bounds made by considering the sedimentation rate, the microtopography of the lava surface and then simply imagining how much sediment might accumulate on that surface in a given amount of time.

It seems clear now that any lava flow that is truly pristine in appearance must be extremely young, probably  $< 5$  years old, and this applies to many flows that have been observed on the Galapagos rift and the EPR. Another lesson from north

**Table 2.** Comparison of Relative Age Scales for Submarine Lavas Based on Their Appearance

Description of Lava Appearance	Scale of <i>Ballard et al.</i> [1981, 1982]	Scale of <i>Macdonald et al.</i> [1988]	Approximate Ages Assigned by <i>Macdonald et al.</i> [1988], years	Revised Age Estimates Based on This Study,* years
No sediment cover, highly vitreous luster	1.0	1.0	<50	<5
Light "peach fuzz" of sediment cover, vitreous luster	—	1.2	<100	<10
Light sediment cover, vitreous luster diminished, no sediment pockets	1.2	1.5	<500	<50
No vitreous luster, very small sediment pockets on and between pillows	1.5	1.7	<1000	<100
Sediment pockets well developed between pillows	1.7	2.0	1000–5000	no estimate
Sediment pockets deep enough to connect between pillows	2.0	3.0	5000–20000	no estimate
Total sediment cover with only isolated pillow outcrops	2.5+	4.0	>20000	no estimate

\* Note these revised age estimates are for areas with pelagic sediment only. In areas with deposits from diffuse hydrothermal venting, even lavas with sediment pockets deep enough to interconnect (age scale 2.0 or 3.0) could be <5 years old. Published sedimentation rates near the southern East Pacific Rise studied by *Macdonald et al.* [1988] and near the southern Juan de Fuca Ridge are similar [*Carson, 1971; McMurtry et al., 1981*], about 1–2 cm/1000 years.

Cleft is that sediment accumulation can be quite variable on a single young flow, ranging from no sediment cover to interconnected pockets between pillows, depending on lava surface texture and the extent of local, diffuse hydrothermal venting. Hydrothermal sediment tends to mask the glassy surfaces of fresh lava, and it is sometimes difficult to distinguish from pelagic sediment in photographs, especially once a dusting of pelagic sediment covers the hydrothermal deposits. This could lead to one flow being assigned a wide range of relative ages depending on where it was observed. Therefore relative age classifications should be made with caution and must be assigned on the basis of a large photographic database that allows one to distinguish local anomalous accumulations from average background sedimentation.

It is interesting to note that similar observations of rapid sediment deposition on young lavas have been made on the EPR at 9°45–52'N, where a more recent eruption has been documented [*Haymon et al., 1991, 1992, 1993*]. *Alvin* visited the site in April 1991, apparently immediately after the eruption and found pristine lava flows, widespread hydrothermal venting, and extensive bacterial mats [*Haymon et al., 1991, 1993*]. However, during an *Alvin* dive series a year later (March 1992) most of the pervasive venting had already ceased, juvenile vent animals had already colonized some of the remaining hydrothermal sites, and the lava flows looked considerably older, mainly due to the deposition of hydrothermal sediment and palagonitization of volcanic glass [*Haymon et al., 1992, 1993*]. The observations at this site on the EPR reflect a different time window on a similar process, illustrating the rapid and variable rates of change of apparent age indicators on young MOR lavas.

Vent organisms seem to be able to colonize new flows very rapidly if suitable hydrothermal sites are created [*Lutz et al., 1992; Milligan and Tunnicliffe, this issue*]. However, this study shows that new biological communities may not last longer than a few years if their heat source is limited to the cooling lava flow itself or a thin underlying feeder dike. The coloniza-

tion of nonvent animals occurs at slower rates and so may be useful for making lava age estimates years or possibly even decades after an eruption. Quantitative mapping at north Cleft shows that the density of sedentary nonvent animals was still an order of magnitude lower on the new lavas than on the surrounding seafloor 2–6 years after the eruption, and sessile nonvent animals were just starting to colonize after 4–8 years [*Milligan and Tunnicliffe, this issue*]. These observations should be continued in future years to further document the rate of biological colonization on the new flows.

How does the north Cleft eruption fit into the context of activity along the global MOR system? Although this was the first documented eruption on the deep MOR, it must have been only one of many such events that occur each year. The frequency of eruptions on the MOR is presently unknown, but *Crisp* [1984] estimated that 75% of the new magma reaching Earth's surface each year is emplaced at the MOR crest, on average ~3 km<sup>3</sup>/yr extrusive and ~18 km<sup>3</sup>/yr intrusive. The north Cleft eruption was thus ~2% of the average annual global volcanic output along the MOR. Following the estimates of *Crisp* [1984], the average extrusive output along the whole Cleft segment is 0.003 km<sup>3</sup>/yr, assuming a full spreading rate of 6 cm/yr, and an extrusive layer of ocean crust 1 km thick. Therefore the volume of the new pillow flows represents 16 years of the average annual volcanic output for the whole Cleft segment.

Clearly, many other eruptions must be occurring every year on the MOR system; the challenge is to devise new techniques to detect them. The quantitative method of comparing repeated multibeam sonar surveys that was used to document the north Cleft eruption [*Fox et al., 1992*] could be used to search for other areas of recent or on-going volcanism anywhere on the MOR system where a baseline survey has already been made. To be detected by this method, new lava flows must be at least 5–15 m thick and 200–300 m in diameter [*Fox et al., 1992*]. Pillow mounds like those described in this paper, are most likely to form at intermediate and slow spreading ridges.

Nevertheless, fast spreading ridges are likely to have more frequent and larger volume eruptions which, even if they produce fluid sheet flows, could build significant new morphological features. This is most likely to occur where a sustained fissure eruption becomes localized at a few discrete vents. For example, Kupaianaha lava shield on Kilauea volcano, Hawaii, was built entirely of fluid pahoehoe lava and is 56 m high and 1 km in diameter, and its lava flows on the south coast of the island are 25 m thick [Heliker and Wright, 1991]. Systematic bathymetric resurveying could help reveal the locations and frequency of significant eruptions along the MOR.

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