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**MAGMA SUPPLY AND ITS RISE AXIS EXPRESSIONS: TIME
CONSTRAINTS FOR MAGMA SUPPLY ON THE EAST PACIFIC RISE AND
JUAN DE FUCA RIDGE**

by

Joan D. Perlin

**A dissertation submitted to the Graduate Faculty in Earth and
Environmental Sciences in partial fulfillment of the requirements for
the degree of Doctor of Philosophy, the City University of New York**

1996

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Abstract

MAGMA SUPPLY AND ITS RISE AXIS EXPRESSIONS: TIME CONSTRAINTS FOR MAGMA SUPPLY ON THE EAST PACIFIC RISE AND JUAN DE FUCA RIDGE

by

Joan D. Perlin

Advisor: Professor Kathleen Crane

The morphology and thermal structure of two distinctly different mid-ocean spreading centers (the East Pacific Rise and Juan de Fuca Ridge) are studied by morphotectonic and water column data analyses to explore variabilities in magmatic cooling processes.

The East Pacific Rise at 10°-13°N, a fast spreading ridge (60 mm/yr half-rate), is marked by robust magma chambers between overlapping spreading centers (OSCs). Analyses of the water column above this section of the EPR reveal undulations in density of water surfaces on the order of several hundred meters. Cold-dense water (colder and denser than ambient water at this depth) is found to reside (at least during the time frame of the survey) over OSCs while warm-buoyant water resides above and adjacent to bathymetrically shallow and broad sections of the rise axis. This suggests that bottom water is being drawn into highly fractured crust at OSCs. In this thesis, a model is developed to explain how entry of cold water into crust near OSCs modifies and cools the neighboring robust magma chamber by percolating along rift-parallel faults into and above the magma chamber at mid-segment, finally exiting by hot-venting into

the ocean. Results from this model predict that the timescale for cooling a magma body along this section of the EPR is approximately 7,500 years.

In contrast, magma supply along the medium spreading rate Juan de Fuca Ridge (29 mm/yr half-rate) can be shown to be intermittent on a much longer timescale of approximately 60,000 years. In this case, bathymetry, temperature and structural data from individual axial segments are used to estimate magma supply. Along the Juan de Fuca Ridge, hydrothermal areas are generally associated with closely-spaced fault zones and shallow lava fields whereas cooler, deeper areas are marked by desiccated remains of previously-active terrain. A near-surface continuously elongate magma chamber has not been detected beneath the Juan de Fuca Ridge so such a chamber cannot supply the axis with a continuous magma source. It is therefore proposed that either magma rises from depth at intermittent intervals or near-surface magma supplies are extremely localized point sources which must migrate along the rise axis.

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Thank you to all the people who helped, understood or just plain listened through the process, with a special thanks to my husband.

This work is dedicated to my friend and mentor, Bob Goldstein, whose support and encouragement made this endeavor possible.

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I. INTRODUCTION

A. Background

The theory of continental drift and plate tectonics was recognized by some scientists as early as the nineteenth century. The puzzle-like fit of the continents on either side of the Atlantic was recognized by Francis Bacon as early as 1620. That lithospheric plates moved around was seen as a convenient way to explain similarities in fossil fauna and flora in places separated by oceans. It was also useful to explain why evidence of ancient glaciation was observed in present-day equatorial areas. Paleontologists, however, took both sides of the continental drift argument.

Finally, in 1928, Arthur Holmes invoked thermal convection in the mantle as the mechanism for continental drift. He proposed that subcrustal convection currents "dragged the two halves of the original continent apart, with consequent mountain building in the front (*of the currents*) where currents are descending, and the ocean floor development on the site of the gap where the currents are ascending." (Press and Siever, 1982). Following World War II the Mid-Atlantic Ridge was mapped and its rift valley was discovered. Consequently, Harry Hess of Princeton University suggested that new sea floor formed by upwelling of mantle material in cracks at divergent sea floor boundaries, called mid-ocean ridges or spreading centers. Work on the magnetic reversal patterns on either side of the divergent plate boundaries by Vine and Mathews (1963) helped to explain Hess' concept, and thus during the 1960s it was theorized that extension occurred uniformly and at a steady state along the respective mid-ocean ridges.

In the 1970s, when definitive proof of hydrothermal activity on the seafloor was obtained (Corliss et al., 1979), marine geologists began to realize that magma intrusion, spreading and extension along the mid-ocean ridges was highly variable over space and time (Ballard et al., 1981, Francheteau and Ballard, 1983 and Crane, 1985). Bathymetry, seismic data and physical oceanography revealed that the mid-ocean ridges were thermally, magmatically and structurally segmented where magmatic centers were spaced like a string of beads along the rise axis (Crane, 1985, Schouten et al., 1985, Langmuir et al., 1986 and Langmuir, 1987). In the 1980s, the application of multibeam echosounding and wide ranging side looking sonar systems allowed for the high resolution investigation of much larger regions of the sea floor than had previously been possible with single-beam echosounders. Overlapping spreading centers (OSCs), which are located at places on the East Pacific Rise (EPR) axis where two portions of the rise axis are situated laterally next to each other, were detected in the process (Macdonald et al., 1984). These were indicators that the rise axis might be structurally as well as magmatically segmented. Structural segmentation of a rise axis is based on differences in morphology and axial structures. The magmatic segmentation of rise axes is related to differential magma supply at places along the rise axis. Langmuir (1987) proposed that OSCs are found at deep, magmatically deprived places. In addition, Langmuir et al. (1986) and Langmuir (1987) suggested that shallower bathymetry can be associated with a well developed axial structure devoid of OSCs, while deeper bathymetry results from a lesser magmatic budget. Controversy still remains about the

wavelengths of structural versus magmatic segmentation and whether or not segmentation is driven by deeper or shallower forces. In this thesis I bring the role of the overlying ocean into the segmentation story and investigate the geometry and the degree to which the percolation of sea water into the mid-ocean ridge may contribute to the cooling of magma bodies along the rise axis.

I will also explore the magmatic time scales on two different mid-ocean ridges (the fast-spreading East Pacific Rise at 10°-13°N and the medium-spreading Juan de Fuca Ridge - JdFR) and from these analyses make hypotheses about magma rates of supply and morphology of spreading centers themselves (Figures i-1, i-2, i-3 and i-4).

B. Hydrothermal Activity

Hydrothermal fields are areas along rise axes where anomalously hot water is being discharged. In the spring of 1977 ridge crest hydrothermal fluids were first sampled by submersible on the Galapagos Spreading Center (Corliss et al., 1979). Two years later hydrothermal solutions of 350°C were found on the EPR at 21°N (Macdonald et al., 1980a). Since then such heat anomalies have been found at other places along the EPR and on many other spreading centers as well.

The study of hydrothermal fields as discharge zones has given some insight into the hydrothermal phenomenon. The zones are elongated in shape, ranging in length from 15-20 km at 8°-14°N on the EPR (Crane, 1985). Elsewhere, such as at locations on the Juan de Fuca Ridge, discharge zones have also been found to be elongate,

perhaps because the vents fields extend for several hundred meters along strike (Feely et al., 1987). Large-scale hydrothermal venting perhaps obscures fine scale fracturing and related structures. However, some fracturing must exist to allow for escaping hot, hydrothermal materials.

While regions of hydrothermal discharge are dramatic, concentrated, and relatively easy to find because of their high temperatures, mineral contents and their association with shallow regions on the rise axis, much less can be said for zones of recharge. It has generally been proposed that steady-state recharge is occurring all along and across the rise axis. This, however, is a rather simplistic view. Recharge taking place over a large region would necessitate a crust which is everywhere similarly permeable to water. Lateral changes in the seafloor near the rise axis, and changes at depth in the seafloor, would certainly tend to argue against slow, regional water recharge (Lister, 1974, Crane and Normark, 1977 and Haymon and Macdonald, 1985).

The sea floor itself is, of course, permeable to water inflow. There is a certain micropermeability that has been perceived to be widespread and perhaps homogeneous and isotropic, particularly beneath the centermost zone of extrusion (Elder, 1967). This idea lends itself to the widespread, diffuse theory of hydrothermal circulation and recharge. However, the sea floor is not, in reality, homogeneous (Crane and Normark, 1977, Haymon and Macdonald, 1985 and Rosenberg et al., 1993) and it exhibits macropermeability in the form of faulting and fissuring which should allow for the inflow of water and serve as conduits for hydrothermal circulation.

Indeed, fault variations across rise axes and along-axis indicate the crust's heterogeneity. Faulting is concentrated at rise axes where fracturing and fissuring is more intense than it is off axis. How magma centers, rise axis faulting, and segmentation interact with the surrounding ocean, and the resultant time scales of segment heating and cooling is a focus of this thesis.

C. Volcanic and Structural Segmentation of the Mid-ocean Ridge

The structural features of the rise axis serve as a major guide to ridge segmentation. While the features described are undoubtedly important, the nature of rise axis tectonic structures provides the key to the processes occurring beneath the rise.

Theories describing rise axis segmentation suggest differences in morphologic, structural and volcanic properties along-axis. The reasons for these differences derive from differences in melt source and availability and resulting differences in spreading processes. Segmentation occurs on spreading centers around the world. Crane (1985) and Schouten et al. (1985) discuss the fact that the spreading center cells, which generally occupy the centers of axial segments, suggest a three-dimensional process of magmatism and crustal accretion. These spreading center cells are spaced evenly along the axial plate boundary, with the cellular structure being on the order of 30-80 km in length for slow spreading ridges and from 90-200 km on the medium to fast spreading ridges (Crane, 1985, Schouten et al., 1985, and Crane et al., 1988).

The variability of rise-axis morphology suggests a variability in the rise axis magmatic budget (Francheteau and Ballard, 1983, Crane,

1985 and Schouten et al., 1985). How a variable magma budget affects the rate of formation of oceanic crust or the deformation of the crust by faulting and spreading (or both) is not well understood. Crane (1985) proposed that the axis of the mid-ocean ridge was both magmatically and structurally segmented at wavelengths differing sometimes by more than a factor of two. Magmatic segmentation was proposed to be dominated by deep-seated diapiric upwelling at wavelengths of about 155 km on the EPR originating from depths of 40-80 km. Crane (1987) also proposed that instability of these diapirs will promote intermediate-scale migration along the rise axis. In contrast, smaller magma cupulas were proposed to be located on top of the larger diapirs generating smaller topographic undulations with wavelengths of about 45 km (Macdonald et al., 1984 and Crane, 1987). This later wavelength corresponds to the wavelength of tectonic segmentation (as illustrated by the spacing of overlapping spreading centers) (Macdonald et al., 1984 and Macdonald and Fox, 1983). Crane (1987) proposed that the tectonic segmentation could be generated by a deviatoric state of stress imposing a preferred cracking wavelength in the crust or by thermal cooling stresses varying along the rise axis.

Macdonald et al. (1988a) discussed segmentation on several scales: a) first-order segmentation of the ridge at intervals of about 300-500 km, with a range for the types of discontinuities (major transform faults and propagating rifts) encompassing 100-1,000 km; b) second-order segmentation, as defined by non-rigidly behaving offsets on a length scale of 50-300 km; c) third-order segmentation (30-100 km length scale) and d) fourth-order segmentation on a

scale of 10-50 km. Macdonald et al. (1988a) describe the segmentation on the scales discussed here as both structural and magmatic. Structural features such as ridge discontinuities including major transforms have been associated with disruption of the sub-axial seismic reflector (which is interpreted to be an acoustically reflective top of an axial magma chamber) which in turn indicates the general lack of magma at discontinuities. Fourth- third- and second-order discontinuities are superimposed between first-order discontinuities. It is quite reasonable to infer that if discontinuities are marked by a severe lack of magma, it should also be possible to delineate the length of axial segmentation based on the dimensions of the regions of maximum magma supply (Crane et al., 1988).

Langmuir et al. (1986), in a discussion of petrological and tectonic segmentation of the East Pacific Rise, stated that wavelength of rise axis segmentation is a smaller scale than the 300 km which is delineated by large fracture zones or similarly large disruptions of the rise axis. This smaller scale segmentation is based on bathymetric undulations on the East Pacific Rise with a wavelength as small as 25 km.

D. Lithospheric Plate Thickness at Divergent Plate Boundaries

The spreading rate of mid-ocean ridges is a manifestation of crustal accretion due to injection and extrusion of magma/lava at the edges of the two accreting plates. This spreading rate varies considerably for different parts of the mid-ocean ridge system. The slow-spreading Mid Atlantic Ridge (MAR) spreads at a rate of about 2.5 to 4.1 cm/yr. The fast-spreading East Pacific Rise (EPR) spreads

at a rate of 9 to 11 cm/yr at the Pacific/Cocos Plate boundary. The spreading rate at the Juan de Fuca Ridge is intermediate and is about 6 cm/yr.

Seismic data from earthquakes on the MAR indicate lithospheric thickness of five kilometers or more (Noltimier, 1974). Lithosphere thicker than five to eight kilometers is cool and more brittle than thinner lithosphere and should support rigid deformation: the rigid interactions form transform faults and fracture zones that scar the ocean floor and fragment the MAR (The Open Univ., 1989).

Lithosphere at and adjacent to the EPR is thin, hot and weak due to its young age and is presumably not rigid enough to maintain a transform fault configuration (Macdonald and Fox, 1983). Thin lithosphere may be as thin as two to three kilometers on average on the EPR (Macdonald and Fox, 1983) and is thought to be less than 0.25 million years old. Lithosphere thicker or older than this may be able to support rigid deformation (as is found on the MAR).

Because of its fast spreading rate, the EPR creates more crust per unit time than a slower spreading ridge. This crust will be much younger (hotter and thinner) at a particular distance off axis than crust at the same distance created at a slow spreading ridge. Thus, the flanks of the slow spreading MAR are comprised of much thicker, cooler and stronger lithosphere at comparable distances off-axis. Crust at and adjacent to intermediate spreading rate ridges does not seem to be as thin and weak as that found near the EPR, thus the types of structures found at medium spreading rate ridges are not those commonly associated with thin, weak crust (such as OSCs).

E. Nature of Faulting On and Off the Rise Axis

Faulting styles on the rise axis may differ from those that occur off-axis. Faulting at the rise axis is formed by axial extension processes and many investigators have proposed the manner in which this faulting occurs and the genesis of off axis faulting.

A good location to investigate spreading center morphologies and processes is Iceland where a divergent plate boundary is superimposed on a large hot spot (Einarsson, 1991). Einarsson (1991) has found that on Krafla, a central caldera of the Northern Volcanic Zone in Iceland, two processes are at work: there is inflow of magma into the crustal magma chamber and rifting at the plate boundary. Tectonic stress builds up slowly with time due to plate movements. Magma flow into a magma chamber can trigger the release of tectonic stress and thus initiate rifting. Fissures then propagate away from the magma chamber to form dikes. Einarsson (1991) further states that magma inflow causes inflation of the crust which leads to extension and normal faulting in the roof above the magma chamber. Deflation events lead to contraction and reverse faulting.

Rubin and Pollard (1988), in investigating dike-induced faulting in Iceland, have concluded that "observation of ... subaerial volcanic rift zones have demonstrated that ... formation of new fissures occur(s) during episodes of dike emplacement." Further, Rubin and Pollard (1988) state that intrusion of magma plays an important role in developing rift-zone geomorphology and topography. The graben floor is seen as representing a zone of pervasive inelastic deformation and there is evidence "the dike

intrusion triggers faulting, and not vice versa." (Rubin and Pollard, 1988).

Faulting on the East Pacific Rise axis as described by Edwards et al. (1991) has been evaluated by determining the width of the fault zone at the rise where faults are being created. Large offset faulting is found to be created within a few kilometers of the rise crest. Thus, Edwards et al. (1991) conclude that the width of active fault formation is located within a few kilometers of the East Pacific Rise crest. It is also proposed that in addition to large-scale plate stresses which affect the tectonic fabric at and surrounding the East Pacific Rise, magmatic inflation and deflation in response to changes in magmatic budget may contribute to local variations in fault parameters (Edwards et al., 1991).

Crane and Ballard (1981), in an investigation of cyclic evolution of the Mid-Atlantic Ridge, suggested that fissures and fractures may actually occur after magmatic injection is complete. They proposed that initial volcanic extrusion covers up previously formed fissures. Thus, after volcanic activity has been reduced, cooling around magma intrusions causes new fissure formation. Further, new fractures may open which allow for egress of hydrothermal waters (Crane and Ballard, 1981).

According to Edwards et al. (1991) new off-axis faulting in the region of the East Pacific Rise is unlikely. Edwards et al. (1991) agree that it is fair to assume "that faults on the flanks of the EPR are created near the rise crest..." Other off-axis expressions of faulting which initially occur on the rise axis itself are abyssal hills. It appears as if the "inward facing slopes of the abyssal hills are well

correlated with the location of faults while the outward facing slopes are not." (Edwards et al., 1991). Edwards et al. (1991) have found the inward facing faults on the flanks of the EPR affect 20% more of the ocean crust than do outward facing faults. Thus, a half-graben model of abyssal hill formation appears to be favored based on the analysis of faulting both on the rise axis and off the rise axis (where the abyssal hills form due to the tectonic fabric produced at the axis) (Edwards et al., 1991).

II. MORPHOTECTONIC FEATURES OF THE EAST PACIFIC RISE

A. Morphology of the East Pacific Rise

The morphology of the East Pacific Rise is marked by segmentation of the rise axis on varying scales. The largest scale of segmentation is marked by features such as fracture zones or transform faults which occur at a spacing of hundreds of kilometers (Macdonald et al., 1988a). Segmentation on progressively smaller scales is determined by the spacing of features like OSCs, and other non-rigidly behaving axial features.

The ridge crest of the EPR in the region of 8°N and 13°N shoals and deepens along strike at a wavelength of about 155 km (Crane et al., 1988). There are small highs of 10 to 50 km in length superimposed on the 155 km wavelength morphology. These smaller highs are truncated by OSCs which have offsets as small as 1 to 2 km and as large as 10 to 15 km (Crane et al., 1988). Near OSCs, where the rise axis deepens, the cross-sectional shape of the rise axis is generally triangular. Other regions are domal in cross-section (Crane et al., 1988). Water depth variation along the axis is commonly 50 to 100m. However, the greatest depth variations (>700m) occur near the intersection of the Clipperton Transform Fault (located between 10°00'N and 10°30'N on the EPR) and the EPR axis.

The axis of spreading on the EPR between 8°N and 13°N is characterized by an axial extension zone which is characterized by a narrow band of braided, tightly-spaced en-echelon faults and fissures (Crane et al., 1988). There are places where no graben-

forming faults are apparent, but the edge of the axial fault zone is still typically sharp and well defined (Crane et al., 1988).

B. Overlapping Spreading Centers and Related Axial Structures

Overlapping spreading centers (OSCs) were reported as a new type of plate boundary interaction by Macdonald et al. (1982). This feature was discovered in a SeaBeam study of the EPR from 8°N to 18°N. Spreading center segments were found to "overshoot" each other and curve toward one another. Fox et al. (1983) describe the structural relationships of OSCs as follows: 1) the slopes that flank each OSC are dominated by volcanic, constructional terrain disrupted by occasional rise-axis parallel faults, 2) the overlap basins are floored by volcanic products and are devoid of structures suggestive of strike-slip deformation, 3) rise axis parallel structures that flank the OSCs are continuous across the zone of overlap, and 4) as OSCs are approached, the depth of the rise axis increases reaching local depth maximums directly at the OSCs on along-strike depth profiles.

OSCs are comprised of two segments of overlapping cusped neovolcanic zones with the overlap basin in between. The original lateral offsets between these zones is commonly less than 15 km (Macdonald and Fox, 1983). Macdonald et al. (1988b) also note that offsets at OSCs are less than 10-15 km. Neovolcanic zones are oriented like two letter "C"s facing each other. The interior overlap basin is elongate subparallel to the neovolcanic zones and in most instances has a length to width ratio (L/W) of 3.0 ± 0.5 . Hummocky volcanic terrain dominates the floor of the OSC basin. In some cases

volcanic cones up to 20-80 m high can be found (Macdonald et al., 1984).

Crane (1987), using SeaMARC I data to evaluate structural evolution on the East Pacific Rise from 13°10'N to 10°35'N, describes the quasi-continuous 250 m to 1.5 km wide zone of axial extension which is generally defined by "closely-spaced fissures and small offset faults numbering more than 6/km in a direction orthogonal to the ridge axis..." Rift tips within the axial extension zone associated with overlapping spreading centers (OSCs) are described by Crane (1987) as having a "density of faults of 10-12.5 faults/km." The zone of axial extension represents the youngest, hottest and thinnest crust on a rise axis. Crane (1987) has described this zone as extensively faulted and fissured and notes that lineaments of the innermost zone of extension are most likely fissures without discernable vertical offsets.

As OSCs may be indicators of rise axis discontinuities that separate magma diapirs, Crane (1985) and Whitehead et al. (1984) proposed that Rayleigh Taylor type magma instabilities may rise harmonically along the axis, translating and migrating through time. Recent along-axis magma flow may create the short propagating axial segments that approach one another, creating OSCs in the process. The degree of lateral propagation determines the degree to which one axial segment may overlap another. Crane (1987) and Macdonald et al. (1988a) applied this theory to explain harmonic spacing of axial highs (interpreted to be magma centers) on the EPR. Between these magma centers the crust is thought to be relatively cool and probably not well-supplied by magma (Langmuir, 1987).

By using seismic tomography McClain et al. (1985), suggested that the depth to the near surface elongate magma chambers is often on average as shallow as 1.1 km. Seismic refraction and reflection data also suggest that the depth to the elongate magma chamber both rises and falls along the rise axis, inferring that recent magmatic intrusion and extrusion occur at the shallow locations and that cool spots may coincide with OSCs (Langmuir et al., 1986 and Sempere and Macdonald, 1986a). They also propose that the position where the magma from two different cells meet is marked by a low point on the axis. In the opinion of Sempere and Macdonald (1986a), OSCs occur where the axial lows do not meet head on and are thus nodal points for volcanic segmentation.

Overlapping propagating neovolcanic zones at OSCs are associated with a high density of faulting and fissuring (Crane, 1987). Fissures and fractures at rift tips are due to a migrating magma front and associated changes in deviatoric stress where dike injection induces faulting and fissuring. Additionally, cooler (brittle) crust at OSCs supports the fracturing found at rift tips.

The East Pacific Rise axis is segmented structurally as well as volcanically, and many different structural wavelengths are represented (Crane, 1987). Macdonald et al. (1988a) divided the East Pacific Rise into segments of varying lengths ranging from 300-500km to the smallest scale of 10-50 km. This segmentation is based on both magmatic budget and the structural breaks in axial continuity which mark the places on the rise axis which are magma starved. The longer wavelengths are marked by major structural interruptions of the axis such as fracture zones. OSCs and other

smaller axial deviations mark the boundaries of the smaller wavelength segments. In many cases it is difficult, based on present-day data, to determine whether offsets in the rise axis are structurally controlled or volcanic breaks in the mid-ocean ridge system.

C. Magma Chambers

The existence, dimensions and stability of magma chambers beneath the EPR has been a topic of interest since the early 1970's. Cann (1974), Orcutt et al. (1975) and Rosendahl et al. (1976) believed the axial magma chamber to be a large body in excess of ten km in width. A wide magma chamber was also hypothesized by Pallister and Hopson (1981), while McClain et al. (1985) reported (based on seismic refraction investigations) a narrow magma chamber about four km wide at the EPR near 12°50'N. In order to clarify which was the more likely chamber width and overall shape, Pallister and Hopson (1981) studied the Oman ophiolite and determined that this ancient magma chamber had a half width of about fifteen kilometers and that it probably was a good analog to the present-day magma chamber underneath the EPR. Sleep (1975) and Kuszniir and Bott (1976) also have reported that intermediate to fast spreading ridges have magma chamber half widths in excess of ten km. On the other hand, Lewis and Garmany (1982) using seismic refraction data suggest that a steady-state chamber may not even be present and if it is present, it does not exceed two kilometers in width. Part of the controversy may lie in the size of the "sub seafloor" footprint that the surveys were seismically imaging. In the earlier cases large area

surveys were carried out and data were integrated over great distances. More recent investigations have aimed at looking at shallower perturbations in the depth of the magma chamber (and thus at a small "footprint"). Langmuir et al. (1986) revealed that between 5°30'N and 14°30'N along the EPR, the shallow magma supply is discontinuous, as evidenced by distinct differences in magma petrology along the axis. Although magma chambers have been generally seen to be quite efficient magma-mixers (O'Hara, 1977), systematic differences in geochemistry between axial lows and highs of all types on the EPR indicate that efficient mixing is not occurring.

Detrick et al. (1987), using multichannel seismic profiling of the EPR between 8°50'N and 13°30'N, found that crustal magma chamber roof depths are located from 1.2 to 2.4 kilometers below the seafloor. From this survey, chamber widths of four to six kilometers could be traced for tens of kilometers along-axis. Generally, though, widths are less than three or four kilometers. Shallow portions of the rise axis, as at the areas between OSCs, exhibit thin crustal lids and are continuous for tens of kilometers. Deep parts of the rise axis, as at the tips of the OSCs at 12°54'N and 12°35'N, show a gap in the axial magma chamber reflector, where the crustal lid is thick. At other axial lows the chamber is either discontinuous or exhibits a prominent dip in the axial reflector. Thus, at OSCs, the magma chamber may be physically disrupted. Indeed, this is compatible with the theory of magma supply at OSCs and other axial lows. However, if this undulating top of an axial magma chamber is really the surface expression of small magma cupulas which upwell from a

much larger and deeper seated magma body, then the "small, shallow footprint" results may be compatible with the "large, deep footprint" results. As an example, Burnett et al. (1989), by studying P-wave travel times, deduced that the shape of the magma chamber below the 12°40'N uplift is about fifteen kilometers wide and three kilometers deep. Thus, different investigators may simply be looking at different depths and different scales of the same magma body.

III. EAST PACIFIC RISE: HYDROLOGY AND STRUCTURE - 12°30'N TO 13°00'N

A. Instrumentation, Methodology and Data Collection

The data set used for the investigation of the East Pacific Rise includes SeaMARC I data collected on the EPR in 1983 (Crane, 1987), SeaMARC II data collected in 1988 (Perram and Macdonald, 1990) and water column temperature and salinity data collected by K. Crane in 1983 and 1986 (during the HYDROFAST expedition) using sensors on dynamic hydrocasts, CTDs (conductivity, temperature and density) and thermistor chains (Crane et al., in prep.) (Figures epr-1 through epr-19).

SeaMARC I AND SeaMARC II are side-looking sonar systems operated via research ships to survey and map the seafloor. They measure backscattered echo intensity on each of a number of narrow beams. Near-vertical incidence of insonified beams results in no useful backscatter being obtained directly beneath the ship. Thus, blank strips showing no image appear along the center or sides of side-scan swaths. SeaMARC I was a 30 kHz side-looking system which surveyed a swath of seafloor up to 5 km in width. The vehicle was towed 300 m above the bottom at about 3 km/hr. Seafloor objects less than a meter in amplitude and spaced less than 20 m apart were resolvable (Vogt and Tucholke, 1986).

SeaMARC II operated as a 11-12 kHz side-looking sonar and as a swath-bathymetric mapping system which was towed at a depth at least 50 m below the surface (Davis et al., 1984). Bathymetry was accurate to within about 3 percent of water depth (Vogt and Tucholke, 1986) while swaths of seafloor up to 10 km in width (with

a resolution of 5 m cross-strike) are imaged by its side-looking sonar (Davis et al., 1984). Side-looking sonar images can reveal different features of the seafloor by backscatter variability. These features include volcanic flows, faults, fissures and sedimentary deposits.

Isopycnals in the water column above the rise axes were constructed through analysis of CTD and thermistor chain data. In addition, a hydrocast built at IFREMER (Institut Français pour l'Exploitation de la Mer), called the Palanque Dynamique (PLD) was used to laterally sample near bottom water over the East Pacific Rise axis (Bougalt et al., 1990 and Charlou et al., 1991). Each PLD hydrocast collected forty bottles of water on each near bottom tow. The PLD consisted of four cages spaced 50 meters apart and was designed such that each bottle would sample over a distance of one nautical mile. Each cage contained ten bottles which filled with water at a rate proportional to ship speed. A heat flow instrument was also attached to each cage. Temperature was measured using four high resolution sensors from conductive heat flow instruments and an attached transmissometer mapped the concentration of particulate matter.

Most of the geochemical samples collected by the PLD and the CTD were processed on board. More than 600 samples were analyzed for SiO_2 , O_2 , salinity, CH_4 and Mn. Measurements for the concentration of ^3He were carried out in Paris by M. Merlivat and P. Jean-Baptiste (Bougalt et al., 1990 and Charlou et al., 1991).

Additional thermal data were obtained by towing two 50 meter long Aanderaa thermistor chains at one time beneath the SAR side-looking sonar. The sonar was navigated using a bottom moored

transponder network and an on-board Global Positioning Satellite System receiver. Each thermistor chain provided eleven temperature measurements between 0 and 60 meters above the sea floor (depending on the altitude of the SAR), with resolution of $\pm 0.01^{\circ}\text{C}$. In addition, thermistor chains were connected to the PLD for one deployment in order to compare the measurements with temperature data collected by heat flow instruments attached to the cable, thus allowing for intercalibration of the thermal data sets.

Results from the PLD allowed Crane et al. (1989), Bougalt et al. (1990) and Charlou et al. (1991) to construct three dimensional chemical, temperature and density maps of the near bottom water from data collected both along and across the strike of the rise axis. While the PLD was operated on ten tows of ten nautical miles each, these were supplemented by twelve vertical CTD profiles spaced strategically to provide vertical constraints on the data as well as off-axial reference values (Figure epr-3). Using all the data collected, three dimensional potential temperature and density maps were constructed for the region between $12^{\circ}25'\text{N}$ and $12^{\circ}57'\text{N}$ (Crane et al., in prep.)

Crane et al. (in prep.) have made estimates of heat flux from hydrothermal fields utilizing thermal data from near-bottom temperature analysis. Crane et al. (1988) reported findings which quantify near-bottom water flow rate and direction. These data, in addition to sonar, CTD and thermistor data are used to model the near axial hydrographic structure.

B. Physical Oceanography and Hydrographic Structure

Figure epr-1 is a potential temperature transect along the rise axis of the EPR based upon data compilation for the years 1974-1983. Hot springs are noted at three places and generally correspond to elevated water column temperature. Crane et al. (1988) describe no short term preferential direction of bottom water flow between 12°40'-50'N. However, bottom water flow measured over 235 days showed a weak mean flow of 1.2 cm/s directed west of south. Current measurements suggested an anticyclonic motion centered around vents at 12°50'N (Crane et al., 1988). CTD casts and current meter observations collected by K. Wishner (pers. comm. via K. Crane) on and near seamounts east of the rise axis suggest large scale westward transport, in agreement with Crane et al.'s (1988) earlier current meter results. Water flow data, especially data which describes sustained overall direction of large scale transport over time, can be useful to help predict the direction of water supply to recharging hydrothermal cells.

Temperature anomalies found by Crane et al. (in prep.) between 12°30'N-13°00'N (shown in Figure epr-2) are from CTD data (Crane et al., in prep.) which were collected both on axis from 11°45'N to 13°20'N and off axis, for comparison purposes. Figures epr-3, epr-4 and epr-5 depict a comparison between potential temperature, salinity and density between the reference CTD (HF51), located 55 km to the west of the ridge, and axial CTDs. (Potential temperature represents the temperature that a parcel of water would obtain if it were brought to the surface). Temperature anomalies are calculated relative to background/reference potential

temperature on the same isopycnal (equal density) surface. ΔT 's (temperature anomalies) compare potential temperature/potential density relationships on the rise axis with the reference CTD. The heat emanating from volcanic activity at the rise axis and the undulating topography of the axis disrupt the temperature profile.

Where the isotherms are most elevated (a negative anomaly corresponding to an accumulation of cold water) coincides with the two OSCs at 12°35'N and 12°54'N and their flanks while positive anomalies are centered near 12°32'N, 12°40'N and 12°40'-43'N where the warmest water occurs (Figure epr-17).

Between 12°40-43'N (centered on the axis at 12°40'-12°41'N) a positive potential temperature anomaly of +0.08°C was detected primarily on the 27.72 isopycnal (Crane et al., in prep.). This high temperature core rises to 2500 m (100 m above bottom) where the anomaly spreads laterally along the axis both north to 12°47'N and south to 12°34'N. The +0.02°C contour is found as high as 2400 m (200 m above bottom) as the warm water rises as a plume. The thermal (hot water) plume levels off between 150-220 meters above the bottom and is bent in a southerly direction and thus it appears as if it were being blown away from its source. Depressed isotherms and isopycnals extend onto both sides of the rise axis where a positive temperature anomaly of +0.04°C can be found out to 8 km both east and west of the axis while the +0.08°C anomaly is constrained to within 1.75 km of the axis in the E-W direction. The absence of any other clear positive temperature anomalies suggests that during the HYDROFAST survey only the one vent field in the

region of 12°41'N was active on this segment of the ridge (Figure epr-17).

Negative potential temperature anomalies as great as -0.04°C were discovered at two anomalously cold regions located above and on the flanks at the OSCs at 12°35'N and 12°54'N. The northern cold water anomaly of up to -0.04°C occupies a region which extends both east and west of the OSC, capping the ridge in a mushroom shape. This anomaly reaches a thickness of 150 m, is 5.5 km wide, and rests near the seafloor at and below the depth of 2600 m (Figure epr-17). The cold water anomaly of -0.02°C above the southern OSC extends from a depth of 2650 up to 2590 m (60 m).

The distribution of isopycnals illustrates the potential density of water in a three-dimensional manner. Elevated isopycnals are indicative of denser water in a region and are generally found at the two OSCs (where nearly 70% of the cold-dense water regions are located) and where depth exceeds 2700 m. By contrast, the off axial reference depth for the 36.94 isopycnal is 2540 meters. Above the two OSCs the 36.94 isopycnal is much shallower (Figure epr-16). Other elevations of the 36.94 isopycnal are found at 12°40'N and 12°45.5'N-12°47.5'N, although none of these elevations rises above the reference depth. At latitudes 12°52-54'N (the northern OSC) high density water (elevated isopycnals) occur on both sides of the axis. At 12°50'N, however, lower density (warmer-more bouyant) water appears to accumulate above the rise axis as the isopycnal depth (for the 36.94 isopycnal) is 2630m. The vent field at 12°41'N corresponds to a distinct deepening of the 36.94 isopycnal (to depths > 2600 m) (Figures epr-9, epr-16).

Water chemistry reveals hydrothermal activity in two regions (12°50'N and 12°38'N - 12°47'N) which are characterized by concentrations of CH₄ 50 times the concentration of normal bottom water (Charlou et al., 1991 and Crane et al., in prep.) (Figures epr-10 and epr-11). This is in contrast to the single positive temperature anomaly suggestive of a hydrothermal plume centered at 12°40'-12°41'N. The methane anomalies are located on or near the seafloor at both 12°50'N and 12°45'N. In contrast, at 12°38'N, closer to the thermal plume, the methane anomaly is suspended 150 m above the bottom. Thus it was either produced elsewhere on the rise axis and advected downstream with the ambient currents or its production mechanism is intermittent and on time scales short with respect to the dissipation time scale of methane in bottom water.

The mismatch between the temperature (dynamic) and chemical (passive) signatures of an active hydrothermal field have been noted by Speer (1989). Here, there is a separation of up to 5 km in the N-S direction between the thermal and chemical anomalies measured on the rise axis. Recent Alvin dives at 9°30'N on the EPR (Haymon and Fornari, pers. comm. to K. Crane) recount the discovery that diffuse lower temperature vents are far richer in methane than higher temperature smoker fields. They suggest that vigorous bacterial communities living on low temperature vents may be responsible for the elevated methane profiles. The higher temperature region near 12°41'N (a high temperature (smoker?) field) may thus correspond to the lower methane level near 12°38'N (which may have originated near 12°41'N and been further transported by currents).

The region surrounding 12°40-41'N is characterized by warm water, a thermal plume, a positive isopycnal anomaly and high concentrations of CH₄. These indicate a region of emanation of hot, hydrothermal fluids. The OSCs at 12°35'N and 12°54'N are marked by cold, high density water which may represent the inflow of water to the rise axis to recharge hydrothermal circulation cells.

C. Faulting Along and Across the Rise Axis

Faulting on the East Pacific Rise can be described both along and across the rise axis. Figure epr-13 illustrates the major seafloor structures in the region of study as derived from SeaMARC I and SeaMARC II sonar images. Figure epr-16 shows the relationship of axial fissure frequency to latitude, depth to the axial magma chamber and depth to the 36.94 isopycnal. The OSCs lying below cold, dense water are marked by a general increase in faulting near the rise axis. Fissure numbers in the axial regions of the OSCs at 12°35'N and 12°54'N are about fourteen, whereas between the OSCs there are from one to seven fissures in the axial zone. South of the 12°35'N OSC fissure number is less than or equal to six. North of the northern OSC, fissure numbers reduce, followed by an increase to fourteen at 13°N, outside the geographic location of this study. The difference in fissure numbers along the rise axis, with the increase at OSCs, is indicative of heterogeneity of the rise axis crust.

A detailed analysis of faulting on the EPR as it varies with distance east and west off the rise axis is presented in Figure epr-20. Generally, in the southernmost part of the survey region, most faulting is concentrated within the central axial kilometer from

12°30'N northward to 12°36'N (at the southern part of an OSC) where there are up to thirteen faults in the central zone. The eastern leg of the OSC at a of latitude 12°37'N is significantly faulted with up to four faults in the 0.5-1.5 km region to the east of the center of spreading. Fault azimuths trend predominately from 345-005° in the region from 12°30'N to 12°35'N (Figure epr-21). The remainder of the 12°35'N OSC north to 12°37'N is cut by faults trending more easterly, mostly at 355-005°, on both the eastern and western legs of the rise axis. Fault lengths south of the OSC to 12°34'N are generally less than 1000 m. There are, however, a few faults up to 2500 m in length (Figure epr-22). Both the eastern and western legs of the 12°35'N OSC are characterized by fault lengths generally less than 500 m, although the western leg has some faults longer than 3000 m long.

Northward, past the OSC, from 12°37'N to a latitude of 12°40'N, most axial faulting (5-6 faults) occurs within the central axis. The exception is at 12°38'N where there is a fault density of five faults in the kilometer east of the central axis (Figure epr-20). The fault azimuths in the region from 12°38'N to 12°40'N are 335-355° (Figure epr-21). At latitudes 12°38'-39'N survey data are not complete enough to quantify fault lengths. However, at 12°40'N most faults are less than 1000 m long (Figure epr-22).

High density faulting within the central axis and the area one kilometer to the west characterizes the axis and its nearby flanks from 12°41-44'N. In this region, there are up to seven faults within the central axial kilometer and up to six faults within the one-kilometer wide region to the west (Figure epr-20). Fault azimuths

range from 335-355° with the majority of faults trending in the more westerly direction of 335-345° (Figure epr-21). Fault lengths in this area are predominately shorter than 1000 m (Figure epr-22). However, there are some faults even longer than 3000 m in this region.

The distribution of faults along the axis from 12°45'-51'N is variable both to the east and west sides of the center of spreading. Although locally, at 12°49'N, there are no faults in the central one kilometer zone around the axis, in the remainder of this region the number of faults in the central kilometer range from one to four. Within a zone one kilometer to the east of the central axis fault number varies from one to nine. In the western kilometer, fault number varies from three to eight. The only place on the axis in this region that has faulting more than 1.5 km off-axis is at 12°51'N where there are four faults in the one kilometer region 1.5 to 2.5 kilometers to the east of the central axis and two faults in the 1.5 to 2.5 km region to the west of the center of the axis (Figure epr-20). The direction of faulting at 12°45'N is predominately 335-345°, but twenty-five percent of the faulting is more westerly (325-335°). Fault azimuths from 12°46'N to 12°47'N rotate towards the more easterly direction of 335-355° and from 12°47'N to 12°50'N to 345-005°. At 12°51'N (at the southern tip of the 12°54'N OSC) all fault azimuths range from 345-355° (Figure epr-21). In this region the edge of the rift valley is characterized by faults longer than 3000 m. The rest of the faults in this region are mostly less than 500 m long (Figure epr-22).

The axis from 12°52'-55'N is dominated by an OSC. The western arm of the OSC at 12°52'N is faulted in its central, eastern and western kilometer wide zones, while the eastern arm is faulted only in its eastern kilometer wide region (Figure epr-20). At 12°53'N both the eastern and western legs of the OSC are faulted in the central and eastern kilometer wide zones, with no faulting in the western kilometer zone on either OSC leg. At 12°54'N the western leg of the OSC is faulted in the eastern, western and central kilometer wide zones, while the eastern leg is faulted only in its central axis. At latitude 12°55'N the OSC is no longer observable and there is faulting across the entire axis. Fault azimuths in this area range from 335 to 355° (Figure epr-21) and faults longer than 3000 m define the edges of the rift valley throughout the region from 12°52'-55'N. Other faults in the region are shorter than 1000 m (Figure epr-22).

D. Observations and Outstanding Questions

The structural and hydrographic data collected along this section of the EPR are useful to model water flow and circulation on and near the ridge. Water flow data are useful to help predict the direction of water supply to recharging hydrothermal cells. For instance, temperature and density anomalies in the water column indicate regions of emanation of hot, hydrothermal fluids as well as regions of proposed inflow of cool bottom water into the crust. Compilation and analysis of both the structural and hydrographic data allow realization of a three dimensional model of water flow into and through the crust at the rise axis.

IV. DATA INTERPRETATION AND MODELLING ON THE EAST PACIFIC RISE

A. The East Pacific Rise: 12°35'N to 12°54'N

The investigation of the rise axis along the EPR between 8°N and 14°N where the Pacific Plate abuts the Cocos Plate has yielded much information about the nature of seafloor spreading centers. The axis of the EPR in this region consists of a long string of elongate volcanic highs which are two to ten kilometers wide, hydrothermally active and linked in many cases by OSCs (Crane et al., 1988). Two of these OSCs are located at 12°35'N and 12°54'N and it is this axial segment that is the focus of the following volcano/tectonic investigation.

Between these two OSCs there is significant hydrothermal activity situated primarily on the widest and shoalest part of mid-segment volcanics (Crane et al., 1988). The region has been under investigation since 1979 when French oceanographers dove on and near the northernmost OSC, finding extensive hydrothermal activity on the neighboring ridge in the process (Hekinian et al., 1985). Vent chimneys are loci for the expulsion of hot water at 12°50'N where water averaging 284°C is expelled at a rate of 152 cm/s (Crane et al., 1988). In 1983 when the SeaMARC I system was used to map the volcanic/tectonic evolution of a "fast" spreading ridge the entire northern OSC was imaged and continuous near-bottom water column measurements were made (Crane et al., 1988) revealing that the bulk of the hydrothermal activity was centered not near the OSC, but close to midway (12°45'N) between the northern and southern OSC. Crane et al. (1988) calculated the volume of water emanating from

these vents to be $4.45 \times 10^6 \text{ cm}^3/\text{s}$, which corresponds to a heat flux of 5,308 MW. However, the bulk of the data were constructed from single track traverses and thus the calculated flux could be in error by a factor of ten. Crane (personal communication) revealed that an expedition in 1986 (Hydrofast) located little or no hydrothermal activity at the 12°50'N site. Instead, the major heat source in this segment of the EPR was located at 12°40'N, covering 3.3 km^2 of actively venting fields.

Based on temperature anomalies in the water column, Crane et al. (in prep.) estimate the 1986 heat flux from the 12°40'N vent site at 3,845 MW. Thus, if the heat flux and volumetric flow are linearly related, then the flow rate at 12°40'N is calculated to be:

$$\frac{3845 \text{ MW}}{5308 \text{ MW}} = \frac{\text{flow rate at } 12^\circ 40' \text{ N}}{4.45 \times 10^6 \text{ cm}^3/\text{s}}$$

$$\text{flow rate at } 12^\circ 40' \text{ N} = \underline{3.22 \times 10^6 \text{ cm}^3/\text{s}}.$$

The proportionality of flow rate to heat flow can be justified because the water temperature of emanation at both vent fields is very close. Thus, volumetric proportionality is probably valid because an established heat flux would correspond to a set amount of water at a set temperature.

B. Hydrothermal Discharge and Recharge Zones

As hydrothermal zones are uniquely associated with the seafloor volcanic/tectonic features beneath, it is logical that zones of recharge should be controlled by similar structures which allow for the easy percolation of water into the crust. If so, the key then is to

locate these structural regions which are the recharge conduits to the hydrothermal system. Indeed, if one closely studies axial features, it is possible to discern probable recharge zones. The seafloor at OSCs shows a pattern of intense fracturing. As OSCs in this region are also characterized by anomalously cold near-bottom water, it is logical to assume that OSCs are the most likely regions where water enters into the subseafloor hydrothermal system (Figures epr-12, epr-13, epr-14, epr-15 and epr-16). The structural framework at OSCs consists of rise-axis parallel faults in addition to volcanic constructional terrain. Essentially, OSCs should be the "path of least resistance" for hydrothermal recharge. Rosenberg et al. (1993) modelled flow and permeability in sea floor hydrothermal systems. They addressed isotropic, nonhomogeneous and layered situations. While these models are useful in that they produce results which show that a highly permeable top layer underlain by a less permeable bottom layer results in a narrow discharge zone, they do not adequately produce results for inhomogeneous media that are necessarily anisotropic (as is fissured and faulted sea floor). Ludwig et al. (1993) have found that porosity decreases dramatically with depth, and so perhaps layered modelling is useful to some extent. However, the faulting that occurs at and near the EPR cannot be construed as consistent with generalized micropermeability that may be found at cooler, less thermally active off-axis sea floor. Macropermeable faulting and fissuring of sea floor at the EPR allows one to assess the effects of axial processes as they relate to hydrothermal water flow.

C. Evidence for Recharge

Hydrographic structure is observed from both potential temperature and density profiles of the near bottom water above the East Pacific Rise. The cold water anomaly at the southern OSC (at 12°35'N) is illustrated in Figure epr-17. Like the hot water anomaly at 12°40'N, the difference in actual water temperature does not seem excessive. However, when it is considered that the rising hydrothermal plume at the vent's source has a temperature anomaly of only 0.08°C, and emanating water of 300°C is its source, it then becomes apparent that small deviation in water temperature over a distance of a few hundred meters in the deep ocean is indicative of extreme heating or cooling conditions at the source of the anomaly.

In addition to the temperature anomalies mapped, isopycnals (contours of constant density) have been constructed. The depth in the ocean at a certain density gives an indication of the buoyancy, and associated heat content, of the water. Thence, given that water is denser with increasing depth, increased depth to an isopycnal is indicative of more buoyant, perhaps warmer water. It follows, then, that lesser depth to the same isopycnal is indicative of less buoyant, perhaps cooler, water. Significantly, Figure epr-13 shows the OSCs to be regions of cold, dense water (the stippled pattern indicates a shallow depth of (<2540 m) of the 36.94 isopycnal). The warm plume between the OSCs indicates a greater depth to the same isopycnal. Figure epr-9 was constructed to illustrate depth surfaces above the EPR axis of discrete density layers in the deep ocean. A glance at Figure epr-9 also reveals a distinct high density anomaly at the 12°35'N OSC. In greater detail, near the seafloor, at any one

depth in the ocean, dense water is more prevalent on the eastern side of the rise axis at the 12°35'N OSC (Figures epr-9 and epr-15). At the western edge of the axis the dense water anomaly is characterized by a steep, nearly vertical isopycnal front coinciding with the onset of the rise axis graben. The linear steep-fronted geometry of the anomaly is suggestive that processes in addition to the normal physical oceanography are influencing the water column above the OSCs. The following flow model attempts to account for this geometry.

The model depends upon the following inputs: 1) structural and volcanic data extracted from SeaMARC I and II records and 2) hydrographic information collected primarily during expedition Hydrofast, 1986 (Bougalt et al., 1990). The structural features mapped by SeaMARC are presented and interpreted in Figures epr-12 and epr-13. Figure epr-14 illustrates the structure at the 12°35'N OSC which coincides with a negative temperature anomaly (Figure epr-2) and density anomaly (Figures epr-9 and epr-15) located at the same OSC.

Figure epr-17 includes a cartoon illustrating proposed recharge and discharge zones along the rise axis where cold water enters at the OSCs and then travels from cooler crust, along the faulted path of least resistance through crustal conduits towards the more robust magma chamber along the axial high. The recharging water is also thought to travel to a maximum depth defined by the upper limit of the magma chamber. Naturally, as the water courses through this plumbing system it is heated. Finally, it emerges at temperatures near 300°C and discharges forming hydrothermal fields.

The following discussion presents calculations representing the magnitude of volumetric flow and heat exchange expected if the above model for hydrothermal recharge is correct.

D. Model Components

The near-bottom water flowing westward over the East Pacific Rise (J.P. Fouchet [pers. comm] and reported by Crane et al., 1988) might get heated everywhere along the ridge, creating less dense, buoyant water at the axis. As the heated, less dense water is transported westward one could expect a linear density boundary over the axis, oriented along the strike of the ridge. Denser ambient water should be found on the eastern flank and less dense hydrothermally warmed water on the west. However, along this segment of the EPR we do not see a linear water density anomaly that extends the full length of the segment. Rather, we clearly see one heated area in the water column between two regions of high density cold water located directly over the OSCs. Why would this be the case? Either convection due to entrainment into the hydrothermal vent or recharge into OSCs could be responsible. If only entrainment were responsible, then the expected recharge geometry of the isopycnals should be somewhat symmetrical about the vent in all directions rather than its asymmetric linear shape. Thus, isopycnal geometry serves as a test that would indicate that flow is not everywhere homogeneous.

I have constructed a model that is reasonably well supported by the physical data presented here including water flow rates and heat output at the 12°40'N field as well as isopycnal maps which

provide clues about the nature of water flow. A few premises basic to the model are assumed. The first premise is that "water in = water out." The hot water that exits at 12°40'N must enter the crust somewhere. It is proposed that the bulk of the recharge water is provided by inflow at both the 12°35'N OSC and the 12°54'N OSC (Figures epr-17 and epr-18).

Rosenberg et al. (1993) discuss the recharge and discharge in rise axis hydrothermal systems. Lateral changes in permeability act to focus flow where such changes are abrupt. The OSC at 12°35'N represents such a change in macropermeability. Indeed, Rosenberg et al. (1993) also postulate that discharge zones are narrower than zones of recharge. The distinct, narrow chimneys within hydrothermal fields serve to supply a smaller egress for water than is afforded water entering at an OSC. Also discussed by Rosenberg et al. (1993) is that hydrothermal convection systems, while not necessarily reaching steady-state flow patterns, do acquire quasi-steady configurations for significant periods of time. The time they calculated to achieve steady-state flow varied from 25,000-150,000 years.

In this model, water enters at OSCs at the same rate it is discharged because the hydrothermal discharge system acts as a pump to pull in cold water. For ease in calculation, the assumption is made that one-half of the water entering at the 12°35'N OSC feeds the hydrothermal field at 12°40'N and the other half feeds hydrothermal field(s) to the south. Likewise, one-half of the flow of cold water into the 12°54'N OSC feeds the 12°40'N field while the

other half feeds field(s) to the north. It is also assumed that the cold water flow into the two OSCs is approximately equal.

Thus if $1/2$ the $12^{\circ}35'N$ OSC flow = $1/2$ the $12^{\circ}54'N$ OSC flow, then either of these flow rates = $1/2$ the hydrothermal field flow at $12^{\circ}40'N$.

Therefore:

$1/2$ ($12^{\circ}35'N$ OSC flow) + $1/2$ ($12^{\circ}35'N$ OSC flow) =
hydrothermal field flow at $12^{\circ}40'N$.

or

the entire cold water flow into the $12^{\circ}35'N$ OSC =
hydrothermal field outflow at $12^{\circ}40'N$.

The second assumption is that inflowing water comes from off the rise axis. The assumption is reasonable because of the general westward flow of water in the region and the fact that the water is cooler and more dense than ambient water, suggesting that it is derived from greater depth. The cold water entering the EPR at the $12^{\circ}35'N$ OSC is colder and, according to the data previously presented, denser than ambient water at an equal depth (Figures epr-2 and epr-9). If calculations are correct and salinities measured at and near the OSCs are representative for the cold water lens above the OSCs, then it can be estimated that the water must rise by one hundred to perhaps a few hundred meters in the water column to reach this position on the EPR rise crest. The source of the cold water

anomaly is thought to originate from the east, as is suggested by the direction of flow of the near bottom currents from east to west across the ridge (Crane et al., 1988) and the previous discussion of isopycnal geometry.

However, both lateral and vertical transport are necessary to raise the cold, dense water up from a greater depth on the flanks of the rise axis to the shallow depth at the OSCs in question. In point of fact, Fox (1990b) suggested that changes in bottom pressure on the Juan de Fuca Ridge are related to the release of a thermal plume, uplifting dense water in the process. In the case of the Juan de Fuca Ridge CTDs monitored a rapid flow of cold dense water over the rise axis immediately following a deflation event. Where sustained upwelling of dense water is suggested at the EPR (although this survey lasted only one month) recharge must be large enough to actually pull water from depth onto the ridge and entrain it into the narrowly focused recharge zone of the hydrothermal system. However, laboratory models rarely, if ever, show that recharge fluids are drawn from depths in excess of the level of discharge (e.g. Rosenberg and Spera, 1990) and thus there are substantial doubts about the nature and the endurance of the cold water lens discovered over the OSCs in question, although the hydrothermal field at 12°40'N does appear to be persistent over the duration of the seafloor investigations in this region. Nevertheless, because upwelling events have been documented in association with MOR volcanic processes, I will assume that these processes have been caught in action on the ERPR and proceed to estimate flux velocities of water in the crust during a recharge process.

If it is assumed that westward flow, discrete zones of recharge and upward entrainment of water into the hydrothermal waters all interact at once and that the region of fractured seafloor above which cold dense water resides is the region of recharge, then it is possible to estimate rates of recharge assuming that the cross-sectional area through which the water flows is a function of both the depth in the crust to which the recharge water reaches and the width of the zone on the seafloor into which seawater enters the system. The assumption here is that water is pulled along the sea floor, is sucked into the faults and fissures of the OSCs, and sinks to a depth determined by large scale permeability and proximity to a heat source.

V. A MODEL OF HYDROTHERMAL RECHARGE ON THE EAST PACIFIC RISE

A. Part A

The equation used to model the flow of near bottom water towards the recharge system is:

$$V = Q/A$$

where

V = velocity of bottom water flow

Q = volumetric flow rate = $3.22 \times 10^6 \text{ cm}^3/\text{s}$

A = cross sectional area of bottom water drawn into the recharge zone

In order to carry out the calculations one must know the cross sectional area of bottom flow. The velocity of bottom flow can then be derived. The maximum length along the rise axis into which the water flows computed here is 5.5 kilometers, which is the north-south length of the most pronounced 36.94 isopycnal density anomaly and the length of the highly fissured region of the seafloor (Figures epr-14 and epr-15). Figure epr-19 illustrates the general geometry of the area of near bottom flow. Of course, it is likely that the cross-sectional geometry of cold water flow is oval and not strictly rectangular in shape (Figure epr-19). Calculations for both geometries are included here and shown in Table epr-1. To obtain a range in flow rates, the thickness of the lens of water drawn towards the axis which supplies the recharge zone is taken to vary in depth from 0.01 meters to 10.00 meters (the lens is located directly above the surface of the sea floor). The velocities involved in flows 0.01 to 1.0 meters thick are high enough to be significant, as they should

represent rates of 0.06-82.0 cm/s. Velocities for flows broader than this would be rather slow and are not only probably non-measurable, but also represent flows too broad for the suction generated to support. The values for water flow along the bottom of the seafloor, before entrainment into the crustal system at an OSC vary from 82.0 cm/s for a lens about 0.01m thick travelling along a 0.50 km portion of the seafloor to 0.01 cm/s for water of 10.00 m thickness travelling over 5.50 km of seafloor. Friction may serve to significantly slow down water flow for the cases where the bottom flow is shallow (lenses 0.01-0.1 meters). Interactions with the seafloor for lenses of greater flow depth less significantly effect flow rates.

B. Part B

An alternative model based on flow into specific faults and fissures of measured lengths and widths (Figure epr-14) was designed to compare the degree of similarity in the derived flow rates of near bottom water to those derived in the previous model. The residence time of water in the rise axis crust is also estimated. Once again, the overall hydrothermal volumetric flux rate of $3.22 \times 10^6 \text{ cm}^3/\text{s}$ is used based on the predetermined discharge rate at the 12°40'N hydrothermal field. In this model, the width of the inlet faults and fissures is taken to vary from 1.0 meters to 5.0 meters. Because the resolution of the SeaMARC I and II systems is not fine enough to resolve widths of this magnitude, the range of widths is chosen because it is representative of fault and fissure widths observed from submersibles in this area (Hekinian et al.,

1985). However, the side looking sonar data do quite well in the resolution of fault length. In this region along the EPR the total length of faults and fissures vary from one to eight kilometers (Crane, 1987 and Perram and Macdonald, 1990) (Figure epr-14). Together, the width and length of these features represent the cross sectional area (A) of recharge into the rise axis hydrothermal system.

Using the equation :

$$V = Q/A;$$

where the parameters V and Q have been previously defined

some representative velocities of water flow into crustal conduits can be calculated. The velocities obtained by this method are included in Table epr-2. Velocities vary from 0.322 cm/s for one kilometer of fissure and fault length at widths of one meter to 0.008 cm/s for eight kilometers of fissure and fault length at widths of five meters.

Assuming that the water enters at the midpoint of the 12°35'N OSC and sinks to a depth of 1.5 kilometers (the depth to the top of the magma chamber as described by Detrick et al., 1987), it then moves along-axis 9.3 kilometers to the north where it travels upward 1.5 km and exits the surface of the sea floor as hot, hydrothermal fluids. Thus, the total distance travelled from recharge to discharge is therefore about 12.3 kilometers. Ludwig et al. (1993), studying porosity in the shallow EPR crust, have determined porosity to be about 30% from the surface down to 50 meter depth, with a severe drop-off to 5% porosity at depths greater than 130 meters. Thus, most likely the widths of faults and fissures do not stay

constant. The area of water inlet probably represents the 30% porosity at the surface. The drop-off to 5% would seem to indicate that only about 17% of porosity at the surface would be available for water flow at depths in excess of 130 meters. Indeed, the trip that water takes in reaching the magma supply is almost entirely beneath depths of 130 meters. Conduits may not, however, narrow to such a degree. The residence time, using the flow rates calculated and assuming that faults and fissures conserve their widths at depth, may vary from 1.5 months to five years. The decrease in porosity discussed by Ludwig et al. (1993) would result in longer residence times, as water would find more resistance to flow during its 12.3 km path through the crust. This would bring residence times to vary between 9 months and 30 years, as indicated in Table epr-2.

C. Cooling Time for the Magma Body

What are the implications for magma cooling time? Burnett et al. (1989), using tomographic imaging and the analysis of P-wave travel times, found that the magma body beneath the hydrothermal plume at 12°40'N is fifteen kilometers wide and three kilometers thick. The length of this axial magma chamber is interpreted to be eighteen kilometers - the distance between the two OSCs - and thus the total chamber volume is about 810 km³.

The solidification of magma which accounts for the heat we see at and near the rise axis can be broken into the latent heat of solidification, conduction of heat through the crust and hydrothermal circulation through sea water. It is possible to estimate the cooling time for a non-steady state magma cupula of the dimensions

mentioned, using 4×10^5 J/kg for the latent heat of solidification and 3.84×10^9 J/s (3,845 MW) for the heat flux from the rising hydrothermal plume. The density of basalt is, on average, $2,700 \text{ kg/m}^3$. Thus the cooling time for a non-steady state magma body is approximately:

$$4 \times 10^5 \text{ J/kg} \times 2700 \text{ kg/m}^3 \times 1 \text{ s} / 3.84 \times 10^9 \text{ J} \\ \times 810 \text{ km}^3 = \underline{7,200 \text{ years}}$$

If the magma chamber basalt is denser, and 2900 kg/m^3 is taken as basalt's density, then the magma body's cooling time is almost 7800 years. Thus, cooling time for the fifteen kilometer wide body is 7500 +/- 300 years.

If Burnett's 1989 estimate for such a wide magma chamber is too large and if the chamber is only three or four kilometers wide as discussed by Detrick et al. (1987), then the resulting time to cool such a (non-steady state) magma chamber would be only about 1800 +/- 300 years, depending upon specific magma chamber width and basaltic density.

Are the magma chambers here steady state, or are there merely periodic injections of magma that cool as discrete units? Perhaps this can be answered by looking at sea floor evidence. Crane et al. (1988) noted that some OSCs seem to have migrated along the rise axis, thus forming relict off-axial structures which strike at angles 285° to 330° to the axis. This evidence may lead one to believe that zones of magma injection do, indeed, migrate along the rise axis. In addition, off axial seamount chains which strike

obliquely to the rise axis are indications of magma pulsing. If areas of low magma supply (at the OSCs) migrate, then the intervening regions of robust supply must also migrate. Areas that were once generously supplied may subsequently become magma starved. Perhaps after only a few incidents of magma intrusion, the intrusion site moves. If this is the case, then the time scale of magma injection may be on the order of less than 10,000 years.

D. Discussion

The cross-sectional region into which cold water flows to recharge the 12°40'N hydrothermal system has been modelled in two ways. Both determine the velocity of water entering the system, but are different in that they represent two different regions of flow.

The first model calculates the rate of near bottom water flow assuming recharge draws from a finite layer of deep ocean water. This recharge rate varies depending upon the vertical cross-sectional area of flow toward the ridge from off axis. It is apparent that flow occurring more than 1.0 meter above the sea floor entrance to the faults probably moves too slowly to be considered.

The second part of the model calculates the rate of inflow through faults and fissures on the eastern side of the OSC at 12°35'N, where water is assumed to enter the hydrothermal system. While the actual length of faults and fissures is 8.05 kilometers for the area indicated (Figure epr-14), it is possible that the inlet area is much smaller than this. It is reasonable to believe that the inlet area lies closer to the rise axis and that the isopychnals indicate water disturbances at some height above the actual point of inflow. If this

is true, then a fault and fissure length of one to three kilometers may be a good estimate of the cold water inlet. Zone A (Figure epr-14) represents the area closest to the rise axis and contains 1.19 kilometers of faults and fissures. A range of water flow velocities from 0.3cm/s through faults one meter in width to 0.06cm/s through faults five meters in width have been estimated for this length of faults and fissures. Similar differences can be seen for Zones B and C.

Although two sets of inflow velocities are determined here, and they both model cold water flow into the same hydrothermal system, the volume of water pulled along the sea floor in Part A (Model 1) is the same as the amount pulled into faults and fissures in Part B (Model 2). The areas across which flow occurs is different in the two models. The velocities derived from the two models are thus also very different due to this, and are in no way required to be equal or even similar.

The modelling undertaken here was meant to put some numerical constraints on flow into cold areas of the EPR that feed hydrothermal systems with recharge systems. These calculations are useful because they present numbers which investigators might test by planning much more detailed and accurate surveys in the region. Further modelling, or even in situ measurement of flow rates, will help to further constrain the velocity (or velocities) of cold water recharge.

The time constraints placed upon magma supply at this portion of the EPR have been based on both the velocity values derived for water flow and on geophysical evidence about the size and shape of magma bodies that underlie the rise axis. Certainly the time it takes

to cool a chamber of the size underneath this section of the EPR is short (a few thousand years). Even short duration hydrothermal events have been observed in a period of only a few years, during which time the center of hydrothermal activity migrated from 12°50'N to 12°40'N (Hekinian et al., 1985). Crane (1985) suggested that axial highs migrate to the south at a rate of 1-1.5 cm/yr in the region from 8°-14°N on the EPR. The actual migration here represents a great distance, given the fact that the entire system is about thirty-seven kilometers along the axis. Thus, a migration of thermal activity amounting to one-half the length of this system may have occurred in only a few short years. It then follows that the underlying magma supply may very well have also migrated in this time. In light of this fact, a non steady state model appears reasonable, if one takes the injection of magma at varying points on the axis to occur in discrete events. A time scale of a few thousand years is reasonable to cool such a transitory feature.

VI. FEATURES OF THE JUAN DE FUCA RIDGE

A. Introduction

The detailed investigation of the structure and evolution of the Juan de Fuca Ridge, a medium spreading rate ridge with a half spreading rate of 2.9 cm/yr, off the northwest coast of the U. S. was initiated in the late 1950's with the discovery of linear magnetic anomalies (Mason (1958), Mason and Raff (1961) and Raff and Mason (1961)). The discovery of these magnetic reversals and their relation to magnetic reversals worldwide across mid-ocean spreading centers was indeed remarkable. The Vine and Matthews (1963) explanation of reversals as an indication of seafloor spreading gained acceptance. Further, Cox et al. (1964) correlated marine magnetic attributes to exact ages based on paleomagnetic age correlations obtained from terrestrial samples.

The symmetric, elongate magnetic reversals observed in the wide ocean basin of the Atlantic are not observed adjacent to the Juan de Fuca Ridge. To explain the curious arrangement of the V- or U-shaped magnetic anomaly patterns Hey (1977) proposed a series of ridge jumps and ridge propagations on the Juan de Fuca Ridge. Hey also proposed that the realignment of lithospheric plates adjacent to the mid-ocean ridge causes the resulting realignment of rise axis segments.

Because of its close proximity to both Canada and the U.S. the Juan de Fuca Ridge has been the focus of a large number of high resolution sea floor surveys. Investigations into axial processes on a small scale, involving magma supply, its accompanying hydrothermal activity and faulting mechanisms have since been undertaken

(Kappel and Ryan, 1986, Barone and Ryan, 1988, Baker et al., 1990 and Lupton, 1990).

The huge Axial Volcano dominates the Juan de Fuca Ridge in the region of 46°00'N (Figure i-4). It is up to 3 km in width and rises to a depth of 1540 meters, whereas the general bathymetry of the surrounding rise axis is 2200-2400 meters (Baker, et al., 1990). This Axial Volcano is part of a seamount chain that intersects the rise axis and is the youngest edifice on the Cobb-Eickelberg Seamount chain (Delaney et al., 1981) which extends toward the northeast from the Juan de Fuca Ridge. Axial Volcano is thought to represent the interaction between a fixed mantle hotspot and the Juan de Fuca Ridge (Desonie and Duncan, 1990). The basalts of the seamounts in the Cobb-Eickelberg chain variably show characteristics of hotspot and mantle plume material and of rise axis basaltic material due to intermittent interaction between the two types of magmatic regimes during the formation of these seamounts.

The Juan de Fuca Ridge not directly affected by Axial Volcano is a medium-rate (6 cm/yr full rate) spreading center. It has undergone clockwise shifts in relative direction of motion at 17 Ma, 8.5 Ma and 5.0 Ma (Kappel, 1985). These shifts have been resolved in the traces of magnetic anomalies surrounding the ridge (Wilson et al., 1984) and appear responsible for reorientation of rift segments and reinitiation of rift propagation. The 5.0 Ma shift resulted in cessation of activity at the northern and southern propagators, giving rise to the Sovanco and Blanco transforms (Wilson et al., 1984) (Figure i-3).

B. Instrumentation, Data Collection and Interpretation

The data set utilized for the Juan de Fuca Ridge consists of SeaMarc I side-looking sonar images collected in 1982 (Crane et al., 1985) (Figures jdfr-2 through jdfr-30). These were compiled into a mosaic of nearly continuous ridge coverage at Lamont-Doherty Earth Observatory. In addition to the structural data, potential temperature and depth data presented by Crane et al. (1985) are combined to provide a clearer picture of heat supply and hydrothermal activity along the medium spreading rate ridge.

SeaMARC I was a side-looking sonar system operated via research ships to survey and map the seafloor. It measured backscattered echo intensity on each of a number of narrow beams. Near-vertical incidence of insonified beams resulted in no useful backscatter being obtained directly beneath the ship. Thus, blank strips showing no image appear along the center or sides of side-scan swaths. SeaMARC I was a 30 kHz side-looking system which surveyed a swath of seafloor up to 5 km in width. The vehicle was towed 300 m above the bottom at about 3 km/hr. Seafloor objects less than a meter in amplitude and spaced less than 20m apart are resolvable (Vogt and Tucholke, 1986).

The potential temperature data were collected by Crane et al. (1985) using a 50 m long Aanderaa Thermistor chain, weighted and towed 200-250 m below the SeaMARC. The data necessary to calculate potential temperature (the temperature of a water sample if it were raised adiabatically to the surface) includes salinity, conductivity and temperature. The thermistors were calibrated to a resolution of 0.01°C and an accuracy of 0.15°C. There are 300

measurements for each thermistor for every 10 km on the map, thus providing a very dense network of data (Crane et al., 1985).

In this thesis, systematic evaluation and interpretation of the side-scan sonar imagery is undertaken. Images are interpreted for: location of faults or volcanics, widths of these zones, the number and size of distinct volcanic edifices, and fault density. Fault density is measured as the number of faults (trending parallel or sub-parallel to the rise axis) per kilometer (measured perpendicular to the rise axis). For each minute of latitude along the north-south trending ridge, volcanic/tectonic data were tabulated from these side-looking sonar images.

In addition to the side-looking sonar data, physical oceanographic information (average potential temperature) is plotted for each minute of latitude. Finally, a determination is made for each minute of latitude about the relative age of the seafloor structures (young and active or old and inactive). Such determinations are made based upon the degree of desiccation or erosion of seafloor features and the degree of backscatter strength. Fresh faults would tend to have crisp, linear edges, whereas aging of extensional seafloor can create a complex network of faults which break up and become less linear, perhaps due to changes in surficial stress. Areas of older crust which have not been subjected to newer, younger, linear faulting would exhibit only the curvilinear, older faults. Perhaps curvilinear faults indicate regions of colder (i.e. thicker) crust where faults may become broken apart, or desiccated. In this case the curvilinear form may represent the shape of the deep boundary of the underlying isotherms and if there are new faults in

old crust, their shape may reflect the shape of the thermal state of the crust subadjacent and adjacent to the rise axis. Faulting that is fresh and not desiccated should indicate recent faulting and extension and fresh axial faults are characterized by a high degree of linearity.

Young volcanic expressions are indicated by acoustically dark, reflective lavas or hot uplifted seafloor (Vogt and Tucholke, 1986). Older volcanics which are less reflective and are seen extensively in the Juan de Fuca Ridge images tend to be cut by faults that superseded the volcanic activity. Fresh volcanic edifices would appear crisp at the edges and highly dark. Older volcanic edifices should have less distinct edges. There are places on the axis where sonar images of volcanics appear as "fishscales" which account for the majority of small volcanic edifices. These are probably desiccation features, as they look like the volcanic features observed on the younger, more uplifted rise segments but are not as reflective and do not have sharp reflective edges.

The combination of the many features available for study and evaluation has allowed me to segment the Juan de Fuca Ridge based upon both seafloor structural features and properties including potential temperature in the water column, bathymetry at the rise axis and hydrothermal activity. There are breaks in continuity along the axis among these features, and careful analysis and assessment provides an image of sub-ridge segmentation and a clarification of the boundaries that delineate these segments.

All of the structural and volcanic data are tabulated along and across the rise axis. The wavelengths of segmentation along the axis

are determined on the basis of the observed harmonic repetitions of surficial structures and volcanics and the inferred distribution of the subjacent magma supply.

The features that contributed most to the determination of axial sections are the nature, location and variability of structural zones along and adjacent to the ridge. Structural segments of the rise axis have been delineated based upon breaks in continuity of structural features. For example, a region which exhibits a single structural zone of faults may transition into a region characterized by a double zone of faults and small volcanic edifices. This may (or may not) accompany a change in bathymetry and potential temperature in the water column directly above the axis. Small structural changes, alone, may not be significant. However, thermal changes accompanying morphologic changes can lead to the delineation of a new axial segment. The dimensions of these axial segments, delineated by "natural" breaks in structure, bathymetry and potential temperature, help to define underlying shape and continuity of magmatic and hydrothermal processes, which in turn offer insights into magmatic timescales.

C. Regional Hydrography, Morphology, Tectonics and Magma Supply

i. Hydrography

Figure jdfr-1 illustrates the thermal structure of the near-bottom water above the Juan de Fuca Ridge. Hydrothermally active areas are characterized by dipping isotherms indicating water at depth which is warmer than is expected in a thoroughly and consistently stratified water column. Profiles of water column data

taken close to and directly over the Juan de Fuca Ridge axis show that water below 1500 m is warmer over the axis than away from the axis. This difference in temperature is apparent even at distances of 20 or 30 km away from the axis. Warming of up to 0.1°C seems to occur, however, in the bottom 200-500 m of the water column. Above this height off bottom, water temperatures off axis are nearly identical to those at the same elevation above the axis (Crane et al., 1985). Crane et al. (1985) propose that high heat convection from the volcanically active ridge is the source of the warmed water above the ridge.

ii. *Morphology*

The morphology of the Juan de Fuca Ridge in the region studied is marked by a continuous, linear elongate crestal ridge that extends from the Blanco Fracture Zone at 44°30'N to 45°12'N. The crestal ridge is about 10km wide and 200-300 m high (Kappel and Normark, 1987). Further northward to 45°40'N, in a region called the Shingle Ridge Segment by Kappel (1985), there is a mixture of smooth and hummocky flow surfaces on the floor of the elongate summit depression. There are also collapsed lava surfaces (Kappel, 1985). The spreading axis is not clearly defined north of 45°40'N (Kappel, 1985) and just south of 46°00'N Axial Volcano dominates the morphology of the ridge with its caldera. The morphology north of Axial Volcano is suggestive of successive draping flows down the ridge flank. Often the volcanics appear to have erupted from a fault or fissure hundreds of meters in length (Kappel, 1985).

Crane et al. (1985) located four bathymetrically uplifted, geothermal regions along the Juan de Fuca Ridge. These regions are spaced about 100 km apart and are up to 20 km in length. They are located at 44°31'N-44°41'N, Axial Volcano and its surroundings, (centered on 46°00'N), 46°41'N-46°55'N and 47°33'N-47°36'N, (which is beyond the region of study).

iii. *Tectonics*

Tectonic studies and theories on volcanic episodicity have been undertaken by researchers such as Kappel and Ryan (1986). Kappel, in her doctoral dissertation (1985) and later in the press (Kappel and Ryan, 1986), discussed the morphology along the Juan de Fuca Ridge, including the variability of rift bathymetry and rift valley dimensions. They proposed that the morphology along and adjacent to the ridge is the result of tectonic super-cycling which includes 1) volcanic construction of a widening crestal ridge prior to the collapse of the summit depression; 2) collapse of the summit of the crestal ridge to form an elongate summit depression during a phase of volcanic inactivity; and 3) renewed magmatism in the elongate summit depression as its floor widens by extension and brittle fracture in the upper crust. Kappel and Ryan did not suggest any time constraints on the individual tectonic episodes or on the cycle as a whole. However, they did make a comparison to volcanic/tectonic events on the EPR at 21°N where the morphology and spreading rate are similar to the Juan de Fuca Ridge.

The mid-ocean ridge which runs through Iceland may also serve as an interesting comparison to the Juan de Fuca Ridge. In

Iceland a divergent plate boundary is superimposed on a large hot spot with a presumed deep root in the mantle (Einarsson, 1991). Einarsson (1991) has found that on Krafla, a central caldera of the Northern Volcanic Zone on Iceland, two processes are at work: there is inflow of magma into the crustal magma chamber and rifting at the plate boundary. Tectonic stress builds up slowly with time due to plate movements. Magma flow into a magma chamber can trigger the release of tectonic stress and thus initiate rifting. Fissures then propagate away from the magma chamber to form dikes. Einarsson (1991) further states that magma inflow causes inflation of the crust which leads to extension and normal faulting in the roof above the magma chamber. Deflation events lead to contraction and reverse faulting.

Rubin and Pollard (1988), in investigating dike-induced faulting in Iceland have concluded that "observation of ... subaerial volcanic rift zones *has demonstrated that ... formation of new fissures occur(s) during episodes of dike emplacement.*" Further, Rubin and Pollard (1988) state that intrusion of magma plays an important role in developing rift-zone geomorphology and topography. The graben floor is seen as representing a zone of pervasive inelastic deformation and there is evidence "the dike intrusion triggers faulting, and not vice versa." (Rubin and Pollard, 1988).

Crane and Ballard (1981), in an investigation of cyclic evolution of the Mid-Atlantic Ridge, have discussed that fissures and fractures may actually occur after magmatic injection is complete. They propose that initial volcanic extrusion covers up previous fissures. Then, after volcanic activity has been reduced, cooling around

magma intrusions causes new fissure formation. Further, new fractures may open which allow for egress of hydrothermal waters (Crane and Ballard, 1981).

Barone and Ryan (1988) have discussed how different faulting styles result from differences in below-axis phenomena at the southern segment of the Endeavour Ridge which lies to the north of the study area of the Juan de Fuca Ridge. They note that tectonism has broken exposed lava surfaces which had become dull due to a thin dusting of sediment. This might indicate that faulting may follow a substantial period of heating and volcanism. Barone and Ryan (1988) continue to discuss that extension resulting from the separation of the Pacific Plate and Juan de Fuca Plate works to break the upper skin of the oceanic crust into slivers. This is perhaps analagous to the zones of faulting and extension observed on the Juan de Fuca Ridge and discussed in this thesis.

There is evidence that plate boundary zone extension is not a steady state feature. Rather, Barone and Ryan (1988) reason that the belt of extension at the plate boundary widens in the absence of significant volcanism. The model of extension at the Juan de Fuca Ridge presented in this thesis also supports a non-steady state model of extension and magma supply, although it agrees with the premises of magma-induced faulting of Einarsson (1991) and Rubin and Pollard (1988).

iv. Magma Supply

The nature of the magma chamber, if any, underlying the Juan de Fuca Ridge has been investigated by Morton et al. (1987). Seismic

reflection data suggest that the top of the Juan de Fuca Ridge magma chamber is about 2.5 kilometers deep. Underneath deeper parts of the axis the distance to the top of the magma chamber from the sea floor increases. Jung and Lewis (1982) reported their findings from seismic refraction studies which were inconclusive about the existence of an axial magma chamber. Lister (1983) proposed that axial magma chambers along the Juan de Fuca Ridge are short-lived and owe their short time span to hydrothermal circulation and cooling.

Lister's theory of the short-lived magma chamber appears to be contradictory to Kappel's (1985) theory of long-term cyclical activity which proposed that specific points on the rise axis are alternatively magma supplied and magma starved. However, Kappel (1985) is not able to hypothesize about the duration of this magma cyclicity. If a slowdown of magmatic activity at the rise axis is indeed indicative of chamber cooling, as Lister suggests, then renewed activity, modelled by Kappel, might necessitate formation of a magma chamber relatively quickly. It may, of course, be possible that magma supplying active parts of the rise axis are fed not by well-developed magma chambers, but rather by large crustal magma conduits that are available intermittently. This is a possible explanation for the lack of sound evidence for an axial magma chamber. It is probable that cooling of crustal magma is achieved over a much shorter period of time than cooling of a large axial magma chamber.

By contrast, Axial Volcano, part of the Cobb-Eickelberg Seamount chain, represents an interaction between a fixed mantle

hotspot and the Juan de Fuca Ridge (Desonie and Duncan, 1990). Thus, the basalts of Axial Volcano have chemical characteristics of both hotspot and spreading ridge melting regimes (Desonie and Duncan, 1990) and Axial Volcano has a more abundant magma supply than the surrounding ridge.

D. Outstanding Questions

The questions that arise in this study relate to magma supply and the time constraints upon it. Dike induced faulting depends upon an abundant magma supply. Analysis of the faulted structure on the rise axis can lead to an estimation of the time over which magmatic injection has occurred. This, in turn, can put constraints on the duration of magmatic activity at varying places on the Juan de Fuca Ridge rise axis.

The following chapter discusses the implications of the magma distribution and magnitude of hydrothermal events on the variability of structure and volcanics along the rise and from this develops hypotheses about periodicity and style of magmatic events along the Juan de Fuca Ridge.

VII. JUAN DE FUCA RIDGE MAGMA MIGRATION AND COOLING - A MODEL

A. Segmentation of the Juan de Fuca Ridge

The delineation of the segmentation of the Juan de Fuca Ridge is dependent upon variations in bathymetry and structures, as well as the variability of hydrothermal activity along the rise axis (Figure i-4).

The rise axis features that are created in response to the subjacent axial processes include faults, fissures and volcanism. Volcanic features can be small-scale cones, sheetflows or large-scale eruptive centers or domes which occupy vast areas. The interplay between and association of these tectonic and volcanic features allows for postulating and describing axial processes and magma supply on the Juan de Fuca Ridge.

Langmuir et al. (1986), Crane et al. (1988) and Macdonald et al. (1988a) have all proposed that rise axes are magmatically segmented and that the wavelengths of the magmatic segmentation reflect the scales and harmonic spacing of magma supply. Crane et al. (1988) used hydrothermally active regions to define a length of magmatic axial segmentation, with hydrothermal activity being associated with an abundant magma supply. Macdonald et al. (1988a) defined segmentation by the distance between rise axis discontinuities which are indicative of points on a rise axis representing distal edges of along-axis magma supply. Crane (1985) and Langmuir et al. (1986) noted undulations in bathymetry on the rise axis to infer differences in magma supply while Langmuir found a short scale petrologic segmentation on the order of 25 km.

Segments delineated on the Juan de Fuca Ridge, for purposes of this thesis, are not based solely on structural or volcanic similarities between segments. Rather, the different segments define portions of the axis where different processes related to magma supply occur. This is why axial segments have different morphologies on the seafloor. For example, some segments are shallow, perhaps indicative of a robust magma supply and this may be observed in the volcanic nature exhibited, evidence of hydrothermal activity or elevated potential temperature in the water column. Deeper segments may be cooler and show no evidence of recent volcanism or hydrothermal activity. Thus, segmentation of the Juan de Fuca Ridge has been based upon a number of factors which are interpreted to be related to magma supply. Depth to the rise axis, degree and location of hydrothermal activity, potential temperature structure of the water column and structural and volcanic features are all considered. There are places on the axis where distinct changes in structural features can be observed. These, combined with other axial features have dictated the segmentation wavelengths that I feel best help to investigate underlying magmatic sources.

B. Structural Segments on the Juan de Fuca Ridge

i. *SEGMENT ONE : 44°28'N to 44°39'N*

This hydrothermally active area (Crane et al., 1985) shoals to depths of 2,220 meters at 44°35'N, the center of the segment, and is twenty-two kilometers long. The potential temperature of the water column directly above the axis at 44°35'N is greater (at 1.77°C) than in the remainder of SEGMENT ONE, which has a minimum potential

temperature at the rise axis of 1.76°C . A likely explanation for this shallow region is that maximum volcanic activity (perhaps in the form of recent magmatic intrusion) occurs here, serving to expand and uplift the crust. The hot, uplifted region is a result of a robust magma supply that also provides the heat necessary to feed the hydrothermal fields. Indeed, the entire length of SEGMENT ONE is bathymetrically uplifted, compared to surrounding terrain on the rise axis (Figures jdfr-1 and jdfr-2).

The region from $44^{\circ}28'\text{N}$ to $44^{\circ}30'\text{N}$ is a single zone characterized by both faults and volcanics. This zone is 1,400 to 1,800 meters in width. The faults, generally ranging from a few hundred meters to 4 km in length, are curved in a sinuous fashion. Fault density ranges from seven to twelve faults per kilometer (as measured perpendicular to the rise axis). Volcanic cone density ranges from fourteen to seventeen small volcanoes per nautical mile (1.85km) of latitude. Potential temperature is 1.77°C in the water column just above the rise axis, at 2270m depth (Figures jdfr-2, jdfr-3 and jdfr-4).

The region from $44^{\circ}31'\text{N}$ to $44^{\circ}39'\text{N}$ is characterized by a double set of inward-facing, en echelon faults. At $44^{\circ}31'\text{N}$ a fault zone develops as a pinched narrow band to the east of the center of spreading whereas on the western side of the axis the fault zone is continuous from the south. The eastern fault zone which emerges at $44^{\circ}31'\text{N}$ widens to 400m at $44^{\circ}32'\text{N}$. Further north this zone varies in width from 600 meters to 1,300 meters, reaching the 1,300 meter width at $44^{\circ}37'\text{N}$, and decreasing to 1,000 meters at $44^{\circ}39'\text{N}$. The western fault zone is more uniform in width. It varies from 800 to

1,400 meters in width, narrowing progressively northward. Fault density (perpendicular to the rise axis) is zero in the eastern fault zone at 44°31'N and 44°32'N. From 44°33'N to the northern extent of SEGMENT ONE's eastern fault zone, fault density is five to thirteen faults per kilometer. Fault density in the western fault zone varies from six to eighteen faults per kilometer. Volcanic cones are few. A neovolcanic zone between the eastern and western fault zones is evident from 44°32'N to 44°36'N.

The faults appear fresh in sonar images due to contrast in acoustic backscatter, creating the crisp edges that are seen on the images. It appears, then, as if desiccation (or breakdown) of fault edges has not had the time to occur. The neovolcanic zone exhibits the dark black, high reflectivity on sonar images, indicating fresh material. Based on the high acoustic backscatter from the rise axis revealed from SeaMARC I imagery, SEGMENT ONE appears to be relatively young (Figure jdf-3). The fresh nature of both the faults and the neovolcanic zone, along with high potential temperature in the water column and the shallow depth of the seafloor support this observation.

The transition to SEGMENT TWO is marked by a change from a double set of axial fissure/fault zones at the northern part of SEGMENT ONE to a single, central fissure/fault zone in SEGMENT TWO.

ii. *SEGMENT TWO : 44°40'N to 44°54'N*

Morphologically, this portion of the Juan de Fuca Ridge is homogeneous in both bathymetry and temperature. It is about twenty-eight kilometers in length and its central fault zone ranges in

width from 2,500 meters to 3,500 meters. Depth at the rise axis varies from 2,280 meters to 2,400 meters. Potential temperature in the water directly above the rise axis (at 2,280 m to 2,400 m depth) varies from 1.76°C to 1.77°C (Figure jdfr-5).

The area is characterized by a single zone of faults aligned along and parallel to the central rise axis. Volcanic activity in this region is minor, at most. The central fault zone ranges in width from 2,500 meters to 3,500 meters. Fault density (perpendicular to the rise axis) varies from four to twelve faults per kilometer. From 44°46'N to 44°51'N fault density is greater on the eastern side of the central fault zone. SEGMENT TWO appears young, as the lavas seen are highly acoustically reflective (Figures jdfr-6 and jdfr-7). The faults appear fresh and are characterized by distinct and sharp edges. The potential temperature is high at depths directly above the shoal sections of the axis; up to 0.10°C higher in SEGMENT TWO than in neighboring SEGMENT THREE (Figure jdfr-1).

The transition from SEGMENT TWO to SEGMENT THREE is marked by an abrupt (250 meter) deepening in the bathymetry. This is accompanied by the onset of a double axial zone; one of axial faults and fissures and one of volcanic material, at 44°58'N.

iii. *SEGMENT THREE : 44°55'N to 45°51'N*

SEGMENT THREE, 104 km long, is varied in its morphotectonic appearance. It can, however, reasonably be considered one segment due to its position between two hot, uplifted regions on the rise axis. Also, it is deep in relation to the adjacent rise axis segments (Figure jdfr-1).

The structural changes within SEGMENT THREE occur over small distances along the axis. The bathymetry is varied, but there are not drastic differences in depth. Local depth anomalies, where the rise axis is deeper than at surrounding segments, are found at 45°05'N, 45°18'N and 45°40'N and are 150 m, 80 m and 130 m, respectively. Water column potential temperatures in these particularly deep regions vary (Figure jdfr-8). At 45°05'N potential temperature in the water column just above the axis is 1.76°C (at a depth of 2620 m), at 45°18'N potential temperature is 1.67°C (at a depth of 2530 m) and at 45°40'N potential temperature is 1.71°C at a depth of 2480m. The near-bottom potential temperature at 45°18'N and 45°40'N is cold relative to the higher potential temperatures above the rise axis directly to the north and south of these latitudes. In these two areas the deepest portions of the rise axis coincide with the coolest potential temperatures. This is in agreement with the idea that hot crust is thermally uplifted and warms the bottom water above it while cool crust is deeper.

Structurally, the rise axis is as varied as the potential temperature in the water column above it. From 44°55'N to 44°57'N there is a zone of morphological transition between the rise axis to the north and the south. The rise to the south is characterized by a centrally faulted zone. In contrast, to the north, both the western and eastern inner rifts are cut by numerous curvilinear faults. Neovolcanic activity is also apparent on the margins of the rift (Figures jdfr-9 and jdfr-10). At the transition between these two structural styles there is a wide single zone of faulting up to 4,000 meters in width where the axis depth does not differ significantly

from the regions directly to the north or south. Potential temperature in the water column just above the rise axis is 1.76°C at depths of 2,200-2,400 m at the transition zone.

The region from $44^{\circ}58'\text{N}$ to $45^{\circ}04'\text{N}$ is characterized by two zones; a western fault zone and an eastern partially neovolcanic zone. The western fault zone is comprised of curvilinear faults with a density of one to eight faults per kilometer (perpendicular to the rise axis). This region of faulting, which is barely visible at its southern terminus, widens to 3,200 meters, at $45^{\circ}02'\text{N}$. The eastern inner rift zone is characterized by neovolcanic activity and is 1,300 to 3,400 meters wide. Fresh lava fields are evident throughout this area, with up to sixteen volcanic edifices per minute of latitude (Figure jdfr-9). Potential temperature just above the rise axis ranges from 1.75°C to 1.77° at depths of 2,400-2,500 m from $44^{\circ}58'\text{N}$ to $45^{\circ}04'\text{N}$. Depth in this region ranges from 2,330 meters at its southern terminus and 2,480 meters at its northern boundary.

The ridge segment from $45^{\circ}05'\text{N}$ to $45^{\circ}14'\text{N}$ (figures jdfr-9 and jdfr-10) is different from the regions already described because unlike its southern counterpart, the spreading center here is characterized by fresh volcanic material and volcanic edifices on both sides of the central rift valley. There is one volcanic zone with a width in excess of 5,000 meters throughout this portion of the segment from $45^{\circ}05'\text{N}$ to $45^{\circ}14'\text{N}$. The width of the SeaMARC image is only 5,000 meters and I am unable to state the true dimensions of the axial volcanic zone because the rise axis in some places was not completely insonified. At $45^{\circ}05'\text{N}$ the western fault zone terminates. Reflective backscatter suggests that unsedimented lava fields lie on

the western side of the center of spreading and many volcanic edifices are also evident (Figures jdfr-9 and jdfr-11). Most are less than 200 meters across at their bases with only two more than 1,000 meters across. Volcano density ranges from eight to forty-one volcanoes per minute of latitude (nautical mile) from 45°05'N to 45°15'N (Figure jdfr-8). However, only minor faulting is seen. In this same region, depth to the rise axis varies from 2,430 meters to 2,590 meters. Potential temperature in the water column above the axis ranges from 1.75°C to 1.76°C at depths of 2,430-2,590 m. This region appears young as evidenced by highly acoustically reflective, fresh volcanic edifices and flows. In addition, the paucity of sediment on these reflective lavas is suggestive of seafloor that is very young and volcanically active.

The axis from 45°15'N to 45°23'N is characterized by a central axial zone, more than 5,000 meters wide, that contains both volcanoes and faults (Figures jdfr-11 and jdfr-12). At 45°15'N ridge-parallel faults lie on the eastern side of the central axis. At 45°16'N the zone of faulting is more central, but is still dominantly eastern. This structural pattern continues to the north. Fault density (as measured in faults per kilometer perpendicular to the rise axis) ranges from four to eleven. The majority of volcanic edifices are less than 200 meters across with none larger than 1,000 meters in diameter at their bases (Figure jdfr-8). Volcanic density approaches about fourteen edifices per minute of latitude, although in places no volcanoes can be found. Depth to the rise axis ranges from 2,450 meters to 2,540 meters. Potential temperature in the water column above the rise axis, as it deepens and shoals, is 1.67°C to 1.72°C. The

lowest potential temperature is correlated with the deepest portion of the rise axis at 45°18'N at a depth of 2,540 meters. In this region the lavas are highly acoustically reflective, faults appear fresh because they are not eroded, and volcanic edifices are clearly visible and not desiccated.

The region from 45°24'N to 45°51'N has a bathymetric range from 1,790 meters to 2,540 meters. The northern end of this region is shoalest. Potential temperature of the water column directly over the rise axis ranges from 1.69°C to 2.1°C (Figure jdfr-8). The axis itself is characterized by old desiccated volcanics, although there is also faulting. The central rift contains both volcanoes and faults that appear to have been the victim of desiccation following extensional faulting (Figures jdfr-13 and jdf-14). This zone exceeds 5,000 meters in width (wider than the SeaMARC I image). A distinct set of faults (crossing the center of spreading) is observed from 45°30'N to 45°33'N. North of 45°33'N to 45°40'N faulting is more common to the east of the central axis, where fault density is nine faults per kilometer, as measured perpendicular to the rise axis. At 45°42'N faulting is more prevalent on the western side. Faulting is almost non-existent from 45°46'N to 45°49'N. From 45°37'N to 45°44'N faults are curvilinear. Along this segment from 45°24'N to 45°51'N fault densities are up to eleven faults per kilometer (perpendicular to the rise axis) although locally there is minor to no faulting. Volcanics are desiccated and include mainly older flows and cones. The edifices appear less distinct here than in the region directly to the south. Based on the criteria already discussed, the curvilinear nature of faulting is indicative of desiccation. However, there are

five to thirty volcanic edifices per minute of latitude. Most are less than 200 meters across at their bases. Only two are larger than 1,000 meters in diameter and the edges of the volcanoes appear eroded. Although the potential temperature of 1.69°C at 45°29'N to 45°32'N supports the idea that this region is magmatically old and cold, the high bottom water temperature of 2.1°C at the northern end of the segment seems anomalous. This seeming contradiction will be resolved in the discussion of magma supply.

The boundary between SEGMENT THREE and SEGMENT FOUR is marked by a transition from desiccated, faulted volcanoes to a region covered by highly reflective, unfaulted lava flows.

iv. *SEGMENT FOUR : 45°52'N to 46°26'N*

The bathymetry of SEGMENT FOUR (65 km in length) is dominated by the enormous Axial Volcano, which rises to an elevation of 1,520 meters at its peak at 45°58'N. Axial Volcano's caldera reaches three kilometers in width (Figures jdfr-1 and jdfr-15). The surrounding region is also bathymetrically influenced by this uplift. Seafloor depths in this area range from 1,520 meters to 2,430 meters; the deeper seafloor lies at the northern boundary of SEGMENT FOUR. The potential temperature signature above the rise axis is also dominated by the volcano. Potential temperature in the water column at the hottest part of the axial volcano is 2.5°C at 45°53'N at a depth of 1,540 meters. Water emanating here is presumably much hotter than this. Interestingly, Baker et al. (in press) found that hydrothermal venting is less than expected from such a "hotspot" mantle plume. It is suggested that perhaps diffuse

percolation serves to dissipate some heat. In fact, other vent fields on the Juan de Fuca Ridge emit more heat and hydrothermal products (Baker and Massoth, 1986, 1987). Potential temperature data in the water column from 46°15'N to 46°25'N is unavailable. Potential temperature elsewhere in AREA FOUR is as low as 1.8°C (at 46°26'N at a depth of 2,420 meters).

The rise axis structure at 45°52'N is characterized by a single volcanic zone uncut by faults in excess of 5,000 meters in width (Figures jdfr-16 and jdfr-17). In this region there are only a few small volcanoes which are generally less than 200 meters in diameter. The region from 45°53'N to 45°59'N contains the Axial Volcano. Superimposed on Axial Volcano are a handful of smaller flank volcanoes, varying in frequency of one to seven per minute of latitude. The structure of the remainder of SEGMENT FOUR is characterized by a single central volcanic axis (with or without faults) in excess of 5,000 meters wide (Figures jdfr-16, jdfr-17, jdfr-18 and jdfr-19). The region from 46°00'N to 46°06'N contains discrete volcanic centers and is cut by a small number of faults up to one or more km in length. The ridge is solely volcanic, characterized by sheet flows and a few cones, and is uncut by faults, from 46°07'N to 46°12'N and from 46°14'N to 46°15'N (Figures jdfr-18 and jdfr-19). The remainder of SEGMENT FOUR is well-faulted with fault densities ranging from four to ten per kilometer (perpendicular to the rise axis).

The entire rise axis of SEGMENT FOUR appears young because the seafloor imaged by the SeaMARC I is acoustically highly reflective (Figures jdfr-16 and jdfr-18). Smaller volcanic edifices,

generally less than 500 meters across, rest atop the highly reflective lava flows of Axial Volcano. To the north of the Axial Volcano, all volcanics imaged are fresh and not eroded. From 46°17'N to the northern terminus of SEGMENT FOUR, straight, fresh faults are observed. In addition, no desiccation appears to have occurred, that is: volcanics do not appear to have been eroded and faults retain their straight edges, not having been eroded over time. The small number of volcanic edifices associated with this region are not eroded.

v. SEGMENT FIVE : 46°27'N to 46°40'N

This small portion of the rise axis (about 30 km in length) coincides with a depth maximum on the Juan de Fuca Ridge (2,660 meters). Overall, along this segment, depth ranges from 2,390 meters to 2,660 meters. The maximum potential temperature is 1.74°C at the southern end of SEGMENT FIVE at a depth of 2,390 meters (at the rise axis) (Figure jdfr-20). This is comparable to other potential temperatures at this depth along the axis (Figure jdfr-1). The potential temperature in the water column above the rise axis falls in some places to as low as 1.63°C. This low potential temperature coincides with the depth maximum at 46°35'N. There is obviously much less thermal uplift occurring here than in SEGMENT FOUR.

Structurally and morphologically this area is characterized by highly faulted and desiccated lava flows. A central axial zone in excess of 5,000 meters across is characterized by both volcanic edifices smaller than 200 meters across and faults which are visible

northward to 46°37'N (Figures jdfr-21 and jdfr-22). At 46°38'N both an eastern volcanic zone and a central fault zone are observed. This eastern zone is more than 2,000 meters across and continues off the edge of the sonar image. Throughout this segment the central zone is faulted and varies in width from 1,400 meters to 2,800 meters. Fault density varies from one to twelve faults per kilometer (perpendicular to the rise axis) along the extent of SEGMENT FIVE. Some faults are not ridge-parallel, but are oriented at a 30° angle to the center of spreading. In addition, the region is characterized by up to sixty-six volcanic edifices per minute of latitude. However, many of these can be described as looking like "fishscales", as the region is dominated by many small (less than 200 meter across) volcanoes grouped together.

Most of SEGMENT FIVE is undoubtedly old in appearance owing to the desiccated volcanics observed, although toward its northern boundary it appears younger, probably due to the heat source to the north. This younger expression is seen in the development of a discernible fault zone, which becomes better-defined in SEGMENT SIX.

The transition from SEGMENT FIVE to SEGMENT SIX is marked by the gradual change from desiccated, faulted terrain to that characterized by fresh lava flows, linear faults and hydrothermal activity.

vi. *SEGMENT SIX : 46°41'N to 47°06'N*

This segment (48 km in length) is hydrothermally active from 46°41'N to 46°53'N (Crane et al., 1985) (Figure jdfr-1). Potential

temperature in the water column directly above the rise axis, which varies in depth from 2,370 meters to 2,730 meters, is divided into two regimes. The hydrothermal area, northward to 46°55'N, has potential temperatures of 1.69°C to 1.73°C at depths of 2,370 meters to 2,560 meters (just above the rise axis). To the north the potential temperature is 1.63°C to 1.68°C at depths of 2,500 meters to 2,730 meters (Crane et al., 1985) (Figure jdfr-23).

The structure of SEGMENT SIX is remarkably constant throughout, despite the differences in water column temperature and depth. A central fault zone which runs throughout the area is nearly devoid of volcanic cones and craters with the exception of the region north of 46°57'N. From 46°41'N to 46°45'N a central fault zone lies adjacent to an eastern volcanic zone which is 1,000 meters to 2,000 meters across (Figures jdfr-24, jdfr-25, jdfr-26 and jdfr-27). The ridge-parallel fault zone becomes well-defined at 46°43'N where it varies from 1,800 to 2,200 meters in width. Fault densities are high, at nine to thirteen faults per kilometer (perpendicular to the rise axis). There are up to twenty-five volcanic edifices per minute of latitude, mostly less than 200 meters in diameter.

Northward, to 46°56'N, three north-south oriented structural zones are apparent where a western volcanic zone joins the two (central and eastern) zones seen to the south (Figure jdfr-25). The western zone is 1,400 to 2,000 meters across. This is in contrast to the eastern (volcanic) zone which is 600 to 1,800 meters across. The central fault zone is 700 to 1,800 meters across. Fault densities are high, at seven to fifteen faults per kilometer. The number of volcanic

edifices in both volcanic zones ranges from sixteen to forty-one. Again, most are smaller than 200 meters in diameter at their bases.

An analysis of water column potential temperature shows that the coolest part of SEGMENT SIX is from 46°57'N to the northern terminus of the segment. The morphology here is similar to the segment directly to the south because there are western, eastern and central zones (Figures jdfr-26 and jdfr-27). However, the central zone (1,500 to 2,300 meters wide) contains volcanic features such as flows and cones in addition to the faults which characterize the central zone to the south. The volcanic edifices are few and small in size. The western (volcanic) zone is 800 to 1,400 meters across, while the eastern volcanic zone ranges from 1,400 to 2,200 meters in width. The central zone is 1,500 to 2,300 meters wide. The greatest volcanic cone density occurs at 47°00'N (42 per minute of latitude), with most cones less than 200 meters in diameter. Fault density within the central rift ranges from seven to ten per kilometer (perpendicular to the rise axis).

Despite the low potential temperatures (as low as 1.63°C) and depths as deep as 2,730 meters in the northern portion of the region, upon viewing the sonar images, one would believe it to be hot and active because the faults are mainly linear without bifurcations (desiccation) and volcanic flows are highly reflective. Due to its acoustic backscatter appearance, all of SEGMENT SIX has been classified as young.

The transition from SEGMENT SIX to SEGMENT SEVEN is marked by a change from linear faults with highly reflective volcanic flows, to dessicated and bifurcated faults and generally nonreflective

volcanics. Otherwise, the northern end of SEGMENT SIX is surprisingly similar to the northern end of SEGMENT SEVEN with respect to the existence of eastern, central and western structural zones.

vii. *SEGMENT SEVEN : 47°07'N to 47°17'N*

SEGMENT SEVEN, the northernmost portion of the JdFR in this study is about twenty kilometers in length and is marked by anomalies in both depth and potential temperature. Depths range from 2,690 to 2,790 meters. Potential temperature in the water column just above the rise axis is 1.61 to 1.63°C, generally cooler than SEGMENT SIX, at the same depths along the axis (Figure jdfr-28). Three volcanic zones are observed from 47°07'N to 47°13'N. The western and eastern zones are only volcanic without noticeable faulting (Figures jdfr-29 and jdfr-30). The central zone is both faulted and exhibits volcanic features, with volcanic edifices seen at the edge of the inner, central fault zone. Flanking volcanoes are abundant and occur in conjunction with somewhat reflective lavas (which may not be especially indicative of very fresh flows, as the near-bottom water in this region is rather cool). From 47°14'N to the northern end of SEGMENT SEVEN there is a single central zone which is both volcanic and faulted. This zone is not well-defined, being in excess of 5,000 meters (the width of the SeaMARC I image) across. Fault density varies from five to eight faults per kilometer (perpendicular to the rise axis) throughout this area, while there are as many as thirty-five volcanic edifices per minute of latitude, with the majority being less than 200 meters across. All of SEGMENT

SEVEN appears old because the central fault zone has been desiccated. In addition, the volcanics do not appear to be fresh, as they are cut by faults north of 47°14'N.

C. Magma Chambers and Axial Magma Supply

As far as can be determined, no well-defined axial magma chamber (AMC) exists beneath the Juan de Fuca Ridge although some suggestion about a possible magma chamber beneath the southern Juan de Fuca Ridge has been made by Morton et al. (1987). The seismic reflector is rather weak and thus has been interpreted to represent the top of a chamber filled with crystalline mush and not fluid magma. This weak reflector is 2.3 to 2.5 km below the seafloor. Thus, in all likelihood a robust, shallow liquid magma chamber does not reside at steady-state underneath the Juan de Fuca Ridge. In contrast, magma probably rises from a deep source through conduits which are intermittently active and undetectable by near-axis seismic studies. The magma rises in conduits and feeds the axis of spreading by injecting magma. Magma injection can occur directly beneath the center of spreading or to one (or both) sides of it. It appears from rise axis morphology, temperature and structure that areas of magma injection are marked by shallow bathymetry, increased potential temperature in the water column directly above the rise axis (indicating an active heat source) and active fault zones (located centrally or off-center). In contrast, the cold, deep areas show desiccation of faults and volcanic structures. Because different segments lie at different depths, the magma supply in the respective segments must be intermittent. Whether or not magma migrates

from segment to segment or erupts through independent channels into stable segments is not well known. However, segments which were once generously supplied by magma and cut by well defined axial faulting may now be cooling, subsiding and extending. The magma conduits may have become blocked by shifts in the crust, or the deep, undetectable magma supply may itself have migrated. This could suggest differences in melting at depth.

Magmatic intrusions into the crust can be evaluated by analyzing faulting and its relation to the associated magma supply to suggest magmatic activity at some depth. While extension is important to faulting processes, it must be stressed that faulting and extension (due to magmatic influences) can be seen to be related. Kastens et al. (1986) suggest that axial fault zone development is controlled by underlying magma. Pollard et al. (1983) found that cracks and normal faults may form over a dike, thus hydrofracturing is probably also a factor here. This results from heated seawater of 400°C-500°C fracturing the crust, leading to black smokers and localized hydrofracturing. Obviously, if a magmatic source is available, then the heat necessary for hydrofracturing to be initiated may also be present.

D. Magma Supply Time Scales

If faulting on the rise axis is a result of hydrofracturing in response to subjacent magma injection (Pollard et al., 1983 and Kastens et al., 1986) then one can estimate the duration of magma injection by utilizing the following information: 1) the width of the axial fissure zone and 2) the average spreading rate.

The information is available for SEGMENTS ONE, TWO and SIX which are volcanically and tectonically young, generally warmer than surrounding regions, generally uplifted and exhibit distinct and relatively dense fault zones. SEGMENT ONE has axial fault densities of seven to twelve per kilometer. Fault density for SEGMENT TWO is four to twelve faults per kilometer, and within SEGMENT SIX fault density ranges from seven to fifteen faults per kilometer. Based on the general concurrence of fresh volcanism (where volcanism is present) with fresh faulting, I assume that the width of the axial fault/fissure zone is proportional to the duration of extension (if one can assume that the overall spreading rate is a good approximation of spreading everywhere along the axis).

The central axis of SEGMENT ONE is structurally divided into a double fault zone. The eastern fault zone ranges from essentially zero width at its southern terminus at 44°31'N to a maximum width of 1,300 meters at 44°37'N and a width of 1,000 meters at 44°39'N, while the western fault zone at this latitude is continuous from the south to the north and ranges in width from 800 to 1,400 meters. This suggests that the magma source feeding the western fault zone extends from SEGMENT ONE to the south, while the eastern fault zone was created by a different set of magma conduits.

Age constraints on the magma supply and episodic extension can be estimated using the following assumptions: 1) each approximately 1,000 meter wide fault zone is fed by distinct and different magma injections, and 2) the half spreading rate (29 mm/yr) for the Juan de Fuca Ridge is representative for this region.

Because each branch of the double fault zone lies on the edge of the inner rift valley, the half spreading rate is a suitable figure to use.

If averaged over time, 1,000 meters of extension (as measured across a fault zone) is equivalent to approximately 34,500 years at a spreading rate of 29 mm/yr. This assumes steady state spreading of each fault zone over the 34,500 year period.

SEGMENT TWO is characterized by a single fault zone formed by the merger of the double fault system in the south creating a 3,000 m wide central axis fault/fissure zone. Perhaps a convergence of magma conduits has occurred or if magma emanates from SEGMENT TWO and is laterally transported towards SEGMENT ONE then the magma source or conduit may bifurcate with distance from its source. To each side of the axis along SEGMENT TWO there is about 1,500 meters (or a total of 3,000 m) of faulted crust representing the surface expression. Once again, using the 29 mm/yr half spreading rate, the duration of the most recent faulting episode in SEGMENT TWO is calculated to be approximately 51,700 years.

At the northern end of SEGMENT TWO the fault zone widens to 3,500 meters (1,750m on each side of axial spreading), corresponding to a magma cooling period of 60,300 years. The southern fault zone narrows to 2,700 meters (1,350m on each side of axial spreading), corresponding to duration of faulting and hence duration of magma injection of 46,600 years. Fresh volcanics are not observed in SEGMENT TWO but the heat source here may be sufficient to fracture and fault the crust, yet be insufficient to produce lava at the surface. It should also be remembered that fracturing can also be caused by

cooling or by caldera collapse, both of which are associated with magmatic processes.

SEGMENT THREE is generally deeper than SEGMENT TWO. This suggests that magma supply here is less robust than for SEGMENT TWO. However, the southernmost portion of this area, just north of a heavily-faulted transition zone at the north of SEGMENT TWO, is covered by reflective lavas. Continuing northward, there is at least some evidence for recent activity in the form of either volcanic edifices or fresh lavas. The exception to these seemingly active regions of SEGMENT THREE is the region from 45°24'N to 45°51'N which accounts for about fifty percent of the segment.

The supply of magma into SEGMENT THREE probably originated by lateral migration from SEGMENT TWO. It is possible that the tip of northward migration coincides with the field of volcanoes at the southern terminus of SEGMENT THREE. In contrast, the colder, northern portion of SEGMENT THREE has not been the locus of recent volcanism. The curvilinear faults of SEGMENT THREE also may indicate present inactivity. This, in conjunction with deeper bathymetry suggests that subsidence has occurred. The amount of subsidence can be estimated by comparing relative segment elevations with relative magma-induced fault zone widths. The difference in average bathymetry between SEGMENT TWO and SEGMENT THREE is 25 meters. If the duration of SEGMENT TWO's magmatically-induced fault activity is 51,700 years, creating 25 m of relative uplift in the process, then the vertical motion (either uplift or subsidence) is estimated to be 4.8×10^{-4} m/yr.

It is unfortunately not possible to estimate the duration of magma injection on the hot and uplifted SEGMENT FOUR because there is no long, continuous fault zone here. Most likely, continuous eruptions cover up any significant faulting by the process of copious extrusion. Magma migration both to the north and south of Axial Volcano probably contributes to sustained hydrothermal activity at these sites.

SEGMENT FIVE is cold and presently amagmatic. Old, desiccated lava flows are evident. Indeed, the "fishscales" that account for the majority of the small volcanic edifices found here are probably desiccation features because they look like the volcanic features observed on the younger, more uplifted rise segments but are not as reflective and do not have sharp reflective edges (Figure jdfr-21). It is suggested that this area is too far from a reliable magma source to be consistently fed. The northern portion of this region is, however, influenced by the warmer SEGMENT SIX. Non ridge-parallel and desiccated faults of SEGMENT FIVE indicate present inactivity. This, in conjunction with deeper bathymetry suggest that subsidence has occurred.

SEGMENT SIX exhibits two distinct thermal regimes. In general, the southern part of the segment is hot and uplifted while the northern terminus is cold and deep. However, a central fault zone runs along its entire axis. Why, then is there such a difference in thermal character? Quite possibly the northern portion has recently cooled. Using the same premise as developed earlier for SEGMENT TWO, the age of the magma supply can be calculated for the narrowest fault zone.

Averaged over time, the 700 meter wide fault zone corresponds to approximately 12,100 years of magmatically induced fault related spreading at 46°51'N (using 350 meters of seafloor created at a rate of 29 mm/yr). It is important to realize that at 46°41'N the central fault zone widens to 2200 meters. This corresponds to a timescale of approximately 37,900 years. Once again only half of the central fault zone width (here, 1,100 meters) is considered.

The subsidence rate can be determined for SEGMENT FIVE by comparing it to SEGMENT SIX. SEGMENT FIVE and SEGMENT SIX have an average bathymetric difference of 60 meters and the southern part of SEGMENT SIX has a magmatically-induced fault related spreading history of 37,900 years, suggesting a subsidence rate of 1.6×10^{-3} m/yr at SEGMENT FIVE.

The northern part of SEGMENT SIX is, on average, 150 meters deeper than the southern part, with the duration of magma injection in the northern part of the southern region calculated to be 12,100 years (at 46°51'N). Thus subsidence at the northern part of SEGMENT SIX is 1.2×10^{-2} m/yr relative to the uplifted south. Part of the warm region (with potential temperature of 1.71°C) is marked by a fault zone so narrow that at one point it is a mere 700 meters across (at 46°51'N), which according to the model used here, makes it the youngest region along the rise axis (with the exception of Axial Volcano). It is also important to note that this narrow fault zone underlies the hydrothermally active zone (46°41'-53'N).

Thus, the age of the southern, wider portion of the fault zone (also associated with the hydrothermal region) is also relatively

young. This suggests that the magma source that recently abandoned the cold northern portion of SEGMENT SIX may have migrated to the south. This would explain the young character of the entire segment and the cold potential temperatures in the water column found north of 46°54'N. It would also indicate that magma supply along the Juan de Fuca Ridge may become active in a region and then abandon it in as little as 12,000 years.

SEGMENT SEVEN is highly desiccated and faulted suggesting that it is a remnant of a previously active narrow fault zone. If this area were freshly abandoned by the magma source (which now lies to the south), subsidence must have occurred very rapidly. SEGMENT SEVEN is, on average, 125 meters deeper than the northern part of a SEGMENT SIX, which has undergone recent magmatically induced faulting during the last 12,100 years. Thus the subsidence rate of SEGMENT SEVEN is about 1.0×10^{-2} m/yr.

In summary, the timescales for magmatically induced faulting along various segments of the Juan de Fuca Ridge vary from 12,000 years to 60,000 years. The youngest region will probably continue to experience crustal faulting and produce hydrothermal fluids for up to 50,000 or more years. The 60,000 year old region also shows no potentially imminent decrease in activity. However, regions such as SEGMENT SEVEN appear to have been recently abandoned by a magma supply creating a rapid cooling and subsidence of the rise axis crust. Thermal subsidence and uplift rates may range from 1.0×10^{-2} m/yr to 4.8×10^{-4} m/yr.

The scale of magmatic segmentation on the Juan de Fuca Ridge can be viewed in two different ways: 1) tectonic segmentation based

predominantly upon the structure of the magmatically active, time-constrained segments, or 2) purely magmatic segmentation. SEGMENTS ONE and TWO may be construed as a single center of magmatic activity separated by structural breaks. SEGMENT SIX is also such a center of magmatic activity. The distance along-axis from the center of SEGMENTS ONE and TWO (at 44°54'N) to the center of SEGMENT SIX (at 46°54'N) is about 235 kilometers. This distance does not take into account Axial Volcano which dominates SEGMENT FOUR. The center of SEGMENT ONE to the center of Axial Volcano (at 45°58'N) is a distance of 120 kilometers. The distance from Axial Volcano to the center of SEGMENT SIX is 105 kilometers.

Crane et al. (1988) determined the spacing between hydrothermal fields on the East Pacific Rise to be on the order of 150 km. On the Juan de Fuca Ridge hydrothermal field spacing was determined to be about 100 km (Crane et al., 1985). Macdonald et al. (1988a) determined East Pacific Rise segmentation length based upon the distance between rise axis discontinuities. It was also determined that first order discontinuities represented by features such as fracture zones are, on average, 300-500 km apart and that second order discontinuities which are represented by non-rigid offsets are 50-300 km apart. Rise axis discontinuities which may represent the distal edges of propagating magma are thought by Macdonald et al. (1988a) to segment the ridge whereas Crane et al. (1988) suggested that the longer wavelength spacing between magma-driven hydrothermal fields was the representative magmatic segmentation on the rise axis.

Segmentation on the Juan de Fuca Ridge may be defined in two ways. The first is a magmatic wavelength of 105-120 km segmentation (including Axial Volcano). The second is a wavelength of 235 km excluding Axial Volcano as a ridge process. It is tempting to use a 235 km wavelength for segmentation on the Juan de Fuca Ridge since Axial Volcano appears to be a mantle hotspot and may only coincidentally be located on the rise axis. However, Crane et al. (1985) determined hydrothermal field spacing on the Juan de Fuca Ridge to be about 100 km because they included Axial Volcano in their segmentation analysis, suggesting that the normal magma supply to the axis within SEGMENT FOUR might be greatly enhanced by the mantle hotspot feeding Axial Volcano.

Thermal subsidence, which varies from 4.8×10^{-4} m/yr to 1.2×10^{-2} m/yr in the region studied, should occur much more rapidly at its onset (the decay is $1/\sqrt{\text{age}}$). However, actual subsidence or deflation rates are often much higher than this. Fox (1990) measured a movement of 10 cm on Axial Volcano over one sixteen day period and three weeks later a second downward displacement of 5-10 cm over thirteen days. Averaged over the course of a year, this is equivalent to a subsidence rate of approximately 2.20 m/yr. The subsidence at Axial Volcano is due to an abrupt change in the magma supply (deflation event) at an active volcano whereas the subsidence rates at SEGMENTS THREE, FIVE and SEVEN should represent longer-term changes in magma supply.

A rise axis that is reorienting and readjusting like the JdFR (Hey, 1977) will indeed experience sudden crustal shifts in axial orientation. Even if rise axis reorientation is not responsible for a

change in magma supply, other processes may be at work; perhaps a minor subsurface shift can cause a decrease in magma supply in a small axial region, changing structural features. Indeed, there has been no evidence found for a robust magma chamber along the Juan de Fuca Ridge, so the magma source is very possibly near-crustal whereby pockets of magma may still lie in the crust, injecting stringers of magma laterally away from the source and causing faulting in the process. The process(es) at work along the ridge segments investigated dictate that intense magmatic activity and its accompanying structural features are episodic through space and time.

VIII. COMPARISON OF TWO RIDGES

The two spreading centers studied here represent different morphologies and magma supplies. Neither has been considered to be steady state as evidenced by migration of OSCs off-axis on the EPR (Crane et al., 1988) and by Kappel and Normark (1987) in their model of episodicity on the Juan de Fuca Ridge. Instead, the magma supplying these ridges is seen to cool and then be replenished. The reasons for considering the magma supply in this way are different for the two regions.

The morphology of the Juan de Fuca Ridge suggests that magma is periodically injected into the upper crust from deeper crustal magma pockets. Periods of injection are followed by periods of relative magmatic and tectonic quiet. The time frame for this periodicity is probably on the order of 60,000 years - the maximum duration of magmatically induced faulting detected. Cool areas of the rise seem to be slated to become warm and active in the future, just as formerly hot areas may now be cool. These cool areas are even postulated to have incredibly high subsidence rates of up to 1.2×10^{-2} m/yr, further supporting the argument for a non steady-state model.

It is less easy to consider the magma supply along the EPR as non steady-state. Of course, any particular spot on the EPR axis is more or less magmatically active in relation to its nearest neighbor, but do the axial structures, and concomittent magmatic activity, migrate along-axis? Do areas that were once characterized by hydrothermal vents later become magma starved and transition into OSCs or devals? If such migration occurs, then the time frames discussed here may be relevant to such migration. The magmatic

processes at depth would, necessarily, be transitory in the same time frame. Crane (1987) suggests that OSCs are formed at crustal weak points and migrate in response to magma body movement. OSC traces found off-axis at angles oblique to the center of spreading support this claim of OSC migration. As OSCs appear at axial lows and are accompanied by intervening hydrothermally active axial highs, it would seem likely that these axial highs would migrate along axis with the OSCs. Thus, hydrothermally active regions at axial highs would also migrate.

The models used to calculate the temporal nature of magma at the two areas studied are different. The validity of comparing the results of such different models may seem unrealistic to some. However, it is interesting to recognize that the values for the longevity of magmatic activity at any particular place on these two ridges differ by generally about a factor of ten (about 2,000 to 7,200-7,800 years for the EPR vs. 12,000 to 60,000 years for the Juan de Fuca Ridge). Considering that the EPR in the area studied spreads at a much greater rate than the JdFR portion evaluated, it seems that the quicker cooling at the EPR may, indeed, be reasonable. The EPR produces more sea floor per unit of time and so must convey more magma-derived material off-axis. In order to do this more magma must cool per unit time. If the magma bodies beneath the two ridges are similar in size, then it would presumably take less time for cooling of a body to occur at the faster-spreading EPR. The magma bodies' size similarity must only be within one order of magnitude for this relationship to be borne out. Thus, the comparison of magma supply at the East Pacific Rise and the Juan de

Fuca Ridge has been shown as valid for the purposes of constraining time relationships, especially in light of the similar magmatic timescales at these two rise axes.

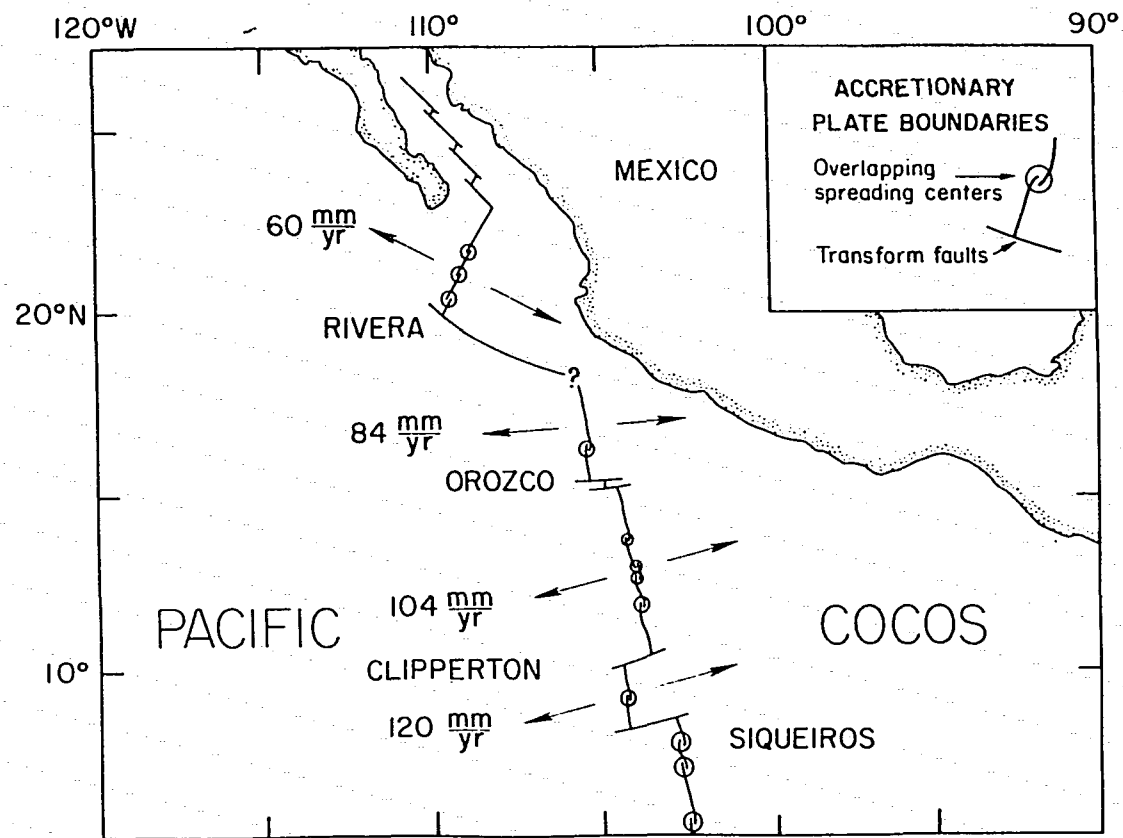


FIGURE i-1
Regional map showing the location of the plate boundary in the eastern Pacific and major rise-axial features. Spreading rates from Antrim et al., (1988).

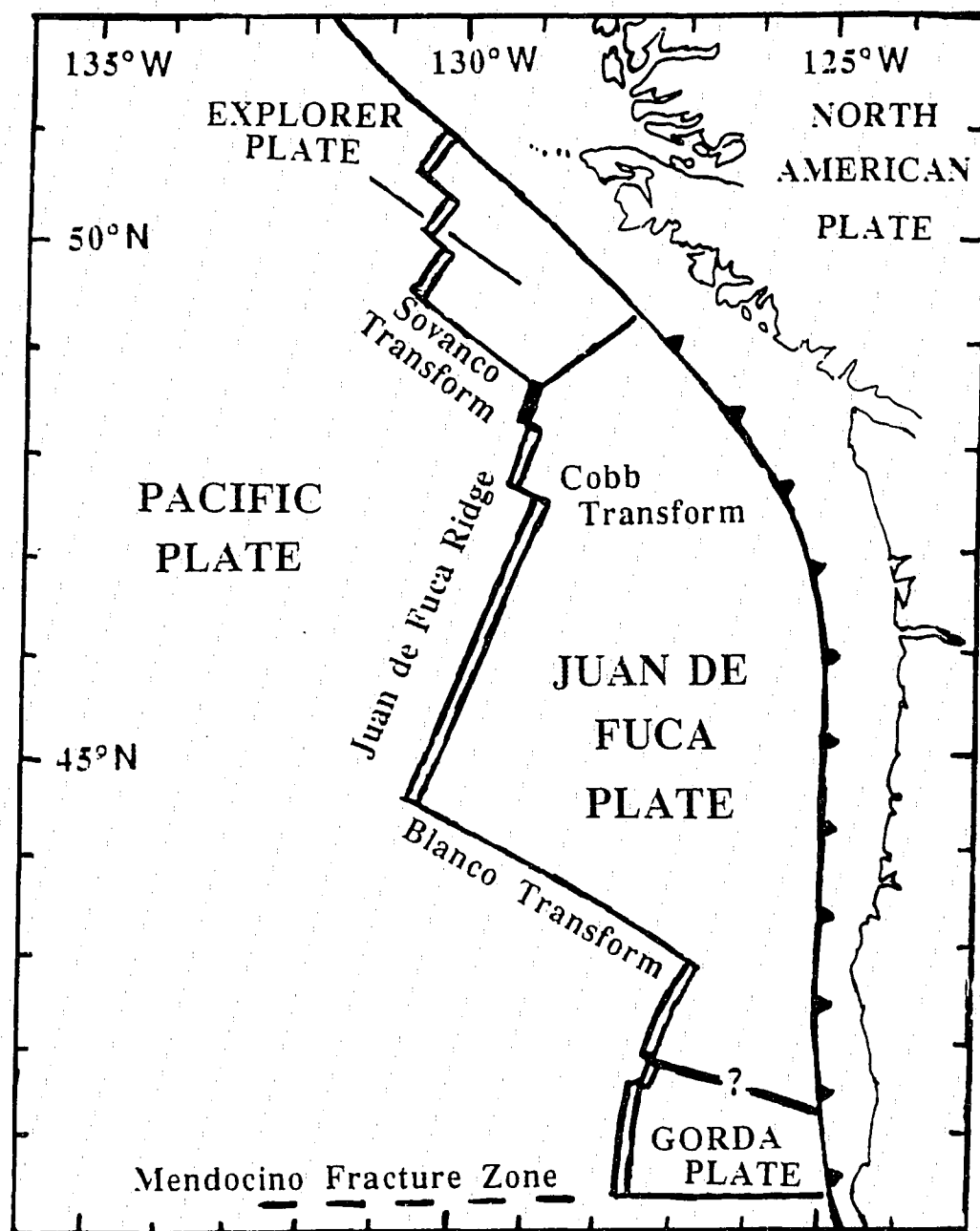


FIGURE i-3
Regional map of the Pacific Northwest Coast illustrating plate boundaries.

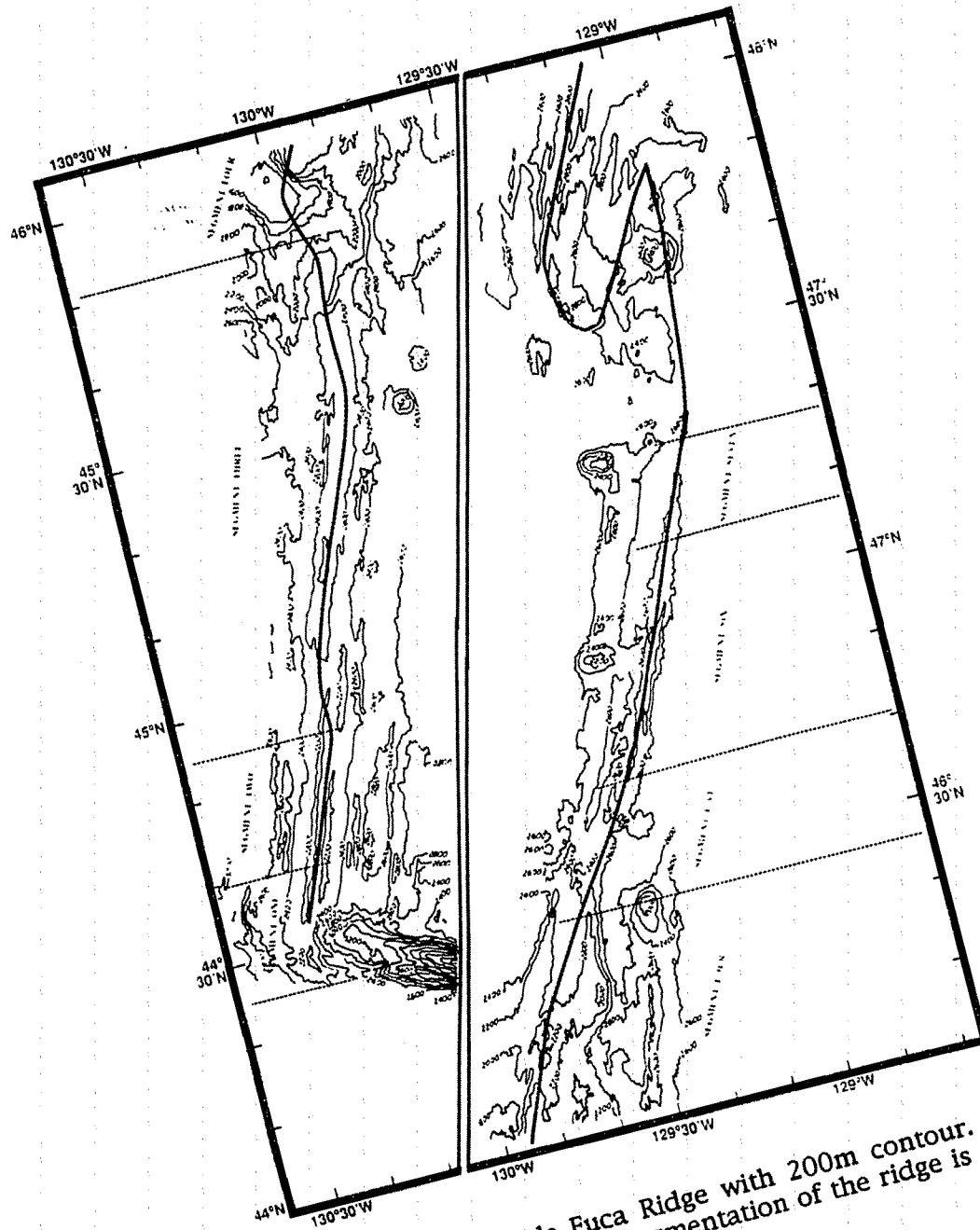


FIGURE i-4
Bathymetry of the Juan de Fuca Ridge with 200m contour. Bathymetric data from Kappel and Ryan (1986). Segmentation of the ridge is also shown.

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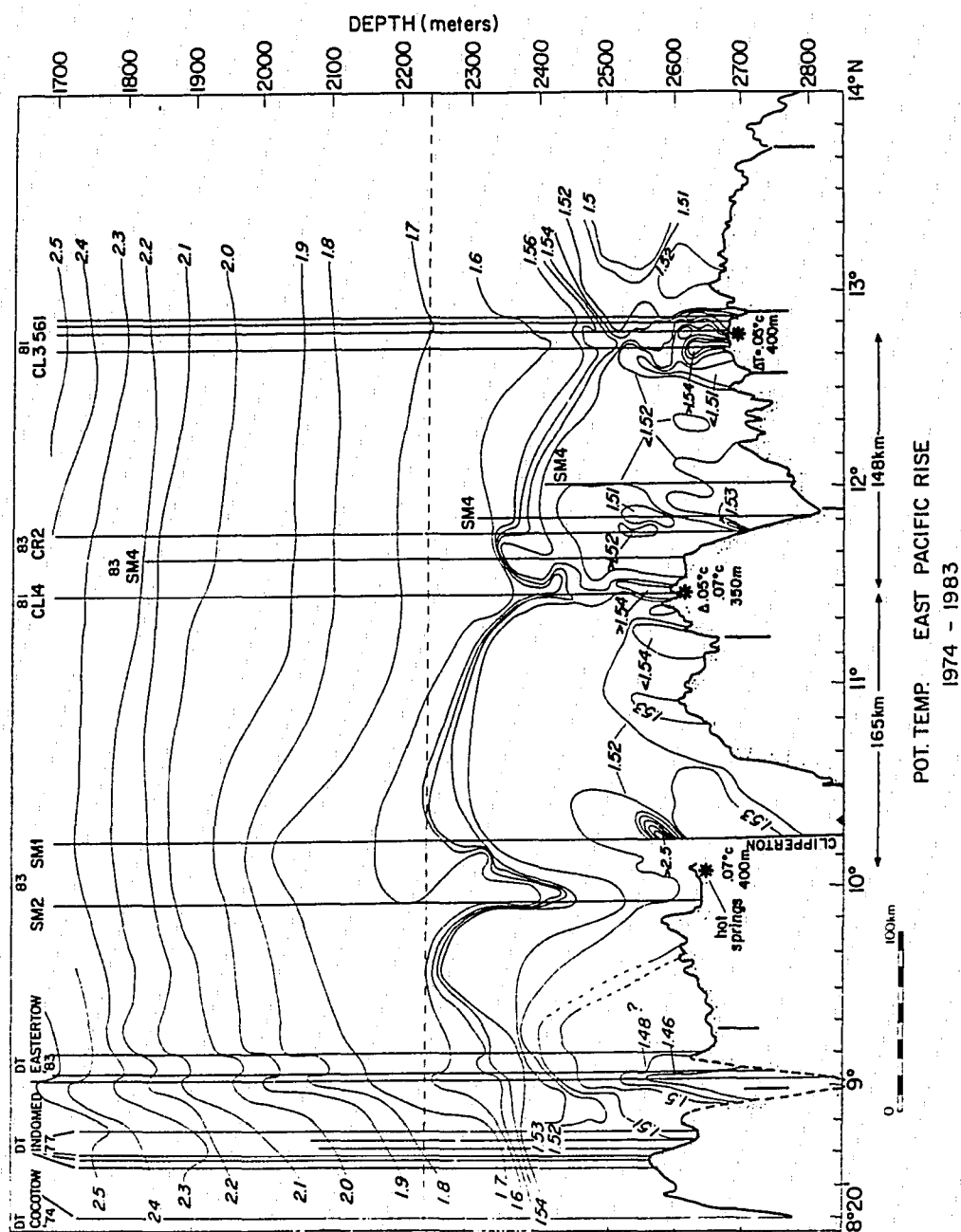


FIGURE epr-1

A compilation of potential temperature data along the EPR rise axis for the years 1974-1983. Vertical lines represent the following: Deep tow CTD profiles (DT), SeaMARC I risings and lowerings (SM), and French hydrocasts (CL). The horizontal dashed line at a depth of 2240 m is the level at which downwards deflected isotherms are noticed. Data from Crane et al. (1988).

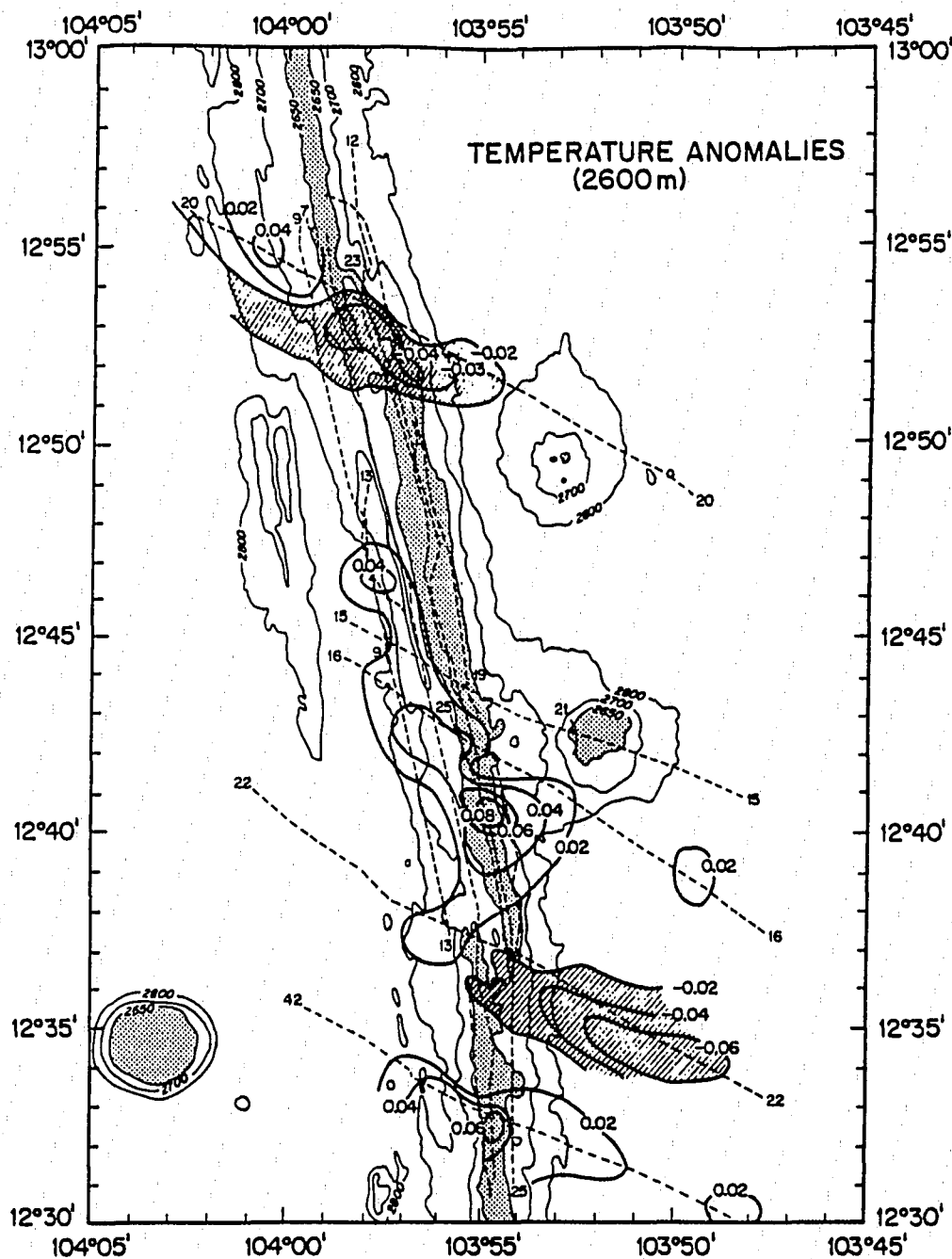


FIGURE epr-2

Temperature anomalies at 2600m depth for EPR 12°30'N-13°00'N. Data from Crane et al. (in prep.).

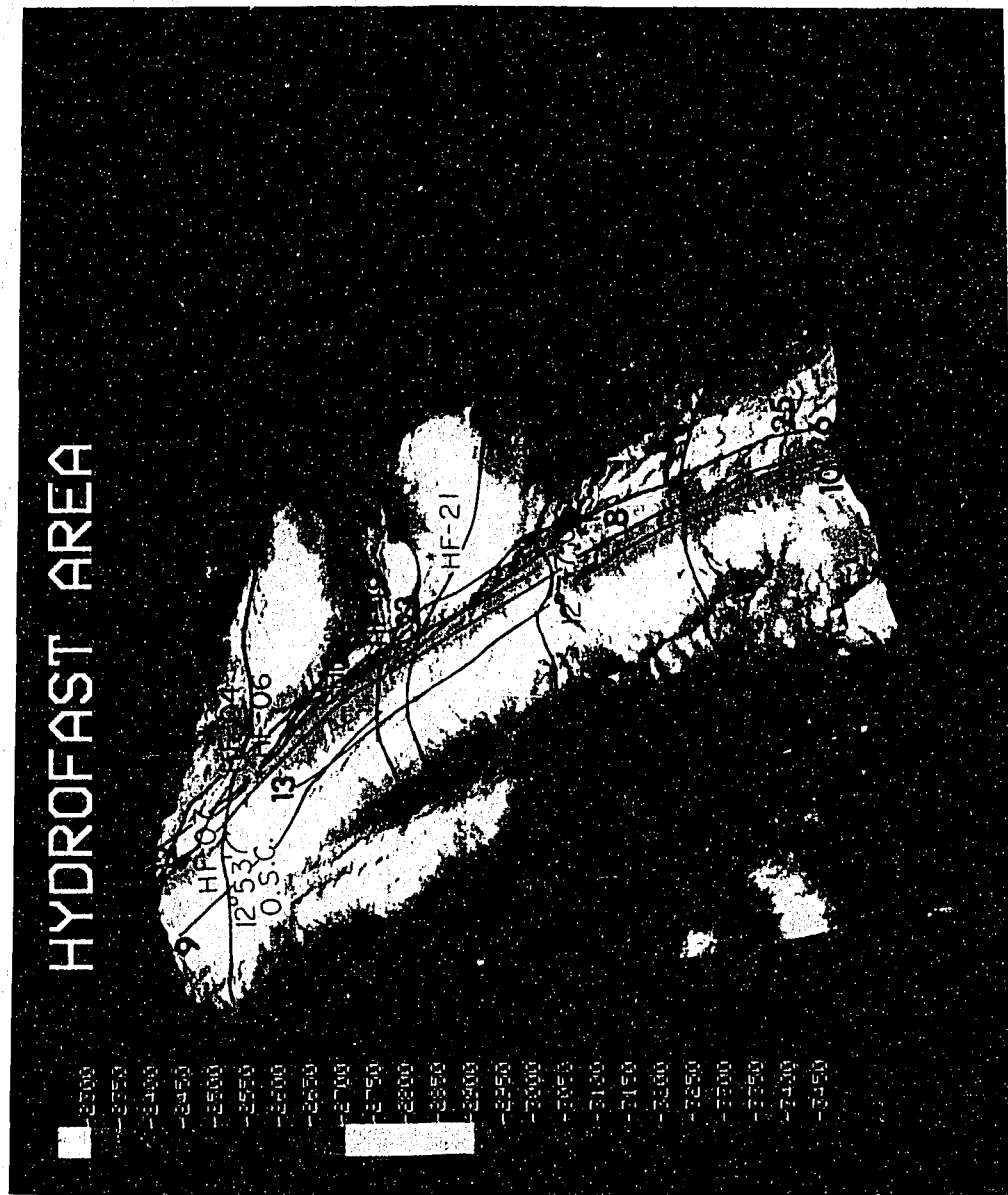


FIGURE epr-3
 Bathymetry of the region along the East Pacific Rise. Superimposed are data tracks (e.g. 10, 12. . .) and CTD locations (e.g. HF-07). Data from Crane et al. (in prep.)

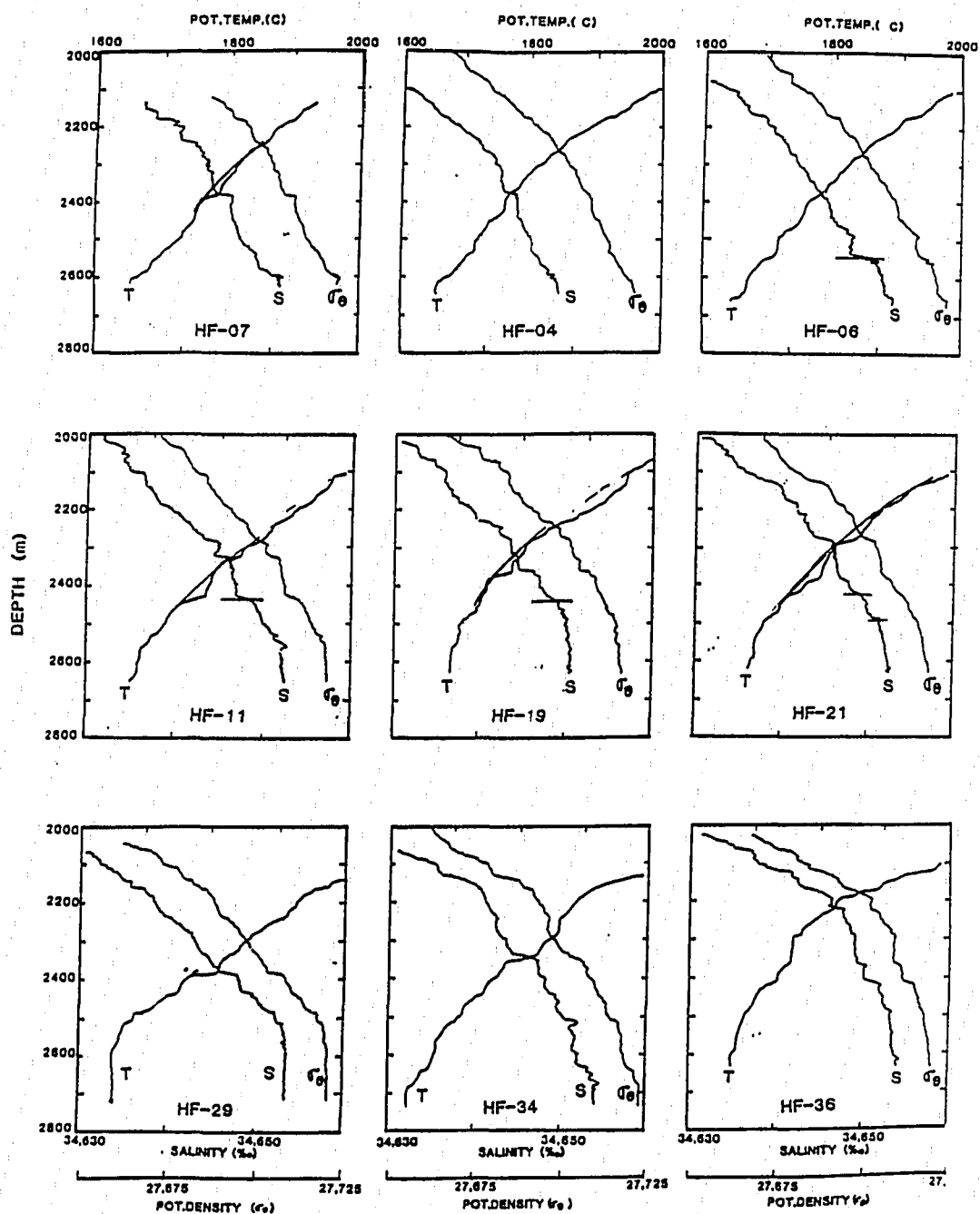


FIGURE epr-4
Potential temperature, Density and Salinity profiles for nine CTD casts along the East Pacific Rise. See FIGURE epr-3 for locations. Data from Crane et al. (in prep.)

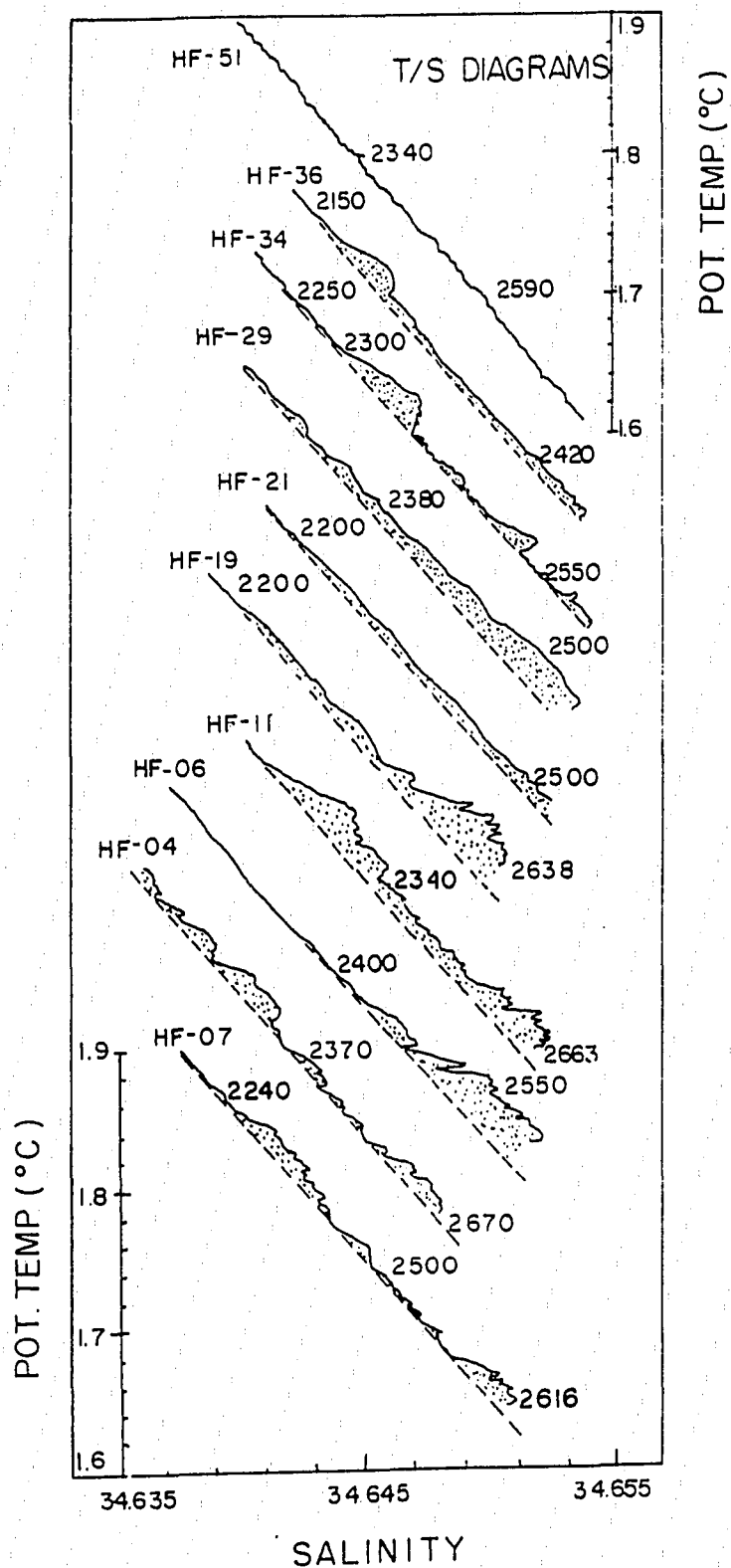


FIGURE epr-5
Potential temperature/salinity diagrams for axial CTDs compared to the reference CTD HF-51 30km west of the axis. Data from Crane et al. (in prep.)

DEPTH TO 1.67°C ISOTHERM

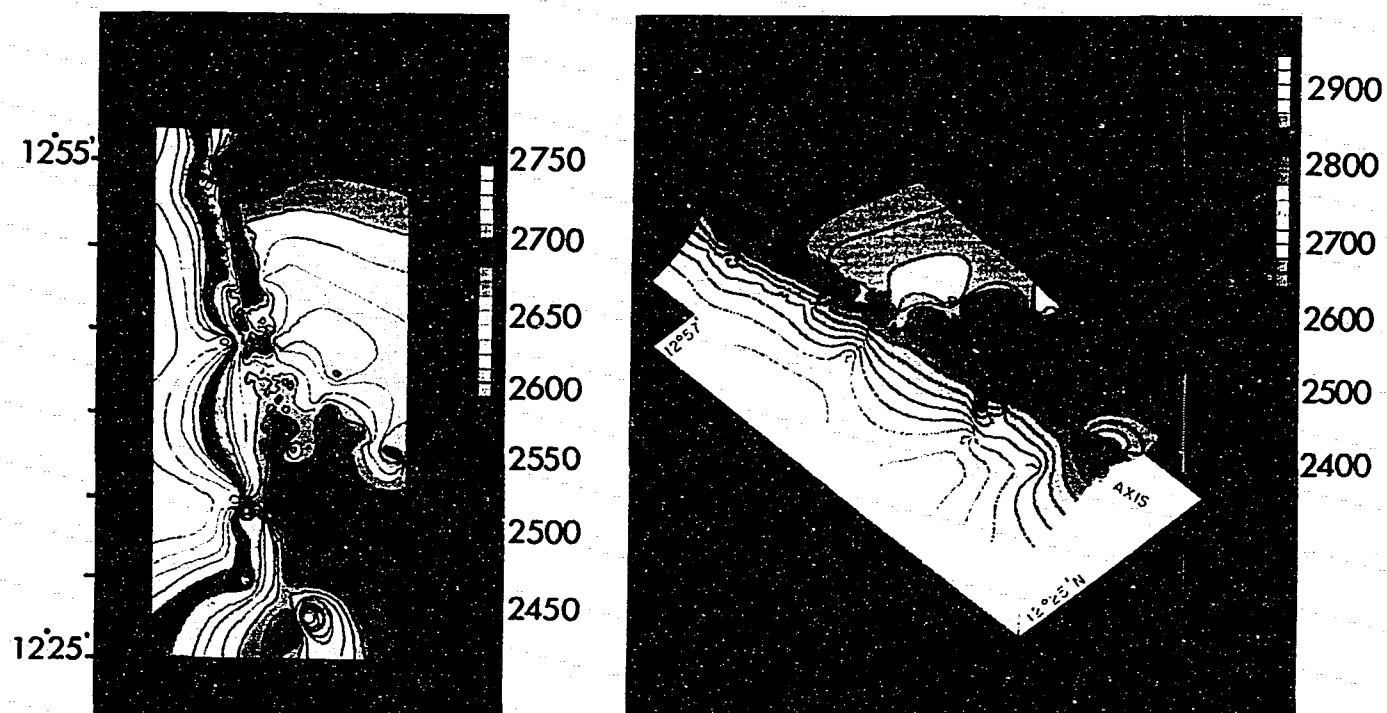


FIGURE epr-6

Depth to 1.67°C isotherm for 12°25'N to 12°57'N on the EPR. Data from Crane et al. (in prep.)

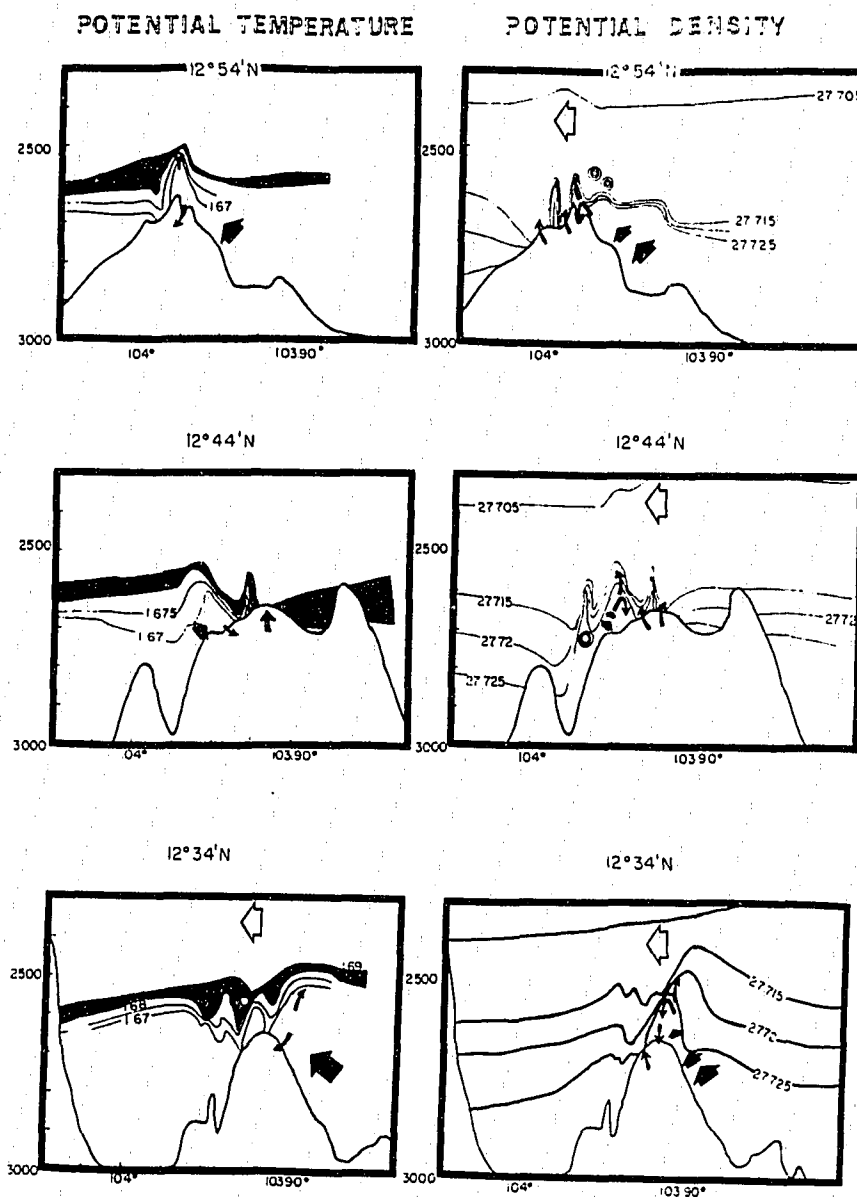


FIGURE epr-7

Potential temperature and density cross sections of the water column above the EPR at latitudes of 12°54'N, 12°44'N and 12°34'N. Arrows indicate inferred flow directions both in the water column and into and out of the subadjacent crust. Data from Crane et al. (in prep.)

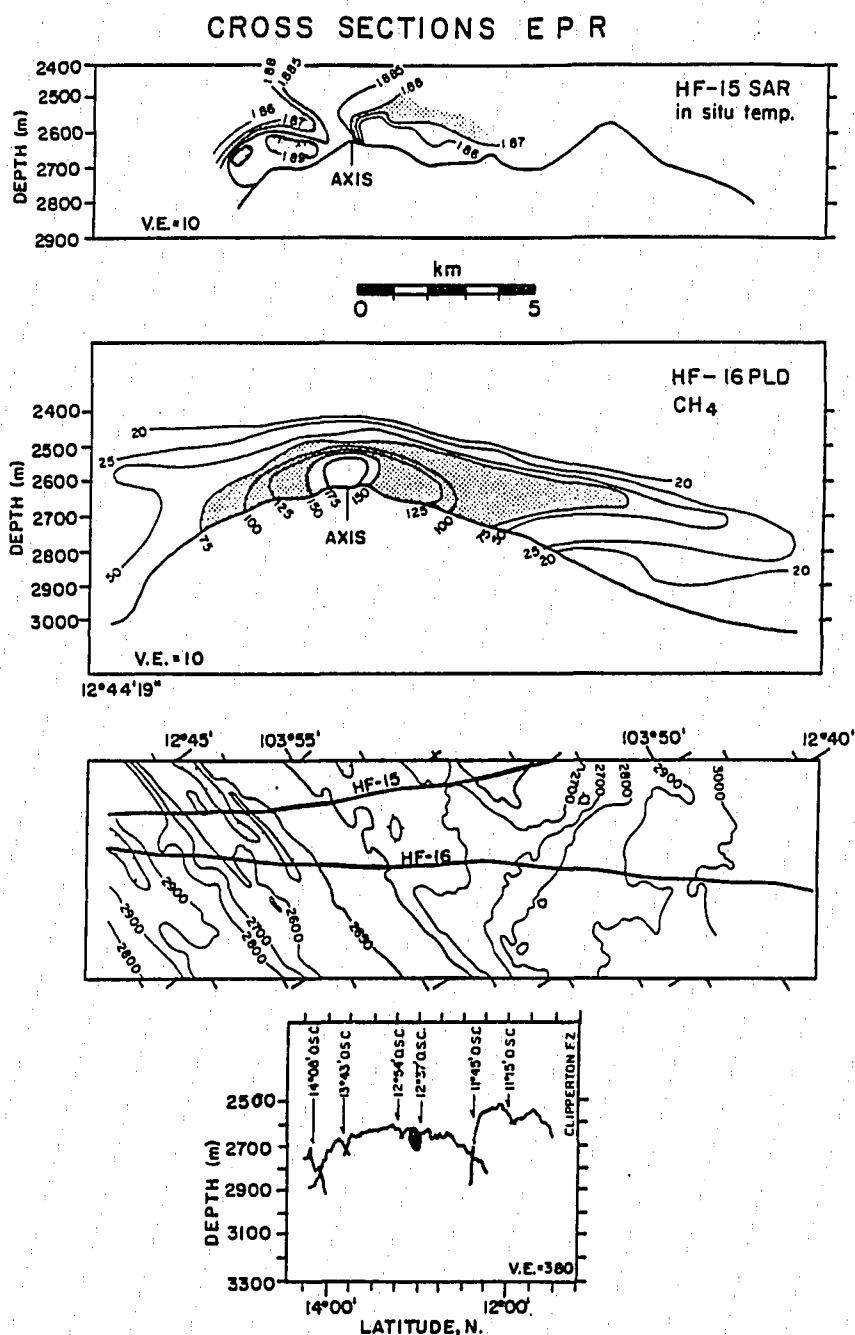


FIGURE epr-8

In situ temperature, methane and bathymetric cross sections of the EPR near 12°40'N. Temperature data suggest that the warm plume is spreading over the rise axis to the west, whereas methane data suggest a spread to the east. Data from Crane et al. (in prep.)

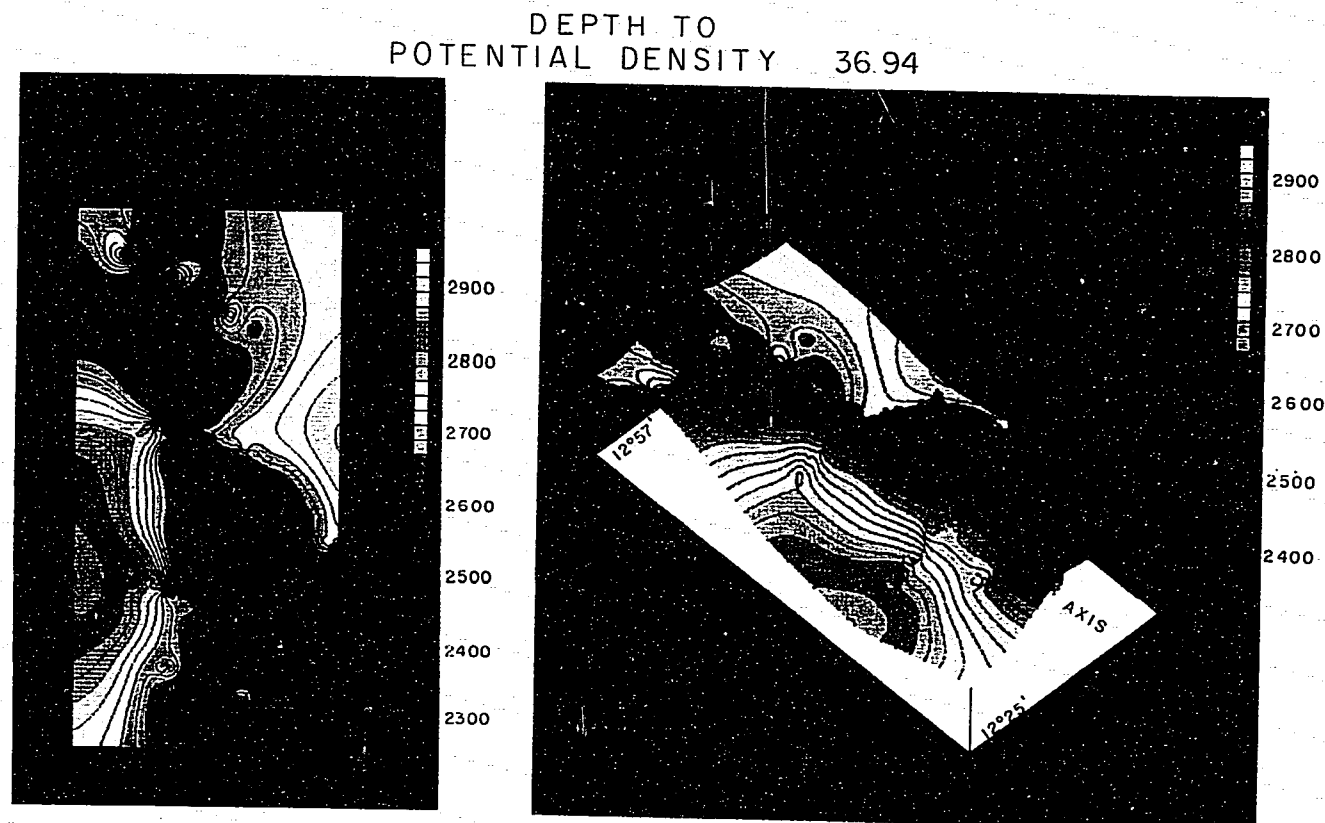


FIGURE epr-9
Plan view of the depth to the 36.94 isopychnal. Data from Crane et al. (in prep.)

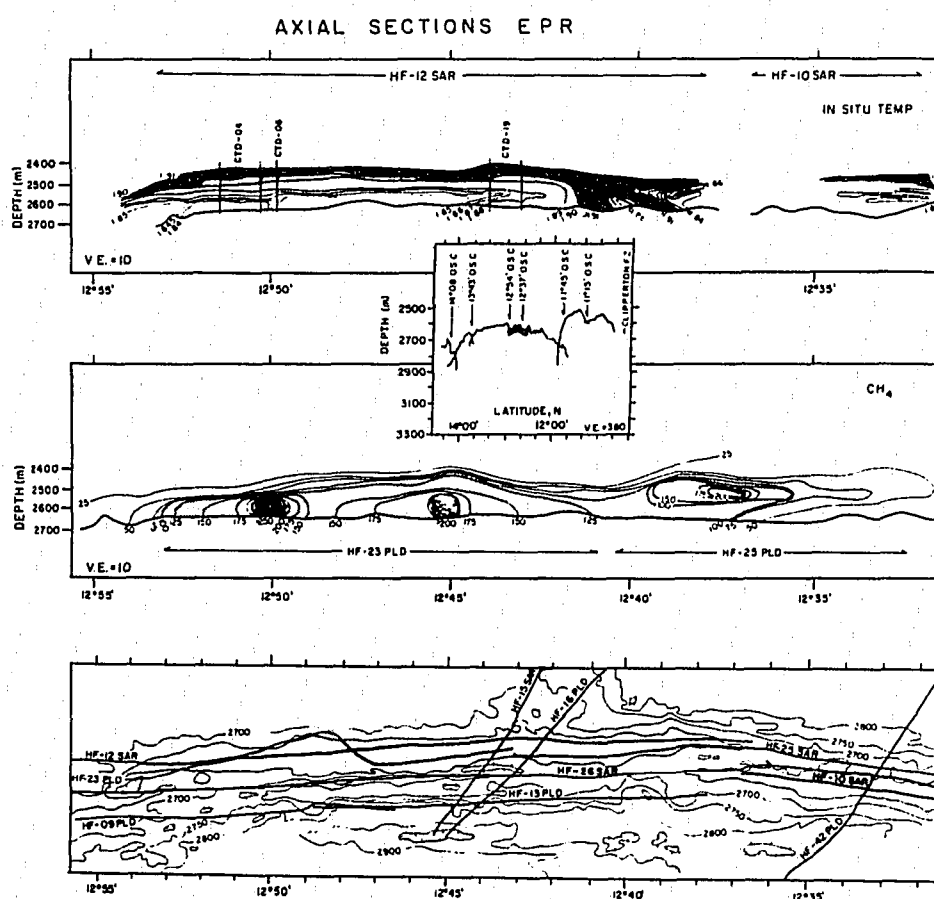


FIGURE epr-10

In situ temperature, methane, and bathymetry along the EPR axis from 12°56'N to 12°30'N. Methane indicates three regions of enhanced concentration (two on the bottom and one lifted off bottom). Data from Crane et al. (in prep.)

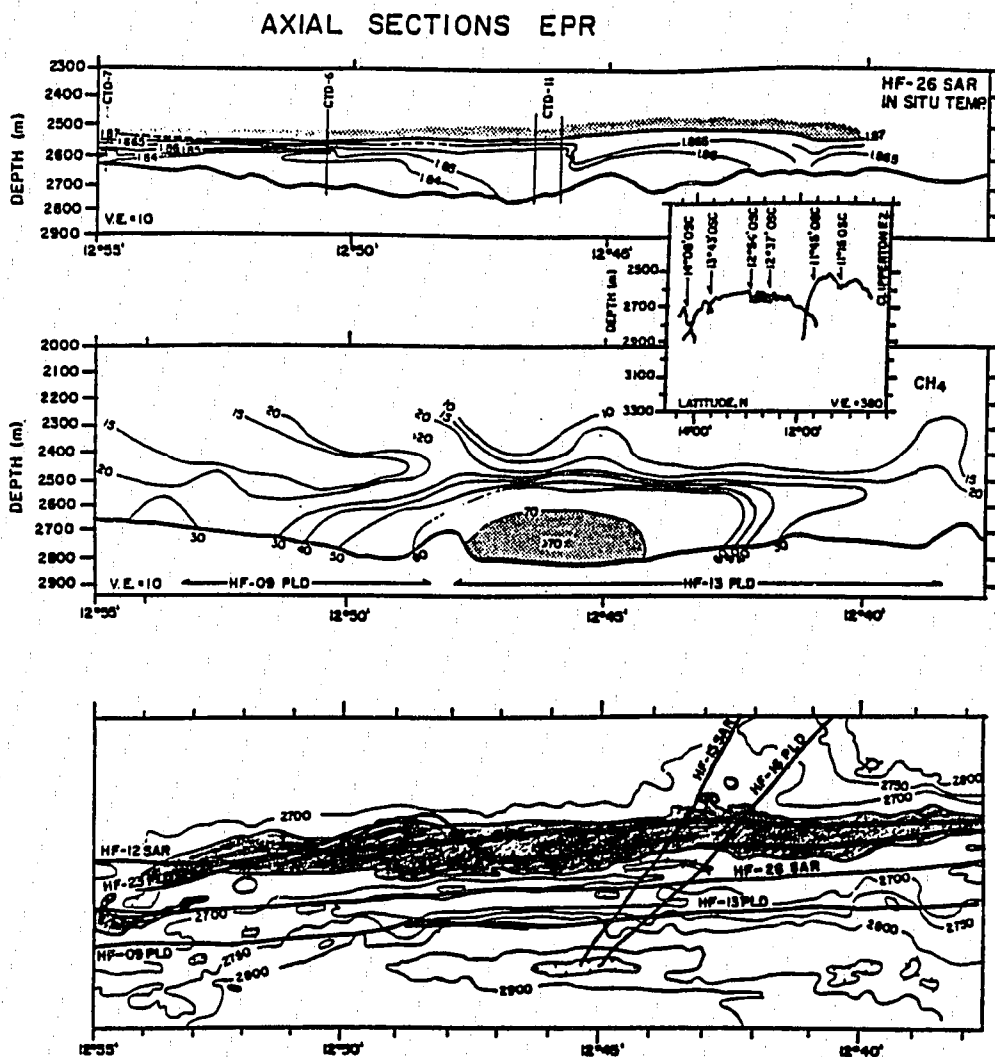


FIGURE epr-11

In situ temperature, methane and bathymetry for tracks of the SAR and PLD taken parallel to, but west of the axis. The downward deflection of isotherms is still detectable near 12°40'N as is the region of anomalous methane concentration. In this section only one region of activity is detected in both the temperature and methane curves. The gray region in the bathymetry indicates the rise axis. Data from Crane et al. (in prep.)

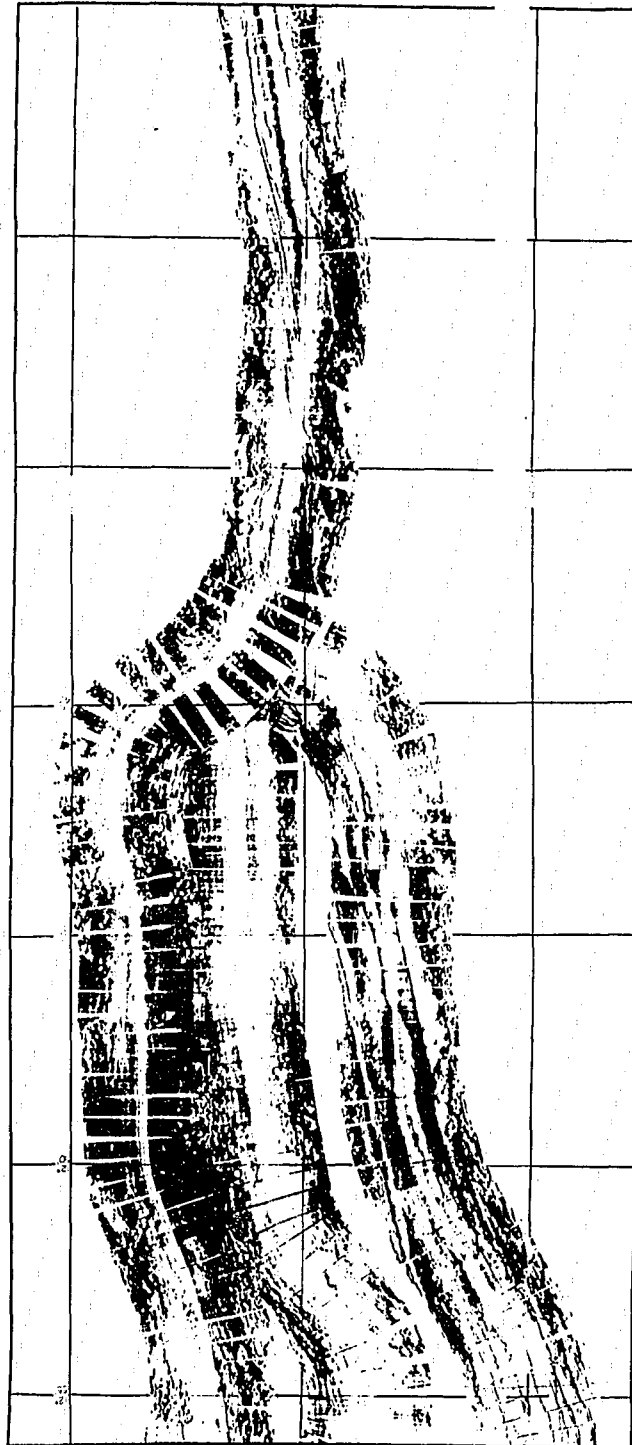


FIGURE epr-12
SeamARC data for 12025'N to 12057'N on the EPR (Data from Perram and Macdonald, 1990).

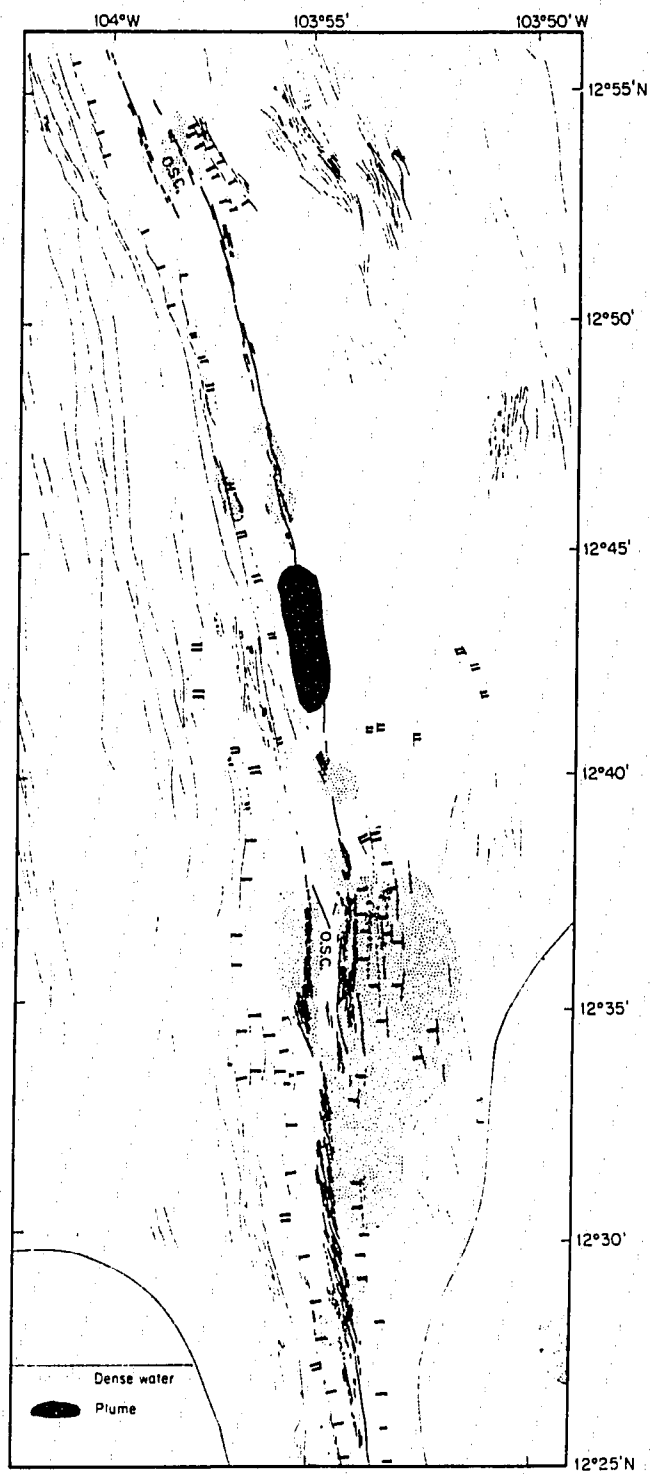
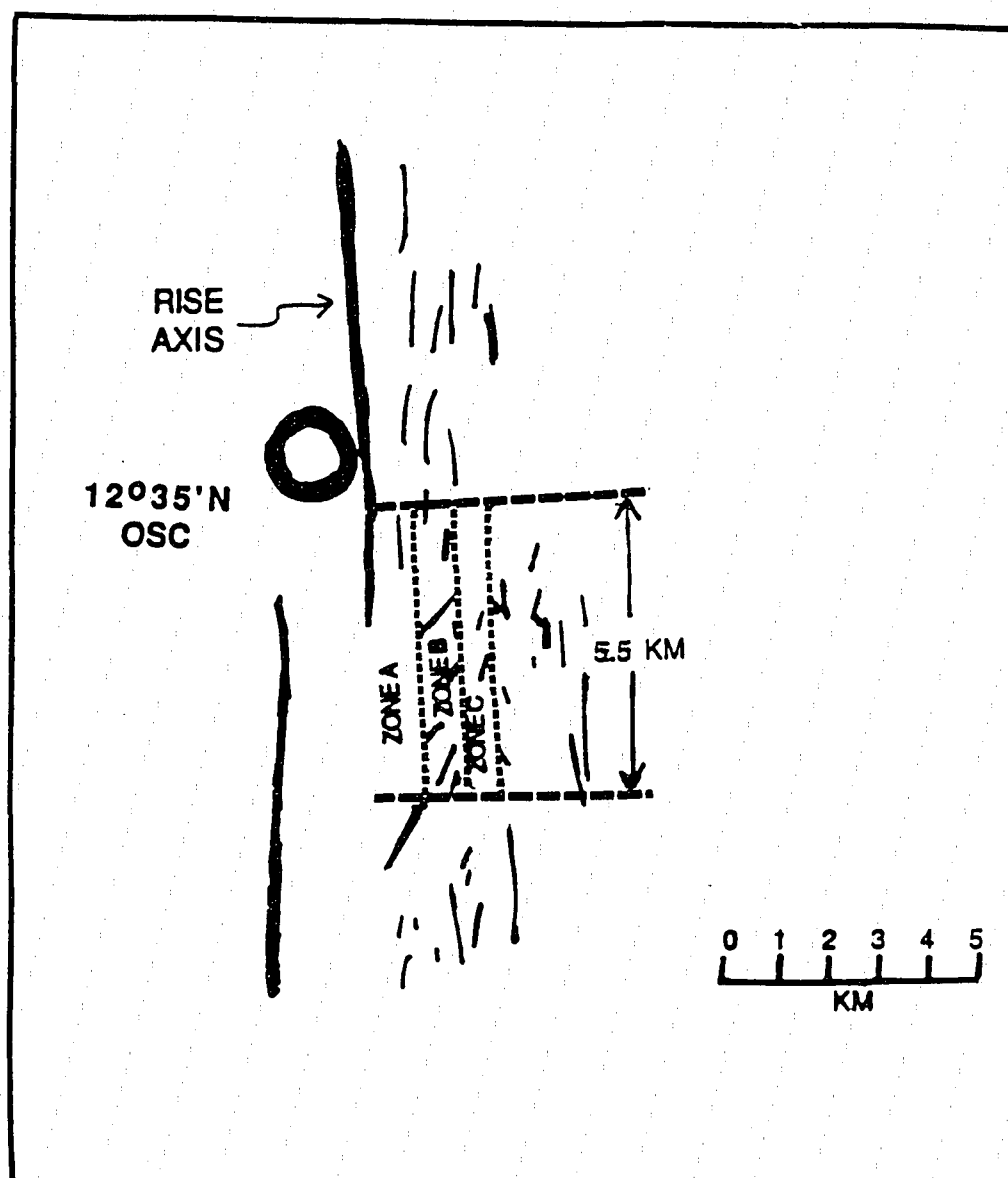


FIGURE epr-13
 Sea floor structures mapped by SeaMARC I and SeaMARC II with locations of a very shallow (2540m) and a very deep (2680m) 36.94 isopycnal surface. Data from Crane et al. (in prep.)



Within the density anomaly extending for 5.5 km three zones are chosen. Zone A has a total fault and fissure length of 1.19km. Zone B has a total fault and fissure length of 5.85 km. Zone C has a total fault and fissure length of 1.01 km.

FIGURE epr-14

Detail of structure at 12°35'N OSC from SeaMARC II data (Data from Perram and Macdonald, 1990). The 5.5km zone is the region of greatest density anomaly.

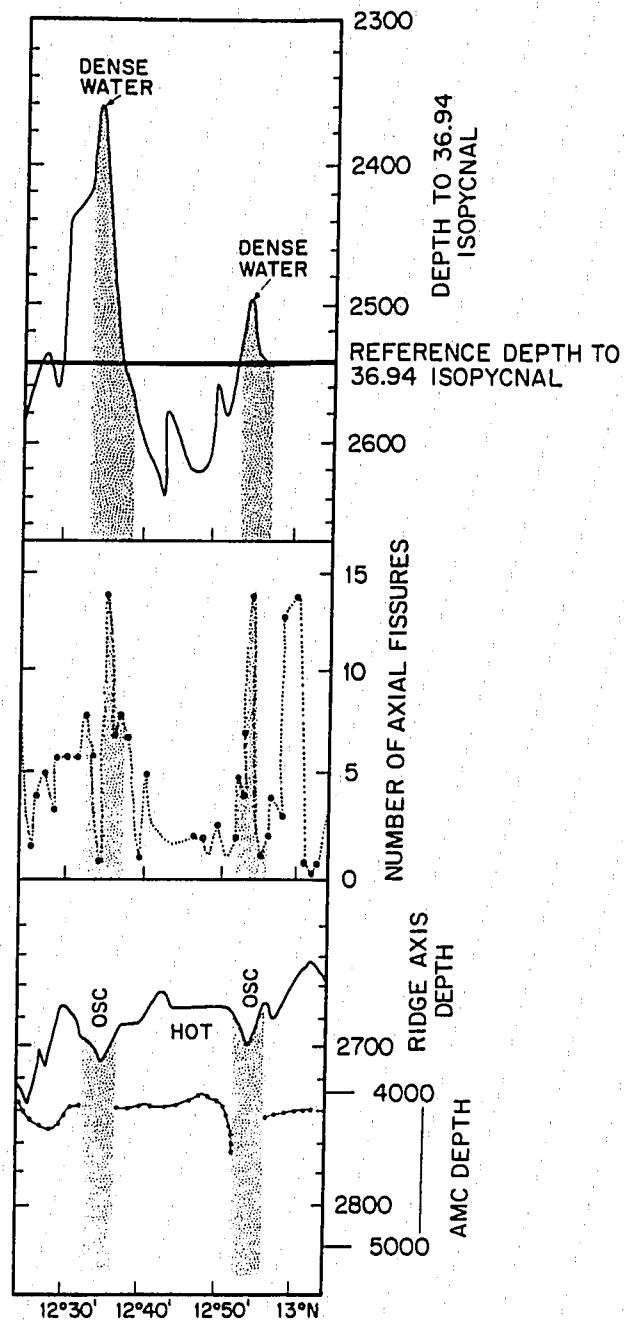


FIGURE epr-16

Comparison of isopycnal depth with number of axial fissures, bathymetry and depth to the subjacent axial magma chamber (AMC) [Data from Detrick et al., 1987] relative to latitude from 12°25'N to 13°N on the East Pacific Rise. Dense water has accumulated over the OSCs which are highly fractured and are locations where the AMC is non-existent. Data from Crane et al. (in prep.)

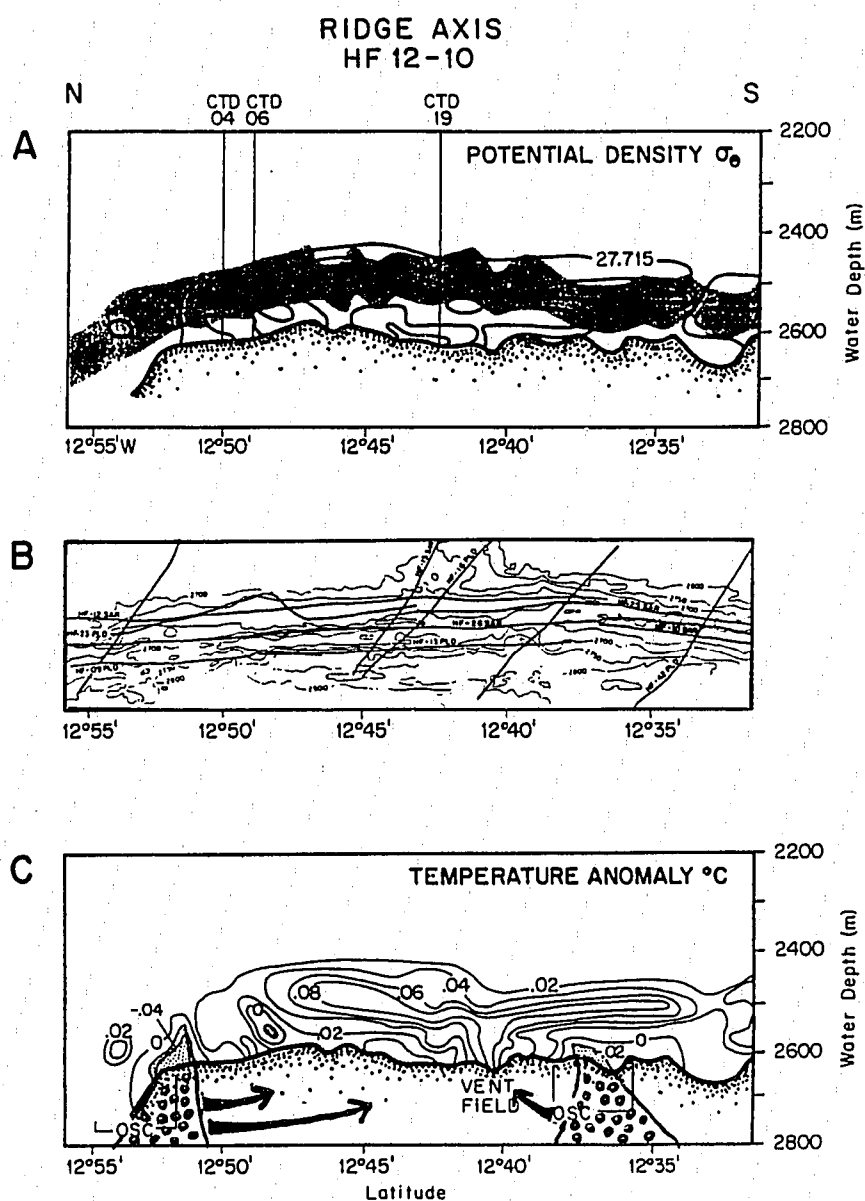


FIGURE epr-17

Axial Cross-sections of temperature anomalies and isopycnals at Overlapping Spreading Centers derived from SAR 8, 12, 20, 22, 26, CTD's 4, 7, and 51 and PLD's 9, 13, 23 and 25. Anomalously cold water funnels are located directly over the Overlapping Spreading Centers while the warm plume is centered at 12°40'N. Proposed recharge is illustrated in the bottom diagram. Data from Crane et al. (in prep.)

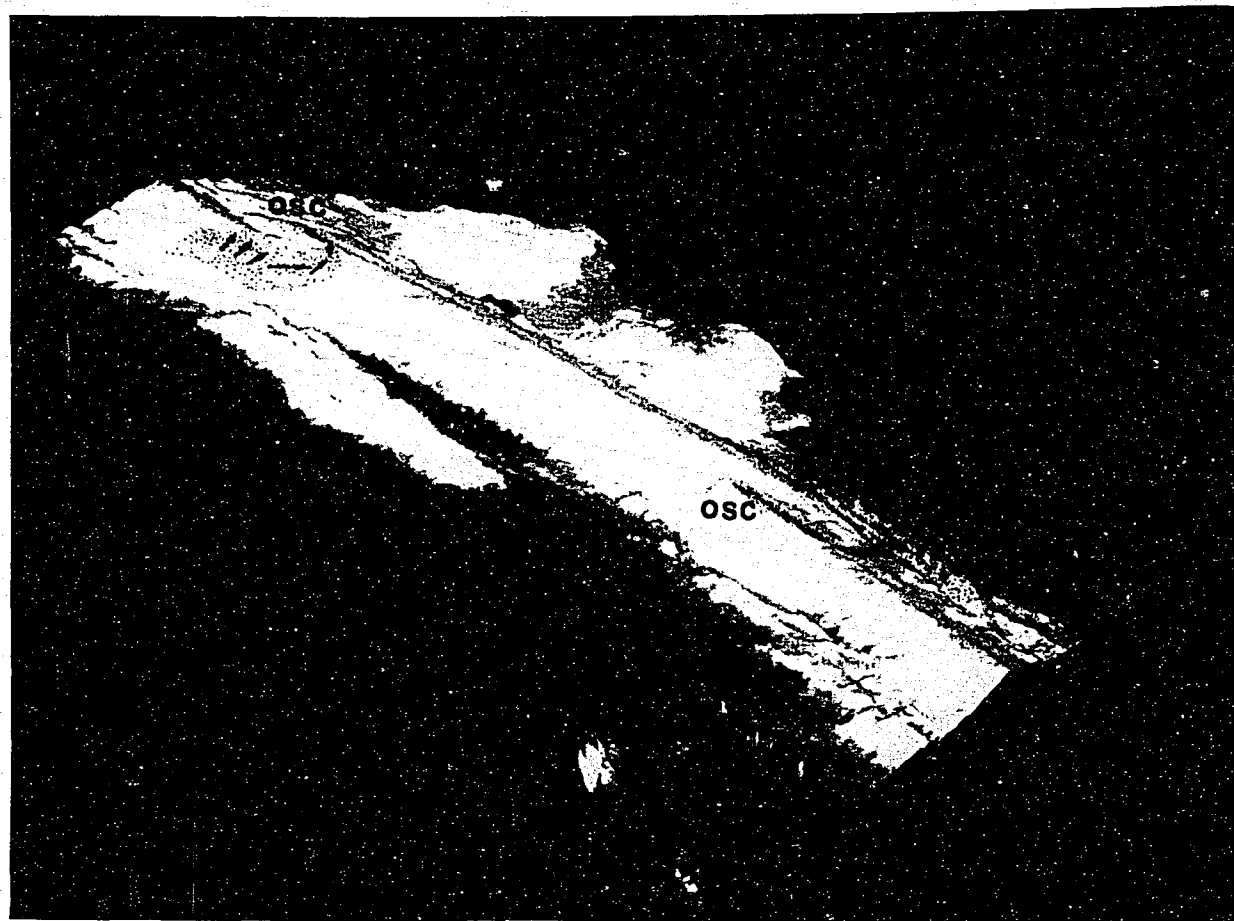


FIGURE epr-18

Proposed pathways of cold water into the crust at OSCs at $12^{\circ}35'N$ and $12^{\circ}54'N$ and beneath the axis between them. Data from Crane et al. (in prep.)

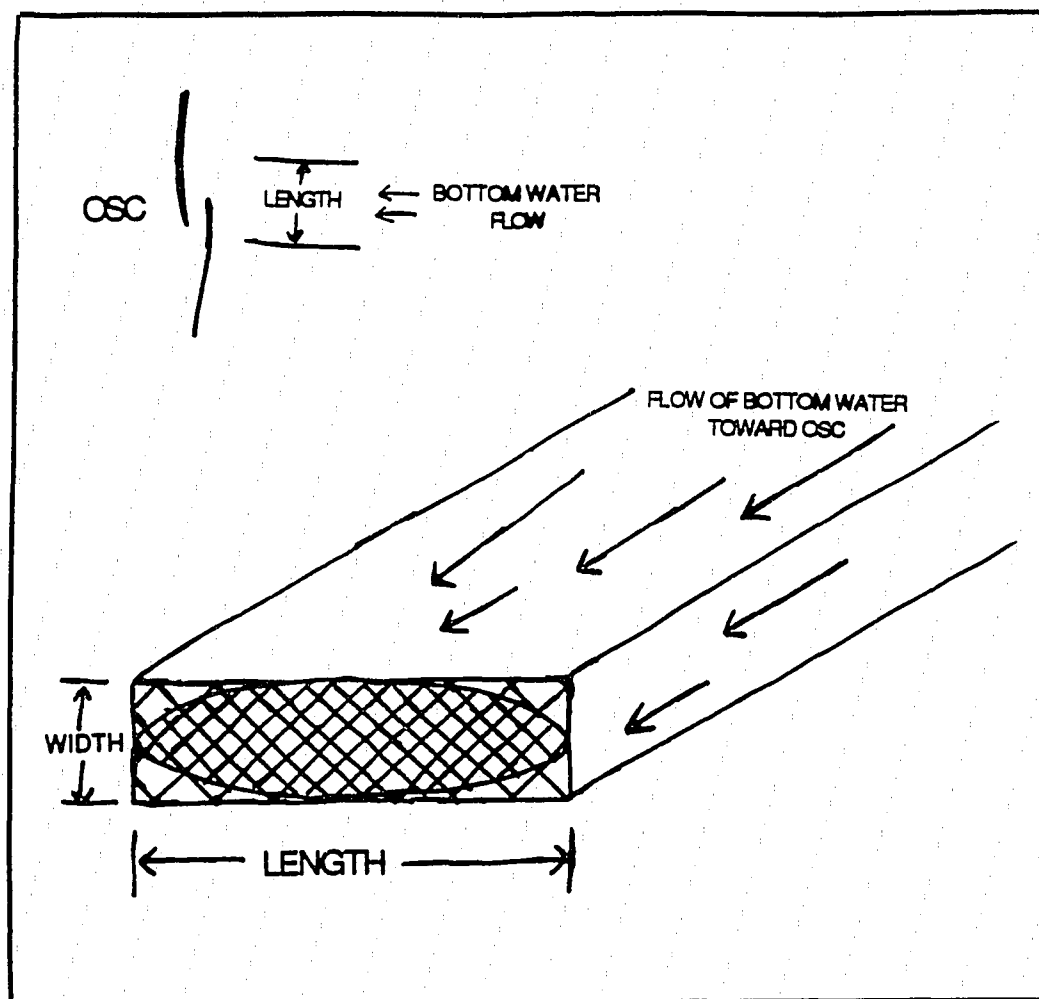


FIGURE epr-19
Cross sectional area of bottom flow represented in both a rectangular geometry and an oval geometry.

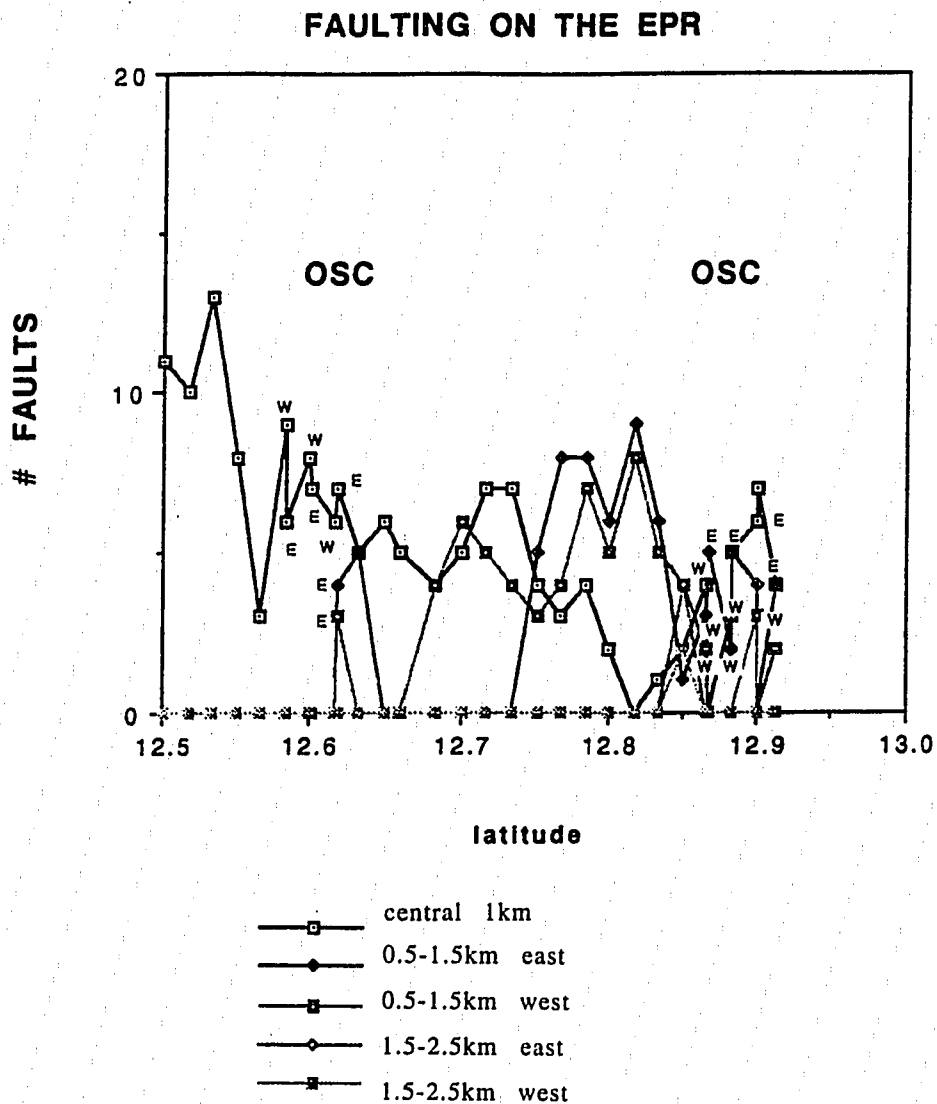


FIGURE epr-20
 Faulting on the EPR. Axial faulting (from SeaMARC data) is shown for the 1km at the center of the rise axis and for 1km to the east and west of the central faulting region. The OSCs are marked. Values for eastern and western legs of the spreading center at the OSCs are marked with 'E' or 'W'.

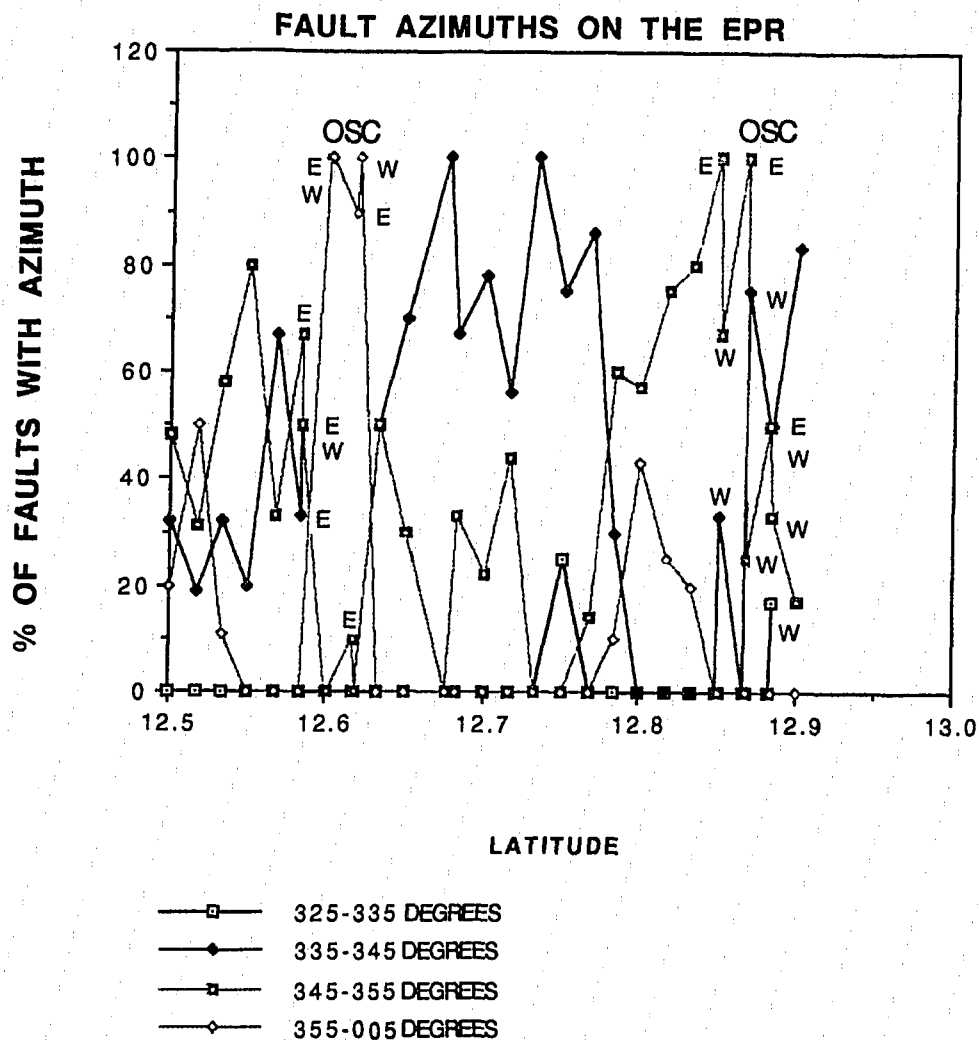


FIGURE epr-21

Fault azimuths on the EPR (from seaMARC data). The percentage of faults in each 10° arc is given for each minute of latitude. The OSC are marked. Values for eastern and western legs of the spreading center at the OSCs are marked with 'E' or 'W'.

EAST PACIFIC RISE FAULT LENGTHS

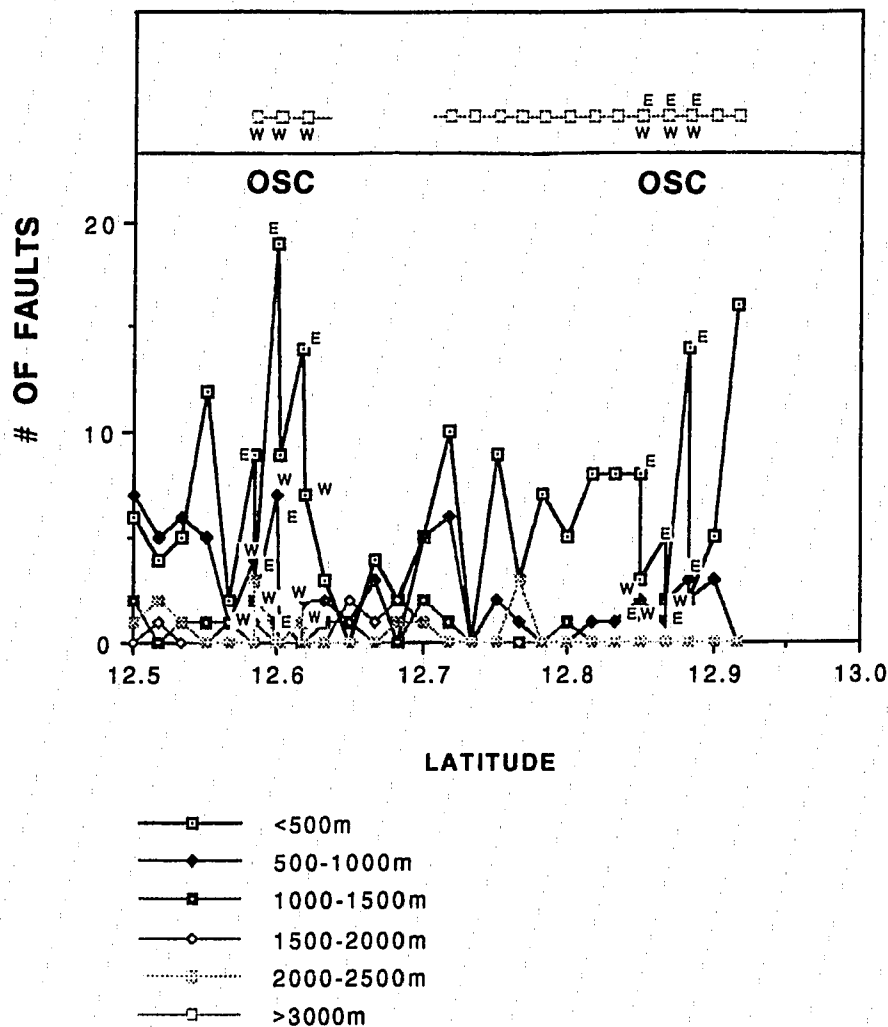


FIGURE epr-22

The number of faults with each length (from seaMARC data) is given at each minute of latitude for faults up to 2500 m. Faulting over 3000 m is indicated, but the number of such faults is not given. These long faults are boundary faults of the rift valley.

LENGTH = 0.50 km

width (m)	RECTANGULAR CROSS-SECTION velocity (cm/s)	OVAL CROSS-SECTION velocity (cm/s)
0.01	64.4	82.
0.10	6.44	8.20
1.00	0.64	0.82
10.00	0.06	0.08

LENGTH = 1.00 km

width (m)	RECTANGULAR CROSS-SECTION velocity (cm/s)	OVAL CROSS-SECTION velocity (cm/s)
0.01	32.2	41.0
0.10	3.22	4.10
1.00	0.32	0.41
10.0	0.03	0.04

LENGTH = 1.50 km

width (m)	RECTANGULAR CROSS-SECTION velocity (cm/s)	OVAL CROSS-SECTION velocity (cm/s)
0.01	21.5	27.3
0.10	2.15	2.73
1.00	0.21	0.27
10.00	0.02	0.03

LENGTH = 2.00 km

width (m)	RECTANGULAR CROSS-SECTION velocity (cm/s)	OVAL CROSS-SECTION velocity (cm/s)
0.01	16.1	20.5
0.10	1.61	2.05
1.00	0.16	0.21
10.00	0.02	0.02

LENGTH = 2.50 km

width (m)	RECTANGULAR CROSS-SECTION velocity (cm/s)	OVAL CROSS-SECTION velocity (cm/s)
0.01	12.9	16.4
0.10	1.29	1.64
1.00	0.13	0.16
10.00	0.01	0.02

LENGTH = 3.00 km

width (m)	RECTANGULAR CROSS-SECTION velocity (cm/s)	OVAL CROSS-SECTION velocity (cm/s)
0.01	10.7	13.7
0.10	1.07	1.37
1.00	0.11	0.14
10.00	0.01	0.01

LENGTH = 3.50 km

width (m)	RECTANGULAR CROSS-SECTION velocity (cm/s)	OVAL CROSS-SECTION velocity (cm/s)
0.01	9.20	11.7
0.10	0.92	1.17
1.00	0.09	0.12
10.00	0.01	0.01

LENGTH = 4.00 km

width (m)	RECTANGULAR CROSS-SECTION velocity (cm/s)	OVAL CROSS-SECTION velocity (cm/s)
0.01	8.05	10.2
0.10	0.81	1.02
1.00	0.08	0.10
10.00	0.01	0.01

LENGTH = 4.50 km

width (m)	RECTANGULAR CROSS-SECTION velocity (cm/s)	OVAL CROSS-SECTION velocity (cm/s)
0.01	7.16	9.11
0.10	0.72	0.91
1.00	0.07	0.09
10.00	0.01	0.01

LENGTH = 5.00 km

width (m)	RECTANGULAR CROSS-SECTION velocity (cm/s)	OVAL CROSS-SECTION velocity (cm/s)
0.01	6.44	8.20
0.10	0.64	0.82
1.00	0.06	0.08
10.00	0.01	0.01

LENGTH = 5.50 km

width (m)	RECTANGULAR CROSS-SECTION velocity (cm/s)	OVAL CROSS-SECTION velocity (cm/s)
0.01	5.85	7.45
0.10	0.59	0.75
1.00	0.06	0.07
10.00	0.01	0.01

TABLE epr-1

Off axis near-bottom flow velocities along the sea floor for up to 5.5km length and up to 10.00m width. See FIGURE epr-19 for illustration of width and length of water intake.

1 METER WIDTH		POROSITY SAME AT DEPTH	POROSITY DECREASE AT DEPTH
length (km)	velocity (cm/s)	residence time (years)	residence time (years)
1	0.322	0.12	0.72
2	0.161	0.24	1.44
3	0.107	0.36	2.16
4	0.081	0.48	2.88
5	0.064	0.61	3.66
6	0.054	0.72	4.32
7	0.046	0.85	5.88

2 METER WIDTH		POROSITY SAME AT DEPTH	POROSITY DECREASE AT DEPTH
length (km)	velocity (cm/s)	residence time (years)	residence time (years)
1	0.161	0.24	1.44
2	0.081	0.48	2.88
3	0.054	0.72	4.32
4	0.041	0.95	5.70
5	0.032	1.22	7.32
6	0.027	1.44	8.64
7	0.023	1.70	10.20
8	0.020	1.95	11.70

3 METER WIDTH		POROSITY SAME AT DEPTH	POROSITY DECREASE AT DEPTH
length (km)	velocity (cm/s)	residence time (years)	residence time (years)
1	0.107	0.36	2.16
2	0.054	0.72	4.32
3	0.036	1.08	6.48
4	0.027	1.44	8.64
5	0.021	1.86	11.16
6	0.015	2.60	15.60
7	0.013	3.00	18.00
8	0.010	3.40	20.40

4 METER WIDTH		POROSITY SAME AT DEPTH	POROSITY DECREASE AT DEPTH
length (km)	velocity (cm/s)	residence time (years)	residence time (years)
1	0.081	0.48	2.88
2	0.040	0.98	5.88
3	0.027	1.44	8.64
4	0.020	1.98	11.88
5	0.016	2.44	14.64
6	0.014	2.79	16.74
7	0.012	3.25	19.50
8	0.010	3.90	23.40

5 METER WIDTH		POROSITY SAME AT DEPTH	POROSITY DECREASE AT DEPTH
length (km)	velocity (cm/s)	residence time (years)	residence time (years)
1	0.064	0.61	3.66
2	0.032	1.22	7.32
3	0.021	1.86	11.16
4	0.016	2.44	14.64
5	0.013	3.00	18.00
6	0.011	3.55	21.30
7	0.009	4.33	25.98
8	0.008	4.88	29.28

TABLE epr-2

Conduit flow velocities and residence time for total inlet length of up to 8km for faults and fissures up to 5m wide. Residence times are calculated based on a 12.3 km journey through the crust. The water travels 1.5km downward to the magma chamber (based on average distance to the AMC), 9.3km along the top of the chamber and another 1.5km back to the sea floor surface. See FIGURE epr-13 for interpretation of length of faults and fissures available for entrance of water.

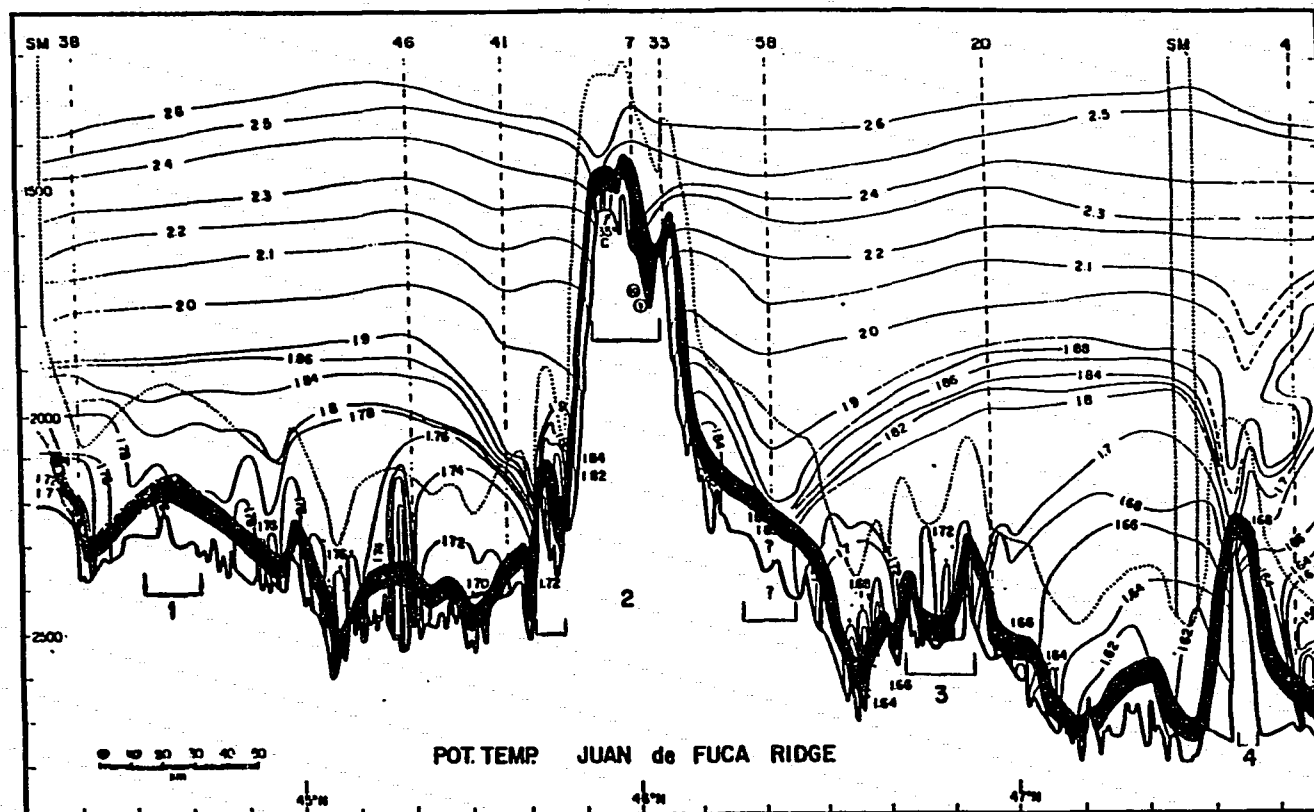


FIGURE jdf-1

Aanderaa, CTD and hydrocast temperature data superimposed over SeaMARC I bathymetry. Four regions marked 1,2,3 and 4 are unusually warm, as indicated by depressed isotherms over the rise. Profiles 38, 46, 20 and 4 are CTD and hydrocast stations taken in the same month as the SeaMARC I survey. Dotted vertical lines indicate SeaMARC I risings or lowerings. These vertical profiles are the control on the upper water isotherms. The gray band near the bottom is the path covered by the Aanderaa thermistor chain. The dotted profile is the SeaMARC I path. Isotherms drawn in between the controls of the vertical profiles and the SeaMARC data are constructed by interpolations of the data sets. Data from Crane et al. (1985).

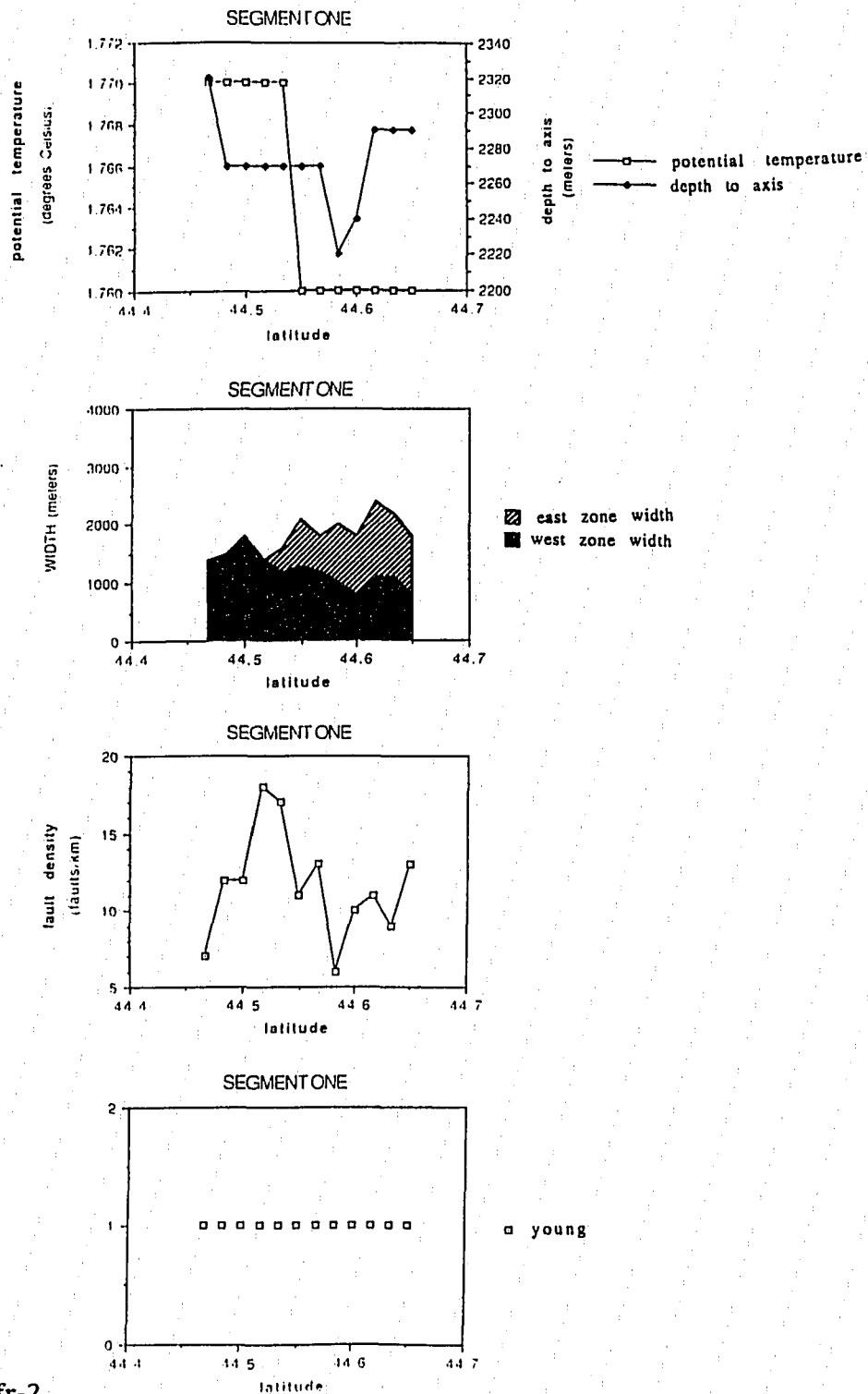


FIGURE jdfr-2
Segment one - Bathymetry and potential temperature from Crane et al. (1985). Zonal width, fault density and age assessment from SeaMARC I images.

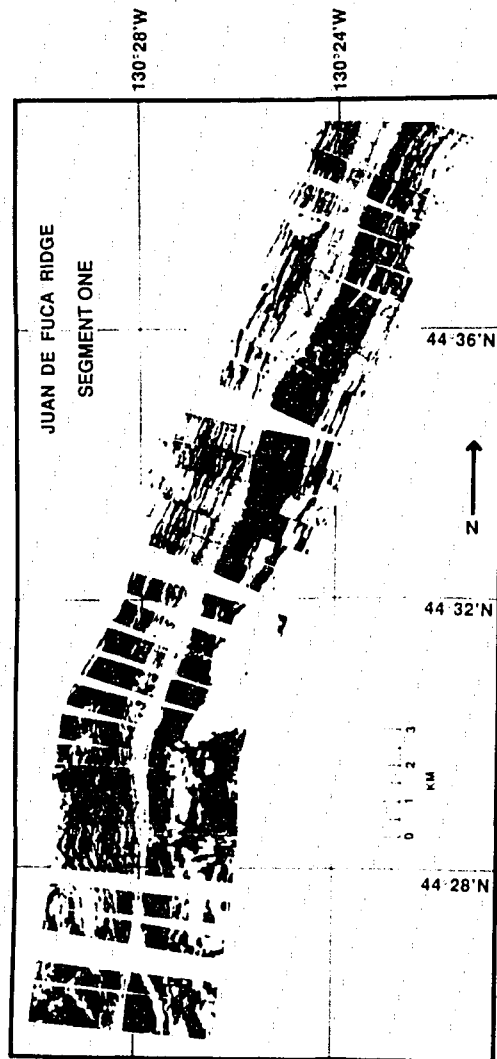


FIGURE jdf-3
Segment one - SeaMARC I image.

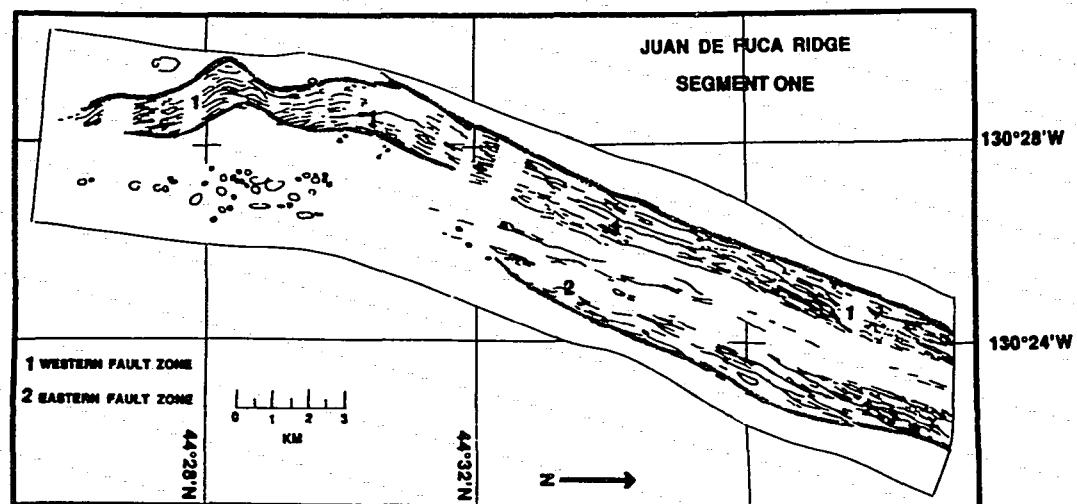


FIGURE jdfr-4
Segment one - Interpretation and evaluation of SeaMARC I image (FIGURE jdfr-3)

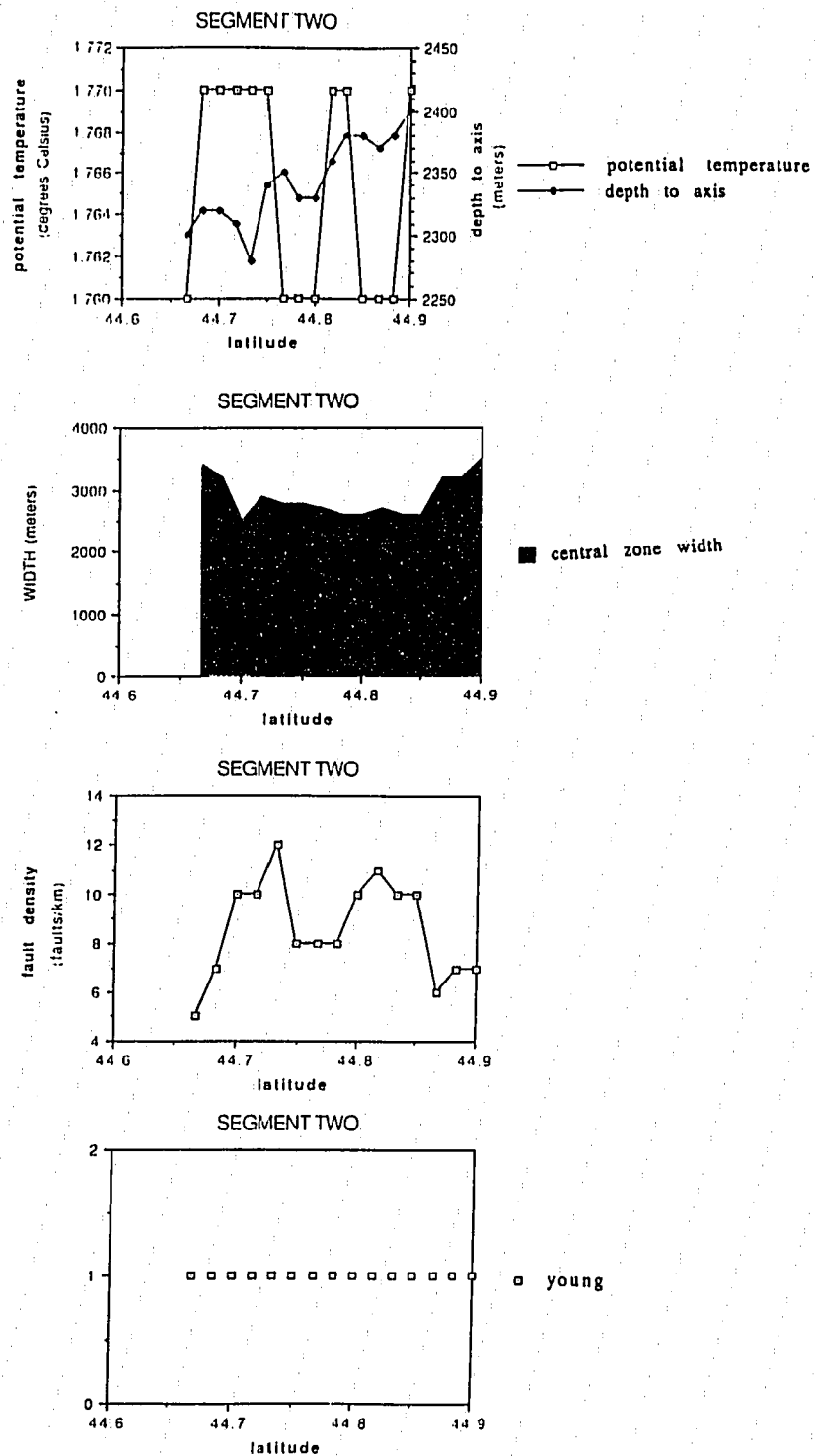


FIGURE jdfr-5
Segment two - Bathymetry and potential temperature from Crane et al. (1985). Zonal width, fault density and age assessment from SeaMARC I images.

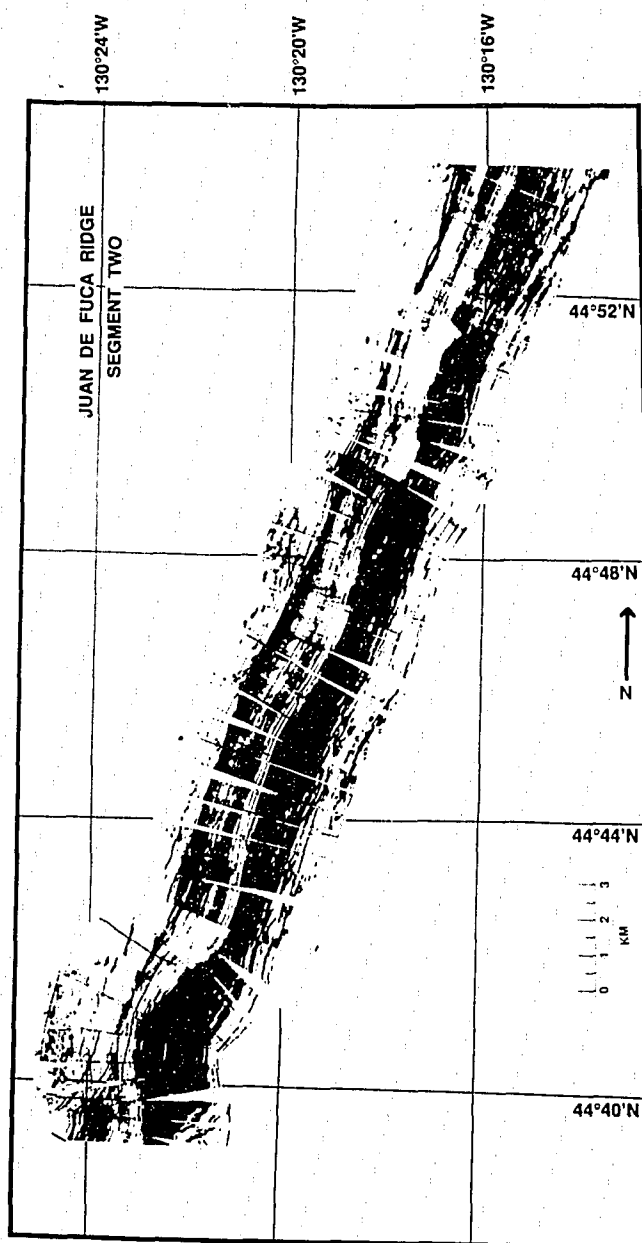


FIGURE jdf-6
Segment two - SeaMARC I image.

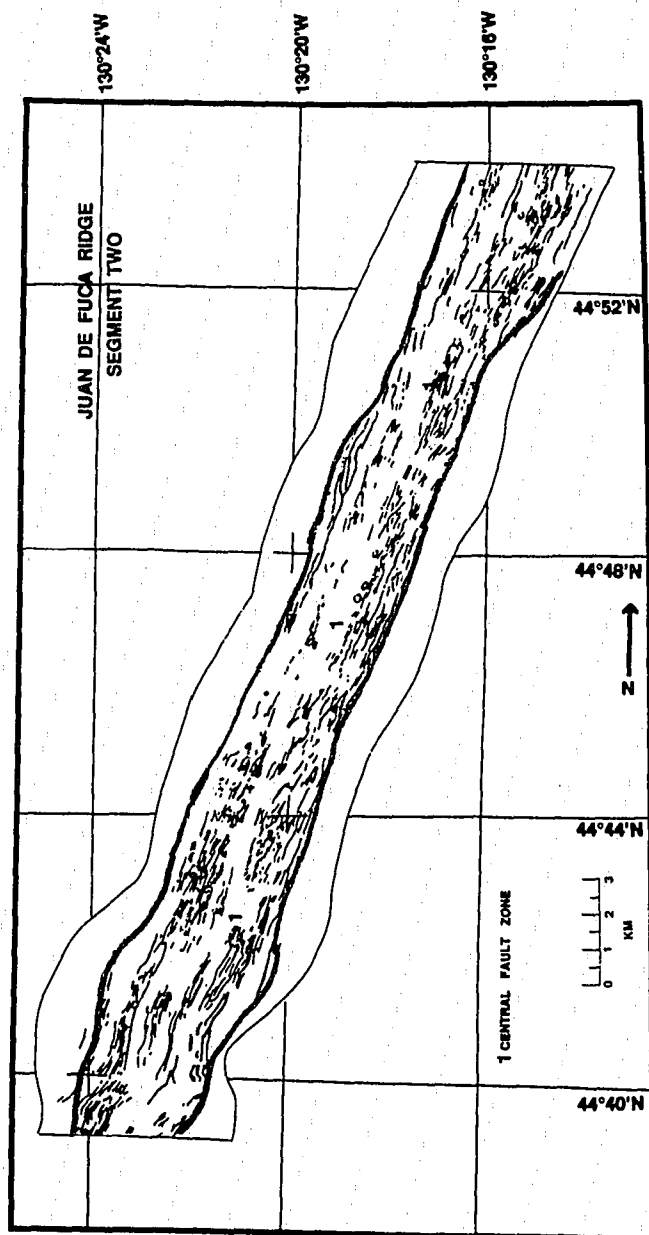


FIGURE jdfrr-7
Segment two - Interpretation and evaluation of SeaMARC I image (FIGURE jdfrr-6)

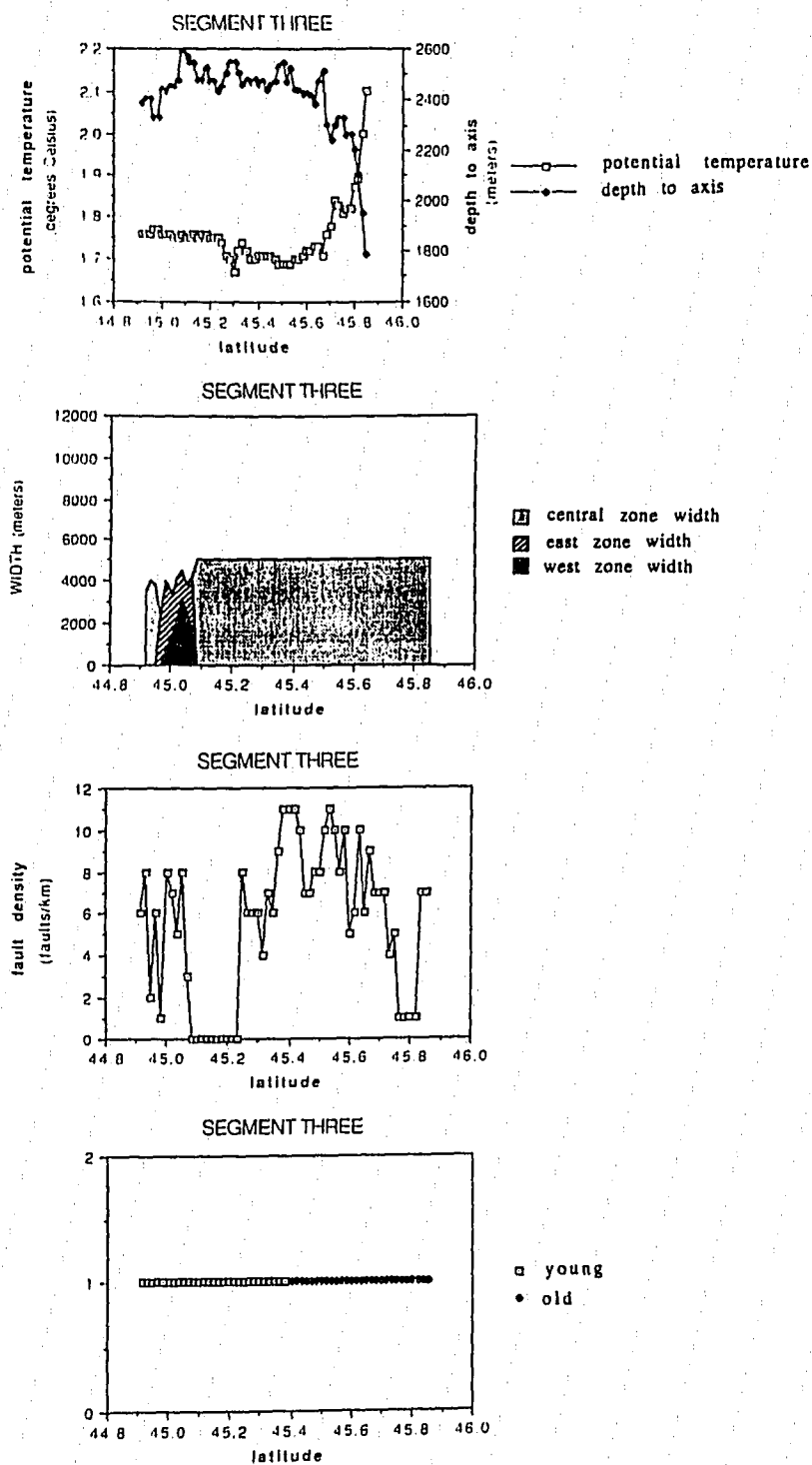


FIGURE jdf8-8
Segment three - Bathymetry and potential temperature from Crane et al. (1985). Zonal width, fault density and age assessment from SeaMARC I images.

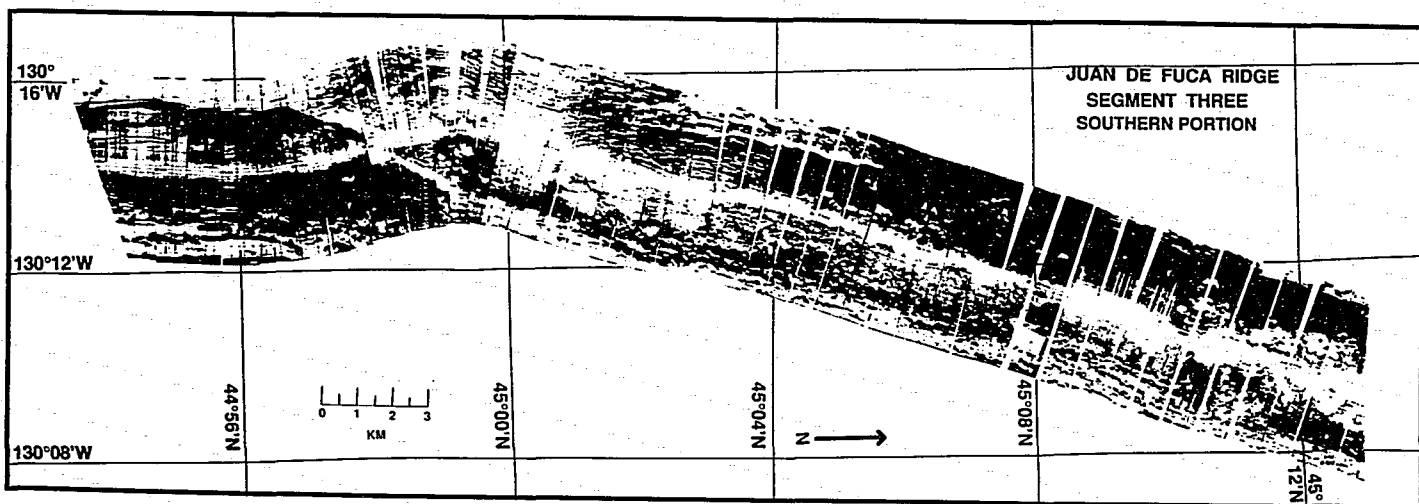


FIGURE jdfr-9
Segment three: southern portion - SeaMARC I image.

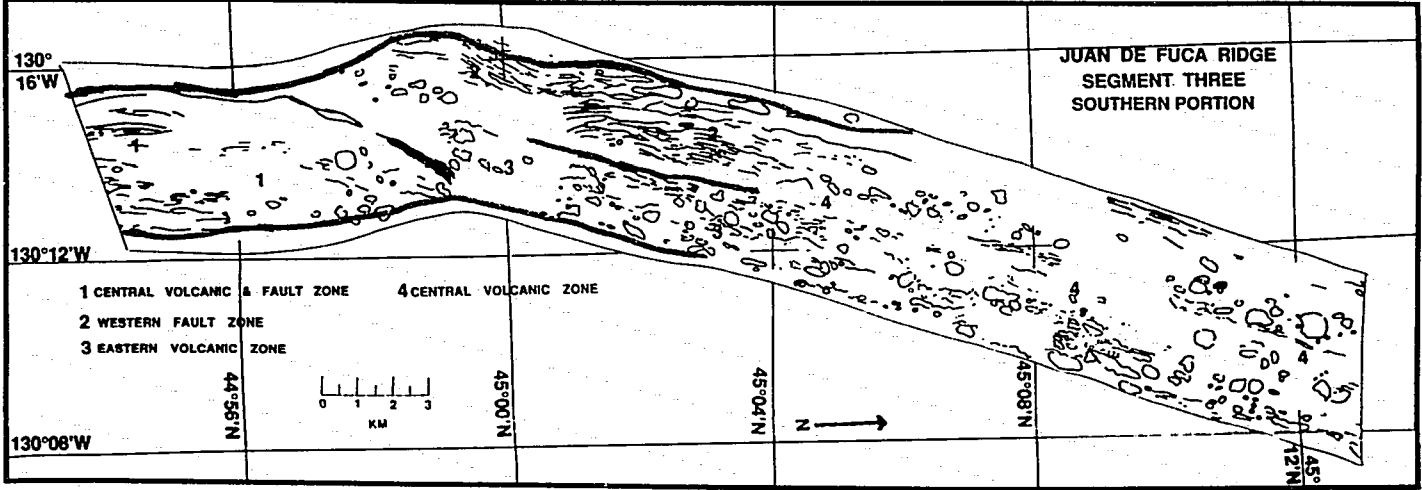


FIGURE jdfr-10
Segment three: southern portion - Interpretation and evaluation of SeaMARC I image (FIGURE jdfr-9)

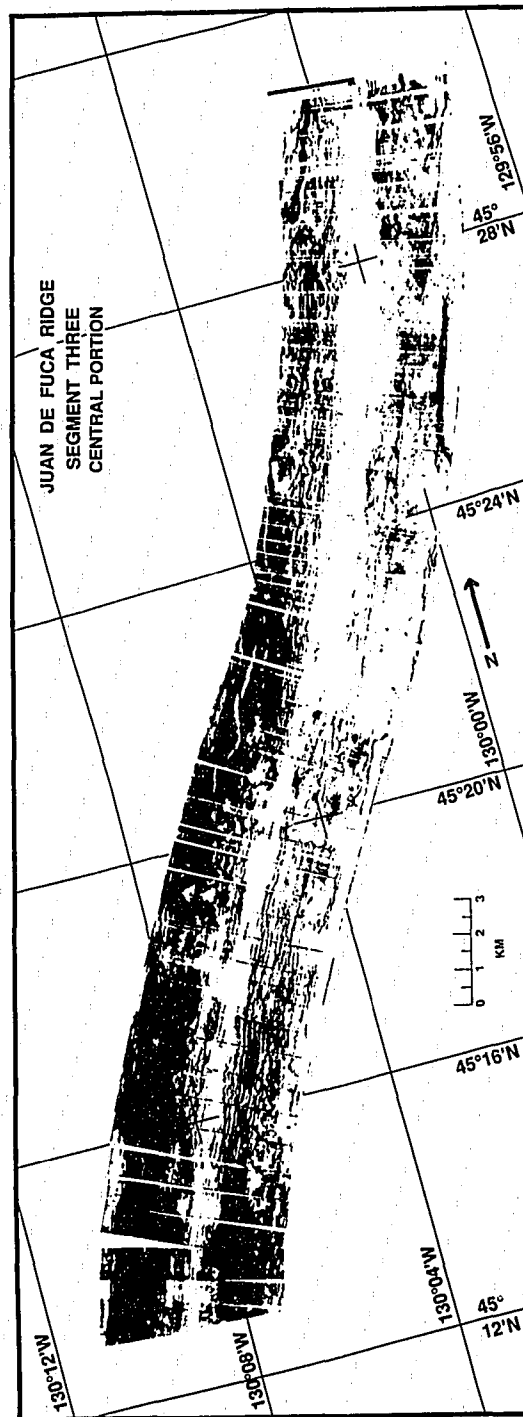


FIGURE jdf-r-11
Segment three: central portion - SeaMARC I image.

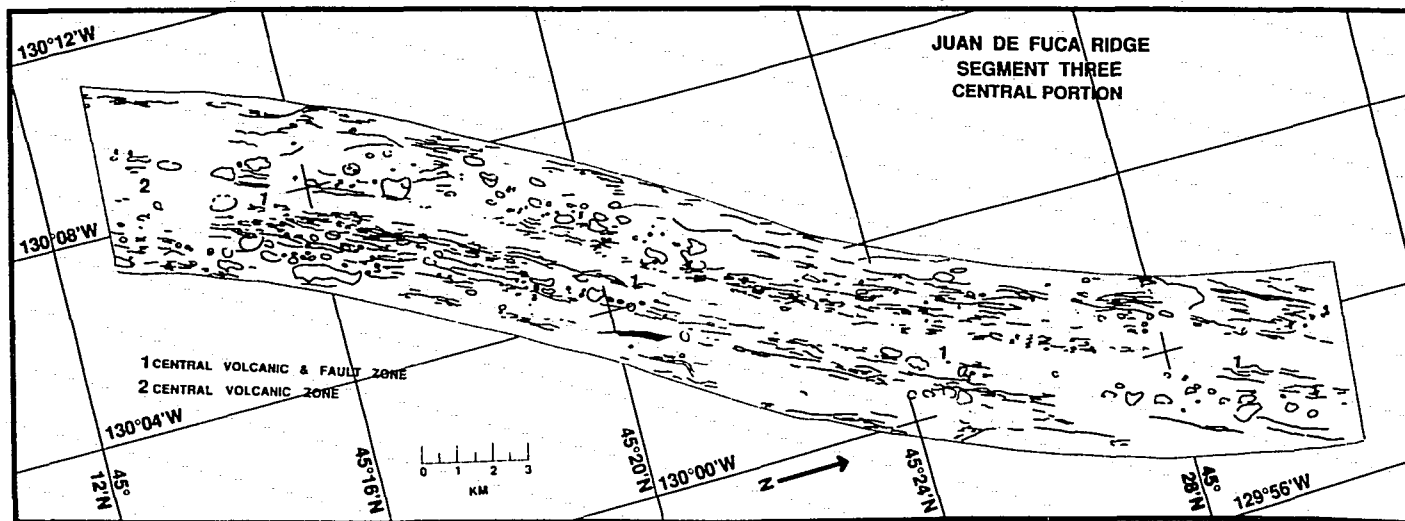


FIGURE jdfr-12
Segment three: central portion - Interpretation and evaluation of SeaMARC I image (FIGURE jdfr-11)

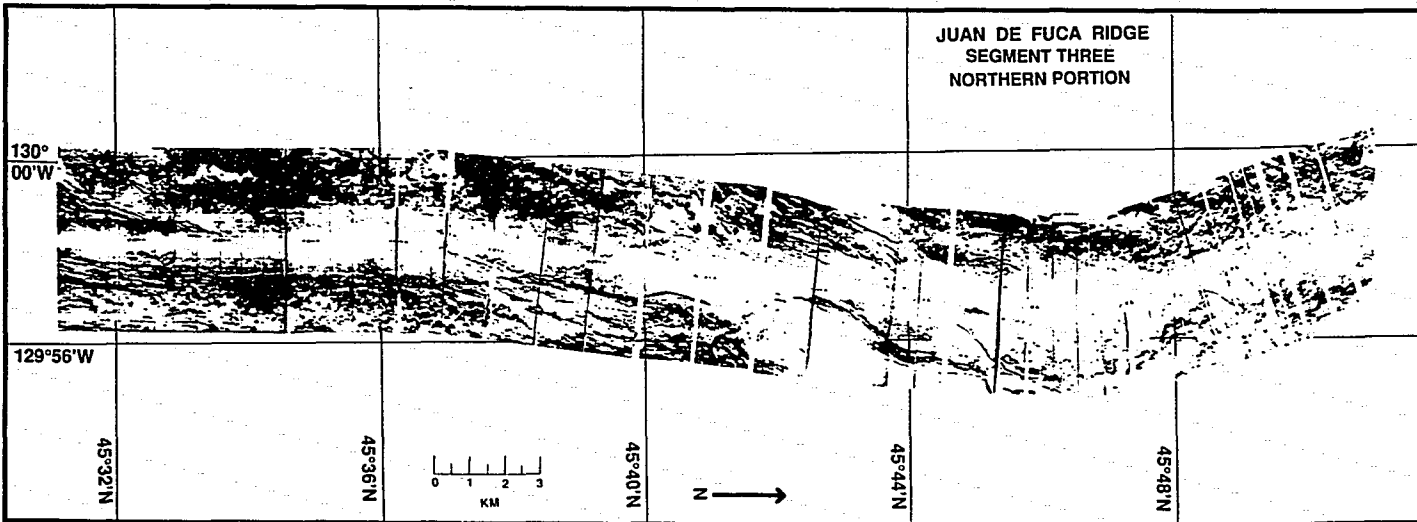


FIGURE jdfr-13
Segment three: northern portion - SeaMARC I image.

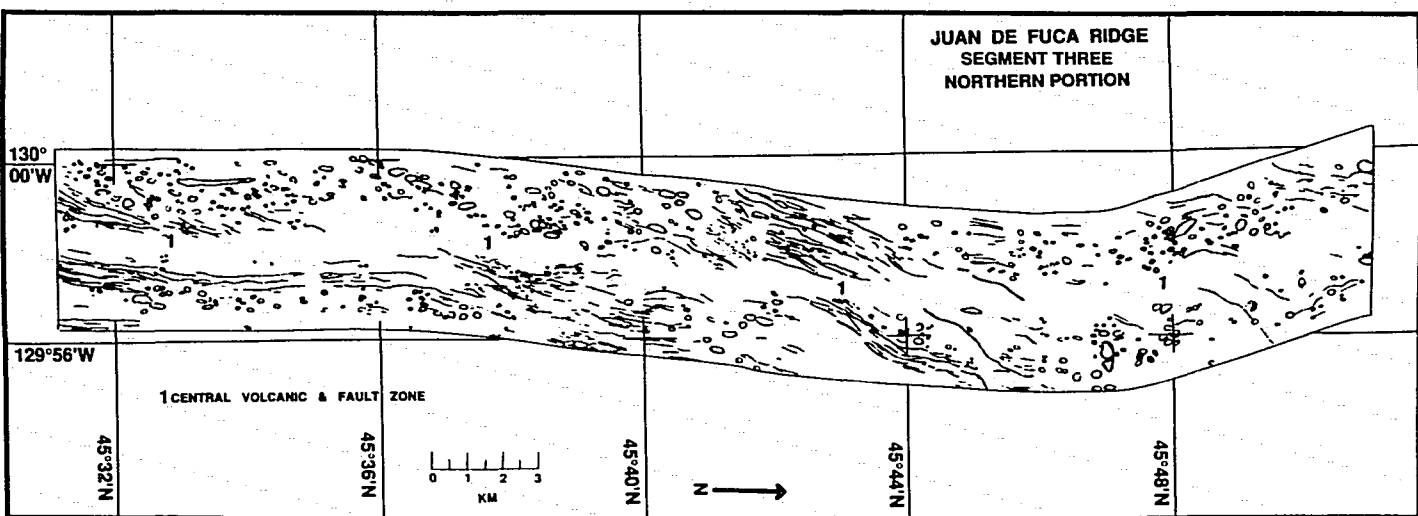


FIGURE jdfr-14
Segment three: northern portion - Interpretation and evaluation of SeaMARC I image (FIGURE jdfr-13)

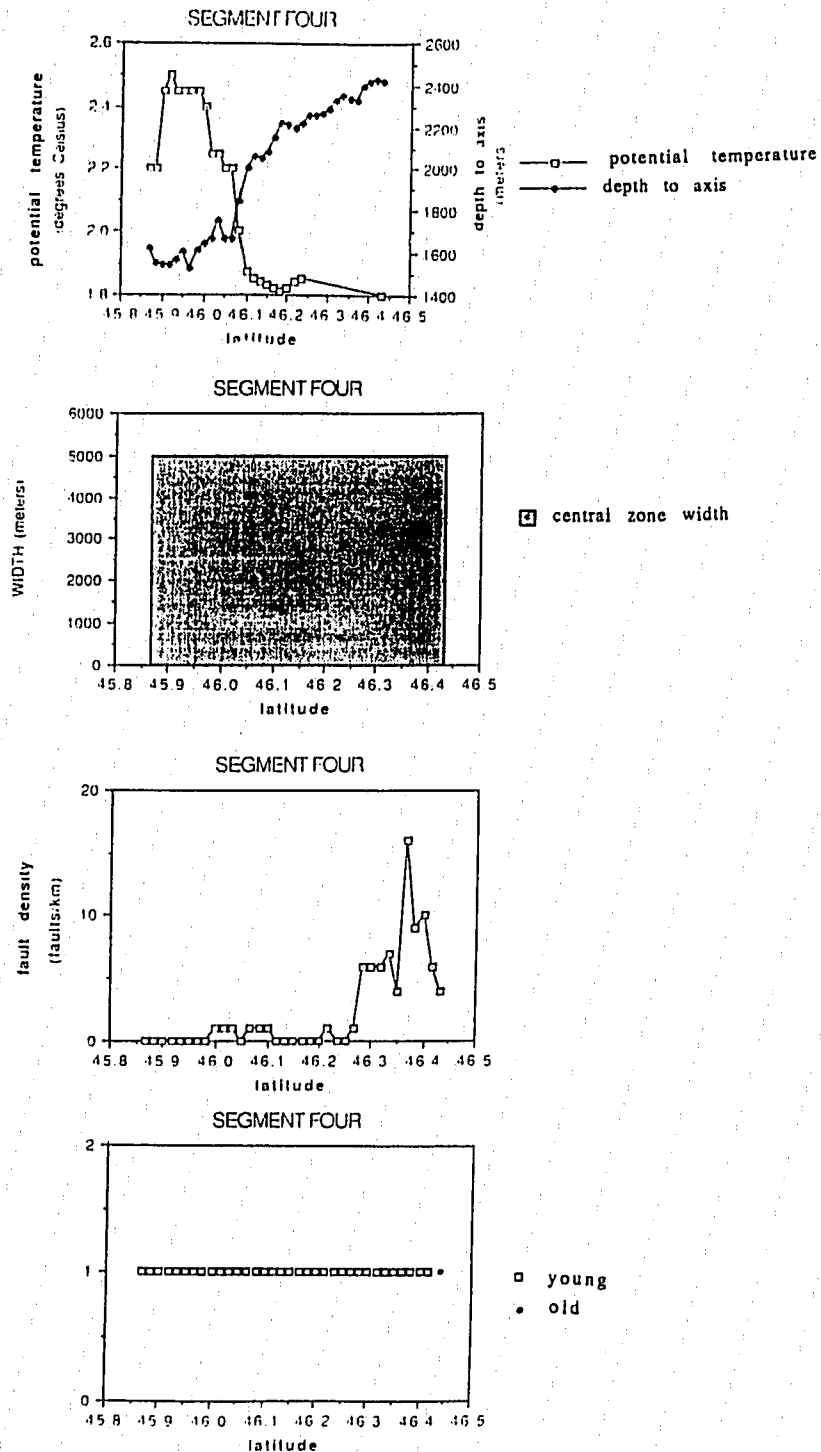


FIGURE jdf-15
Segment four - Bathymetry and potential temperature from Crane et al. (1985). Zonal width, fault density and age assessment from SeaMARC I images.

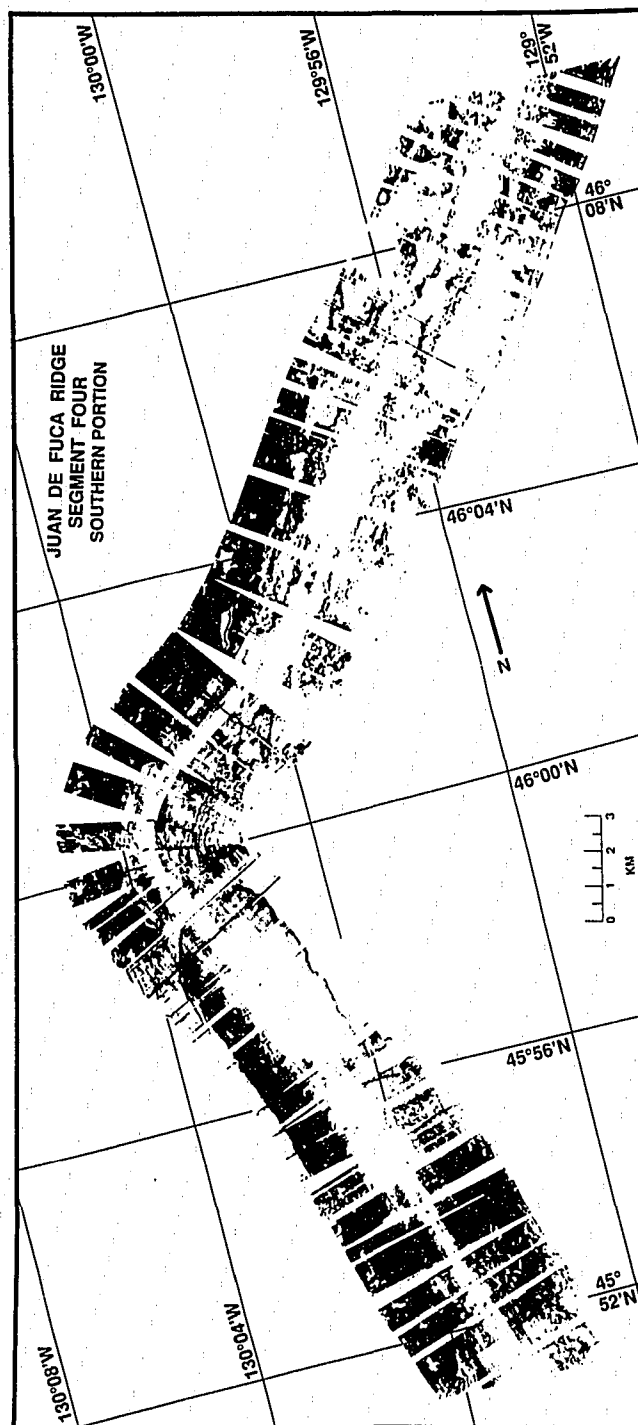


FIGURE jdf-16
Segment four: southern portion - SeaMARC I image.

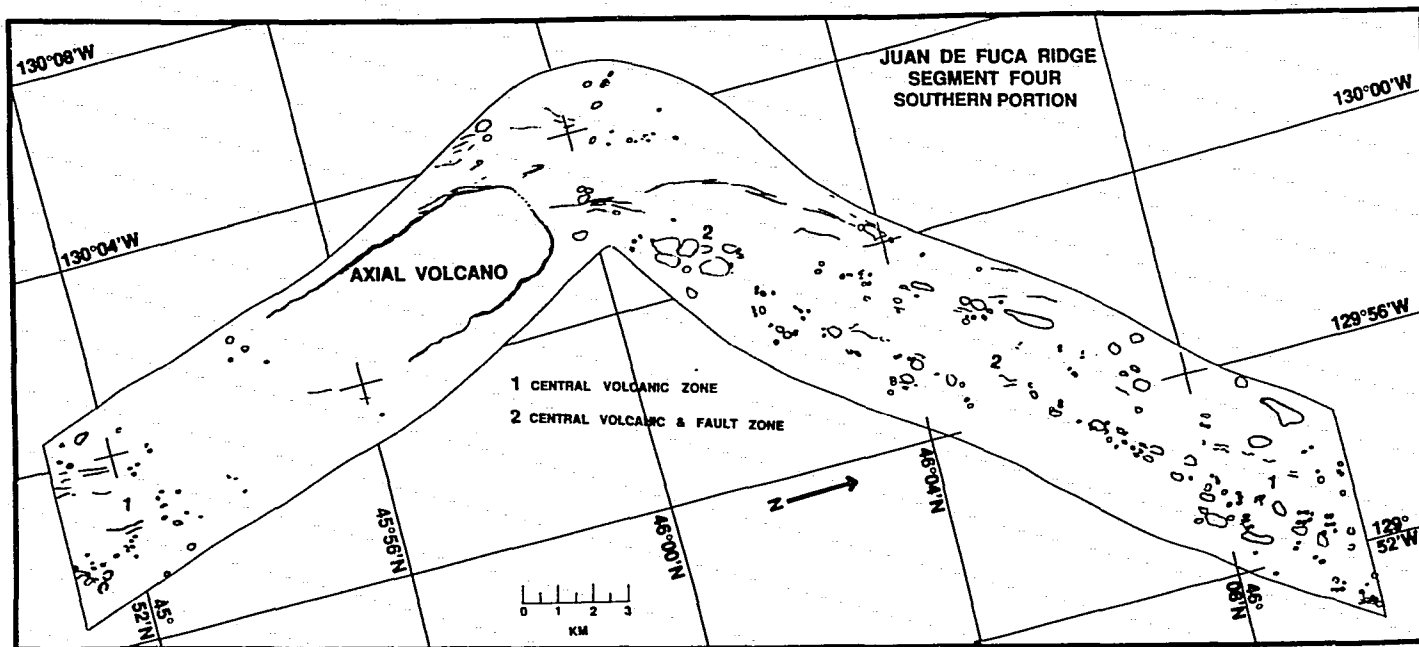


FIGURE jdfr-17
 Segment four: southern portion - Interpretation and evaluation of SeaMARC I image (FIGURE jdfr-16)

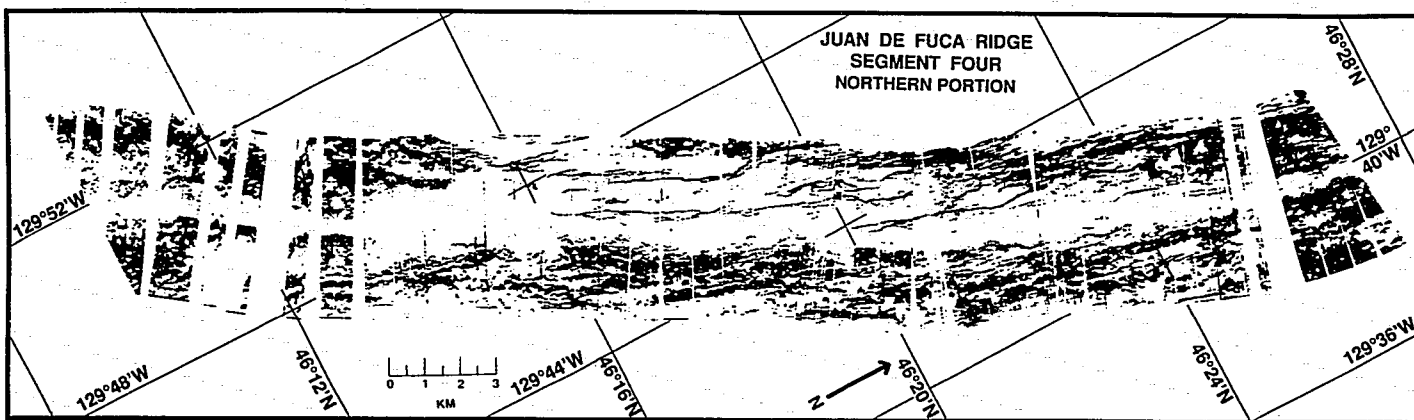


FIGURE jdfr-18
Segment four: northern portion - SeaMARC I image.

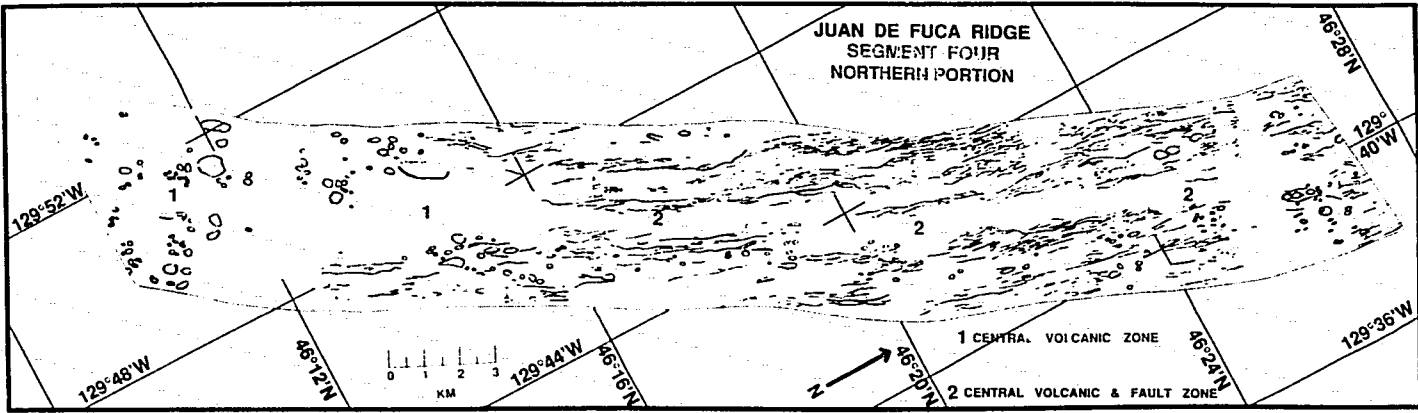


FIGURE jdfr-19
Segment four: northern portion - Interpretation and evaluation of SeaMARC I image (FIGURE jdfr-18)

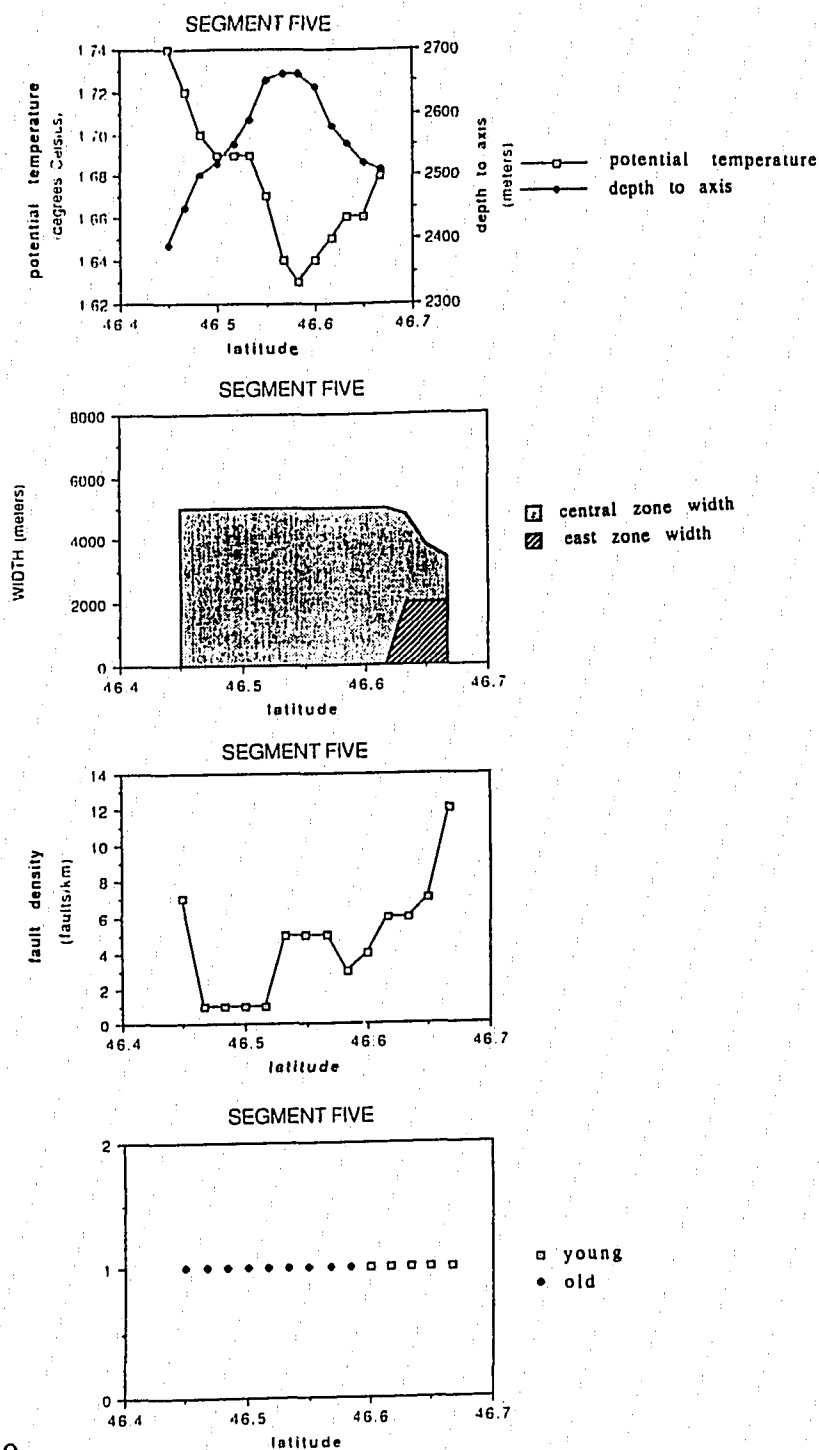


FIGURE jdfr-20
Segment five - Bathymetry and potential temperature from Crane et al. (1985). Zonal width, fault density and age assessment from SeaMARC I images.

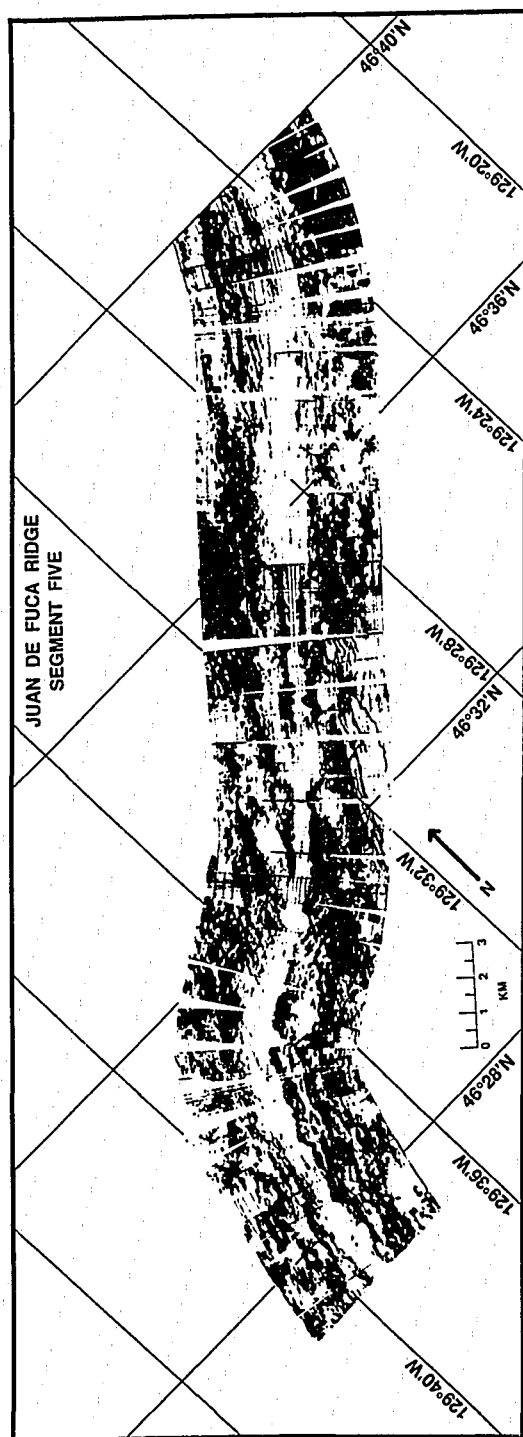


FIGURE jdftr-21
Segment five - SeaMARC I image.

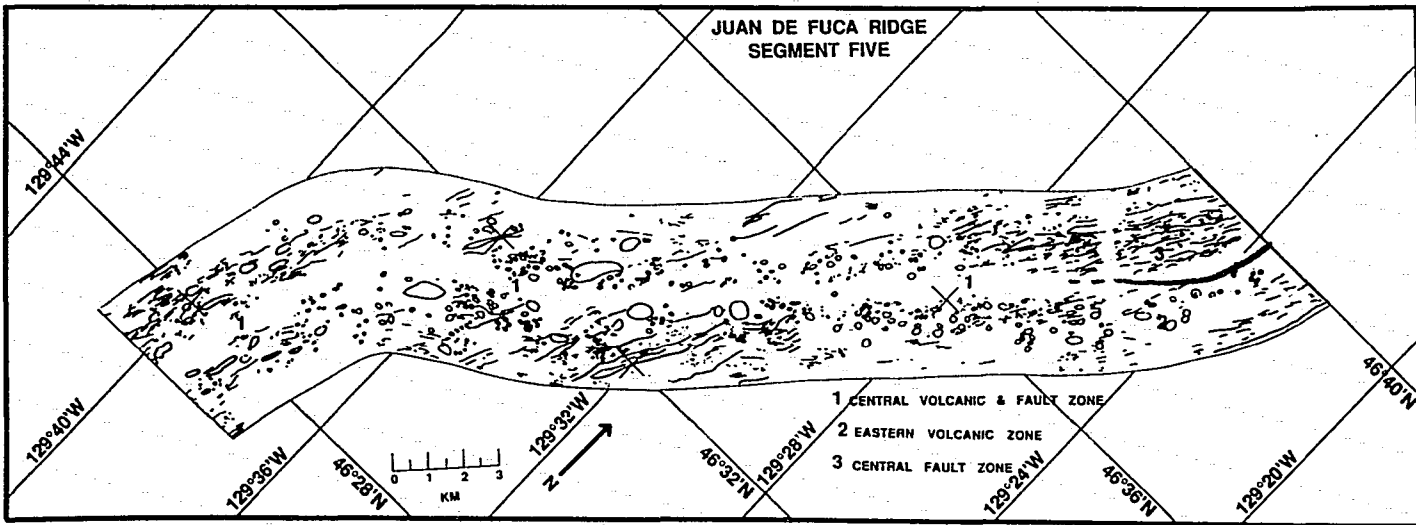


FIGURE jdfr-22
Segment five- Interpretation and evaluation of SeaMARC I image (FIGURE jdfr-21)

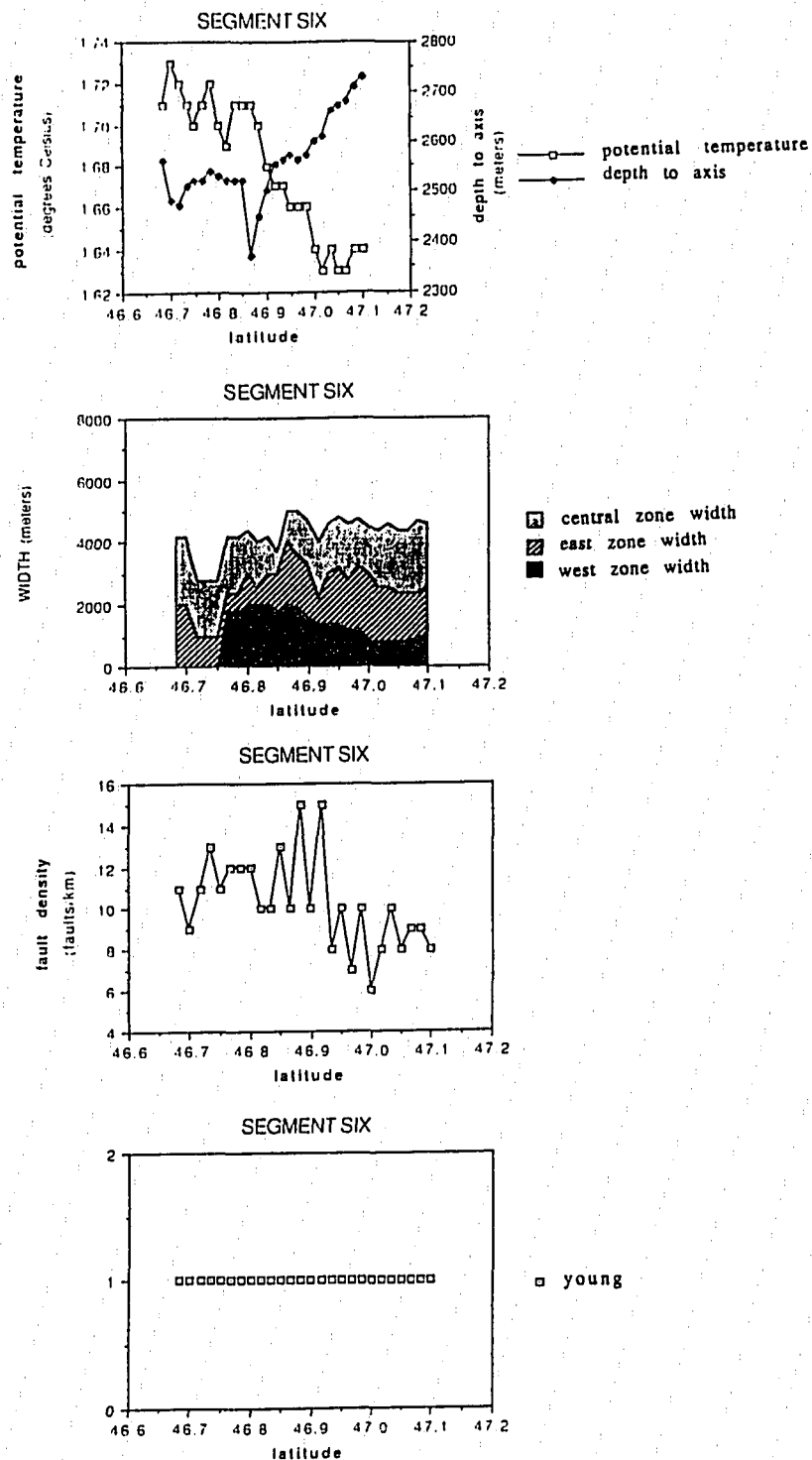


FIGURE jdfr-23
Segment six - Bathymetry and potential temperature from Crane et al. (1985).
Zonal width, fault density and age assessment from SeaMARC I images.

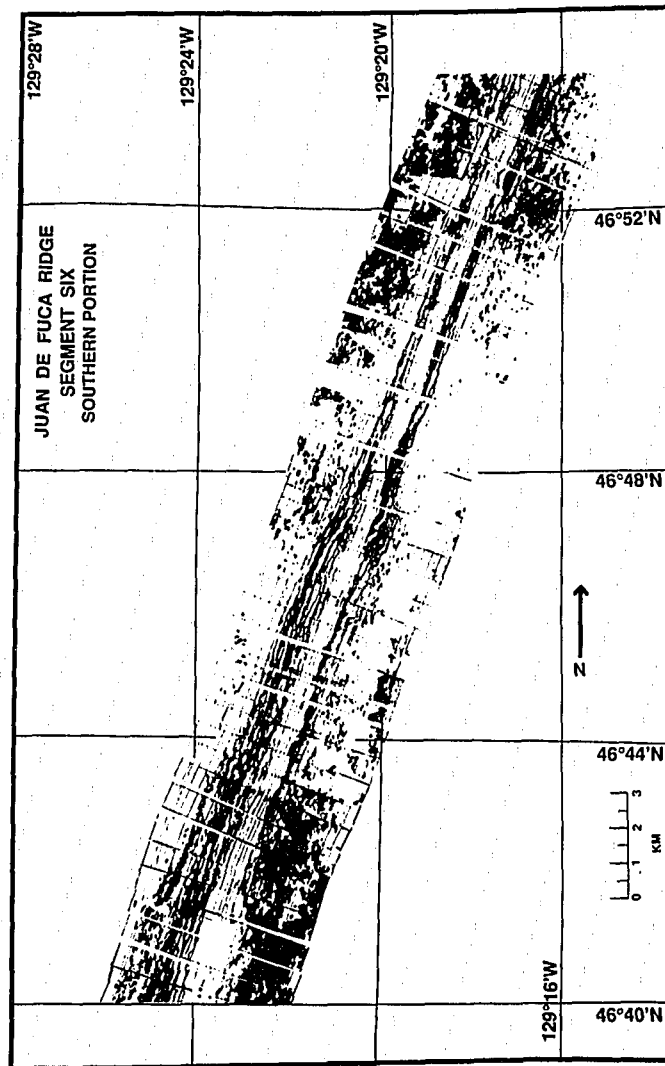


FIGURE jdf-24
Segment six: southern portion - SeaMARC I image.

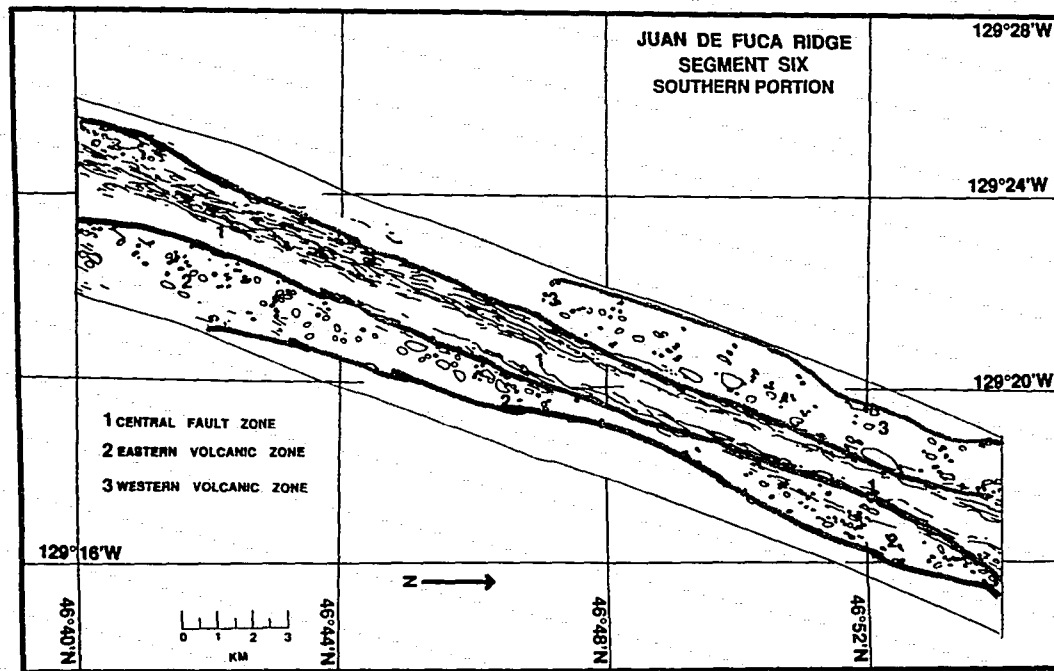


FIGURE jdfr-25
Segment six: southern portion - Interpretation and evaluation of SeaMARC I image (FIGURE jdfr-24)

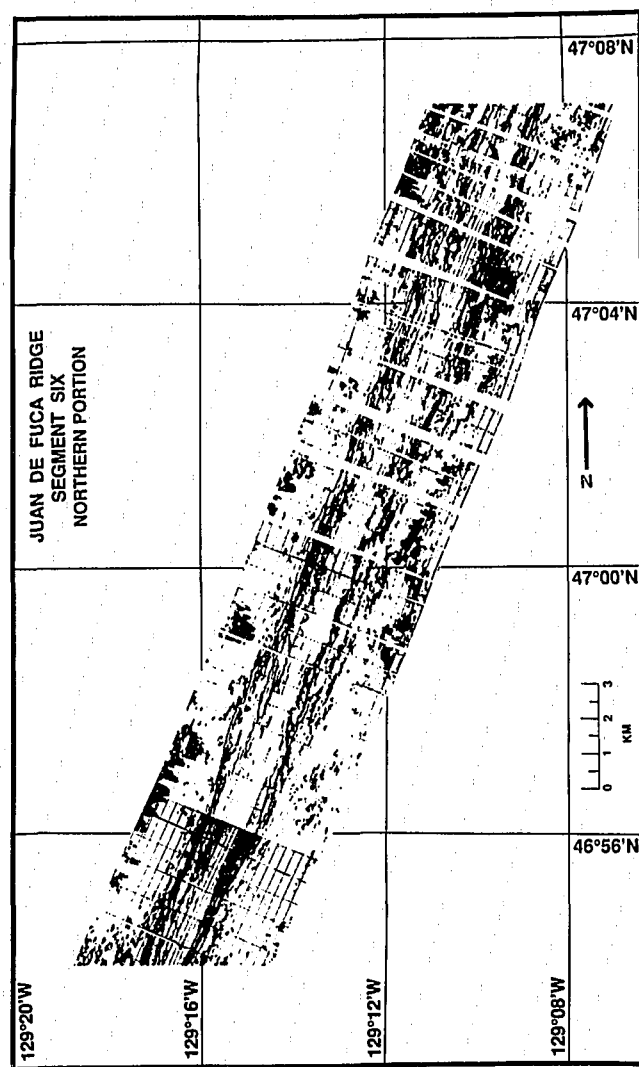


FIGURE jdfir-26
Segment six: northern portion - SeaMARC I image.

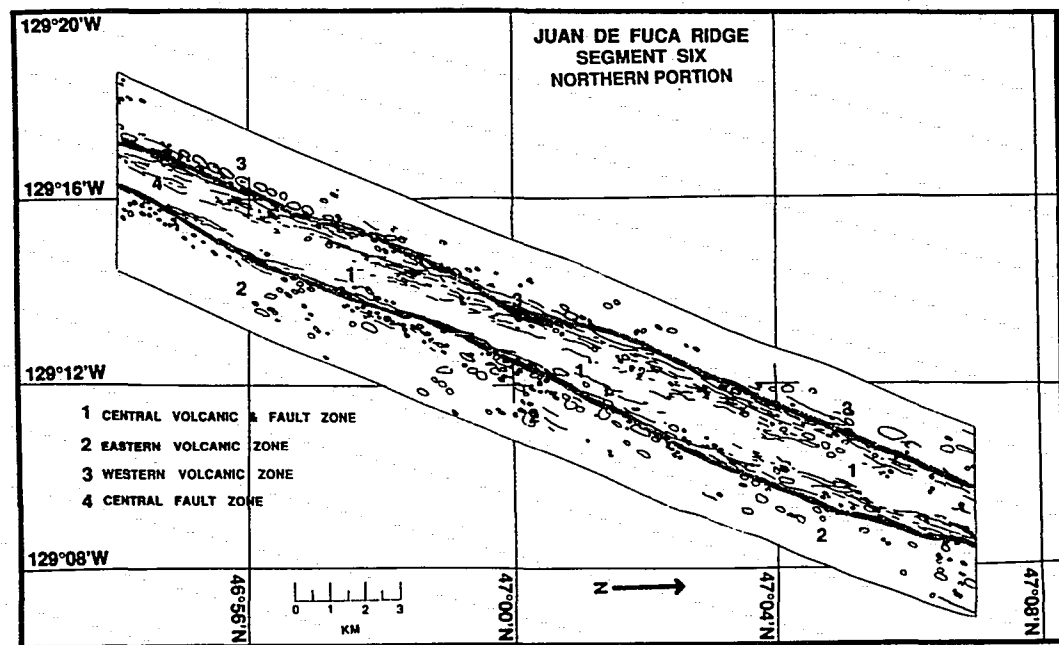


FIGURE jdfr-27
Segment six: northern portion - Interpretation and evaluation of SeaMARC I image (FIGURE jdfr-26)

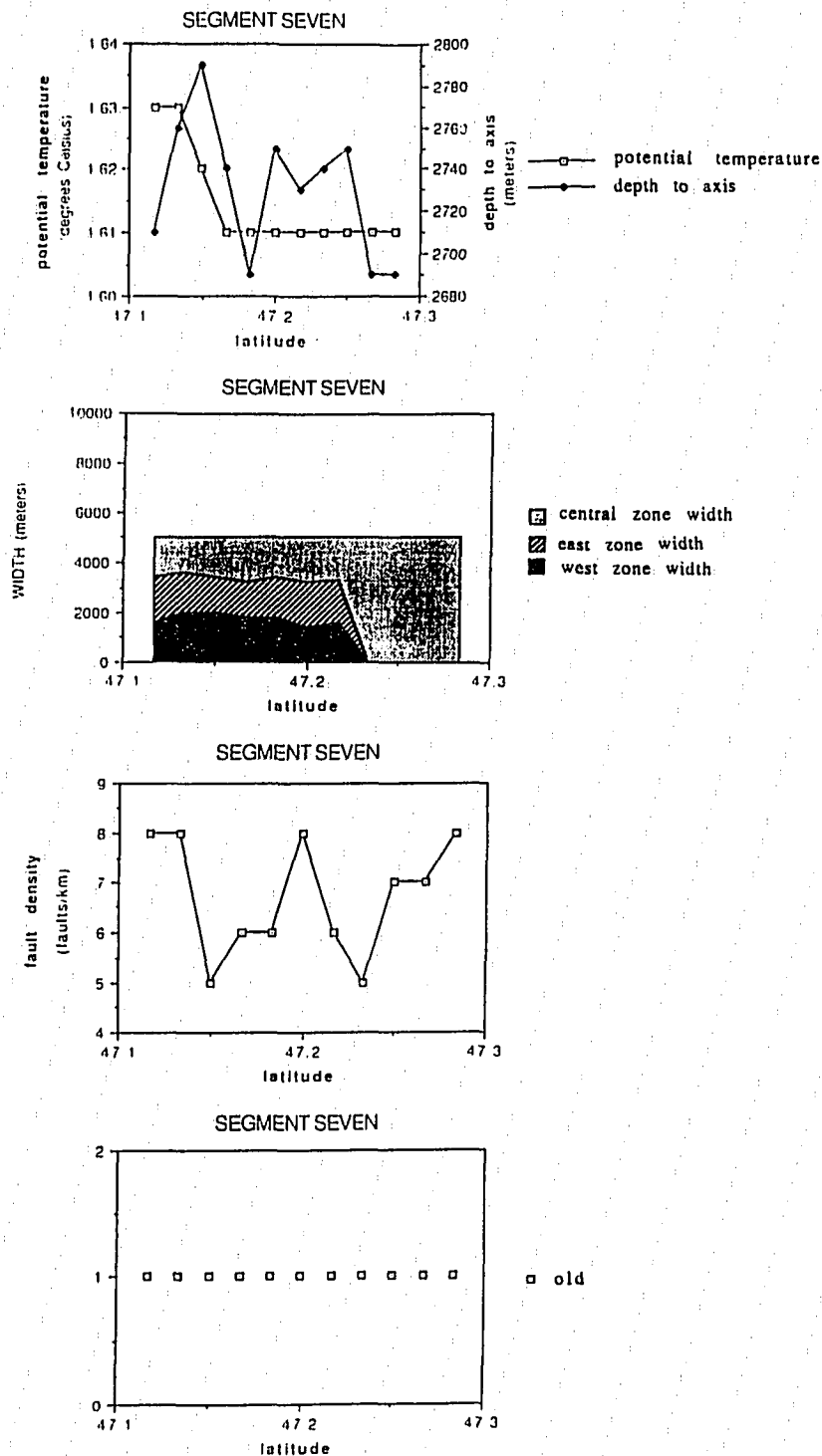


FIGURE jdf-28
 Segment seven - Bathymetry and potential temperature from Crane et al. (1985). Zonal width, fault density and age assessment from SeaMARC I images.

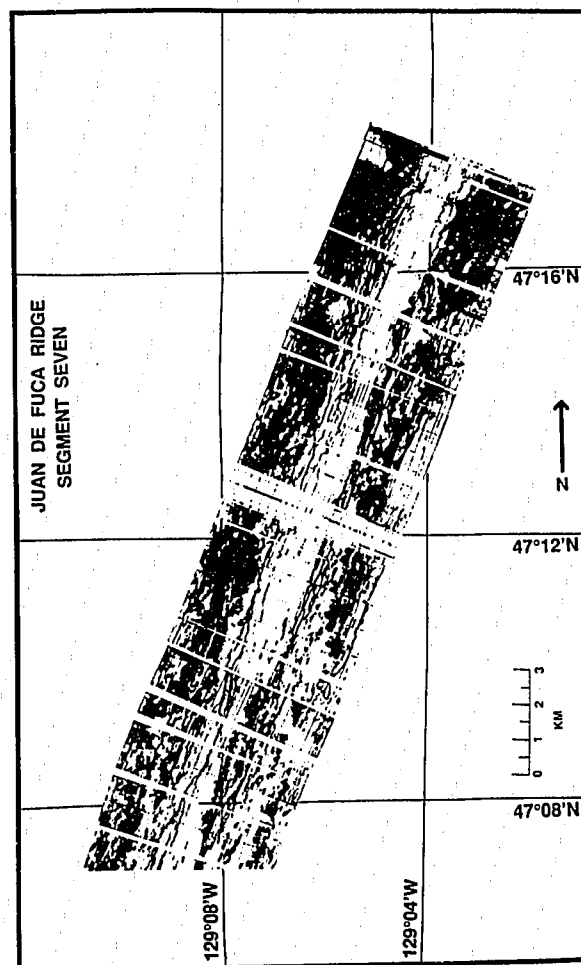


FIGURE jdf-29
Segment seven - SeaMARC I image.

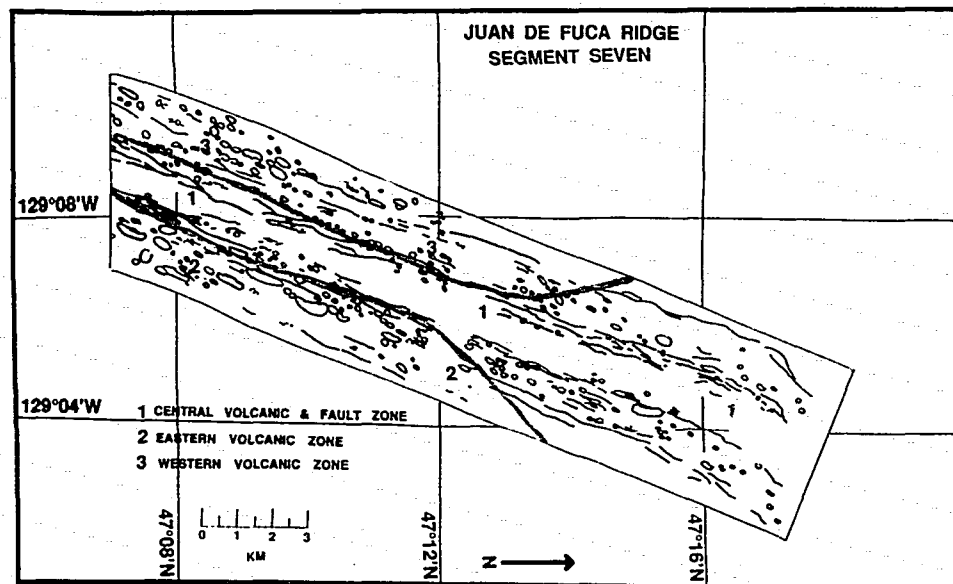


FIGURE jdfr-30
Segment seven - Interpretation and evaluation of Seamarc I image (FIGURE jdfr-29)

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