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A geophysical investigation of the Nova-Canton Trough: The key to the Late Cretaceous evolution of the central Pacific

Joseph, Devorah Dee, Ph.D.

University of Hawaii, 1993

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A GEOPHYSICAL INVESTIGATION OF THE NOVA-CANTON TROUGH: THE KEY TO THE LATE CRETACEOUS EVOLUTION OF THE CENTRAL PACIFIC

A DISSERTATION SUBMITTED TO THE GRADUATE DIVISION OF THE UNIVERSITY OF HAWAII IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE OF

DOCTOR OF PHILOSOPHY

IN

GEOLOGY AND GEOPHYSICS

DECEMBER 1993

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Frontispiece: 3-D perspective image of the SYS09 Nova-Canton Trough bathymetry viewed from an azimuth of N70°E and elevation of 20°. V. E. = 5X. This image was produced from digitized contours gridded using the GMT System, Version 2.1 [Wessel and Smith, 1991].
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ABSTRACT

Free air gravity anomalies derived from satellite altimetry data show that the major Pacific fracture zones, from the Pau to Marquesas, are co-polar about an Euler pole located at 150.6°W, 34.4°S for the period preceding Chron 33 and including a large portion of the Cretaceous Normal Superchron. They also show continuity of the Clipperton Fracture Zone through the Line Islands to the Nova-Canton Trough, and that this Canton-Clipperton trend is co-polar to the same pole. Sidescan-sonar and bathymetry data in the Nova-Canton Trough region reveal N140°E-striking abyssal hill topography south of the Trough's N70°E-striking structures, and crustal fabric striking normal to the Trough (N160°E) to the north. I conclude that the Nova-Canton Trough is the Middle Cretaceous extension of the Clipperton Fracture Zone. I propose that the anomalous depths (7.0 to 8.4 km) of the Trough between 167°30'-168°30'W are the result of a complex plate reorganization. Accretion terminated on the Pacific-Phoenix spreading axis shortly after M-O time and portions of the Phoenix plate were trapped on the Pacific plate when the triple junction jumped south to the nascent Manihiki Plateau. Pacific-Farallon spreading south of the Nova-Canton Trough jumped westwards, initiating transcurrent motion along the western deep of the Nova-Canton Trough which subsequently became the western end of the Clipperton (Pacific-Farallon) Transform. An analysis of free air gravity data collected by various research cruises in the Nova-Canton Trough region indicates that crustal structure proximal to the western deep is symmetric about the ridge-trough region, but that sub-surficial structure adjacent to the eastern Nova-Canton Trough is asymmetric, with 1 to 2 km thinner crust to the south. This evidence, and the polarity of the geoid step suggest younger crust to the south indicating that the Nova-Canton Trough was a right-lateral offset of the Pacific-Farallon spreading axis.
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LIST OF ABBREVIATIONS AND SYMBOLS

\[ g/cm^3 \] grams per cubic centimeter
\[ Hz \] Hertz
\[ in^3 \] cubic inches
\[ kHz \] kilohertz
\[ km \] kilometers
\[ km/s \] kilometers per second
\[ m \] meter(s)
\[ mGal \] milligals
\[ m.y. \] million years
\[ Ma \] mega annum
\[ sec \text{ (or s)} \] seconds
\[ V. E. \] Vertical Exaggeration
\[ \rho \] density
\[ \angle \] angle
\[ \Delta \] triangle
CHAPTER I

New Sidescan-Sonar and Gravity Evidence that the Nova-Canton Trough is a Fracture Zone

Introduction

The Nova-Canton Trough, an anomalous deep of controversial origin, lies in the magnetic quiet zone of the central equatorial Pacific basin between longitude 165° and 169°W, striking ~70° from latitude 1°30'S to the equator (Plate I). Because of its proximity to an inferred Early Cretaceous triple junction between the Pacific, Phoenix, and Farallon plates, determining its character and origin may be the key factor in reconstructing the Late Mesozoic evolution of the central Pacific basin.

The Nova-Canton Trough exhibits asymmetric and variable ridge-trough morphology, with depths exceeding 8 km in some regions, the greatest depths in any ocean not associated with subduction zones. Bordering ridges rise to a minimum of 2.2 km, producing occasional relief of 5.8 km. Major structures appear to terminate east of the Phoenix Islands, but 73° to 80°-striking bathymetric depressions stretch from the Phoenix Islands to 180°W. A 75°-striking ridge (designated the Nova-Canton ridge) spans the region between the Nova-Canton Trough and the Line Islands sediment apron. Scant crustal age information is available for the Nova-Canton Trough region because Deep Sea Drilling Project sites do not lie close enough to the trough to give a reliable age estimate (Plate I). Low amplitude magnetic anomalies suggest that the crust here was formed during the Cretaceous Normal Superchron, but relatively fresh volcanic cinder and glass have been recovered from its scarps.
Plate I: Bathymetric map of the central Pacific basin showing the location of the Nova-Canton Trough and other major seafloor structures. Diamonds indicate DSDP Sites 65 and 66 [Winterer, Riedel et al., 1971], Sites 164, 165, 166, and 167 [Winterer, Ewing et al., 1973], and Sites 315, 316, and 317 [Scientific Staff, 1974]. Thick gray lines indicate the Phoenix magnetic-anomaly lineations [Nakanishi et al., 1992]. CPFZ: Central Pacific fracture zone; PFZ: Phoenix fracture zone. The thin red line indicates the route of the Seafloor Surveys International/OTC Limited PacRim cable route cruise from Honolulu to American Samoa, including the Nova-Canton Trough survey. The red box indicates the location of the region shown in Plate III. The color map was produced from ETOPO-5 data [National Geophysical Data Center, 1988].
Several authors have suggested a spreading center origin for the Nova-Canton Trough, which is subparallel to the adjacent Phoenix magnetic-anomaly lineations (Plate I). A recent survey in the eastern Nova-Canton Trough region, however, revealed pronounced abyssal hill fabric trending into, not parallel to, trough structures. This evidence precludes crustal accretion at the eastern Nova-Canton Trough and favors earlier interpretations that it is the western extension of the Clipperton Fracture Zone. Both the continuity of the Nova-Canton Trough and Clipperton Fracture Zone free air gravity anomalies and the co-polarity of the Canton-Clipperton trend with other coeval Pacific-Farallon fracture zones further support the fracture zone interpretation.

Previous Work

Menard [1967] first reported the ridge-trough topography present in the generally flat equatorial central Pacific basin and postulated that these structures are the western extension of the Clipperton Fracture Zone. He further noted that several of the major Pacific fracture zones, from the Mendocino Fracture Zone in the north Pacific to the Marquesas Fracture Zone near 10°S, extend into the central Pacific, beyond the mid-plate swells created by the Hawaiian, Line, and Marquesas seamount chains, but that the fracture zones branch west of these island groups (Plate I, Figure 1.1). Rea [1970] suggested that the fracture zones, rather than branching, change trend by 10° to 15° throughout a 125 to 150-km wide strip extending as a great circle from ~156°W, 21°N to ~174°W, 43.5°N, which he named the "bending line." Noting the concave southward expression of the Pioneer, Murray, Molokai, and Clarion fracture zones (Figure 1.1), Sager and Pringle [1987] proposed an Euler pole located at 155°W, 11°S to describe motion on the Pacific-Farallon paleo-spreading axis responsible for these fracture zone traces.
Figure 1.1: Mercator projection of the slope of the free-air gravity field. The major features in the Pacific Ocean are located for ease in interpreting Figure 1.8. Note that at the change in trend of the Murray, Clarion, and Clipperton Fracture Zones, there are small "Y's" in the fracture zone trace. The white arrow indicates the location of the N80°E-striking lineated bathymetric depressions which are discussed in the text. Compare the strike of these features in this Mercator projection to their strike when projected relative to our estimated pole of opening for the Middle Cretaceous. This image was produced from GEOSAT free-air gravity data.
A series of dredge hauls in the 1960's sampled the scarps of the Mendocino, Molokai, Clarion, and Clipperton fracture zones, and a deep trough, named the Argo Trench by Engel and Engel [1970] located by them at 178°W, 3°S. The authors describe this trench as plunging to depths of 8400 m with slopes averaging more than 32°. This description fits the Nova-Canton Trough, and as Rosendahl [1972] points out, there may be some confusion as to the location of this feature, because no such extreme relief is encountered at the location described by Engel and Engel. Dredges of the scarps returned relatively fresh tholeiitic and alkali basalts, as well as gabbros altered in part to greenschist facies. One of the rocks yielded an age of ~70 Ma, but the authors warned that this age estimate could be equivocal owing to alteration of the minerals contained.

In 1970, the R/V Mahi surveyed this region, naming the linear trough feature the Canton Trough [Halunen et al., 1970; Rosendahl and Halunen, 1971]. Rosendahl [1972] and Rosendahl et al. [1975] explained the extreme relief and elongate magnetic anomaly highs as indicating the presence of ENE-striking linear intrusive bodies in an extensional environment. He concluded that slow spreading within the Nova-Canton Transform Fault during a slight reorientation of the Pacific-Farallon spreading axis, or volcanism and faulting during a later period of rejuvenation, created the severe relief. He argued that the Nova-Canton Trough mainly exhibits structures suggestive of fracture zone morphology, but the "occurrence of relatively fresh intrusive rocks and some instances of faulted sediments" [Rosendahl, 1972] suggest that recent rejuvenation (volcanism and faulting) may have occurred.

Northwest of the Nova-Canton Trough, Larson et al. [1972] identified the Phoenix magnetic lineations (Plate I), which record Early Cretaceous spreading between the Pacific plate to the north, and Phoenix plate to the south. Larson and Chase’s [1972] tectonic reconstruction of the central Pacific basin interpreted the Nova-Canton Trough as a Pacific-Farallon transform fault at ~100 Ma, north of a Pacific-Phoenix-Farallon triple junction.
The identification of the Phoenix lineations by Larson et al. [1972], led Winterer et al. [1974] to hypothesize that the Nova-Canton Trough originated as "a deep slot associated with a spreading direction change" and, later Winterer [1976a, 1976b] postulated a spreading center origin for the Nova-Canton Trough. In Winterer's reconstruction of Early Cretaceous Pacific basin tectonics, the Nova-Canton Trough forms the western axis of a ridge-ridge-ridge triple junction between the Pacific, Phoenix, and Farallon plates at about anomaly M0 time. Winterer et al. [1974] considered the Trough to be an extension of ~N80°E-striking linear bathymetric depressions lying between 180°W and the Phoenix Islands (Plate I, Figure 1.1, arrow). They proposed that an abrupt change in spreading geometries preserved the extreme relief found in the Trough when the triple junction jumped south to the vicinity of the Manihiki Plateau. A spreading center origin of the Nova-Canton Trough overcame several objections to Menard's [1967] fracture zone hypothesis, including 1) the lack of a regional depth change across the Nova-Canton Trough [Rosendahl et al., 1975]; 2) the absence of ridge-trough morphology east of 165°W; and 3) the conjectured relationship of the linear bathymetric depressions in the region west of the Phoenix Islands to the Nova-Canton Trough, implying a crustal swath with an eastern Pacific spreading pattern which the Phoenix magnetic anomaly trends clearly prohibit [Rosendahl et al., 1975].

Mammerickx and Sandwell [1986] later characterized the Nova-Canton Trough as a rift abandoned by a long-distance spreading center jump, noting that the new spreading site is not evident in the surrounding bathymetry, but may be buried under the Line Islands volcanic edifice.

Magnetic anomaly mapping by Nakanishi et al. [1992] of Early Cretaceous lineations in the western Pacific identified conjugate anomalies M1 through M3 in the Phoenix lineation set between the Central Pacific and Phoenix fracture zones (Plate I). These conjugate isochrons imply a cessation of spreading on the western portion of the Pacific-
Phoenix spreading center just after M1 time. No conjugate anomaly lineations have been identified in the region between the eastern Phoenix lineations and the Nova-Canton Trough [Cande et al., 1989; Nakanishi et al., 1992; Yamazaki and Tanahashi, 1992]. In addition, Nakanishi et al. [1992] have identified NNW-trending M-series anomalies in the region between 170°W and the Line Islands (Plate I). The presence of this magnetic bight confirms the existence of a triple junction between the Pacific, Phoenix, and Farallon plates prior to M0 time.

Based on gravity and magnetics investigations, Yamazaki and Tanahashi [1992] support a spreading-center origin of the Nova-Canton Trough. They argue that although trough-parallel lineated magnetic anomalies can be explained by topographic relief, these anomalies possess the same skewness parameter as the M-series Phoenix lineations, thus having experienced the same rotational movement since accretion. Their gravity models suggest slight crustal thickening in the trough axis, which, given results of very thin crust observed at Atlantic fracture zone troughs [e.g., Detrick and Purdy, 1980; Ludwig and Rabinowitz, 1980; Sinha and Louden, 1983; Cormier et al., 1984; White et al., 1984; Louden et al., 1986; Whitmarsh and Calvert, 1986; Calvert and Whitmarsh, 1986; Potts et al., 1986a, 1986b] appear to contradict a fracture zone origin of the western Nova-Canton Trough.

It is necessary at this point to address the confusion regarding the name and precise location of the feature referred to as the Nova-Canton Trough, to differentiate it from other structures in the general region that will be examined in this study. Engel and Engel [1970] first applied the name "Argo Trench" to a trough, described as plunging to depths of 8400 m with slopes averaging 32°, located at 3°S, 178°W. However, current seafloor charts do not show any such extreme relief at the location described by these authors. The designation "Canton Trough" first appears in an abstract submitted to EOS by Halunen et al. [1970], applied to "a deep trough in excess of 8000 m located northeast of Canton.
Island at approximately 1°S, 168°W (striking) N68°E and (extending) for more than 200 km.” This designation (also “Nova Trough”) appears in the Gazetteer of Undersea Features [Defense Mapping Agency, 1981]. However, according to this publication, the feature is located at 2°30'S, 173°W, a site in the middle of the Phoenix Island group. The Scripps maps of the Pacific [Mammerickx et al., 1971; Mammerickx and Smith, 1982] attach the name "Nova-Canton Trough" to a series of bathymetric lineaments, approximately 5800 m deep, in the region between the Phoenix Islands and 180°, which span over 5° north to south (Plate I, Figure 1.1, arrow). These charts designate Halunen et al.'s "Canton Trough," the "Nova Trough." However, the 1985 edition of Bathymetry of the Northcentral Pacific [Mammerickx and Smith, 1985] calls the "Canton Trough" as defined by Halunen et al. [1970], the "Nova-Canton Trough," in addition to the region between the Phoenix Islands and 180°. I prefer to retain the designation "Nova-Canton Trough" when referring to the deep trough system extending east from approximately 169°W, because this has become common usage. However, I do not consider the lineated depressions lying in the region between the Phoenix Islands and 180°W (Plate I) to have been formed by the same mechanism responsible for the extreme relief found in the region east of 169°W, and therefore will refer to these bathymetric features by their location. These lineaments are designated on Figures 1.1 and 1.8 by a white arrow.

**SYS09 Sidescan-Sonar Survey of the Eastern Nova-Canton Trough**

The Seafloor Surveys International/OTC Limited PacRim cable route cruise in January 1990 mapped 6700 km² in the eastern region of the Nova-Canton Trough between latitude 0° and 1°15'S, longitude 166°22' to 166°54'W. Data collected during the survey include five parallel, N-S trending lines of SYS09 sidescan-sonar images (swath width ~14 km) and swath bathymetry (swath width ~3.4 times towfish altitude) [Blackinton et al., in3
Figure 1.2: SYS09 sidescan-sonar mosaic of the Seafloor Surveys International/OTC Limited PacRim Nova-Canton Trough survey (swath width ~14 km). The SYS09 sidescan data were reprocessed at the Hawaii Institute of Geophysics using SeaMARC II sidescan enhancement programs to refine the fabric information contained in the reflectivity data [Sender et al., 1989]. Gains are first normalized to a common level, then a global contrast map is generated to ensure that gray scales are uniform throughout the survey. In this image, light gray tones indicate low reflectivity, whereas very dark gray tones indicate high reflectivity. The arrow indicates the location of the N140°E-striking abyssal hill fabric.
1991] (Figure 1.2, Plate II), single-channel seismic reflection profiles (60 to 200 Hz, 40 in$^3$ airgun, Figures 1.3, 1.4), 3.5 kHz echo-sounder bathymetry, and gravity data. This survey was part of the larger PacRim survey conducted by SSI for OTC Ltd. and AT&T to map the seafloor for a proposed fiber optic cable route between Honolulu, Hawaii and Auckland, New Zealand. The survey area can be divided into four structural sub-provinces. From north to south these are: (1) the northern plain, (2) a series of three sub-parallel ridges (central ridge) enclosing the 6.1 km deep eastern trough, (3) the central trough, and (4) the southern plain.

1. Northern plain

The northern region is a flat, sedimented plain standing at \(-5.3\) km below sea level. A chain of \(-400\) to 2400 m diameter volcanoes trending \(072^\circ\) lie in the extreme northern part of the survey area, paralleling the strike of the ridges (Figure 1.2, Plate II). Seismic reflection profiles (Figures 1.3, 1.4) show a 0.05 to 0.08 s thick acoustically transparent unit overlying an \(-0.4\) to 0.6 s thick sequence of turbidites [Kroenke, 1970] resting on acoustic basement. Pelagic sediments drape the volcanoes without internal disruption of the layering, suggesting deposition after formation of the seamounts.

2. Central ridge and eastern trough

The bathymetric map (Plate II) shows a trough more than 6 km deep on the eastern edge of the survey area, similar to the deep trough surveyed by the R/V Mahi farther to the west [Rosendahl, 1972; Rosendahl et al., 1975]. A series of 60° to 80°-striking ridges separate this deep trough from a southern shallower trough that spans the central part of the surveyed region.

The northernmost ridge, striking \(066^\circ\), lies at an average depth of 4.85 km, with peaks rising to 4.5 km. Sidescan sonar images of this ridge indicate some slumping, but generally little sediment cover (Figure 1.2). The northern ridge is separated from the southern extent of the central ridge province by a passage averaging 5.1 km deep. The
Plate II: Color image of SYS09 bathymetry of the Seafloor Surveys International/OTC Limited PacRim Nova-Canton Trough survey. The original shipboard bathymetry mosaic was hand-contoured by D. M. Hussong at 50 m intervals and digitized by the Seafloor Surveys International staff. The swath width is ~3.4 times tow-fish altitude; bathymetry is accurate to 2% of tow-fish altitude. The color image shows a plan view of SYS09 bathymetry. The unusual color change interval was used to create the most drastic color changes at important bathymetric levels, and to correspond to the color palette intervals used in Plate III.
generally little sediment cover (Figure 1.2). The northern ridge is separated from the southern extent of the central ridge province by a passage averaging 5.1 km deep. The center ridge strikes 060° in the northeastern part of the study area, but changes strike to 079° in the southwestern region, where it joins the southernmost ridge. Depths on the center ridge average 4.8 to 4.9 km, but isolated peaks rise to 4.1 km. South of the center ridge, on the extreme eastern edge of the study area, lies a small trough plunging to more than 6.1 km. The eastern trough is elongated in a NE-SW direction, and continues eastwards beyond the surveyed region. Depths decrease sharply to the SW in a V-shaped pattern. The point of the "V" marks the intersection of the center and southern ridges, and the location at which the center ridge changes strike to E-W (Plate. I.II). The southern ridge is the widest of the three (12 to 22 km) and exhibits the most variable topography. It strikes 073° in the west of the survey, changing to 081° in the east. Isolated peaks trend in various directions, rising to 3.0 km below sea level. The average depth of the ridge is 4.2 km, but increases to ~5.1 km in the east. Seismic reflection profiles show ponded transparent sediments in depressions between the peaks with strong reflectors at the bottom of the sections which may be talus shed from the ridges (Figures 1.3, 1.4). Sidescan sonar images show scant sediment cover on the bathymetric highs (Figure 1.2).

3. Central trough

The southern ridge slopes steeply into a trough which widens from 7.5 km in the east to 12 to 14 km in the west. These slopes strike from 060° in the southwest to almost E-W east of 166°45'W. Basin depths range from 5.6 km to over 5.9 km. The low reflectivity exhibited by the sidescan images (Figure 1.2) indicates sedimentation throughout the trough which seismic reflection profiles show to be primarily transparent sediments (average thickness ~0.6 s), resting on buried fault blocks (Figure 1.4, B-B'). The buried fault blocks and the steep slopes into the basin from the north imply that this trough is structurally controlled.
Figure 1.3: Survey track lines of the SYS09 Nova-Canton Trough survey showing locations of the following seismic reflection profiles. Time markings are shown for two hour divisions.
Figure 1.4: Seismic reflection profiles collected aboard the R/V *Moana Wave* during the PacRim cable route survey. Profiles are lettered A-A' through E-E' to correspond to lettering shown on Figure 1.3. Two-way travel time intervals in seconds are shown. Vertical exaggeration = 6X.
4. Southern plain

The slopes from the basin up to the southern plain change trend from 090° in the eastern, to 067° in the western study area (Figure 1.2). Abyssal hill topography striking N140E° lies on a broad arch in the southern part of the surveyed area. Seismic reflection profiles (Figure 1.4, A-A', B-B', C-C') show that this crustal fabric is generated by normal-faulted blocks rising ~100 to 250 m above the surrounding seafloor. The strike of this inferred spreading fabric is distinct from all other structures in the survey, which strike from 60° to 90°. The R/V Mahi seismic reflection profiles (Rosendahl, 1972; Rosendahl et al., 1975) show a similar broad arch extending to 150 km west of the region mapped by SYS09, with an identical normal-faulted character. I infer that this indicates abyssal hill fabric extending to ~168°W. The N140E°-striking crustal fabric is evident on sidescan data to ~2°S (Figure 1.5). Lower lying areas are blanketed by a thin (average 0.06 s) transparent sediment layer which smoothes the underlying topography.

Other major features in the southern plain include a large (>20 km diameter) seamount (Dolmah Seamount), located at 0°58'S, 166°51'W, rising to 2.65 km, and a smaller (<6 km diameter) seamount in the southeast (Figure 1.2, Plate II). Dolmah Seamount exhibits higher reflectivity than the surrounding seafloor, indicating scant sediment cover, as do flows radiating from its summit (Figure 1.2). The timing of the most recent activity on Dolmah Seamount is uncertain, and no samples were recovered during the PacRim cruise. However, the R/V Mahi survey recovered fresh volcanic cinder and glass ~100 km from Dolmah seamount [Rosendahl, 1972; Rosendahl et al., 1975]. The R/V Mahi may have sampled debris from this or other Neogene seamounts.

North of the study area, thick accumulations of turbidites cover the seafloor, however crustal fabric striking ~N160°E can be detected, though muted by sediments (Figure 1.6). At the extreme northern edge of the study area, a large-throw normal fault striking N162°E steps into the northern plain (Figure 1.2, Plate II). The crustal fabric
Figure 1.5: Composite image of single-channel seismic reflection profiles, SYS09 sidescan-sonar and bathymetry of the region immediately south of the SYS09 Nova-Canton Trough survey. The bathymetry is contoured at 100 m intervals, white arrows indicate the location of the N140°E-striking spreading fabric, and the seismic profiles show this fabric in profile. The bathymetry is surfaced and gridded from SSI digitized contours using the GMT System, Version 2.1 [Wessel and Smith, 1991].
Figure 1.6: Composite image of single-channel seismic reflection profiles, SYS09 sidescan-sonar and bathymetry of the region between 1°10'N to 2°10'N. The bathymetry is contoured at 100 m intervals. The original crustal fabric is heavily obscured by turbidites, however, abyssal hill fabric striking ~N160E° (white arrows) can be detected, as shown in the seismic profile and reflectivity image. The bathymetry is surfaced and gridded from SSI digitized contours using the GMT System, Version 2.1 [Wessel and Smith, 1991].
north of the study area is orthogonal (±7°) to the overall N70°E strike of Nova-Canton Trough structures.

Regional Seafloor Depths Adjacent to the Nova-Canton Trough

Using the seismic reflection profiles collected during the PacRim survey to estimate sediment thickness, I backstripped sediments from the traverses leading into and out of the Nova-Canton Trough survey (Plate I) to determine if a depth to basement difference exists across the Nova-Canton Trough once it has been adjusted to account for sediment loading. Densities of 1.03, 2.0, and 2.7 g/cm$^3$ were used for water, sediments, and basement respectively, and a velocity of 1.8 km/s was used to determine sediment thickness from the time sections.

The northern line exhibits an average sediment thickness of 240 m, composed primarily of turbidites. Sediment thickness was measured to the top of a series of very strong parallel reflectors, which may be basalt flows postdating the age of seafloor formation, as Winterer [1976a, 1976b] has suggested (profiles were not digitally recorded and are generally of poor quality; I am unable to recognize a deeper basement beneath them). The average water depth along this traverse is 5.267 km. Backstripping the sediments yields an average unloaded depth to basement of 5.405 km ± 0.0898 (Table 1.1).

The seafloor along the southern traverse (average water depth 5.373 km) shows an average sediment thickness of only 68 m, and greater extremes in relief than the northern traverse. The average backstripped depth to basement was calculated to be 5.399 km ± 0.228, essentially indistinguishable from that north of the Nova-Canton Trough despite the 100 m variation in depth to seafloor (Table 1.2).
Table 1.1. Backstrip Chart for the Region North of the Nova-Canton Trough

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<th>time (year/Julian day/hour)</th>
<th>depth (seconds)</th>
<th>depth (meters)</th>
<th>basement depth (seconds)</th>
<th>sed. thick. (seconds)</th>
<th>sed. thick. depth (meters)</th>
<th>backstrip amount (meters)</th>
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Table 1.1. (continued) Backstrip Chart for the Region North of the Nova-Canton Trough

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avg. backstripped depth
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Table 1.2. Backstrip Chart for the Region South of the Nova-Canton Trough

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Table 1.2. (continued) Backstrip Chart for the Region South of the Nova-Canton Trough

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Table 1.2. (continued) Backstrip Chart for the Region South of the Nova-Canton Trough

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Table 1.2. (continued) Backstrip Chart for the Region South of the Nova-Canton Trough

<table>
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<th>time (year/Julian day/hour)</th>
<th>seafloor depth (seconds)</th>
<th>seafloor depth (meters)</th>
<th>basement depth (seconds)</th>
<th>basement depth (meters)</th>
<th>sed. thickness (seconds)</th>
<th>sed. thickness (meters)</th>
<th>backstrip amt. (meters)</th>
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avg seafloor depth 5372.90

avg. backstripped depth 5398.56
Testing Menard's Hypothesis

The free air gravity field (Figure 1.1) derived from GEOSAT (W. F. Haxby, personal communication, 1990) altimetry data shows a lineament stretching from the Clipperton Fracture Zone, through the Line Islands volcanic edifice and sediment apron, to the Nova-Canton ridge and Nova-Canton Trough. I interpret this lineament to indicate structural continuity between the Clipperton Fracture Zone and Nova-Canton Trough. Several Pacific fracture zones extend into the central Pacific. Recent GLORIA surveys west of the Hawaiian Islands have insonified two strands of the Molokai Fracture Zone stretching at least as far as 165°W [Searle et al., 1993]. Free-air gravity data show the Clarion Fracture Zone extending to 165°W, where it is obscured by the Line and Crosstrend seamounts. The Murray Fracture Zone extends westward, beyond the Hawaiian-Emperor seamount chain, and the Marquesas and Galapagos fracture zones extend west of the Line Islands (Figure 1.1). Given this evidence, it appears unlikely that the Nova-Canton Trough, lying between these fracture zone extensions, should have originated by a different mechanism.

A more conclusive test of Menard's fracture zone hypothesis lies in basic plate tectonic principles. If the Nova-Canton Trough and Nova-Canton ridge are indeed the western extension of the Clipperton Fracture Zone, these structures must be co-polar with other coeval Pacific fracture zone traces; that is, all western fracture zone extensions should form small circles about the relevant stage pole of Pacific-Farallon opening. By analyzing the trends of western fracture zone extensions from the Pau to the Marquesas fracture zones (Figure 1.1), I have derived a congruent pole of Pacific-Farallon opening for the period prior to and including Chron 34. Because I am analyzing only fracture zone trends, without the constraining influence of spreading velocities during the Cretaceous Normal Superchron, a precise pole of opening is difficult to determine [Francheteau et al., 1970];
however, a pole located at approximately 34.4°S, 150.6°W fits the fracture zone traces very well. The 95% confidence ellipse is shown in Figure 1.7 and the fracture zone trend data used to constrain this pole are given in Table 1.3. The oblique projection of the free air gravity field employing this pole (Figure 1.8) shows that the western fracture zone extensions do, indeed, form small circles about this pole in that, in the oblique projection, the fracture zone traces are horizontal and parallel. Although several shorter duration stage poles may better fit these fracture zone traces, this pole works very well as a first-order approximation, and serves to corroborate Menard's hypothesis regarding the relation between the Nova-Canton Trough and Clipperton Fracture Zone.

The objections to Menard's fracture zone origin hypothesis outlined above may now be addressed:

1) Although data from the PacRim survey do not show a clear offset of basement depths across the Nova-Canton Trough, even when sediment loading is taken into account, the absence of a regional depth change is not unexpected for a Mesozoic fracture zone [Wessel and Haxby, 1990]. For example, I point out that there is no systematic depth offset across the Central Pacific Fracture Zone in the Phoenix lineations, although the magnetic isochrons are offset by 180 km (Plate I). Further, a depth change of only 50 m is predicted across a 100 m.y. old fracture zone with a 5 m.y. offset [Parsons and Sclater, 1977]. In a region as thermally perturbed as the central Pacific basin, this amount of depth offset across a fracture zone is within the "noise."

2) McNutt et al. [1989] interpret the Nova-Canton ridge, spanning the region from 165°W to the Line Islands (Plate I) as the surface expression of weak hot spot volcanism produced by passage over the Marquesas hot spot plume ~30 m.y. ago. According to McNutt et al. [1989], the Marquesas hot spot plume is too weak to penetrate normal oceanic crust, but the passage of a zone of weakness over the plume may allow the hot spot magmas to penetrate the surface. A ridge on the Marquesas Fracture Zone, a similar ridge
Figure 1.7: The 95% confidence region for an Euler pole located at 34.4°S, 150.6°W (diamond), for Pacific-Farallon opening during the Cretaceous Normal Superchron. The long axis of the ellipse is oriented NNW, being unconstrained by spreading velocity information. This confidence ellipse was calculated by W. W. Sager for 26 observations with a standard deviation of 2° (see Table 1.3).
Table 1.3. Observation Site Locations and Model Results Used in Euler Pole (150.6°W, 34.4°S) Calculations for Pacific-Farallon Opening During the Cretaceous Normal Superchron.

<table>
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<tr>
<th>Site*</th>
<th>Site Latitude (°N)</th>
<th>Site Longitude (°E)</th>
<th>Obs. Trend (degrees)</th>
<th>Calc. Trend‡ (degrees)</th>
<th>Difference (degrees)</th>
<th>Weight</th>
</tr>
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<td>-4.3</td>
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</tr>
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<td>78</td>
<td>77.2</td>
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<td>0.03</td>
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* Sources of fracture zone trend observations as follows: Galapagos FZ, [Mammerickx, 1992]; Clipperton and Clarion FZs, [Mammerickx and Smith, 1982; Mammerickx and Smith, 1985]; Molokai FZ, [Searle et al., 1993]; Murray, Pioneer, and Pau FZs [Mammerickx, 1989a; Mammerickx, 1989b; Atwater and Severinghaus, 1989a; Atwater and Severinghaus, 1989b].
‡ Standard deviation of 2° used in calculations.
on the Galapagos Fracture Zone, and the Nova-Canton ridge, lie in the predicted Pacific hot spot track [McNutt et al., 1989, Figure 6]. If McNutt et al. [1989] are correct in their hypothesis, a zone of crustal weakness underlies the Nova-Canton ridge, indicative of a fracture zone trace, implying structural continuity between the Nova-Canton ridge and Clipperton Fracture Zone.

3) The linear topographic depressions stretching from the Phoenix Islands to 180° span a much wider area than the fairly narrow Nova-Canton Trough (Plate I, Figure 1.1, arrow) implying a different mechanism of formation. I suspect that these lineaments are related to the M-series Pacific-Phoenix spreading regime, rather than a western extension of the Nova-Canton Trough. In support of this interpretation, Nakanishi et al. [1992] have identified conjugate anomalies M1 through M3 in the region between the Central Pacific Fracture Zone and Phoenix Fracture Zone in the Phoenix lineation set, indicating that spreading ceased on this segment of the Pacific-Phoenix spreading axis at ~M1 time. The conjugate anomalies lie in the region of the linear bathymetric depressions. In addition, these lineations appear to have been formed about a different pole of opening than the ridge-trough structures associated with the Nova-Canton Trough as shown by their disparate trend when projected about the best fit Euler pole for the central Pacific fracture zones. These lineations form neither small nor great circles about the pole located at 34.4°S, 150.6°W. Unfortunately, the paucity of bathymetric data available for this area prevents more specific characterization of its morphology.

Summary

I propose that the eastern Nova-Canton Trough and Nova-Canton ridge, extending eastward from 169°W to the Line Islands, are the western extension of the Clipperton Fracture Zone. The asymmetric profiles of ridge-trough structures [Rosendahl, 1972] are
Figure 1.8: Oblique Mercator projection of the slope of the free air gravity field from GEOSAT altimetry data, using the best-fit middle Cretaceous Pacific-Farallon pole (34.4°S, 150.6°W). The western fracture zone extensions (between the white diamonds) are generally parallel and horizontal, forming small circles around this pole. In this projection, the pole change following Chron 34 (white vertical lines) is especially apparent. The white arrow shows the location of the lineated bathymetric depressions inferred to have formed at the abandoned Pacific-Phoenix spreading axis at ~M-0 time. Note the difference in the strike of these lineations, as compared to the strike of the Nova-Canton Trough and Ridge further to the east. Compare the strike of these lineations in this oblique projection to their strike in a real world projection in Figure 1.1.
characteristic of fracture zone morphology, with bordering ridges that are neither continuous nor symmetric. The juxtaposition of the N140ºE-striking spreading fabric to the south and N160ºE-striking crustal fabric to the north, indicates that no significant amount of spreading has occurred at the trough. Although the tectonic spreading fabric imaged by SYS09 south of the Nova-Canton Trough (Figure 1.2, Plate II) is not orthogonal to trough structures, I do not believe that this precludes a fracture zone origin. Oblique spreading fabric is commonly reported adjacent to Atlantic fracture zones [e.g., Atwater and Macdonald, 1977; Tucholke and Schouten, 1988] at kinks in the fracture zones traces, and has been observed near the Siqueiros Fracture Zone on the East Pacific Rise [Fornari et al., 1989]. This trend continues over 100 km south of the study area (Figure 1.5), indicating that the obliquity is not simply a "bending" effect due to proximity to the fracture zone, but must have formed in association with a more regional feature, such as a propagating spreading center. Localized extremes of relief may indicate that this feature was formed near a reorganization of the post-M1 age Pacific and adjoining plates, or that subsequent tectonism occurred, initiated by thermal rejuvenation of the region, as proposed by Rosendahl [1972] and Larson [1991].

That the Nova-Canton Trough is the middle Cretaceous extension of the Clipperton Fracture Zone has major implications for the history of Pacific basin evolution. The location and co-polarity of the western fracture zone extensions (Figure 1.8) imply that a Pacific-Farallon spreading pattern was in effect throughout a large part of the Cretaceous Normal Superchron, not only in the north Pacific, but also in the central Pacific, west of the Line Islands seamount chain and east of the Manihiki Plateau.
CHAPTER II

Examination of the Morphology and Sedimentary Units of the Nova-Canton Trough Region

Introduction

The SYS09 sidescan-sonar/bathymetry survey in the eastern region of the Nova-Canton Trough [Joseph et al., 1992] and a compilation of echo-sounder profiles from the National Geophysical Data Center database and the Geological Survey of Japan (GSJ) provides an excellent opportunity to examine the regional setting of this complex seafloor province. I contend that the eastern Nova-Canton Trough is the central Pacific extension of the Clipperton Fracture Zone, based on the Trough's co-polarity to coeval central Pacific fracture zones and the sub-orthogonal strike of abyssal hill fabric abutting trough structures. However, there are elements in the structural expression of this complex feature that suggest a more complicated evolution. For example, the extreme depths observed in the western Nova-Canton Trough does not conform to observed fracture zone or abandoned spreading center models.

The Nova-Canton Trough comprises a system of deep, linear troughs, intermittently flanked or interrupted by steep ridges. Bathymetric coverage is sparse west of 169°W, however trough structures appear to extend to the Phoenix Islands at ~171°30'W [Rosendahl, 1972; Rosendahl et al., 1975]. At ~165°W the ridge-trough system is replaced by the Nova-Canton ridge [Winterer, 1976a], spanning the region from the ridge-trough
system to the Line Islands. East of the Line Islands, trough morphology reappears and continues eastward across the equatorial Pacific as the Clipperton Fracture Zone.

Various attempts to define the nature of the Nova-Canton Trough have characterized it as the western extension of the Clipperton Fracture Zone [Menard, 1967; Larson and Chase, 1972; Rosendahl, 1972], as a failed rift left by a change in spreading velocities [Winterer et al., 1974; Rosendahl et al., 1975], as the western spreading axis of a Pacific-Phoenix-Farallon triple junction at M0 time [Winterer, 1976a; Winterer, 1976b; Yamazaki and Tanahashi, 1992], again as an extension of the Clipperton Fracture Zone [Joseph et al., 1992], and most recently as a complex structure involving both spreading ridge and transform fault tectonics [Joseph et al., 1993].

This chapter examines the morphologic expression of the Nova-Canton Trough region in detail, describes the various sedimentary layers and structural provinces, and interprets morphologic elements in terms of the proposed evolution of Nova-Canton Trough.

Data Analysis

The bathymetric map shown in Plate III was created by applying a continuous minimum curvature surfacing algorithm [Smith and Wessel, 1990; Wessel and Smith, 1991] to 3.5 kHz echo sounder data compiled from the National Geophysical Data Center database, 3.5 and 12 kHz echo-sounder values collected during 1981 and 1982 GSJ R/V Hakurei-Maru cruises [Nakao, 1986; Usui, 1992], and the high resolution SYS09 swath bathymetry collected during the PacRim cable route survey [Joseph et al., 1992] (Figure 2.1). An examination of the error between crossing points of the various cruises yielded a mean error of 36 m [Wessel, 1989]. Given the extreme relief found in the Nova-Canton Trough (~2.2 km to >8 km), an error of this magnitude is insignificant, however many data points from the Eurydice cruise across the western trough were removed because the depth
Plate III: Bathymetry of the Nova-Canton Trough region (shown by the red box in Plate I) projected about a pole located at 34.4°S, 150.6°W produced from 3.5 and 12 kHz echo sounder data, and SYS09 swath bathymetry. Note the N140°E-striking spreading fabric extending across the region south of the Nova-Canton Trough. Data sources include the National Geophysical Data Center as well as the GSJ (cruises GH80-1, GH81-4, and GH82-4). The data were surfaced and gridded using the GMT System, Version 2.1 [Wessel and Smith, 1991]. Contour lines are drawn at 500 meters depth intervals. Letters refer to structural provinces discussed in the text.
Figure 2.1: The cruises included in the gridding process used to produce the bathymetric map in Plate III, viewed in the same projection. The PacRim contoured swath bathymetry is shown as thin black lines (R/V Moana Wave); gray diamonds represent data values collected during GSJ cruises conducted aboard the R/V Hakurei-Maru; black crosses are R/V Mahi cruises; small black triangles are R/V Conrad and Scripps' Eurydice Expedition data points; squares represent bathymetric values collected during R/V Dmitri Mendeleyev cruises; the thick line shows the location of R/V Kana Keoki data points. In the gridding process, the PacRim cruise (swath bathymetry) was weighted at 1, the R/V Hakurei-Maru (GSJ) cruises at 2, the R/V Kana Keoki at 3, the R/V Dmitri Mendeleyev and Eurydice Expedition at 4, and the R/V Conrad and R/V Mahi cruises at 5.
errors appear to be due to positioning errors. Depths in this region are better constrained by more recent GSJ data. Because of the uneven and sparse distribution of depth values across the deep western trough and surfacing artifacts resulting therefrom, the region was gridded in two segments. Bathymetric values were projected around the best-fitting Euler pole for this region at 34.4°S, 150.6°W [Joseph et al., 1993] to translate the ridge-trough structures into an E-W orientation. The region was gridded with an anisotropic grid cell 5 km by 2 km north of Dolmah Seamount elongated in the oblique east-west direction, and an isotropic grid cell 1 km by 1 km across the entire southern region. This procedure created some inconsistencies at the juncture between the two segments. The misfit values were removed by hand-contouring through the juxtaposed area, and inserting the corrected values back into the surfaced data. The elongate 5 km by 2 km grid cell used north of Dolmah Seamount perhaps overly smoothed the northernmost region but there is little data to better constrain seafloor relief in this area (Figure 2.1).

Topographic profiles were selected from the research cruises traversing the Nova-Canton Trough (Figure 2.1) to examine the topography in cross-section. Figure 2.2 shows the location of selected profiles and the data density. The selected profiles were projected perpendicular to a line drawn through the strike of the deep western trough to the edge of the study area to show the variation in ridge-trough morphology along strike (Figure 2.3).

Morphology of the Nova-Canton Trough Region

The western trough (Plate III, A; Tanahashi's [1992] nomenclature will be used throughout) attains a depth of 6.6 km at the westernmost traverse (profile gh8a1, Figures 2.2, 2.3), flanked to the south by a steep ridge (Plate III, D) with a smaller ridge paralleling the trough to the north. Progressing eastwards, profiles c1006 through mah02.5 (Figures 2.2, 2.3) cross the western trough at its maximum width (~10 km) and
Figure 2.2: (Top) Bathymetry of the Nova-Canton Trough region projected about a pole at 34.4°S, 150.6°W. The contour interval is 200 m and ship tracks for bathymetric profiles presented in this analysis are superimposed on the bathymetry. (Bottom) Ship tracks are shown with location of data values (small circles). The prefixes used for the various cruises follow L-DEO nomenclature [Wessel and Smith, 1991]: c1006: R/V Conrad; dmm24, dmm28: R/V Dmitri Mendeleyev; gh8a1, gh8b1, gh82a, gh82b: R/V Hakurei-Maru (GSJ); kk090: R/V Kana Keoki; mah02: R/V Mahi; mw902: R/V Moana Wave (PacRim cable route survey).
Figure 2.3: Bathymetry from 3.5 and 12 kHz echo-sounder data for the ridge-trough province projected normal to a line drawn through the deep western trough to the edge of the study area. The small dots indicate the position of projected data values. The profiles are arranged consecutively from west to east to show the variation in ridge-trough morphology along strike from (a), the westernmost profile, to (ab), the easternmost profile. From profile gh8a1 (a) to profile gh82a.5 (d), there is a single ridge bordering the trough to the south. At profile mah02.6 (e), a parallel ridge to the north appears and continues along strike until the western trough disappears at profile mah02.3 (l), replaced by the central ridge. The central trough begins at profile gh82a.5 (d) and continues eastwards to the edge of the study area at profile gh82b.1(ab). The eastern trough first shows bathymetric expression as a small cleft in the central ridge at profile mw902.2 (v), and deepens progressively eastwards to profile gh82b.1 (ab).
deepest point (~8 km depth) where the bordering ridge reaches a minimum depth of 2.2 km but decreases in elevation progressively eastward. The central trough (Plate III, B) borders the ridge to the south, beginning at 167°45'W (Figures 2.2, 2.3, profile gh82a.5) continuing past the termination of the ridge to the edge of the study area (Figures 2.2, 2.3, profile gh82b.3). The central trough is ~6 km wide and has a minimum depth of 5.4 km in the west, but deepens progressively eastwards to ~5.9 km and widens to 11.5 km south of the central ridge (Plate III, E). Seismic reflection profiles collected during the PacRim survey show that the central trough contains an average of ~0.6 s of transparent sediments with several strong reflectors at the top and midway through the sediment column. Crustal blocks, similar to the blocks in the southern region, are not truncated at the trough (Plate III, J, Figure 1.4, B-B') but continue into the central trough, underlying the sediments.

The eastern trough (Plate III, C, Figures 2.2, 2.3, profiles mw902.2, mw902.1, gh82b.2, and gh82b.1) begins at 166°35'W at ~6.0 km depth, increases to 6.2 km depth to the east and widens to ~13 km at the edge of the study area. It is flanked to the north and south by ridges, the central ridge to the south, and another, smaller ridge to the north which is paralleled by a second ridge ~10 km further north. Transparent sediments 0.4 to 0.5 s thick with a few strong reflectors near the top and midway through the sediment column fill the trough bottom [Tanahashi, 1992].

The central ridge (Plate III, E) begins at 168°W, flanking the western trough to the north (Figures 2.2, 2.3, profiles mah02.6, dmm28.1, gh82a.4, kk090, mah02.4, and gh82a.3). The ridge continues past the end of the western trough as the trough shoals along strike to the east (Figures 2.2, 2.3, profiles mah02.3, gh82a.2). At ~167°20'W, the ridge broadens and bifurcates, encompassing the entire region between the western and eastern troughs. Between 167°20'W and 166°35'W, the sole trough structure along strike is the central trough, bordered to the north by the central ridge. At 166°40'W, the dual ridges join and change trend to E-W, paralleling a bend in the central trough (Plate III).
Turbidite layering covers the region north of the ridge-trough topography (Plate III, F) to varying depths, but averages ~0.6 s thick [Joseph et al., 1992; Tanahashi, 1992; Kroenke, 1970; Orwig, 1981]. The turbidite layers are overlain by ~0.08 s of transparent sediments. SYS09 sidescan sonar and bathymetry [Joseph et al., 1992, 1993] show a normal fault striking N160°E, orthogonal to the overall Nova-Canton Trough trend (Figure 1.2, Plate II) at the extreme northern edge of the SYS09 survey. A chain of small seamounts protrudes through the sediments, paralleling the Nova-Canton Trough trend (Figure 1.2, Plate II).

The GSJ surveyed the region south of the Nova-Canton Trough in detail in 1982 [Usui, 1992]. Profiles from GSJ cruises gh82a and gh82b trending parallel to the Nova-Canton Trough (Figure 2.2) were projected perpendicular to a line drawn along a N140°E-striking elongate high at 167°W, 2°S (Plate III, I, Figure 2.4).

The seafloor south of the Nova-Canton Trough exhibits a regional depth change across the N140°E-striking elongate ridge (Plate III, I). West of the ridge, a turbidite-filled basin striking NW-SE lies at ~5.6 km depth (Plate III, G); east of the ridge, the seafloor is generally rougher, covered with small seamounts and abyssal hills at a mean depth of 5.3 km. The depth difference across the ridge prompted Yamazaki and Tanahashi [1992] to propose a transform fault origin of the elongate ridge.

The western turbidite-filled basin (Plate III, G, Figures 2.2, 2.4, profiles gh82a.e, gh82a.d, gh82a.c, gh82a.b, and gh82a.a) [Tanahashi, 1992] is a generally flat expanse of seafloor ~65 km wide, covering the region south of the deepest expression of the western trough and bordering ridge, to the southern edge of the study area. Seismic reflection profiles collected during GSJ cruises gh82a and gh82b [Tanahashi, 1992] show that the sedimentary section in the basin changes from intercalated transparent sediments and turbidites in the south to a central region of small, domed basement highs with ponded sediments between the highs, to semi-opaque sediments immediately south of the southern
Figure 2.4: Projected bathymetry profiles across the region south of the ridge-trough province. 3.5 kHz echo-sounder data were projected normal to a line drawn through the eastern edge of the ridge at 167°W, 2°S and parallel to the spreading fabric in the southeastern region. Small dots show the location of projected data values.
ridge. A second NW-SE-trending turbidite-filled basin ~30 km wide lies immediately east of the elongate ridge (Plate III, H). The central turbidite-filled basin exhibits an average 0.6 s thick column of deformed turbidite layering surrounding small conical basement highs [Tanahashi, 1992].

The seafloor east of the elongate high (Plate III, J) exhibits only a thin (average ~0.06 s) veneer of transparent sediments, and is generally much rougher than the region west of the ridge (Figures 2.2, 2.4, profiles gh82b.b, gh82a.c, gh82a.f, gh82a.g, gh82a.h, gh82b.c, and gh82b.d). A seafloor grain trending N140°E is apparent throughout this region (Plate III, J). SYS09 sidescan sonar and bathymetry (Figure 1.2, Plate II) reveal this structural grain to be N140°E-striking abyssal hills. The additional bathymetric coverage provided by GSJ and R/V Dmitri Mendeleyev cruises (Figure 2.2) reveal Dolmah Seamount to be double-peaked, lying immediately south of the central trough (Plate III). Many smaller circular bathymetric highs are interspersed throughout the region.

Western Extent of Ridge-Trough Morphology

Free air gravity values derived from GEOSAT altimetry data (W. F. Haxby, personal communication, 1990) can aid in determining the western extent of the ridge-trough province, and its relationship to lineated bathymetric depressions west of the Phoenix Islands that have been conjectured to form the western extent of the Nova-Canton Trough [Winterer et al., 1974]. Because of the sparse shipboard bathymetric coverage west of ~168°30'W, I examined the GEOSAT gravity field to determine if these data will aid in recognizing structures similar to the ridge-trough morphology in the Phoenix Islands area.

Seismic reflection profiles collected aboard the R/V Mahi in 1970 extend to 170°30'W indicating that the western trough and southern ridge are still well developed to
Figure 2.5: Enlarged image of the slope of the free air gravity field derived from GEOSAT altimetry data for the Nova-Canton Trough region cut from Figures 1.1 and 1.8. (a) The upper Mercator projection shows the entire Nova-Canton Trough region, including the lineated depressions lying west of the Phoenix Islands and the Clipperton Fracture Zone to the east. ETOP05-derived contours for depths less than 3000 m are superimposed on the figure to aid in locating structures on the lower oblique projection. (b) GEOSAT data for the Nova-Canton Trough region projected about the best-fit pole for the Cretaceous Normal Superchron at 34.4°S, 150.6°W.
169°W although the relief is greatly reduced, and the western trough extends to 170°30′W as a shallow trough with small bordering ridges [Rosendahl, 1972, Figure 5]. The slope of the free air gravity field derived from GEOSAT altimetry data (Figure 2.5a) show lineaments extending west from the deep western trough through the Phoenix Island chain. When the GEOSAT altimetry data are projected about the best-fit Euler pole for the Cretaceous Normal Superchron at 34.4°S, 150.6°W (Figure 2.5b), it is clear that the lineaments extending west from 171°30′W strike at a very different angle implying formation about a different pole.

To determine if these lineaments were formed by Pacific-Phoenix spreading, I projected the GEOSAT altimetry data about the Pacific-Phoenix pole of opening for the latest M-series magnetic anomaly isochrons [Engebretson et al., 1985] at 74°E, 17°S (Figure 2.6). Spreading-related lineaments should form great circles about the Euler pole. The lineations deviate by ~8° from paralleling the great circle trend about this pole (Figure 2.6). This may indicate that the plate geometry was changing when these depressions were formed, or that the pole is slightly in error. However, it is interesting to note that the overall trend of the seamount chain crossing the bathymetric depressions is orthogonal to the pole of opening except at the extreme northern and southern ends, where the trend deviates in opposite directions (Figure 2.6). This implies that the pole of rotation changed during the time interval between formation of the two seamount chain trends. The northern seamounts cross the easternmost M1-M3 isochrons in the Phoenix lineation set (Plate I). The conjugate M1-M3 isochrons may lie in the region of the southern seamounts (~173°W, 8°S), however recent magnetic anomaly mapping in the central Pacific basin [Nakanishi et al., 1992] has not identified any M-series isochrons in this southern region (but few survey lines cross the area). If the lineaments west of the Phoenix Islands were not formed by steady-state Pacific-Phoenix spreading, they may have resulted from changing ridge crest geometry through rift propagation, as their slightly disparate trends suggest.
Figure 2.6: The slope of the free air gravity field derived from GEOSAT altimetry data for the Nova-Canton Trough region projected about the Pacific-Phoenix pole of opening for the latest M-series lineations, located at 74°E, 17°W [Engebretson et al., 1985]. Spreading-related trends are oriented vertically. Note that the lineated depressions lying west of the Phoenix Islands deviate by ~8° from following a great circle trend about this pole. The strike of the seamount chain is orthogonal to the pole of opening except at the distal ends of the chain, where the strike deviates in opposite directions.
Sedimentary Units of the Nova-Canton Trough Region

Because no seafloor samples were recovered during the PacRim survey I shall briefly describe the sedimentary units present in this region based on previous studies. During the GSJ R/V Hakurei-Maru cruise in 1982 several piston cores, box cores, and free-fall grabs of surface materials were recovered from the region south of the ridge-trough province [Nishimura and Ikehara, 1992]. The sediments north of the Nova-Canton Trough were sampled during a GSJ cruise in 1981 [Nishimura, 1986] and various other locales in the region sampled during the GSJ’s Wake-Tahiti transect in 1980 [Nakao and Mizuno, 1982]. The sedimentary sequences delineated from seismic reflection profiles have thus far been described as 1) a transparent layer, 2) a series of closely spaced parallel reflectors, probably turbidites, and 3) a semi-opaque layer.

The transparent layer defined from seismic reflection profiles occurs in the southeastern region, overlying the stratified layer in the northern plain, and filling the three troughs. Direct sampling indicates that the transparent layer is composed of siliceous mud or siliceous ooze, locally rich in radiolarian and diatom tests [Nishimura and Ikehara, 1992, Nakao and Mizuno, 1982]. Although the Carbonate Compensation Depth (CCD) presently lies at ~4900 m below sea level in the central equatorial Pacific [Berger et al., 1976], calcareous sediments are often observed several centimeters from the surface north of the Nova-Canton Trough indicating a deeper CCD in the past [Nakao and Mizuno, 1982].

Three basins in the immediate Nova-Canton Trough region exhibit the highly stratified acoustic layering characteristic of turbidite deposition. Turbidite sections have been recovered from the western and central basins south of the ridge-trough province (Plate III, G, H) and from a small NW-SE-striking basin at 167°N, 1°30'S [Nishimura and Ikehara, 1992]. The calcareous composition of the turbidite units indicate that they were formed above the CCD, suggesting the Manihiki Plateau to the southeast as the source
The northern basin (Plate III, F) also exhibits turbidite sequences, however the relief of the ridge-trough province probably prevented deposition from the southeast. A probable source for the turbidite layers north of the Nova-Canton Trough is the Line Islands or Central Basin Rise [Rosendahl, 1972; Rosendahl et al., 1975; Kroenke, 1970; Orwig, 1981; Lineberger, 1975], the only regions in the eastern central Pacific basin with sufficient relief to generate turbidity currents. In addition, the turbidite units are present in the northeastern section of the Nova-Canton Trough region but disappear further to the west. Turbidite units are not visible on seismic reflection profiles for traverses gh8a1 or gh82a.6 (Figure 2.2) suggesting that flow originated in the east or northeast, in agreement with the turbidite distribution pattern defined by Orwig [1981].

There is no direct information to determine the composition of the semi-opaque unit observed in the deep western trough and immediately south of the southern ridge. Given the composition of the transparent layers present in the eastern and central troughs (siliceous mud and ooze), I surmise that the semi-opaque acoustic layers may represent these pelagic sediments mixed with talus derived from the southern ridge. That the semi-opaque layers occur intermittently throughout the section and show some disturbance [Nishimura and Ikehara, 1992] may indicate that structural modification occurred after primary genesis of the western region. Strong reflectors within the transparent layer overlying the semi-opaque layer in the western trough are interpreted by Nishimura [1992] to represent interbedded turbidites.

**Discussion**

The combined bathymetry gleaned from cruise data available through the National Geophysical Data Center, GSJ [Tanahashi, 1992; Yamazaki and Tanahashi, 1992], and the SYS09 swath-mapped data indicate pervasive N140°E trending crustal fabric south of the
Nova-Canton Trough as far west as ~167°30'W (Plate III). This bathymetric fabric strikes at a high angle to the overall N70°E strike of the Nova-Canton Trough, paralleling the elongate ridge at 167°10'W, 2°S (Plate III) and the fault structures of the northeast margin of the Manihiki Plateau (Plate I). The trend of the crustal fabric throughout the southeastern region appears to support my contention that the eastern Nova-Canton Trough is a fracture zone. However, the western trough is not juxtaposed by this fabric trend, nor is any other crustal fabric trend evident in the bathymetric data to aid in constraining the western extent of the fracture zone model of the Nova-Canton Trough (Plate III). Crustal fabric is obscured to the west of the elongate ridge by turbidites [Tanahashi, 1992]. Seismic reflection profiles collected during the 1970 R/V Mahi survey show faulted crustal blocks creating horst and graben structures in the region north of the western trough [Rosendahl, 1972; Rosendahl et al., 1975] that appear to parallel the ridge-trough structures. However, this megascopic evidence is not conclusive regarding the seafloor fabric adjacent to the western trough. Clearly, fine-scale swath-mapped bathymetry and reflectivity information is needed to determine the trend of terrain elements, if original crustal fabric is not obscured by sediments. The extreme depth and high, bordering ridge observed at the western trough, together with the diverse suite of samples recovered by a dredge of the southern ridge [Rosendahl, 1972; Rosendahl et al., 1975] and faulted, intercalated pelagic and talus-derived sediments within the trough [Tanahashi, 1992] suggested to previous researchers that a later stage of tectonism and/or volcanism might have been responsible for the anomalous relief. The entire central Pacific, from the Nauru Basin in the west, to the Line Islands in the east, is hypothesized by many researchers [e.g., Menard, 1964; Crough, 1978; Schlanger et al., 1981] to have experienced a thermo-volcanic event in the Late Jurassic which resulted in thermal rejuvenation and uplift of the entire region (Menard's "Darwin Rise"). More recently, Larson [1991] proposed a "superplume" episode beginning at ~125 Ma in the mid-Cretaceous south Pacific that so
perturbed the thermal structure at the core-mantle boundary that the earth's magnetic field ceased reversing for more than 40 m.y. [Larson and Olson, 1991]. The central Pacific basin formed ~4000 km southeast of where it presently lies. The gradual northwest movement of the Pacific plate during the last 125 m.y. would have necessitated movement of the central Pacific basin over the Darwin Rise (or "superplume") at some time in its history, which lends credence to Rosendahl's [1972] theory of thermal rejuvenation of the Nova-Canton Trough region. Indeed, as noted by others [e.g., Menard, 1964; Crough, 1978; Schlanger et al., 1981; McNutt and Fischer, 1987] ocean floor depths in the central Pacific basin are over 1000 m shallower than depths predicted by the Parsons-Sclater [1977] subsidence curve for seafloor of comparable age.

Although the trend of crustal fabric in the southeastern region is approximately orthogonal to the ridge-trough province, the Nova-Canton Trough's fracture-zone expression is not conformable to simple fracture-zone models [Fox and Gallo, 1986; Fox and Gallo, 1989]. The three troughs comprising the Nova-Canton Trough fracture-zone system range in depth from 5.9 to 6.2 to more than 8.0 km and are separated by high flanking ridges producing vertical relief of 5.8 km in the western trough and 2.8 km at the eastern trough. Such extreme relief is not normally observed at fracture zones elsewhere, although the Romanche Fracture Zone plummets to more than 7.7 km [Heezen et al., 1964]. Maximum ridge-trough relief at the Siqueiros and Orozco transform faults is on the order of 1.5 km, a fraction of what we observe at the western Nova-Canton Trough [Trehu and Purdy, 1984; Madsen et al., 1986; Fornari et al., 1989].

From the available data I cannot unequivocally state that the western trough was produced purely by transcurrent motion between two offset ridge segments (Plate III, D). However, its great depth and extreme, though asymmetric relief is more suggestive of transform fault origin than spreading ridge origin. The central trough and ridge are also problematic (Plate III, B and E). The abyssal hill fabric south of the central trough, rather
than curving through a broad arc to parallel the trend of the trough, as is observed at most fracture zones and transform faults (e.g. Fox and Gallo, 1989; Fornari et al., 1989; Searle et al., 1993), continues into the trough, underlying the sediments (Plate II, Figure 1.4, B-B'). This may indicate that the central trough is not a true fracture zone trough, once accommodating strike-slip movement, but may be the result of flexural response of the lithosphere adjacent to a fracture zone [Haxby and Parmentier, 1988]. However, this trough is over 6 km deep when sediments are removed, therefore ~700 m deeper than regional depths of ~5.3 km. If this trough did not accommodate strike-slip movement, it would be difficult to reconcile the role of the central ridge. However, similar morphology is observed at the Clipperton Transform Fault [Gallo et al., 1986], where a 40 km long median ridge, bounded to the north by a parallel trough, is interpreted as an extensional relay zone, accommodating transcurrent motion [Fox and Gallo, 1989; Gallo et al., 1986]. The bathymetric deep to the north of the median ridge is not considered to be the locus of transcurrent motion, nor is a similar sinuous deep paralleling the transform-fault zone to the south. The central ridge of the Nova-Canton Trough may be analogous to the median ridge located at the present Clipperton Transform Fault and may have been the locus of transform motion during the time interval between motion along the western and eastern troughs, reflecting a compressional stress environment during the central ridge’s active period. Abyssal hill fabric adjacent to the northern deep at the Clipperton Transform Fault continues parallel to the East Pacific Rise to the edge of the deep [Gallo et al., 1986, Figure 4], but curves into a transform-parallel trend within the deep. An analogous situation may exist in the Nova-Canton Trough’s central trough, because blocks similar to the blocks forming the crustal fabric south of the trough are observed beneath the trough sediments (Figure 1.4) but the sediments obscure the trend of the fabric. The abyssal hill fabric adjacent to the southern deep at the Clipperton Transform Fault appears to continue into the
basin without a change in strike, similar to what is observed at the junction between the central trough and southern basin of the Nova-Canton Trough (Figure 1.2, Plate II).

North of the eastern trough (Plate III, C) lies a N162°E-striking normal faulted block (Figure 1.2, Plate II) and further to the north abyssal hill fabric trends ~N160°E±7° (Figure 1.6), approximately orthogonal to the N70°E-strike of the Nova-Canton Trough. There is nothing in these data to contradict a transform fault origin for eastern portion of the Nova-Canton Trough, although the eastern trough is offset slightly to the north of the strike of the western trough (Figure 2.3). This offset may indicate a slight change in the pole of relative motion between periods of transcurrent movement that formed the two troughs. The overall copolarity of the entire ridge-trough province eastwards from ~171°30'W to coeval Pacific fracture-zone traces supports a transform-fault origin (Figure 1.8), although the bathymetric data in the western trough region, due to their sparseness, cannot confirm this.

The elongate ridge in the region south of the Nova-Canton Trough (Plate III, I) has been proposed to be of transform-fault origin due to the ~300 m depth difference across it [Yamazaki and Tanahashi, 1992]. However, the occurrence of abyssal hill fabric striking parallel to this structure contradicts a transform-fault origin for this structure. An alternate interpretation is that the ridge is a rifted margin juxtaposing crust created by pre- or syn-M0 Pacific-Phoenix N-S spreading to the west, and syn- or post-M0 Pacific-Farallon NNE-SSW spreading to the east. This interpretation is supported by the parallel trends of crustal fabric east of the ridge and the northeastern margin of the Manihiki Plateau, and the upward flexed profile of the ridge (Figure 2.4) [Weissel and Karner, 1989]. However, the shallower seafloor east of the ridge, the slight obliquity of the abyssal hill fabric, and the generally rougher seafloor of the southeastern region may indicate that this portion of seafloor is the trapped remnant of a microplate active during the emergence of the Manihiki Plateau. Indeed, the most seemingly obvious modern-day analogs to the Nova-Canton
Trough are the Hess and Dietz deeps associated with the Galapagos microplate [Lonsdale, 1988; Francheteau et al., 1990] and the Pito Deep at the northeastern boundary of the Easter microplate [Naar and Hey, 1991; Naar et al., 1991; Martinez et al., 1991] at the East Pacific Rise. Furthermore, an intervening microplate in this region could explain the lack of identified conjugate M-series magnetic anomaly isochrons south and east of the Nova-Canton Trough and the curious linear deeps west of the Phoenix Islands.

Summary

On the basis of compiled bathymetric data integrating NGDC-archived and GSJ cruises, and swath bathymetry collected during the PacRim cable route survey in the Nova-Canton Trough region, I have drawn the following conclusions regarding the structural expression of the Nova-Canton Trough's ridge-trough province and adjacent seafloor:

1. The western trough extends from ~170°30'W to 167°30'W, changing in structural expression from west to east from a shallow trough bordered to the north and south by small ridges, to a >8.0-km-deep trough bordered to the south by a 2.2-km-deep ridge, but shoaling further eastwards before being replaced by the central ridge. The western trough is bordered to the north by a narrower, deeper ridge, beginning at ~167°50'W which joins the central ridge to the east. The western trough's asymmetric profile is more characteristic of fracture-zone morphology than an extinct spreading-ridge profile, however the occurrence of faulted, intercalated sediments within the trough and the occurrence of gabbroic rocks recovered from the lower reaches of the southern bordering ridge suggest that magmatic or tectonic rejuvenation of the region may be responsible for the extreme relief.

2. The central trough borders the western trough south of the southern ridge eastwards from ~168°W, past the termination of the western trough to parallel the central ridge from
167°30'W to 166°30'W. For this ~100 km stretch of seafloor, the central trough is the only trough structure in the region. Transparent sediments fill the central trough to ~700 m depth, covering block structures similar to the blocks comprising the abyssal hill fabric evident in the southeastern region.

3. The eastern trough, beginning at ~166°W, reaches a maximum depth of 6.2 km in the study area, is filled with ~400 m of transparent sediments, and is bordered by a small ridge to the north. It is offset slightly to the north of the strike of the deep western trough, possibly indicating a slight change in the relative pole of motion in the intervening period between formation of the troughs. The strike of a normal-faulted block to the north of the eastern trough and the overall strike of seafloor-spreading fabric throughout the northern region (N160°W±7°) indicates that spreading was oriented normal to the eastern trough, in support of a transform-fault origin.

4. Tectonic spreading fabric in the southeastern region is oriented N140°E, slightly oblique to the overall N70°E strike of the ridge-trough province. This fabric trend pervades the southeastern region, paralleling an elongate ridge at 167°W, 2°S and the northeastern margin of the Manihiki Plateau.

5. Seafloor depths vary by ~300 m across the elongate ridge, changing from a smooth 5.6-km-deep turbidite-filled basin to the west to a rougher, shallower region to the east. The occurrence of the parallel spreading fabric contraindicates a transform-fault origin for this ridge, but supports a rift origin, as does the upward flexed profile of the ridge.

6. When projected about the best-fit Euler pole for the Cretaceous Normal Superchron at 34.4°S, 150.6°W, the strike of lineated bathymetric depressions west of the Phoenix Islands differ by ~30° from the small circle formed by the Nova-Canton Trough. The depressions may be related to the shutdown of the pre-M0 Pacific-Phoenix spreading system although they deviate by ~8° from forming great circles about the latest M-series Pacific-Phoenix pole of opening. However a seamount chain centered about this pole...
forms a small circle with opposite deviations at the chain's far ends, implying that the pole of relative motion changed during formation of the seamount chain.

7. An alternate explanation for the slight obliquity of spreading fabric in the southeastern region, the varying trends of the lineated depressions west of the Phoenix Islands, and the extreme relief encountered at the deep western trough is the invocation of an intervening microplate formed during the genesis of the Manihiki Plateau that was later transferred to the Pacific plate. Modern day analogs such as the Pito Deep at the Easter microplate boundary and the Hess and Dietz deeps at the Galapagos microplate boundary suggest a similar mechanism of formation. Unfortunately, supporting bathymetric evidence for this interpretation must await further reconnaissance of the Nova-Canton Trough region.
CHAPTER III

The Nova-Canton Trough and the Late Cretaceous Evolution of the Central Pacific

Introduction

While great progress has been made over the past twenty years in reconstructing the tectonic evolution of the Pacific basin, plate movements during the Cretaceous Normal Superchron remain uncertain. The absence of magnetic reversals during this period in the evolution of the central Pacific basin has hampered attempts to explain the change in spreading configurations from those inferred from the N80°E-trending Mesozoic Phoenix magnetic-anomaly lineations (Plate I) to spreading trends evidenced by the long, linear scars of the eastern Pacific fracture zones (Figure 1.1, Plate IV). The magnetic bight between the Phoenix and Hawaiian lineations indicates a three-plate system in effect in the central Pacific basin prior to MO time. Following the Cretaceous Normal Superchron, however, a two-plate configuration is observed in the region east of the Hawaiian, Line, and Marquesas seamount chains. The subsequent passage of the central Pacific basin over the south Pacific hotspots compounds the problem by obscuring original crustal fabric with a profusion of seamount chains and volcanic islands.

Like previous attempts to unravel the genesis of plate geometries in the central Pacific during the Cretaceous Normal Superchron, I re-examine the two major structures in that region, the Nova-Canton Trough and Manihiki Plateau (Plate I), and their relation to the Phoenix lineations. I use GEOSAT altimetry data (W. F. Haxby, personal communication, 1990) to show continuity of the Clipperton Fracture Zone through the Line Islands volcanic
Plate IV: Lambert azimuthal projection of the Pacific Basin centered on 150°W. Magnetic-anomaly lineations located on the Pacific plate are shown to indicate the extent of the Cretaceous Normal Superchron (QZ) throughout the entire central Pacific. The red box shows the location of the central Pacific basin shown in Plate I.
chain and sediment apron to the N75°E-striking Nova-Canton ridge and the N70°E-striking Nova-Canton Trough. I have also shown that the Nova-Canton Trough-Clipperton Fracture Zone trend is co-polar about a pole located at 34.4°S, 150.6°W with other coeval Pacific-Farallon fracture zone segments, from the Pau to Marquesas fracture zones.

This fracture-zone interpretation is corroborated by high-resolution data from recent sidescan-sonar [Joseph et al., 1992] and bathymetry [Tanahashi, 1992] surveys of the Nova-Canton Trough region. Both surveys reveal pronounced crustal fabric paralleling the trend of the northeast margin of the Manihiki Plateau (N140°E) and at a high angle to Trough structures. The combined evidence supports the interpretation that the Nova-Canton Trough is the Middle Cretaceous extension of the Clipperton Fracture Zone and that the northeast edge of the Manihiki Plateau is a rifted margin. On this basis I propose a new model for the transition from the Mesozoic three-plate geometry to the two-plate configuration that spans the central and northeastern Pacific.

Sidescan-Sonar and Bathymetric Evidence

The strike of abyssal hill fabric may be used in the absence of magnetic-anomaly lineations to define the strike of a paleo-spreading axis. Abyssal hill topography striking N140°E lies on a broad arch southeast of the Nova-Canton Trough. Seismic reflection profiles show that this crustal fabric is composed of normal-faulted blocks rising ~100 to 250 m above the surrounding seafloor (Figure 1.4). SYS09 sidescan data indicate that the N140°E-striking crustal fabric extends to at least 2°S (Figure 1.5). An elongate ridge with a pronounced flexural profile lies at 167°W, 2°S, paralleling the trend of the abyssal hill fabric.

North of the Nova-Canton Trough, thick accumulations of turbidites cover the seafloor muting crustal fabric striking ~N160°E, orthogonal to the N70°E-striking Nova-
Canton Trough (Figure 1.6). At the extreme northern edge of the SYS09 survey area, a normal fault striking N162°E steps into the northern plain paralleling the abyssal hill trends further to the north. These seafloor fabric trends suggest that north of the eastern portion of the ridge-trough system, spreading was oriented orthogonal to the Nova-Canton transform fault (Figure 1.2, Plate II).

Southeast of the Nova-Canton Trough lies the Manihiki Plateau, a large, heavily sedimentsed region of anomalously shallow, thick oceanic crust [Winterer et al., 1974; Hussong et al., 1979] (Plate I). Deep troughs, hypothesized to be of fault origin, divide the plateau into three distinct morphological provinces: the High, Western, and North plateaus [Winterer et al., 1974]. The trends defined by these structural boundaries led Winterer et al. [1974] to posit a Cretaceous triple junction in the northern Manihiki Plateau near the junction between the North, Western, and High plateaus. The High and Western plateaus are composed of narrow ridges and sediment-filled troughs forming a sub-orthogonal pattern along NNE- and NW-trends, interpreted by Winterer et al. [1974] as offset ridge-axis segments. The northeast margin of the North plateau is bounded by a region of N140°E-striking tilted fault blocks, similar to the topography bounding the eastern edge of the High plateau [Sharman and Mammerickx, 1990; Hill et al., 1991; Munch et al., 1992].

Tectonic Reconstruction of the Central Pacific Basin

The Pacific-Phoenix-Farallon magnetic bight (Plate I) indicates that the plate configuration in the central Pacific basin at M1 time was characterized by a ridge-ridge-ridge triple junction between the Pacific, Phoenix and Farallon plates (Figure 3.1a) [Larson and Chase, 1972; Winterer et al., 1974; Winterer, 1976a, 1976b]. The southernmost identified magnetic-anomaly lineations become younger from the Nauru Basin (M11)
[Larson, 1976; Nakanishi et al., 1992] eastward to the spreading segment between the Central Pacific and Phoenix fracture zones, where spreading ceased just after M1 time as evidenced by conjugate anomalies M1 through M3 [Nakanishi et al., 1992] (Plate I). It is possible that spreading ceased on all of these western Phoenix-Pacific spreading axes by M1 time, although the absence of isochrons as young as M1 on the most westerly spreading segments (Plate I) suggests that spreading ceased progressively eastwards. I concur with Engebretson et al. [1991] that accretion continued on the easternmost segment of the Pacific-Phoenix spreading axis until about M0 time. A spreading center, either abandoned shortly before M0 time or decreasing in spreading rate to zero just after M0 time, would account for the absence of an M0 magnetic anomaly in this region [Cande et al., 1989; Nakanishi et al., 1992; Yamazaki and Tanahashi, 1992]. I postulate that as the triple junction moved southeastward, the Pacific-Phoenix spreading axis lengthened to include the westernmost region of the Nova-Canton Trough (the region of >7 km depths). The ~N80°E-striking linear bathymetric depressions lying between the Phoenix Islands and 180° (Plate I), which also have some expression in the gravity data (Figures 1.1, 1.8, arrow), may be associated with the abandoned Pacific-Phoenix spreading axis further west along this segment. When projected about the best-fit Euler pole at 34.4°S, 150.6°W (compare Figure 1.1 to Figure 1.8, arrows), it is obvious that these depressions, if they are indeed related to spreading, were not formed about the same pole as the Nova-Canton Trough or the western fracture-zone extensions of the entire central Pacific. A chain of small seamounts stretches symmetrically away from the inferred remnant spreading axis (Plate I, Figure 2.6).

I propose that formation of the Manihiki Plateau coincided with, or immediately preceded, cessation of spreading on the easternmost Pacific-Phoenix spreading axis. This event may have initiated the shutdown of the M-sequence spreading system in the central Pacific basin. An 40Ar-39Ar incremental heating plateau age of basalt recovered at Site
317 (Plate I) on the Manihiki Plateau indicates that it formed at 123 ± 1.5 Ma [Mahoney et al., 1993, citing R. A. Duncan], immediately prior to M0 time, now dated by Mahoney et al. [1993] at 122.4 ± 0.8 Ma based on Ontong Java Plateau basalts and overlying microfossils [Kroenke, Berger et al., 1991]. The sharp relief bordering the southeast and northeast margins of the Manihiki Plateau and the orthogonal trends defined by ridge-trough structures within the Plateau suggest a more complex origin and subsequent evolution than that generated by a surfacing mantle plume head, a hypothesis that has been applied to other oceanic plateaus and continental flood-basalt provinces [Richards et al., 1989; Duncan and Richards, 1990; Duncan and Richards, 1991; Mahoney and Spencer, 1991; Tarduno et al., 1991]. I suggest that upon the shutdown of the ~M0 Pacific-Phoenix spreading ridge, the triple junction jumped ~500 km south to the nascent Manihiki Plateau. The tilted fault-block morphology of the northeast margin of the Manihiki Plateau [e.g., Hill et al., 1991] strikes N140°E. The similarity in strike of this boundary to both the abyssal hill fabric south of the Nova-Canton Trough and the elongate ridge at 167°W, 2°S suggests a related origin [Tanahashi, 1992]. I postulate that the Pacific-Farallon spreading axis south of the Clipperton (Nova-Canton) transform jumped westwards, rifiting the northeast margin of the Plateau and forming the oceanic crust south of the Nova-Canton Trough (Figure 3.1b). This rift extended into the abandoned Pacific-Phoenix spreading axis, transforming the eastern end into a transform fault, and creating the anomalous depths observed in the western deep of the Nova-Canton Trough. I have insufficient bathymetric data east of the PacRim and R/V Hakurei-Maru cruises to determine the eastern extent of the oblique spreading fabric, but assume that the ridge reoriented later to produce spreading fabric orthogonal to the Clipperton Fracture Zone.
Figure 3.1: (a) Tectonic reconstruction of the Central Pacific Basin showing the ridge-ridge-ridge triple junction active at M0 time. The ~M0 spreading axis between the Pacific and Phoenix plates lies in the region between the lineated bathymetric depressions near 180°W and the westernmost deep of the Nova-Canton Trough. Further west, spreading ceased between M1 and M4 time. The Manihiki Plateau is in the process of forming, but has not yet undergone any structural modification. (b) Tectonic reconstruction of the Central Pacific Basin during the early part of the Cretaceous Normal Superchron. The triple junction moved through the Manihiki Plateau [see Winterer et al., 1974, Figures 13 and 14], and jumped south of the Plateau trapping Phoenix plate on the Pacific plate. A rift bordering the northeastern boundary of the Manihiki Plateau extended into the western deep of the Nova-Canton Trough and initiated spreading between the nascent Nova-Canton (Clipperton) and Galapagos transform faults. Similarly, spreading between the Galapagos and Marquesas transform faults was initiated by a rift propagating along the southeastern boundary of the Manihiki Plateau [Sharman and Mammerickx, 1990; Hill et al., 1991; Munch et al., 1992].
The recent identification of ENE-trending conjugate M22 through M29 magnetic-anomaly lineations in the Samoan Basin south of the Manihiki Plateau [Engebretson et al., 1991] suggests that a large portion of the Phoenix plate was annexed to the Pacific plate when the spreading center jumped south (Figure 3.1b). The formation of the Manihiki Plateau and Robbie Ridge probably overprinted many of the younger conjugate Phoenix magnetic lineations.

The mode of formation of the Manihiki Plateau is a problem that remains unsolved, but is presently receiving a great deal of scrutiny, together with other oceanic plateaus. Mahoney [1987] and Mahoney and Spencer [1991] find that, although Pb, Nd, and Sr isotopic ratios are within the range of oceanic island basalts, samples from DSDP Site 317 (Plate I) [Jackson et al., 1976] also possess mid-ocean ridge basalt affinities, suggesting that the basaltic flood event responsible for the formation of the Manihiki Plateau may have been generated by the initiation of a mantle plume in a near ridge-crest setting [Mahoney and Spencer, 1991]. Various theories advanced to explain the origin of the Shatsky Rise in the northwest Pacific also invoke the interaction of triple-junction movements [Sager et al., 1988; Sharman and Risch, 1988] and hotspots [Nakanishi et al., 1989]. Indeed, the confluence of mantle plume initiation, ridge-crest processes, and concomitant rifting seems necessary to explain both the thickened oceanic crust, and internal, as well as external, structural expression of the Manihiki Plateau. I cannot confidently suggest any tectonic configurations for the intervening period between the shutdown of the Pacific-Phoenix spreading system (Figure 3.1a) and the initiation of the Pacific-Antarctic-Farallon triple junction to the south (Figure 3.1b) [Winterer et al., 1974], however, the Pacific-Phoenix-Farallon triple junction may have jumped south in discrete steps through the Manihiki Plateau, trapping more portions of the Phoenix plate [Winterer et al., 1974, Figures 13 and 14], while the Galapagos and Marquesas fracture zones organized in the wake of the southward movement of the Pacific-Farallon spreading ridge. A "Phoenix microplate"
between the Pacific and Antarctic plates may be a viable alternative [see also Engebretson et al., 1991], but there is not enough information to constrain its motion.

I considered the possibility that the N140°E-striking crustal fabric south of the Nova-Canton Trough was the result of Phoenix-Farallon spreading south of a ridge-ridge-transform (R-R-F) triple junction (Figure 3.1a). This could explain its sub-normal trend with respect to the Nova-Canton transform, while keeping the Nova-Canton trend co-polar to the Pacific-Farallon pole. A R-R-F triple junction with the transform between the Pacific and Farallon plates is not stable, given the known spreading rates [Nakanishi et al., 1992]. Furthermore, the Pacific-Farallon fabric trend is not compatible with the ESE Pacific-Farallon spreading direction predicted by the vector diagram of such a triple junction (Figure 3.1a). It is oblique to both Pacific-Farallon and Phoenix-Farallon expected spreading fabric. I prefer the former interpretation, given the similar strike of the seafloor fabric to the rifted northeast Manihiki Plateau margin, and that the Plateau formed while Phoenix-Pacific spreading ceased. In addition, I cannot discount the possibility that the Nova-Canton Trough formed on the Phoenix plate and was later transferred to the Pacific plate by a ridge jump, however, this interpretation would not explain the co-polarity of the Nova-Canton Trough and ridge to the entire Cretaceous Normal Superchron system of fracture zone trends formed by Pacific-Farallon spreading (Figure 1.8).

North of the Nova-Canton Trough, many small ridge segments offset by fracture zones define the Early Cretaceous Pacific-Farallon plate boundary (Plate I). The profusion of small spreading segments probably resulted from the reorganization of this plate boundary following the transfer of the Magellan microplate [Tamaki and Larson, 1988] to the Pacific plate. I suggest that by the Late Cretaceous these had been transformed into one or two long spreading segments between the Clarion and Clipperton fracture zones.

Many of the suggestions raised in this study could be addressed by a high-resolution sidescan-sonar survey of the region between the northern Manihiki Plateau and the Nova-
Canton Trough. This region has received very little attention in the past but may hold clues regarding the precise nature of the method by which plate geometries were affected by the widespread flood basalt event which created the Manihiki and Ontong-Java plateaus. In addition, more sidescan-sonar coverage of the region east of the SYS09 survey area could clarify the eastern extent of the oblique seafloor spreading fabric, and possibly the manner in which the Cretaceous Normal Superchron Pacific-Farallon spreading axis between the Canton-Clipperton and Galapagos fracture zones reoriented to produce the fracture zone-orthogonal fabric evident east of the Line Islands. The extremely sparse bathymetric coverage of the area west of the Phoenix Islands prevents more definitive characterization of the nature of the curious lineated bathymetric depressions and their possible role in the shutdown of the Early Cretaceous plate boundary.

**Summary**

By analyzing free air gravity anomalies derived from GEOSAT altimetry data, I have shown co-polarity of the eastern Pacific fracture zones, from the Pau to Marquesas, about a pole located at 34.4°S, 150.6°W, through most of the Cretaceous Normal Superchron. This pole also fits the Nova-Canton Trough and ridge, which I conclude are the western extension of the Clipperton Fracture Zone, as Menard [1967] originally proposed. This interpretation is supported by sidescan sonar and bathymetry data which show abyssal hill topography at a high angle to the N70°E-striking Nova-Canton Trough. The crustal fabric strikes N140°E over a large area to the south of the Trough and limited data suggest a N160°E trend to the north.

Recently identified magnetic lineations better define the magnetic bight north of the Nova-Canton Trough, indicate (a probably eastward progressive) cessation of Pacific-Phoenix spreading at M3 to M0 time [Nakanishi et al., 1992], and also show that much of
the Phoenix plate was trapped west of the Manihiki Plateau when spreading initiated on the Pacific-Antarctic Ridge south of the Samoan Basin [Winterer et al., 1974; Engebretson et al., 1991]. Aptian volcanism along the Manihiki Plateau and Robbie Ridge overprinted the trapped crust during the southward jump to the Pacific-Antarctic-Farallon triple junction. This transition may have involved a series of jumps, as Winterer et al. [1974] suggested, and/or the short existence of a "Phoenix microplate" between the Pacific and Antarctic plates.

The block-faulted northeast and southeast margins of the Manihiki Plateau are rifted margins that form the western limit of Pacific-Farallon spreading between the Canton-Clipperton, Galapagos and Marquesas fracture zones. I suggest that the 7000 to 8400 m deep western part of the Nova-Canton Trough initially formed the eastern limit of the failed Pacific-Phoenix spreading axis, but later became the western limit of the Canton-Clipperton transform fault.
CHAPTER IV

Gravity Analysis of the Nova-Canton Trough Region

Introduction

The tectonic synthesis presented in the previous chapter proposes that the great depths observed in the western region of the Nova-Canton Trough are the result of a complex plate reorganization occurring early in the Cretaceous Normal Superchron. Prior to the advent of the magnetic quiet zone, a three-plate tectonic configuration is observed in the Central Pacific Basin between the Pacific, Phoenix, and Farallon plates as evidenced by the trends of the Phoenix magnetic-anomaly lineations (Plate I). However, following the Cretaceous Normal Superchron, a two-plate configuration is observed east of the Line Islands as indicated by the Clarion, Clipperton, Galapagos, and Marquesas fracture zones (Figure 1.1, Plate IV). Joseph et al. [1992] used sidescan-sonar information and free air gravity profiles derived from satellite altimetry data to show that the eastern Nova-Canton Trough is a fracture zone. However, transcurrent motion along a transform fault separating two plates would not produce the extreme depths (~8.0 km) found in the western Nova-Canton Trough. To explain this anomalous relief, Joseph et al. [1993] proposed a tectonic reconstruction which characterized the western Nova-Canton Trough initially as the eastern terminus of the ~M0 Pacific-Phoenix spreading axis. The thermal event which generated the outpouring of flood basalts comprising the Manihiki Plateau, a large region of anomalously shallow, thick oceanic crust [Winterer et al., 1974; Hussong et al., 1979] so perturbed regional tectonics that the pre-M0 Pacific-Phoenix-Farallon triple junction shut
down, forcing the Early Cretaceous triple junction to jump southward to the Manihiki Plateau. In Joseph et al.'s [1993] interpretation of Early Cretaceous Superchron tectonics, subsequent structural modification throughout the Manihiki Plateau included a rifting event originating along the northeastern margin of the plateau terminating at the Nova-Canton Trough. This rift, which eventually became the Phoenix-Farallon spreading axis, caused transform motion to begin in the vicinity of the Nova-Canton Trough. Therefore the Nova-Canton Trough is a complex structure; initially an \(-080^\circ\)-striking spreading axis, later the termination of a NW-SE-striking rift, and finally an \(-070^\circ\)-striking transform fault and fracture zone.

To test this interpretation of the tectonic evolution of the Nova-Canton Trough, I shall investigate the regional gravity field to determine if this explanation is supported by sub-surficial structure and crustal thickness. In Joseph et al.'s [1993] interpretation, the \(-6 \text{ km}\) eastern deep of the Nova-Canton Trough was formed by transcurrent motion along a transform fault and is not complex, as the western, deeper (\(-8 \text{ km}\)) trough is proposed to be. The occurrence of tectonic spreading fabric trending into, not parallel to, trough structures in the eastern region and the copolarity of the Nova-Canton Trough with other coeval fracture zones in the central Pacific, from the Pau to Marquesas, substantiates the contention that the eastern portion of the Nova-Canton Trough has always been a fracture zone, and therefore may show a sub-surficial structure indicative of transform fault origin. The genetic complexity of the western deep may leave its origin equivocal if its sub-surficial structure is neither suggestive of transform fault nor spreading axis origin.

To date, there have been no fine scale investigations of the gravity field in the central province and eastern trough of the Nova-Canton Trough region. The free air gravity field in the western Nova-Canton Trough was examined in detail by Rosendahl [1972] and Rosendahl et al. [1975] to investigate sub-surficial structure in the region of the deep western trough, however gravity data collected aboard the R/V *Mahi* is sparse by
current standards (Figure 4.1). Free-air gravity values were determined at 292 stations along 1400 km of ship track and averaged over 15 minute intervals.

The Geological Survey of Japan (GSJ) surveyed the Nova-Canton Trough region in 1981 and 1982, collecting echo-sounder bathymetry, gravity and magnetics data, and seismic reflection profiles [Yamazaki and Tanahashi, 1992; Tanahashi, 1992]. The R/V Hakurei-Maru cruises utilized a LaCoste & Romberg S-63 gravimeter; data were logged every 30 seconds, then decimated to five minute averages. Interpretation of the gravity data was limited to one profile crossing the deep western trough [Yamazaki and Tanahashi, 1992].

To examine the entire region from 166° to 169°30'W, I have compiled all free air gravity data available through the National Geophysical Data Center and included gravity values collected during the PacRim cable route survey in 1990 [Joseph et al., 1992] (Table 4.1). The PacRim survey employed a LaCoste & Romberg S-33 gravimeter; data values were logged every 30 seconds then averaged to one minute values. Positioning was obtained from a combination of Global Positioning System (18 hours/day), transit satellites, and dead reckoning.

The purpose of this study is to compile and analyze sea surface free air gravity data for the entire Nova-Canton Trough region. This analysis will first examine the change in the free air gravity field along the strike of the ridge-trough province from west to east to determine if there are characteristic differences in sub-surficial structure that suggest a different genesis of the structures along strike. This may aid in clarifying the tectonic evolution of the ridge-trough province. The region south of the ridge-trough province will be addressed in a separate section. Selected shipboard gravity profiles are forward modeled by applying various crustal geometries and density structures to corresponding bathymetric profiles, and these interpreted in terms of the genesis and possible later alteration of primary structures.
Figure 4.1: (Top) Bathymetry of the Nova-Canton Trough region projected about a pole at 34.4°S, 150.6°W. The contour interval is 200 m and ship tracks for gravity profiles presented in this analysis are superimposed on the bathymetry. (Bottom) Ship tracks are shown with location of data values (small triangles). The larger triangles denote profiles analyzed in this study. The prefixes used for the various cruises follow the L-DEO naming convention: c1006: R/V Conrad; dmm24, dmm28: R/V Dmitri Mendeleyev; gh8a1, gh8b1, gh82a, gh82b: R/V Hakurei-Maru (GSJ); kk090: R/V Kana Keoki; mah02: R/V Mahi; mw902: R/V Moana Wave (PacRim cable route survey).
Table 4.1. Gravity and Bathymetric Data Available in the Nova-Canton Trough Region.

<table>
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* Eotvos correction
† Tares
‡ Drift
Crossover Error Analysis

The gravity data were first examined for crossover errors between cruises. Significant errors were expected due to the span of years over which the gravity values were collected (implying varying accuracy of positioning), the various gravimeters employed in the surveys, differing theoretical gravity formulae and reference systems used to reduce the gravity values, and corrections applied (Table 4.1). I computed the error and standard deviation for each crossing point (Figure 4.2a), then edited the gravity data to remove turns and outliers [Wessel, 1989; Wessel and Watts, 1988]. Cruise gh81b [Nakao, 1986] was chosen as the standard due to the high number of internal crossovers and low values of average internal error and standard deviation. The average errors between gravity values collected during cruise gh81b and all other cruises were then added to other cruise gravity values, reducing the majority of crossover errors to within ±5 mGal (Figure 4.2b). The crossover correction procedure greatly reduced obvious offsets, although the paucity of data across the deep western trough necessitated hand-contouring the gravity values to produce Figure 4.3.

Gravity Analysis of the Ridge-Trough Province

Several profiles were selected based on the ship traverse trend across the N70°E-striking Nova-Canton Trough (Figure 4.1); the bathymetric and gravity values were projected perpendicular to a line drawn through the deep western trough to the edge of the study area. The projected profiles (Figures 2.3, 4.4) show the variation in bathymetry and the free air gravity field along the trough's strike from west to east. Maximum negative values of -130 mGal occur over the deepest bathymetric expression of the western trough (Figures 2.3, 4.4 - profiles c1006 and gh82a.6). These strong negative values persist until
Figure 4.2: (a) Histogram showing the number versus magnitude of all crossing point errors in the compiled gravity data set prior to the crossover error correction procedure. (b) Histogram after the cruise gravity data had been edited to remove turns and outliers and corrected. Note that the majority of crossing point errors now lie within ±5 mGal.
Initial Crossover Errors

- # Cross-overs = 910
- Mean = -4.2515
- Standard deviation = 11.911

Errors After Analysis

- # Cross-overs = 817
- Mean = 0.1394
- Standard deviation = 6.5899
the western trough begins to shoal at ~168°25'W longitude (Figures 2.3, 4.4 - profile gh82a.3), however negative values remain until the western trough is replaced by the central ridge at 167°W (Figures 2.3, 4.4 - profile mah02.3). The ridge bordering the southern side of the western trough attains a maximum positive value of ~78 mGal (Figures 2.3, 4.4 - profile gh82a.6), producing a maximum ridge-to-trough free air gravity anomaly of over 200 mGal. The central trough begins to show expression in the gravity profiles at 168°W (Figures 2.3, 4.4 - profile gh82a.5) although gravity values remain positive until the central trough deepens to 5900 m at 166°50'W and never exceed -50 mGal (Figures 2.3, 4.4 - profile mw902.4). The eastern trough begins at ~166°35'W (Figures 2.3, 4.4 - profile mw902.2), reaches a maximum depth of ~6200 m (Figures 2.3, 4.4 - profile gh82b.1) with an accompanying negative gravity value of -68 mGal at its widest point. By this point, the central trough has lost most of its bathymetric definition and all expression in the gravity field. Persistent positive gravity values of 10 to 20 mGal dominate the region south of the Nova-Canton Trough east of 167°W.

Although gravity data density has increased significantly since the analyses of Rosendahl [1972] and Rosendahl et al. [1975], the data distribution will still not allow 3-D modeling because the along-strike wavelength in the free air anomaly values is, in most cases, less than the along-strike data density. 2-D gravity models were constructed from 12 cruise tracks projected perpendicular to the ridge-trough topography (Figure 4.1). Endpoint depths for each profile were extended 500 km beyond the center point to avoid edge effects and then sampled at 0.5 km intervals.

The Grav2D program for the Macintosh® [Hurst, 1989] was used to calculate gravity models for various crustal geometries. The Grav2D program applies a line integral calculation to n-sided polygon bodies following the method Talwani et al. [1959] (see Appendix).
Figure 4.3: Mercator projection of the free air gravity field in the Nova-Canton Trough region contoured at the 10 mGal level from the corrected cruise track data.
The similarity of the free air gravity profiles (Figure 2.3) to the corresponding bathymetric profiles (Figure 4.4) across the ridge-trough province suggests that the water/crust density interface is the major contributor to the free air gravity field. The combined effect of greater depth and smaller density difference of the crust/mantle interface produces a contributing, though smaller, influence. The differences between the observed free air gravity profiles and the gravitational attraction produced by the seafloor topography alone is due to variations in relief of the crust/mantle interface or other lateral density changes.

To determine the topographic component of the observed gravity, I modeled the bathymetric profiles with a flat Moho at 12 km depth. Varying this singular depth to Moho will simply produce a constant offset, therefore choosing a 12 km depth is arbitrary, but within reasonable limits, given an average 5.5 km seafloor depth and ~6 km crustal thickness. I included sediment bodies delineated from seismic reflection profiles in these crustal models, using densities of 1.03, 2.70, 2.00, and 3.30 g/cm³ for the water, crust, sediments, and mantle respectively [Carlson and Raskin, 1984]. Modeling the Moho as a flat interface will produce variations in gravitational attraction due to the water/crust, water/sediment, and sediment/basement boundaries alone because these are the only changing density interfaces. The difference between the observed gravity and gravity calculated from this crustal geometry is the simple Bouguer anomaly, which shows the residual attraction of the crust/mantle interface alone unless intracrustal density variations are present. I then modeled the crust as a constant thickness, constant density body 6 km thick and calculated the gravity due to this crustal geometry. The deviation of the modeled gravity from the observed profile is the mantle Bouguer anomaly, showing departures in the Moho relief from a constant density, constant crustal thickness configuration. By comparing the simple Bouguer and the mantle Bouguer residuals, the range of Moho depths from point to point and relative amplitude of the crust/mantle interface for a single
Figure 4.4: Projected gravity values corresponding to the bathymetric profiles shown in Figure 2.3. The small triangles indicate the positions of projected gravity values. Note the high degree of similarity between the gravity profiles and their corresponding topographic profiles, indicating that a large amount of the gravity signal is due to the density interface at the seafloor.
density crust is further constrained. Positive residuals indicate a mass excess, while negative residuals indicate mass deficits. The residuals may be due to lateral variations in crustal thickness or may indicate the presence of a heterogeneous density structure within the crust or the mantle, although in the following calculations a homogeneous mantle density structure is assumed. In the interests of simplicity, a single density (2.70 g/cm$^3$) was applied to the crust, rather than delineating separate seismic layer 2 and layer 3 densities. This simplification is probably the largest source of error in the modeled profiles, however the absence of any sub-surficial crustal parameters for this region render overly complex modeling too ambitious. No seismic refraction or other high resolution reflection surveys have been conducted in this area to define crustal thickness or velocities, therefore an extremely simplistic analysis must be initially applied. Neither a flat Moho nor constant density, 6 km thick crust is expected to replicate reality but by comparing the amplitude of the simple Bouguer and mantle Bouguer residuals, the depth of the density interface at the Moho can be approximated if a homogeneous crustal density exists in this area. If the simple (single crustal density) modeled gravity profiles show large residuals, this may indicate that the crust is heterogeneous. Indeed, a region as topographically severe as the Nova-Canton Trough would be expected to possess a heterogeneous density structure.

The Nova-Canton Trough does not exhibit local isostatic equilibrium. Airy isostatic compensation of seafloor features describes a crustal geometry in which every seafloor relief change corresponds to a relief change at the Moho proportional to the difference in density at the water/crust and crust/mantle interfaces. This departure from local isostatic equilibrium is not unexpected because topography across fracture zones is hypothesized to be uncompensated for wavelengths of 50 km or less [Mulder and Collette, 1984]. Given a 1.67 g/cm$^3$ density difference at the seafloor and a 0.6 g/cm$^3$ density change at the Moho, the ratio of proportionality is 2.78. Thus, for every meter of seafloor relief change, there
will be a corresponding relief change at the Moho of 2.78 meters in the opposite sense from some pre-defined reference depth. Deep regions will be compensated by thinner crust and elevated regions will have crustal roots. To check for Airy isostatic compensation of seafloor relief, I tested profile cl006 (Figure 4.5). Using a 6 km thick crust north and south of the ridge-trough province and an average seafloor depth of 5.7 km, it is clear that the depth of the trough cannot be isostatically compensated from a reference Moho depth of 11.7 km (average seafloor depth + 6 km thick crust). This configuration (Figure 4.5a) shows crust protruding through the seafloor in the trough axis. Employing a different Moho reference depth in order to more reasonably fit the trough would correspondingly create an unreasonably thick crust under the southern ridge and away from the ridge-trough region (Figure 4.5b). A density difference at the Moho for mantle to a seismic layer 3 density (2.90 g/cm$^3$) would give a ratio of 4.17, and predict a more unrealistic crustal structure.

Although there are no universally prototypical models of spreading ridge or fracture zone gravity signatures, three possible variations in crustal thickness and density are presented which reproduce the observed gravity signal and which may be construed, given prior evidence, to simulate various tectonic environments: a variable thickness/constant density model, and two variable thickness/variable density models. The variable density models employ 1) a low density body (2.50 g/cm$^3$) in the trough axis to simulate the fractured then later serpentinized crust found in fracture zone troughs [e.g., Christensen and Salisbury, 1975; Robb and Kane, 1975; Bonatti, 1978; Louden et al., 1986; Potts et al., 1986a] and 2) a higher density body (2.90 g/cm$^3$) in the trough axis to simulate a gabbroic, frozen magma chamber [Hall et al., 1986]. These two density values also produce a $\pm 0.2$ g/cm$^3$ symmetry about the average crustal density value of 2.70 g/cm$^3$.

This analysis addresses the progressive change in the gravity field from west to east across the ridge-trough province. Four profiles crossing each of the three major sub-
Figure 4.5: An Airy isostatic compensation approximation applied to profile c1006. (a) Given a 1.67 g/cm$^3$ density difference at the Moho and a 0.6 g/cm$^3$ density difference at the seafloor, the ratio of proportionality is 2.78. Therefore, for every meter of relief change at the seafloor, there will be a corresponding 2.78 m change in relief at the Moho from the reference depth of 11.7 km in the opposite direction. As shown, local isostatic compensation in the Airy sense is impossible using the reference depth of 11.7 km because the depth of the trough rises above the seafloor depth. (b) Using a deeper reference (11.3 km crustal thickness) depth to accommodate the trough would result in unrealistic crustal thickening beneath the southern ridge (~21 km) as well as an overly thick crust away from the ridge-trough region.
Reference thickness = 6.0 km, for 5.7 km average seafloor depth

Reference thickness = 11.3 km, for 5.7 km average seafloor depth
provinces of the ridge-trough morphology are examined and modeled. Sediment thicknesses were measured from seismic reflection profiles, where available. Seismic reflection profiles for lines c1006 and dmm28.1 were not available, however, sediment thicknesses measured from adjacent GSJ survey lines can be reasonably extrapolated [Tanahashi, 1992]. Employing a sediment density of 2.0 g/cm$^3$ produces a gravity signature of $\sim$3 mGal for a 100 m thickness, and 7 mGal for a 200 m thick section. The calculated error for the GSJ, R/V Mahi, and PacRim gravity profiles is $\pm$5 mGal. The expected error information for the R/V Dmitri Mendeleyev and R/V Conrad profiles are not available, but may be expected to be at least $\pm$5 mGal or higher. Therefore I shall ignore sediments in the following models if the section is less than 200 m thick.

Western Trough

Profiles gh8a1, c1006, gh82b.6, and dmm28.1 (Figure 4.1) cross the deep western trough from its westernmost extent (gh8a1) to the point where it begins to shoal, traversing almost 100 km from west to east. Within the trough, disturbed, semi-opaque sediments vary in thickness from $\sim$550 m in the deepest part of the trough to less than 100 m at profile gh8a1. North of the trough, turbidite layering is apparent only adjacent to profile dmm28.1, however there is a turbidite-filled basin south of the southern ridge, but sediment thicknesses are only significant south of the region under scrutiny.

Profile gh8a1 lies at the westernmost edge of the study area crossing the western trough (6.5 km depth) and bordering southern ridge (3.2 km depth). Sediment thickness is negligible in this region and will be ignored in the following gravity models. The free air gravity profile (Figure 4.6a) shows a maximum ridge-trough difference of 135 mGal. By removing the effect of the topography, it appears that a mass deficit is present in the trough and a short wavelength mass excess remains over the ridge. Immediately north of the trough, a mass excess of $\sim$20 mGal remains after the topographic contribution is removed.
Figure 4.6: (a) Gravity model for the westernmost profile, gh8a1. The residual anomaly after accounting for the topography is shown above the observed and modeled gravity profiles. The topography does not account for the observed gravity over the trough, nor over a small seamount south of the ridge-trough province, but otherwise approximates the gravity field to within 10 mGal. (b) Modeled gravity profile due to a constant 6 km thick crust of constant density (2.70 g/cm³). The mantle Bouguer anomaly (observed gravity values minus model values) is shown above. Note that the large mass deficit across the trough is greatly reduced by the changing density interface at the Moho and the slightly increased depth to Moho.

NOTE: For these profile, and all subsequent gravity profiles examined in this analysis, the absolute values of the free air gravity field are not shown. Refer to Figure 4.4 for absolute gravity values. The free air anomaly values for all ridge-trough province profiles are in 25 mGal divisions, the residual profiles are in 10 mGal divisions. The observed values in the free air anomaly profile are small circles, the modeled profile is a line. Each gravity model shows the crustal geometry and densities used for each body to produce the free air gravity model. The seafloor topography is taken from the profiles in Figure 2.3, however the Grav2D program [Hurst, 1999] allows only 30 vertices per body, therefore the topography has been somewhat smoothed, and in some cases several bodies with the same density are joined to produce a single layer.
South of the bordering ridge, the residual gravity signal oscillates within ~20 mGal of zero. The amplitude of the mantle Bouguer anomaly (Figure 4.6b) over the trough and southern ridge shows that the 6 km crustal thickness comes very close to modeling the observed profile besides a 10 mGal mass deficit at the southern edge of the trough. The major deviations in the mantle Bouguer anomaly occur north and south of the trough. The 6 km thick crust under the basin 30 km north of the trough produces a positive residual of ~28 mGal indicating that a mass excess is present in this region, and a slightly smaller mass excess remains in the region south of the trough. By varying the depth to the crust/mantle interface (Figure 4.7a), a slight amount of crustal thickening under the trough and upwarping of the Moho under the bordering ridge will reproduce the observed free air anomaly. A low density ($\rho = 2.50 \text{ g/cm}^3$) body in the trough axis (Figure 4.7b) requires slightly less thickening, but a higher density ($\rho = 2.90 \text{ g/cm}^3$) body in the trough axis needs further crustal thickening (Figure 4.7c). All variable crustal thickness models require a slightly shallower, upwarped Moho north and south of the ridge-trough region to replicate the observed gravity signal, producing a symmetric Moho structure about the ridge-trough feature.

Profile c1006 crosses the western trough at its maximum depth of 8.0 km. The adjacent ridge, at 2.2 km depth, produces ridge-trough relief of 5.8 km over a 16 km distance (>30° slope). Although I do not have access to seismic reflection profiles for cruise c1006, the seismic reflection profiles for the adjacent line, gh82a.6, have been published [Tanahashi, 1992] and sediment thicknesses may be extrapolated to this profile. Profile gh82a.6 shows a 90 m thick transparent layer and a 550 m thick semi-opaque layer covering the trough floor [Tanahashi, 1992]. I have modeled the sediments in the trough as a ~600 m thick triangular-shaped body with a density of 2.0 g/cm³. The simple Bouguer anomaly (Figure 4.8a) shows a large (~40 mGal) mass deficit in the trough axis and a slight (~12 mGal) deficit immediately south of the bordering ridge. The increased
Figure 4.7: Three models with varying crustal thickness or lateral density structure that will reproduce the observed gravity profile. (a) Variable thickness, single density crust; (b) variable thickness, variable density crust, using a lower density body in the trough axis; and (c) variable thickness, variable density crust using a higher density gabbroic body in the trough axis. In all three models, crustal thickening is necessary in the trough axis, less so when employing a low density body. All three models also require uplift of the Moho under the southern ridge and slightly thinner crust immediately adjacent to the ridge-trough province. Note the symmetry of Moho relief north and south of the ridge-trough province.
crustal thickness below the trough and corresponding decrease under the southern ridge provided by the constant crustal thickness model (Figure 4.8b) generally accounts for the deviations, leaving small residual mass deficits under the ridge and northern edge of the trough. By varying the crust/mantle relief (Figure 4.9a), or introducing variable density bodies below the trough (Figure 4.9b, 4.9c), thus creating larger density differences from point to point, the width of the negative anomaly in the trough is reproduced. In the three variable thickness models, the regions north and south of the trough are symmetric.

Profile gh82a.6 lies immediately east of profile c1006, and besides being limited in extent which introduces high residual mass excesses adjacent to the ridge-trough topography, the ridge to trough relief and free air anomaly amplitude is similar. The simple Bouguer anomaly (Figure 4.10a) shows that the topography and sediments do not account for the entire gravity anomaly; a 28 mGal residual remains in the trough after the effects of the water/crust and crust/sediment interfaces are accounted for. The constant crustal thickness approximation resembles the observed free air profile (Figure 4.10b), however there is still a 10 mGal mass deficit in the trough, slightly offset to the south of the trough axis. This skewness may indicate a failure of the 2D approximation because the gravity field is affected by adjacent structure, or could suggest that the crust is thicker on the southern edge of the trough as shown in the variable thickness model (Figure 4.11a). The variable density model using a low density body in the trough axis (Figure 4.11b) must be extended deeper on the southern edge of the trough to reproduce the observed free air gravity profile whereas a high density body (Figure 4.11c) is relatively symmetric but extends even deeper into the mantle. Profile gh82a.6 is too limited in horizontal extent to address the regions north and south of the ridge-trough structures.

Profile dnm28.1 crosses the western trough at a point where it begins to narrow and shoal. The trough has shoaled 500 m and the southern ridge has decreased in height to 3.6 km below sea level. A parallel ridge 4.7 km below sea level borders the trough to the
Figure 4.8: cl066. (a) Free air gravity due to seafloor topography and sediment fill within the trough. The only major residual is the mass deficit across the trough. (b) Free air gravity due to a constant thickness crust of 2.70 g/cm$^3$ density. The only significant residual remaining is due to the failure of the model to approximate the width of the free air anomaly across the trough. A small residual remains over the southern ridge.
a. simple Bouguer anomaly

b. free air anomaly

c. mantle Bouguer anomaly

d. free air anomaly

depth (km)

1.03 2.0 2.70

-80 -40 0 40 80

distance (km)

-50.0 -0.0 50.0

mGal

-8.0 -4.0 0.0 4.0

-16.0 -12.0 -8.0 -4.0

114

c1006
Figure 4.9: Variable thickness model. Crustal thickening in the trough axis is necessary to reproduce the observed gravity, as well as extreme thinning on the southern edge of the trough. (b) Introducing a lower density body or (c) a higher density body in the trough axis reduces the need for extreme thinning of the lithosphere on the southern boundary of the trough. Note the symmetry of the Moho relief adjacent to the ridge-trough province.
Figure 4.10: gh82a.6. (a) Free air gravity model for the seafloor density interface and sediments. A 27 mGal mass deficit remains across the trough, and a small mass excess across the bordering ridge. (b) Free air gravity model for a constant density 6 km thick crust. Note that the magnitude of the residuals across the ridge-trough structure is greatly reduced by the changing density interface at the Moho. This profile was modeled by Yamazaki and Tanahashi [1992] with a 5 km thick crust with a density of 2.67 g/cm³ with similar results. The remaining residuals north and south of the ridge-trough province are due to the lack of constraining bathymetric data.
3.30 a

50.0 mantle Bouguer anomaly

-50.0

mGal

-80

free air anomaly

mGal

-80

-40

0

40

80

distance (km)

-4.0

1.03

2.0

2.70

-8.0

depth (km)

-12.0

-16.0

mantle Bouguer anomaly

-50.0

mGal

-80

-40

0

40

80

distance (km)

-4.0

1.03

2.0

2.70

-8.0

depth (km)

-12.0

-16.0

free air anomaly

gh82a.6
north. The amplitude of the free air gravity ridge to trough relief has decreased to ~100 mGal, reflecting the gentler topography. Seismic reflection profiles are not available for this profile, however a GSJ ship track crosses the western trough about 5 km east of profile dmm28.1 (Figure 4.1) and the sediment thicknesses shown on the seismic reflection profiles for line gh82a.4 may be reasonably extrapolated. Removing the topographic component of the gravity signal leaves a large mass deficit across the ridge-trough feature (Figure 4.12a) and shorter wavelength mass discrepancies in the region north and south of the trough. The constant crustal thickness approximation increases the amplitude of the residuals (Figure 4.12b). By varying the crust/mantle relief (Figure 4.13a) or the lateral density structure (Figures 4.13b, 4.13c), the ridge-trough amplitude can be modeled, however the Moho relief in these models is extreme.

Central Trough and Ridge

Four profiles cross the central trough and ridge: mw902.5, mw902.4, gh8b1, and mw902.3 (Figure 4.1). Seismic reflection profiles show turbidites covering the plain north of the central ridge to an average depth of 550 m under a 60 m layer of transparent sediments. South of the central trough, a 60 m thickness of transparent pelagic sediments drape the abyssal hills. A variable thickness layer of semi-opaque sediments fill the central trough, increasing from ~400 m at line mw902.3 to ~700 m at line mw902.5.

Profile mw902.5 traverses the central trough at a depth of 5.8 km, and crosses the eastern peak of Dolmah Seamount at 2.8 km depth and the central ridge at 3.4 km depth (Figure 2.3). Over 600 m of semi-opaque sediments fill the central trough, and the northern plain exhibits a 550 m turbidite thickness under a 40 m layer of transparent sediments. Two free air gravity highs of 72 mGal over the seamount and 45 mGal over the central ridge surround the minimum value of -32 mGal across the trough (Figure 4.4). The major contribution to the gravity field is the topography, as shown by the small anomalies
Figure 4.11: gh82a.6. (a) Model for a variable thickness crust with a constant density. Increased depth to Moho under the trough axis and an uplifted Moho under the ridge is required to reproduce the magnitude of the observed gravity values. Note that very thin crust is necessary to replicate the observed profile away from the ridge-trough province, however this is an artifact due to the lack of topographic data away from the central region. (b) Model for a lower density, serpentinite body in the trough axis, which requires less penetration onto the mantle, and (c) for a higher density gabbroic body in the trough axis which requires more depth to Moho.
Figure 4.12 dmm28.1. (a) Modeled gravity for the topographic and sedimentary contribution to the free air gravity field. Large residuals remain across the ridge-trough province and the northern plain, however the southern plain is fit well. (b) Gravity due to a constant thickness, constant density crust. The residuals are increased by the variable Moho relief across the ridge-trough province (compare to Figure 4.12a).
Figure 4.13: dmm28.1. Models for various crustal geometries and densities. (a) Extreme thinning under the southern ridge and northern plain, and thickening under the trough are required to model the observed values. (b') A lower density serpentinite body in the trough axis shows less penetration into the mantle than the variable thickness, constant density model, but (c) a higher density gabbroic body which requires 13 km penetration into the mantle. All three models have similar Moho relief away from the ridge-trough province, with uplift under the southern ridge, northern plain, and on the southern side of the ridge. The crustal geometries are very extreme and I question the capability of these simple models to reproduce the structure along this transect. The northern bordering ridge shows no expression in the free air gravity field.
remaining after removing the topographic signature (Figure 4.14a). Small mass residuals remain in the trough and across the peak of Dolmah Seamount. The introduction of a Moho relief change in the constant crustal thickness model (Figure 4.14b) corrects these mass residuals, but introduces larger residuals across the central ridge and northern basin. By varying the depth to Moho across the traverse (Figure 4.15a), decreasing Moho depth under the central ridge, and increasing the depth to Moho under the northern plain, these residuals can be removed. The shape of the crust-mantle interface is slightly changed under the central trough and seamount, although the amplitude is the same. Introducing a serpentinite body in the trough (Figure 4.15b) and a slender gabbroic body to simulate a volcanic conduit [Wedgeworth and Kellogg, 1987; Kellogg et al., 1987] eliminates the need for the extreme Moho upwarping under the seamount, although a thinned crust is necessary at the southern edge of the trough. A higher density gabbro body beneath the trough axis requires a slightly deeper Moho (Figure 4.15c). The limited horizontal extent of this profile renders it an unreliable model for crustal structure away from the central province, and the adjacent high amplitude gravity signal due to Dolmah Seamount brings the modeled values in the trough into question.

Profile mw902.4 crosses Dolmah Seamount on its lower flanks at a depth of 4.7 km (Figure 2.3) and the corresponding free air gravity value has decreased to 34 mGal (Figure 4.4). The central ridge has deepened slightly to 4.0 km and the free air gravity value decreased to 38 mGal. The central trough is 5.9 km below sea level, filled with ~600 m of semi-opaque sediments and the free air gravity value has decreased to ~40 mGal. The contribution of the water/crust interface accounts for most of the observed free air gravity signature (Figure 4.16a), however a small mass deficit remains over the central trough, and a large mass excess of ~40 mGal is evident over the seamount flanks. The 6 km constant crustal thickness model (Figure 4.16b) almost corrects the central ridge to trough gravity
Figure 4.14: mw902.5. (a) The westernmost profile examined crossing both the central ridge and trough also crosses Dolmah Seamount near its crest. The topography accounts for most of the observed profile with small residuals across the central trough and the seamount. (b) Gravity due to a 6 km thick crust with a density of 2.70 g/cm$^3$. The changing Moho relief causes misfit across the central ridge and northern plain, however the trough is fit by this geometry.
Figure 4.15: mw902.5. (a) The variable thickness model shows that the gravity across the seamount must be fit with an upraised Moho, implying that the seamount is not in local Airy isostatic equilibrium. The central trough requires slight crustal thickening, and the central ridge must be fit with a slightly upraised Moho. (b) Variable density, variable thickness model using a serpentinite body in the trough axis to simulate the fractured and later hydrothermally altered crust found in fracture zone troughs. Note that the inclusion of a high density (2.90 g/cm$^3$) body simulating a volcanic conduit in the seamount obviates the need for the upwarping shown in Figure 4.15a, however some crustal thinning is still necessary at the southern edge of the trough. (c) A higher density gabbro body in the central trough axis requires a 6 km thickness.
change by creating a 3 km change in Moho relief, but the mass excesses to the north and south of the central province are not corrected by the constant crustal thickness Moho relief, which is essentially flat. The mass excesses north and south of the central province may be corrected by creating extreme lateral crust/mantle relief variations (Figure 4.17a). The variable density model (Figure 4.17b) requires less crustal thickness (~6 km) below the central trough than the variable thickness model, but creates an unreasonably thin crust south of the trough to approximate the observed gravity high there. The higher density gabbro body below the trough axis must be almost 12 km thick to reproduce the observed gravity values (Figure 4.17c). The crustal thicknesses required by both serpentinite and gabbro models is possibly overestimated due to the gravitational attraction of the seamount.

Profile gh8b1 is the only available long profile across the central province and, as such, will aid in constraining possible Moho configurations to account for the variations in the free air gravity anomaly across the central province and the crust adjacent to it. The central trough is 5.7 km deep (Figure 2.3) with a 450 m thick sediment fill, and an accompanying free air gravity value of -27 mGal (Figure 4.4). The central ridge rises to 3.5 km below sea level along this traverse, with a free air gravity high of 38 mGal. However, the most interesting aspects of this profile are the varying gravity values north and south of the central province. Free air gravity values across the northern turbidite plain reach -25 mGal before returning to zero ~60 km north of the center line where depths decrease to the regional average of 5.3 km below sea level. A 10 to 15 mGal high persists over the region immediately south of the central province. The topographic and sedimentary contribution to the gravity field agrees with the observed profile for the northern region and to within 10 mGal over the central ridge (Figure 4.18a), however a 27 to 40 mGal mass excess remains over the southern area. The constant crustal thickness approximation (Figure 4.18b) increases the amplitude of the residual values, but does not change the residual shape. In order to replicate the observed gravity profile, extreme
Figure 4.16: Contribution to the gravity field from the topography and sediments. Note the large mass excess across the flanks of the seamount. The topography alone produces much smaller gravity values. This offset is probably due to the failure of the 2D approximation. Although this profile crosses the seamount at a much lower point, the height of the seamount still affects the gravity. Slight mass residuals occur across the central trough and ridge and the northern plain. (b) Gravity model for a constant thickness, single density crust. The small amount of Moho relief does not alter the residual across the seamount’s flanks, but the mass excess across the central ridge is removed by the changing Moho relief.
simple Bouguer anomaly

free air anomaly

mantle Bouguer anomaly

free air anomaly
Figure 4.17: mw902.4. (a) Variable crustal thickness model. Note the extreme Moho upwarping necessary to reproduce the observed gravity values and the crustal thickening under the trough. (b) A serpentinite body in the trough axis reduces the crustal thickening necessary in the trough axis, but very thin crust under the seamount's flanks is still needed. (c) Mantle penetration to over 16 km is required if a high density body is used beneath the trough axis.
changes in Moho relief or lateral density structure are necessary (Figure 4.19), especially adjacent to the central trough. The variable thickness (Figure 4.19a) and variable thickness/variable density (Figure 4.19b and 4.19c) models all require thinner crust below the southern region relative to the northern plain to replicate the observed values. Broad crustal thickening is necessary across the central ridge in the three models shown in Figure 4.19, however the crustal thickness below the central trough changes in accordance with the density used, from 4.8 km thick when using a low density serpentinite body (Figure 4.19b), to 6.8 km thick for the average crustal density of 2.7 g/cm$^3$ (Figure 4.19a), to 11.6 km thick to fit a high density gabbro body (Figure 4.19c).

Along profile mw902.3, the central trough lies at a depth of 5.7 km (Figure 2.3) with a 550 m thick sediment infill. The central ridge stands at 3.6 km below sea level south of the 5.3 km deep northern plain where a thin (35 m) layer of transparent sediments overlay ~600 m of turbidite sequences. Abyssal hills cover the southern plain, but the average depth stands at about 5.2 km. The free air gravity profile shows a high of 41 mGal over the central ridge with an adjacent low of -32 mGal across the central trough (Figure 4.4). Removing the contribution of the water/crust interface and sediment bodies leaves residual mass excesses over the north and south basins and small mass deficits over the central trough and the northern flank of the ridge (Figure 4.20a). Applying a constant crustal thickness model increases the mass excess over the northern flank of the central ridge but cancels out the mass excess over the northern turbidite-filled basin (Figure 4.20b). The mass excess over the southern plain is slightly increased by the Moho relief change south from the central trough. To fit the observed gravity profile, a large mass change is necessary, either through a large relief change between the central trough and southern plain (Figure 4.21a) or a gentler crust/mantle relief, but with a lateral density change (Figures 4.21b and 4.21c). In the variable thickness/constant density model, the crustal thickness below the trough axis stands at ~5.6 km, whereas a serpentinite body in
Figure 4.18: gh8b1. (a) This profile samples the largest horizontal extent across the central province and can help constrain crustal geometries in this region and point out artifacts due to the limited extent of parallel profiles. The topography and sediment layers account for most of the observed signal across the northern turbidite plain and central trough, however the central ridge and southern plain show mass deficits and mass excesses, respectively. (b) The constant crustal thickness approximation does not remove the residuals and increases residuals in some areas. The uplift under the central ridge clearly increases the mass deficit implying a thicker crust. The northern plain is still fit by this geometry, however a large mass excess is present under the southern plain. Note the 35 to 40 mGal difference between observed values across the adjacent plains.
Figure 4.19: gh8b1. (a) The variable thickness model which reproduces the observed gravity profile shows some thickening under the central ridge and the ridges at the extreme northern edge of the profile. A severely thinned crust is necessary to reproduce observed values across the southern plain. (b) A serpentinite body in the trough axis reduces mantle penetration but increases crustal thickening under the central ridge. (c) A gabbro body in the trough axis requires unrealistic crustal thickening, however the crust proximal to the trough is very thin (< 4 km).
the trough axis requires a 4.4 km thick crust. A gabbroic body in the trough axis (Figure 4.21c), because of its higher density, requires a 7.6 km thick crust to replicate the free air gravity values. The regions north and south of the central province must be fit with significantly different crustal thicknesses, the southern area being 1 to 1.5 km thinner than the northern region, similar to the crustal geometry shown for profile gh8b1 (Figures 4.18, 4.19).

Eastern Trough

The eastern trough begins at ~166°35' W longitude and extends to the end of the study area. From west to east, profiles mw902.2, mw902.1, gh82b.2, and gh82b.1 cross the eastern trough and the western extents of the central ridge and trough (Figure 4.1). Turbidite layering in the northern basin continues uninterrupted to 0° latitude, but ends abruptly against a block faulted structure stepping up to ~4.6 m depth. Sediments in the eastern trough thicken from less than 200 m at line mw902.1 to over 400 m at profile gh82b.1 where the trough is 10 km wide. The strike of the eastern trough is slightly offset to the north of the line through the deep western trough (Figures 2.3, 4.4).

The westernmost extent of the eastern trough, crossed by profile mw902.2, is 5.1 km deep (Figure 2.3) and exhibits only a thin layer of transparent sediment infill. The simple Bouguer anomaly shows the same large mass excess across the southern plain (Figure 4.22a), and is similar in other respects to the residuals across line mw902.3. The constant crustal thickness model (Figure 4.22b), as observed on profile mw902.3, accentuates the residuals. Large lateral mass changes at the Moho are necessary to reproduce the observed free air gravity profile, including a very thin crust south of the central province. Crustal thinning is necessary at the southern edge of the central trough (Figure 4.23) and either a slightly thinned crust (~4.5 to 5 km) in the trough axis in the lower density models (Figures 4.23a and 4.23b) or thickened crust (~7.5 km) in the higher
Figure 4.20: mw902.3. (a) Gravity due to seafloor topography and sediment layers. The residuals show the general profile for this province: mass excess across the northern and southern plains, and small mass deficits across the central trough and ridge. (b) The constant density, constant crustal thickness approximation increases the amplitude of the residuals (also shown on Figure 4.25 for profile gh8b1) across the profile, except for the northern plain, but this is probably due to increased values over the central ridge.
Figure 4.21: mw902.3. (a) A variable thickness model that can reproduce the observed gravity profile requires extreme crustal thinning south of the central province, a slightly thinner than 6 km crust under the central trough, and thickening under the central ridge. (b) Introducing a low density body in the trough axis changes the gravity profile only slightly, however note the difference in crustal thickness north and south of the central province. (c) The high density gabbro body requires 7.6 km crustal thickness, but does not affect the crust/mantle interface away from the central trough.
density model (Figure 4.23c) to reproduce the observed signal. The southern region must be modeled with a 1 to 1.5 km thinner crust than that to the north of the central province.

By profile mw902.1, the eastern trough is accompanied by a gravity low of -14 mGal; free air gravity value over the central trough is slightly lower, at -20 mGal (Figure 4.4). The simple Bouguer anomaly (Figure 4.24a) indicates that mass excesses remain over the north and south regions and the central ridge, whereas the central trough is accompanied by a small mass deficit when the seafloor topography and sediments have been taken into account. The mantle Bouguer anomaly (Figure 4.24b) shows that the variation in Moho relief created by mirroring the topography 6 km below the seafloor removes the residual over the central ridge, but creates a mass deficit over the eastern trough and widens the residual across the central trough. The mass excesses over the northern and southern plains are slightly decreased, but a mass deficit is created over the small ridge immediately north of the eastern trough. In order to remove these residuals, sharp relief changes with a greatly thickened crust are necessary under the two troughs and thinner crust under the central ridge (Figure 4.25a). If serpentinite bodies are introduced beneath the two troughs (Figure 4.25b), the relief of the crust/mantle boundary can be modeled as a much gentler interface and the severe crustal thinning under the central ridge is unnecessary, however, higher densities show even greater mantle penetration (Figure 4.25c). The crustal thinning necessary south of the central province required in the preceding three profiles appears to be unnecessary on profile mw902.1 to reproduce the observed gravity values.

By profile gh82b.2, the central trough has narrowed and shoaled. The free air gravity values over the eastern trough have decreased to -25 mGal reflecting the deepening of the trough along this traverse, and the gravity values over the central ridge have likewise decreased (Figure 4.4). Residual gravity values after removing the topographic contribution (Figure 4.26a) show mass deficits across the central province and eastern
Figure 4.22: mw902.2. (a) Gravity due to seafloor topography and sediment layers. The mass excess over the central ridge changes closer to the crest of the ridge than on profile mw902.3 (Figure 4.20a) due to the eastern trough, which at this point is still a small cleft in the broader high of the ridge. Mass excesses similar to those shown on preceding profiles across the southern plain are evident.

(b) Modeled gravity due to a 6 km thick crust with a constant density. Moho upwarping under the central ridge causes a mass deficit implying that the Moho is actually deeper than the 6 km thick crust predicts. Mass excesses across the central trough and southern plain indicate either higher density bodies or a shallower Moho are necessary to reproduce the observed gravity profile.
free air anomaly

simple Bouguer anomaly

mantle Bouguer anomaly

free air anomaly

depth (km)

distance (km)
Figure 4.23: mw902.2. (a) The profile due to a variable thickness crust can reproduce the observed gravity profile if depth to Moho is reduced across the northern and southern plains, and the crust is thickened under the central ridge. (b) The addition of a serpentinite body in the central trough reduces the crustal thickness under the trough but requires a thickened crust under the central ridge and thinning at the southern edge of the trough. (c) High density gabbro bodies in the trough axes require mantle penetration beyond the depth of the surrounding crust/mantle interface.
Figure 4.24: mw902.1. (a) The gravity due to the seafloor density interface and sediment bodies shows mass residuals similar to the preceding profiles. The dual troughs across the central region produce symmetric lows about the central ridge.

(b) This approximation of the gravity field due to a constant 6 km thick crust is the first such profile across the central province that does not show increased residuals across the traverse. The amplitude of the relief across the central trough and ridge are better reproduced and the mass excess across the northern plain is reduced.
trough and the same mass excesses over the adjacent plains as in the preceding profiles.
The constant crustal thickness model (Figure 4.26b) decreases, but does not remove, the
mass deficit over the central province, and increases the mass excess over the northern and
southern regions. Figure 4.27 shows that, again, steep relief along the density interface at
the Moho is necessary in both the variable thickness and variable density models to model
the observed gravity variation along the profile.

The easternmost profile examined, gh82a.1, crosses the central trough at the point
where it has lost most of its expression in both the topography and in the gravity field
(Figures 2.3, 4.4). The eastern trough, however, has widened to over 10 km and its relief
dominates the free air gravity profile. The simple Bouguer residual indicates that the
topography and sediment bodies do not account for the entire observed anomaly; mass
residuals remain over the eastern trough and adjacent plains (Figure 4.28a). The mantle
Bouguer anomaly shows that the mass deficit across the eastern trough is reduced by
modeling increasing Moho relief beneath the trough, however a 10 mGal mass deficit
remains (Figure 4.28b). The constant density/variable thickness model (Figure 4.29a) and
the variable thickness/variable density model employing a high density gabbro body below
the trough axes (Figure 4.29c) require greatly thickened crust, however the model using a
lower density serpentinite body in the trough axes simulates crustal thickness more
reasonably (Figure 4.29b).

Interpretation of Gravity Models for the Ridge-Trough Province

The purpose in examining the gravity field along the ridge-trough strike is to
determine if the free air gravity data can aid in classifying features by drawing inferences
about changes in Moho relief or intracrustal density structure that may be peculiar to a
particular tectonic environment. Are there characteristic differences between the deep
Figure 4.25: mw902.1. (a) The variable thickness model for profile mw902.1 shows that increased crustal thickness is necessary under the two troughs to reproduce the observed values, and extremely thin crust under the central ridge, a result not favored by the preceding gravity models. (b) By employing two serpentinite bodies in each trough axis, the extreme thickening in the trough axes is not necessary, or is the upwarped Moho under the central ridge. Slight crustal thinning is necessary at the southern edge of the central trough and northern plain. (c) A gabbro body in the beneath the central trough axis requires a 8.8 km thick crust, however because of the limited horizontal extent of the eastern trough on this traverse, the gabbro body beneath the eastern trough requires less penetration beyond the surrounding Moho to reproduce the observed values.
Figure 4.26:  gh82b.2. (a) Gravity modeled for the effect of the seafloor topography and sediment layers shows a small mass excess across the northern turbidite plain, and a small mass deficit across the eastern trough. (b) By approximating a constant 6 km thick crust, the residuals are somewhat increased, especially across the northern plain.
Figure 4.27: gh82b.2. (a) Broad crustal thickening across the entire central region can reproduce the observed gravity values. (b) Serpentinite bodies in the trough axes decreases depth to Moho across the central ridge and generally produces a more realistic crust/mantle relief across the northern plain. The southern plain is not realistically addressed on this profile. (c) As shown on the preceding profiles, the high density gabbro body in the trough axes requires unrealistic thickness to reproduce the observed signal.
western trough and the central and eastern regions that can aid in identifying the genesis of the deep western trough? In addition, is there a preferred model for the ridge-trough province crustal structure determined from the preceding models?

The most obvious difference between the eastern and western regions is the amplitude of the bathymetric and free air gravity relief. In the western Nova-Canton Trough, free air gravity variations range from 135 mGal at the westernmost profile, gh8a1 (3.4 km ridge to trough relief), to \( \sim 200 \) mGal at profiles c1006 and gh82a.6 which cross the greatest elevation difference (5.5 to 5.8 km), to \( \sim 100 \) mGal at profile dmm28.1, crossing the western trough where ridge to trough relief has decreased to 4.1 km. The amplitude of the simple Bouguer residuals show that the topography and sediments account for over 70% of the observed free air gravity field in the west, except for profile dmm28.1, where the residual shows that the topography account for only half of the observed gravity. Profiles traversing the central ridge and trough in the eastern region of the Nova-Canton Trough cross \( \sim 2.2 \) to \( 2.4 \) km of bathymetric relief with accompanying free air gravity relief of 65 to 75 mGal. The topography and sediments account for over 80% of the gravity across the ridge-trough relief. Across the easternmost part of the Nova-Canton Trough, the average ridge to trough relief is 2.2 km and free air gravity relief is \( \sim 50 \) mGal. The topography accounts for over 85% of the observed signal on lines mw902.1 and gh82b.2, and 80% of the free air gravity across line gh82b.1. The largest residuals in the entire eastern region are south, and to a lesser extent, north of the ridge-trough structures.

In order to gain a better understanding of the gravity signature across a Mesozoic fracture zone, I examined free air gravity profiles across a section of the Murray Fracture Zone, at 158°W longitude (profile c1208, Figure 4.30a) and 159°W longitude (profile c1210, Figure 4.30b) that is approximately coeval to the eastern Nova-Canton Trough. The amplitude of the residuals over the ridge-trough topography is similar to that observed across the eastern Nova-Canton Trough, although the relief across the Murray Fracture
Figure 4.28: gh82b.1. (a) The easternmost profile analyzed in this study shows that the eastern trough dominates the gravity field and the central trough has very little expression left in the observed gravity profile. The central ridge has also decreased in height. The topography and sediment layer cannot account for the gravity field across the region, and mass residuals are evident across the entire region. (b) The changing Moho relief addressed by the constant crustal thickness model better approximates the values across the eastern trough, however the northern plain shows a mass excess.
50.0 simple Bouguer anomaly

-50.0 free air anomaly

-80 distance (km) -40 0 40 80

1.03

-16.0

3.30

50.0 mantle Bouguer anomaly

-50.0 free air anomaly

-80 distance (km) -40 0 40 80

1.03

-16.0

3.30
Figure 4.29: gh82b.1. (a) A variable thickness model requires an unrealistically thick crust in the eastern trough axis, however, (b) by employing a lower density body in the trough axis, less thickening is necessary, although the Moho beneath the eastern trough shown on this profile is greater than any variable density models of the central trough. (c) Although a gabbro body beneath the central trough does not require unreasonable thickening, the same density beneath the eastern trough shows Moho at almost 16 km depth.
Zone is more subdued. In addition, there is a mass residual across the region south of the fracture zone although in the opposite sense to the residual observed south of the eastern region of the Nova-Canton Trough. These residuals are not removed by approximating a constant thickness crust (Figures 4.31a, 4.31b). The right-lateral Murray Fracture Zone exhibits a depth increase of ~300 m on the southern, younger side of the fracture zone that is not accounted for by the correspondingly shallower Moho of the constant crustal thickness model. The mantle Bouguer residuals indicate a mass deficit on the southern side, implying a thicker crust or higher density crust or Moho south of the fracture zone. The mass deficit could also be due to proximity to the Musician Seamounts further south.

Western Trough

The constant crustal thickness models for the deep western trough (Figures 4.6b, 4.8b, 4.10b) show that a 6 km thick, constant density crust comes very close to fitting the observed gravity, although slight crustal thickening is required under the trough (Figures 4.7a, 4.9a, and 4.11a) if 2.70 g/cm³ density material underlies the trough axis. Using a 2.70 g/cm³ crustal density, but ignoring sediments, Yamazaki and Tanahashi [1992] conclude that profile gh82a.6 (Figure 4.1) can be reasonably fit with a homogeneous density, constant 5 km thick crust. However employing an average crustal density of 2.88 g/cm³ and Moho density of 3.44 g/cm³, Rosendahl [1972] and Rosendahl et al. [1975] concluded that a crustal root is developed under the deepest expression of the western trough with Moho ~4.5 km below the trough axis for topographic and gravity profiles collected aboard the 1970 R/V Mahi cruise (Figure 4.1, profiles mah02.5 and mah02.6). The observed free air anomaly can be also be reproduced by using a higher density body (ρ = 2.90 g/cm³) beneath the trough to simulate a frozen magma chamber remaining after the cessation of spreading. This interpretation is in agreement with the model of Madsen et al. [1990] for the East Pacific Rise based on gravity analyses and multichannel seismic
Figure 4.30: (a) Seafloor topography contribution compared to the observed gravity across the Murray Fracture Zone at 158°W longitude. The residuals are very small, less than 10 mGal, except at the extreme southern end of the profile, where a mass deficit implies a change in the crust/mantle relief. (b) Similar to (a), but crossing the Murray Fracture Zone at 159°W.
50.0 simple Bouguer anomaly

-50.0 free air anomaly

distance (km)

-80 -40 0 40 80 100

1.03

2.70

3.30
c1208

50.0 simple Bouguer anomaly

50.0

-50.0

free air anomaly

distance (km)

-80 -40 0 40 80 100

1.03

2.70

3.30
c1210

b.
Figure 4.31: (a) Constant crustal thickness model corresponding to Figure 4.30a across the Murray Fracture Zone. The residuals are barely changed by the varying Moho relief, but the mass deficit at the southern end of the profile is not removed by the subtle Moho undulations. (b) as above, but corresponding to Figure 4.30b. In this case, the residuals are somewhat increased by the constant 6 km thick crust, especially at the southern end of the profile.
reflection profiling for the horizontal extent and depth of a crestal magma chamber, and the
results of Hall et al. [1986] regarding the gravity signal due to crustal geometry and density
frozen in at the cessation of spreading at an extinct spreading center. But a low density
serpentinite body also can reproduce the observed gravity signal and requires less crustal
penetration into the mantle. The only result of the variable thickness models across the
western region that favors a spreading center origin over a transform origin is the symmetry
of the crust/mantle interface adjacent to the ridge-trough province on profiles gh8a1 and
c1006 (Figures 4.7, 4.9).

All three variable thickness models require an uplifted Moho beneath the bordering
ridge to model the observed values across the western trough. The upwarping of the
crust/mantle density interface necessary to model the gravity high across the ridge and the
flexural profile of the ridge is consistent with observations at fracture zone transverse
ridges [e.g., Abrams et al., 1988]. Petrographic evidence for the composition of the
southern ridge also does not provide any clear-cut insight into its genesis. The 1970 R/V
Mahi cruise dredged the northern slope of the ridge at ~7.1 km depth. The diverse suite of
specimens collected include tholeiitic, transitional, and alkalic basalts, and a plutonic
assemblage including coarse-grained diabases, pegmatitic gabbros, and cumulate gabbros,
all showing varying degrees of alteration, from fresh to extensively altered [Rosendahl,
1972; Rosendahl et al., 1975]. Rosendahl was convinced that this aggregation of rock
types was evidence that the original structures had been modified by subsequent tectonism
resulting in rejuvenation and overprinting of primary structures. Free air gravity profiles
cannot distinguish between original features and those later altered by secondary tectonism.
If the western trough was originally part of a spreading ridge, as has been previously
hypothesized [Winterer et al., 1974; Rosendahl et al., 1975; Winterer, 1976a; Winterer,
1976b; Mammerickx and Sandwell, 1986; Yamazaki and Tanahashi, 1992; Joseph et al.,
1993], and is suggested by the symmetry of Moho relief to the north and south of the
ridge-trough structures, then some form of subsequent alteration is necessary to produce the asymmetry of the ridge-trough province, with a high bordering ridge on one side, similar to a fracture zone transverse ridge. Although sections of slow-spreading mid-ocean ridges often exhibit asymmetrical bordering ridge heights (for example, see Hall et al., 1986, Figure 3), the difference in scale (~3 km difference between bordering sides of the Nova-Canton Trough at profile c1006 to 0.5 to 1 km for portions of the South Atlantic ridge [Hall et al., 1986]) and the necessity of an upwarped Moho to fit the gravity profiles are more consistent with a transverse ridge origin.

The tectonic synthesis proposed by Joseph et al. [1993] suggests that the shutdown of the ~M-0 Pacific-Phoenix spreading axis was caused by the relocation of the spreading system south to the emerging Manihiki Plateau. This event and concomitant riftting of the northeastern boundary of the Manihiki Plateau along a trend ending at the Nova-Canton Trough and the lithospheric stresses generated by plate motions changing from extensional to transcurrent may be responsible for the great depths observed in the western deep and the western trough's characteristics that suggest both spreading ridge and transform fault structures. While the gravity models presented here for the western trough do not, nor indeed cannot, unequivocally prove such a scenario, the symmetry of the crust/mantle interface away from the ridge-trough province surrounding a highly asymmetric ridge and trough are consistent with this tectonic history.

It is clear that the high density body (2.90 g/cm$^3$) used to model crustal structure beneath the trough axis (Figures 4.7c, 4.9c, 4.11c) is excessive. The high density, gabbroic body creates unrealistic crustal thickening, from 7.2 to 10 km, with crustal rock penetrating the mantle far beyond the surrounding crust. Although 2.90 g/cm$^3$ is probably a good approximation of crustal density beneath an extinct ridge, models of crustal structure at currently-active and abandoned ridges based on multichannel seismic reflection and refraction results including well-constrained gravity modeling indicate that the crestal
magma chamber extends to 8 to 9 km below sea level [Hall et al., 1986; Madsen et al., 1990]. The depths attained by the gabbro bodies in Figures 4.7, 4.9, 4.11 (up to 18 km below sea level) are clearly unreasonable. It is more difficult to eliminate either the low density serpentinite model (2.50 g/cm$^3$, Figures 4.7b, 4.9b, 4.11b) or the average crustal density model (2.70 g/cm$^3$, Figures 4.7a, 4.9a, 4.11a). Neither model can be eliminated due to unrealistic thickness; the serpentinite model shows sub-trough crustal thicknesses of 4.4 to 6 km, while the average crustal density model shows thicknesses ranging between 5.2 to 6.7 km.

Simple and mantle Bouguer residuals for profile dmm28.1 show that neither the topography alone (Figure 4.12a) nor a constant crustal thickness model (Figure 4.12b) can replicate the observed gravity. Profile dmm28.1 crosses the western trough at a point where a northern bordering ridge appears and where the trough begins to narrow. This profile also marks the intersection of the Nova-Canton Trough with the trend of the northeastern margin of the Manihiki Plateau and the elongate ridge at 167°W, 2°S (Figure 4.1) which Joseph et al. [1993] suggest may mark the a transition zone between the older, originally north-south spreading Pacific-Phoenix regime to the west, and the subsequent east-west spreading Pacific-Farallon regime to the east. Large amounts of crustal thickening and widening under the trough (Figure 4.13a) and a severe Moho discontinuity under the southern ridge are required to model the observed gravity. The variable density models (Figures 4.13b, 4.13c) also do not show reasonable Moho configurations. The high density gabbroic body requires 9.7 km thickness in the trough axis (Figure 4.13c), and the low density serpentinite body must also penetrate further than shown on profiles gh8a1, c1006, and gh82a.6 (Figures 4.7b, 4.9b, and 4.11b) for a similar density structure. In addition, Moho relief adjacent to the ridge-trough province is not symmetric about the trough, however the same uplifted Moho is required to model the gravity high proximal to the bordering ridge. The location of this profile in the "transition zone" and its unusual
characteristics may be due to overprinting of original structures by the tectonic reorganization proposed in Chapter 3 and by Joseph et al. [1993].

Central and Eastern Troughs

In contrast to the western trough, constant crustal thickness models applied to the central and eastern regions generally increase the amplitude of the residuals implying that Moho relief is more subdued than the topographic relief. The southern extents of profiles mw902.5 (Figure 4.14a) and mw902.4 (Figure 4.16a) are dominated by Dolmah Seamount and thus are not conclusive about Moho relief or density changes across the entire region. The failure of the 2D assumption is probably responsible for the high observed gravity values on the southern extent of profile mw902.4. The topography on profile mw902.4 across the seamount's flanks is relatively subdued compared to profile mw902.5, which crosses near the crest of the seamount. However the gravity high south of the central trough on profile mw902.4 may indicate that the seamount's relief is still affecting the gravity values, because the topography on the lower flank of the seamount is not sufficient to cause the observed gravity, whereas the topographic contribution of the seamount alone (Figure 4.14a for profile mw902.5) accounts for most of the gravity high.

In general, the relief of the density interface at the Moho in both the variable thickness and variable density models across the entire eastern area is highly asymmetric about the ridge-trough province, in contrast to the symmetry shown adjacent to the deep western trough (Figures 4.6 through 4.9). The southeastern region shows consistent 10 to 20 mGal highs, whereas the northern turbidite-filled basin exhibits gravity values in the range 0 to -15 mGal (Figure 4.3). This asymmetry about the ridge-trough region is expected at transform faults and young fracture zone extensions that have not yet thermally equilibrated, however it is surprising to find this frozen-in asymmetry at an old fracture zone, especially considering that there is no appreciable depth offset across it. However,
as shown in Figures 4.30 and 4.31 across a Cretaceous Normal Superchron portion of the Murray Fracture Zone, there is a distinct difference in the gravity values north and south of the fracture zone. The 10 mGal high over the southern region of the Nova-Canton Trough extends to \(-1^\circ 30'\)S latitude (Figure 4.3) but appears to decrease to the range 0 to 10 mGal further to the south, although values remain positive for the entire region south of the Nova-Canton Trough and east of the basin bordering the elongate ridge at 167°W. The variable thickness models for the eastern region must be fit with a 1 to 2 km thinner crust south of the ridge-trough structures to reproduce the observed signal. But is this a realistic model of the crust adjacent to a 115 m. y. old fracture zone?

There is a disturbing paucity of experiments conducted across fracture zones in the Pacific basin to aid in constraining crustal thickness, seismic velocities, and densities (Table 4.2). Seismic refraction experiments across the large offset Mendocino Fracture Zone in the northeast Pacific [Raitt, 1963] suggest that the crust south of the fracture zone is over twice as thick as crust to the north. Gravity models based on these refraction results and conversion of seismic wave velocities using Nafe and Drake [1957] velocity/density relationships infer crustal thickening under the fracture zone trough and a continued increase in total crustal thickness to the south [Dehlinger et al., 1967; Dehlinger, 1969]. The Rivera ocean seismic experiment (Project ROSE) [Tréhu and Purdy, 1984; Ewing and Meyer, 1982; Ewing and Purdy, 1982; Ouchi et al., 1982; Tréhu and Solomon, 1983; Tréhu, 1984] examined crustal structure in the vicinity of the Orozco Transform Fault, however the \(~5.4\) km crustal thickness determined by seismic refraction data does not specifically characterize the transform fault itself. Tréhu and Purdy's [1984] line 1 traverses a region sub-parallel to an extensional relay zone or intra-transform spreading center, the Mid-Orozco spreading center [Madsen et al., 1986] and does not appear to cross any of the en echelon transform segments that characterize the southern Orozco transform boundary.
### Table 4.2. Crustal Parameters for Various Fracture Zones from Wide Angle Seismic Experiments

<table>
<thead>
<tr>
<th>Fracture Zone</th>
<th>Spreading Rate (mm/yr half-rate)</th>
<th>Offset Distance (km)</th>
<th>Sense of Offset</th>
<th>Age of Offset (my)</th>
<th>Age at Sampling Line (Ma)</th>
<th>Southern Crustal Thickness (km)</th>
<th>Northern Crustal Thickness (km)</th>
<th>Trough Crustal Thickness (km)</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Pacific Ocean</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mendocino</td>
<td>55</td>
<td>1620</td>
<td>left-lateral</td>
<td>25-30</td>
<td>32 Ma S 4 Ma N (128°30'W)</td>
<td>6.6</td>
<td>2.63</td>
<td>not modeled</td>
<td>A, B, C, D</td>
</tr>
<tr>
<td>Molokai (N &amp; S)</td>
<td>KQZ*</td>
<td>460 (N)</td>
<td>left-lateral</td>
<td>KQZ</td>
<td>6.5-8.2</td>
<td>5.5-6.1</td>
<td>no info</td>
<td>E, F</td>
<td></td>
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<td><strong>Atlantic Ocean</strong></td>
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<td>Kane</td>
<td>15</td>
<td>160</td>
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<td>10</td>
<td>7.1-18.1 (across FZ)</td>
<td>5.25 (75 km S) 4.4 (45 km S) 3.9 (15 km S)</td>
<td>3.5 (17 km N) 4.5 (67 km N) 2.3 (FZ)</td>
<td>G, H</td>
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| Vema Line A            | 24                               | 320                  | left-lateral   | 26                | transverse ridge near RTI young FZ valley | --- | --- | 4.3 km (TF) | 3.5-5.0 km (RTI) | I, J | L
<p>| Vema Line B            |                                  |                      |                |                   |                          |                               |                               |                               |        |
| Tydeman-Line A         | 18                               | 110                  | right-lateral  | 6                 | 53-56 (N-young side) 6.25 (20 km S) 6.8 (40 km S) 53-71 | 4.4 (12 km N) 5.75 (30 km N) | --- | 3.0-5.0 km (FZ) | K, L, M |
| Tydeman-Line B and E1  |                                  |                      |                |                   |                          |                               |                               |                               | L, M |</p>
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<th>Fracture Zone</th>
<th>Spreading Rate (mm/yr half-rate)</th>
<th>Offset Distance (km)</th>
<th>Sense of Offset</th>
<th>Age of Offset (Ma)</th>
<th>Age at Sampling Line (Ma)</th>
<th>Southern Crustal Thickness (km)</th>
<th>Northern Crustal Thickness (km)</th>
<th>Trough Crustal Thickness (FZ or TF)</th>
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<td>9.4</td>
<td>130</td>
<td>right-lateral</td>
<td>10</td>
<td>along transform</td>
<td>---</td>
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<td>2.0-5.5 (TF)</td>
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<tr>
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<td>9.0-10.0</td>
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<td>---</td>
<td>---</td>
<td>4.5 km (near RTI)</td>
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<td>---</td>
<td>---</td>
<td>N</td>
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<tr>
<td>Oceanographer- Line C2</td>
<td>15</td>
<td>---</td>
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<td>---</td>
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<tr>
<td>Charlie Gibbs Line 10607 (OBS 6)</td>
<td>11.3</td>
<td>125</td>
<td>left-lateral</td>
<td>10.1</td>
<td>-3</td>
<td>5.64 (20 km S, eastern RTI)</td>
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<td>5.98</td>
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<tr>
<td>Charlie Gibbs Line 10617 (OBS 2)</td>
<td>5.8</td>
<td>8.0</td>
<td>---</td>
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<tr>
<td>Charlie Gibbs Line 10617 (OBS 6)</td>
<td>5.1</td>
<td>---</td>
<td>4.0-5.0 (35 km N)</td>
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<tr>
<td>Charlie Gibbs Line 10615 (OBS 2)</td>
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<td>3.48 (TF)</td>
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<tr>
<td>Charlie Gibbs Line 10615 (OBS 6)</td>
<td>along transform</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>3.01 (TF)</td>
<td>O</td>
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Table 4.2. (continued) Crustal Parameters for Various Fracture Zones from Wide Angle Seismic Experiments

<table>
<thead>
<tr>
<th>fracture zone</th>
<th>spreading rate (mm/yr half-rate)</th>
<th>offset distance (km)</th>
<th>sense of offset</th>
<th>age at sampling line (Ma)</th>
<th>southern crustal thickness (km)</th>
<th>northern crustal thickness (km)</th>
<th>trough crustal thickness (FZ or TF)</th>
<th>source</th>
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<td>13</td>
<td>12</td>
<td>left-lateral</td>
<td>4</td>
<td>141-145</td>
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<td>2.0-2.4</td>
<td>P</td>
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<tr>
<td>ESP 2</td>
<td>7.3</td>
<td></td>
<td></td>
<td></td>
<td>---</td>
<td>---</td>
<td></td>
<td>Q</td>
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<tr>
<td>Blake Spur</td>
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<tr>
<td>ESP 13</td>
<td>7.2 (under transverse ridge)</td>
<td></td>
<td></td>
<td></td>
<td>---</td>
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<td></td>
<td>Q</td>
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<tr>
<td>ESP 5a</td>
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<td>3.5 km to Moho</td>
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<td>Q</td>
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<td>Blake Spur</td>
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<td>(15 km N)</td>
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<td>Q</td>
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<td>7.75 km</td>
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<td>Q</td>
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<td>ESP 8</td>
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<td></td>
<td></td>
<td>---</td>
<td>(30 km N)</td>
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</table>


* Cretaceous Quiet Zone
Results from a multichannel seismic reflection and refraction experiment conducted west of the Hawaiian Island chain [Brocher and ten Brink, 1987; ten Brink and Brocher, 1988] suggest that seismic layer 2 is $1.0 \pm 0.5$ km thinner north of the Molokai Fracture Zone than to the south (Table 4.2). However Lindwall [1991] cautions that these results may be suspect, and inspection of subsequently acquired GLORIA data for this region [Searle et al., 1993] show that the northern end of ten Brink and Brocher's [1988] expanding spread profile 2 grazes the southernmost trace of the Molokai Fracture Zone and expanding spread profile 3 projects well into it. Lindwall comments that the low seismic velocities reported by ten Brink and Brocher [1988] and their interpretation of a thinner layer 2 may be influenced by the proximity of their refraction lines to the Molokai Fracture Zone. Results from modeling sonobuoy records [Lindwall, 1991] indicate a crustal thickness of $\sim 6.2$ km south of the fracture zone, however, he suggests that unusually low velocities recorded by sonobuoy 4 for the upper 1.5 km of seismic layer 2 may have resulted from its proximity to the Molokai Fracture Zone. Neither Lindwall [1991] nor ten Brink and Brocher [1988] specifically characterize the crustal structure within the fracture zone.

Seismic refraction and multichannel seismic reflection experiments conducted across fracture zones in the North Atlantic have generally shown low crustal velocities, the absence of seismic layer 3, 2 to 4.5 km thick crust within the fracture zone trough, and anomalous crustal thicknesses up to 30 km from the fracture zone [e. g., Detrick and Purdy, 1980; Ludwig and Rabinowitz, 1980; Sinha and Louden, 1983; Cormier et al., 1984; White et al., 1984; Louden et al., 1986; Whitmarsh and Calvert, 1986; Calvert and Whitmarsh, 1986; Potts et al., 1986a] (Table 4.2). These experiments were conducted either across the active transform domain or crossed the inactive fracture zone near the spreading ridge, thus characterizing crustal structure across the active plate boundary or its relatively young, inactive trace. Two wide-angle seismic experiments have been conducted
across older portions of Atlantic fracture zones. A reversed seismic refraction experiment employing ocean-bottom seismometers and free-floating sonobuoys was conducted at the Tydeman Fracture Zone crossing crust aged 53 Ma on the northern (younger) side of the fracture zone to crust aged 59 Ma on the older, southern side at 26°20'W longitude, 36°N latitude [Calvert and Potts, 1985; Potts et al., 1986b]. Crustal thickness measured from ray-tracing across the fracture zone indicates ~1 km thinner crust on the younger side at comparable distances from the fracture zone trough. Seismic layer 3 is present north and south of the trough but disappears beneath the trough axis where crustal thickness ranges from 3.5 to 5.0 km with velocities characteristic of seismic layer 2 (Table 4.2) [Potts et al., 1986b, Figure 11]. Multichannel seismic evidence [Mutter et al., 1984] and expanding spread profile and seismic reflection experiments conducted across and parallel to the small-offset Mesozoic Blake Spur Fracture Zone in the western Atlantic [Minshull et al., 1991] indicate that the crust adjacent to the fracture zone exhibits seismic velocities characteristic of normal oceanic crust and the presence of a 4 to 5 km thick seismic layer 3 persisting to within 10 km of the median trough, including the transverse ridge. Within the fracture zone trough, a 2 to 4 km thick high velocity gradient layer overlies a 2 to 4 km thick layer with velocities of 7.2 to 7.6 km/s. Minshull et al. [1991] conclude that partially serpentinized upper mantle peridotite may overlie seismic Moho.

The available evidence indicates that crustal thickness may vary across a fracture zone according to the age offset, although this relationship is not implicit in, nor required by lithospheric cooling models [e.g., Sclater and Francheteau, 1970; Parker and Oldenburg, 1973; Davis and Lister, 1974; Parsons and Sclater, 1977]. There is scant evidence that oceanic crust thickens significantly with either age or spreading rate [Goslin et al., 1972; McClain, 1981; McClain and Atallah, 1986; Chen, 1992; Mutter and Mutter, 1993]. Seismic refraction evidence from the 25 to 30 m. y. age offset, left-lateral Mendocino Fracture Zone in the northeastern Pacific indicates that the thickness varies by a
factor of two from the younger (thinner) side to the older side at 128°30'W across crust aged from 4 Ma to 32 Ma [Raitt, 1963; Dehlinger et al., 1967; Dehlinger, 1969; Cazenave et al., 1982], and the crust thins by ~1 km on the younger side of the Molokai Fracture Zone [ten Brink and Brocher, 1988] (Table 4.2). However, with the exception of the Tydeman Fracture Zone which exhibits ~1 km thinner crust on its northern, younger side [Calvert and Potts, 1985; Potts and Calvert, 1986b] (Table 4.2, Figure 4.32), fracture zones in the north Atlantic appear to juxtapose slightly thicker crust on the younger side to thinner crust on the older side (Table 4.2, Figure 4.32). In these cases, however, the thickness difference appears to be a function of the rate the crust resumes normal thickness away from the fracture zone (Figure 4.32), however the resolution of Moho depths determined in these wide-angle seismic experiments is uncertain to within 1 to 1.5 km. This variation in thickness from thinner crust on the younger side in the Pacific to thinner crust on the older side in the Atlantic may be due to the different thermal regimes in effect at slow-spreading, magma-starved Atlantic spreading centers and the robust, fast-spreading Pacific ridges. The location of transverse ridges bordering fracture zones also varies; in the Pacific, transverse ridges border the younger side of the fracture zone [Sandwell and Schubert, 1982a; Sandwell, 1984], however, in the Atlantic the converse is true [Abrams et al., 1988].

Although I have thus far assumed that the Canton-Clipperton transform was a right-stepping offset of the paleo-spreading axis (Figure 3.1), in the absence of magnetic-anomaly lineations offset across the Nova-Canton Trough and an obvious depth difference across the trough, I have had no method of determining the offset sense. The present-day Clipperton Transform Fault and the inactive fracture zone trace is right-lateral to immediately east of the Line Islands [Eittreim, 1992]. However, the extreme crustal thinning in the southern region of the eastern Nova-Canton Trough should be expected on
Figure 4.32: Crustal thicknesses proximal to three north Atlantic fracture zones from sources listed in Table 4.2. Note that the crust appears somewhat thinner on the older sides of the fracture zones at comparable distances, however this may be due to a difference in the rate of return to normal crustal thickness away from the anomalous fracture zone subsurface structure.
the younger side, if the evidence from studies of the Molokai and Mendocino fracture zones can be extended to include the Canton-Clipperton region.

The age contrast across fracture zones provides a unique locale to investigate the thermal and mechanical properties of oceanic lithosphere. The change in geoid amplitude across fracture zones has been used to determine the effective elastic thickness of the lithosphere [e.g., Crough, 1979; Cazenave et al., 1982], to test the various lithospheric cooling models [e.g., Detrick, 1981; Sandwell, 1984; Marty et al., 1988; Marty and Cazenave, 1988], and to investigate the flexural response of the lithosphere to the age/temperature contrast juxtaposed at fracture zones [e.g., Sandwell and Schubert, 1982b; Parmentier and Haxby, 1986; Haxby and Parmentier, 1988; Wessel and Haxby, 1989, 1990].

By examining the shape and amplitude of the geoid step across the Nova-Canton Trough, I can reasonably infer that the southern side is younger, that is, the Canton-Clipperton transform was a right-lateral offset of the paleo-spreading axis (Figure 4.33). The step-like change in geoid height across a fracture zone is normally an indicator of the difference in age juxtaposed at the offset, although the amplitude of the step decreases with increasing lithospheric age [Detrick, 1981]. Figure 4.33 shows that the geoid height increases to the south of the Nova-Canton Trough. The northeastern Pacific fracture zones all show this same increase in geoid height from the older to the younger side of the fracture zone [Marty et al., 1988]. Given this evidence, it may well be possible that the southern region of the Nova-Canton Trough possesses an extremely thin crust, given the above evidence and as shown in the gravity models across this region (Figures 4.15, 4.17, 4.19, 4.21, 4.23, 4.25, 4.27, and 4.29). However, the 4 to 5 km crustal thicknesses modeled for profiles across the southeastern Nova-Canton Trough (Figures 4.15, 4.17, 4.19, 4.21, 4.23, 4.25, 4.27, and 4.29) seem underestimated given an average oceanic crustal thickness of 7±1 km [White et al., 1992], although the thinning immediately south
Figure 4.33: (a) Location of the four GEOSAT profiles overlain on bathymetry of the Nova-Canton Trough region projected about a pole located at 34.4°S, 150.6°W. (b) Ascending GEOSAT profiles 128 and 214 and descending profiles 171 and 257 with south located to the left. Note that the geoid step decreases from south to north across the Nova-Canton Trough topography, implying that the southern side of the trough is younger than the crust to the north.
of the central trough is consistent with observations from the wide-angle seismic experiments mentioned above. If the models had used two separate density layers to approximate oceanic crust, that is, a seismic layer 2 \( (\rho = 2.60 \text{ g/cm}^3) \) and seismic layer 3 \( (\rho = 2.90 \text{ g/cm}^3) \), the same offset difference would be preserved unless the thickness of each layer were varied independently, a crustal geometry that is inconsistent with observation. The southern region could exhibit a more reasonable thickness, but the northern region would have to thicken correspondingly.

Another explanation for the difference in crustal thickness across the eastern Nova-Canton Trough may be that the Nova-Canton Trough marks the boundary of a tectonic corridor [Kane and Hayes, 1992] across which parameters such as subsidence rate, ridge crest depth, and geoid rate change along the mid-ocean rise system. Kane and Hayes [1992] have defined eight tectonic corridors bounded by tectonic flowlines, or fracture zones, in the south Atlantic. Although these authors do not include crustal thickness or (implied) robustness of the axial magma chamber, these parameters may also vary along the ridge.

The various crustal thickness and density models presented in Figures 4.15, 4.17, 4.19, 4.21, 4.23, 4.25, 4.27, and 4.29 indicate that the high density body \( (2.90 \text{ g/cm}^3) \), Figures 4.15c, 4.17c, 4.19c, 4.21c, 4.23c, 4.25c, 4.27c, and 4.29c) underlying the central and eastern troughs requires far too much crustal thickness (average 8.5 km) to be reasonable applications of crustal structure. The average crustal density models (Figures 4.15a, 4.17a, 4.19a, 4.21a, 4.23a, 4.25a, 4.27a, 4.29a) show crustal thicknesses in the trough axes ranging from 4.8 to 8 km, within reasonable limits. The low density body \( (2.50 \text{ g/cm}^3) \), Figures 4.15b, 4.17b, 4.19b, 4.21b, 4.23b, 4.25b, 4.27b, and 4.29b) requires thicknesses of 4 to 5.7 km to model the observed free air gravity profiles, not unreasonable given evidence from north Atlantic fracture zone seismic surveys. All of the north Atlantic fracture zone evidence suggests that the crust within the fracture zone
trough exhibits low velocities (hence low density) and less than average oceanic crustal thickness. Given the lack of available evidence describing Pacific fracture zone crustal structure, I cannot assume that the eastern Nova-Canton Trough should be different, although the vastly different magmatic budgets and spreading rates characteristic of the Pacific would imply different crustal structures from the slow-spreading, magma-starved Atlantic.

Gravity Analysis of the Region South of the Nova-Canton Trough

In order to investigate the apparent crustal thinning southeast of the Nova-Canton Trough and the general character of the entire southern region, I examined profiles from GSJ cruises gh82a and gh82b [Usui, 1992] crossing the region south of the Nova-Canton Trough. Seafloor morphology south of the ridge-trough province changes dramatically from west to east. A large, NW-SE-trending turbidite-filled basin (Figure 4.1) lies in the westernmost region separated from the central turbidite-filled basin (Figure 4.1) by a N140°E-striking elongate ridge at 167°W, 2°S. Sediment depths range from ~400 m at profile gh82a.e (Figure 2.4) to ~150 m deep ponds between basement highs at profile gh82a.c to ~200 m at profile gh82a.a [Tanahashi, 1992]. West of the elongate ridge, the seafloor becomes generally rougher, covered with N140°E-striking abyssal hills and dotted with small seamounts. Seafloor depths are 200 to 300 m shallower east of the elongate ridge (Figure 2.4), suggesting to Yamazaki and Tanahashi [1992] that the ridge is of transform fault origin. Joseph et al. [1993] hypothesize that the elongate ridge is a rift originating at the northeastern boundary of the Manihiki Plateau; topographic profiles crossing this ridge (Figure 2.4 - profiles gh82a.e, gh82b.a, and gh82a.d) appear to support a rift origin of this feature because of the ridge's characteristic upward flexed outline. At
the western limit of the southeastern basin, seafloor depths begin to shoal, but there is no information for the region west of the study area.

The free air gravity profiles show -10 to -30 mGal gravity lows surrounding a 20 mGal gravity high over the elongate ridge (Figure 4.3, Figure 4.34 - profiles gh82a.e, gh82a.d, and gh82a.c). Values across the western turbidite-filled basin range from zero at profile gh82a.e to -20 mGal at profile gh82a.a (Figure 4.34). Gravity lows across the turbidite-filled basin bordering the N140°E-striking ridge to the east reach ~30 mGal. Gravity highs over the entire eastern region average ~15 mGal.

Topographic profiles were constructed by projecting cruise track 3.5 kHz echo-sounder data for lines gh82a.a through gh82a.h and gh82b.a through gh82b.d (Figure 4.1) perpendicular to a line drawn along the trend of the elongate ridge (Figure 2.4 and 4.34), and sampled at 0.5 km intervals. Free-air gravity models were constructed for four of these profiles (Figure 4.1) to evaluate the topographic contribution and the gravity due to a constant thickness and a variable thickness crust. Sediment thicknesses on these four profiles are under 200 m, and were therefore ignored. In order to bring the modeled gravity profiles into agreement with the profiles across the ridge-trough province, I used the same reducing constant for the southern profiles and changed the constant crustal thickness to force the match prior to varying the crustal thickness from point to point. No other densities were used in the southern region models because I can find no geological justification to vary the density from the oceanic average of 2.70 g/cm³.

The southernmost extents of profiles crossing from the central and eastern regions of the Nova-Canton Trough show high mass excesses implying either a significantly thinner crust south of the ridge-trough topography (Figures 4.19, 4.21, 4.23, 4.25, 4.27), or a high density (> 2.70 g/cm³) body at depth.

Profile gh82a.a and gh82a.h cross the entire southern region from the western turbidite-filled basin to the abyssal hill province (Figure 4.1) including the northward
Figure 4.34: Gravity data projected normal to the strike of the elongate ridge at 167°W, 2°S corresponding to the projected bathymetric profiles shown in Figure 2.4. The small triangles show the location of projected data values.
extension of the N140°E-striking ridge. The topography (Figure 4.35a) fails to model the -15 to 20+ mGal gravity variation across the southern area despite the depth difference. A 5.2 km thick crust most closely reproduces the observed gravity signal after the reducing constant is removed but does not replicate the amplitude of the negative to positive change from west to east (Figure 4.35b). A Moho discontinuity under the ridge simulating ridge flank uplift, such is observed adjacent to rifted structures [Weissel and Karner, 1989] and a 4.5 to 5 km thick crust in the eastern region can successfully reproduce the observed free air anomaly (Figure 4.35c).

Profile gh82a.f crosses a typical portion of the southeastern region away from the influence of Dolmah Seamount (Figure 4.1). Free air gravity values range from near zero in the west to ~15 mGal in the center of the profile (Figure 4.34). The topography predicts a 25 mGal high over a 4.5 km deep peak in the west and a 12 mGal high over a 4.75 km deep peak near the center of the line (Figure 4.36a), although the observed free air gravity values show no such highs. A 5 km thick crust most closely agrees with the observed free air gravity profile. However, the constant crustal thickness approximation (Figure 4.36b) does not correct these deviations from the observed values, but approximating isostatic compensation beneath the two small seamounts and a 4.5 km thick crust at the eastern end of the profile (Figure 4.36c) reproduces the observed gravity signature. If this crust/mantle interface configuration is correct, it is the first instance of locally compensated topography in the entire region.

More examples of possible isostatic compensation and crust/mantle upwarp are evident on profile gh82a.c (Figure 4.37) which crosses the western turbidite-filled basin, the elongate ridge, and the westernmost part of the abyssal hill province (Figure 4.1). The topography alone accounts for some of the observed gravity (Figure 4.37a), however the gravity predicted by the topography overestimates the gravity high across seamounts at either end of the profile and values across the western turbidite-filled basin. The constant 5
Figure 4.35: gh82a.a, gh82a.h. (a) Observed free air gravity (small circles) and modeled profile (solid line) for a flat Moho to show the gravity due to seafloor topography only. Note that the change in depth across the elongate ridge is not sufficient to create the higher gravity values in the southeastern region, nor is the amplitude of the change at the ridge modeled by the topography alone. (b) The changing Moho relief of the constant crustal thickness model better approximates the gravity change from west to east but again does not successfully reproduce the observed gravity about the elongate ridge. (c) A somewhat thinner crust is required east of the elongate ridge to reproduce the observed gravity, however thickness increases at the far western edge of the profile. A severely upwarped Moho is necessary under the ridge and crustal thinning on the eastern edge.

NOTE: All models for the southern region show the free air anomaly in 10 mGal divisions.
mGal

depth (km)

-16.0
-12.0
-8.0
-4.0

-80 -40 0 40 80 distance (km)

2.70

3.30

1.03

a.

mGal

depth (km)

-16.0
-12.0
-8.0
-4.0

-80 -40 0 40 80 distance (km)

2.70

3.30

1.03

b.

gh82a.a, gh82a.h

mGal

depth (km)

-16.0
-12.0
-8.0
-4.0

-80 -40 0 40 80 distance (km)

2.70

3.30

1.03

c.
Figure 4.36: gh82a.f. (a) Across the southeastern region, topography accounts for most of the observed anomaly, except across two seamounts. The topography predicts much higher gravity values than shown by the observed profile. (b) The constant 6 km thick crust composed of a 2.70 g/cm$^3$ layer does not remove the mass deficits over the two seamounts, and creates a mass excess at the extreme eastern end of the profile. (c) Local isostatic compensation is required under the two seamounts to reproduce the observed gravity values by increasing the depth to the crust/mantle interface in an amount proportional to the density differences at the seafloor and Moho (~3:1).
km crustal thickness approximation more closely reproduces the values across the turbidite-filled basin and western seamounts (Figure 4.37b), but overestimates values across the eastern seamount and underestimates the values across the ridge. By increasing crustal thickness in an amount proportional to the density ratios (~3:1) at the eastern seamount, the observed gravity is replicated (Figure 4.37c). In addition, sharp, upwarped Moho relief below the N140°E-striking ridge and a thinned crust is necessary to approximate the free air gravity values across this structure. Interestingly, the small seamount immediately west of the higher, broader seamount at the eastern end of the profile (Figure 4.37c) requires decreased Moho depth to reproduce the small step in the broader gravity high of the larger seamount but this may be due to the very different sizes of the two seamounts.

Profile gh82b.d parallels the central trough immediately south of Dolmah Seamount (Figure 4.1). The residuals remaining after modeling seafloor topography (Figure 4.38a) show mass excesses and deficits implying considerable Moho relief. I started with a 4.5 km thick crust for the constant crustal thickness model because this thickness gives the smallest residuals after applying the reducing constant required by the Grav2D program [Hurst, 1989]. The gravity profile calculated using a constant 4.5 km thick crust shows a 10 to 15 mGal misfit along most of the profile, with the exception of the central area (Figure 4.38b). The variable thickness model (Figure 4.38c) requires a significantly thinner crust near the western end of the profile, but the uplifted Moho used to model the observed gravity is probably artificial, as in profile mw902.4 (Figure 4.17). Profile gh82b.d skirts the edge of Dolmah Seamount at this point (Figure 4.1) and the effect of the seamount high is reflected in the gravity (failure of the 2D approximation). Crustal thickness ranges from 4.0 to 5.5 km on this profile, which agrees with the crustal thicknesses modeled for profiles gh8b1 and mw902.2 (Figures 4.19 and 4.23) using the same crustal density structure south of the central trough.
Figure 4.37: gh82a.c. (a) Gravity due to seafloor topography for profile gh82a.c.. Gravity values for the seamount at the far eastern end and the western turbidite plain are overestimated by the seafloor relief. (b) Decreased depth to Moho under the eastern seamount increases the residual, and the variable Moho relief better fits the western turbidite-filled basin but creates mass excess across the elongate ridge. (c) By assuming local isostatic compensation under the eastern seamount, and crustal thinning under the elongate ridge, the observed profile can be replicated.
Figure 4.38: gh82b.d. (a) Seafloor contribution to the gravity field for profile gh82b.d, crossing the southeastern abyssal hill region proximal to the central trough. The undulations of the modeled profile do not reproduce the observed profile in amplitude or shape. (b) The changing Moho relief shown in the constant crustal thickness approximation also does not reproduce the observed profile, however the departures are between 10 to 15 mGal. (c) A extremely changeable crust/mantle boundary is necessary to reproduce the observed gravity values. Very thin crust (< 4.5 km) is necessary near the eastern end of the profile, before the crust must thicken to > 6 km to model the seamount at the eastern edge. The severe thinning near the western end of the traverse is probably an artifact due to proximity to Dolmah Seamount.
distance (km)

-4.0

depth (km)

-8.0

-12.0

a.

mGal

distance (km)

60 80 100 120 140

1.03

-4.0

2.70

3.30

b.

mGal

distance (km)

60 80 100 120 140

1.03

-4.0

2.70

3.30

c.

gh82b.d
Interpretation of Gravity Models for the Region South of the Nova-Canton Trough

The modeled gravity profiles for the region south of the Nova-Canton Trough suggest that slightly less than normal crustal thicknesses are found here with a few notable exceptions. On a large scale, Moho relief follows the topography, indicating that the region is not locally compensated. A few seamounts appear to be locally compensated by crustal roots (Figures 4.36, 4.37) but the gravity low across the western turbidite-filled basin (Figure 4.36), for example, suggests more, rather than less, depth to Moho. The most interesting deviation from the constant crustal thickness assumption is the elongate ridge at 167°W. The variable thickness models for profiles gh82a.a and h and gh82a.c (Figures 4.35c, 4.37c) show that crustal thinning is necessary beneath the eastern (steeper) side of the ridge, an expected Moho configuration at a rifting site. The upward flexed profile of the ridge (Figure 2.4) further substantiates this interpretation of the morphology, suggesting flexural response of the lithosphere to the unloading accompanying rifting [Weissel and Karner, 1989].

The local compensation of some seamounts in the southern region implies formation at or near a ridge crest (Figures 4.36, 4.37) [e. g., Watts and Ribe, 1984]. In contrast, Dolmah Seamount appears to be largely uncompensated (Figure 4.15). To model the observed free air values across Dolmah Seamount, higher density bodies must be present closer to the seafloor, either through an uplifted Moho beneath the seamount or a high density pipe within the seamount (Figures 4.15a, b, c) [Kellogg et al., 1987]. The age of Dolmah Seamount is problematic; high backscatter values shown on SYS09 sidescan-sonar images of the edifice infer little or no sediment cover on the seamount crest or its flanks (Figure 1.2), however this evidence is not conclusive because the abyssal hill region in general exhibits thin (~60 m) sediment cover. Given the location of the Nova-
Canton Trough region in the equatorial high productivity belt, the lack of a considerable sediment cover is unexpected. Antarctic Bottom Water (AABW) funneled through the Samoa Passage is considered responsible for the thin sediment cover, either through non-deposition or erosion of existing sediments [Hollister et al., 1974]. However Dolmah Seamount may have been formed long after the Nova-Canton Trough structures. Based on the recovery of samples of fresh volcanic glass, Rosendahl [1972] and Rosendahl et al. [1975] suggest that the region may have experienced a second stage of tectonism. However, those samples were recovered from the southern slope of the deep western trough about 100 km from Dolmah Seamount, therefore not unequivocally diagnostic of the seamount's composition. Recent theories regarding a mid-Cretaceous intraplate volcanic event spurred by a surfacing mantle "superplume" [Larson, 1991] may lend more credence to Rosendahl's [1972] and Rosendahl et al.'s [1975] rejuvenation hypothesis. In any case, the modeled gravity across seamounts in the region south of the ridge-trough province suggests both locally compensated (Figures 4.36, 4.37) and uncompensated (Dolmah Seamount, Figure 4.15) seamounts.

Gravity models imply that crustal thickness immediately south of the central trough in the eastern region (Figure 4.38) is slightly less than the regional average of 5.0 to 5.5 km. The best-fitting crustal geometry employing a single density crust (Figure 4.38c) shows an average 10.0 km depth to Moho, or a 4.8 km thick crust. This crustal thinning is only evident very close to the fracture zone.

Summary

On the basis of forward 2-D gravity modeling of the Nova-Canton Trough region I can draw the following conclusions and make suggestions regarding the crustal thicknesses and intracrustal density structure of the region:
1. The deep western trough and adjacent crust can be reasonably approximated by a constant density 6 km thick crust with slight crustal thickening in the trough, as previously reported by Yamazaki and Tanahashi [1992]. However, a variable thickness, variable density crustal structure can more closely approximate the observed gravity, with an upwarped Moho under the southern bordering ridge, similar to the Moho structure observed at fracture zone transverse ridges. Whereas the ridge-trough province is highly asymmetric, both bathymetrically and in the gravity field, crustal structure to the north and south of the ridge-trough province is very symmetric. I conclude that although the presented gravity models cannot substantiate a spreading center later overprinted by transform motion, this interpretation is not contradicted (Figure 4.39).

2. In contrast to the symmetry shown adjacent to the deep western trough, the crustal structure in the eastern Nova-Canton Trough, as interpreted from gravity models, is extremely asymmetric north and south of the central ridge and trough and the eastern trough. This asymmetry is implied by a 10 to 20 mGal gravity high across the region south of the eastern Nova-Canton Trough as opposed to a 0 to -15 mGal low across the northern turbidite-filled basin. The region immediately south of the central trough is 1.5 to 2 km thinner than to the north (Figure 4.39).

3. The ridge-trough province across the eastern Nova-Canton Trough is best approximated by a laterally heterogeneous crustal density structure including a low density (i.e., serpentinite) body in the trough axes, and some crustal thickening beneath the central ridge (Figure 4.39). Crustal thickness based on gravity models across the central ridge are within reasonable limits and do not indicate the need for a density different from an average layer 2 density (2.70 g/cm³).

4. The polarity of the geoid step across the Nova-Canton Trough implies an initial right-lateral offset of the paleo-spreading axis, therefore the crust decreases in age from north to south. Evidence from wide-angle seismic experiments across the Molokai and Mendocino
Figure 4.39: Preferred gravity profiles for models of crustal structure along the strike of the ridge-trough province.
fracture zones in the north Pacific support this interpretation by analogy, because they indicate that the crust is thinner on the younger side of the fracture zone.

5. Free air gravity values across Dolmah Seamount and models based on seafloor topography suggest that Dolmah Seamount is not locally compensated by a deep crustal root. However some seamounts in the southeastern region do require crustal roots, implying local Airy compensation and possible formation at or near a ridge crest setting.

6. In general, the region south of the ridge-trough province exhibits slightly less than normal crustal thickness. However, the elongate ridge at 167°W, 2°S requires an upwarped Moho and some crustal thinning beneath the eastern edge of the ridge. This crustal geometry and the flexed profile of the ridge are consistent with the flexural response of the lithosphere to unloading accompanying rifting.
Appendix

According to Hubbert [1948], the vertical component of gravitational attraction due to a two dimensional body is

\[ 2Gp\int zd\theta. \] (1)

Talwani et al. [1959] show that the horizontal component of gravitational attraction is given by

\[ 2Gp\int xd\theta. \] (2)

For an n-sided body (Figure A), the vertical and horizontal components of gravitational attraction can be calculated as follows: Let P (Figure A) be the origin of a coordinate system with x defined positive to the right and z positive downwards. The n-sided body lies in the xz-plane. For a line segment on the n-sided body, say AB, the attraction can be evaluated by constructing right triangles, \( \Delta pjX_j \) and \( \Delta QjX_j \). \( \Delta pjX_j \) subtends angle \( \angle \theta_i \) with the x-axis. Similarly \( \Delta QjX_j \) creates \( \angle \phi_i \) with the x-axis. Because both triangles are right triangles, it is apparent that the z component at j is given by

\[ z = X_j \tan \theta_i, \] (3), and

\[ z = (X_j - PQ) \tan \phi_i. \] (4).

Therefore

\[ X_j \tan \theta_i = (X_j - PQ) \tan \phi_i. \]

Substituting \( X_j = z \tan \theta_i \),

\[ (z/\tan \theta_i) \tan \theta_i = (z/\tan \theta_i - PQ) \tan \phi_i, \] or

\[ z = \frac{PQ \tan \theta_i \tan \phi_i}{\tan \phi_i - \tan \theta_i}. \]
Substituting this into 1 gives
\[ \int_{AB} zd\theta = \int_{A}^{B} \frac{PQ \tan \phi_i \tan \theta_i}{\tan \phi_i - \tan \theta_i} d\theta \equiv Z_i. \]

By solving for \( X_j \) in equations 3 and 4 gives
\[ X_j = \frac{PQ \tan \phi_i}{\tan \phi_i - \tan \theta_i}, \]

and substituting this into equation 2 gives
\[ \int_{AB} X_j d\theta = \int_{A}^{B} \frac{PQ \tan \phi_i}{\tan \phi_i - \tan \theta_i} d\theta \equiv X_i. \]

The vertical and horizontal components of gravitational attraction for the entire polygon can be calculated by summing the attraction of each side:
\[ V = 2G\rho \sum_{i=1}^{n} Z_i, \]
\[ H = 2G\rho \sum_{i=1}^{n} X_i. \]
Figure A.1: An n-sided polygon body showing the parameters used to calculate horizontal and vertical gravitational attraction as discussed in the text.
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