PROJECT SUMMARY

US-JAPAN COLLABORATIVE RESEARCH: MULTI-SCALE SEISMIC IMAGING OF THE MARIANA SUBDUCTION FACTORY

We propose to carry out marine multi-channel seismic reflection, controlled-source wide-angle reflection/refraction, and passive recording of local and teleseismic earthquakes to provide a comprehensive velocity, attenuation, structural and stratigraphic image of the Mariana island-arc system, from the subducting Pacific Plate to the backarc, at 15° to 18°N. This will be the first completely integrated seismic study of any active arc.

Principal objectives are to determine:
- the velocity and attenuation structure of the mantle as a proxy for temperature/partial melting of the mantle wedge below the backarc spreading center and the active arc and for hydration and metamorphism below the forearc;
- the large-scale pattern of flow in the mantle wedge, as reflected by seismic anisotropy, which controls the mantle magma supply to the arc and back-arc;
- the precise location and velocity structure of subducting oceanic crust, which will place constraints on the depth of various devolatization/metamorphism reactions and the basalt-eclogite transformation;
- the velocity and density structure of the crust as a proxy for the composition of an intra-oceanic arc, with implications for models of continental growth and crustal recycling to the mantle;
- the seismic stratigraphy and structure of the forearc, arc and remnant arc which, calibrated by existing drill hole data, record a 50 m.y. history of intra-oceanic sedimentation, magmatism and deformation involving subduction initiation, episodic arc volcanism – rifting – back-arc spreading, and serpentinite diapirism.
- the possible identification of magma chambers below active volcanoes and of the conduits beneath forearc serpentinite seamounts;
- the possible existence of an intermediate depth double seismic zone and the relationship between slab seismicity and island arc volcanism;
- the updip and downdip limits of the seismogenic zone and implications for the largely aseismic subduction in the Mariana convergent margin.

Our study will provide the baseline seismic information required for the MARGINS Subduction Factory experiment in the Mariana system. We therefore plan to collect the data necessary to create images detailed enough to guide future geochemical measurements and proposed ODP drilling to understand the material fluxes input at the trench and output in the forearc, volcanic arc, and backarc.

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Note to Reviewers: With the concurrence of the MARGINS Program Managers, we are formally submitting this as a "Collaborative Proposal". It is collaborative between researchers at multiple US institutions, as well as internationally between US and Japan. The format of a Collaborative Proposal is up to ten pages of overall project description plus up to five pages for each institutional project description. Given 8 US co-PIs, we provide abbreviated "Results from prior NSF funding" in each major section and incorporate information from several previous projects directly into the text.

1.1 Overview

Subduction zones are fundamental to Earth recycling, controlling the return of crustal materials to the mantle and the partitioning of some fraction back to the surface. Volcanic arcs contain the record of that fractional recycling. Understanding the processes and products involved in the initiation, growth and rupture of arcs is essential to assessing the geochemical and dynamic evolution of the mantle, the origin of hypothesized continental building blocks, and the partitioning of material fluxes through the subduction system.

The processes internal to the subduction factory are, however, essentially hidden. Good petrological constraints indicate that island arc magmatism results through the interaction of volatiles from the slab with hot material in the mantle wedge [e.g. Gill, 1981; Tatsumi and Eggins, 1995; Pearce, 1995]. However, the physical transport processes involved are poorly understood [Davies & Stevenson, 1992; Davies, 1999]. Seismic imaging and anisotropy studies of the upper mantle have the potential to help constrain the dehydration locations and depths and mantle flow patterns associated with these processes. Similarly, geochemical and petrological studies suggest some interaction between island arc and backarc magma source regions [e.g. Stolper & Newman, 1994], but the details of this interaction and how it is related to mantle flow patterns is not understood.

The production rate and composition of arc magmatism are also disputed. The two best crustal structure images of island arcs (Izu-Bonin and Aleutian) are at odds [Suyehiro et al., 1996; Fliedner & Klemperer, 1999; Holbrook et al., 1999]. The Izu-Bonin data are consistent with a tonalitic middle crust, the Aleutian data with a more basaltic one. The parallel geochemical controversy, between an andesitic versus basaltic bulk arc composition [Arculus, 1981; S.R.Taylor, 1977; McLennan & Taylor 1982; Kay & Kay, 1986; Smithson et al., 1981] therefore remains unresolved by geophysical data.

Thirdly, the extent, timing and processes by which an arc is stretched and thinned, intruded by magma and serpentine, and tectonically eroded or accreted, await better seismic stratigraphic and structural images in order to be resolved.

The Izu-Bonin-Mariana (IBM) region is the classic example of an intra-oceanic arc – trench – back-arc system [Karig, 1971a,b]. The MARGINS Program has identified the IBM system as a focus area for interdisciplinary investigations of the "Subduction Factory" (see http://soest.hawaii.edu/margins/SubFac.html). There, the history of subduction input, volcanic output and back-arc spreading are well studied (including 8 ocean drilling legs: 6, 31, 58, 59, 60, 125, 126, 185; Figure 1); there is no continental influence on sedimentation or magmatism, and the present subduction of seafloor materials is virtually complete [Taylor, 1992]. Furthermore, in the Mariana convergent margin, a complete cross section of modern surface outputs from the "factory" can and have been sampled: from the serpentine seamounts in the forearc to the spreading center in the backarc [Figures 2 and 3; Fryer, 1996].

Therefore, we propose a nested, multi-scale seismic experiment to better image the processes and products of the Mariana Subduction Factory. We include marine multi-channel seismic (MCS) reflection, controlled-source wide-angle reflection/refraction, and passive seismic imaging, to provide a comprehensive velocity, attenuation, structural and stratigraphic image of the Mariana island-arc system, from the subducting Pacific Plate to the backarc, at 15° to 18°N. These experiments are required if we are to test and refine existing models of arc formation and evolution, crustal growth and recycling, and mantle flow and melting.

1.2 Tectonic and Geologic Background

The IBM system is the product of 50 Ma of subduction and associated magmatism [e.g., Cosca et al., 1998]. Currently, old Pacific plate lithosphere (late-Cretaceous to early-Jurassic) subducts beneath the Philippine Sea plate at rates increasing northward from 2.4 cm/yr at 12°N to 6.1 cm/yr at 34°N [Seno et al., 1993]. A well-developed backarc spreading center exists west of the Mariana arc, and shows a morphology similar to a slow spreading mid-ocean ridge (Figure 4). Estimates for the spreading rate range from 3 to 4.3 cm/yr [Bibee et al., 1980; Hussong & Uyeda, 1981; Seama and Fujiwara, 1993], making the total convergence rate at the Mariana Trench on the order of 7 cm/yr. The repeated occurrence of back-arc spreading along the IBM arc through the Cenozoic may be related to the general retreat of the Philippine Sea plate in the mantle reference frame [e.g., Chase, 1978].

The Pacific plate subducts beneath the Mariana arc with an unusually steep dip, and is essentially vertical between depths of 200 and 600 km (Figure 2). Low resolution tomographic images indicate that the Mariana slab penetrates through the 670 km discontinuity and descends nearly vertically into the lower mantle, whereas the Izu-Bonin slab lies recumbent immediately above the 670 km discontinuity [van der Hilst et al., 1991]. This difference may be due to the absolute motion of the respective arcs; the Mariana arc has remained nearly stationary in the mantle reference frame over the past 45 Ma, whereas the Izu-Bonin arc has moved toward the northwest [van der Hilst & Seno, 1993]. In the Mariana system, slab seismicity occurs throughout the entire upper mantle between 16°N - 20°N, whereas south of 15 °N seismicity is restricted to depths shallower than 200 km (Figure 5).

The IBM arc began with middle and upper Eocene supra-subduction zone magmatism (boninites and island arc tholeiites) similar to many ophiolites [Taylor, 1992; Stern & Bloomer, 1992; Bloomer et al., 1995]. A modern-style island arc with volcanism concentrated along a volcanic front developed by the latest Eocene (35 Ma). In the Mariana system, this is represented by the Guam-Saipan frontal arc ridge (Figure 4). The island arc was episodically rifted and backarc spreading successively isolated the remnant arcs (Palau-Kyushu Ridge and West Mariana Ridge), forming the Parece Vela and Shikoku Basins from 30-15 Ma and the Mariana Trough since ~8 Ma [Fryer, 1995; Hussong & Uyeda, 1981; Kroenke, Scott et al., 1980; Kobayashi et al., 1995; Mrozowski & Hayes, 1979; Okino et al., 1998, 1999].

The Mariana arc tephras record temporal variations in volcanic output, from low-K (Oligocene and early Miocene), to high-K (late Miocene), and subsequently medium-K compositions (Plio-Quaternary) – most likely, some argue, as a function of varying subducted sediment input [Bryant et al., 1999]. High field strength element data indicate that the degree of mantle melting is more-or-less constant at 25-30% along the active arc, and this is consistent with evidence from the heavy REE data for an absence of garnet in the melt residue [Bloomer et al., 1989b; Peate & Pearce, 1998]. Peate & Pearce [1998] estimate that ~10% of the melting may be attributed to decompression and ~15-20% to fluid addition. Using these estimates and the data of Stolper & Newman [1994], the mantle source of the arc melts would need 0.4% water, significantly more than the upper end (0.25% water) of their estimate for the mantle sources of Mariana Trough glasses.

The outer IBM forearc lacks all but the most ephemeral accretionary prism at the toe of slope. Instead, it has been cited as the type example of a forearc being subjected to tectonic erosion (e.g., Hussong and Uyeda, 1981), but this suggestion remains controversial (Karig and Rankin, 1983). Resolving whether there is net addition, subtraction or stability of the forearc is fundamental to quantifying the material fluxes through the whole system. Furthermore, the outer forearc is the locus of extensive remobilization of serpentine derived from hydration of mantle peridotite [Fryer et al.., 1985, 1995; Fryer & Fryer, 1987; Fryer, 1992a,b, 1995, 1996; Fryer et al., in press; Horine et al., 1990].

Although the 50 Ma tectonic and magmatic history of the IBM arc is arguably better known than any other convergent margin, the velocity, attenuation, structure and stratigraphy of the Mariana arc is very poorly known. The perspective view of the central Mariana system (17-18°N, Figure 3)

summarizes the crustal structure inferred from 1970s refraction work using sonobuoys and two OBSs [Bibee et al., 1980; LaTraille & Hussong, 1980; Sinton & Hussong, 1983; Ambos, 1984]. Much of the subsurface structure is extrapolated rather than measured. Only one regional transect of two parallel MCS reflection profiles exists (at 17°45'-18°N). These R/V *Conrad* 2006 profiles were collected with a 24 channel streamer and 4x466 in^3 airguns, and were not migrated [Mrozowski et al., 1981]. Although there are many commonalities between the Izu-Bonin and Mariana arcs (outlined above) that suggest their seismic character should be similar, there are important differences that make it impossible to extrapolate the Suyehiro et al. [1996] seismic data from 32°N to the Mariana system at 15°-18°N. These differences include the late Miocene splitting of the Mariana arc; the curvature and N-S extension of its forearc, the different subduction inputs and volcanic and serpentine outputs, and the varying slab dips.

1.3 Proposed field experiment

We propose below an integrated multi-scale seismic program to definitively image the Mariana Subduction Factory (Figure 6). The proposal consists of three main components, including passive recording to image upper mantle structure, controlled source wide-angle reflection/refraction to constrain arc crustal structure, and MCS reflection profiling to image the seismic stratigraphy and structure of the sediments and basement. We concentrate the experiment along the Mariana arc between 15°N - 18°N for the following reasons:

1) Seismicity occurs throughout the upper mantle in this region, yielding many crossing paths for the passive tomography and mantle imaging. This region is near a "nest" of deep earthquakes at 18°N (Figure 5).
2) The arc volcanoes are subaerial here and are highly active. They show interesting local end-member geochemical variations (e.g., between Guguan and Agrigan), inferred by some to be related to aqueous fluid versus sediment melting processes [e.g. Elliott et al., 1997].
3) The frontal arc high is well developed from Saipan to Farallon de Medinilla. This may represent the thickest arc crust section, and thus is vital for the wide-angle active source component. The forearc is highly anomalous and distended from 17°30'-19°30'N (Figure 4).
4) The back arc spreading center is well organized and relatively uncomplicated [Stuben et al., 1998].
5) A large, active serpentine seamount (Celestial) is located in the forearc [e.g., Fryer, 1995, Fryer et al., in press], which is a target for future ODP drilling.

We first describe the three components of the field program, including their scientific rationale. We then describe the integrated field program and data analysis techniques, and why we expect results superior to any previous experiments. We conclude with an overview of the project responsibilities and human resources.

2. Large scale imaging of seismic structure, mantle flow patterns and seismicity: The passive component

Results from Prior NSF Funding -- D.A. Wiens and J. A. Hildebrand:

D. A. Wiens, PI, EAR-9219675, $317,451, 5/1/93 - 10/31/97, "Collaborative Research: A passive broadband seismic experiment for study of subduction zone and back-arc structure and tectonics in the S. Pacific"

D. A. Wiens, PI, OCE-931446, $147,979, 6/1/94 - 5/31/98, "Seismic study of the Lau back-arc spreading center and Tonga Island Arc using ocean bottom seismographs"

J. A. Hildebrand, PI, OCE-9314399, $641,651, 6/1/94 - 5/31/98, "Seismic study of the Lau back-arc spreading center and Tonga Island Arc using ocean bottom seismographs"

These grants funded a two year deployment (Nov, 1993 – Dec, 1995) by Wiens of 11 broadband digital seismic stations in Fiji and Tonga, and a collaborative OBS deployment involving Hildebrand, Wiens, and Dorman. 14 papers have been published on the results of these experiments (see references). Results include:

1. **Seismic tomography** using teleseismic, local, and OBS arrivals clearly images a slab with velocity 6% faster than the surrounding mantle, and suggests that slow velocity anomalies extend to depths of 400 km beneath the backarc spreading center [*Zhao et al.,* 1997].
2. **Attenuation tomography** using regional waveforms shows a low Q zone beneath the Lau spreading center [*Roth et al.,* 1999a]. Comparison of attenuation and velocity tomographic images provides an empirical relationship between V and Q that is consistent with lab results [*Roth et al.,* 1999b].
3. **Regional waveform inversion** suggests upper mantle velocity heterogeneity of up to 16% between adjacent tectonic regions, with exceptionally slow seismic velocities in the backarc basins, extending to depths of about 180 km [*Xu & Wiens,* 1997].
4. **Forward modeling of the slab travel-time anomaly** using thermal and petrological models for the Tonga slab suggest that the anomaly can be well fit by simple thermal models and accepted values of $\Delta V/\Delta T$. Initial results show little evidence of a metastable olivine wedge [*Koper et al.,* 1998]
5. **Shear wave splitting** along paths from deep events to Fiji show ~ 1.5 s splitting, with fast axes parallel to the convergence direction at the trench. Inversion of local paths and teleseismic *SKS* results show that azimuthal anisotropy is limited to the upper 400 km [*Fischer & Wiens,* 1996].
6. **Multiple ScS phases** that interact with the 400 km and 670 km discontinuities suggest that the 670 km discontinuity is depressed by 20-30 km in the region of the Tonga slab [*Roth et al.,* 1999].
7. **One of the largest deep earthquakes** occurred beneath our array in March, 1994. This event shows that deep aftershocks and mainshock rupture zones [*Wiens et al.,* 1994; *McGuire et al.,* 1997] can extend outside the active Benioff zone, suggesting that models of deep earthquake occurrence through transformational faulting in a metastable wedge are too simplistic. The aftershock sequence demonstrates that deep earthquake aftershocks align along the mainshock fault plane [*Wiens et al.,* 1994; *Wiens & McGuire,* 1999].

2.1 Velocity and attenuation structure of the volcanic front and mantle wedge: constraints on magma source and transport regions

2.1.1 Scientific questions. The source region and transport process of magma for island arc volcanoes and the backarc spreading center is one of the primary questions of the subduction factory. It is generally accepted that island arc magmas are produced through the interaction of volatiles from the slab with hot material in the mantle wedge [Tatsumi, 1989; Peacock, 1990; Tatsumi & Eggins, 1995], but the exact process and its geometry are uncertain. There is considerable debate about which hydrous phases are involved, the depth of dehydration, the transport process and geometry between the slab and the near-surface, and what controls the position of the volcanoes relative to the slab [e.g. Davies & Stevenson, 1992; Ulmer & Trommsdorf, 1995; Schmidt & Poli, 1998, Davies, 1999].

Similar questions are associated with backarc spreading centers. Presumably backarc spreading centers show many similarities to mid-ocean ridge spreading centers, in which the major element partial melting occurs between depths of about 80 and 20 km in the mantle [e.g. Shen & Forsyth, 1995]. The melt generation region in the mantle is quite broad for fast-spreading ridges [Toomey et al., 1998], but may be much narrower and have a more 3-D geometry beneath slower spreading ridges [Magde & Sparks, 1997]. The mechanism by which melt is focused along a mid-ocean ridge is still uncertain. A backarc spreading center, located immediately above highly energetic seismic sources in the slab, is the ideal location to image spreading center processes.

In addition, there is also a question of the relationship of the backarc and arc magma source regions. Backarc spreading centers generally show some geochemical signature of the slab [Fryer et al., 1982; Hawkins & Melchior, 1985; Stern et al., 1990], which can be accommodated by assuming input of a volatile component derived from the slab [e.g. Stolper & Newman, 1994]. Whether this interaction occurs directly at some distinct depth interval, or whether it simply involves fluxing of the entire mantle wedge has not been determined.

2.1.2 Relation of material properties and seismic observations. *P* and *S* wave velocities are strong functions of temperature, and the temperature derivatives are relatively well known from laboratory experiments [Anderson et al., 1992]. Volatiles such as water and CO_2 can greatly lower the melting temperature of mantle materials and are likely to have a major effect on seismic velocities in the wedge [Nolet, 1995; Zhao et al., 1997]. The effect of volatiles on seismic

velocity can probably be quantified by using the homologous temperature (i.e. the ratio of the temperature to the solidus temperature) [Sato et al., 1989]. The effect of small percentages of partial melt is to further lower the seismic velocities and to perhaps alter the ratio of *P* and *S* wave velocity anomalies [Schmeling, 1985; Forsyth, 1993; Faul et al., 1994]. Temperature, volatiles, and partial melt have an even more pronounced effect on seismic attenuation [Kampfmann & Berckhemer, 1985; Sato et al., 1989; Forsyth, 1993]. Attenuation shows an exponential relation with temperature, whereas velocity shows a linear relation, suggesting that combined observations of velocity and attenuation may provide the strongest contraints on material properties [Roth et al., 1999].

2.1.3 Previous mantle imaging work. A dense local deployment of seismographs is required to image island arc structure with sufficient resolution to answer questions about mantle wedge temperature structure and the distribution of melt and volatiles. Higher resolution studies carried out in Japan [Hasagawa et al., 1991; Zhao & Hawagawa, 1993] reveal a dipping zone of low seismic velocities above the subducting slab and beneath the volcanic front to a depth of at least 120 km. Japan and most other arc regions that have been imaged by tomography, such as the Aleutians [Zhao et al., 1995], lack an associated active backarc region to investigate the relationship of backarc to arc magma sources.

Our previous work in Tonga [Zhao et al., 1997] resulting from a combined land-OBS experiment has produced the best tomographic image of an island arc with an associated active backarc (Figure 7). Very strong (5-7%) slow velocity anomalies are found in the upper 100 km of the mantle beneath both the active backarc spreading center and the Tonga island arc. We interpret these regions as representing the primary source regions for island arc and backarc melts, consistent with petrological constraints [Tatsumi, 1989; Shen & Forsyth, 1995]. The low velocity anomalies are distinct at shallow depths (100 km), but appear to join at greater depths, possibly indicating some physical connection between the source regions. The low velocity anomaly extends to depths of ~400 km, possibly indicating that the mantle wedge is fluxed by fluids expelled as the slab dehydrates at these depths [Thompson, 1992; Nolet, 1995].

The only previous tomographic images of the Mariana region were obtained as part of large scale regional [van der Hilst et al., 1991; van der Hilst & Seno, 1993] or global tomographic inversions [Inoue et al., 1992, Grand et al., 1997]. These images use data only from sparse permanent stations on islands in the region, and are limited to resolutions of 100-200 km, which is insufficient to answer the questions discussed in this proposal.

2.1.4 Proposed study. We propose to conduct the highest resolution passive tomographic study of an active island arc and backarc region, using as sources the earthquakes in the downgoing Mariana slab. The Mariana subduction zone is one of the few subduction zones with earthquake activity throughout the upper mantle. Table 1 shows the average number of Mariana earthquakes per year from global catalogs. Many more earthquakes will be recorded by the proposed array that are currently too small for teleseismic detection. The deep earthquakes are concentrated between 16° - 20°N, immediately beneath the passive array (Figure 5).

Table 1: Mariana earthquakes / year (14°-20° N)

Depth (km)	No. $m_b > 4.5$	Total number
0-70	22	82
70-200	16	37
200-400	4	25
400-670	5	15

The earthquakes will be recorded by our array of 70 long-term ocean bottom seismographs (50 US and 20 Japanese) and by the 20 land seismographs. The array is designed with sparse OBS coverage in the outer regions, to help locate the intermediate and deep earthquakes and provide a low resolution large-scale 3-D image. A high density of passive OBSs are concentrated along a profile perpendicular to the arc near 17°N. High density 3-D arrays are located in the immediate

vicinity of the backarc spreading center and several arc volcanos to resolution of about 15 km in these regions. These arrays will enable high resolution imaging of the melt source and trasport regions beneath these features.

Arrival times will be determined using either conventional or cross correlation techniques, and the P and S wave structure determined using 3-D tomographic inversion incorporating an iterative ray-bending method. In addition, the wide-angle experiment and Japanese explosion sources will provide accurate crustal structure which can be stripped off, and Pn paths with known source times and locations that can be directly incorporated into the tomography. Both the Wiens and Hildebrand research groups have considerable experience in tomographic analysis [Wiggens et al., 1996; Zhao et al., 1997; Roth et al., 1999]. Dense OBS coverage will also allow a detailed attenuation tomographic model to be determined. Unlike teleseismic arrivals (used by the MELT experiment), local arrivals from slab earthquakes have sufficient S/N ratios and high frequency content to produce an accurate attenuation model [Roth et al., 1999]. The combination of P, S, and attenuation models will enhance constraints on the material properties at depth, such as the existence of small amounts of partial melt [Roth et al., 2000].

2.2 Seismic Anisotropy: Constraints on the large-scale pattern of flow in the mantle wedge

Detailed mapping of seismic anisotropy provides an opportunity to relate seismic observations to past and present deformation processes, including large-scale mantle flow. Where anisotropy can be associated with crystal alignment it can be related to mechanical deformation [Nicolas & Christensen, 1987] resulting from either past [Silver, 1996] or present day [Vinnik et al., 1989; Wolfe et al., 1998] mantle flow.

Modeling of the strain resulting from flow coupled to the downgoing plate predicts a fairly uniform pattern of anisotropy paralleling the absolute plate motion [McKenzie, 1979; Ribe, 1989], as indicated by shear-wave splitting measurements at island stations in backarc areas [Bowman & Ando, 1987; Fischer & Wiens, 1996; Fouch & Fischer, 1996; Fischer et al., 1998]. However, more complex patterns are possible, and large scale deviation of mantle flow due to a subducted slab was postulated by Alvarez [1982] and reported by shear wave splitting studies in South America [Russo & Silver, 1994]. Geochemical constraints indicate a complex pattern of mantle source regions and thus mantle flow in some arcs [e.g., Turner & Hawkesworth, 1998]. Numerical modeling of the likely induced lattice preferred orientation of olivine and orthopyroxene produces results that are non-unique and may only be fully tested with a more detailed mapping of the back-arc system [Fischer et al., 1999].

Spreading center processes also induce anisotropy. In the oceans both Pn [Hess, 1964] and SKS [e.g. Wolfe et al., 1996] have indicated fast azimuths sub-parallel to the spreading directions. However, ridge-parallel observations have been made at both Easter Island [Wolfe & Silver, 1998] and Iceland [Bjarnason et al., 1996]. In surface waves ridge-perpendicular fast azimuths are seen far from the ridge but reduced or ridge parallel anisotropy is seen closer to the ridge axis [Forsyth et al., 1998]. This is consistent with predictions if the olivine a-axes preferentially align vertically due to the upwelling flow [Blackman et al., 1996]. To understand these processes better we propose to use S-wave splitting and surface wave observations to map the anisotropy and thus test models of mantle flow in the backarc system.

S-wave splitting is the observation of two time-separated, orthogonally polarised S-waves from near-vertical arrivals. S-wave splitting observations are insensitive to depth variations but give geographically constrained estimates of anisotropic fast direction. We have recently used S-wave splitting estimates from the 1994 Tonga-Fiji OBS deployment to better resolve the anisotropic structure of the Lau backarc basin [Smith et al., 1998] (Figures 8-9). In the western part of the basin fast directions are sub-parallel to the direction of absolute plate motion. However, closer to the plate boundary a much more complex anisotropic signature is observed with many of the readings becoming more trench parallel.

This variable pattern of anisotropy is not easily explained by simple models of mantle strain resulting from flow coupled to the downgoing plate [Ribe, 1989]. Instead it indicates more

complex flow patterns, as suggested by the modeling of Buttles & Olson [1998], where the flow patterns change radically due to the angle of convergence. This is also consistent with the southward migration of isotope anomalies indicative of the Samoa hotspot source region [Turner & Hawkesworth, 1998]. Although large (1s) splitting times are observed at either end of the basin, much smaller measurements are recorded in the center of the basin. This is not predicted by the modeling of Fischer et al., which does not include a spreading center, and we infer that this effect may be due to vertical flow associated with the spreading center, consistent with the modeling of Blackman et al. [1996]. The spacing of OBSs in the Lau backarc was insufficient to map the details of the mantle flow pattern near the back-arc spreading center.

We propose to carry out a more extensive study of the mantle flow patterns in the Mariana backarc. In Tonga we had a very limited deployment of OBSs, primarily transverse to the spreading ridge which limited our ability to adequately map and thus interpret the observed signature. The much larger deployment of stations across the Mariana system should also allow us to examine how the flow varies along strike giving a much clearer indication of the detailed flow occurring through-out the entire backarc region. The density and distribution of OBSs in this study will be such that shear-wave splitting observations will be complemented by Pn and surface-wave observations, providing better estimates of the depth of anisotropy. Active source Pn measurements (explosion and airgun) will be particularly useful, since there is no uncertainty resulting from source mislocation. provide invaluable constraints on this aspect of the analysis.

2.3 The velocity structure of subducting oceanic crust: constraints on the depth of dehydration reactions and the basalt-eclogite transformation in the slab

Mineralogical reactions as the slab descends at a subduction zone include the dehydration of oceanic crust and sediments [Anderson et al., 1978; Delany & Helgeson, 1978], and the basalt/gabbro to eclogite reaction [Ringwood, 1982; Hacker, 1996]. These reactions are fundamental to the formation of island arc magmas within the mantle wedge [Tatsumi, 1989; Davies & Stevenson, 1992].

Seismological constraints on the down-going crust have involved either phases that convert at the interface, or phases that travel along the strike of the slab as guided waves. Observations in different regions have produced a striking divergence of opinion about the state of subducting crust. Observations of high frequency precursors in New Zealand from earthquakes in Tonga [Ansell & Gubbins, 1986; Smith et al., 1994; van der Hilst & Snieder, 1996] suggest a thin high velocity layer which may represent oceanic crust transformed to eclogite [Gubbins et al., 1994]. However, in Japan [Iidaka & Mizoue, 1991] and Alaska [Abers & Sarker, 1996], no high frequency precursor is seen; instead, a low frequency phase arrives first, and waveform dispersion calculations suggest a thin low velocity layer to depths of 100-150 km [Abers & Sarker, 1996]. Converted phases in Japan [Matsuzawa et al., 1986; Helffrich & Stein, 1993] and the Aleutians [Helffrich & Abers, 1997] also provide strong evidence for a thin low velocity layer at the top of the slab to depths of at least 150 km

We propose to use both the broad array of OBSs deployed near 17°N and a tighter array of nine land seismographs deployed on the islands of Saipan and Tinian to identify various later phases, such as *PS* and *SP* conversions from the slab interface (Figure 10b). An example of how these arrays can detect and study converted phases is shown in Figure 10, which shows finite difference synthetic seismograms calculated at a generic array with an aperture of 60 km using a velocity model including a subducting slab. We used a slab model which is based on the inferred thermal and mineralogical characteristics of the Tonga slab [Koper et al., 1998], and placed an 8-km-thick low velocity crustal layer at the top of the downgoing slab. The low velocity crust produces distinct secondary arrivals (Figure 10c), although these are not generally identifiable on individual seismograms. The array allows the data to be transformed to the T-p domain, where the arrivals can be distinguished by their arrival time and ray parameter. In particular, we observe S-to-P and P-to-S conversions from the subducting crust, and an S wave reverberation off the top of the subducting crust (Figure 10c-d). The amplitude of such reflections and conversions in the real data will place important constraints on the mineralogy of the subducting crust. We propose to use a forward modeling approach along with a range of mineralogical models for subducted oceanic crust

[Peacock, 1996; Hacker, 1996] to replicate the observed arrival times and amplitudes and place constraints on the mineralogy of subducted crust.

2.4 The updip and downdip limits of the seismogenic zone and implications for largely aseismic subduction in the Mariana arc system.

The Mariana subduction zone has traditionally represented the far end-member case of "decoupled" subduction, with backarc extension, very steep slab dip and a near-complete absence of large shallow thrust faulting earthquakes [Kanamori, 1977; Ruff & Kanamori, 1980; Pacheco et al., 1993]. One of the main scientific questions to be addressed in the Seismogenic Zone (SEIZE) component of MARGINS is what causes differences in the seismic character of subduction zones. Ruff & Kanamori [1980] proposed they result from differences in plate ages and convergence rates. Lay & Kanamori [1981] suggested that differences in the size of primary contact zones, or "asperities", play an important role. Certainly the thermal structure of the subduction zone and the width of the seismogenic zone plays a role in limiting the sizes of the largest earthquakes [Kao & Chen, 1991; Hyndman & Wang, 1993; Tichlaar & Ruff, 1993; Wiens, 1993]. Peacock & Hyndman [1999] suggest that hydrous minerals limit the downdip extent of the seismogenic zone. McCaffrey [1997] suggests that the variations result from the effect of plate convergence rate on the recurrence time of major earthquakes, given the short period of observation.

The MARGINS-SEIZE program has naturally targeted seismogenic zones showing high seismicity (Nankai and Costa Rica), but study of an aseismic end-member subduction zone is also necessary to understand the natural variation in subduction zone processes. We propose to use the array of OBSs and land stations to perform a detailed study of the seismicity of the shallow thrust zone of the Mariana Trench. Of particular interest are:

The updip and downdip extent of the seismogenic zone. One explanation for the lack of large magnitude thrust zone seismicity in the Mariana forearc is the narrow width of the seismogenic zone, which is not well constrained from teleseismic studies but could be by OBSs located in the forearc. These results can then be compared with various thermal and rheological models [e.g. Hyndman et al., 1995, Peacock & Hyndman, 1999] to better understand what controls seismogenic zone width. In particular, if the downdip extent of the seismogenic zone is raised by hydrous minerals, we would expect a shallow extent in the Mariana arc where the forearc is highly serpentinized.

Association between small earthquakes and serpentine diapirs. Serpentine seamounts in the forearc show a complex relationship with seismicity. In general, the mud volcanoes that appear active are seismically quiet, whereas sedimented ones, lacking mudflows at the surface, are associated with clusters of earthquakes (with one exception). The existing data are from earthquakes larger than magnitude 4.5. We know nothing of the possible presence of smaller magnitude earthquakes associated with these edifices. It stands to reason that there may be microseismicity associated with the eruption of the active mud volcanoes and it may be possible to trace an individual conduit within an edifice as serpentine muds protrude to the seafloor. There is no way to record such activity without an array of OBS instruments.

2.5 The existence of an intermediate depth double seismic zone and the relationship between slab seismicity and island arc volcanism.

Pressures at the region of intermediate-depth earthquakes (70-300 km depth) should be too great for brittle fracture under normal, dry conditions. Fluids from slab dehydration reactions may facilitate intermediate depth ruptures by reducing the normal stresses on pre-existing faults [e.g. Meade & Jeanloz, 1991; Kirby et al., 1996]. This suggests a relationship between intermediate-depth earthquakes and island-arc volcanism, which probably results from the liberated fluids interacting with the mantle wedge [e.g. Tatsumi & Eggins, 1995].

Intermediate-depth earthquakes may also be directly associated with island-arc volcanism. Recent U-series disequilibria measurements of Mariana volcano lavas suggests very rapid (<40,000 yrs) transport of fluids from the dehydrating ocean crust to the magma source region beneath the volcanic front [Elliot et al., 1997]. It is difficult to understand how fluids can travel rapidly, since diffusion processes are slow and rock permeability is thought to be low. One recent proposal is

that the occurrence of intermediate-depth earthquakes interconnect water-filled cracks and give rise to a hydrofracture [Davies, 1999]. The hydrofracture itself is probably not detectable seismologically but earthquakes might also be generated in the mantle wedge along the route of a large hydrofracture.

Intermediate depth double seismic zones have been found in a large number of subduction zones, but their interpretation is still uncertain [Hasagawa, 1978; Fujita & Kanamori, 1981; McGuire & Wiens, 1995]. The upper zone is generally coincident with the location of subducting oceanic crust, and thus likely the result of crustal dehydration, whereas the origin of the lower zone is less clear it being located near the center of the slab and being much more discontinuous [Abers, 1996].

Because of its remote location, very little is known about the intermediate depth seismicity of the Mariana slab. A double seismic zone was reported by Samowitz & Forsyth [1981], but it was based on only a few teleseismically located earthquakes. We propose to use the 90 station passive array to investigate the existence of a double seismic zone, and to map the spatial relationship between intermediate depth earthquakes and the major island arc volcanoes. The high density of stations near the center of the array might permit the detection of small earthquakes associated with hydrofractures above the subducting slab.

2.6 Passive seismic imaging – ancillary objectives.

The passive seismic array will collect a unique dataset that will be useful for a variety of important studies that are not directly part of the MARGINS program. Also, the entire dataset, including OBS data, will be available to the general community after two years. Of particular interest are:

Deep earthquakes. The mechanism of deep earthquakes is a major question. Various proposals include transformational faulting associated with phase transitions [Green & Houston, 1995; Kirby et al., 1996], the role of fluids [Meade & Jeanloz, 1991; Silver et al., 1995], and friction induced melting [Kanamori & Heaton, 1997]. The 90 station array (70 OBSs and 20 land stations) will be positioned over one of the major deep earthquake clusters in the world, facilitating detailed study of these deep earthquakes.

Slab penetration into the lower mantle. Previous work suggests that the Mariana slab penetrates into the lower mantle [Creager & Jordan, 1986; van der Hilst et al., 1991], but the lack of local stations has severely limited the resolution of these results. The 90 station array proposed here, combined with data from a 20 station Japanese passive OBS deployment across the Philippine plate in year 2000, will enable much higher resolution tomography and better constraints on the ultimate fate of the Mariana slab.

3. Velocity structure and composition of the crust and uppermost mantle: The wide-angle active-source component

Abbreviated Results from Prior NSF Funding to Klemperer
"Seismic velocity structure of the Aleutian Arc and Bering Shelf and the Composition of Continental Growth" NSF EAR-92-04998 $245,102

We collected and interpreted new onshore/offshore seismic refraction data around the eastern Aleutian Islands, to measure the seismic velocity of the lower crust and hence to interpret the composition of an active island arc. We used two-dozen portable PASSCAL seismic recorders in remote sites along the volcanic arc and forearc islands to record the airgun signals of UNOLS ship R/V *Ewing* as it recorded MCS data in a separate but related NSF project. (A third NSF grant funded deployment of OBSs and OBHs on two arc-perpendicular transects [Holbrook et al., 1999].) The geometry of along-arc deployment and along-arc and cross-arc shooting allowed us to tomographically invert our dataset for a 3D seismic-velocity model for the Aleutian arc. Our results provide the first 3D velocity model, and the first S velocities (Figure 11, left) for any intra-oceanic arc, hence providing better regional and lithological constraints (Figure 11, right) here than elsewhere [Fliedner & Klemperer, 1999]. Mafic rocks form the bulk of the Aleutian arc which has average P-velocities ranging from 6.7 to 6.8 km/s, clearly in excess of the average seismic velocity of continental crust which is 6.45 km/s [Christenson & Mooney, 1995]. The arc has a near-continental thickness of c. 30 km. Apart from the shallow crust, intermediate in composition, the arc is made

up exclusively of mafic rock types. Our lithological interpretation is consistent with the exposed Kohistan arc [Miller & Christensen, 1994], though in the Aleutians we also recognize a high-velocity mid-crustal layer possibly of remnant oceanic crust, as in the model of Kay & Kay [1985].

Our result shows a discrepancy between the mafic composition of the Aleutians, typical of the island arcs commonly inferred to be the building blocks of continents, and the mean composition of the continents, inferred from seismic velocities to be intermediate in composition. However, whereas the oceanic part of the Aleutian arc has higher seismic velocities than the continental average throughout the crust, the Peninsula section of the Aleutian arc is closer to the continental average in the upper 20 km of the crust [Fliedner & Klemperer, 2000]. Resolution of the discrepancy between the Aleutian oceanic arc and average continental crust seems to require repeated episodes of arc magmatism to produce a felsic-to-intermediate upper crust as observed in the continents, coupled with delamination of the most mafic lowermost crust and upper mantle during arc accretion [Fliedner, 1997; Fliedner & Klemperer, 2000].

3.0 The most important scientific question – arc composition and continental assembly

The lithological cross-section that we will produce for the Mariana arc is intended to resolve the paradoxical difference between the composition of oceanic arcs and the composition of continents. Phanerozoic continental crust is believed to grow mainly by the accretion of magmatic arcs [eg. Hamilton, 1981]. However, petrological studies of arcs, whether models of melting of subducted oceanic crust [eg. Marsh, 1976] or melting of a mantle wedge fluxed by fluids from the subducting slab [eg. Nicholls & Ringwood, 1973], imply a bulk basaltic/gabbroic composition [eg. 48-52% SiO_2, Kay & Kay, 1994], whereas continental crust is in contrast believed to be intermediate (andesitic/dioritic) in composition [eg. Rudnick & Fountain, 1995] with SiO_2 content of 57% [Taylor & McLennan, 1985] to 63% [Shaw et al., 1986]. The most commonly proposed resolution of this paradox is delamination of mafic-ultramafic arc lower crust during arc accretion [eg. Kay & Kay, 1993; Taira et al., 1998], but this cannot be evaluated without detailed knowledge of the degree of fractionation within the arc crust, and hence whether sufficiently dense (high-velocity) lower crust exists in oceanic arcs. Velocity resolution in seismic experiments is typically worst in the critical lower-crustal section, but the improved source and receiver shooting geometries and greater attention to S-wave recording that we propose can overcome this problem to obtain a well-constrained cross-section of an oceanic arc.

3.1 Previous crustal wide-angle data on intra-oceanic arcs

Previous experiments: The two best crustal wide-angle experiments on intra-oceanic arcs are the studies of the eastern Aleutian islands reported by Fliedner & Klemperer [1999, 2000], Holbrook et al. [1999] and Lizarralde et al. [2000] (see "Abbreviated Results from Prior NSF Funding to Klemperer" above), and the Suyehiro et al. [1996] and Takahashi et al. [1998] study of the Izu-Bonin arc at 32°N. A newer study, in the South Sandwich islands has not yet been fully analysed, but was less well instrumented than the other two studies [Larter et al., 1998; R. Larter, pers. comm., 12/99]. An older study of the Ryukyu island arc provides velocity models for the backarc and forearc, but over the active arc their data "constrain only the upper crustal structure" [Iwasaki et al., 1990]. A German OBS study across the Mariana arc-backarc at 15°N [Lange, 1992] provides a preliminary velocity model and a crustal thickness of only 16 km but, due to the small aperture of the experiment beneath the arc and forearc, lacks velocity resolution in the basement and any structural resolution of distinct intermediate and mafic layers.

Different velocities: One of the principal motivations of our proposed study is to understand the extremely different results obtained from the Aleutian and the Izu-Bonin arcs (Figure 12) (we show Holbrook's 2D model as more closely comparable to Suyehiro's 2D model, but Fliedner & Klemperer found similar results over a 3D volume of the Aleutian arc, Figure 11). Possibly the most important difference in subduction parameters between the two arcs is that the Aleutian arc is 'accretionary' whereas the Izu-Bonin-Mariana arc is 'non-accretionary' [Fryer, 1996], suggesting that different volumes of sediment and fluid involvement in the magma generation process may result in distinctive arc compositions.

However, these differences preclude at present any possibility of providing a single answer to: “what is the composition of crustal growth at an intra-oceanic island arc?,” or to the related question, “which parameters of the subduction system control the composition and volume of the overlying magmatic arc?.” Suyehiro et al. found a middle crustal layer with Vp=6.0-6.4 km/s forming 25% of the volume of the arc (Figure 12), believed to represent tonalitic composition. Holbrook et al. found no such layer and suggest that the Aleutian crust is composed exclusively of basaltic lithologies, thus suggesting that very distinct magmatic processes exist in the different arcs. Insight into the lithology of the middle crust beneath the IBM arc is provided by several exposures. In the Tanzawa Mountains west of Tokyo, the collision of the Izu-Bonin arc with Honshu accreted a crustal slice of the arc in which a 26x8 km plutonic body of tonalite and lesser gabbro (with 11-5 Ma cooling ages) intruded middle Miocene submarine volcanics and volcaniclastics [Kawate & Arima, 1998]. Dredges from fault scarps bounding the rifted eastern edge of the Palau-Kyushu Ridge at 30°N reveal extensive submarine exposures of tonalite [Taira et al., 1998]. Likewise, similar silicic rocks were dredged by P. Fryer from major fault scarps in the Mariana forearc southeast of Guam. Given these exposures, Taira et al. [1998] interpret the Suyehiro et al. [1996] 6.0-6.4 km/sec layer to be tonalitic (i.e., felsic). A similar 6 km/sec layer was also found near 31.5° & 33°N by Hino [1991], and on the northern Palau-Kyushu Ridge, indicating its regional provenance in the northern Izu-Bonin system [Taira et al., 1998].

Missing crustal reflectivity: In addition to a specific average composition [Rudnick, 1995; Rudnick & Fountain, 1995], continental crust is also typically characterised by reflective lower crust. The Aleutian MCS profiles are lacking lower-crustal reflectivity, despite evidence for signal penetration to lower-crustal depths [Holbrook et al., 1999; McGeary, 1997]. The Suyehiro study did not report crustal-penetrating MCS data from the Izu-Bonin arc, so that our MCS profiles will be the first to test whether the apparently more evolved Izu-Bonin crust (the Vp=6.0-6.4 km/s layer) is also crust with characteristic “continental” reflectivity. If we find both low-velocity tonalitic material and lower-crustal reflectivity, it will suggest that some arcs are capable of producing continental-type material in a single cycle of arc magmatism; whereas others (the Aleutians) require substantial modification during or after accretion to a continental margin to become “continental” in character.

Along-arc changes in forcing functions: Although the Suyehiro et al. study is from the same arc system as our proposed work, it is >1500 km along-strike to the north, so a comparison with this study will be illustrative of along-strike variation in a single arc system. The variation in convergence rate, the decrease in age of subducted crust, and the changes in sediment and seamount input, from our field area to the Suyehiro 32°N study area provide changing forcing functions that will either prove to be irrelevant to arc composition, or can be modelled as potential causes of changes in crustal composition.

3.2 Technical improvements over previous wide-angle studies

3D vs. 2D: Our proposed measurements are conceived as a 3D study, rather than 2D observations that may be affected by cross-arc structures between active volcanic islands. The Aleutian islands are typically separated by crustal-scale strike-slip faults [Geist et al., 1988, 1992], and the Mariana system by radial extensional structures in the forearc that may propagate back through the active arc [Wessel et al., 1994], so that possibly the 2D Suyehiro et al. and Holbrook et al. results represent unusual crossings of the arcs (though of course the Fliedner & Klemperer 3D study confirms that the Holbrook et al. result is representative of a significant segment of the Aleutian arc). Fliedner & Klemperer [1999] relied on land stations off the line of the active-source profile (so no short offsets, hence no precise upper-crustal velocities, were obtained), and suffered from the lack of OBSs collocated with the along-arc MCS profile (so that this MCS profile was not particularly useful in interpreting the wide-angle results, nor vice versa, both deficiencies remedied in the present proposal.

Denser station spacing: The Aleutian land stations were restricted to accessible volcanic islands, and to c. 100 km station spacing, around MCS lines designed for structural studies, not wide-angle work. Both the Holbrook et al.[1999] and the Suyehiro et al. [1996] studies used OBS/OBH spacings of c. 20 km. In contrast, our proposed experiment will have OBS spacing of

only 10 km across the magmatic arc from the backarc to the forearc high, and no greater than 15 km elsewhere. Our MCS source and OBS receiver distribution is designed for a 3D crustal study, including multiple lines of OBSs (some shared with the passive-imaging deployment) to allow a better tomographic study. Our clustered OBS sites around active volcanoes Alamagan, Guguan and Sarigan, complemented by island recording sites, will offer a better chance of detecting low-velocity mid-crustal magma chambers (or zones of partial melt), if any, than previous experiment.

OBS vs. OBH: Whereas the Aleutian OBS/OBH study had only c. 25% of the instruments as OBSs (due to limited availability of such instruments) we will use exclusively OBSs in order to record S-waves, and hence make more reliable lithological assignments from the combination of P and S velocities. S velocities are particularly useful for recognizing magma chambers [e.g. Makovsky & Klemperer, 1999], and in distinguishing serpentinite from diabase or peridotite in the forearc [e.g. Holbrook et al., 1992; Christensen, 1999]. Our long-offset refraction shooting should allow us to make major improvement in our knowledge of the forearc mantle composition [cf. Peacock & Hyndman, 1999].

More appropriate seismic source: The main seismic source for our wide-angle work will also be improved. The Suyehiro et al. [1996] Izu-Bonin experiment used a source of only 3640 in^3. In the Aleutian study the R/V *Ewing* standard 8400 in^3 tuned airgun array was fired on a 20 s repeat schedule, and was successfully recorded to offsets of over 200 km on the OBSs deployed in the Aleutians, but with the serious problem of "previous shot noise" obscuring secondary arrivals. In contrast, we will re-shoot our OBS lines with a longer repeat schedule (100 s to allow the previous shot noise to dissipate), and with a larger source (11,000 in^3) tuned to lower frequencies for improved propagation. This large source will also be recorded using the 6-km streamer to acquire 12-fold 90-s records, potentially capable of tracing the subducting slab to depths of over 200 km true depth. (In the Aleutian experiment, slab reflections were seen to the full 17-s penetration, 45 km true depth, of the reflection profiles [Holbrook et al., 1999].) In addition, our Japanese colleagues will detonate a series of dynamite shots, in order to ensure that good Pn velocities are recorded along and across the arc to the full length of the long profiles.

Integrated, targeted experiment design: In contrast to the Aleutian MCS, OBS/OBH, and land station deployments funded through three separate grant proposals, our proposal will provide a study integrating both marine receivers and land observations. We have focussed on an exclusively intra-oceanic arc segment (the Aleutian experiment crossed from intra-oceanic arc to continental margin arc, so that at least part of those data reflect the more complex interaction of arc magmatism with the continental margin). Our profiles extend west of the arc across the Mariana Trough, in order to test the continuity of oceanic crust into the magmatic arc (the Aleutian MCS and wide-angle profiles did not extend far enough into the back-arc to properly test whether the high-velocity 6.5-6.8 km/s layer in the mid-crust could properly be identified with oceanic crust on which the arc was built).

3.3 Important questions and aims of the controlled-source wide-angle study

•Is there a thick Vp=6.0-6.4 km/s "tonalitic" layer present at 15°-18°N, >1500 km south of Suyehiro's prior observations? Or is there a thick Vp=6.4-6.8 km/s "basaltic" layer as in the Aleutian arc?

•What is the crustal thickness, hence magma production rate along the arc? Is the Holbrook et al. [1999] measurement of c. 2 km^3/km/Ma in the Aleutian arc, double the c. 1 km^3/km/Ma typically assumed [Reymer & Schubert, 1984] also typical of the Mariana arc? If so, we need to find more efficient crustal recycling mechanisms.

•How uniform is the crustal velocity structure of the arc along strike, and hence how uniform is the magmatic process generating arc crust beneath and between the active volcanoes?

•Can we distinguish velocity differences indicative of magma chambers below active volcanoes?

•Can we distinguish velocity differences indicative of different magmatic processes in different volcanoes correlative with the different fluid components (hence mantle wedge vs.

sedimentary input) observed on different volcanoes in our segment of the Mariana arc [Elliot et al., 1997; Ishikawa & Tera, 1999]?

•Can we trace a coherent high-velocity mid-crustal layer in the arc rearward of the arc into coherent back-arc oceanic crust, thereby supporting the view from the Aleutians that such material is still recognizable as an intact density barrier to magma segregation and emplacement [Fliedner & Klemperer, 1999; Holbrook et al., 1999]?

•Can we recognize "lower-crustal layering" in the Mariana arc, suggesting that the Izu-Bonin-type arc [Suyehiro et al., 1996] is truly a continental precursor? or is it absent (as apparently in the Aleutian arc) implying that oceanic arcs require additional tectono-magmatic evolution before becoming typical continental material?

•Can we recognize seismic bright spots on or just above subducting slab (in principal to >100 km depth using the 90 s MCS records), similar to the Aleutian bright spots [McGeary, 1997] and if so tie them to velocity anomalies that may indicate overpressured aqueous fluids, or shallow melting of sedimentary rocks?

•How does the Pn velocity of the mantle wedge vary across the arc, and hence how does mantle temperature vary across the arc? Does seismic velocity support the lower temperatures estimated from thermal modelling [Peacock & Wang, 1999], or the higher temperatures petrologically inferred from the observed magmatic products of wedge melting [e.g., Turner & Hawkesworth, 1997].

•What is the extent of serpentinite in forearc crust and mantle, recognisable from its characteristic Poisson's ratio (S-wave velocity) and is there enough present to materially affect the rheology of the forearc crust and upper part of the mantle wedge?

3.4 Methodology for wide-angle data analysis

In year 1 we will build 2D velocity models along the principal arc-parallel and arc-crossing lines, where we will have best model resolution in our shooting and recording lines. Our preferred inversion methodology will be forward modelling based on ray-tracing to obtain initial models for both P and S-wave arrivals [e.g. Luetgert, 1982], using the preliminary structural models from the MCS stack sections. This will be followed by tomographic inversion using specific layer boundaries/reflector geometries, for which Colin Zelt's code [Zelt & Smith, 1992; Zelt, 1999] is currently being used at Stanford [cf. models Figure 12, bottom]. In year 2, we will begin to build 3D velocity models based not only on the shots along 13W and 18W recorded along 18W and 13W, but also on the shots recorded on the passive, broad-band OBSs. We plan also to use John Hole's code [Hole, 1992; Hole & Zelt, 1995] in this stage, as an efficient way to handle the very large model volumes we are investigating [cf. models of Figure 11], as well as Zelt's 3D code [Zelt & Barton, 1998]. We will interpret our final P and S velocities, coupled with temperature estimates for the Mariana arc, in terms of lithology [cf. methodology of Fliedner & Klemperer, 1999] and compare with density cross-sections available from modelling of shipboard gravity data. This lithological structure of the Mariana arc and forearc will then be used to constrain the petrologic origin of this crust, and to determine which parts of the crust are stable against delamination during future accretion events. For specific portions of the data – for example if we observe possible magma chambers – we will use full-waveform inversion methods, as pioneered by Pratt et al. [1996] in the frequency domain, and by Singh et al. [1999] in the time domain. Currently these methods are too computer intensive to be used for the entire data-set, but are valuable when trying to place precise constraints on petrophysical properties such as melt percentages [cf. Singh et al., 1999; Makovsky & Klemperer, 1999].

4. Seismic stratigraphy and structure of the sediments and basement: The multi-channel seismic reflection component

Abbreviated Results from related Prior NSF Funding
"Multichannel seismic investigation of the Izu-Bonin arc-trench system"

B. Taylor and G.F. Moore: OCE 86-14687, $417,208; 1987-1989

This project investigated the volcanism, rifting, sedimentation and serpentine diapirism of the Izu-Bonin arc-trench system, and served as the MCS site survey for ODP legs 125 and 126. Results of this research were published by the PIs and their students in 12 papers (see references) and are discussed in the text below.

"Side-scan sonar and geophysical survey of the southern Mariana arc: a step toward understanding geochemical mass balance at convergent plate boundaries"

P. Fryer and G.F. Moore: OCE-9633501, $320.214; 10/01/97-09/30/00

The August, 1997 field program for this ongoing project produced HMR2 sidescan and bathymetry, 6-channel seismic reflection, magnetics and gravity data. We surveyed the outer half of the Mariana forearc from 13 to 17°N, imaging all of the serpentine mud volcanoes present in this region, and the arc/backarc region SW of Guam, covering half of the Mariana Trough south of 13°40'N to 11°45'N. We collected dredge samples and cores from backarc volcanoes and from forearc seamounts. Sediment and pore fluid studies confirm the presence of active venting of slab-derived fluids from Celestial Seamount and of the presence of serpentine muds at the summit of Turquoise Seamount. The southern Mariana spreading center is an inflated axial volcanic high, rather than the more typical axial valley north of 14°N, as a result of additional magma supply from arc sources. Five papers (see references) have resulted to date.

4.0 Reflection Profiling of the Forearc/Arc/Backarc

Our seismic reflection profiles of the Mariana system will provide the highest resolution images of its tectonic, volcanic and sedimentary history. They will provide the cross-sections that (1) reveal the variation of the system along- and across-strike, (ii) place outcrop and drill sites in context and (iii) allow the geologic evolution to be better interpreted. Understanding this evolution is essential if the MARGINS Subduction Factory initiative is to quantify the modern and average material fluxes through the Mariana system.

4.1 Arc Growth Versus Rupture

MCS profiles along and across the inner forearc, arc and remnant arc will image the structures and stratigraphy resultant from three stages of arc growth and two episodes of arc rupture known from seafloor drilling studies of the Mariana system. An integrating hypothesis to be tested (in concert with the wide-angle studies) is that the *different crustal structures and compositions of intra-oceanic island arcs reflect their different histories of arc rifting and backarc spreading* (and are not simply a function of varying slab input and mantle melting).

4.1.1 Background. Tephra studies and measures of explosivity index versus age reveal three Mariana volcanic maxima at 35-24, 18-11 and 6-0 Ma (Lee et al., 1995). These three periods are likely related to significant magmatic differentiation and crustal growth. With a grid of MCS profiles in the Izu-Bonin system, we showed that equivalent variations in volcanic activity there correlate with recognizable seismic stratigraphic packages that can be regionally mapped (Taylor et al., 1990; Taylor, 1992). The expectation is that we can likewise regionalize the history of arc production known from a few drill sites and islands in the Mariana system.

These periods of arc growth are to be contrasted with the two periods of arc rupture that ultimately formed the Parece Vela Basin and Mariana Trough (Figure 1). Studies of lithospheric rheology [e.g., Kusznir and Park, 1987] suggest that the thick crust and high heat flow of the volcanic line constitute a rheologic weak zone that will tend to focus rifting in its vicinity [Molnar and Atwater, 1978]. However, the exact location of arc rupture can vary with respect to this line [Hawkins et al., 1984]. Detailed swath bathymetry surveys show that most of the Palau-Kyushu Ridge south of 25°N is a flexed West Philippine Basin rift-flank bounded by faults on the east and with a few cross-chain volcanoes to the west (Okino et al., 1998, 1999). This is significant because it proves that the split of the Oligocene arc occurred on its *backarc* side, leaving the Eo-Oligocene arc (Guam-Saipan ridge etc) and forearc to the east of the Parece Vela Basin. In contrast, the split of the Miocene Mariana arc occurred on its *forearc* side. Opening of the Mariana Trough isolated a majority of the Miocene arc in the remnant West Mariana Ridge (Figure 4) and the currently active Mariana arc had to build anew on the eastern edge of the Mariana Trough [Bloomer et al., 1989a].

These observations have important implications for our study of the Mariana system:

(a) In the current Mariana forearc, arc and remnant arc (West Mariana Ridge) we have virtually the complete record of 50 m.y of Mariana arc output (excepting only the minor cross-chains of the Palau-Kyushu Ridge).

(b) Because of the relative rheological strength of ophiolitic forearcs and oceanic backarcs, the rifting region of intra-oceanic arcs is confined (compared to continents) so that the limits of extension and the rates of strain can be well defined [Taylor et al., 1999].

(c) In the Mariana system we can study the structures and stratigraphy associated with the complete break-up of an arc and the initiation of backarc spreading. The syn-rift stage of arc rupture has been studied with MCS techniques in the Okinawa Trough [e.g., Sibuet et al., 1998], Bransfield Strait [Barker and Austin, 1994, 1998] and Izu-Bonin system [Taylor et al., 1991; Klaus et al., 1992]. The syn-rift structures associated with these systems, including both high- and low-angle rift-bounding normal faults, have been well imaged. What is missing are good images of the structures, and understanding of the processes, associated with the transition from backarc rifting to spreading.

4.1.2 Important Questions. Our MCS studies of the Mariana inner forearc to remant arc are designed to answer the following questions:

• What is the extent of rifted arc crust versus backarc basin crust in the Mariana Trough? *Major* differences in interpretations exist on this issue [e.g., Martinez et al., 1995; Stern et al., 1996, 1997; Yamazaki et al., 1993; Yamazaki and Murakami, 1998], in part because the low geomagnetic latitude and northerly orientation of spreading segments make magnetic anomaly lineations hard to define. No good MCS images exist in the Mariana Trough to determine the extent of oceanic versus rifted arc crust. Calc-alkaline gabbro breccias recovered from DSDP Site 453 confirm the exposure by faulting of consolidated magma chambers from deep within the West Mariana Ridge [Natland, 1981]. The structural context of these exposures (e.g., whether by normal detachments) remains unknown.

• Are the active arc volcanoes built on/through rifted arc crust [Bloomer et al., 1998] or backarc crust? The latter scenario presents the possibility of sandwiching oceanic crust directly within arc crust. Because of the thick (1-2 km) volcaniclastic wedge mantling the eastern side of the basin, existing geophysical data have been unable to image the rift or oceanic structures beneath the arc line. The new MCS system on R/V EWING should readily image into the basement along lines shot between the volcanoes. For example, using even the older system on *Ewing* we have imaged a crustal-scale detachment in the rifting continental arc of the western Woodlark Basin beneath 2 km of rift sediments to depths of 9 km (Figure 13; Taylor et al., 1999]

• Does the Mariana forearc contain early Oligocene rift basins equivalent to those we imaged and drilled in the Izu-Bonins [Taylor, 1992]? And, can the volcaniclastic products from the three maxima of explosive activity of the arc be recognized as seismic-stratigraphic packages and therefore regionally correlated and volumetrically quantified? Existing MCS data in the Mariana inner forearc [Figure 3, Mrozowski et al., 1981] do not penetrate through the sedimentary cover to image basement, and the area has not been drilled. In the Izu-Bonins, these rift basins contain 2/3rds of the forearc sedimentary section and represent a substantial record and product of volcaniclastic output. Proving their existence in the Mariana forearc would also substantiate the proposal that by the early Oligocene the IBM arc had not developed a sufficiently thick crustal root to localize rifting, and that the rifting was more distributed as a result [Taylor, 1992].

• Do the different crustal structures and compositions of intra-oceanic island arcs reflect their different histories of arc rifting and backarc spreading? The Aleutian arc crust is 30-33 km thick and has not experienced backarc spreading [Holbrook et al., 1999]. The Izu-Bonin arc crust is 20-22 km thick, has experienced one cycle of arc rupture and backarc spreading, and is actively rifting again [Suyehiro et al., 1996]. The Mariana arc crust has not been well imaged, but existing OBS-refraction data [Lange, 1992] suggest an even thinner crust (16 km). If this is correct, then it is likely that a primary control on arc crustal thickness (or rather, thinness) is the prior history of arc rupture (where, how often and when). This correlation may also explain the velocity/composition and reflectivity of the lower crust, as the arc may be underplated with high-Mg magmas during the rifting-spreading transition. Our deep near-vertical MCS reflection images, together with the wide-angle reflection profiles, will directly test this hypothesis. In particular, our

proposed reflection profiles across the rifted eastern margin of the West Mariana Ridge (remnant arc) should be able to provide full crustal reflection images to Moho.

4.2 Outer Forearc Tectonics

Understanding the magmatic, tectonic and metamorphic evolution of the Mariana forearc is essential if the MARGINS Subduction Factory initiative is to quantify the modern and average material fluxes through the Mariana system. Several known processes compete in this environment [e.g., von Huene & Scholl, 1991]. Tectonic accretion or underplating will strip materials from the subducting Pacific plate and cause seaward growth and/or uplift of the forearc, whereas frontal or basal tectonic erosion will cause the opposite. Collision or passage of subducting seamounts may locally uplift then collapse the forearc. Hydration of mantle peridotites produced by slab dewatering may cause volume increases and/or serpentinite intrusion/protrusion that will uplift the forearc. Arc magmas often intrude into the forearc [R. Taylor et al., 1995]. The net product of these processes (i.e., the net flux of material from its first input at the trench to passage beneath the arc), is recorded in the stratigraphy and structure of the forearc sediments and basement.

Our proposed MCS transects are placed to image the outer forearc in three areas with varying tectonic forcing. We propose to image the forearc in perhaps the least complex and disrupted area (ESE of Saipan), where there is an active serpentine mud volcano and subducting ridge (east of Sarigan), and along the DSDP Leg 60 drilling transect near maximum forearc disruption at 18°-19°N (Figure 6). Our MCS studies of the Mariana outer forearc are designed to answer the following questions:

• Although there is little evidence for modern tectonic accretion [e.g., Taylor, 1992], has there been significant tectonic erosion, and/or did accretion occur previously? The Mariana forearc has been cited as the type example of a forearc being subjected to tectonic erosion [e.g., Hussong and Uyeda, 1981], but this suggestion remains controversial [Karig and Rankin, 1983]. The key to resolving the past history of forearc uplift or subsidence, and hence the record of net basement gain or loss, are the seismic stratigraphic relations (onlap, downlap, etc) of the basal sedimentary sequences. We propose to collect MCS data to image these sequences and definitively resolve this 20-year-old controversy.

• How extensive is pre-Eocene oceanic crust within the Mariana forearc? Incorporated in the Eocene within the IBM foundation terrane *to unknown extent* are remnants of pre-existing Philippine Sea Plate oceanic crust [e.g., N-MORB at 32°N, DeBari et al., 1999] and trapped or accreted Mesozoic Pacific oceanic crust [e.g., at 20°N, Johnson et al., 1991; Pearce et al., 1999]. MCS profiles over the outer forearc will characterize the basement structures and may provide a means of recognizing if there is an arcward limit to these remnants that, to date, have only been sampled in the outer forearc.

• How widespread are arc magmas intruded into the Mariana forearc? Leg 125 ODP drilling at Conical Seamount (~19.2°N) unexpectedly encountered young (Plio/Pleistocene) basaltic sills over 100 km trenchward of the arc [Marlow et al., 1992]. In fact, ODP drill holes encountered Neogene forearc sills in all three (also Izu-Bonin and Tonga) forearcs drilled [R.Taylor et al., 1995]. Previously unreported, these intrusives appear to be a common feature of the western Pacific intra-oceanic forearcs. The geochemistry of the sills most closely resembles the subjacent volcanoes along the Quaternary arc. R.Taylor et al., [1995] infer that they were most likely fed by dikes that propagated from the arc in the basement. MCS profiles of the Izu-Bonin forearc reveal significant sill reflectors (one 21 km long) [Taylor, 1992; R.Taylor, 1995]. How much "arc" magmatism has penetrated the Mariana forearc, and how it gets >100 km from the arc, is unknown for want of quality reflection images.

• What is the internal structure of an active serpentine mud volcano? Fryer [1992] postulates that the forearc seamounts have long-lived conduits that channel rising serpentine muds and fluids to the seafloor. The fluids are slab-derived [Mottl, 1992]. Furthermore the evidence of high-pressure/low-temperature metamorphic rocks and crystal fragments in the muds indicates that the conduits tap depths as great as the top of the subducting slab [Maekawa et al., 1995]. Understanding the structure of the forearc basement is critical to understanding the evolution of these serpentine mud volcanoes and their relationships to faulting and seismicity in the region. Our

proposed experiment includes MCS profiles across and surrounding the actively venting Celestial Seamount. This is one of the seamounts for which Fryer et al. have a SCICOM-ranked ODP proposal (#505-Rev) awaiting additional MCS site surveys before OPCOM scheduling for drilling. Our MCS profiles will provide the necessary site survey for such drilling and the regional context in which to interpret the data.

• Can we map, and stratigraphically date, structures associated with the circumferential stretching of the forearc resultant from the successive episodes of backarc spreading in the Parece Vela Basin and Mariana Trough that increased its radius of curvature? This stretching [Wessel et al., 1994], apparently caused major extension of the basement. It is one reason for the massive intrusion by serpentine at 18°-19°N, where the forearc also lacks the Eo-Oligocene frontal arc ridge (Figure 4). Mrozowski and Hayes [1980] recognized substantial normal faulting of the 17°45'-18°N region on the two R/V Conrad 2006 MCS lines (Figure 3), but were unable to correlate the structures between the lines. Prior to our survey, we will reprocess and migrate these data and we will position our new lines so as to better interpolate between all four.

4.4 MCS Data Processing and Analysis

From previous experience on the *Ewing*, an extensive amount of initial data processing can be done on computers during the cruise. We will conduct a "first pass" of routine processing using a commercial package (ProMAX from Landmark), as was done successfully on the recent Nankai 3-D seismic cruise. Field tapes (in SEGD format) will be read, converted to SEG-Y and written both to DLT tape (for archive and land-based processing) and to disk (for shipboard processing). The disk files will be used for semblance velocity analysis, bandpass filtering, NMO and stack, and post-stack migration. Based on prior experience, we will have each line fully processed within one day after the line is acquired. The stacked and migrated lines will be available for preliminary interpretation. Post-cruise, advanced processing will consist of dip moveout (DMO), refined velocity analysis after DMO, deconvolution, pre-stack and migration plus depth migration. See Taylor et al. (1996, 1999) and Goodliffe et al. (1999) for examples of MCS data so processed.

Analysis of the MCS data will consist of four parts related to (1) the shallow velocity structure of the forearc, arc, backarc and remnant arc; (2) the deformation structures; (3) the seismic stratigraphy and (4) deep crustal reflection imaging. The 6-km streamer is long enough to permit resolution of velocities within the sedimentary section, which are necessary for good structural and stratigraphic imaging as well as to feed into the OBS component of the program. We will obtain a rough velocity estimate through NMO-semblance analyses and a refined estimate using pre-stack depth migration where there are sufficient sedimentary reflections. Structural and stratigraphic analysis will follow standard procedures. Line ties through the DSDP sites will provide essential age and lithology information for geologic interpretation.

Sections 5, 6 and 7 of the original proposal submitted to NSF discussed the proposed experiment, including logistics and roles of each P.I. These details have subsequently changed considerably as a result of changed budgets and logistic support, as agreed with program managers at NSF. Rather than mislead potential collaborators about our intentions, we invite you to contact us for details of our current plans.

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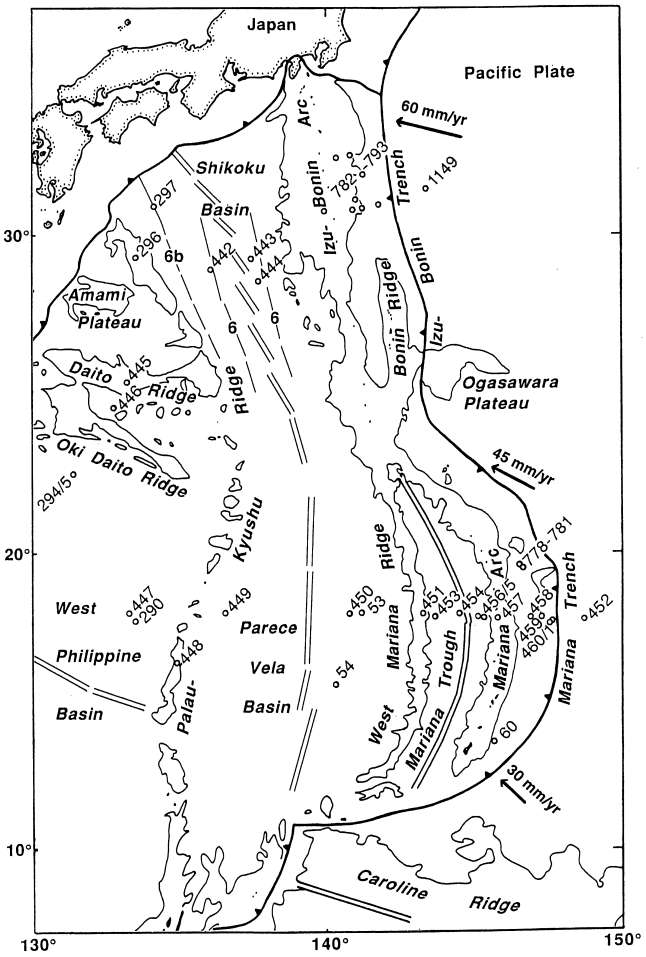


Figure 1. Volcanic arcs, backarc basins, and seafloor drilling sites of the Philippine Sea region (modified from Taylor, 1992). Relative plate motions from Seno et al. (1993).

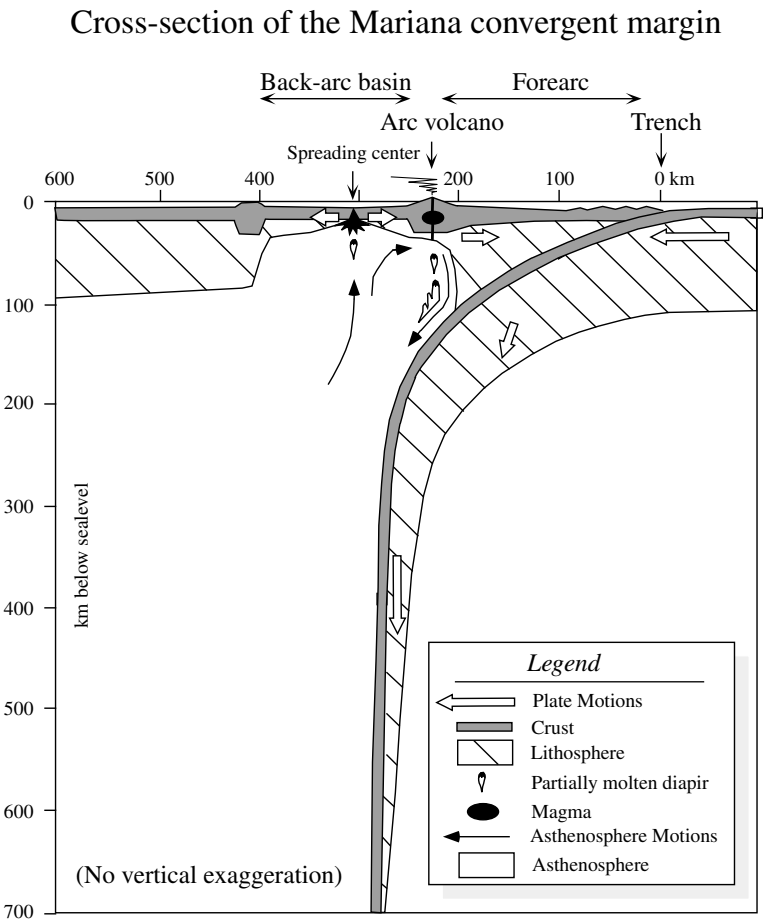


Figure 2. A textbook X-section of the central Mariana margin & postulated aspects of mantle flow and melt distribution (from Stern, 1998). Subducted slab position is based on teleseismicity.

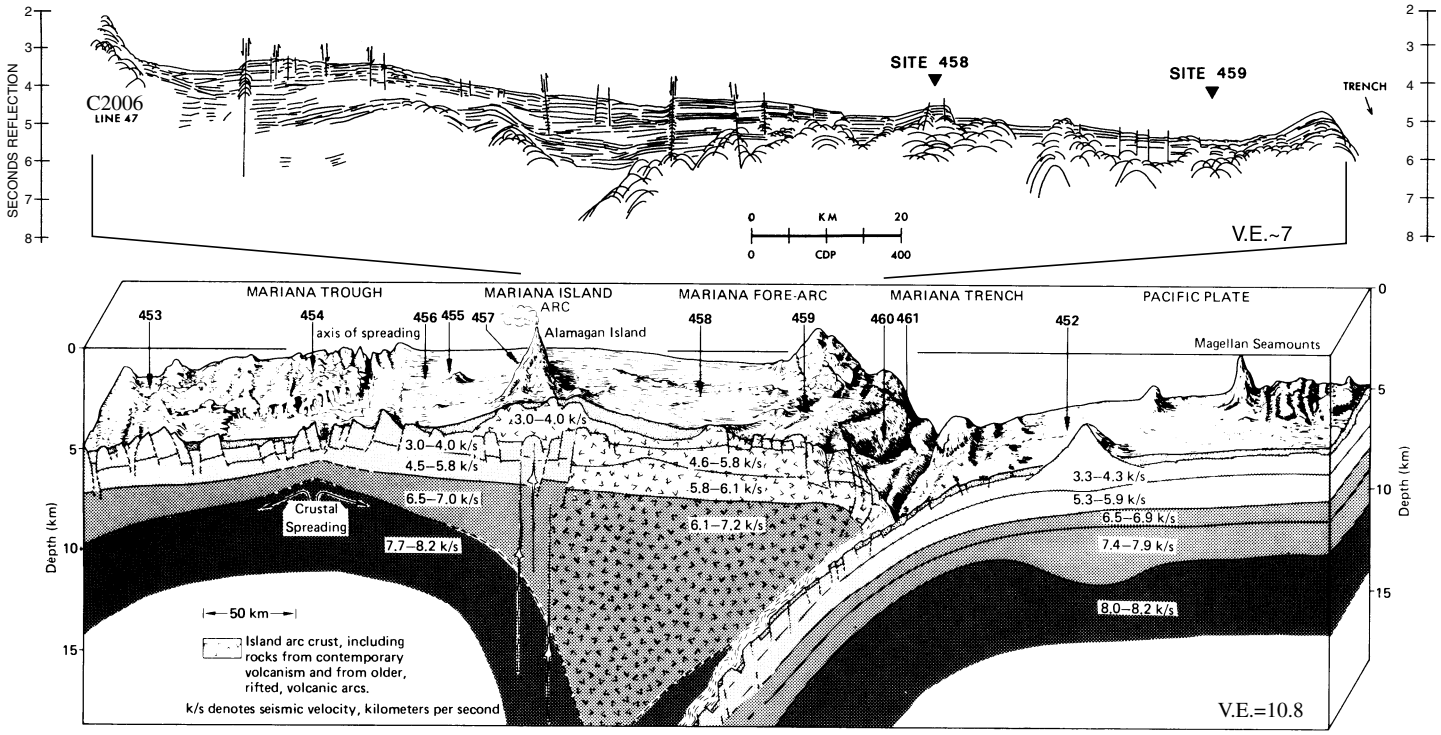


Figure 3. Block diagram of the central Mariana convergent margin from the Pacific plate on the east to the Mariana Trough on the west showing inferred crustal structure and location of DSDP Leg 60 sites (from Fryer and Hussong, 1981). Line Drawing of MCS profile above shows forearc stratigraphy and structure (from Mrozowski and Hayes, 1980).

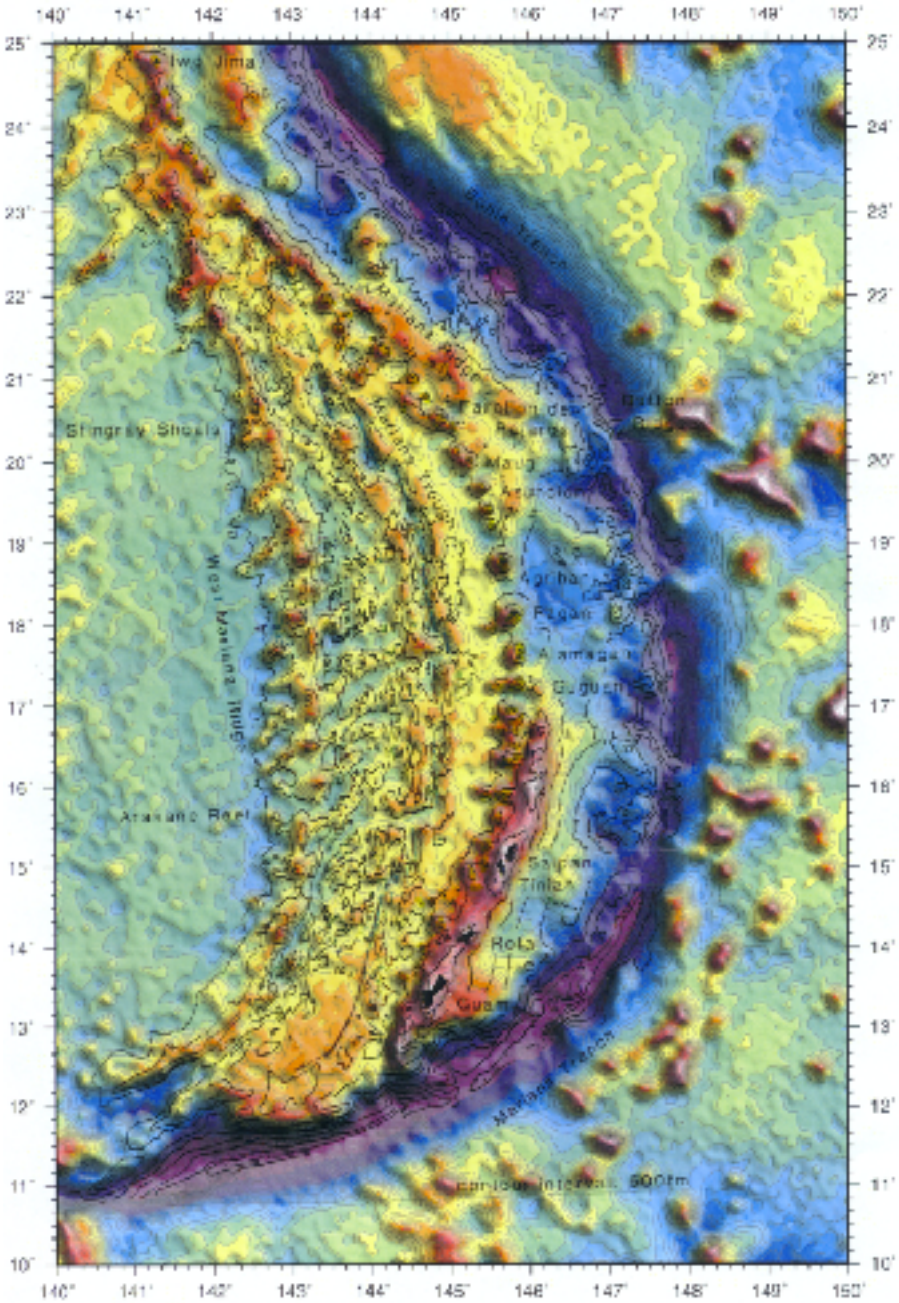


Figure 4. Bathymetry contours from Navy multibeam surveys (Smoot, 1990) overlain on satellite derived free air gravity data. Note the gravity low (lack of Guam-Saipan high) in the forearc between 17.5°-19.5°N.

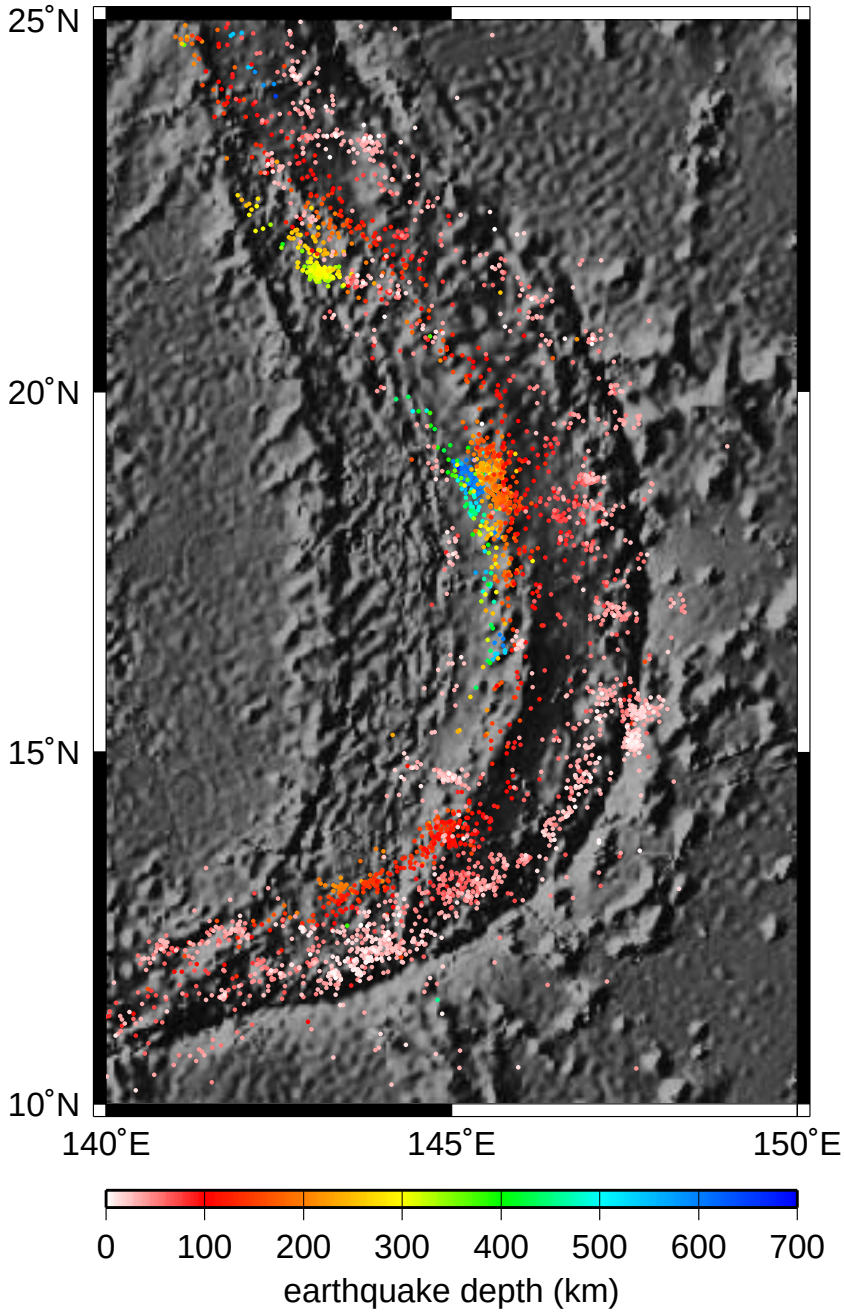


Figure 5. Earthquake seismicity (Engdahl et al., 1998) of the Mariana convergent margin plotted on a background of sunlit bathymetry. Note the deep earthquakes concentrated at 16°-20°N just west of the active volcanoes.

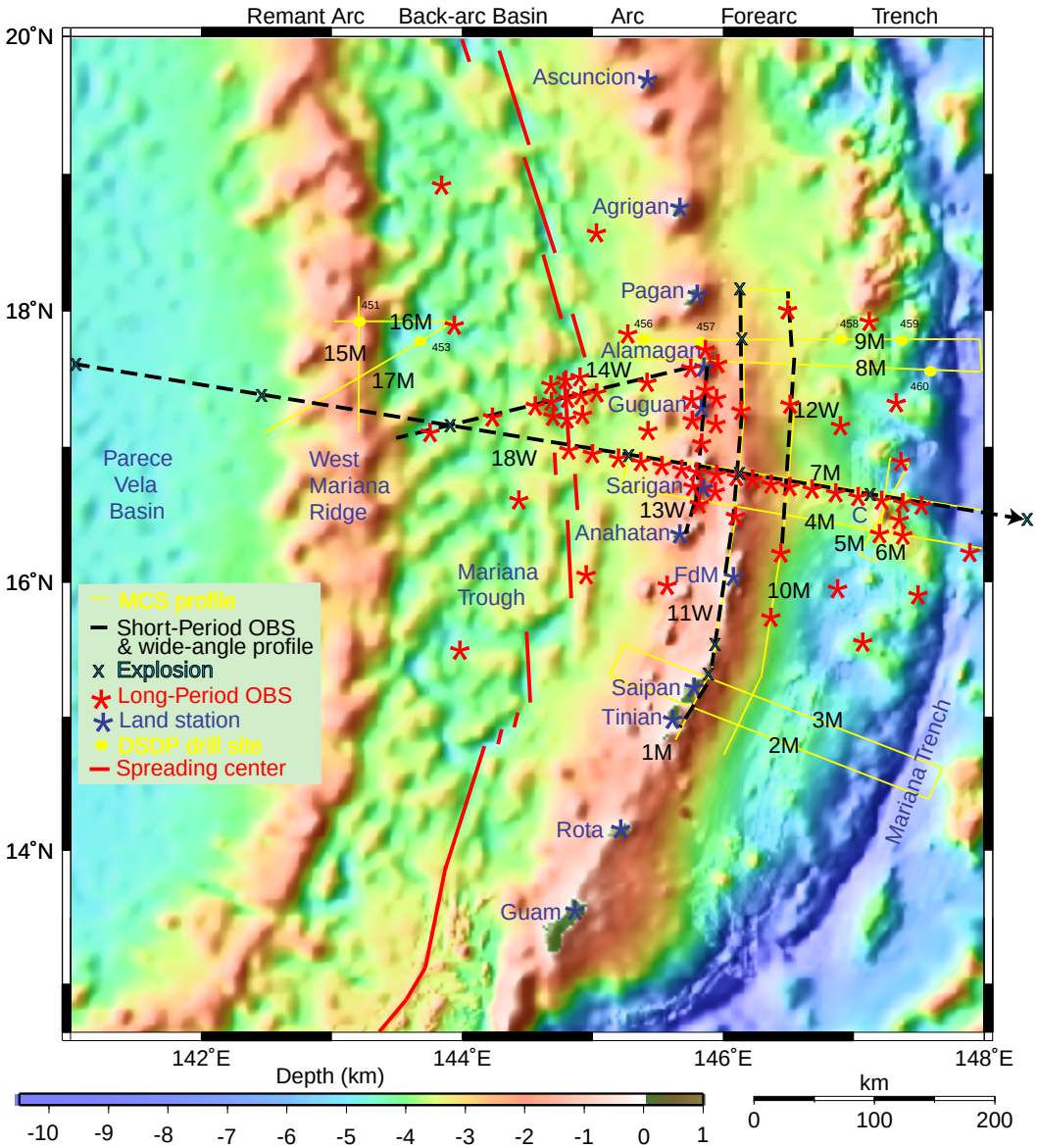


Figure 6. Location of proposed tracklines for MCS profiles (3075 km 60-fold data) (yellow), wide-angle profiles (1860 km; 100 SP Japanese OBSs, 11 explosions) (dashed black), long-period OBSs (50 US OBSIP LP instruments, 20 Japanese LP OBSs) (red stars), and land seismographs (broadband PASSCAL instruments on 12 islands) (blue stars), on sunlit bathymetry. Line 18W extends to 149°E. The average spacing of all stations (LP + SP OBS + land) along wide-angle profiles is 12.5 km, and will be 10 km between the Trench and Mariana Trough axis. C: Celestial serpentinite seamount. FdM: Farallon de Medinilla.

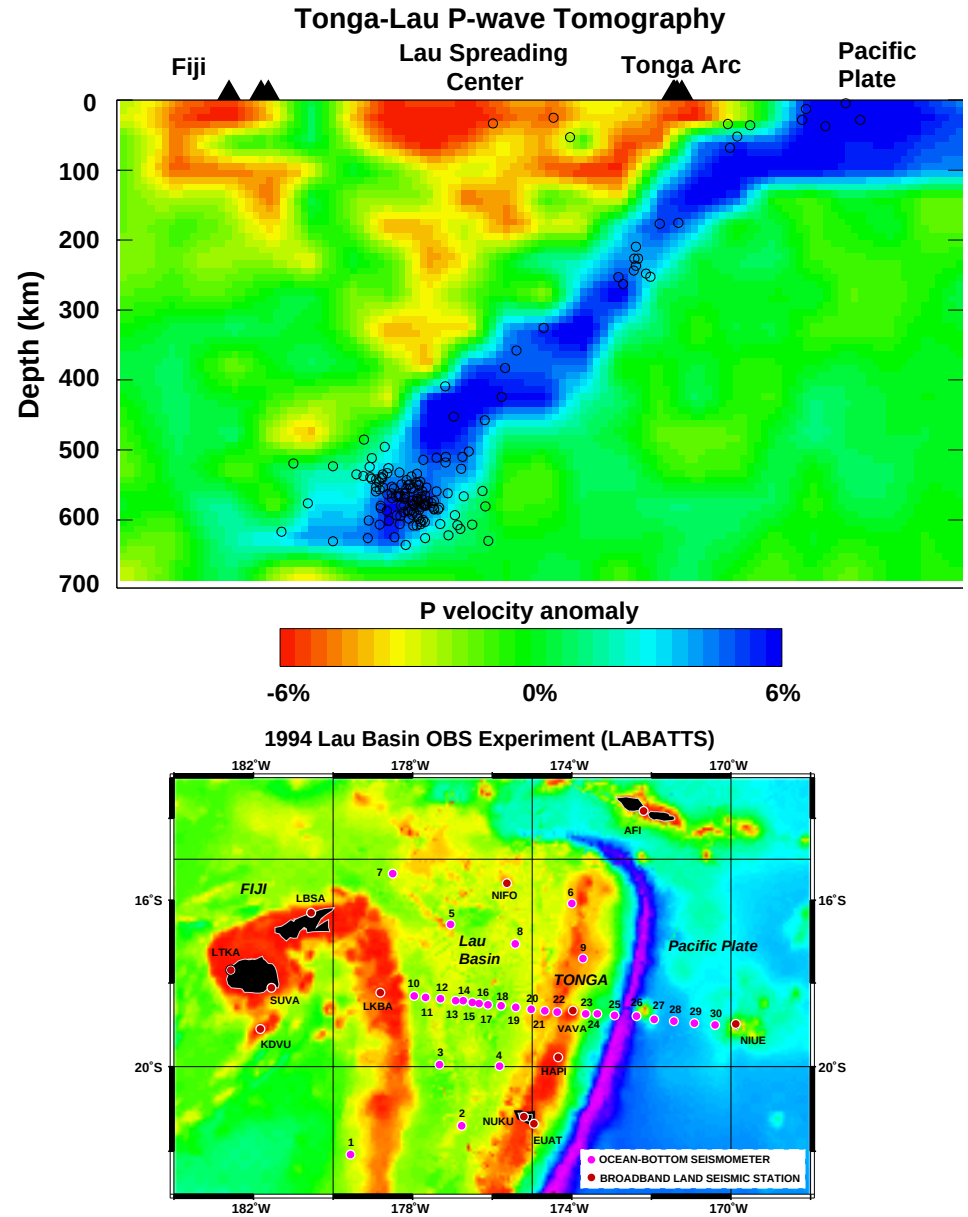


Figure 7. (upper) Seismic velocity profile of the Tonga-Fiji region obtained from P wave travel time tomography using data from the LABATTS experiment (Zhao et al., 1997). (lower) The Labatts experiment geometry on a map of digital bathymetry. This experiment consisted of 12 land seismic stations and 30 ocean bottom seismographs, which were deployed for three months. The approximate OBS spacing along the main line was 50 km. We expect much higher resolution in the Mariana experiment to result from our smaller station spacing and better 3-D station distribution (Figure 6).

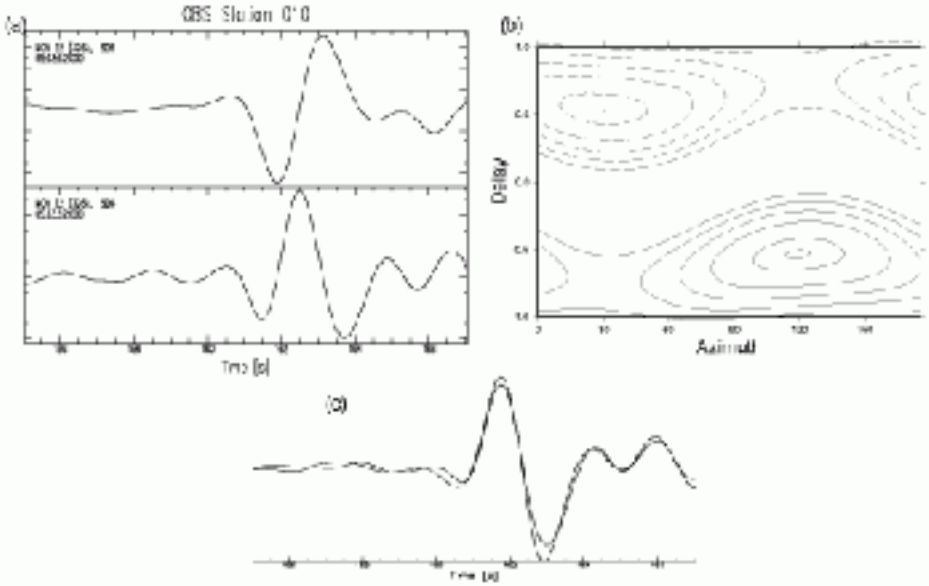


Figure 8. (a) S wave arrivals recorded at OBS station 0 10. (b) Contour plot showing the correlation as a function of rotation and time shifts. Solid lines are positive correlation, dashed lines show negative correlations. Note the maximum at 120°, -0.5s. (c) The two horizontal components after rotation and time shifting.

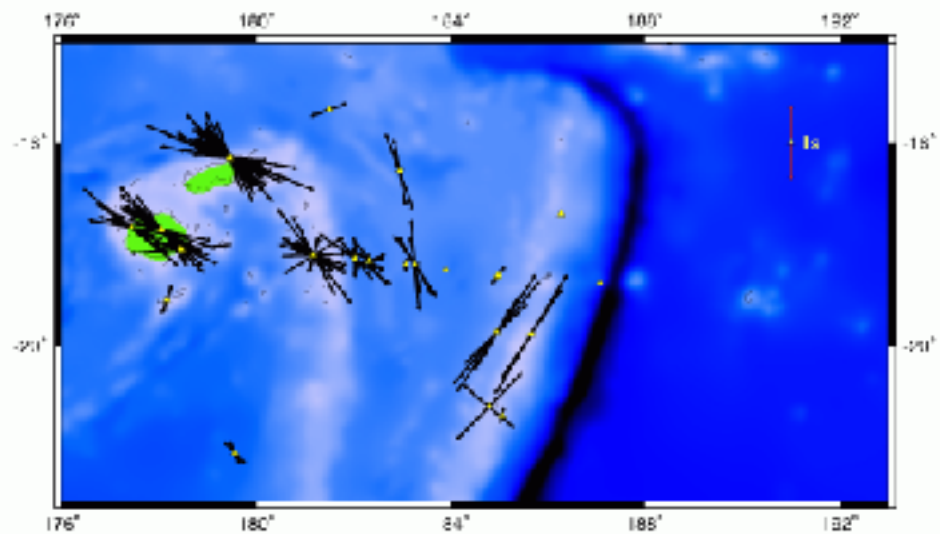


Figure 9. All splitting measurements from SPASE and LaDetts stations. Star one are represented by triangles. Splitting vectors are aligned parallel to the fast direction and their lengths are proportional to the splitting time.

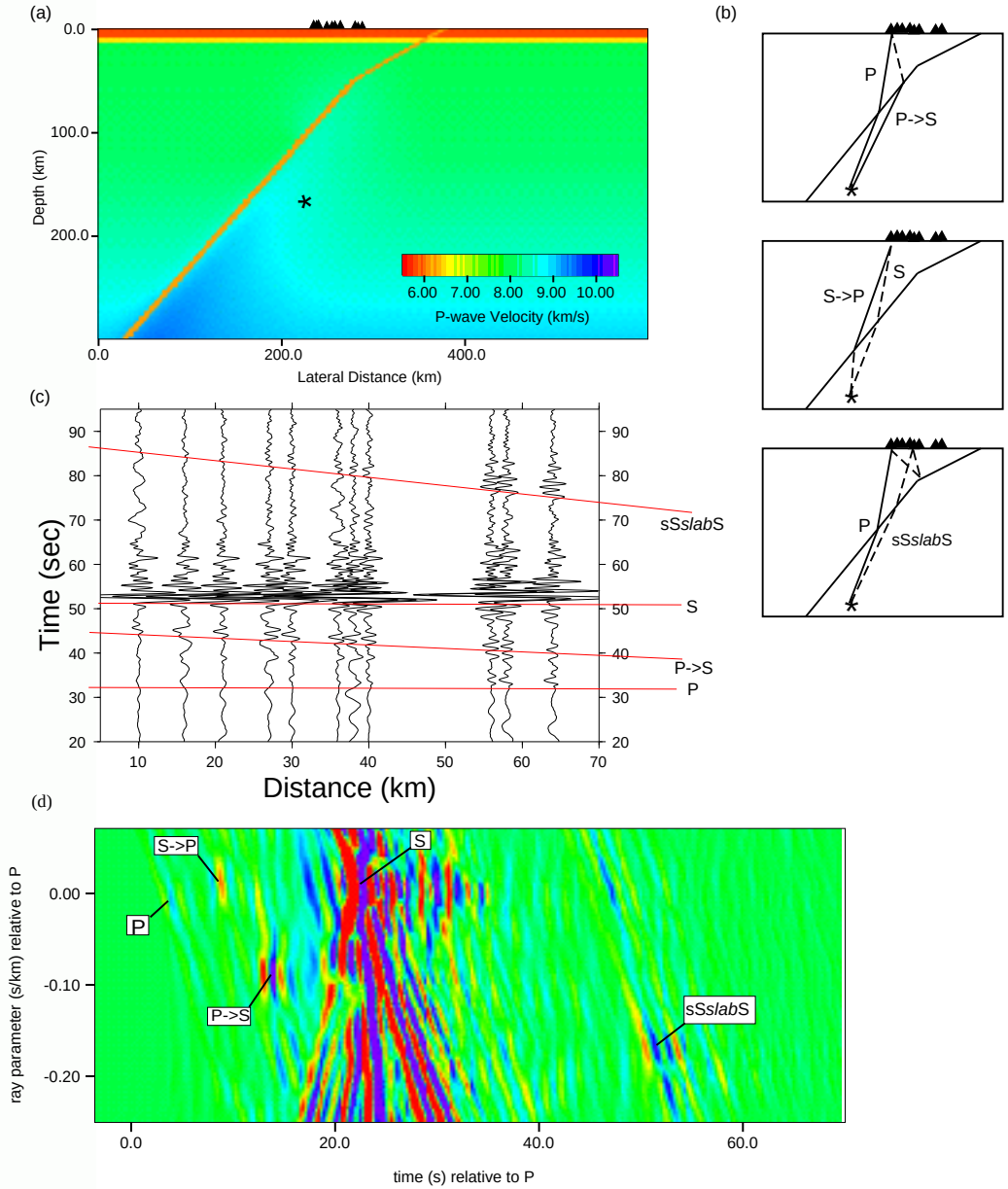


Figure 10. (a) Structural model used in calculating finite difference synthetics. The model is constructed from the inferred thermal and mineralogical characteristics of the slab and a 8km thick subducting crust. (b) Ray path geometries (c) Horizontal component finite difference synthetics generated using the model in (a) and the locations of the stations shown above. Note that real noise data has been randomly added and the P-arrivals have been aligned in time (d) Slant stacking of the synthetic traces shown in (c). Despite the randomly added real noise several slab phases are clear in the array processing which are not immediately evident from examination of the individual seismograms

Figure 11: *Velocity, Poisson's Ratio, & Lithology of the Aleutian Arc.*

Left: P-wave velocities & reflectors along 200 km of arc (for color-scale see Figure 12); & Vp/Vs ratio derived from equivalent S-wave velocity model (not shown).

Right: Lithologic column deduced from combined P-wave & S-wave velocities.

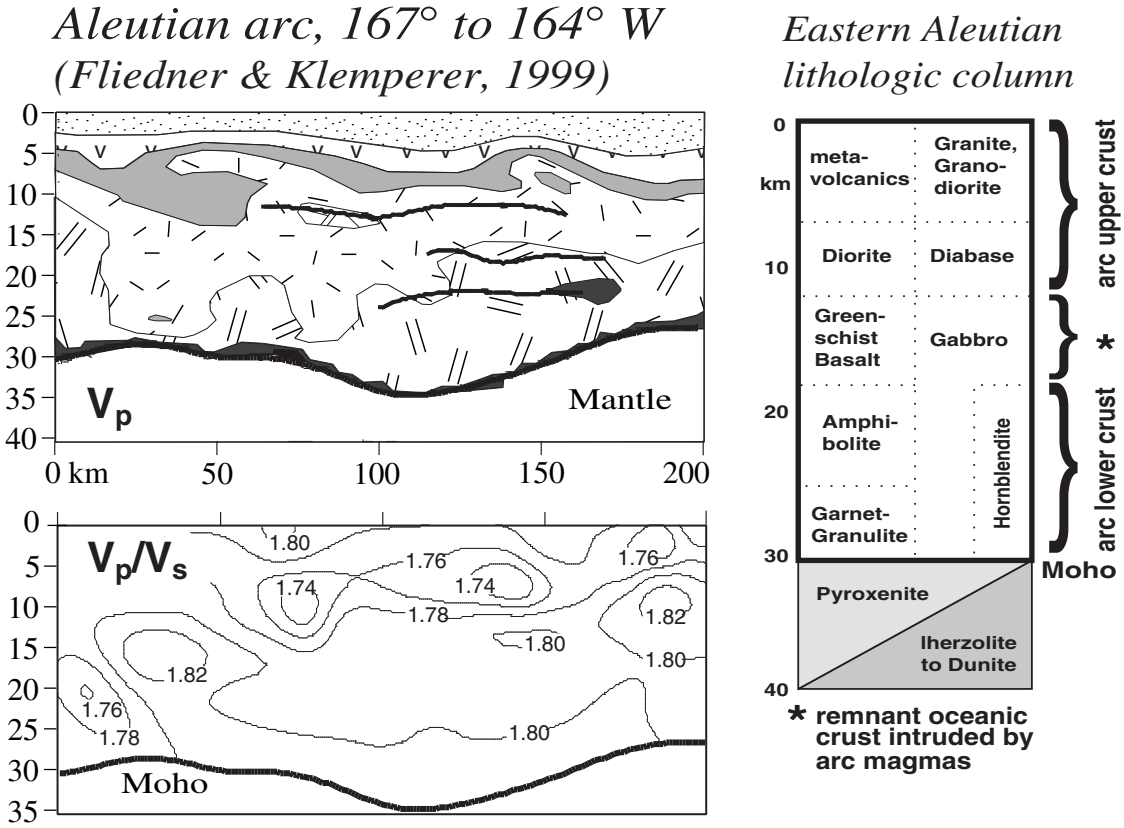
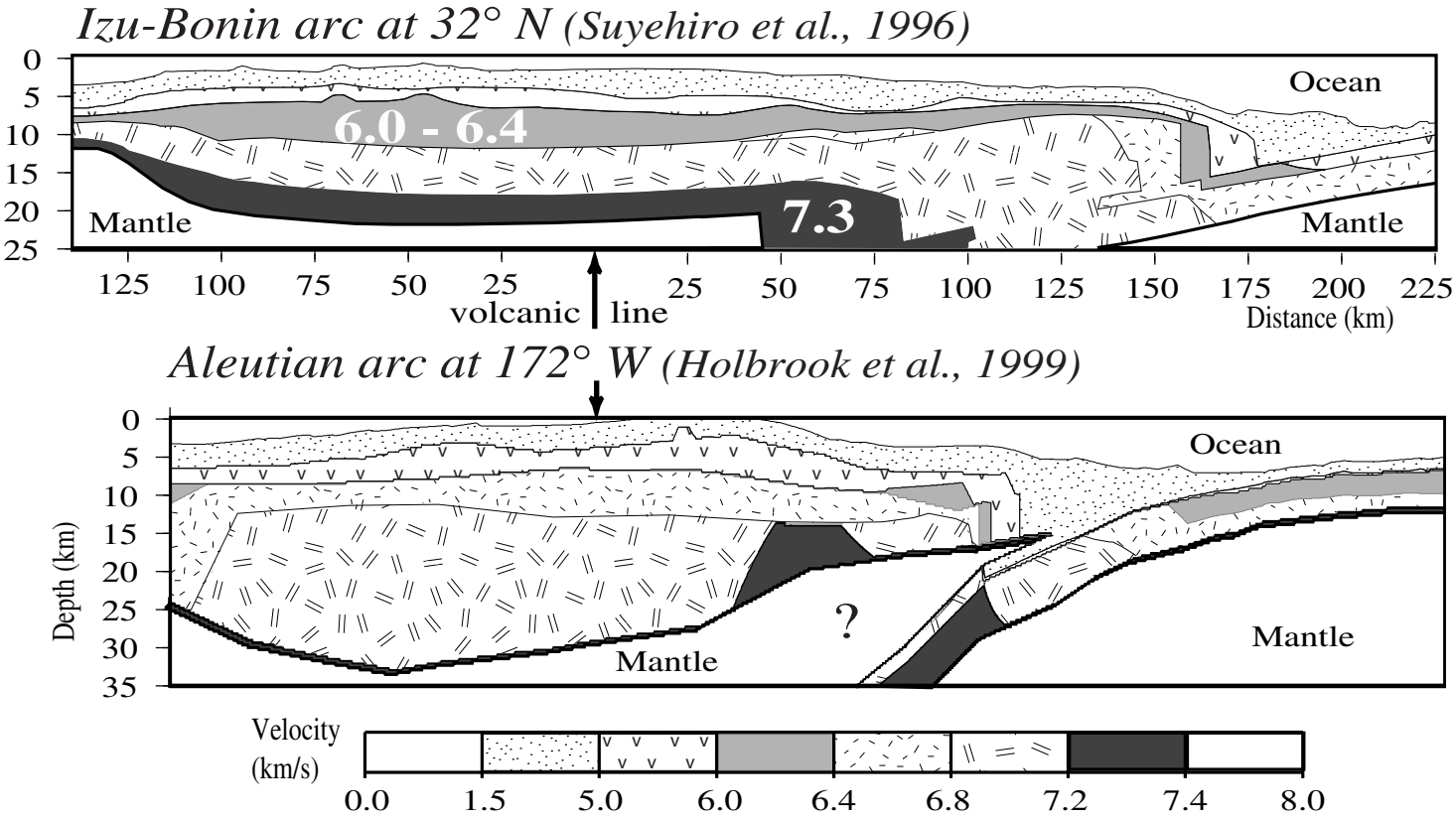


Figure 12: *Comparison of velocity structure of two intra-oceanic island arcs.*
Note Izu-Bonin is characterized by an upper-crustal, continental-like silicic-intermediate layer (gray, 6.0-6.4 km/s) and by a base-crustal highly mafic layer (dark gray, 7.2-7.4 km/s) (which is likely to be delaminated during arc accretion onto a continent). The Aleutian arc possesses neither layer, despite having a basement thickness double that of the Izu-Bonin arc.



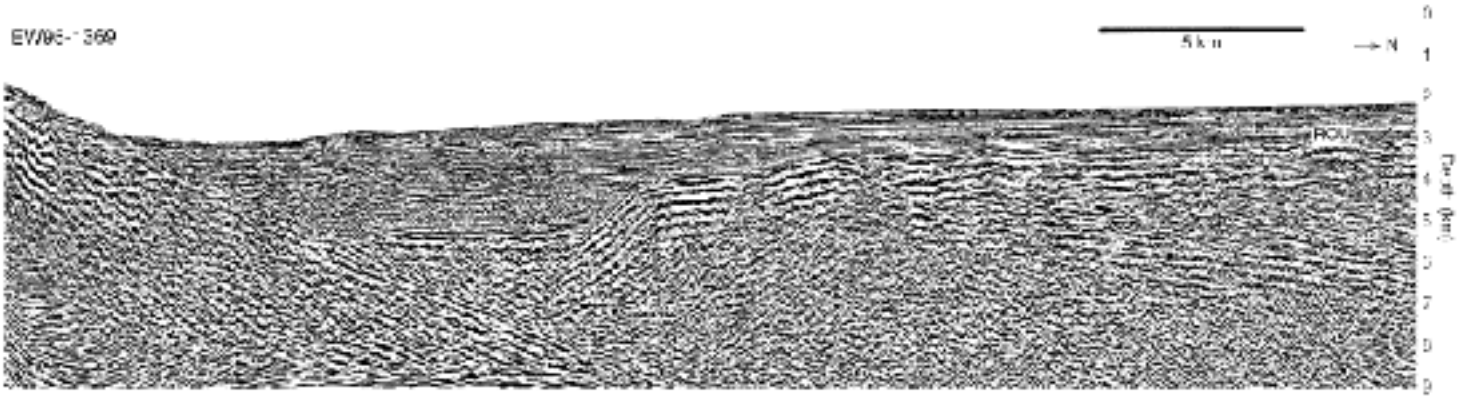


Figure 13. Stacked, migrated and depth-converted Ewing 95 MCS profile across the western Woodlark Basin rift basin [Taylor et al., 1999]. The normal detachment beneath the 2 km syn-rift section is imaged to 9 km - the base of the teleseismic zone.