The Blanco fracture zone contains many features which are not consistent with simple strike-slip movement along a transform fault. These features consist of areas of reverse faulting, normal faulting, uplift, folding and subsidence. Sedimentological analysis of piston cores indicate that Cascadia depression is subsiding at an average rate of 1.0 cm/yr while abyssal plain turbidites adjacent to this depression have been uplifted over 450 meters at a rate of 2.0 cm/yr.

This study presents single channel seismic reflection profiles acquired in the central portion of the fracture zone which can be used to document tectonism. The interpretation of these reflection profiles shows that the abyssal hill adjacent to the depression is composed almost entirely of uplifted, folded turbidites. The Blanco ridge, a 165 km long ridge that forms a barrier to turbidite deposition on the southeast portion of the fracture zone, is a fault bound feature containing uplifted basement and turbidite sediments. An uplifted
sediment bench on the north side of the Blanco ridge can be traced as a linear feature over a distance of 40 km. The south side of the Blanco ridge contains at least 750 meters of structurally deformed turbidite sediments in a subsiding trough. This trough apparently overlies an area of convergence and subsidence of the Pacific plate adjacent to the Juan de Fuca plate along the Blanco fracture zone. Seismic reflection profiles also show Cascadia depression to be an area of active subsidence, accompanied by stratigraphic thickening and normal faulting of sediments in this basin.

Evaluation of the geologic and geophysical data presented here suggests that a compressional component present in the stress field, oriented normal to the Blanco fracture zone, is controlling the morphology and structure in the central portion of the fracture zone. This compressive stress is the result of the clockwise rotation in spreading directions of the Juan de Fuca and Gorda spreading ridges as well as a differential northward motion between the Pacific and Juan de Fuca plates. This differential motion is inferred by the overriding of the Juan de Fuca plate by the Pacific plate along the Mendocino fracture zone to the south. The extension present in Cascadia depression is the result of an 18 to 20 km offset in the trend of the Blanco fracture zone structures, forming two, en echelon fault zones.
The Structure and Tectonics of the Blanco Fracture Zone

by

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THE STRUCTURE AND TECTONICS OF
THE BLANCO FRACTURE ZONE

INTRODUCTION

Mid-ocean spreading ridges and their associated transform faults are among the most striking features of the ocean basins. The plate boundaries of the northeastern Pacific have only recently been studied in the detail necessary to resolve some of the problems present in the area. The Blanco fracture zone, which connects the Gorda and Juan de Fuca ridges, is one of the larger tectonic features in this area and one of the least studied and understood transform faults in the Pacific ocean (Figure 1).

Many authors have stated that the Blanco fracture zone is anomalous. Some of the most notable aspects are the west northwest trend of the fracture zone and its complex, irregular morphology. The west northwest trend is anomalous because of its diversion from the east-west trend of the Mendocino fracture zone, and most other major Pacific fracture zones. The complex morphology is anomalous when compared to the relatively simpler structure of most other fracture zones. This study presents geological and geophysical data which helps answer many of the questions surrounding the origin and kinematics of the Blanco fracture zone.

As determined in this study, the Blanco fracture zone is not a simple transform fault representing pure strike-slip motion. While strike-slip motion may predominate, there is also a large component of
Figure 1. General tectonic features of the northeast Pacific.
compression present in the stress field, which is actively influencing the morphology and geology of the fracture zone. We will point out or emphasize how this compressional component relates to the overall relative motions of the Pacific, Juan de Fuca and North American plates.

One note of clarification is required at this point in regard to the name "Juan de Fuca" plate. Throughout this paper, the author will refer to the Juan de Fuca plate as a single entity, recognizing that smaller subplates, such as the Explorer Plate and possibly a Gorda subplate, are also present (Figure 1).
BACKGROUND

Morphology

The Blanco fracture zone is a west northwest trending bathymetric feature on the ocean floor, located off the southern Oregon coast (Figure 1). The fracture zone is approximately 460 kilometers long and consists of a series of elongate, subparallel ridges and troughs (McManus, 1967a, 1967b; Duncan, 1968). The fracture zone is terminated at its western and eastern ends by a juncture with the Juan de Fuca and Gorda ridges, respectively. Associated with these terminations and ridge intersections are the deepest topographic features of the fracture zone, troughs of greater than 4000 meters water depth (Figure 2) compared to an average sea-floor depth of about 3000 meters in the area.

The Blanco fracture zone is one of two basic types of Pacific fracture zones. The first type is the Mendocino fracture zone type; it exhibits a single large elongate ridge which separates sea-floor of greatly differing regional depths on either side of the ridge (Figure 3A). The second type of Pacific fracture zone is the Blanco type, which consists of small, subparallel, elongate ridges and valleys with little or no change in regional depth across the fracture zone.

Both Pacific types of fracture zones are different from the basic Atlantic type fracture zone. The latter fracture zone typically consists of a deep, essentially continuous, central trough bordered on one or more sides by narrow, high ridges. The Romanche fracture zone and Vema fracture zone, described by van Andel et al. (1967), van Andel (1971) and Bonatti and Honnorez (1976), are excellent examples of the Atlantic type fracture zone (Figure 3B, 3C).
Figure 2. Bathymetry of the Blanco fracture zone. (after Wilde et al., 1978 and Wilde et al., 1979 with modifications using unpublished OSU data.)
Figure 3. Profiles of:

3A Pacific type fracture zone. The Mendocino fracture zone is a typical example.

3B, 3C Atlantic type fracture zone as illustrated by the Romanche (B) and Vema (C) fracture zones. (modified after Bonatti and Honnorez, 1976.)
The Blanco fracture zone, then is one of the more complicated classes of transform faults from a morphologic standpoint. The fact that transform faults can exhibit radically different types of morphology probably also implies a kinematic or tectonic difference. Van Andel (1971) suggested that the difference between the Atlantic type and the two Pacific types of transform faults is possibly due to the different spreading rates between the two mid-ocean ridge systems (i.e., slow in the Atlantic and fast in the Pacific). He also suggested that since the East Pacific rise is not changing its orientation with time, as are parts of the mid-Atlantic ridge, this could be a factor in the morphologic differences between the transform faults in the two ocean basins. It will be shown later in this paper that the most important factor influencing the morphology of the Blanco fracture zone is the change in orientation or spreading direction of the Gorda-Juan de Fuca spreading system.

Age of the Blanco Fracture Zone

The age of the Blanco fracture zone can be inferred from geophysical and sedimentologic data. Based upon the recovery of sediments from the fracture zone which contained Foraminifera of Pliocene age, McManus (1967b) suggested a late Tertiary age for the fracture zone.

Atwater (1970) used the sea-floor magnetic anomaly patterns, numbered according to the time scale of Heirtzler, et. al. (1968), to reconstruct the plate motions of the northeast Pacific back to about 30 million years ago. In this reconstruction, Atwater showed the first offsets in
magnetic lineation pattern due to the presence of the "proto" Blanco transform fault occurred about 10 million years ago. Elvers, et. al. (1973) show that the first magnetic lineation offset by the "proto" Blanco was anomaly 5A on the NOS Seamap series magnetic anomaly map (map No. NOS 12042-12M) (Potter et al., 1974). Ness et al. (1980) suggest a date of about 12 million year b.p. for this anomaly. While more recent theories on the formation of oceanic crust at spreading centers and its effect on magnetic anomaly patterns (Hey, 1977) may alter this date, it is the best estimate available from geophysical interpretation of the magnetic anomalies.

Gravity and Magnetics

Gravity

Dehlinger, et. al. (1970) have summarized what is known about the gravity field of the Blanco fracture zone and there has been little additional work done on a scale which is detailed enough to significantly change the earlier gravity interpretations. In general, the entire area of the Blanco fracture zone exhibits a negative free-air anomaly (Figure 4). One area exhibits a very large negative free-air anomaly (-30 m gal.). It is located over the Blanco trough, the deep trough (greater than 4000 m.) at the intersection of the Juan de Fuca ridge with the Blanco fracture zone. Dehlinger, et. al. (1970) attribute this large negative anomaly solely to topography.

Melson (1969) also obtained gravity data in this area and calculated a larger free-air anomaly over the Blanco trough. He suggested that the -50 m gal. anomaly is not as large as it should be for such a deep structure, meaning that some type of compensation has occurred in
Figure 4. Free-air gravity anomaly map west of Oregon. (after Dehlinger et al., 1970.)
this area, possibly by narrow upwelling of mantle material along the axis of this trough. This would be similar to the emplacement of material along a spreading center.

**Magnetics**

The magnetic survey of Raff and Mason (1961) off the west coast of North America still remains one of the most heavily used data sets with respect to plate tectonic interpretations in this area. Additional magnetic surveys were made in the area by Melson (1969) and NOAA in 1971 (Potter et al., 1974). The later survey produced the NOS Seafloor series of maps, within which map number NOS 12042-12M covers the area of the Blanco fracture zone. This map contains nearly all the features of the data presented by Raff and Mason (1961). As was mentioned earlier, the detailed interpretation of the geologic history represented by these sea-floor magnetic anomalies is a current subject of intense research. One of the more exciting new interpretations involves the "propagating rift" hypothesis first suggested by Couch (1969) and expanded upon by Hey (1977). It is not clear at this time what impact the phenomena of propagating rifts has on large transform faults such as the Blanco fracture zone.

Melson (1969) obtained detailed magnetic data in the area west of the intersection of the Blanco fracture zone with the Juan de Fuca ridge, across a possible westward extension of the fracture zone (Figure 5). Melson's data indicate that the magnetic anomaly pattern from the Juan de Fuca ridge extends southward across what would have been the extension of the Blanco fracture zone. However, anomalies located to the south of the western edge of the Blanco, originating from the Gorda ridge, are disturbed.
Figure 5. Magnetic anomaly map of the region near the Juan de Fuca-Blanco fracture zone juncture (after Melson, 1969). Solid lines are positive anomalies, dashed lines are negative.
The disturbed Gorda ridge anomalies are attributed to a possible clockwise rotation of the trend of the Blanco fracture zone through time (Melson, 1969).

The NOS Seanimap series magnetic map (Potter et al., 1974), whose anomalies are described by Elvers et al. (1973) is exactly opposite to that of Melson (1969) in that the Elvers et al. map does indicate a westward extension of the Blanco fracture zone. This is probably due to the increased amount of data collected by NOAA which allowed correlation of anomalies over a much greater area. This results in better definition of anomalies produced by the Gorda ridge versus those produced by the Juan de Fuca ridge. Elvers et al. (1973) also suggest clockwise rotation of both the Gorda and Juan de Fuca ridges, asymmetric spreading of the Juan de Fuca ridge, and possible subduction along the Blanco fracture zone, based on their interpretation of the magnetic anomaly pattern. Both the clockwise rotation of the ridge systems and possible subsidence of the Pacific plate along the Blanco fracture zone are supported by geological and geophysical data presented later in this study.

Melson (1969) also notes a strong positive (greater than 1400 gammas) magnetic anomaly over the Blanco trough. This anomaly is also seen on the NOS Seanimap series map. Melson (1969) suggests that this strong positive magnetic anomaly may be associated with recent volcanic activity.

More recently, Detrick and Lynne (1975) applied three dimensional magnetic models to the Blanco trough in order to explain this strong positive anomaly. This three dimensional modeling is based on the recovery of highly magnetized basalts dredged from the Blanco trough,
which are highly magnetized due to a high content of iron and titanium, probably as a result of extensive shallow fractionation of a normal oceanic tholeiite (Detrick and Lynne, 1975). The resulting models imply that the high intensity magnetic anomaly of the Blanco trough is due to the possible presence of a block of highly magnetized basalt which Detrick and Lynne (1975) suggest may have leaked out of the Juan de Fuca ridge.

This leakage of material into the Blanco fracture zone from the Juan de Fuca ridge, however, cannot be used to imply that the Blanco is a "leaky transform". As pointed out by Vogt and Byerly (1976) in their elaboration on the presence of high Fe, Ti basalts as the cause of the high magnetic amplitudes, a "leaky transform" can only occur in a clockwise rotating transform fault with a right lateral ridge offset (or vice versa). The Blanco fracture zone is rotating clockwise, but has a left lateral ridge offset. Apparently some other mechanism, as yet unknown, must be invoked to emplace any anomalous material into the Blanco fracture zone.

Sediments and Tectonics

Sediment Type and Distribution

The general distribution of the different types of terrigenous sediments in the area of the Blanco fracture zone are described by Duncan (1968), and Phipps (1974), while Selk (1978) has described in detail the metaliferous sediments of the Blanco trough. The sediment types and their general distributions are shown in Figure 6. An analysis of
Figure 6. Distribution of sediment types in the vicinity of the Blanco fracture zone. The basins at the eastern end of the fracture zone contain a very thin cover of hemipelagic sediment with a high volcanic debris content. The basin at the west end of the fracture zone contains layered metaliferous sediments (Selk, 1978).
sediment type and distribution along with age dating can provide important information on several different aspects of fracture zone tectonics and volcanism.

The presence of abundant volcanic glass and debris in the hemipelagic sediment can indicate recent submarine volcanism. Piston cores 6601-1, 6609-8 and W7707-2 (all located at or near the junction of the Blanco fracture zone and Gorda ridge) contain sediments of this type. The Blanco trough is thought to be an area of recent hydrothermal activity, based on the recovery of iron-manganese rich sediments from this trough (Selk, 1978). The presence of recent volcanic debris and Fe-Mn rich sediment layers suggest that the intersections of the Blanco fracture zone with the Gorda and Juan de Fuca ridges are both experiencing recent submarine volcanic activity. In contrast, the sediments of the central portion of the fracture zone show no evidence of Pleistocene or post-Pleistocene volcanic activity. At the intersection of the Juan de Fuca ridge and Blanco fracture zone, the hydrothermal activity may be associated with a heat source located below the fracture zone itself. This would be consistent with the magnetic and gravity data discussed earlier, both of which indicate that this intersection is anomalous in that it has a very large (-30 to -50 m gal.) negative free-air gravity anomaly and has a very large, positive magnetic anomaly (greater than 1400 gammas). The volcanic glass and debris found in the sediments near the Gorda-Blanco intersection (Figure 6) may be indicative of recent volcanic activity on the Gorda ridge only, or may imply actual volcanic activity in the deep trough located at this intersection. This deep trough (greater than 4000 m) is nearly devoid of sediment,
which may be attributed to its great depth (CaCO₃ dissolution increases dramatically below 3500 meters) and possible recent submarine volcanic flows. The ridge bordering the north side of the trough must have also predated the trough and acted as a barrier to terrigenous sediment input from Cascadia deep sea channel.

Accumulation Rates and Basement Age

Some selected sediment intervals in several piston cores were dated by Duncan (1968) and these dates have been used to calculate sedimentation rates. This age data can also be used to give some minimum ages of the basement if the total thickness of sediment is known.

The sediment overlying the crust on the southeast side of the Blanco fracture zone appear from seismic reflection profiles to be chiefly transparent, hemipelagic sediment (Figure 7). An exception occurs near Cascadia Gap, where turbidites are present (Figure 6). One seismic reflection record along profile EE' (Figure 7), shows 260 meters of hemipelagic (transparent) sediment overlying the acoustic basement assuming a seismic velocity of 1.7 km/s for this type of sediment (Kulm et al., 1973). The average Pliocene sediment accumulation rate for hemipelagic sediments in this area is approximately 10 cm/1000 years (Duncan, 1968; Heath et al., 1976). This gives an approximate age of 2.6 million years for the sediment immediately overlying the crust. The age of the basaltic crust in this area is 2.5 to 3.0 million years from magnetic anomalies using the magnetic time scale of Ness et al. (1980).
Figure 7. Seismic reflection profile EE' showing 260 meters of hemipelagic sediment overlying volcanic basement. See Figure 9 for location.
There are several piston cores in the Blanco fracture zone which can be used to document tectonic movement. The first, core 6609-19, is located in Cascadia depression. Cascadia depression is a bathymetric basin, about 70 meters deep, located in Cascadia channel just downstream of Cascadia gap (Fig. 6) (Duncan, 1968). This core, and others down channel from it contain volcanic ash from the eruption of Mt. Mazama in southern Oregon, 6600 years ago (Duncan, 1968, Griggs and Kulm, 1970). Cores down channel from this depression show a significant increase in hemipelagic deposition compared to cores upchannel of the depression. The core located in Cascadia depression shows a marked increase in post-glacial sedimentation rate, while down channel cores show a significant decrease in post-glacial sedimentation rate (Duncan, 1968). This implies that the depression is acting as a trap, preventing turbidity current flows from proceeding out onto Tuft's abyssal plain to the west. Because Mazama ash is present both up and down channel, the depression must have formed during the past 6600 years (Duncan, 1968). Given the average depth of the depression of 70 meters, the calculated subsidence rate is about 1.0 cm./year.

Core 6609-20 was taken on a small ridge on the north side of the Blanco fracture zone adjacent to Cascadia gap (Figure 6). The sediments in this core show a progressive change from sand turbidite deposits at the base, gradually reducing in number and thickness up core, to totally pelagic, foraminiferal sediment near the top. The core was taken at a water depth of 2600 meters. The present depth of the adjacent Cascadia channel thalweg is 3300 meters. If the assumption is made
that turbidity currents cannot deposit sand higher than 250 meters above the channel floor (Griggs and Kulm, 1970), then the last turbidite flow in core 6609-20 must have been tectonically uplifted about 450 meters. A radiocarbon date immediately above the youngest turbidite flow shows an age of 21,950 years (Duncan, 1968). This gives us a rate of uplift of about 2.0 cm./year, which is approximately twice the subsidence rate in Cascadia depression. It is important to note that the rates of uplift calculated here are probably absolute rates, not relative rates. That is, the turbidites in core 6609-20 have risen 450 meters above the regional depth of turbidite deposition on the surrounding abyssal plain.

These areas of anomalous, vertical tectonic activity first described by Duncan (1968) are not consistent with simple strike-slip motion along the Blanco fracture zone, as will be discussed later.

**Earthquake Seismicity**

Tobin and Sykes (1968), Bolt et al. (1968), Chandra (1974), Couch et al. (1974) and Rogers (1979) have determined or compiled the focal mechanisms of number of earthquakes in the northeast Pacific Ocean. These earthquake mechanisms exhibit predominately right lateral strike slip solutions (Fig. 8) which has been used to confirm the transform nature of the Blanco fracture zone. However, some solutions, as pointed out by previous authors (Bolt, et al., 1968 and Tobin and Sykes, 1968), show normal faulting and left lateral strike-slip motion.
Figure 8. Earthquake mechanisms for the Blanco fracture zone (after Chandra, 1974). Interpretation by Chandra, 1974, is right lateral strike-slip mechanisms.
STRUCTURE AND TOPOGRAPHY

Evaluation of the shallow structure and topography of the Blanco fracture zone provides one of the best indications of the nature of the tectonics occurring there. The structure of the Blanco fracture zone has been examined through the use of single channel air gun reflection profiles collected in the area by Oregon State University and NOAA (Figure 9). The Blanco fracture zone is especially suited to study by seismic reflection because Cascadia deep sea channel has provided a source of seismically reflective sediments (turbidites) which can be used as indicators of tectonic displacements. Interpretation of the available seismic reflection data shows several areas of tectonically disturbed sediments in the vicinity of Cascadia depression and along the Blanco ridge.

Cascadia Depression

Cascadia depression, discussed earlier with respect to sedimentological evidence of subsidence, is crossed by seismic reflection Profile AA' (Figure 10, a north-south trending line, and Profile A'B' (Figure 11), a northwest-southeast trending line.

Profile AA' provides an excellent confirmation of the subsidence in this depression. The basin contains a very thick section of turbidites which thicken markedly towards the south. The actual thickness of the sediment present in this depression cannot be determined as the profile does not receive any basement reflections. Assuming a seismic velocity of 1.8 km/second (Kulm and Van Huene et al., 1973) for the sediments, at
Figure 9. Location of seismic reflection and bathymetric profiles used in this study. All profiles were acquired by Oregon State University except Line 28 which was acquired by NOAA in 1971.
Figure 10. Seismic reflection profile AA'. This profile shows stratigraphic thickening and normal faulting in Cascadia depression, both indicators of recent subsidence. See Figure 9 for location.
Figure 11. Seismic reflection profile A'B'. This profile shows normal faulting in sediments on the border between Cascadia depression and the Blanco ridge. See Figure 9 for location.
least 750 meters of sediment is penetrated on this profile. The sediments are
truncated to the north by a steep ridge which shows no internal structure
and is believed to be volcanic in nature. In the center of the basin
several small normal faults are observed which actually have a surface
expression, that suggests a graben-like structure. The normal faulting
and stratigraphic thickening of the sediment towards the center of the
basin provide excellent evidence of recent subsidence.

Profile A'B', a northwest-southeast trending profile (Figure 11)
traverses the western end of the Blanco ridge and the Cascadia depression
(Figure 9) and exhibits approximately 700 meters of strongly reflecting
sediments that data from piston core 6609-20 (Figure 6) suggests are
probably turbidites. As in Profile AA', basement cannot be seen in this
profile.

The reflection record (Figure 11) clearly shows that this western
end of the Blanco ridge is cut by normal faults. They are indicated
by both offset seismic reflectors and by scarps apparent in the present
topography. Because the west side of these faults is the hanging wall
(i.e., the west side is down relative to the east side), these faults
support the previous sedimentological evidence of subsidence in Cascadia
depression.

Uplifted Ridge

On page 19 of this paper an area of vertical uplift was described
that was based on sedimentological evidence from a piston core obtained
from the top of an abyssal hill east of Cascadia depression (Figure 6).
Profiles BB' and CC' cross this abyssal hill and show a thick section of
reflecting sediments that suggest uplift, which is in complete agreement
with the sediment data.

Profile BB' crosses the west end of the uplifted abyssal hill (Figure 12). The region to the north of the hill is essentially sediment-free terrain of probable volcanic origin. The profile crosses Cascadia gap to the south, a steep, narrow trough through which Cascadia Channel flows. South of Cascadia gap is the Blanco ridge which, in this profile, appears to be composed of volcanic material. The profile clearly shows that the abyssal hill is composed of about 500 meters of very strongly reflective sediments, interpreted to be turbidites.

The large scarp in the sediments on the north side of the hill probably represents a loss of material through submarine slumping since reflecting horizons in the hill terminate abruptly at the scarp. The hill itself appears to be a block of sediment composed of turbidites that were originally deposited by Cascadia channel and subsequently uplifted and folded. As was pointed out earlier, these sediments are at least 450 meters higher than the regional depth of turbidite sedimentation from Cascadia Channel in this area. The rate of uplift as documented by Duncan (1968) from sedimentological evidence is on the order of 2.0 cm/year.

Profile CC', a north-south line (Figure 13), crosses the uplifted abyssal hill 9 km farther east than Profile BB'. This profile also crosses Cascadia gap and the Blanco ridge to the south. The reflection profile shows moderately reflective sediment on top of the hill which have been folded and uplifted. The Blanco ridge (south of Cascadia gap) has no apparent reflectors in it. This implies that the ridge is composed of basalt and, therefore, no direct evidence of uplift
Figure 12. Seismic reflection profile BB'. Profile BB' portrays the uplifted abyssal hill adjacent to Cascadia depression. It is composed largely of uplifted turbidites from Cascadia channel. See Figure 9 for location.
Figure 13. Seismic reflection profile CC' shows the uplifted turbidites seen in figure 14 and also shows deformed turbidite sediments on the south side of the Blanco ridge. Note the stratigraphic thickening in this subsiding trough. See Figure 9 for location.
can be documented here. The steepness of the ridge and speed of the ship, however, could also make the detection of sediment on top of this ridge very difficult.

The sediment basin south of the Blanco ridge is shown in profile CC'. There is more than 700 meters of very reflective sediment present in this basin. The sediment exhibits slight thickening at the center of the basin and obvious structural deformation. The tectonic implications of this basin will be discussed later.

**Blanco Ridge**

The Blanco ridge is the most prominent positive topographic feature of the Blanco fracture zone (Figure 2). It is approximately 150 km in length and averages about 5 km in width. Profile A'B' shows that the western end of the Blanco ridge is normally faulted and is at least partially composed of turbidite sediments (Figure 11). Profile DD' and Line 28, however, provide excellent evidence to show that the ridge is a fault bounded, uplifted ridge, that may have been formed by oblique compression across the Blanco fracture zone.

Profile DD' (Figure 14) is a north-south profile which starts at the regionally deposited spill-over deposits of Cascadia Channel to the north, crosses a series of small hills, then crosses Cascadia Channel and finally the Blanco ridge to the south. The spill-over sediments to the north of Cascadia Channel in Cascadia Basin, show the typical excellent reflectivity of turbidite sediments. The deposits were confirmed to be of turbidite origin by Duncan (1968). Acoustic basement can be clearly seen beneath these sediments and is interpreted to be
Figure 14. Seismic reflection profile DD'. Profile DD' illustrates the uplifted and faulted nature of the Blanco ridge as shown by the uplifted sediment bench on the north side of the ridge. See Figure 9 for location.
oceanic basalt (Layer 2). In contrast, basement is not detected beneath Cascadia channel. There is more than 650 meters of sediment present in the Channel.

A small volcanic hill protrudes through the turbidites south of Cascadia Channel. Between this small hill and the Blanco ridge is a section of highly reflective sediments which are somewhat deformed and dip noticeably towards the north. These sediments are most likely turbidites that have been uplifted about 200 meters above the regional depth of turbidite deposition. This suggests that this portion of the Blanco ridge was uplifted and the overlying sediments with it. There is a noticeable bench produced in the topography of the Blanco ridge by these uplifted sediments. Figure 15 shows a series of bathymetric profiles from several OSU cruises which cross this part of the Blanco fracture zone. These profiles clearly show that the bench formed by the uplifted sediments is a linear feature some 40 kilometers in length.

Line 28 (Figure 16) is a portion of a single channel reflection profile, acquired by the NOAA vessel Surveyor in 1971, which is oblique to the Blanco fracture zone structures. All of the profiles acquired by NOAA, during this cruise, show a consistent loss of definition of shallow reflectors and better penetration to acoustic basement. This is probably due to the use of larger airguns compared to those that produced the OSU profiles and to the lowpass filtering (31-98 Hz) used by NOAA during the recording of the data. Line 28, in Cascadia Channel shows more than 900 meters of sediment present (assuming an average velocity of 1.8 km/s).
Figure 15. Bathymetric profiles showing the presence of a bench believed to be the uplifted and faulted sediment wedge seen in figure 16. This bench can be traced for more than 40 km in a series of profiles (1-4). See Figure 9 for location.
Figure 16. Seismic reflection profile Line 28. It shows the faulted, uplifted Blanco ridge and uplifted sediment wedges on the north and south sides. On the south side of the Blanco ridge a subsiding trough is present overlying the Pacific plate, which is also subsiding and possibly subducting beneath the Juan de Fuca plate. See Figure 9 for location.
The extremely irregular nature of the topography in the channel is caused by the oblique crossing over several meanders in the channel at this location. Many of the features seen in Profile DD' (Figure 14) can also be observed in Line 28. The uplifted section of turbidite sediments is clearly visible on the north side of the Blanco ridge. On Line 28, however, the nature of the Blanco ridge is better defined. Acoustic basement can be observed below the uplifted sediments suggesting that the Blanco ridge is composed of uplifted oceanic basalt. The sediments which are uplifted along with the ridge are now more than 400 meters above the regional depth of sand turbidite deposition which is shown on the eastern portion of Line 28.

The south side of the Blanco ridge also contains several very interesting features (Figure 16). Line 28 shows the depth-to-basement on what is actually the Pacific plate. This profile indicates that the Pacific plate is becoming markedly deeper as it approaches the Blanco fracture zone. The sediments overlying the Pacific plate in this area show a dramatic stratigraphic thickening and some structural deformation adjacent to the Blanco ridge. These same features are shown in profile CC' (Figures 9, 13) made perpendicular to the fracture zone. Line 28 and Profile CC' both show that the Pacific plate is actively subsiding adjacent to the Blanco fracture zone. Subsidence is indicated by the stratigraphic thickening of the sediments in the axis of this structural trough. The near-surface sediments were determined to be turbidites from piston cores taken by Duncan (1968).

In contrast to the subsidence present in this trough there is a
small wedge of sediments plastered onto the south side of the Blanco ridge (Figure 16). While these sediments appear to be transparent, they are probably turbidites, since the sediments in the subsiding trough south of the ridge show excellent reflectivity in OSU profiles (Figure 13). This suggests that their transparent nature in Line 28 is due to the technical features of the NOAA profiles discussed above. The small sediment wedge, then, appears to have been uplifted and deformed from its original position on the abyssal plain. On the south side of the Blanco ridge in Profile CC' (Figure 13) there is a distinct break in slope which could represent a similar uplifted wedge of sediment seen in other profiles. The amount of uplift is on the order of 300 meters above the present day sediment level of the abyssal plain.
DISCUSSION

The present study of the Blanco fracture zone indicates that there are several anomalous features present in its morphology and structure. The fracture zone has a very complex morphology when compared to the much simpler Atlantic or Mendocino type fracture zones (Figure 3). The Blanco fracture zone shows evidence of intense uplift and subsidence accompanied by faulting and folding of the oceanic crust and overlying sediments. Several hypotheses are proposed here to explain the evolution of the morphology and structure of the Blanco fracture zone. They are outlined below and evaluated on the basis of the geologic and geophysical data available for the Blanco fracture zone. An evaluation of the processes occurring at the Blanco fracture zone may provide a better understanding of the interactions between the lithospheric plates in the northeast Pacific ocean.

**Hypotheses on the Origin of the Morphology and Structure of the Blanco Fracture Zone**

**Submarine Volcanism**

Many of the positive topographic features on the ocean floors, such as seamounts, are believed to be the direct result of submarine volcanism of the type described by van Andel and Ballard (1979). It is conceivable that the ridges and abyssal hills of the Blanco fracture zone could also have formed by extrusive, submarine volcanic activity.

If this mechanism were active in the Blanco fracture zone there should be basaltic volcanic glass present in the sediments, as was found by Duncan (1968) on the Gorda ridge. There should also be
evidence of Fe/Mn oxide enrichment which Bonatti et al. (1976) have stated commonly occurs on mid-ocean ridges. To be consistent with van Andel and Ballard's (1979) observation and interpretation of the Galapagos rift, which were based on high resolution bathymetry and submersible observations, the fracture zone ridges should be discontinuous volcanic centers or cones, sometimes with the flanks of one or more cones merging together. Under this hypothesis the movement of material upward and into the volcanic centers along the fracture zone could produce basins and depressions via the "head loss" mechanism of Sleep and Biehler (1970), where the movement of fluid upwards along the walls of a fracture or conduit is accompanied by a loss of hydraulic head in the central part of the conduit. This loss of hydraulic head causes formation of a basin or depression in the center of the conduit.

**Serpentinite Intrusion**

The ridges of the Blanco fracture zone may also be the result of the intrusion of mantle-derived serpentinized peridotite into the shallow crust, similar to the model invoked for the Vema fracture zone by Bonatti and Honnorez (1976). This intrusion of altered mantle material could be accompanied by minor submarine volcanism and deep hydrothermal circulation and related metallogenesis. The sediments overlying the area of uplift may show piercement structures, uplift and deformation on the flanks of the ridges. This type of intrusion may expose large bodies of ultramafic rocks and may be associated with distinctive gravity and magnetic anomalies which Bonatti and Honnorez (1976), describe is common in Atlantic fracture zones.
Alteration and Expansion of Underlying Mantle

Very deep fracturing may allow fluids to circulate to mantle depths (Lister, 1972, 1974) causing hydration and serpentinite formation below the Blanco fracture zone. Gumma (1974) has postulated this type of mechanism for the Rivera fracture zone in the eastern Pacific ocean based on his interpretation of the gravity and magnetics of the area.

This process involves expansion of the mantle underlying the fracture zone due to serpentinization of peridotite. This expansion would cause uplift across the two structurally independent plates resulting in crustal dilatation and the generation of a void space between the plates. The uplift would also cause faulting along the areas near the void forming a central valley bordered by raised parallel ridges (Gumma, 1974). Deep hydrothermal circulation would probably also be accompanied by precipitation of Fe/Mn oxides (Bonatti et al., 1976) and anomalous heat flow.

While individual features, such as troughs and ridges, may exhibit large gravity anomalies, the overall anomaly would suggest that the area is compensated at depth as Gumma (1974) indicated for the Rivera fracture zone.

Compressional Tectonism

Faulting, folding and uplift along the Blanco fracture zone could be a response to a component of compression normal to the fracture zone. This component of compression could result from a clockwise rotation of the spreading directions of the offset ridge segments (Wilson, 1965; Menard and Atwater, 1969; Dewey, 1975; Fox et al., 1976) or, in the case of the fracture zones of the Juan de Fuca plate, from differential
northward motion of the Pacific plate relative to the Juan de Fuca plate.

Evaluation of Hypotheses

Submarine Volcanism

Submarine volcanism is not considered as a viable hypotheses for the evolution of the morphology and structure of the Blanco fracture zone for the following reasons:

- With the exception of the juncture of the Gorda ridge with the Blanco fracture zone, the sediments of the fracture zone show no indication of recent volcanic activity (Duncan, 1968).
- In at least two cases, the Blanco ridge and the uplifted abyssal hill (pg. 27), positive topographic features are shown to be either partially or completely composed of turbidite sediments (Figs. 11, 12, 13), which has been uplifted 250 to 450 meters at 2.0 cm/year (Duncan, 1968).
- The Blanco ridge was shown by Duncan (1968), to be a continuous, narrow ridge. This differs from the volcanic ridges studied by Van Andel and Ballard (1979) which were made up of discontinuous volcanic cones with one or more cones sometimes merging together.
- Seismic reflection profiles show the Blanco ridge to be fault bounded and uplifted (Figs. 14, 16).

While submarine volcanism does not appear to be the origin of the major features of the central portion of the Blanco fracture zone, it may be an important process at the junctures of the fracture zone with the Juan de Fuca and Gorda spreading ridges based on the presence of basaltic glass in cores near the Gorda ridge (Duncan, 1968). Also, some
of the small seamounts seen in reflection profiles, such as in Figure 14, may be volcanic in nature, although it is likely that they formed much earlier when the plate was nearer to the spreading ridge.

**Serpentinite Intrusion**

While the intrusion of the serpentinized peridotites into the shallow crust appears to be common in Atlantic fracture zones, this mechanism does not appear to be occurring in the Blanco fracture zone on the basis of the following evidence:

- Structural deformation of the sediments in the Blanco fracture zone is not confined to the flanks of uplifts, as would be expected for simple piercement structures. Instead, folding of sediment extends for several kilometers (5-10 km) away from uplifted features (Fig. 13).
- The large gravity and magnetic anomalies expected for such intrusive bodies are not seen in the Blanco fracture zone.
- No ultramafic rocks have been recovered along the Blanco fracture zone, whereas ultramafic rocks are common in Atlantic fracture zones (Bonatti and Honnorez, 1976).
- Uplift rates of 2.0 cm/yr have been documented for areas of the Blanco fracture zone. This is an order of magnitude larger than the 1 mm/yr rate of serpentine intrusions along the mid-Atlantic ridge axis, calculated by Bonatti and Honnorez (1976).

Although no ultramafic rocks have been recovered from the Blanco fracture zone, Duncan (1968) recovered greenstone fragments in a piston core south of the Blanco ridge and Melson (1969) dredged greenstones from a ridge in the same area. This provides evidence of shallow
crustal movement, exposing rocks buried 750 to 1000 meters (K. Scheidegger, personal communication), but does not necessarily imply exposure of more deeply buried material.

**Alteration and Expansion of the Underlying Mantle**

The serpentinization and resulting hydration of the mantle underlying the Blanco fracture zone does not appear to be a major process controlling the morphology and structure in the central portion of the fracture zone. The following characteristics of central portion of the Blanco fracture zone are inconsistent with this model:

- Individual topographic features, such as the Blanco ridge, show no apparent relation to the gravity of the area (an exception is the deep troughs at the ends of the fracture zone).
- The morphologic features of the Blanco fracture zone are confined to a long, often very narrow zone (4 kilometers). This would appear inconsistent with a broad area of uplift (30 km) due to deep swelling of the mantle as Gumma (1974) has shown may exist in the Rivera fracture zone.
- There is no central deep trough in the mid-portion of the Blanco fracture zone, which is a characteristic of the model presented by Gumma (1974), where the central trough forms by the uplift and separation of the two adjacent plate edges.
- The sediments in the central portion of the fracture zone show no evidence of hydrothermally enriched Fe-Mn deposits (Duncan, 1968; Phipps, 1974), as would be expected in a zone of intense, hydrothermal circulation and alteration (Bonatti et al., 1976)
Deformation of sediment, especially the folding of sediments seen in Figure 13 and 16 is consistent with the broad (30 km) uplift and related extension faulting and downdropping of blocks which is used by Gumma (1974) to explain some morphologic aspects of the Rivera fracture zone.

Bottinga and Allegre (1976) have summarized the geophysical, petrological and geochemical models of the ocean lithosphere and suggest that extensive hydration at upper mantle depths beneath mid-ocean ridges is unlikely based on present plate tectonic thermal models and the projected temperatures that would be present at depths of 35 kilometers. Bottinga and Allegre (1976) attribute the seismically observed wedge of upper-mantle material close to the ridge axis which shows increased absorption and low velocities as being produced by decomposition of uprising asthenospheric material.

Isotopic composition of oxygen from some ophiolites indicate a meteoric origin for the water, rather than a seawater origin (Lewis, 1978). This would seem to support Bottinga and Allegre's (1976) objection to extensive hydration of the upper mantle.

Lewis (1978) states that less than 1% water is available in the water/rock fracture system to cause serpentinization. This low amount of water would probably require convective cooling waters, flushing out some products of serpentinization (Lewis, 1978) and would probably also require long time intervals for extensive serpentinization.

The large time scale which is probably involved in extensive serpentinization is contrary to Gumma's (1974) hypothesis, which
calls for relatively rapid expansion rates to account for
topography being formed at the ridge-fracture intersection.

While the above data seem to rule out this hypothesis for the
central portion of the fracture zone, there is insufficient data
available at the junction of the Blanco fracture zone with the Gorda
and Juan de Fuca ridges to properly test this hypothesis for these regions,
since the morphology and evidence of hydrothermal circulation (Selk, 1978) do suggest that this hypothesis is able to account for some
of the features seen at these junctures.

Compressional Tectonism

The data presented in the previous sections of this paper appears
to be most consistent with the hypothesis that the morphology and struc-
ture of the Blanco fracture zone is created by a component of compres-
sion normal to the trend of the fracture zone. The features which
support this hypothesis are listed and discussed below:

- Many of the individual morphologic and structural features of
  the Blanco fracture zone are long, narrow, uplifted and fault
  bounded. This type of structural setting is consistent with
  compressional faulting and folding.

- Seismic reflection profiles show folded and uplifted sediments
  across a zone of possible convergence between the Pacific and
  Juan de Fuca plates (Figs. 14, 16). These sediments are de-
  formed for up to 10 kilometers away from the uplifted Blanco
  ridge, consistent with compression normal to the fracture zone.

- Figure 16 shows subsidence of the Pacific plate adjacent to
  the Juan de Fuca plate at the Blanco fracture zone. Such an
area of subsidence could be expected due to oblique convergence between these two plates.

Earthquake mechanism studies presently contain a high degree of ambiguity and cannot be used to rule out a compressional component of stress across the Blanco fracture zone.

Compression across the fracture zone could result from two possible mechanisms. Many authors have stated that the clockwise rotation of the spreading direction or spreading ridges across a transform fault with a left lateral ridge offset will result in compression across the transform fault (Wilson, 1965; Menard and Atwater, 1969; Dewey, 1975; Fox et al., 1976). Compression could also result from a differential northward motion between the Pacific and Juan de Fuca plates, resulting in convergence between the two.

The Blanco fracture zone, then, is apparently characterized by oblique convergence of the Juan de Fuca and Pacific plates. The situation can be compared to the subduction zone along the Oregon trench where the Juan de Fuca plate is underthrusting the American plate. Figure 17A is a schematic cross section drawn perpendicular to the Blanco fracture zone using information projected from Line 28 and Profile CC' (Figs. 13 and 16). This situation is strikingly similar to Figure 17B, which is a seismic reflection profile across the Oregon trench, and though the intent is not to imply large scale subduction exists across the Blanco fracture zone, it does imply subsidence of the Pacific plate may be due to partial convergence between it and the Juan de Fuca plate.
Figure 17. 17A Schematic cross-section showing deformation and possible subduction south of the Blanco fracture zone.

17B Seismic reflection profile K-L showing subduction of the Juan de Fuca plate beneath North America along the Oregon trench.

Both A and B drawn to same scale.
Earthquake mechanism studies such as those of Chandra (1974) presented above (Fig. 8), generally show strike-slip mechanisms for the Blanco fracture zone. This data can, however, be reinterpreted as shown in Figure 18. In this interpretation four of the six earthquake mechanisms are now interpreted as examples of thrust faulting, rather than strike-slip faulting. These reinterpreted earthquake mechanisms are similar to what would be expected if there were a component of compression present, normal to the Blanco fracture zone, superimposed on a dominantly strike-slip regime (Fig. 19). The reinterpretation of the earthquake mechanisms of Chandra (1974), shown here, are not intended as proof that compression normal to the Blanco fracture zone exists. Rather, this analysis is intended to show there is a large degree of uncertainty present in the available data set and that several interpretations are possible. In fact, Bolt et al. (1968) obtained normal fault solutions for events 15 and 18.

It is clear that either a clockwise rotation in spreading direction of the Juan de Fuca - Gorda ridge system or differential absolute motion between the Pacific and Juan de Fuca plates can provide a compressional component to the stress regime of the Blanco fracture zone.

Figure 20A shows the initial, orthogonal configuration of a given transform fault and spreading ridge pair. The trend of the transform fault can be rotated in a clockwise direction by a change from perpendicular spreading to oblique spreading (Fig. 20B) or by an actual rotation of the spreading ridges themselves (Fig. 20C).

Atwater (1970) reconstructed the past motions of the Pacific and Juan de Fuca plates from magnetic lineations. She showed that in the last 5 million years the Blanco fracture zone has undergone clockwise
Figure 18. Reinterpreted earthquake mechanisms (shown in figure 8) for the Blanco fracture zone. Events 16 and 17 remain unchanged from Chandra's (1974) original interpretation. All other events can be interpreted to show a component of compression across the fracture zone. (original data from Chandra, 1974)
Figure 19. Two possible patterns for earthquake mechanisms on the Blanco fracture zone. Figure 19A illustrates the expected pattern for simple strike-slip motion; 19B represents the expected pattern for oblique compression. In both figures the shaded area represents compression and the white dilation.
Figure 20. Compression across a left lateral, originally orthogonal (A) transform fault as the result of a clockwise rotation of; the spreading direction (B) or; the spreading ridges (C).
rotation, as has the spreading direction of the Juan de Fuca and Gorda ridges. More recent studies by Riddihough (1977) and Delaney (personal communication), agree with this clockwise rotation of the Juan de Fuca - Gorda spreading system. The very recent structural features described in the present study may imply that this rotation is still occurring at present.

Another reason for expecting compression across the Blanco fracture zone is illustrated by studies of the Mendocino fracture zone to the south. On the basis of several types of geophysical and geologic data (gravity, earthquake studies, bathymetry) (Seeber et al., 1970; Dehlinger et al., 1970; Silver, 1971a, 1971b; Couch, 1980) it appears that the Pacific plate is overriding the Gorda - Juan de Fuca plate in the area of the Mendocino fracture zone (Fig. 21). If we assume that the oceanic crust south of the Blanco fracture zone is not decoupled from that part of the Pacific plate south of the Mendocino fracture zone, and that there is no north-south extension in the Gorda basin, then if overriding and subduction are occurring at the Mendocino fracture zone due to differential northward movement of the Pacific plate relative to the Juan de Fuca plate, convergence and subsidence should also be occurring at the Blanco fracture zone. The above assumptions are consistent with all interpretations of the plate geometry in this area.

It is not immediately clear how or if the two mechanisms described above are related. One possible clue to the relation between these mechanisms lies in the overall trend of the ridges and fracture zones of the Juan de Fuca plate compared to the trends of the San Andreas transform fault and Queen Charlotte Islands transform fault. These two very large transform faults represent the direction of motion of the Pacific
Figure 21. The general tectonic features of the northeast Pacific ocean illustrating the probable area of overriding of the Juan de Fuca plate by the Pacific Plate.
plate relative to the North American plate. If one looks at the trends of the transform faults of the Juan de Fuca plate, from south to north (i.e., Mendocino, Blanco, Sovanco, Revere-Deiwood), a noticeable clockwise rotation in these fracture zones can be seen (Fig. 21). It seems as though the fracture zones of the Juan de Fuca plate are gradually becoming more aligned with the trends of the San Andreas and Queen Charlotte Islands transform faults. The shorter, younger transforms to the north are more closely aligned than the longer, older Blanco and Mendocino fracture zones to the south. The realignment of the transform faults of the Juan de Fuca plate is probably a result of this small, young Juan de Fuca plate being caught in a very unstable position between the motion of the Pacific plate on one side and the North American plate on the other.

**Extension in Cascadia Depression**

While much of the data presented here can be interpreted to represent compressional tectonics normal to the fracture zone, there are additional data suggesting extension in the same area, that is, Cascadia depression. Profile AA' (Fig. 10) shows that Cascadia depression is an actively subsiding basin with sediments thickening markedly towards the center of the basin. The area exhibits extensive normal faulting (Fig. 10, 11) and Duncan (1968) has documented sedimentological evidence which suggested that the depression was subsiding at a rate of about 1.0 cm/yr. The question which arises is how can this subsidence (extension) be explained in a partly compressional strike-slip stress regime?

If one studies the bathymetry of the Blanco fracture zone (Fig. 2)
to determine its trend or find the location of the major strike-slip fault, it is apparent that the fracture zone is actually composed of two en echelon fault zones, offset by approximately 18-20 km (Figs. 22A, Fig. 2). The offset is located in the position of Cascadia depression. This significant offset of the Blanco fracture zone in conjunction with the relative motions of the Juan de Fuca and Pacific plates will produce a small area of extension (Fig. 22A). This type of structural setting is observed in continental strike-slip faults. Death Valley, California is a large continental example of this type of extensional tectonics in a transcurrent fault (Fig. 22B) (Burchfiel and Stewart, 1966).

Directions for Future Studies

In order to gain a more complete understanding of all the processes occurring along the Blanco fracture zone and how they are related, several types of additional studies are needed in this area. Geophysical investigations, such as seismic refraction and heat flow measurements, would be extremely useful and would add much to the knowledge of the area. Geologically, further studies of the composition and history of the topographic features along the fracture zone would be important. This would require intense sampling programs, including dredging for basement outcrops and additional attempts to collect sediment samples, especially from the tops or flanks of topographic highs. To date, this has only been done on one abyssal hill which is described in this study and was described by Duncan (1968).
Figure 22. 22A Schematic diagram showing en echelon nature of the Blanco fracture zone. The stress field shown will produce oblique compression along the fracture zone and extension in Cascadia depression.

22B This figure shows a comparable stress regime for a continental transcurrent fault, Death Valley, California. (after Birchfiel and Stewart, 1966)
SUMMARY AND CONCLUSIONS

The Blanco fracture zone is approximately 460 km long and consists of a series of subparallel, elongate ridges and troughs. The fracture zone is a northwest-southeast trending feature which has a small north-south offset present in the area of Cascadia depression. The Blanco fracture zone has been described as being different from other fracture zones. Special attention was given to its anomalous trend (with respect to other major Pacific fracture zones) and complicated bathymetry. It is now clear that most of the anomalies of this transform fault can be explained within the framework of basic plate tectonic theory as it pertains to rigid plates.

The Blanco fracture zone exhibits sedimentologic and structural characteristics which indicate that the fracture zone is not a simple transform fault, but has a component of compression normal to the fracture zone superimposed on its strike-slip stress regime. Sedimentological data have shown that in the area north of Cascadia gap, turbidites originally deposited by Cascadia Channel have been uplifted more than 450 meters at an average rate of 2.0 cm/year. Several single channel seismic reflection profiles across the fracture zone indicate uplift, faulting and deformation of sediments and oceanic crust, and convergence and subsidence along the southern boundary of the Blanco fracture zone. While this component of compression explains the features of the central portion of the fracture zone, it is recognized that other factors may be of importance at the fracture zone - spreading ridge junctures, in addition to compressional tectonics.

Subsidence in the area of Cascadia depression is explained as the
result of a small offset in the trend of the fracture zone forming two en echelon faults (Figure 22A). This offset in a dominantly strike-slip regime forms a small area of extensional stress. It is not clear, at present, what the eventual fate of this type of feature will be. A reasonable assumption is that, in a simple strike-slip regime, the area would become a probable area for magmatic intrusion. It would possibly even represent the beginnings of a new, small spreading ridge segment. When the compressional stress present in the Blanco fracture zone is considered, however, it is also quite possible that this offset will be a very ephemeral feature of the fracture zone.

The compressional component present across the Blanco fracture zone is probably a result of the differential relative motion between the Pacific and Juan de Fuca plates. It appears that the Pacific plate is moving northward at a rate which is greater than that of the Juan de Fuca plate, resulting in compression and possible overriding or subduction along the southern boundaries of the Juan de Fuca plate. The consequence of the compressional stress which is being exerted on the Juan de Fuca plate, is the realignment and clockwise rotation of its associated transform faults and spreading ridges. Once these features have been rotated to a position where the transform faults are aligned with the small circle described by the trend of the Queen Charlotte Islands and San Andreas faults, then the plate geometry in this part of the Pacific Ocean will become more stable.
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