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Distribution of recent volcanism and the morphology of seamounts and ridges in the GLIMPSE study area: Implications for the lithospheric cracking hypothesis for the origin of intraplate, non–hot spot volcanic chains

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[1] Lithospheric cracking by remotely applied stresses or thermoelastic stresses has been suggested to be the mechanism responsible for the formation of intraplate volcanic ridges in the Pacific that clearly do not form above fixed hot spots. As part of the Gravity Lineations Intraplate Melting Petrology and Seismic Expedition (GLIMPSE) project designed to investigate the origin of these features, we have mapped two volcanic chains that are actively forming to the west of the East Pacific Rise using multibeam echo sounding and side-scan sonar. Side-scan sonar reveals the distribution of rough seafloor corresponding to recent, unsedimented lava flows. In the Hotu Matua volcanic complex, recent flows and volcanic edifices are distributed over a region 450 km long and up to 65 km wide, with an apparent, irregular age progression from older flows in the west to younger in the east. The 550-km-long Southern Cross Seamount/Sojourn Ridge/Brown Ridge chain appears to have been recently active only at its eastern end near the East Pacific Rise. A third region of recent flows is found 120 km north of Southern Cross Seamount in seafloor approximately 9 Myr old. No indication of lithospheric extension in the form of faulting or graben formation paralleling the trend of the volcanic chains is found in the vicinity of recent flows or anywhere else in the study area. Thermoelastic cracking could be a factor in the formation of a few small, very narrow volcanic ridges, but most of the volcanic activity is broadly distributed in wide swaths with no indication of formation along narrow cracks. The Sojourn and Brown chains appear to begin as distributed zones of small seamounts that later develop into segmented ridges, perhaps under the influence of membrane stresses from self-loading. We suggest that the linear volcanic chains are created by moving melting anomalies in the asthenosphere and that lithospheric cracking plays at most a secondary role.

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1. Introduction

[2] One hypothesis for the formation of oceanic, intraplate, volcanic chains like the Puka Puka ridges is that they develop where tectonic cracks in the lithosphere allow asthenospheric melt to escape to the seafloor. The Puka Puka chain comprises numerous, individual en echelon ridges and in total extends for more than 2000 km from near the Tuamotu islands in the west eastward to near the East Pacific Rise, where it broadens into the Rano Rahi seamount field (Figure 1). It is aligned approximately in the direction of motion of the Pacific plate in the hot spot reference frame and may lie in a low of a cross-grain gravity lineation. Age determinations for the basalts indicate that it could not have formed by motion of the Pacific plate over a fixed hot spot [*Sandwell et al.*, 1995]. Two suggestions have been advanced for the origin of the hypothesized tectonic cracking: lithospheric stretching and boudinage [*Sandwell and Dunbar*, 1988; *Sandwell et al.*, 1995] and thermal contraction of the plate [*Gans et al.*, 2003; *Sandwell and Fialko*, 2004].

[3] The stresses required for lithospheric stretching could possibly be transmitted in the plate from the remote pull of subducting slabs [*Sandwell et al.*, 1995], although the stretching mechanism has been challenged on the grounds that at least 10% extension is required to initiate boudinage [*Dunbar and Sandwell*, 1988] and the expected increase in separation between preexisting fracture zone traces has not been found [*Goodwillie and Parsons*, 1992; *Gans et al.*,

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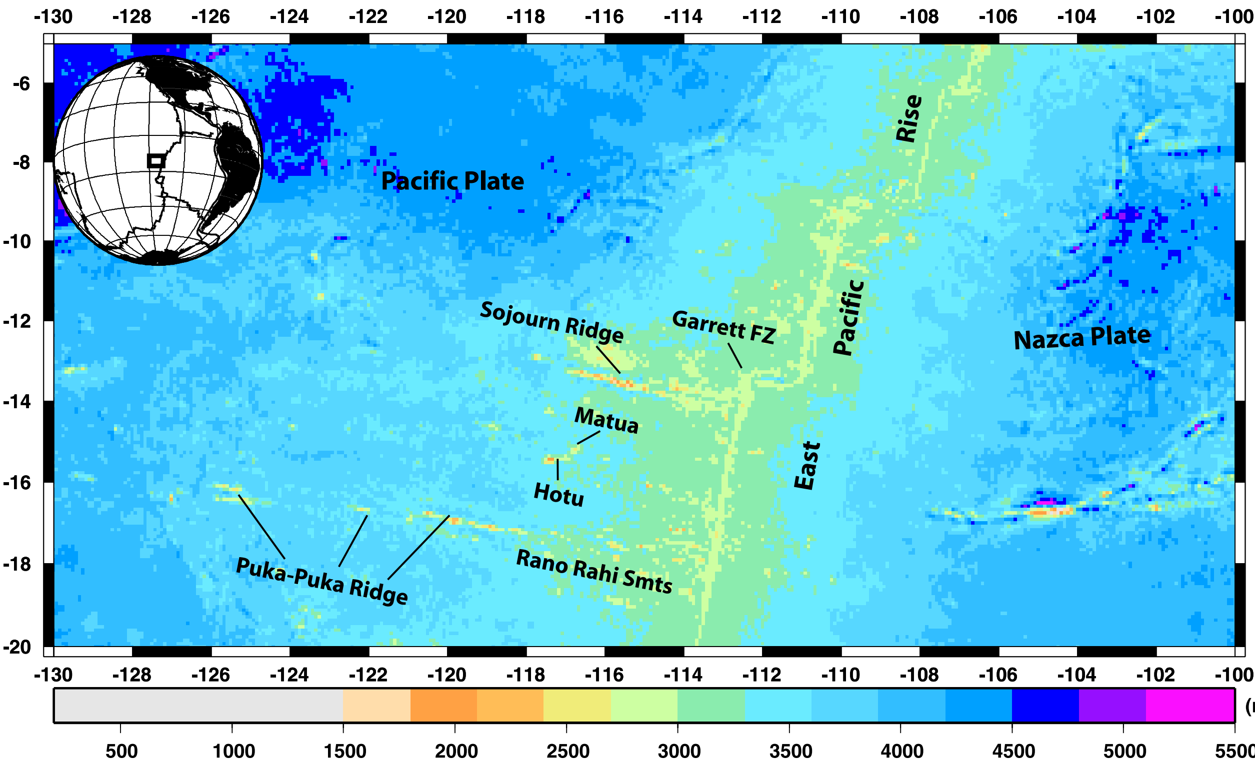


Figure 1. Predicted bathymetry of the East Pacific Rise in the vicinity of the GLIMPSE study area based on satellite altimetry [*Smith and Sandwell*, 1997]. The broad shallow area immediately north of the Sojourn Ridge at about 116°W does not exist (see Figure 2); it is an artifact stemming from a geoid or gravity anomaly that does not have corresponding bathymetric expression. The inset map of the globe shows the location of the GLIMPSE study area in the box, with heavy black lines indicating plate boundaries.

2003]. *Gans et al.* [2003] suggested instead that cracking of the lithosphere is caused by bending of the plate under thermal stresses associated with cooling of the lithosphere. *Sandwell and Fialko* [2004] showed that thermal stresses would lead to a preferred wavelength of bending and cracking, which could reproduce the initial 150 to 200 km wavelength of cross-grain gravity lineations [*Haxby and Weissel*, 1986]. Both of these variations on the tectonic cracking hypothesis are passive models in which the volcanic ridges form in response to cracking; either the crack lets preexisting asthenospheric melt escape to the surface or the small amount of local extension induces some upwelling that causes pressure release melting.

[4] Alternative models for the origin of the volcanic ridges involve anomalous melting in the asthenosphere associated with small-scale convection, minihot spots, or return flow of the asthenosphere to the East Pacific Rise that incorporates compositional and/or thermal anomalies within the returning flow field. In these models, the volcanic ridges are caused by active processes beneath the lithosphere that create temperature variations and pressure-release melting. The ridge-like nature of the individual edifices could be attributed to dikes propagating under the influence of a remotely applied stress field or under the stresses associated with loading of the lithosphere by a linear chain of seamounts [*Hieronymus and Bercovici*, 2000].

[5] One way to distinguish between lithospheric cracking and asthenospheric models for the origin of the volcanic lineations is to examine in detail the morphology of the seafloor and the distribution of eruptive centers and lava flows. One could argue that the evidence for cracks preceding volcanism has been obscured by subsequent burial beneath the built-up edifices, but if we look at the process in the early stages of formation, we should be able to recognize which type of model is more likely by searching for faults or linear features and by establishing whether the initial volcanic activity is localized or widespread. The Puka Puka chain is now a fossil feature, except perhaps for some scattered volcanism in the Rano Rahi seamount field [*Scheirer et al.*, 1996a, 1996b], but we have recently mapped two other neighboring, linear volcanic features that are still actively forming to the west of the East Pacific Rise.

[6] This paper describes the distribution of recent intraplate volcanism and the morphology of the volcanic features in the Gravity Lineations Intraplate Melting Petrology and Seismic Expedition (GLIMPSE) study area, with an emphasis on distinguishing between lithospheric cracking and asthenospheric origins for the volcanic activity. In this area just to the north of the Rano Rahi seamount field (Figure 2), young lava flows and the Hotu and Matua seamounts were first discovered in the early 1990s during mapping in preparation for the Mantle Electromagnetic and Tomography (MELT) Experiment. The existence of the Sojourn Ridge was first noted on high-resolution images of satellite free-air gravity anomalies [*Smith and Sandwell*, 1997], and it was partially mapped from shipboard for the first time in

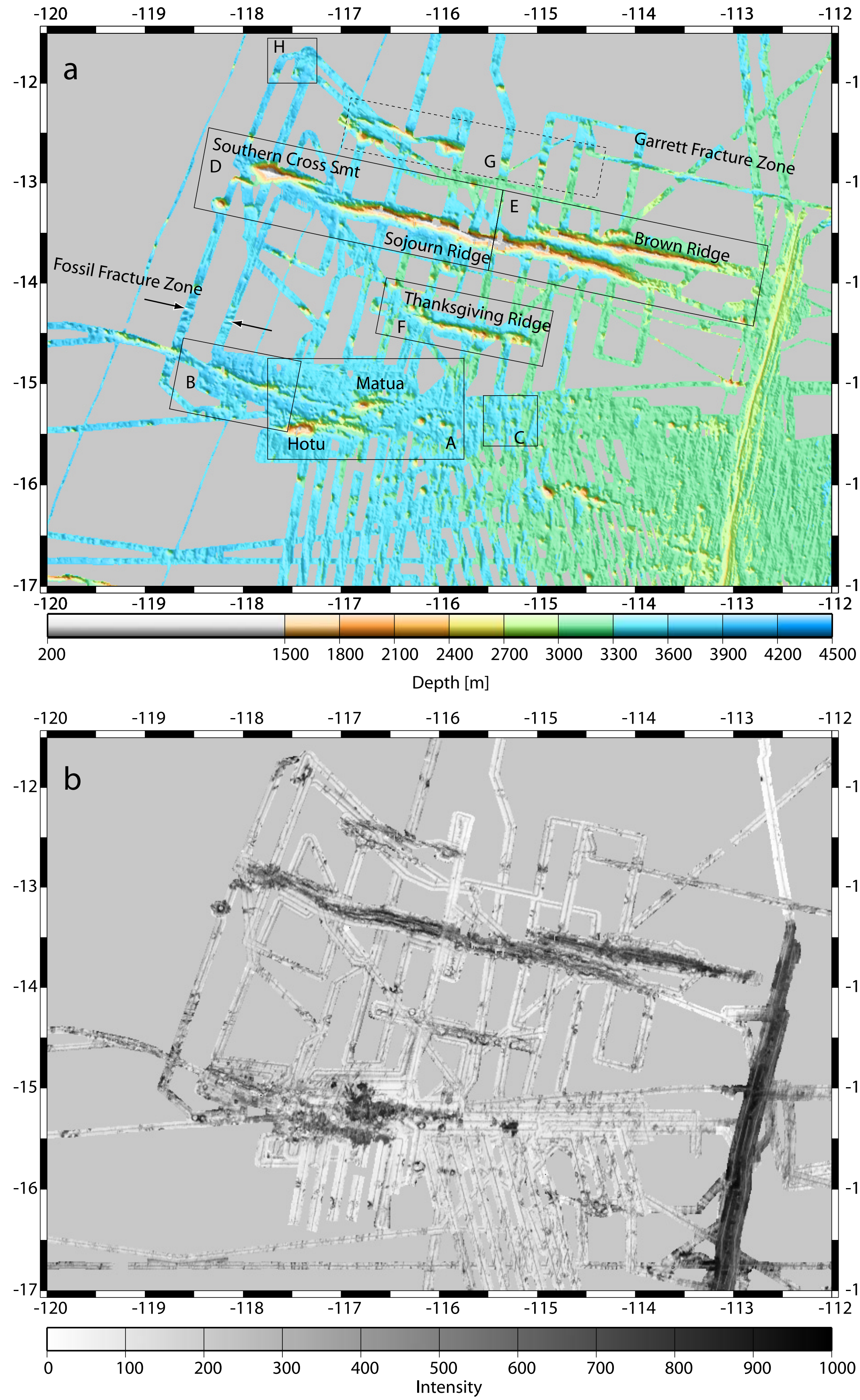

a
H
G
Garrett Fracture Zone
Southern Cross Smt
D
E
Sojourn Ridge
Brown Ridge
Fossil Fracture Zone
Thanksgiving Ridge
F
B
Matua
Hotu
A
C
Depth [m]
b
Intensity

Figure 2

1996. The Sojourn Ridge/Brown Ridge/Southern Cross Seamount volcanic lineament extends about 550 km perpendicular to the East Pacific Rise and is the most prominent volcanic feature on the Pacific plate between the equator and the seamount chain at ∼25°S created by the Easter hot spot. The GLIMPSE Experiment in 2001 and 2002 investigated the origin of these volcanic features using underway geophysical techniques (gravity, magnetic anomalies, bathymetry and side-scan sonar), seismic reflection and refraction, passive seismic tomography, rock sampling and geochemical, petrological, and geochronological analysis. In this paper, we use the bathymetry and side-scan sonar data to demonstrate that there is no indication of tectonic extension or cracking preceding volcanism.

2. Data Acquisition and Processing

[7] The foundation of our maps and interpretations is SeaBeam2000 multibeam bathymetry and side-scan sonar data gathered on the R/V *Melville* during cruises VANC04MV (2002), COOK16MV (2001), SOJN01MV (1996), BMRG01MV (1995), GLOR03MV (1993), and GLOR02MV(1992). In addition, we incorporated bathymetry data from Sonne Cruise 145-1 [*Grevemeyer et al.*, 2002] and other, older cruises, as compiled by *Scheirer et al.* [1996a, 1996b, 1998]. During the 2001 and 2002 GLIMPSE cruises, we were guided by satellite free-air gravity anomaly and predicted bathymetry maps [*Sandwell and Smith*, 1997; *Smith and Sandwell*, 1997] and by prior multibeam coverage. Ship tracks were designed to cross virtually every identifiable seamount within the study area, generally following chains trending approximately perpendicular to the East Pacific Rise. Cross track lines parallel to the spreading center provided views of nonseamount seafloor characterized by "normal", abyssal hill morphology, but normal seafloor was undersampled relative to the seamount-dominated areas (Figure 2). Few tracks were oriented in optimal locations or directions for identifying magnetic anomalies created at the spreading center. Navigation of all of the *Melville* tracks employed P-code Global Positioning System (GPS) data, so no track adjustments were needed to reconcile bathymetry on crossing tracks.

[8] SeaBeam2000 bathymetry nominally covers a 120° cross-track angle yielding a swath ∼3.4 times as wide as the water depth. The entire swath could usually be used in areas of rough topography or unsedimented seafloor, but in flatter, sedimented areas where the returns from the outer beams were weaker, we often had to eliminate the outer ∼20% of the swath to avoid artifacts. Crossing tracks were useful in identifying systematic artifacts, but a large percentage of the artifacts were automatically eliminated using a despiking algorithm [*Scheirer et al.*, 2000]. The SeaBeam2000 flagging system for bad data was not employed. After despiking, all swaths were hand edited for any remaining outliers or obvious artifacts. The quality of the data was generally high, but was dependent on sea state, ship heading, and, probably, degree of fouling of the ship's bottom. With careful editing and corrections for cross-track bias, roll and yaw, the system is capable of <2 m bathymetry precision for the inner 2/3 of the beams [*Dunn et al.*, 2001], but as routinely operated in this study, the standard errors of individual beam depths is probably on the order of 5 to 10 m. Bathymetry was compiled on a 200-m grid. All beams within 200-m square bins were averaged without regard to quality other than the editing process described above. In very detailed maps based on this gridded data set, some small, along-track artifacts are still visible where cross-track biases were imperfectly removed or where there are slight offsets in estimates of the absolute values of depths between cruises due to changes in velocity of the water column.

[9] Side-scan sonar data are particularly useful for detecting areas of recent intraplate volcanism. High side-scan backscatter is associated both with slopes facing the ship, such as fault scarps or seamount flanks and with areas of seafloor that are rough at a length scale similar to the sonar wavelength of ∼10 cm. The spreading center itself is highly reflective, but as it becomes covered with a veneer of sediment, the backscattered energy decreases (Figure 2b). There is a strong correlation of reflectivity with time since the last lava flow, but the reflectivity cannot be precisely calibrated in terms of age because it also depends on sedimentation rate and the initial roughness of the lava flow. Nevertheless, our experience is that we can recover fresh, glassy basalts by dredging the areas that are most strongly reflective, while less reflective areas yield sediment-encrusted samples or come up with empty dredges. The steep slopes of seamounts or volcanic ridges are also rough and reflective, but can usually be distinguished from fresh flows by a patchier side-scan appearance that is more dependent on viewing angle than the side-scan character of flows, which tends to be uniformly reflective from all viewing angles.

[10] Because the side-scan data are dependent on viewing angle relative to seafloor slopes and somewhat on sonar acquisition parameters, such as gain settings and pulse length, that differ from cruise to cruise, we use a different approach to generate maps of side-scan intensity than we use for bathymetry. We employ an equalizing algorithm [*Scheirer et al.*, 2000] to reduce the systematic variations in amplitude across the swath, although the near-vertical specular returns are always distinguishable in a swath even after this compensation. We also adjust the side-scan levels to obtain approximate agreement between cruises crossing

Figure 2. (a) Bathymetry of the GLIMPSE study region mapped using multibeam echo sounding. The major bathymetric features in the region are labeled. The boxed regions indicate areas of interest discussed in the text; all but G are shown in more detail in subsequent figures. Topography is illuminated from the south-southwest. (b) Seafloor side-scan reflectivity of the GLIMPSE study region mapped using the SeaBeam2000 system on the R/V *Melville*. The areas of high intensity of acoustic backscattering (dark) are interpreted to be regions of young, fresh, unsedimented basalt in flat areas and seafloor slopes on the flanks of seamounts and ridges. Note the high reflectivity of young seafloor adjacent to the East Pacific Rise axis. The regions surrounding Matua Seamount and the Brown Ridge also have high intensity despite modest topographic relief, indicating that they are in an active volcanism stage.

Table 1. The ^{40}Ar-^{39}Ar Incremental Heating Plateau Ages[a]

Sample	Latitude	Longitude	K_2O, %	Age, Ma	2 SD, Ma	^{39}Ar, %	N	MSWD	Plate Age	Age Difference
D3 −1	−14.970	−116.794	0.81	3.28	0.39	67	4/6	1.86	6.1	2.8
D4 −1	−15.240	−116.903	0.54	6.13	0.32	100	6/6	0.76	6.1	0.0
D6 −3	−15.349	−117.217	1.29	0.61	0.06	87	4/6	2.15	6.5	5.9
D6 −2	−15.349	−117.217	1.29	0.28	0.12	100	5/5	0.34	6.5	6.3
D7 −1	−15.235	−116.750	1.89	0.29	0.10	100	6/6	0.19	5.9	5.6
D9 −1	−15.293	−116.034	0.89	0.34	0.11	100	5/5	0.54	4.8	4.5
13 −1	−13.457	−115.961	0.73	2.13	0.38	100	5/5	0.06	5.4	3.3
15 −2	−15.486	−116.777	1.02	0.82	0.08	85	4/6	0.34	5.8	5.0
15 −1	−15.486	−116.777	1.00	0.79	0.13	100	5/5	0.13	5.8	5.0
17 −1	−15.429	−116.911	4.25	0.16	0.01	80	4/7	0.86	6.1	5.9
19 −1	−15.386	−117.161	0.65	1.36	0.17	46	3/6	0.62	6.4	5.1
31 −4	−13.594	−115.324	0.67	1.62	0.12	92	4/6	0.74	4.4	2.8
38 −2	−13.861	−113.203	1.01	0.32	0.13	100	5/5	1.49	1.1	0.8
39 −2	−13.521	−114.763	1.30	0.63	0.04	100	5/5	0.52	3.6	2.9
58 −2	−15.020	−117.807	2.04	2.11	0.05	64	4/6	0.55	7.6	5.4
61 −1	−13.388	−116.197	0.50	1.04	0.22	96	4/5	2.10	5.8	4.8

[a]Ages calculated using decay constants of *Steiger and Jaeger* [1977]; ^{39}Ar is percent of total gas released represented by the plateau; N is number of heating steps in plateau/total number of steps; MSWD is mean square of weighted deviations, an F statistic with critical value for significance ~2.5. Expected plate ages are interpolated from *Müller et al.* [1997].

the same area, using different cross-axis equalizations, although this adjustment is also imperfect. In addition, noise levels vary within and among cruises, depending on sea state and other environmental conditions. We construct maps from 200-m grids as with bathymetry, but instead of averaging all intensity values within a bin, we select a single value based on a priority table for swaths, established on the basis of signal-to-noise ratio, favorable azimuth for illuminating features, and compatibility with adjacent swaths.

[11] Rock samples were collected both by dredge and wax (glass) core on many of the seamounts and their surroundings, although wax coring proved to be relatively unsuccessful. Radiometric ages of whole rock basalt samples obtained by dredging were determined using the ^{40}Ar–^{39}Ar incremental heating method [*Dalrymple et al.*, 1981; *Duncan and Hogan*, 1994]. Well-crystallized portions of rocks, away from glassy rims, were chosen to avoid contamination from trapped, mantle-derived ("excess") Ar. After neutron irradiation in the Oregon State University TRIGA reactor facility, argon was extracted from samples in a series of heating steps and isotopes were measured by mass spectrometry. Ages are calculated from the isotopic composition of Ar released in each step. Plateau ages and uncertainties (Table 1) are calculated from the weighted mean and scatter of concordant step ages.

[12] The incremental heating method effectively separates atmospheric ^{40}Ar, dominant in low-temperature steps, from radiogenic ^{40}Ar, dominant in midtemperature to high-temperature steps. The higher proportions of radiogenic ^{40}Ar lead to higher precisions on the measured ages. Isochrons, derived from the plateau-forming step isotopic compositions (i.e., ^{40}Ar/^{36}Ar versus ^{39}Ar/^{36}Ar), yield estimates of the initial composition of Ar in the sample that reveal whether excess ^{40}Ar is a significant contaminant. Details of the experimental procedure are given by *Duncan and Hogan* [1994], *Duncan et al.* [1997], and *Sandwell et al.* [1995]. Despite the improvements of the ^{40}Ar-^{39}Ar incremental heating method over conventional K-Ar dating it can be difficult to determine useful and reliable dates for young samples with low K content because the concentration of radiogenic ^{40}Ar is so low. In this study, we found increasing formal uncertainties with lower K content, in extreme cases larger than the measured ages. For this reason, the results reported in Table 1 are restricted to samples with $K_2O > 0.5\%$ and formal errors (2 standard deviations) less than 0.4 Myr. Of the 27 sample analyses, 8 were rejected based on both criteria, 2 were rejected because of low K_2O, and 1 was rejected because of excessive error.

3. Hotu Matua Volcanic Province

[13] The Hotu Matua volcanic province is a complex group of conical and flat-topped seamounts, knife-like narrow ridges, botroidal mounds of rounded cones, and recent lava flows with little topographic expression (Figures 2, 3, 4, and 5). Altogether, the complex extends for at least 450 km from 119°30′W to 115°15′W. At its widest point, near 116°45′W, relatively recent flows and seamounts extend 65 km from north to south. In general, there is an increase in side-scan reflectivity from west to east indicating that the volcanism has propagated eastward with the most recent flows near the eastern end.

[14] The youth of some of the features is indicated by the dredging recovery of fresh, glassy basalt with no indication of manganese encrustation or sedimentation and by the Ar-Ar dating of samples as young as 160 ka. In addition, an array of ocean bottom seismometers deployed as part of the GLIMPSE experiment for a period of 4 weeks in 2001 recorded a number of microearthquakes just north of Matua, indicating that volcanic activity may be continuing today [*Llenos et al.*, 2003]. The general age progression from west to east is not strictly linear, as indicated by the presence of very reflective seafloor scattered along a line from Matua eastward 180 km to about 115°15′W and by scattered radiometric ages (Table 1 and Figure 3).

[15] Matua and Hotu seamounts are the tallest volcanic edifices in the complex. Matua is dominated by a single, steep-sided, crudely conical feature standing more than 2000 m above the surrounding seafloor (Figure 3). A ridge extends eastward for ~30 km from the center of Matua, gradually decreasing in elevation. Although there is no continuous topographic feature, reflective seafloor contin-

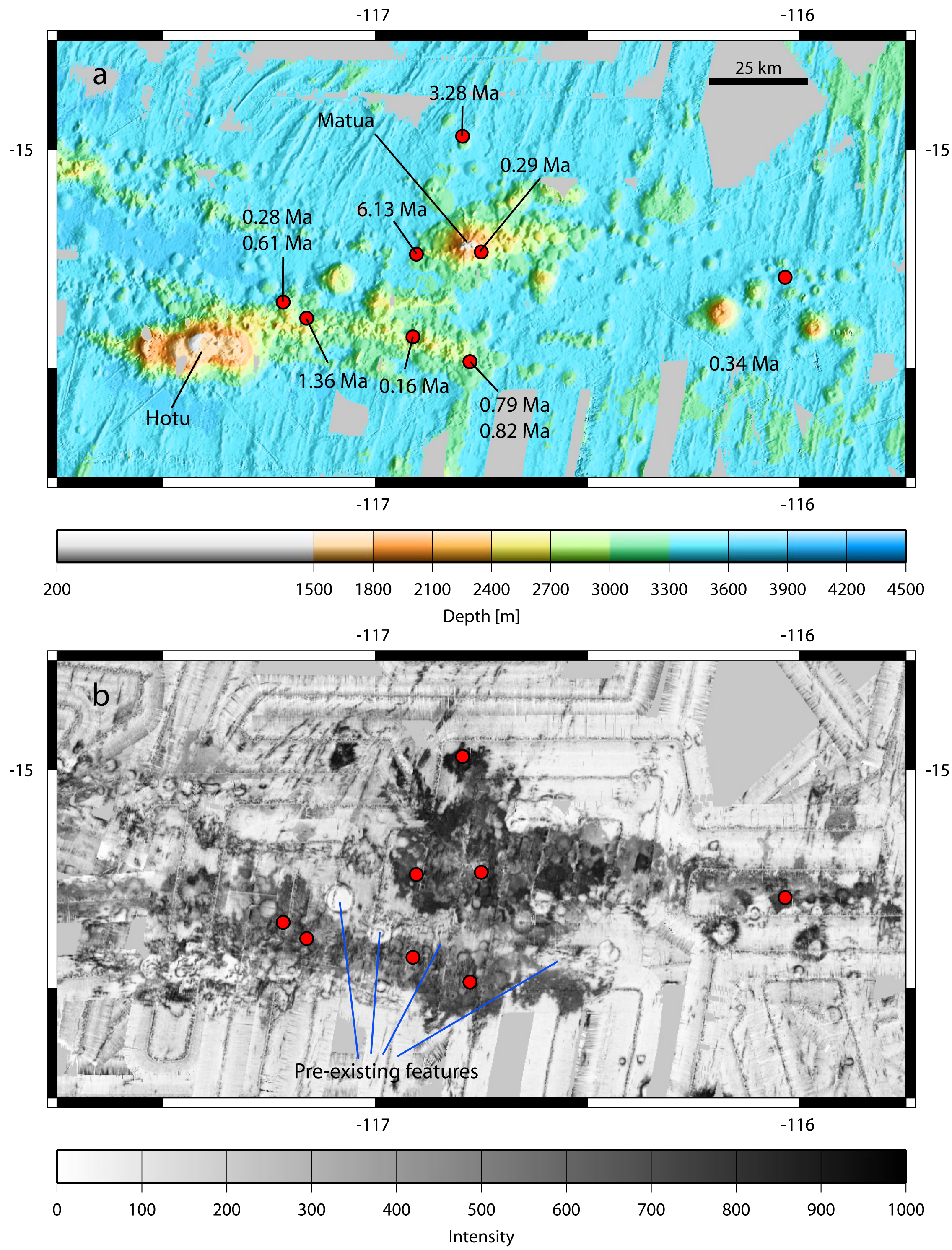


Figure 3. (a) High-resolution bathymetry and (b) side-scan sonar images of Hotu Matua region(box A in Figure 2). Matua is dominated by a single conical feature, while Hotu is a coalescence of at least four flat-topped volcanic centers. The most reflective patches representing the roughest and presumably youngest seafloor are scattered around and to the east of Matua. An example of botroidal topography is found to the east of the main edifice of Matua. Blue lines indicate older, less reflective, bathymetric features that separate the Hotu and Matua recent flows and which probably formed nearer the East Pacific Rise (EPR); they are labeled as preexisting features. Dredge locations with Ar/Ar ages are shown as red dots with age determinations indicated. In this and subsequent detailed images, pixel size is 200 m. Artificial illumination of the bathymetry is from the east.

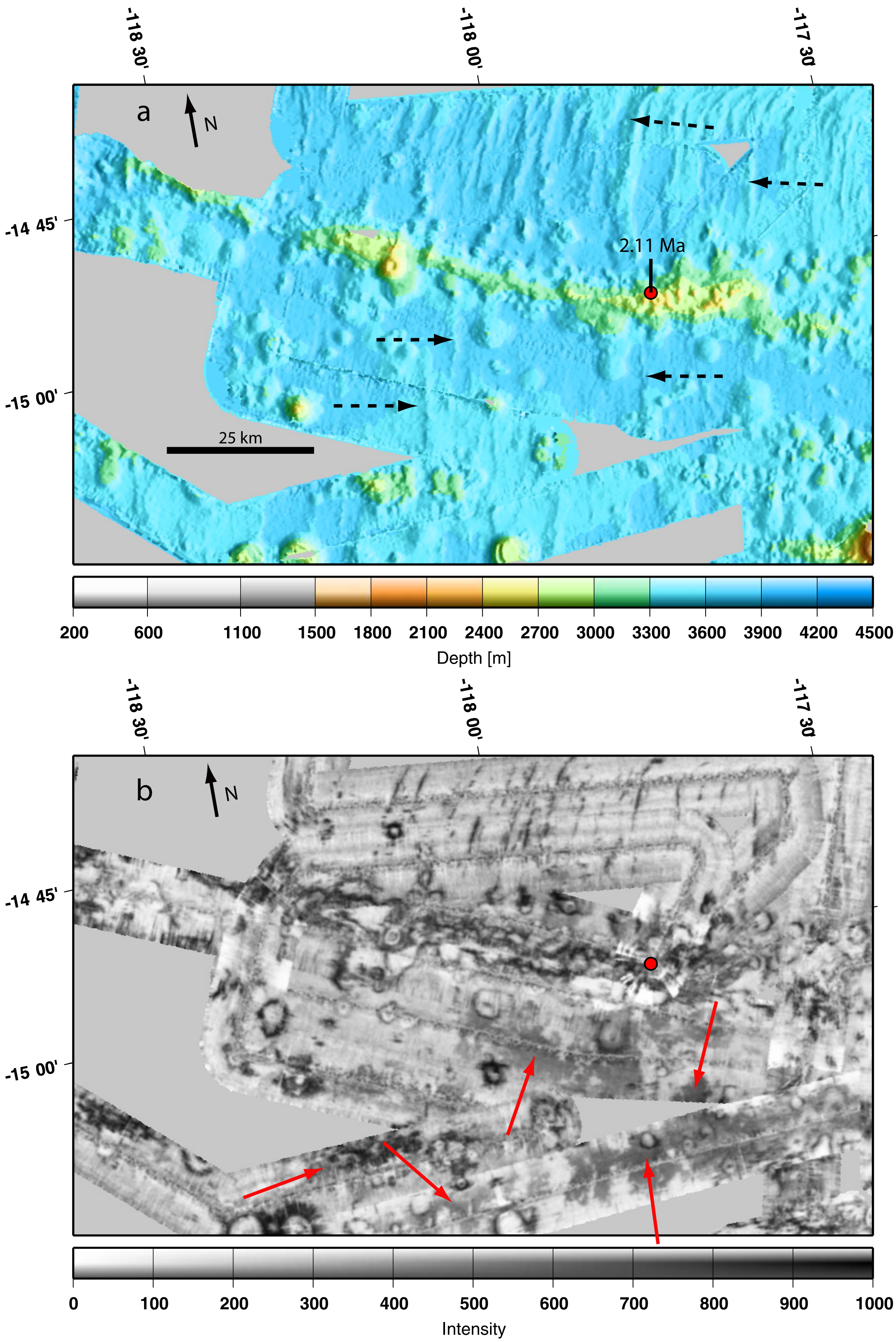


Figure 4. Detailed (a) bathymetry and (b) side-scan sonar images of region north and west of Hotu showing narrow volcanic ridge that could originate from a lithospheric crack or a dike propagating away from small volcanic centers (box B, Figure 2). Dashed black arrows point to examples of abyssal hills. South of the ridge, much of the abyssal hill fabric has been obscured by numerous rough and smooth seamounts and by lava flows. The red arrows indicate young flows in the region that have been partially covered by sediment. While the existence of the narrow knife-like ridge suggests a lithospheric crack, the widespread resurfacing of the seafloor topography and widely dispersed volcanism is inconsistent with narrow focused volcanism predicted by a fissure. Artificial illumination of the bathymetry is from the east.

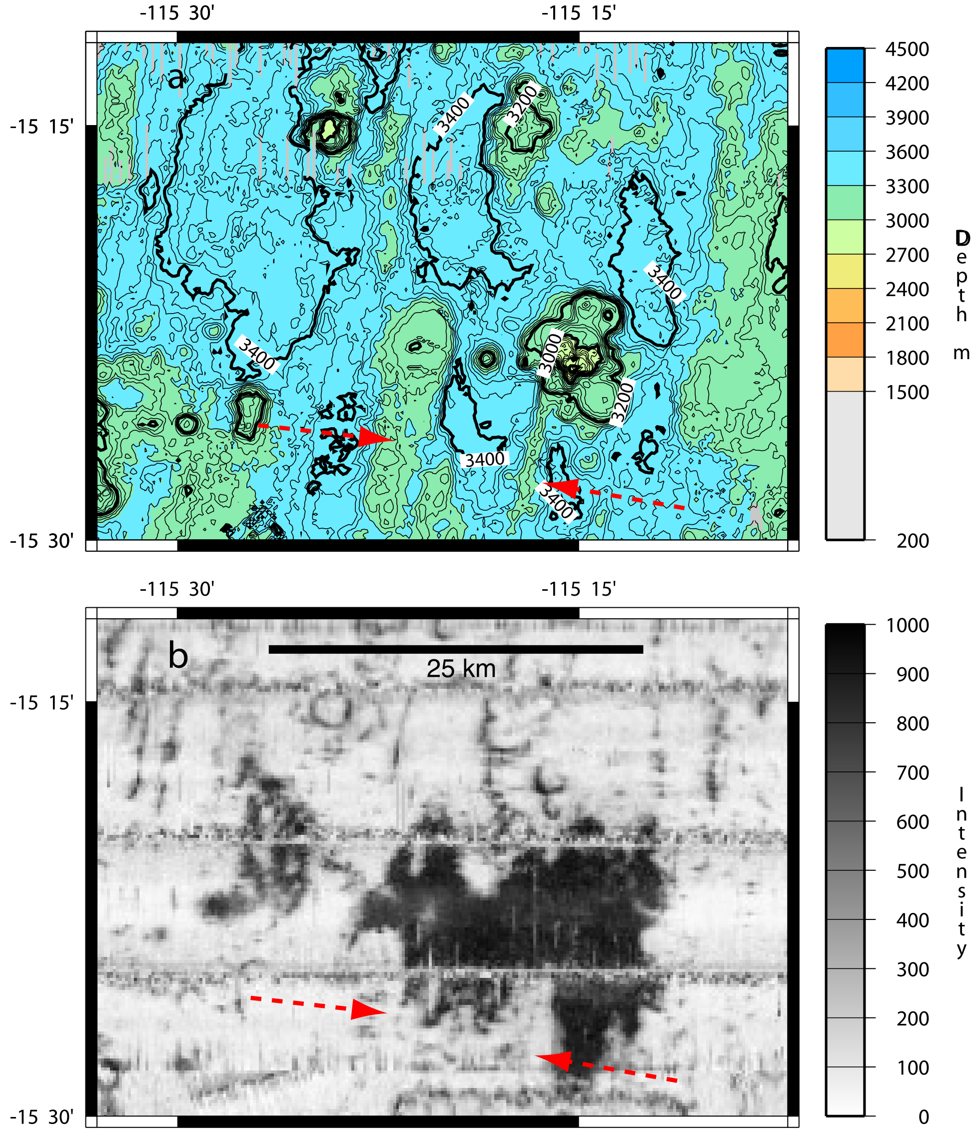


Figure 5. Detailed (a) bathymetry and (b) side-scan sonar images of an isolated area of recent volcanic activity near at the eastern end of the Hotu Matua trend (box F, Figure 2). To show finer details, bathymetry is contoured at 25 m interval instead of illuminated as in the other figures. Dashed red arrows indicate older abyssal hill topography. The strong reflectivity of the seafloor in this region is an indication that the seamounts and surrounding flows in the center of the figure are very young. However, east and west of the recent flows, there is no indication of a normal fault scarp or a graben oriented perpendicular to the spreading center or along the Hotu Matua trend that would indicate that extension of the lithosphere occurred parallel to the EPR preceding volcanism.

ues along the same line for another 80 km, indicating resurfacing by lava flows and small seamounts. The easternmost reflective patch in the complex also lies along this line, with a gap of about 30 km between it and the end of the Matua reflective line. There are a few smaller reflective areas scattered 10 to 20 km away from the main reflective line, such as the one centered at about 15°08′S, 115°54′W. Some of the most reflective flows and the earthquake activity are scattered in a broad region extending 25 to 30 km to the north of the peak of Matua. The highly reflective flows, fresh glassy basalts, a very young radiometric age (290 ± 100 ka), and microseismicity indicate that Matua (and surroundings) is the most active feature in the entire study area.

[16] Two dredges from near Matua yielded age determinations substantially older than others in this area and much older than expected from the seafloor side-scan reflectivity (Figure 3). One of the ages, 6.13 Ma, is almost exactly the expected age of the plate at this location based on magnetic isochrons. Although dredged from a high-reflectivity area, we may have dredged a scarp of the underlying seafloor. The new flows on the flanks of Matua may be relatively thin and thus may not fully cover preexisting bathymetric features. The second unexpectedly old sample (3.28 ± 0.39 Ma), 3–1 in Table 1, is from a small conical seamount

within another reflective area that could have formed off axis at an earlier date. We are somewhat skeptical about the reliability of the date of this second sample, however. It was one of our youngest appearing samples, with fresh glass at the surface, no palagonite or manganese coating or indication of sediment, it was dredged from one of the most reflective areas, it has relatively low K_2O and the largest formal error of any of the plateau ages that passed our selection criteria. In retrospect, perhaps our sampling strategy should have concentrated more on the rough, nearly flat flows than on the traditional dredge targets with relatively steep slopes.

[17] In contrast to Matua, Hotu is an amalgamation of four distinct, flat-topped volcanic centers (Figure 3) that currently appear to be inactive. The flat tops are presumably caused by a combination of caldera collapse and outward building of the volcanoes due to limits on the maximum height to which magma can rise from a chamber, as they are too deep (~1500 m) to be submerged atolls. No fresh basalts were dredged from Hotu. To the south of Hotu, abyssal hill morphology is recognizable in the lightly sedimented seafloor that deepens into a flexural moat flanking Hotu and there is no indication of recent flows or volcanic resurfacing of the seafloor (Figure 3). On the north side, a band of reflective seafloor extends tens of km both east and west of Hotu, from which rocks dated as young as 160 ± 10 ka were dredged. This band forms an en echelon pattern with the Matua reflective band, but none of this seafloor is as reflective as the more reflective side-scan patches within the Matua trend; thus the most recent eruptions in this band are probably somewhat older than in the Matua band. There is no obvious age progression within this band based on radiometric dates, and no successful age determinations were obtained for samples from Hotu itself. On the basis of the side-scan reflectivity and freshness of the samples, we believe that Hotu and some other smaller edifices (marked in Figure 3) are older features that were formed on seafloor closer to the East Pacific Rise axis and are unrelated to the current episode of volcanism that has created the numerous small cones in the reflective band. In some places, more recent flows seem to have dammed against these older features.

[18] To the west of Hotu and Matua there is a broad area, about 50 km across, that has been largely resurfaced by off-axis volcanism covering most of the abyssal hill topography (Figure 4). There are numerous small cones and other constructional features, as well as recognizable individual lava flows filling in some of the local topographic deeps. The region as a whole is not as reflective as that within the Matua trend to the east. In spots, there are vestiges of abyssal hill topography, indicating that the overall, average thickness of the flows resurfacing the area is probably only on the order of 100 m or less. Because of geographic limitations of the survey, we are not sure how far to the west the resurfaced area extends. Within this area, there is a very narrow (<2 km wide) volcanic ridge, roughly 200 to 300 m high, that links some of the larger seamount edifices. This is the feature in the entire GLIMPSE study area that is most likely to be associated with a single crustal crack. We were not very successful in our attempts to dredge fresh samples from this narrow ridge or the edifices along it; we recovered mostly sediment and pumice, with only one basalt sample yielding an age of 2.11 ± 0.05 Ma, significantly older than rocks dredged to the east of Hotu but significantly younger than the underlying seafloor. This feature could have formed along a tectonic fissure or it could represent eruption along a dike extending from a volcanic center. There is no indication of faulting or graben formation adjacent to the narrow ridge. Magma emerging from a fissure beneath the narrow ridge cannot be feeding the entire, 50-km-wide, resurfaced area, because many of the cones are at depths equal to or shallower than the ridge itself, and there is an intervening, deeper area between the ridge and the southern part of the field; lava cannot simply flow down hill to fill in the deeps. The lava flows have the appearance of emerging from local centers.

[19] At the eastern end of the Matua trend, where the highly reflective seafloor and the lack of a major edifice suggest that volcanism is just beginning, there is no indication of rifting, fissuring, or diking along a linear trend (Figure 5). At the finest resolution of raw or binned SeaBeam data from an individual swath, we should be able to recognize any linear feature more than about 5 m in amplitude in the form of a step offset or in the form of a valley with width more than 200 or 300 m. There is a mounded cluster of seamounts 100 to 300 high, and associated lava flows fill in the deeps, damming up against abyssal hill topography. Outside this resurfaced area, however, there are no features detectable trending east-west or perpendicular to the East Pacific Rise. Figure 5 is constructed from swaths with east-west trend, so there are some artifacts beneath the ship tracks on the side-scan map and subtle ones where the swaths merge on the bathymetry map, but we have crisscrossed this area several times on a variety of headings and none of the individual swaths give any indication of rifting preceding volcanism.

4. Southern Cross Seamount–Sojourn Ridge–Brown Ridge

[20] Southern Cross seamount, Sojourn Ridge, and Brown Ridge collectively form a quasi-linear volcanic complex beginning within 50 km of the East Pacific Rise and extending for about 550 km to the west approximately in the direction of spreading and absolute plate motion (Figures 2 and 6). The westernmost feature in the complex, Southern Cross seamount, is diamond-shaped with roughly perpendicular "ribs" or axes. The long axis trends about N75°W. Southern Cross is the tallest feature in the GLIMPSE study area, standing about 3500 m above the surrounding seafloor and rising within 200 m of the sea surface. It does not have a flat top, suggesting that it never was at sea level, which, in turn, indicates that it formed well away from the spreading center because the surrounding seafloor is more than 800 m deeper than the axis of the East Pacific Rise and the seamount must have subsided less than 200 m. Magnetic anomalies indicate that Southern Cross seamount is not uniformly magnetized, so it must have formed during a period spanning at least one polarity reversal. There is no indication of recent volcanic activity on or near Southern Cross. A small, apparently separate ridge immediately south of the eastern tail of Southern Cross trends about N67°W and overlaps slightly with the

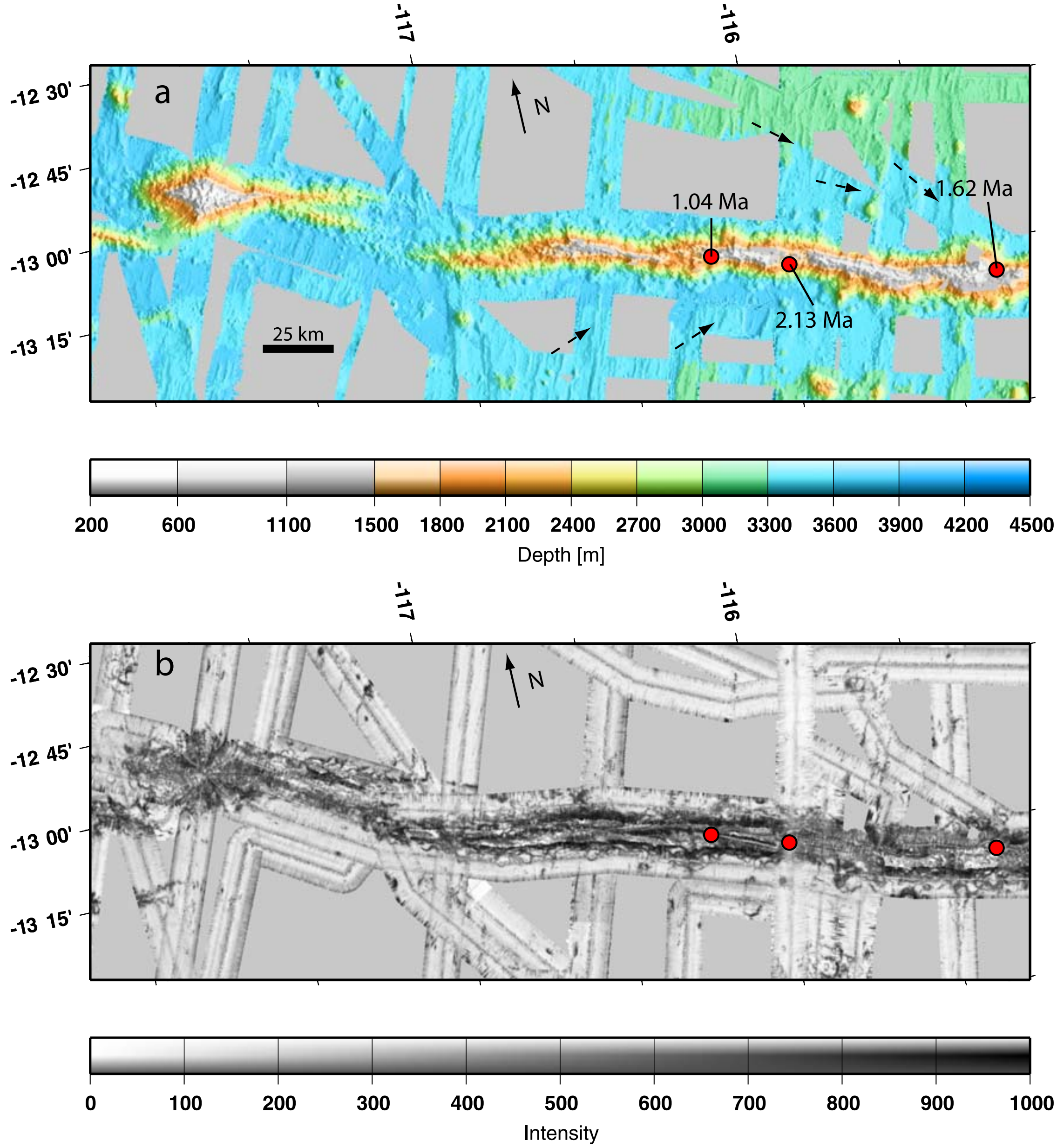


Figure 6. (a and c) Bathymetry and (b and d) side-scan sonar images of the Sojourn Ridge system (boxes D and E in Figure 2). Blue areas are the developing flexural moat on either side of the Sojourn Ridge. Flows or slumps, smaller seamounts, and sediments partly fill the moats. Increased biological activity, due to increased nutritional content of the deep currents forced to upwell over the 2 km height of the ridge, creates more sediments flanking the ridge than elsewhere in the region. Several individual ridge sections are visible in the bathymetry data forming the en echelon pattern. The dashed black arrows point to examples of abyssal hill scarps near the Sojourn that have not been buried by sediments or resurfaced by intraplate volcanism. Blue arrows indicate the distributed volcanic features at the ends of both the Sojourn and Brown ridges, while the red arrows draw attention to recent lava flows on the Brown Ridge. The most recent volcanism, i.e., highest acoustic reflectivity, appears focused on the eastern end of the Brown Ridge, indicating that portion of the ridge is actively building. Ar/Ar age locations are shown as red dots. Artificial illumination of the bathymetry is from the east.

westernmost end of Sojourn Ridge, forming a link between these two major topographic features.

[21] The Sojourn and Brown ridges both comprise several en echelon volcanic ridges that are topographically continuous at lower elevations, but have trends that are distinct from the overall trend of the ridges (Figures 6a and 6c). Sojourn trends N76°W overall with individual volcanic ridges having azimuths of 67 to 75°W of north. The overall trend for Brown is N78°W, with individual segments trending about N76°W. In comparison, the spreading direction for the East Pacific Rise is N77°W [*DeMets et al.*, 1990, 1994], and the immediately adjacent spreading seg-

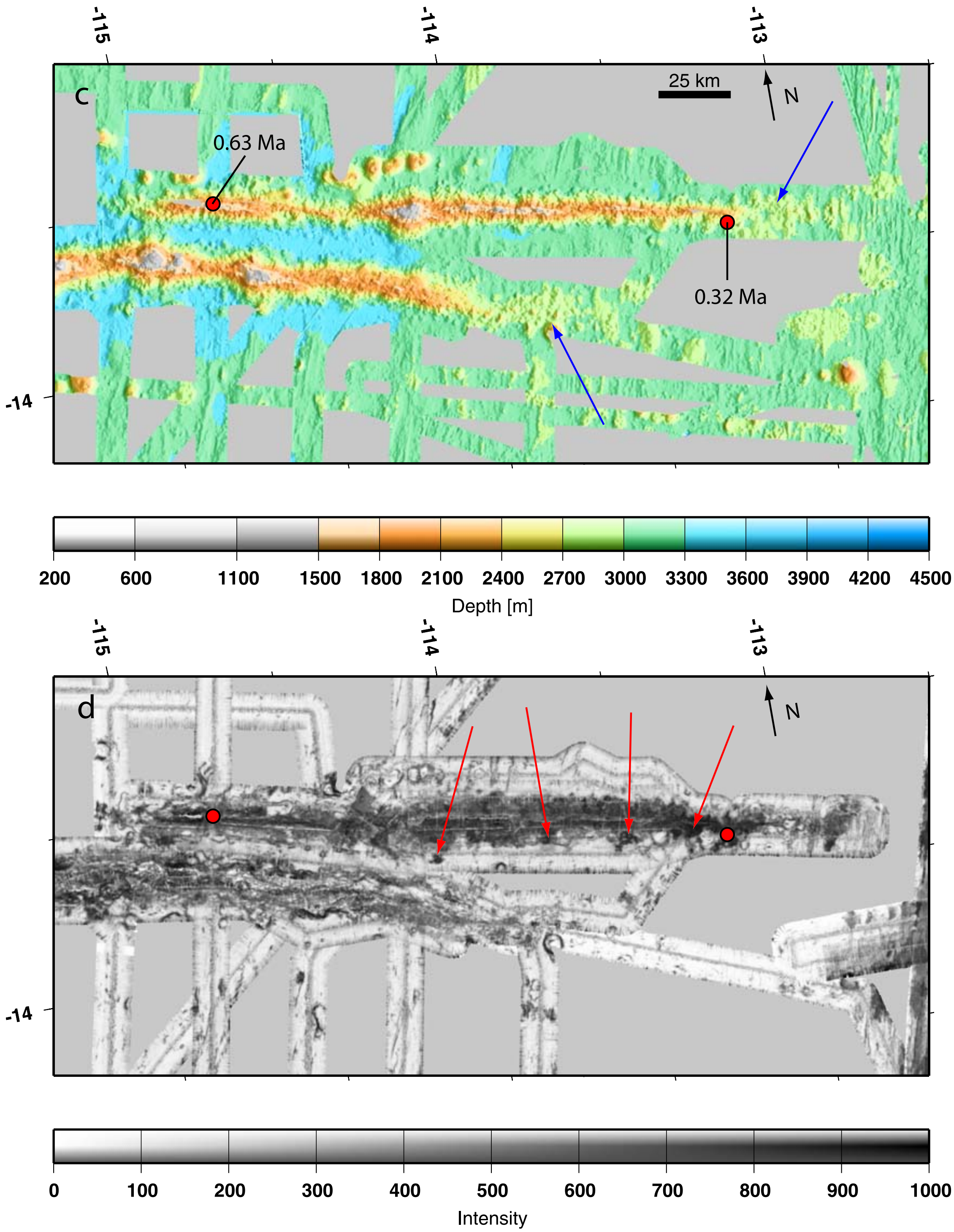


Figure 6. (continued)

ment trends N14°E. The motion of the Pacific plate in this area in the hot spot coordinate frame is indistinguishable from the spreading direction [*Scheirer et al.*, 1998]. None of the primary segments of the Brown and Sojourn ridges has a typical quasi-circular seamount shape. Some are elongate ridges of nearly uniform width with shallowest point approximately in the middle. Others bulge out somewhat, with a semiconical center and flanking ridges extending out in both directions. Typical depths to the peaks of each ridge, which we interpret as volcanic centers, are 1200 to 1500 m below sea level. Altogether, there are at least 14 recognizable individual ridges or segments within the Sojourn and Brown ridges, yielding an average spacing between volcanic centers of about 35 km.

[22] The only recent lava flows recognizable from the side-scan sonar are along the eastern part of the Brown ridge (Figure 6d), thus suggesting a west-to-east propagation of volcanism along this linear volcanic complex. While we do not have very good constraints on the rate of propagation, we do have three indications that the locus of volcanism has

moved closer to the spreading center with time. First, the radiometric dates on the Sojourn Ridge are distinctly younger than the seafloor on which it sits, with greater age contrast than between the currently active part of Brown ridge and its surroundings (Table 1). Second, there is a progressive change in apparent flexural rigidity along the chain. The eastern Brown Ridge has no recognizable flexural moat, typical of seamounts that form on very young seafloor close to the ridge axis as in much of the Rano Rahi seamount field [*Scheirer et al.*, 1996b]. They are essentially locally compensated and do not flex the surrounding seafloor enough to create a detectable moat. The Sojourn Ridge has a moat that is partially filled with sediments and is easily recognized in both topography and satellite gravity. Southern Cross seamount has a wider moat that is not as easily recognizable as the moat flanking Sojourn because the greater width leads to gentler moat slopes, but gravity/topography admittance analysis indicates that it is sitting on significantly stiffer seafloor than Sojourn [*Harmon et al.*, 2006a]. Third, rare earth element (REE) ratios indicate that there is a progression in average depth of melting from Brown to Southern Cross, with a higher proportion of melt being generated in the garnet stability field beneath Southern Cross, as would be expected if it is created by pressure release melting beneath an older, thicker lithosphere [*Donnelly et al.*, 2003].

[23] At the eastern (young) ends of both the Sojourn and Brown ridges, there are numerous, small cones scattered across a region approximately 15 to 20 km wide (Figure 6c). In these areas, there is no apparent ridge or narrow line of volcanism. Moving farther west, a central ridge emerges and builds up, but there is no further widening of the topographic ridge or broadening of the area of abyssal hills that is covered by lava flows beyond that initially established at the young end. The very small seamounts flanking the Sojourn ridge at spots (Figures 6c and 6d) may thus represent this initial stage of volcanism, with later eruptions concentrating along a single, central line.

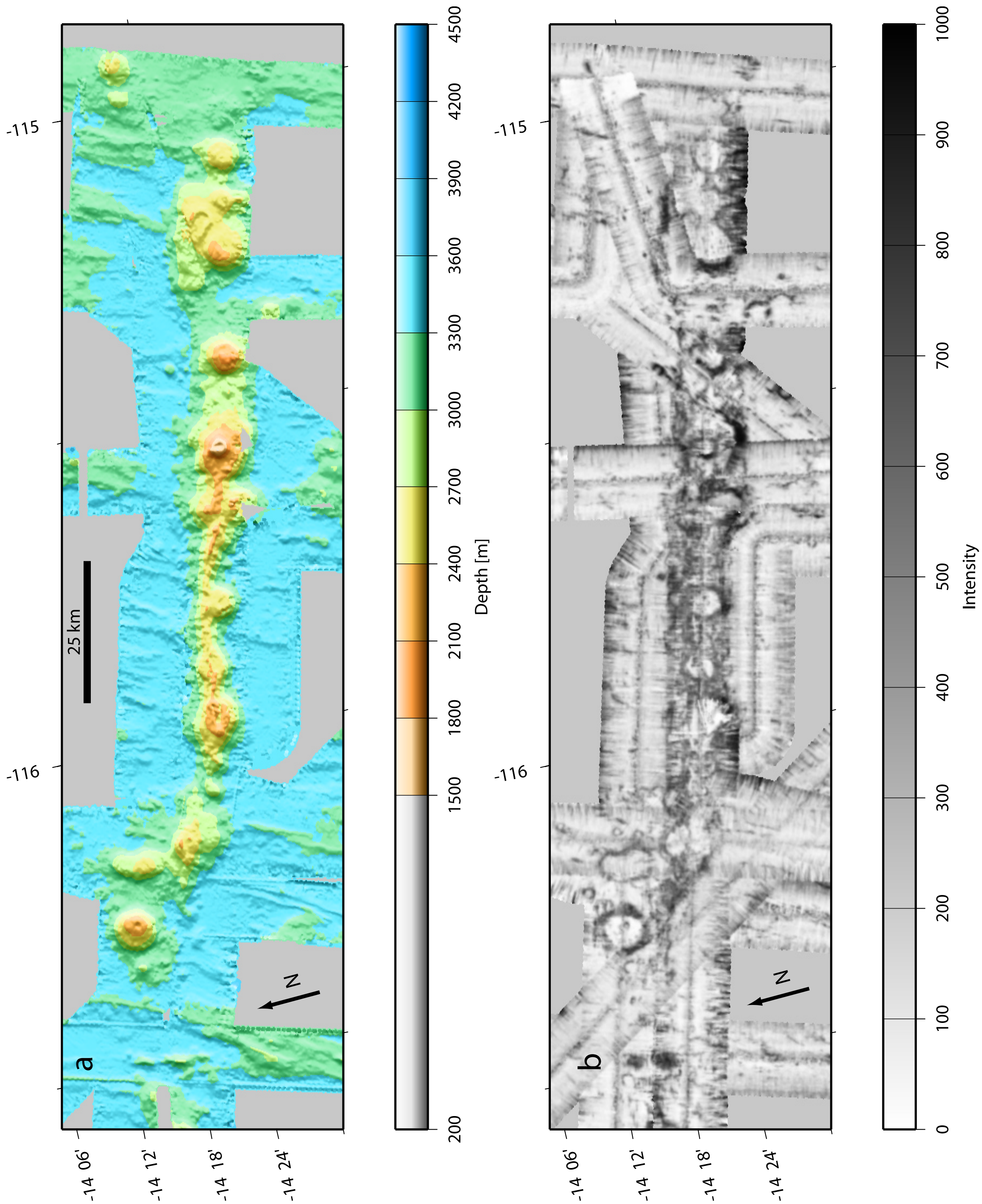
5. Other Volcanic Features

[24] There are many small seamounts scattered throughout the study area, usually approximately equant in outline, occurring as individual edifices or in short chains. We infer that these seamounts formed relatively close to the ridge axis, because no recent flows are associated with them and the population density doesn't change significantly with distance from the ridge axis. Although we singled them out above due to their size and proximity to the currently active Hotu Motua line, the Hotu seamount(s) probably also belong in this category. There is a more dense group of seamounts beginning near the East Pacific Rise at about 16.7°S and continuing west to 115°W that lie along a line approximately coincident with the overall trend of the Hotu Matua line. In this Providence and Santa Barbara seamounts group [*Scheirer et al.*, 1996a, 1996b], recent lava flows are found as far as 80 km from the spreading center [*Shen et al.*, 1993], but it is not known whether there is any genetic relationship to the Hotu Matua line because there is a gap between the two from 115 to 116°W and because the eastern part of the Hotu Matua line has a somewhat different trend.

[25] The Thanksgiving seamounts lie approximately halfway between the Sojourn Ridge and the Hotu Matua complex (Figure 2). It is a much shorter chain, 140 km in total length, and none of the seamounts stand more than 1500 m above the surrounding seafloor. A long, narrow ridge links the largest of the Thanksgiving seamounts with smaller members of the chain, but other seamounts in the group are not linked (Figure 7). Here the total width of resurfaced seafloor is only about 10 km. Except for locally steep slopes, the seafloor around the Thanksgiving seamounts exhibits relatively low, side-scan backscatter amplitudes, indicating that the chain is inactive and probably formed closer to the East Pacific Rise.

[26] About 60 km north of the Sojourn Ridge, there is a series of ridges and elongate seamounts that may be nearly continuous from about 114°30′W to 117°W. The eastern part of this chain is within 10 km of the extension of the Garrett transform fault (box G in Figure 2). The off-axis trace of the Garrett fracture zone between the East Pacific Rise and 115°W is subtle, with a bathymetric signature of parallel, low-amplitude ridges and valleys trending in the seafloor spreading direction and an offset in depth from younger to older seafloor of only 100 to 200 m. Magnetic anomalies on the few track lines that cross the Garrett fracture zone exhibit distinct signatures related to the contrast in polarity of magnetic anomalies in seafloor of different age on either side of the fracture zone. East of 115°W, there is no indication in the sparsely surveyed bathymetry or in satellite gravity of any seamounts or significant volcanic ridges associated with the fracture zone. Obvious traces of the Garrett die out at ~115°30′W, but *Goff and Cochran* [1996] interpreted the satellite gravity anomalies associated with the ridges and seamounts in this area as part of the Garrett fracture zone. The seafloor in this area was formed shortly after the cessation of spreading on the Galapagos Rise and the reorganization of the Pacific-Nazca plate boundary [*Lonsdale*, 1989], and it is not entirely clear whether the Garrett transform was newly formed at this time or if it connected continuously to a preexisting fracture zone linking the East Pacific Rise and the Galapagos Rise [*Rea*, 1981]. The only other fracture zone in the study area is a fossil feature at about 14°20'S that disappears at about 117°30'W and does not connect to any current offset of the East Pacific Rise. This fracture zone may be the conjugate to the fossil Bauer fracture zone mapped on the Nazca plate [*Goff and Cochran*, 1996]. Additional detailed surveying would be required to establish

Figure 7. (a) Bathymetry and (b) side-scan sonar images of the Thanksgiving Ridge (box F Figure 2). There is a narrow ridge that connects several seamounts in the chain. As was the case with the blade-like ridge northwest of Hotu, there are also several seamounts that are not connected to the ridge. The low reflectivity of the region suggests this feature formed earlier on younger seafloor near the spreading center. Background abyssal hill fabric trending parallel to the East Pacific Rise is visible throughout the area up to the base of the Thanksgiving Ridge. Artificial illumination of the bathymetry is from the east.


a
b
-115
-116
-14 06'
-14 12'
-14 18'
-14 24'
25 km
N
Depth [m]
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3000
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3600
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4500
Intensity
0
100
200
300
400
500
600
700
800
900
1000

Figure 7

whether the elongate ridges near the Garrett are continuous and have any genetic relationship to the fracture zone, but we are skeptical that they do because neither the mapped fracture zones nor the migrating, overlapping spreading centers in this area [*Grevemeyer et al.*, 2002] have seamounts associated with them.

[27] Another region of recent lava flows is located in the northwestern corner of the GLIMPSE study area (Figure 8). Centered at about 11°50′S and 117°30′W, reflective flows are scattered about a region at least 35 km by 50 km in extent, in some cases flowing around preexisting topographic highs. A dredge recovered highly vesicular, relatively fresh, alkalic basalt from a seafloor depth of about 3600 m; these were "popping" rocks that disintegrated loudly on deck at atmospheric pressure. Some of the samples had a flat side indicating that thin flows covered a previously sedimented surface. These flows occur in a tectonically complex area that was part of the western boundary of the microplate formed by the simultaneous spreading on the Galapagos Rise and East Pacific Rise at about the time that spreading ceased on the Galapagos Rise [*Goff and Cochran*, 1996]. Some of the abyssal hills are curved, suggesting the passage of a propagating rift or migration of an overlapping spreading center. The more prominent, straight abyssal hills are rotated 10 to 12° clockwise from the current orientation of the East Pacific Rise. Although the area lies on a line with the extension of the Garrett fracture zone, there is no indication of an established fracture zone in this area. The small seamounts in the area are irregular in shape and stand about 450 m above the surrounding seafloor at their peaks. It is not clear what caused the renewed volcanism in this area, but there is no indication of any rifting along a line perpendicular to the East Pacific Rise or parallel to the trend of the gravity lineations.

6. Discussion

[28] Nowhere in the study area is there any indication of faulting or graben formation with a trend perpendicular to the ridge axis that would be expected if lithospheric extension were responsible for the formation of linear bands of seamounts and ridges and associated gravity lineations. Combining this morphological observation with the constraints on the amount of possible extension from the spacing between fracture zones [*Goodwillie and Parsons*, 1992; *Gans et al.*, 2003] and the absence of crustal thinning to the east of Matua [*Harmon et al.*, 2006b], we join other recent investigators [e.g., *Shen et al.*, 1993; *Sandwell and Fialko*, 2004] in rejecting the diffuse extension or boudinage hypothesis [*Sandwell et al.*, 1995].

[29] There are some very narrow ridges that could be interpreted as originating as fissures or cracks in the lithosphere resulting from thermal contraction and bending or thermoelastic cracking, particularly in the 6 to 8 Ma seafloor west of Hotu and Matua seamounts and north of the western end of Sojourn. The thermoelastic cracking hypothesis is more difficult to test than the boudinage model using morphology alone because the amount of extension required at the surface is minimal and any fissuring that does occur could be covered by volcanism, provided melt is available to migrate up the crack. Adopting *Sandwell and Fialko*'s [2004] example of 130 km spacing between cracks resulting in 230 m dynamic elevation relief through bending, we can calculate the amount of opening in a crack. We assume the depth to the base of rheologically brittle plate is 20 km, corresponding to the 800°C isotherm in 8 Ma seafloor and the maximum depth of seismicity beneath Matua seamount [*Llenos et al.*, 2003]. (This value of 20 km is substantially greater than the effective elastic thickness of the plate in this area inferred from admittance analysis of gravity and topography [*Harmon et al.*, 2006a], but will yield a conservative, maximum estimate of the width of the crack at the base) Assuming zero width of the crack at the surface, the width of the crack at 20 km based on Sandwell and Fialko's model is about 280 m. It would be very difficult to detect such a crack or cracks adding up to this amount of extension unless melt emerged to form a volcanic edifice, in which case it would likely be obscured by the volcanic products.

[30] There are several problems with the thermoelastic crack model, however, as an explanation for the origin of the intraplate volcanism: it does not explain why volcanism is found over broad regions; it does not explain why volcanism has migrated rapidly from west to east, rather than staying in a fixed position relative to the ridge axis; it does not explain the origin of the melt; and it does not provide an explanation for the orientation of the volcanic lines on the Pacific plate.

[31] Thermoelastic stresses due to cooling of the oceanic lithosphere have no preferred orientation. Near the spreading center, stresses normal to the ridge can be relieved by contraction at the plate boundary, but away from the ridge axis itself, thermal stresses will accumulate. To provide a preferred orientation, *Sandwell and Fialko* [2004] appeal to an applied regional stress field, saying that the cracks will align perpendicular to the least compressive stress within the plate. They note that in older seafloor the gravity lineaments and volcanic ridges do not trend in the fossil spreading direction, so it is clear that the orientation is not controlled by the fossil spreading direction or age gradient. Because the orientation is relatively uniform in the Pacific plate despite a change in the seafloor spreading direction, Sandwell and Fialko suggest that the gravity lineaments originated within the past 5–7 Myr. Thermoelastic stresses should accumulate gradually as a function of the cooling of the plate in a steady state process. Because there is rapid propagation of volcanism toward the spreading center as we observe qualitatively in the GLIMPSE study area and has been reported for the Puka-Puka ridge group [*Sandwell et al.*, 1995; *Janney et al.*, 2000], and because there is preferred orientation of the cracking, some other mechanism must be involved.

[32] A mechanism other than remotely applied stresses that could provide a preferred orientation for thermoelastic bending and cracking is weakening of the plate along preferred lines. A preferred line of weakness could be created if melt is generated along a line in the asthenosphere, perhaps by small-scale convection. The rising melt could induce magma fracturing, weakening the plate, and the thermoelastic bending stresses might then enhance the cracking, aiding the passage of magma. In this scenario, thermoelastic cracking could play an enabling role, but the geometry of the ridge and the melt production itself is controlled by some process other than bending of the plate.

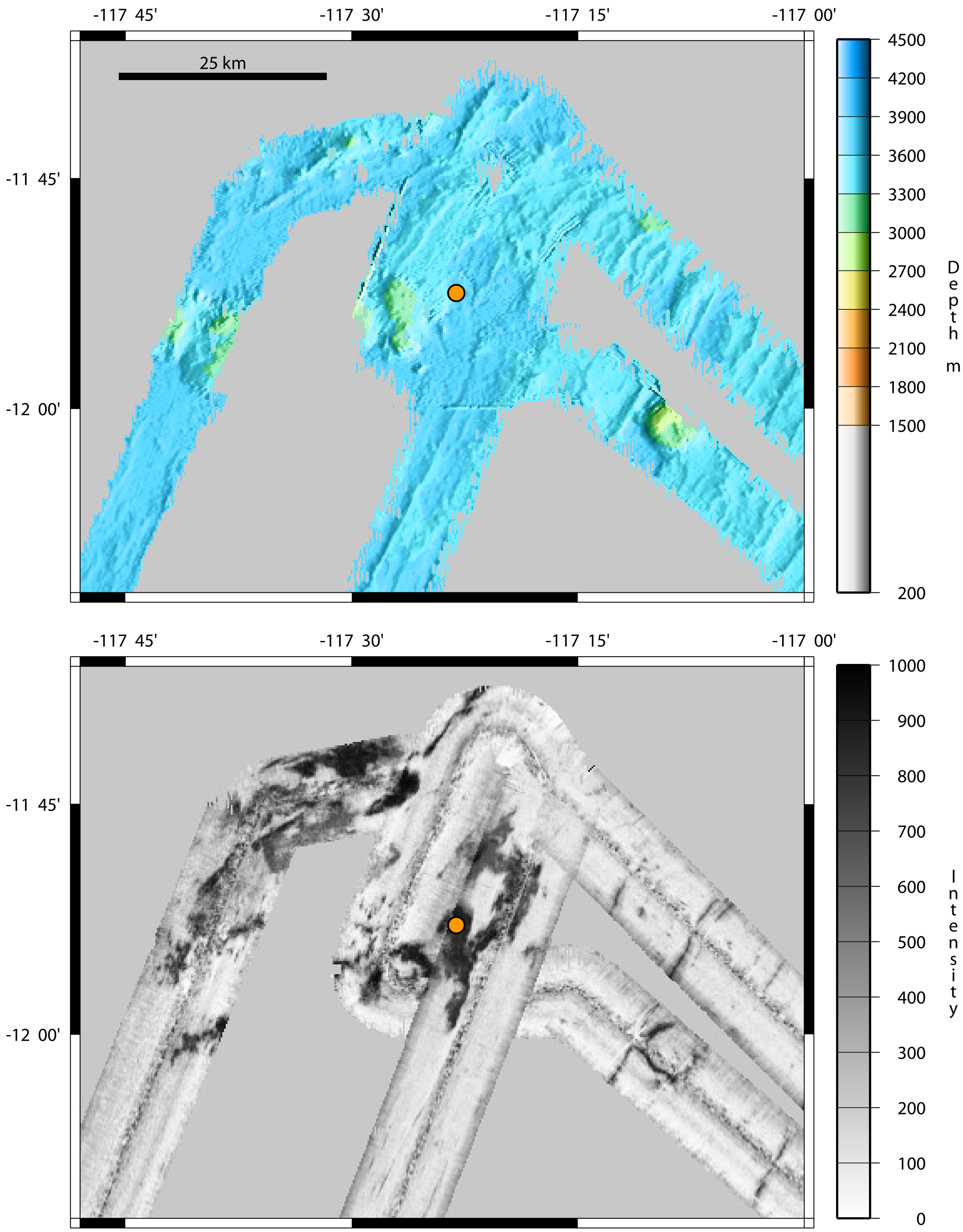

Figure 8. (a) Bathymetry and (b) side-scan sonar images of a relatively low relief, high acoustic reflectivity region in the northwest corner of the GLIMPSE study region (box H in Figure 2). The lava flows inferred from the side-scan sonar image appear to collect in topographic lows in the region. There is no clear indication of cracking parallel to spreading direction associated with renewed volcanism, and there is no clear relationship to any of the other bathymetric features in the region. There is a complex pattern of abyssal hill topography around 11°45′S reflecting the location near a dying pseudofault and near the possible origin of the Garrett transform. Relatively fresh, highly vesicular, "popping rocks" were recovered from one of the flows (orange dot). Artificial illumination of the bathymetry is from the east.

[33] It is unlikely that the magma arises from widespread, preexisting melt stably residing in the asthenosphere that is passively tapped by the formation of cracks in the lithosphere. Uniformly distributed melt would provide no preferred orientation, so remote stresses would be required to generate volcanic lineations. Active melt production by upwelling seems necessary. Melt stably residing in the asthenosphere is not actively collecting and migrating through fractures. It is unlikely that sufficient melt could be extracted to produce large features like the Sojourn ridge and the larger ridges of the Puka Puka group. In the absence of active convection beneath the plate, there is no known mechanism that could focus a large volume of magma to a crack from a widely distributed region in which a small, preexisting melt fraction resides. None of the mechanisms suggested for melt focusing at mid-ocean ridges apply to this setting: there is no corner flow that could create pressure gradients, no strain or stress field that would create anisotropic permeability, no sloping bottom of the lithosphere to channel flow upward toward the crack along a freezing front; and no concentration of dynamic upwelling into a narrow channel through melt retention and reduced viscosity.

[34] There also does not seem to be any passive mechanism for creating a sufficient volume of melt within or directly beneath a thermoelastic crack. No significant pressure release melting is predicted; the deflections of the plate are small and downward in the vicinity of the crack. Downward deflection of the plate would also tend to guide any rising melt away from the downward deflected region. There is a small volume of mantle that would flow upward to fill the crack. If we adopt the crack described above extending 20 km downward, widening to 280 m, assume that it opens rapidly so that cooling to the surrounding lithosphere doesn't occur, and assume pressure release melting at a rate of 0.4% per km of upwelling [*Langmuir et al.*, 1992], then the total amount of magma that could be produced by partial melting of the mantle flowing into the crack is sufficient to produce a volcanic ridge 1 km wide at the base and about 150 m high. This volume is 2 orders of magnitude smaller than the volume represented in a typical cross section of the Sojourn ridge. It is possible that the very narrow, small, linear ridges west of Hotu and Matua seamounts and north of Sojourn could be created in this manner. Both areas are in settings where other mechanisms could weaken the lithosphere; the first is in a broad zone of volcanism and the second is near or within the extension of the possible fossil continuation of the Garrett fracture zone. The more voluminous volcanic ridges and seamounts, however, seem to require some active, localized means of generating the melt that is independent of thermoelastic bending. *Natland and Winterer* [2005] have suggested that melt ponds beneath the lithosphere in streaks created by melting of embedded heterogeneities that have been elongated by shearing within the asthenosphere and that these ponded melts are subsequently tapped by lithospheric cracking. In this view, cracking controls the surface morphological expression, but the location of the ridges would be determined by where melt is produced.

[35] Finally, one of the strongest arguments against a thermoelastic cracking mechanism for the intraplate volcanism is the width of the areas of volcanic activity. The area around 12°S, 117.5°W is at least 35 km across, with no narrow ridge-like feature. The Hotu Matua field is up to 65 km across. On the same age Pacific seafloor at about 18.5°S, the Apitoka field of fresh lava flows is about 40 km wide [*Scheirer et al.*, 1996b]. The individual Sojourn and Brown ridges are only about 10 km across, but the volcanism appears to localize after initial activity is spread across a 15- to 20-km-wide region. Where the Brown and Sojourn ridges overlap over a distance of about 125 km, the total width is about 40 km. The Puka Puka ridges are very similar morphologically to the Brown and Sojourn ridges. *Lynch* [1999] describes a characteristic sequence of development, beginning with a wide swath of small volcanoes, followed by a narrower, scattered cluster of domed volcanoes that eventually build up into a 10- to 15-km-wide ridge. Where the Puka Puka ridges are en echelon, the total width can be as great as 70 km. *Sandwell and Fialko* [2004] describe one pair of Puka Puka ridges separated by 30 km that overlap for nearly their entire length of 150 km; a geometry that seems better described by a wide zone of volcanic activity than en echelon crack behavior. Morphologically none of these features appear to have initiated as eruptions from narrow cracks. Only the very narrow ridge west of Hotu and Matua seems to be consistent with thermoelastic cracking, and it lies within a broader field of small seamounts and lava flows that must have originated in some other way.

[36] It could be argued that the initial thermoelastic cracking occurs in random locations, so it might be broadly distributed. Once a crack develops, however, it relieves stress in the neighboring lithosphere, tending to suppress adjacent cracks, so it should rapidly localize if the site of the crack remains weaker than the surrounding lithosphere. The cracks should lengthen and propagate, so that misaligned initial locations of cracking could link up in elongate, en echelon ridges, like the Sojourn and Puka Puka ridges. However, random misalignments do not explain why the Sojourn en echelon segments consistently step in the same direction. It would appear that lithospheric stresses may control the orientation of the individual subsegments of the ridge, but that some underlying mechanism controls the overall trend or controlled the initial locations. A passive, preexisting zone of weakness in the lithosphere does not seem to be a likely explanation, because the volcanic activity of both the Sojourn and the Hotu Matua trends continue back nearly to the spreading center in the form of small seamounts on recently formed seafloor [see also *White et al.*, 2006], where they link up with more robust sections of the East Pacific Rise characterized by broader axial highs. Thermoelastic cracks are not expected to propagate back to the ridge axis and should not affect the morphology of the spreading center.

[37] If we eliminate lithospheric boudinage and thermoelastic cracking on morphological grounds, what other mechanisms remain which could be responsible for the extensive intraplate volcanism in this part of the Pacific? The volume of some of the ridges and seamounts and the breadth of the fields suggest that melt is actively being generated, rather than just passively tapped at cracks. Melt generation requires upwelling, but there are two forms of convective flow that seem possible. One possibility is that return flow to the East Pacific Rise contains heterogeneities

in composition or temperature that lead to anomalous melting as the asthenospheric flow shallows beneath a plate that thins approaching a spreading center. As D. Weeraratne et al. (Rayleigh wave tomography of the oceanic mantle beneath intraplate seamount chains in the South Pacific, submitted to *Journal of Geophysical Research*, 2006, hereinafter referred to as Weeraratne et al., submitted manuscript, 2006) point out, viscosity heterogeneity associated with compositional or thermal variations could lead to the return flow being concentrated in channels, which could further enhance localization of melt production. Any associated density heterogeneity could also create some component of local, convective upwelling. Another viable possibility is small-scale convection generated by the negative buoyancy and instability of the cooling lithosphere [*Haxby and Weissel*, 1986; *Buck and Parmentier*, 1986]. In order to explain the rapid propagation of volcanic activity from west to east, a convective cell would either have to be transported back toward the ridge axis in return flow in the asthenosphere or the instability would have to propagate from older seafloor where the lower thermal boundary layer is thick enough for convection to begin [*Buck and Parmentier*, 1986; *Marquart*, 2001; *Huang et al.*, 2003; *Korenaga and Jordan*, 2003] into younger seafloor where it is metastable. These possibilities will be discussed further in the companion papers [*Harmon et al.*, 2006a, 2006b; Weeraratne et al., submitted manuscript, 2006], but note that asthenospheric return flow has been suggested to explain the asymmetry in seafloor subsidence and other physical characteristics of the East Pacific Rise in this area [*Phipps Morgan et al.*, 1995; *MELT Seismic Team*, 1998; *Toomey et al.*, 2002] and it has been predicted in global models of mantle flow [*Gaboret et al.*, 2003].

[38] Metastable propagation from older seafloor into young seafloor could also be invoked by advocates of cracking models. Stresses could accumulate, eventually becoming large enough to crack the plate, and then the crack might propagate back into younger seafloor. Neither metastable convection nor metastable crack propagation, however, seems to be a likely explanation for the rapid propagation of the Puka Puka ridges for over 2000 km. Gravity lineations are found ubiquitously west of the East Pacific Rise from the equator to about 20°S. If metastable propagation of a crack is required to penetrate into younger seafloor, how does the characteristic wavelength between lineations decrease nearer the ridge axis [*Haxby and Weissel*, 1986]?

[39] Either transport of heterogeneities in the return flow or the alignment of small-scale convection by shear beneath the plate [*Richter and Parsons*, 1975] could explain the orientation of melt production in lines approximately perpendicular to the East Pacific Rise and in the direction of motion of the plate in the hot spot coordinate frame. Once the line of volcanism is established from below, thermal weakening of the lithosphere and the membrane stresses associated with flexure due to self-loading of a line of seamounts [*Hieronymus and Bercovici*, 2000] could be responsible for the development of ridges rather than isolated seamounts. Remotely applied stresses or thermal stresses could also play a role in determining the orientation of dikes, but we do not think that these stresses play a dominant role in controlling the formation or orientation of the ridges. The Puka Puka and Sojourn/Brown systems seem to develop into ridges only after an initial stage of distributed flows and seamount formation, suggesting the importance of self-loading. The western Sojourn ridge (Figure 6) consists of en echelon, linear segments oriented at an angle of about 10° different from the overall trend of the ridge. If remotely applied or thermal stresses control the orientation of the individual segments, some other process, such as the distribution of melt production in the mantle, must control the overall trend of the volcanic centers that form the ridge. Finally, a series of 32 intraplate earthquakes in the Rano Rahi seamount field were all normal faults, but with random orientation of the fault planes, consistent with release of thermal stresses and extension in all directions, not extension dominated by the influence of a remotely applied stress field [*Hung and Forsyth*, 1996].

7. Conclusions

[40] 1. The Hotu Matua volcanic province is a broad zone of recent intraplate volcanism extending for at least 450 km from 119°30′W to 115°15′W. It includes fossil features like the Hotu seamount group that were probably formed near the East Pacific Rise; a still active volcano, Matua seamount, that stands 2000 m above the surrounding seafloor and is surrounded by recent lava flows spread out over a region 65 km across; a long, very narrow ridge to the west of Matua that extends for more than 100 km; many small, scattered seamounts, both domed and irregular in shape; and recent flows that are isolated from major constructional edifices.

[41] 2. The Southern Cross Seamount/Sojourn Ridge/Brown Ridge system is 15 to 20 km wide and extends for about 550 km perpendicular to the East Pacific Rise. Within the Sojourn and Brown ridges there are at least 14 recognizable, separate volcanic centers rising an average of about 1500 m above the surrounding seafloor. Collectively they form an en echelon set of ridges oriented a few degrees off the overall trend of the system. The Southern Cross Seamount is the largest individual feature, standing 3500 m above the surrounding seafloor.

[42] 3. There are many other small seamounts and volcanic ridges in the GLIMPSE study area, most of which were probably formed close (<50 km) to the East Pacific Rise. The Thanksgiving seamounts form a 140-km-long chain, with several of the edifices linked by a continuous ridge. Other ridges and elongated seamounts are found in the vicinity of the projected westward extension of the Garrett fracture zone, although the genetic relationship to the fracture zone is unclear at this time. Near the intersection of this continuation of the Garrett fracture zone with the tectonically complex zone where the current East Pacific Rise took over spreading from the Galapagos Rise, there is another zone of broadly scattered, recent lava flows.

[43] 4. There is no indication of faulting or graben formation that would be expected if lithospheric stretching were responsible for the formation of lines of intraplate volcanism.

[44] 5. Thermoelastic cracking associated with the cooling of the plate could contribute to the development of some of the volcanic ridges, but it does not appear to be the dominant process. Magma does not erupt primarily from

narrow cracks, but instead is distributed in broad zones. Even the relatively narrow Sojourn and Brown ridges, like the Puka Puka ridges, appear to develop from an initial stage of densely scattered seamounts; later they build up into distinct ridges, perhaps under the influence of stresses from the loading of the plate by other edifices along the volcanic line. Because thermoelastic stresses have no preferred orientation away from the plate boundary, some other mechanism is required to control the orientation of the ridges. It is unlikely that passive tapping of uniformly distributed, preexisting asthenospheric melt by cracks could supply enough magma to form the larger volcanic edifices.

[45] 6. Active melt production along a line in the asthenosphere could explain the orientation of the volcanic chains, the volume of melt produced, and the width of the region in which eruptions occur. Pressure release melting could be associated with small-scale convection aligned by plate motion or with thermal and compositional heterogeneities being transported back toward the spreading center by asthenospheric return flow.

[46] **Acknowledgments.** We thank the captains and crews of the R/V *Melville* for their dedication and proficiency. This research was supported by the National Science Foundation under grant OCE-9911729.

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