

Subduction Zones and Related Volcanism in Northwestern South America

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ABSTRACT

The Eastern Equatorial Pacific is subject to spreading mechanisms from the ocean floor leading to subduction. A hotspot is observed to be the cause of the Galapagos Islands as well as other aseismic features on the segmented plates being subducted. A portion of the ocean crust has been obducted along the Colombia-and-Ecuador coast resulting in the Western Cordillera. Some ophiolite studies and rock sample measurements show agreement with marine refraction data. A synthetic seismic model contributes to know the oceanic crustal structure. Volcanic rocks in the northern part of the Andes are correlated with dipping of the Benioff zone showing agreement between the andesitic volcanism observed and subduction actually taking place in this complex area of northwestern South America.

Keywords: Colombia, Ecuador, hotspots, subduction, Benioff zone, vulcanism, Galapagos Islands.

ZONAS DE SUBDUCCIÓN Y VULCANISMO ASOCIADO EN EL NORORIENTE DE SURAMÉRICA

El oriente del Pacífico ecuatorial está sujeto a mecanismos de separación en la placa oceánica generando subducción. Se observa que un *hotspot* es la causa de la formación de las islas Galápagos así como de otras características asísmicas en las placas segmentadas que están siendo subducidas. Una parte de la corteza oceánica ha sido abducida a lo largo de las costas de Colombia y Ecuador dando origen a la cordillera occidental. Algunos estudios de la ophiolita y muestras de roca indican consistencia con datos de refracción marina. Un modelo sísmico sintético contribuye al conocimiento de la estructura de la corteza oceánica. Las rocas volcánicas en la parte norte de la cordillera de Los Andes están correlacionadas con la inclinación de la zona de Benioff la cual muestra coherencia entre el vulcanismo andesítico observado y la subducción que se está presentando en esta compleja área del noroccidente de Suramérica.

Palabras clave: Colombia, Ecuador, *hotspots*, subducción, zona de Benioff, vulcanismo, Islas Galápagos.

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INTRODUCTION

The Eastern Equatorial Pacific is considered, tectonically, a very complex area compared to regions where a simpler spreading behavior of the ocean crust is taking place at the Eastern Pacific rise and the subduction mechanism along the Chile-Peru trench.

Results from laboratory experiments and measurements in rocks, and pressure-temperature, density and synthetic seismic models are used to show how lithosphere being subducted under Colombia and

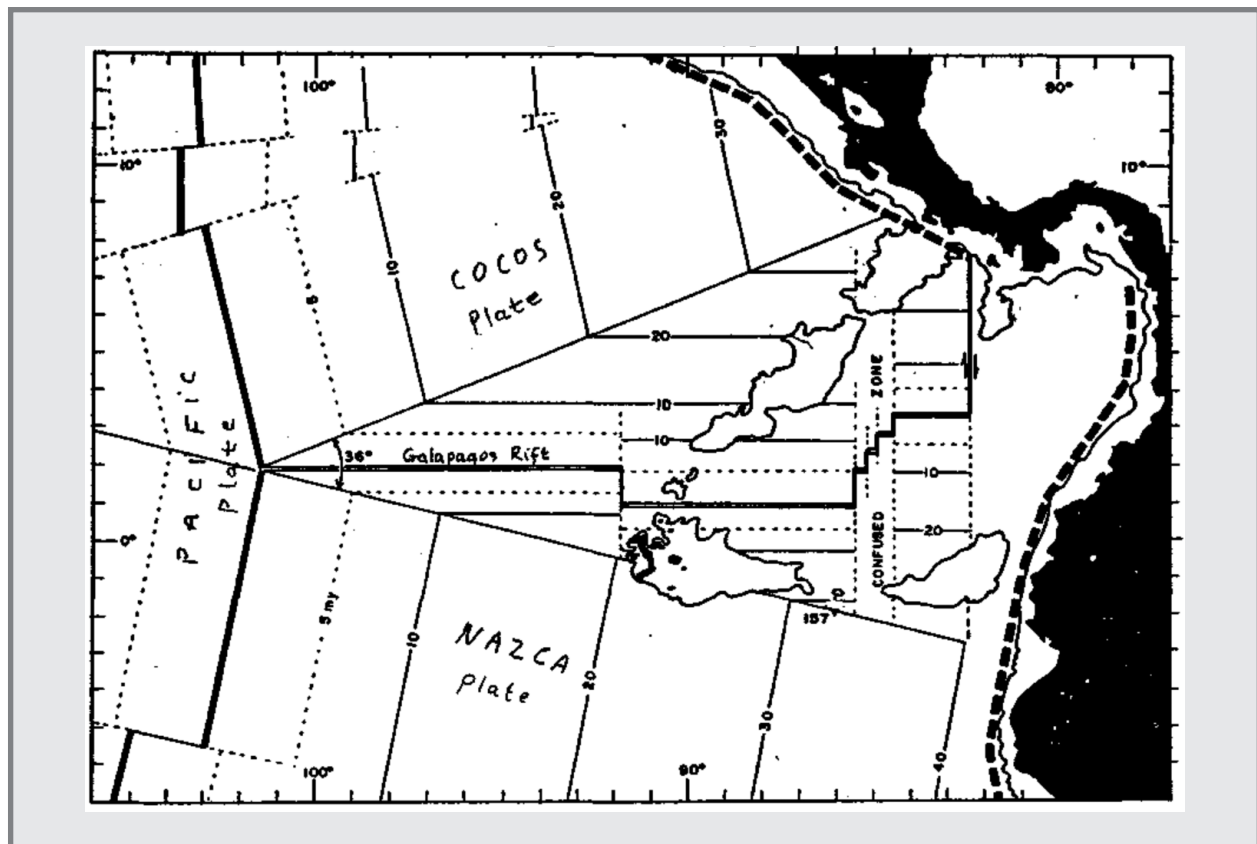
Ecuador is the cause of the formation of the volcanic chain for the Northern Andes.

REGIONAL TECTONIC SETTING

THE GALAPAGOS GORE

A complexity is observed within the Galapagos Gore (see figure 1), a triangular region extending to the eastern side of the rise Galapagos Triple junction where we find its apex.

Figure 1. The Galapagos Gore.



Source: Holden and Dietz (1972).

This area has been developed by a slow spreading rate impressing in the east-western region, oriented-topographic features with marked-magnetic anomalies.

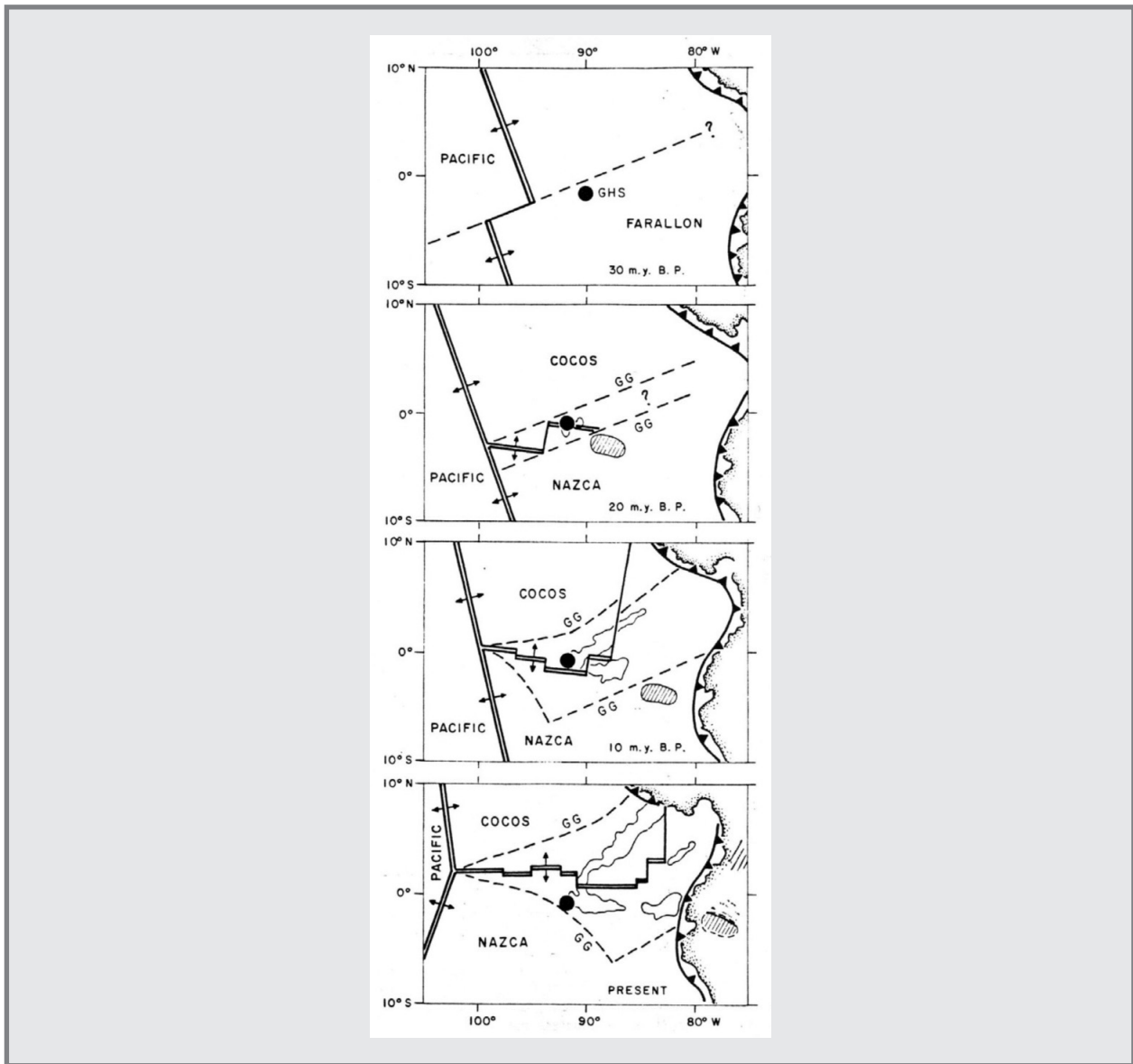
The Galapagos Gore began around 30-25 my BP by shearing the Farallon plate (Holden and Dietz, 1972). Its triangular aspect is explained by an

asymmetric spreading along the Galapagos Rift Zone, caused partly by discrete jumps of the rise axis. Splitting the Farallon Plate formed the Cocos-Nazca-Pacific triple junction, by projecting this triple junction backwards along the Nazca isochron flexure while holding the Galapagos hotspot fixed, (Holden and Dietz, 1972), which depict clearly

the evolution stages of Galapagos Gore up to its beginning (figure 2).

The Panama Fracture Zone serves as a guideline for basis of the gore with a 1200 km length. It closes the gore, limited by two sides of the triangle, both approximately 2200 km length.

Figure 2. Evolution of the Galapagos Gore.



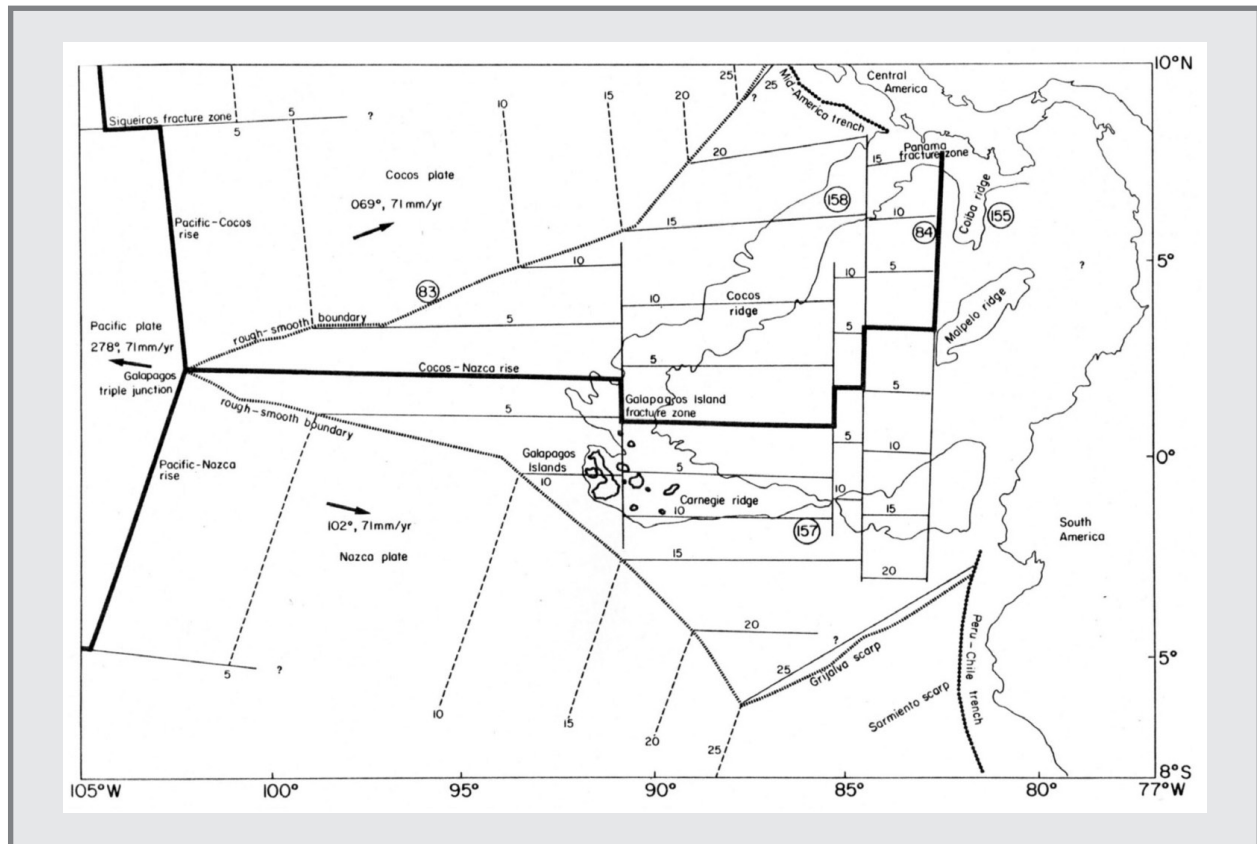
Source: Holden and Dietz (1972).

THE GALAPAGOS RIFT ZONE

The Galapagos Rift is referred by several authors using different names and the most common are Galapagos Rift Zone, Galapagos Fracture Zone, Galapagos Rift, Galapagos Rise, Galapagos Spreading

Center, Cocos-Nazca Rise, and Galapagos Ridge. The rift is assumed to have taken its origin along a zone of weakness called Grijalva Fracture Zone through which the Farallon Plate was broken. This zone can now be recognized as a scarp along the southern wing of the Galapagos Gore (see figure 3).

Figure 3. Schematic diagram of Galapagos area.



Source: Johnson et al., 1976. Schematic isochrones dating million years ago and shown by numbered lines—solid for strong magnetic anomalies areas and dashed for weak magnetic anomalies areas. Heavy lines = active plate boundaries. Light lines = inactive fracture zones. Hachured lines = rough-smooth boundaries. Aseismic ridges delineated by 1.200 fm isobath. Circled numbers = DSDP holes. Azimuths and rates of plates from model PAM1 and relative to triple junction.

From Nazca-Cocos-Pacific triple junction the Galapagos rift extends to eastern to reach the Panama-Fracture Zone. It is not a continuous straight rift; it is trisected but each segment being offset at particular fracture zones, and becoming shorter as they go eastern. At their turn, they are multisectioned but the western section—called the Western Rift—is the

longest measuring 1045 km. The Central Rift has a length of 700 km, and the Eastern rift 222 km. Between the central and the eastern portions there is a discontinuity at a place called confused zone which is 144 km wide. The whole length of Galapagos Rift is, including the confused zone, approximately 2120 km.

An interesting observation is the western branch of the rift involving at Lat. 2°12' N, Lon. 101° 35' W, the deepest known axial valley in the world (Hey et al., 1972), with 5480 km depth. This section has a current spreading rate of 1.5 cm/yr and the eastern section 2.8 cm/yr.

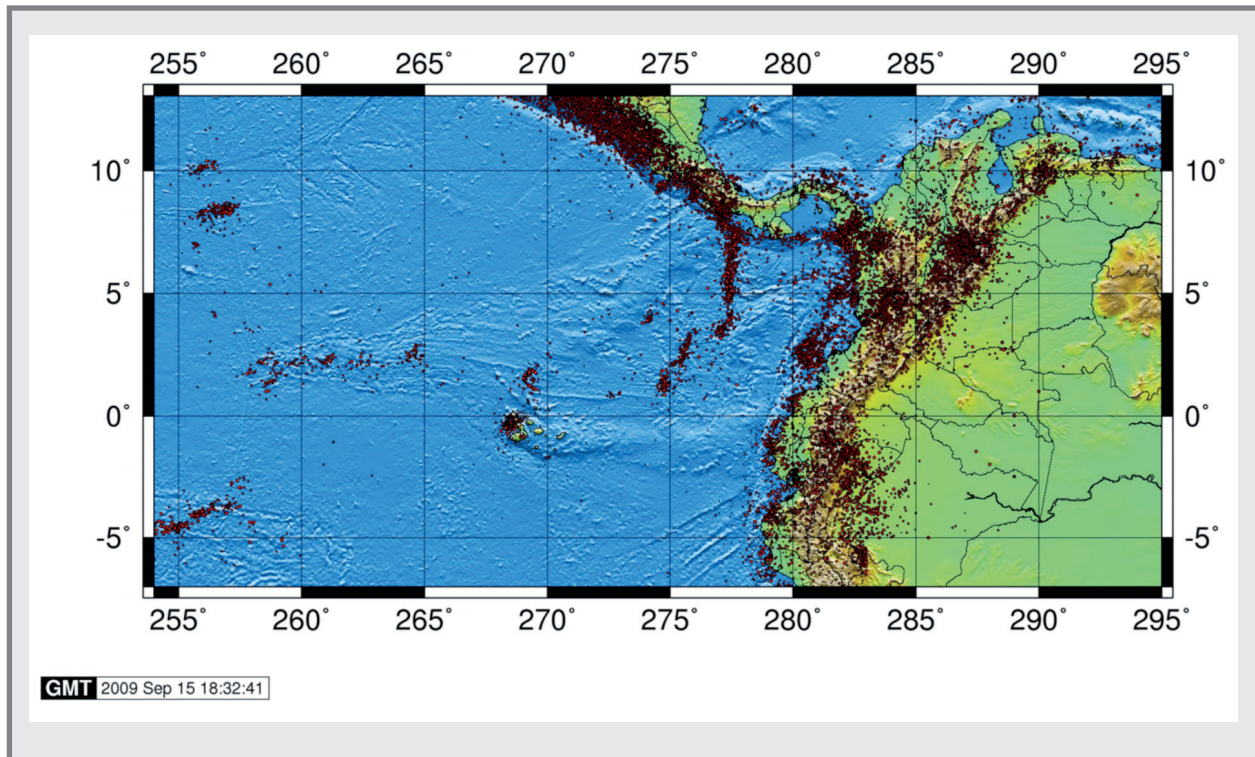
THE GALAPAGOS HOTSPOT

A hotspot, as defined by Johnson and Lowrie (1971), and gathering some elements from Morgan (1971), is a narrow plume (pipe) of mantle material rising and spreading out radially in the asthenosphere. As the oceanic plate moves over the upwelling, a continuous outpouring of basalt yields a linear aseismic ridge on the sea floor. A hotspot could be, according to Wilson (1963), the cause of oceanic aseismic

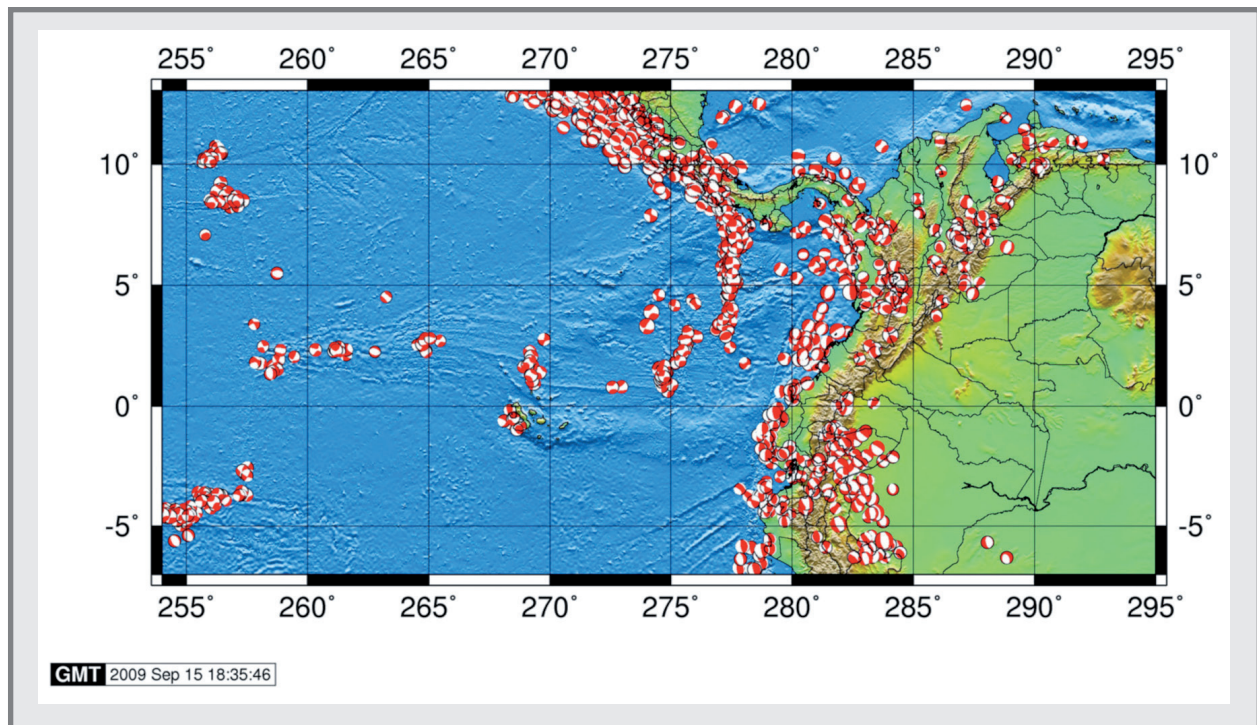
ridges. Johnson and Lowrie (1971) recognize that Galapagos Islands seem to be situated directly over a great upwelling of mantle material, and its outpouring formed these islands as well as the Cocos and Carnegie ridges (figure 4).

The Galapagos Hot Spot is given an age of 40 my BP by Holden and Dietz (1972) but they cannot sustain the evidence with data. Hey (1977) says it must be much younger than that because based on Cox and Dalrymple's (1966) analysis of K-Ar dating that show the islands are all Pliocene and younger, the hot spot could be 30 to 25 my BP. Lonsdale and Klitgord (1978) put the origin of the hot spot some 22 to 17 my BP based on magnetics, bathymetry, refraction profiles, and thicknesses of oceanic crustal layers.

Figure 4. Epicenters and focal mechanisms at the analyzed area.



Sources: Author's.



Sources: Epicenters since 1541 until 15/09/2009, from USGS (2009). Focal mechanisms since 01/01/1976 until 15/09/2009, from Harvard CMT (2009). Topography and bathymetry from University of Alaska (Lindquist et al., 2004).

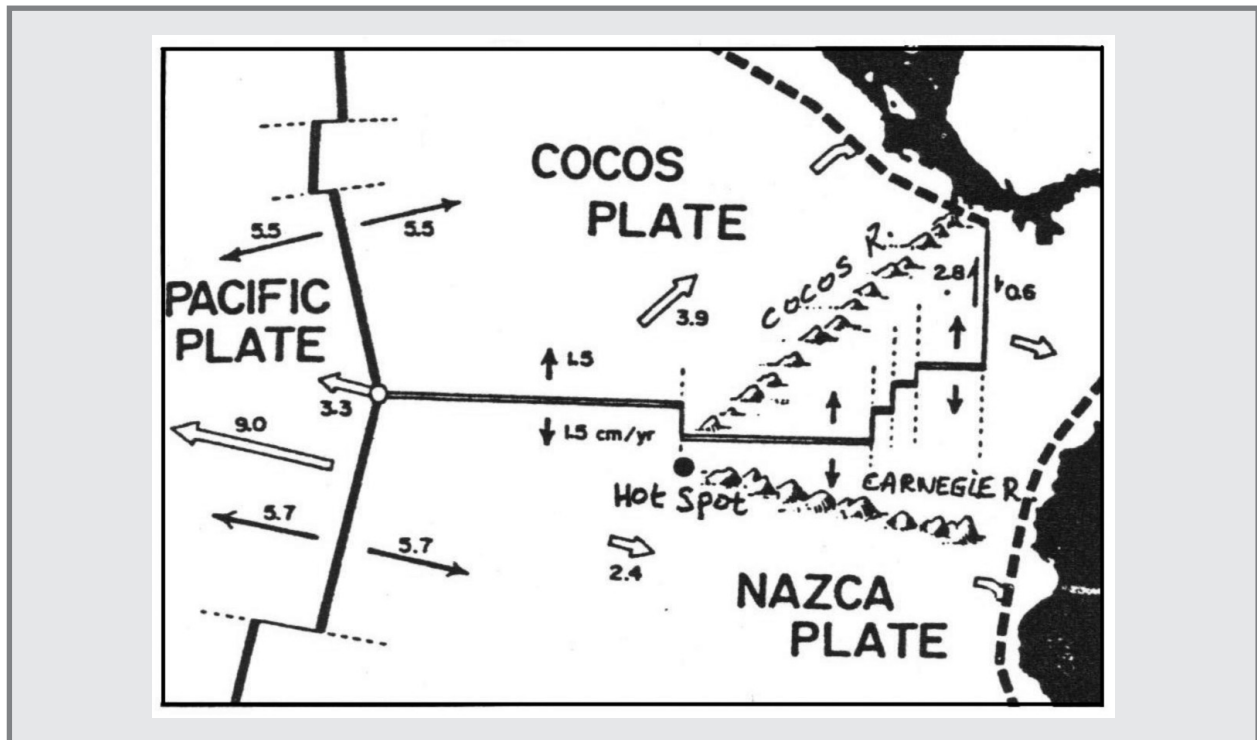
The oldest sedimentary deposits on both are almost identical; stratified sequences are very uniform, ending in a wide-spread erosion unconformity which implies uplift a faulting on both ridges.

There are two hypotheses concerning these ridges: the Ancestral Ridge and the Hot Spot hypotheses. Firstly, Van Andel et al. (1971) state both ridges were once laying side-by-side and then were separated by rifting along the central axis. Rifting started earlier in the eastern side and moved slowly

to the western, and was associated with a jumping transform fault.

Secondly, the Hot Spot hypothesis –sustained by Hey (1977)– shows that “Cocos and Carnegie Ridges were formed from outpourings of basalt from the Galapagos hot spot on to the Cocos and Nazca plates, respectively, and thus mark the azimuths of motions of these plates relative to the hot spot”. So each ridge was successively simultaneous but independent from a common origin (figure 5).

Figure 5. Formation of the Carnegie and Cocos Ridges.



Source: Hey (1977).

THE MALPELO RIDGE

Laying south-eastern and parallel to the Cocos Ridge and equidistant to both the Cocos and Carnegie Ridges, this ridge was an enigma to several researchers (figure 3). Holden and Dietz (1972) call it a possible remnant from the Cocos Ridge. Hey (1977) argues it formed the north-eastern extension of the Cocos Ridge and the Cocos Plate, so was transferred to the Nazca Plate by migration to the western of the Cocos-Nazca-Caribbean Triple junction. According to Lansdale and Klitgord (1978), the segment of the Malpelo rift had already moved away from the hot spot about 16 my BP. There are though “reported occurrences of continental rocks on Malpelo Island” (McBirney, 1970), personal communication to Van Andel), Van Andel et al. (1971).

THE COIBA RIDGE

This small ridge is located eastern side of the Panama Transform Fault against the Panama continental margin. This morphological unit differs greatly in structure and origin from all the other ridges within the Panama Basin. It is not a hot spot outpouring. As interpreted by Van Andel et al. (1971), based on East-Western traverses, it is oceanic. The western border of the Coiba Rift was uplifted about 3 km and the Southern part only 1.5 km during the late Miocene-and-Pliocene ages.

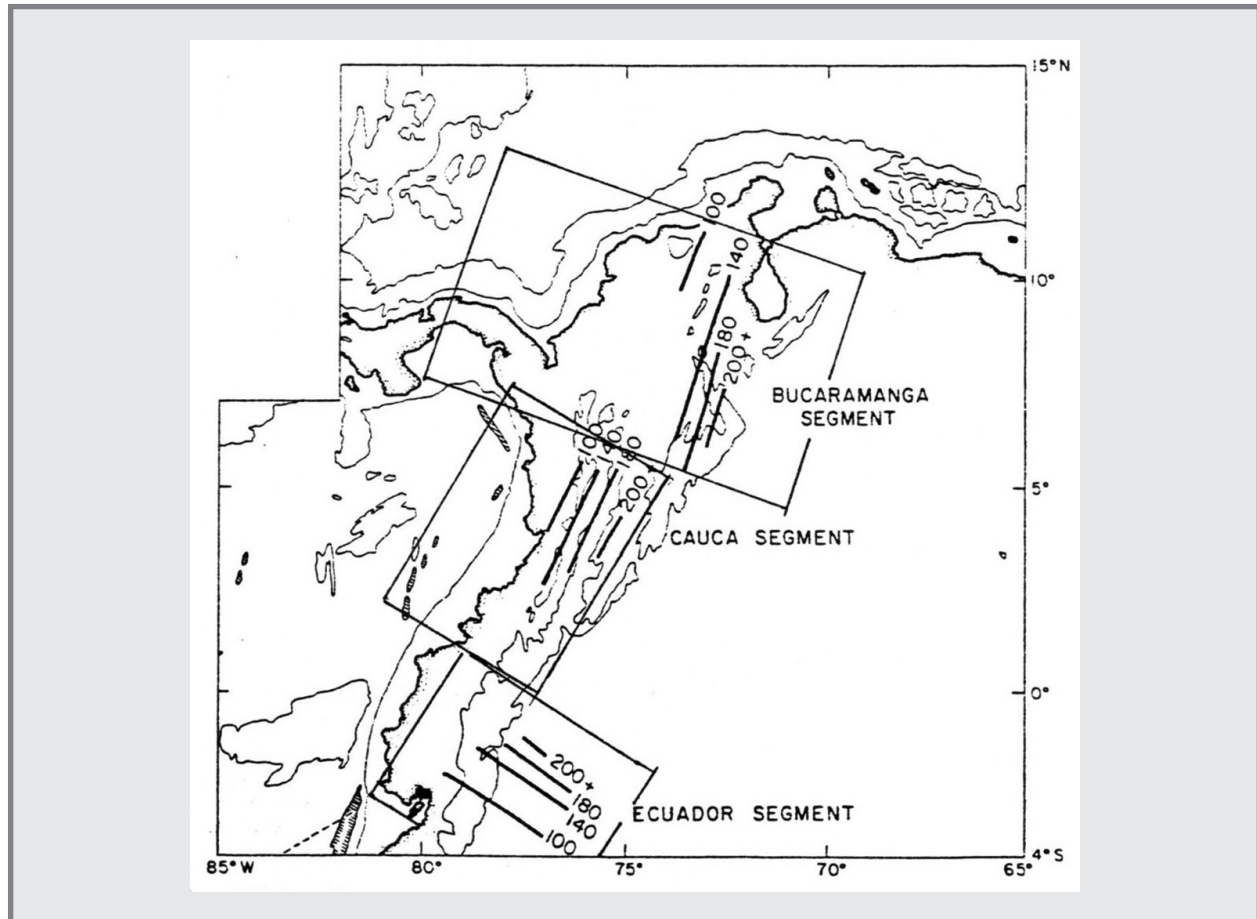
THE PANAMA FRACTURE ZONE

An important contribution by Molnar and Sykes (1969) is that Panama Fracture Zone is a portion of the

main boundary between the Cocos and Nazca plates. Van Andel (1971) has an interesting explanation showing how the eastern boundary from the Cocos Nazca plates has been shifting to the west by jumps, breaking the Cocos plate, so the broken segments of

the Cocos plate have become part of the Nazca plate. Lonsdale and Klitgord (1978) assume that broken portions of the Cocos plate instead of becoming part of the Nazca plate have formed a separate platelet.

Figure 6. Contours to the Benioff zones beneath Northwestern South America (labeled km). The three segments of subducted lithosphere identified.



Source: Pennington (1981).

THE SUBDUCTION ZONE

Plate Segments

Beneath Northwestern South America three plate segments are being subducted (figure 6) (Pennington, 1981), from regional seismicity and focal plane solutions, discriminated as follows:

- 1) The Bucaramanga segment as a portion of the lithosphere common to Panama and the Caribbean sea floor. It is subducted under the northern most part of Colombia.
- 2) The Cauca segment crosses the central and southern region of Colombia and it is oriented N

120E, dipping at 35 degrees. This is a portion of the Nazca plate.

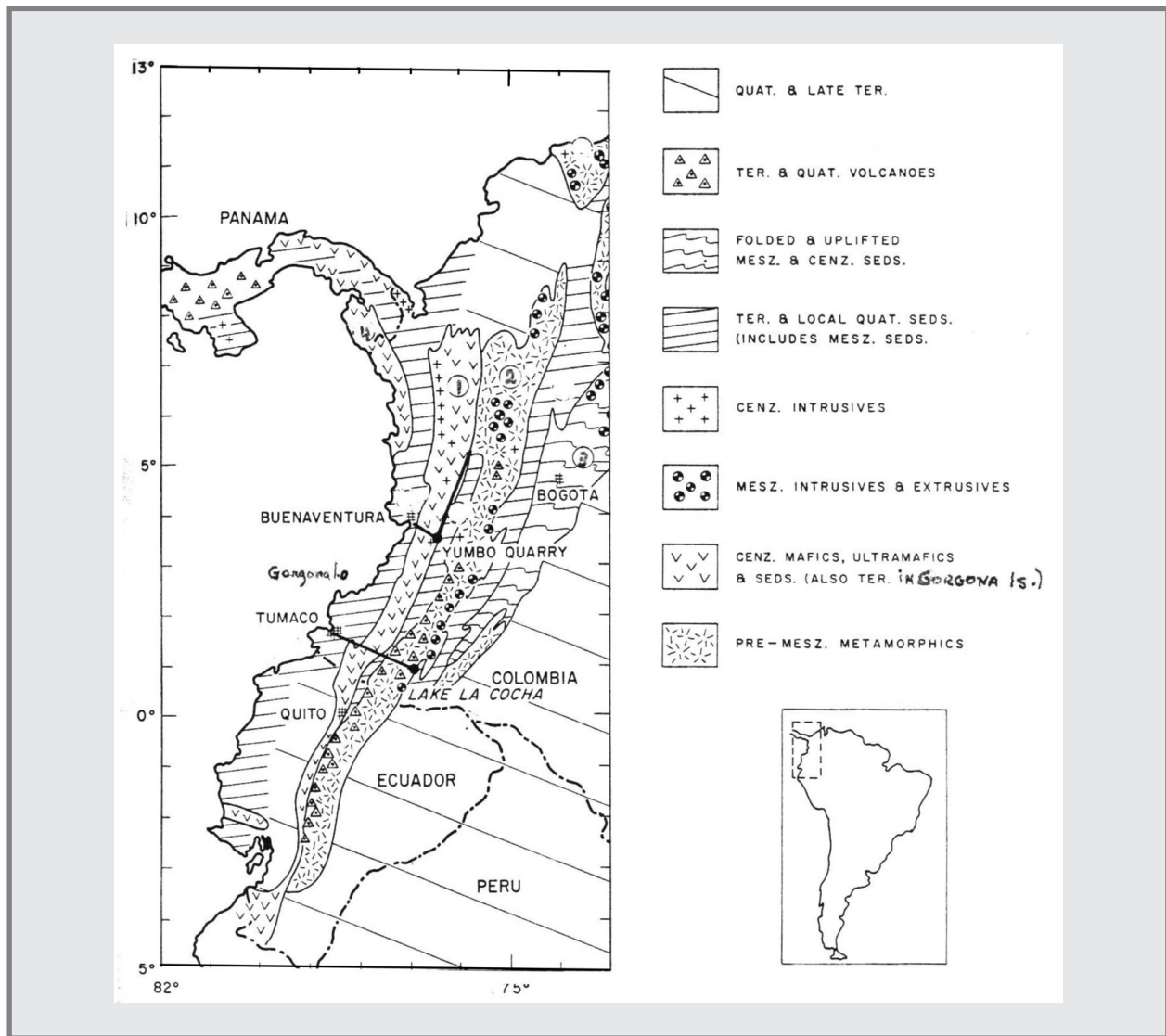
- 3) The Ecuador segment, also a part of the Nazca plate, is being subducted beneath Ecuador, with the striking peculiarity that dipping segment changes dramatically to N 35° E, keeping the same dip angle of 35 degrees.

In this study we are mainly concerned with the last two segments which are responsible for the North-Andean volcanism.

Ophiolite emplacements

The Western Cordillera, both in Colombia and Ecuador show all the features of the ocean crust (figure 7).

Figure 7. (1) Western Cordillera. (2) Central Cordillera. (3) Eastern Cordillera. (4) Coastal Serrania de Baudo. Show Ophiolites along the Western Cordillera and Serrania de Baudo.



Source: Ramírez and Aldrich (1977).

STRUCTURE AND COMPOSITION OF THE OCEANIC CRUST

Upwelling of upper mantle material partially melted flows through the Central Galapagos Rift valley. Large sheet flows of tholeiitic basalt (< 47.K20) as long as 7.0 km and bulbous pillow lavas 1-2 m diameter have been described by Van Andel and Ballard (1979). A better understanding of the composition of the oceanic crust and the facies of metamorphic rocks was obtained by means of laboratory experiments and the development of theoretical models.

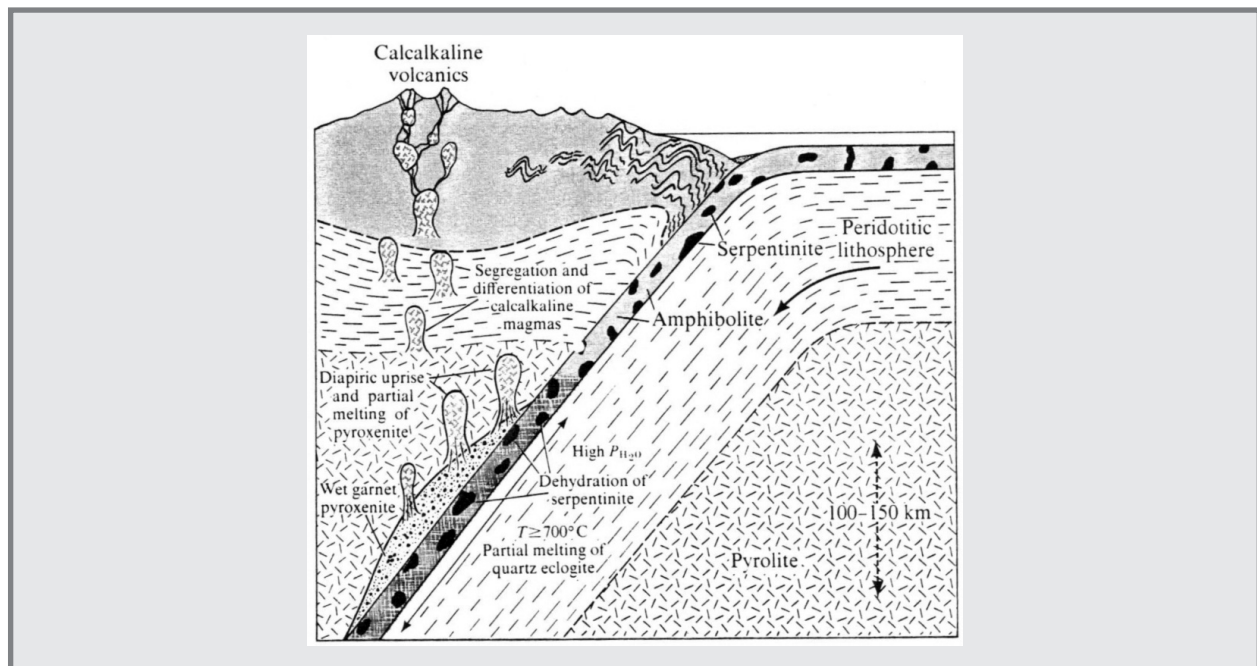
THE PRESSURE-THERMAL MODEL

It seems that processes governing the formation of new crustal active ridges lead to conditions similar to those ones conducted by experiments. Factors affecting particular melts change mineralogy, chemistry,

or texture for each basalt. As the newly formed basaltic crust mainly by fractional crystallization of peridotitic or eclogite, is carried away from the active ridges toward the subduction zones, it is hydrated and enriched with other mineral sex changed in the presence of sea water. It is highly altered specially at the top 1-3 km. Landward distribution of lavas from island arcs and active margins shows rocks presenting differences in composition of potassium-to-sodium ratios. These differences have called the attention of geologists (Clark and Ringwood, 1966; and Green and Ringwood, 1970) who have proposed several explanations for basalts formation.

Also differences of pressure and temperature, partial melting, dehydration, fractional crystallization, rise of material to the surface, could account for the variety of rocks formed from other rocks and minerals coming from the mantle (figure 8).

Figure 8. The later phase of island-arc development. Dehydration of serpentine results in partial melting of quartz eclogite; the melt reacts with the overlying mantle to form a garnet pyroxenite. with diapiric rise, partial melting, and later fractionation, so calc-alkaline magma is produced.



Source: A. E. Ringwood, 1974, *Journal of the Geological Society*, and Ehlers-Blatt, 1980

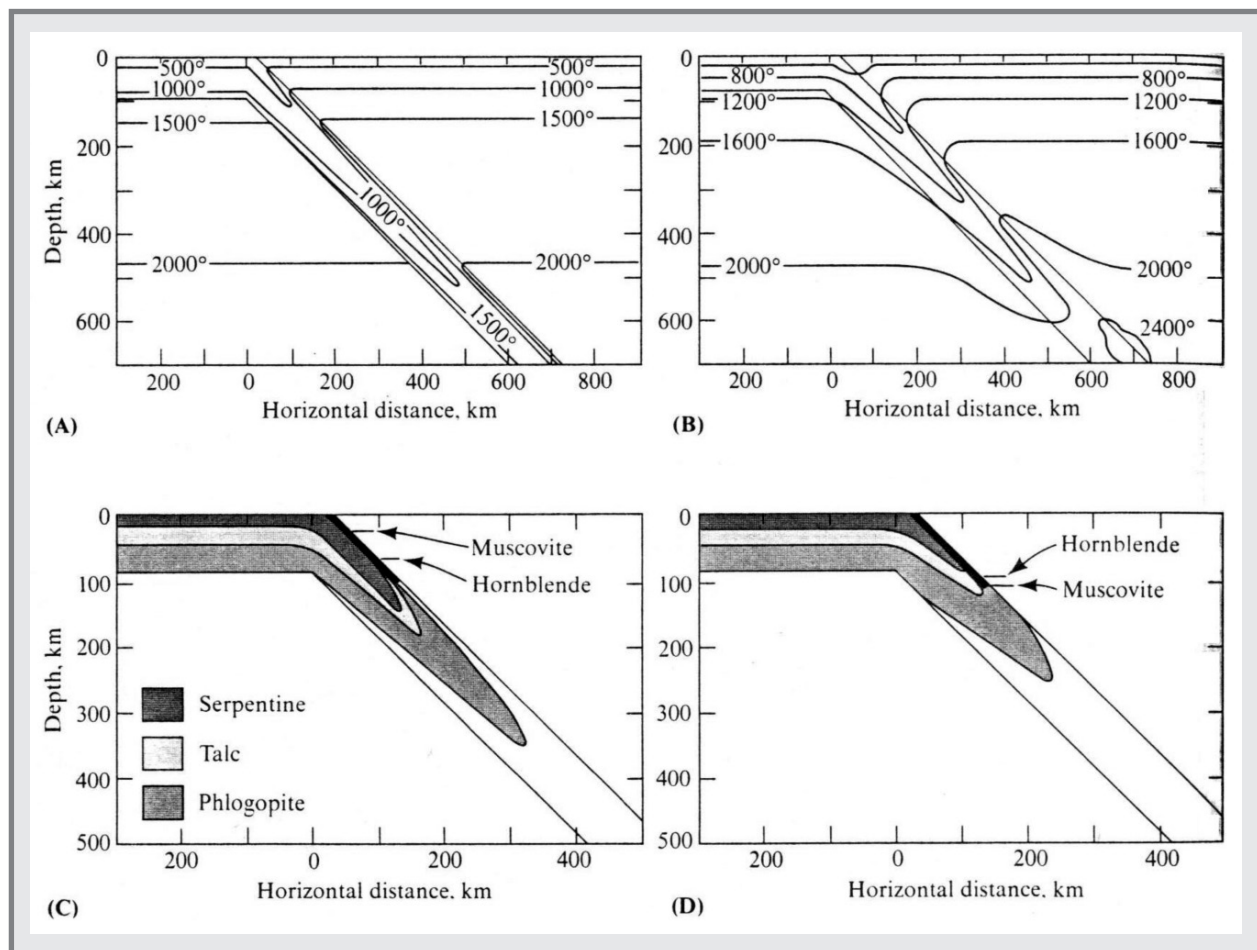
The presence of specific types of rocks in volcanic areas where subduction is taking place, have led to relate them with changes in conditions of downgoing slab. It is observed from seismicity that Wadati-Benioff zone becomes deeper landward. The potassium-to-sodium ratio in volcanic rocks resulted in depths corresponding to depths of the Benioff zone but also increases. Laboratory experiments of peridotites made by Green and Ringwood (1970) show the behavior of this rock as a function of pressure and temperature.

Three main groups of volcanic rocks have been recognized from subduction related volcanism as follows:

- 1) Tholeiitic rocks (basalts), low in potassium, and a high FeO to MgO ratio. Rocks of this type are found at points closest to the trench in subduction areas, either on young volcanic islands or continents.
- 2) Calc-alkaline rocks (basalts richer in aluminum, andesites, dacites, rhyolites). The potassium contents in these rocks are higher than in the previous type. They are lower in iron and higher in aluminum. They are found further inland.
- 3) Potassic rocks. They still show a higher ratio of potassium (up to 4%) according to its SiO₂ contents. They are detected on distant volcanic chains away from the trench, corresponding to a deeper point of the Benioff zone. Ringwood (1974) showed it is possible to obtain these rocks from a pyroxene fractionation. Minear and Toksoz (1970) had provided the thermal models for a subducting plate (figure 9).

The plate –as going down– carries within a temperature colder than its surroundings, and its rocks are subjected to more pressure. Under these conditions, stability regions are created far rocks being dehydrated. The bond-cracking power of the water helps the formation of the partial melting of the mantle above and into the slab itself.

Figure 9. A and B. Thermal models for the crust and mantle in a subduction region. C and D: The stability regions of serpentine, talc, and phlogopite according to models A and B, respectively.



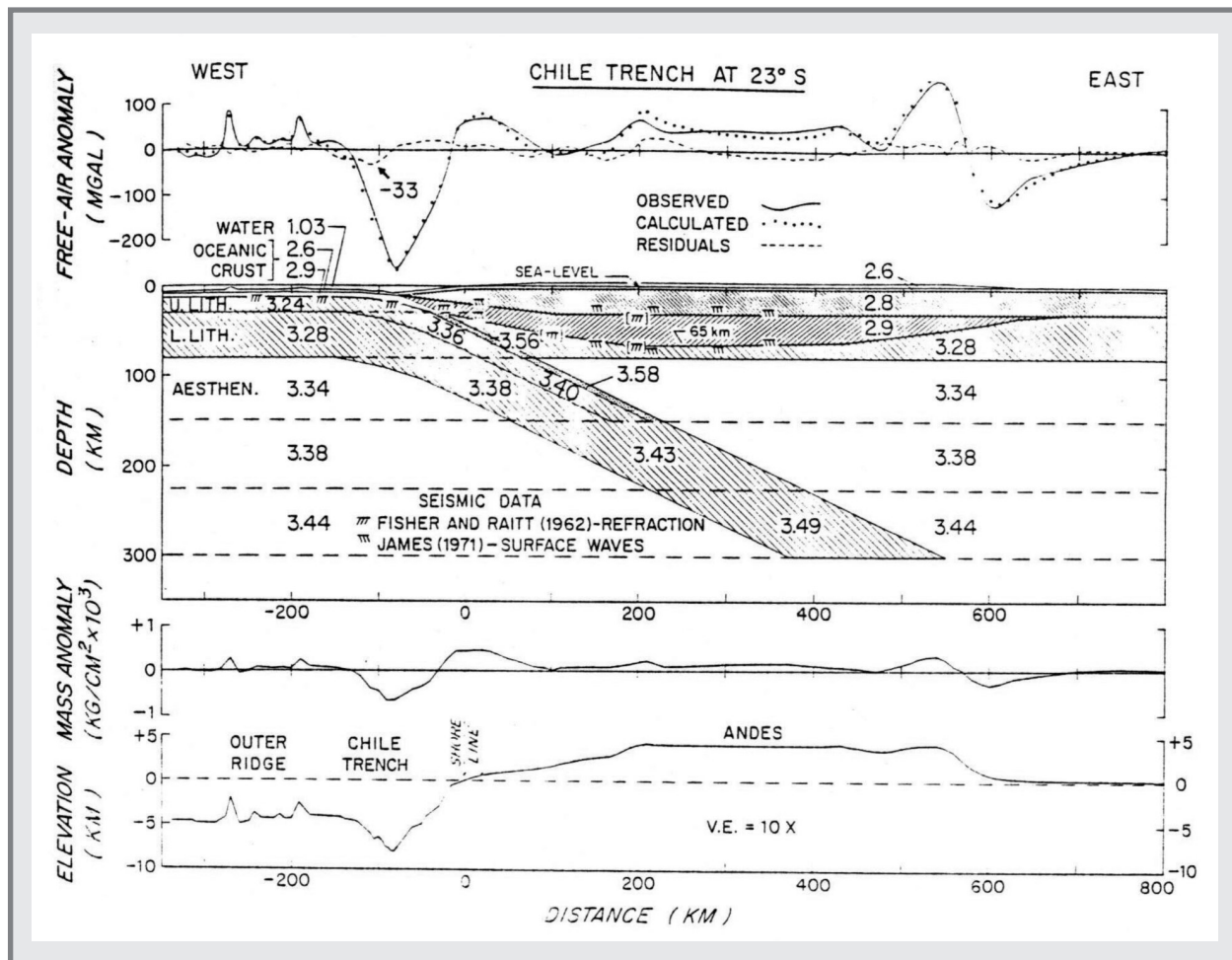
Source: A and B from J. W. Minear and M. N. Toksoz, 1970, *Journal of Geophysical. Res.*, 75, Figs. 6, 10. C and D from P. J. Wyllie, 1973, *Tectonophysics*, 17, Fig.7.

GRAVITY-DENSITY MODEL

Grow and Bowling (1975) provided the rock density model for a subduction zone, using data from petrology, temperature, gravity, and seismic observations. This model assumes anomalies instead of irregular changes in the thicknesses of the downgoing slab (figure 10).

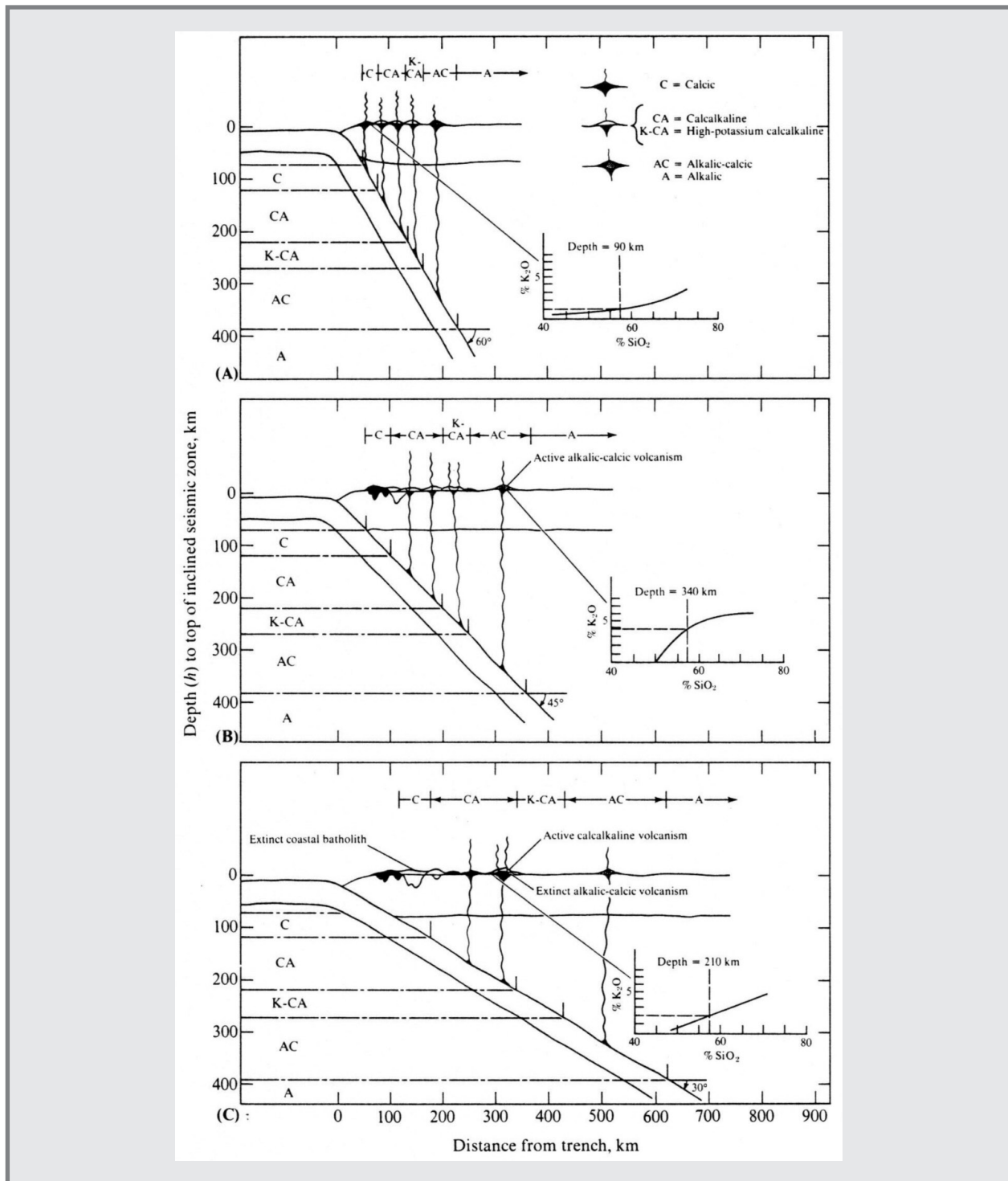
Barazangi and Isacks (1976) related volcanism, seismicity, and crustal structure due to Benioff zone dipping with different kinds of volcanic lavas distributed on the surface (figure 11). The extension on surface of rocks with different composition will reflect the Benioff zone underneath. The greater the horizontal range of these rocks, the smaller the angle of the subducting slab. The closer the rocks, the greater the angle dips of the Benioff zone.

Figure 10. Gravity model of Chile trench and Andes using density structure for mantle and descending slab based on thermal and petrologic data for a periodotite mantle. The oceanic crust is assumed to be basalt, which can transform to ecogilite (3.56 g/cm3) at 30 km depth.



Source: Marine gravity and topography from Hayes (1966). West of the trench the oceanic crustal layers are interpolated. Moho among refraction data from Fisher and Raitt (1962).

Figure 11. Diagrams showing the relationship between subduction dip angle and location of various igneous rock types. Inserts show K_2O/SiO_2 variation plots that show trend lines typical of calcic, alkalic-calcic, and calc-alkaline associations in A, B, and C, respectively.



Source: S. R. Keith, 1978, *Geology*, 6.

SEISMIC MODEL

Physical properties of rocks, such as electrical conductivity, magnetism, density, and seismic wave transmission velocity are the main concern of geophysicists. A seismic survey data in the ocean compared to laboratory real observations using the same properties in rock samples, gave rise to the idea that if similar velocities were obtained for both sets of data, the same type of rock could be presumably inferred to exist in the oceanic crust.

A seismogram shows the wave arrival time, at a particular station, coming from a source. A field work provides sufficient data from several stations. Distances and times are measured. A theoretical model is created which would reflect the conditions existing within the Earth. Thicknesses and dipping of the layers are determined according to calculations of the rays which have to travel with certain velocities, through different media, to simulate the same arrival times that real observations data have for the same distances. In this way a set of synthetic seismograms is obtained.

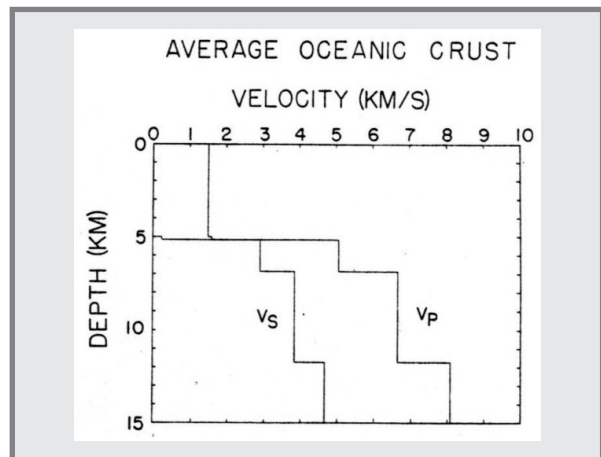
The resulting model is applied to the real situation of the Earth, so shapes, sizes, and materials are assumed to be from the same kind of those fitting better the values provided by the laboratory samples and field data.

This method has some limitations since it is based on physical features of the rocks, so the interpretation is not always unique. Depending on the size of the source, different frequency ranges can be obtained. The smaller the size of the explosion (or event), the finer the structure which can be determined. In the Antarctic, 1 gram-size charges were used to create wave lengths short enough to be able to scan ice layer thicknesses of 1 cm or more. This was secured by the absence of background seismic noise in other areas

so much bigger charges must be used to overcome the attenuation of the waves with distance and local seismic noise conditions. Lower frequencies thus generated a limit to capability of finer layering detection.

A valuable contribution from seismology to know the structure of the oceanic crust was given by Spudich and Orcutt (1980), by using synthetic seismograms compared to finer refraction seismic, stratigraphy, and petrology data. Previous assumptions modeled a very homogeneous, uniform, flat (sometimes dipping) oceanic layered crust which would satisfy marine refraction data. A restraint was however that only first arrivals could fit the model, An average model of this kind widely used as a reference in marine work, is shown in figure 12,

Figure 12, An average oceanic crust model. Top two layers are water and unconsolidated sediment. Crustal 5 velocities are derived from the assumption of a 0.25 value of crustal Poisson's ratio.



Source: *P* velocities and layer thicknesses are taken from *Raitt* [1963].

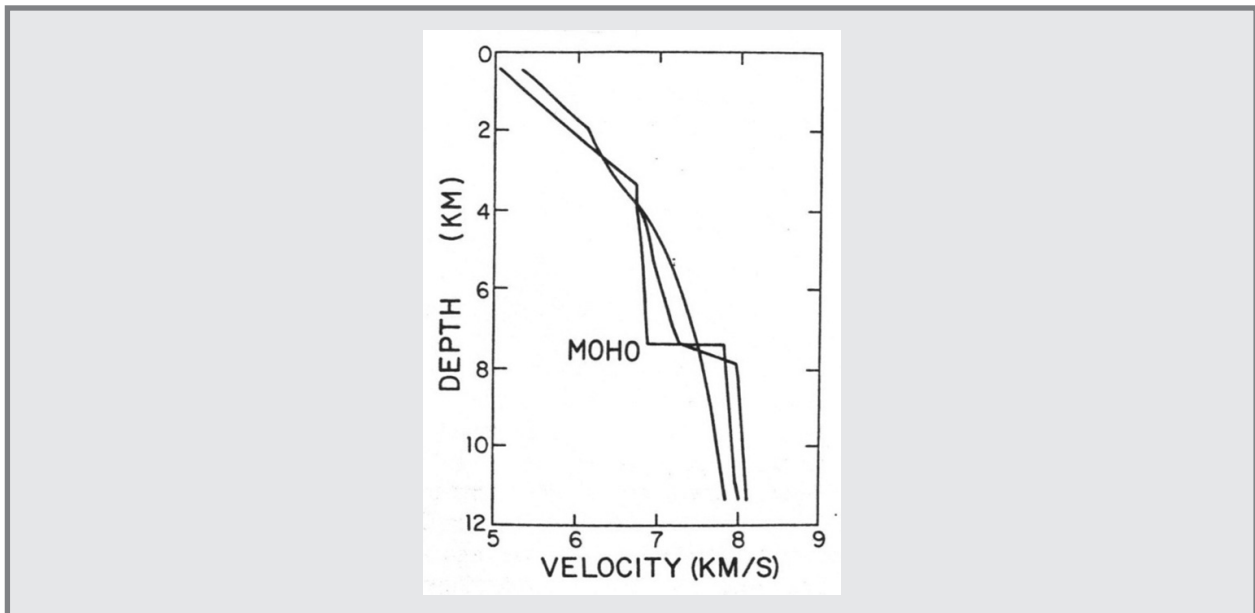
This model was composed by averaging together all the layered crustal models available in 1963. Very different velocity profiles could be developed to fit perfectly the seismic data. Figure 13 shows three

velocities profiles which would satisfy the arrival in figure 14.

Spudich and Orcutt's model improvement is achieved by being able to make use, not only first but also second, third, and other arrivals from reflections

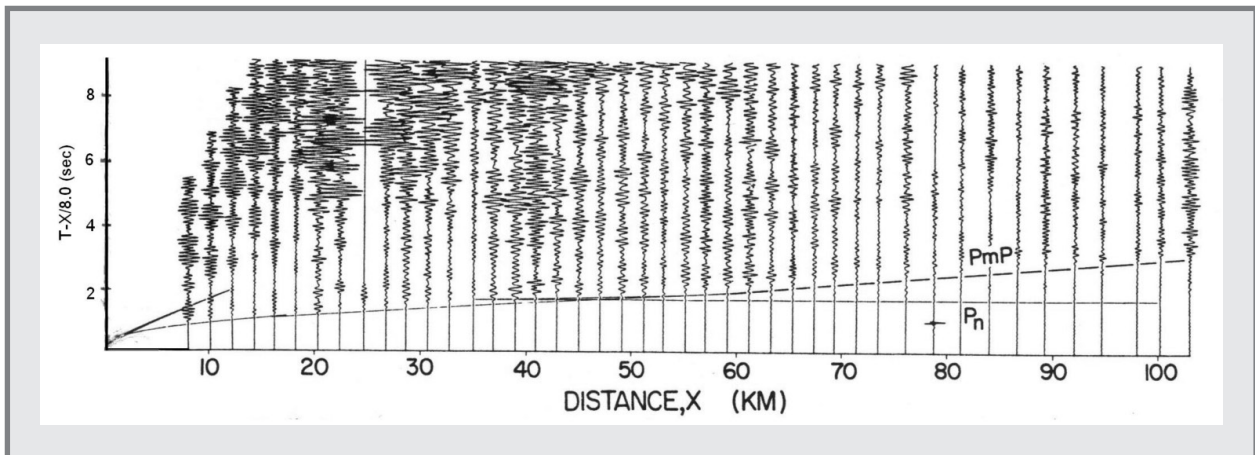
and refractions, further constraining the synthetic seismogram model. In addition, the refined analysis of a massive data leads to a rather unique model. An example is shown in figure 15 where four velocity profiles overlay each other.

Figure 13. Three velocity profiles which fit the observed travel times of *P* waves in figure 14. Note the different Moho transitions at 8 km depth which is measured from the top of the crust.



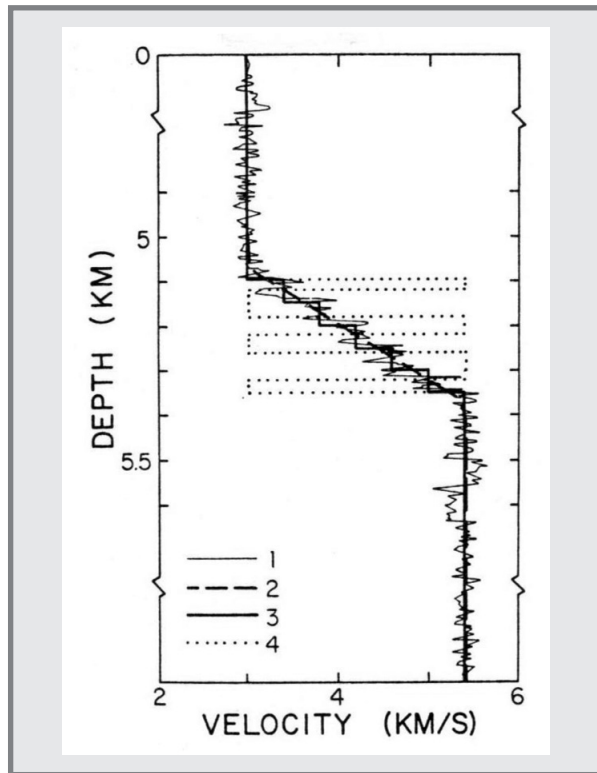
Source: Spudich and Orcutt (1980).

Figure 14. Data from which models in figure 13 were derived. *PmP* is the mantle reflection.



Source: Spudich and Orcutt (1980).

Figure 15. Four velocity profiles which are indistinguishable from each other as they are examined by using 1-km-long seismic waves.



Source: Spudich and Orcutt (1980).

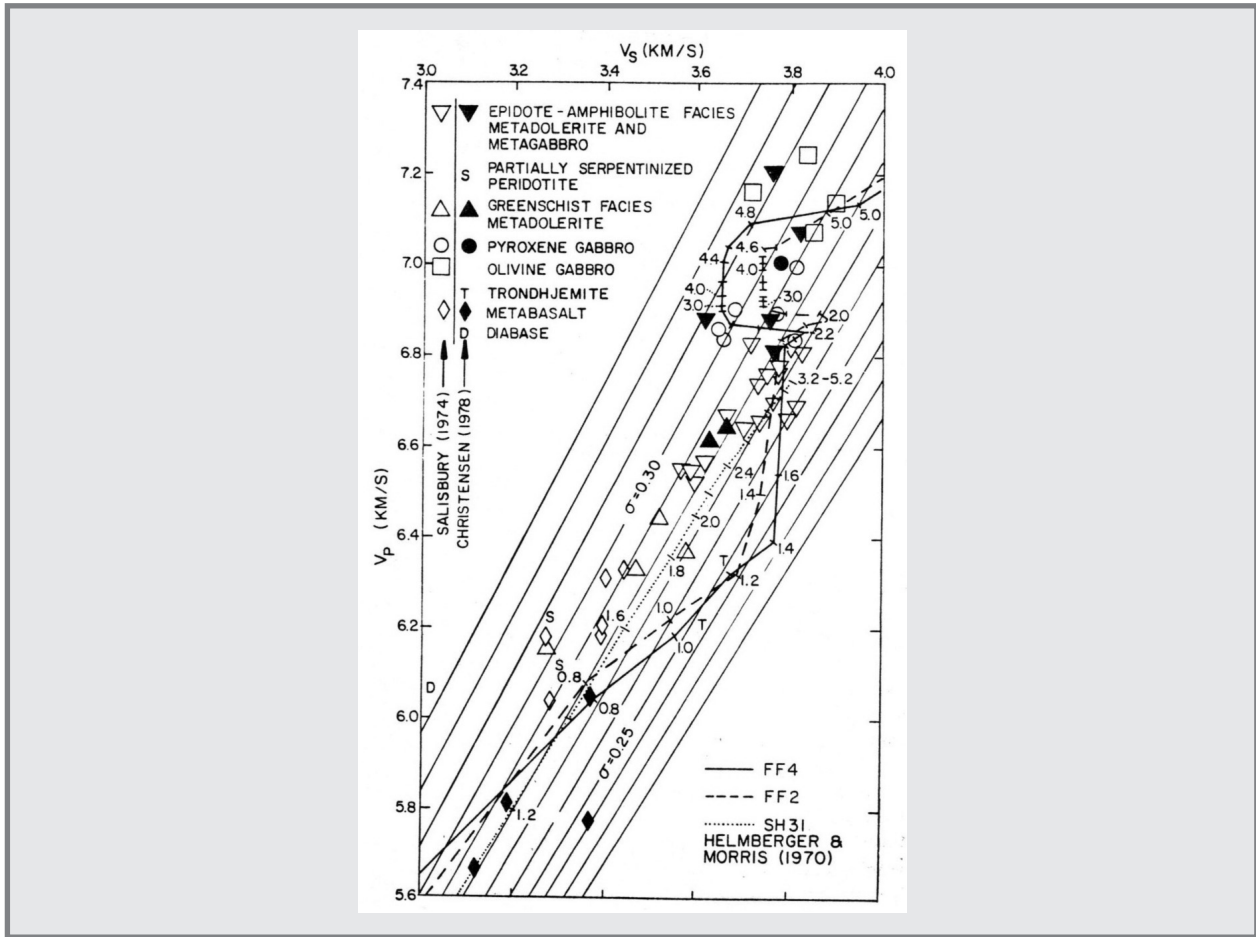
Spudich and Orcutt's final model is checked versus real seismic data and petrology of the ophiolite complex at the Bay of Islands, they agree very smoothly. From this analysis each one of the oceanic crustal layers is discussed as follows.

- Layer 1 –a thickness roughly up to 1.7 km, consists of unconsolidated sediments.
- Layer 2 –a finer layering. Usually three layers are distinguished –2A, 2B, and 2C.
 - Layers 2A and 2B are fissured, cracked, filled with water, sediments and pores. Velocities range from 2.56 km/s to 5.52 km/s, controlled mainly by porosity gradients. Layer 2C is more homogeneous. As pores are modeled flat, velocity decreases; as they are spherical, velocity is higher. Metamorphic grade and mineralogical transitions (prehnite/pumpellite to Greenschist, epidote to amphibolite) show changes per seismic velocities but do not coincide with pillow basalt and dike strata boundaries.
 - Layer 3 is rather uniform. Laboratory velocity measurements of samples, velocity measurements at sea and in ophiolites are similar. This layer has a gentle gradient in velocity 0.0 to 0.1 s, maximum 0* to 0.2 s far Vp and Vs < compressional and shear waves respectively. This gradient at depth depends both on pressure and mineralogy. Figure 16 shows this relationship.

Correlation of velocities of P (Primary) waves and S (Shear) waves in ophiolites, ocean rocks samples, and seismic wave arrivals, shown detailed. The top (0 - 2.4 km) is made of metadolerite. Velocities agree with those of samples of hornblende and metagabbro. The middle part (2.4-4.4 km) agrees with pyroxene gabbro. The bottom section (4.4 km to mantle) agrees with olivine gabbros.

Lewis and Snysman (1979) and Stephen et al. (1979), report a Low Velocity Zone for P waves at the base of layer 3 for Cocos plate region. Spudich (1980) finds a Low Velocity Zone for shear waves near Guadalupe Island, Mexico. Lateral heterogeneities of layer 3 from explosion seismology are detected. The thickness of the crust-mantle transition is 0. to 3.0 km.

Figure 16. The V_p - V_s paths for FF2, FF4 and SH31 compared to ophiolite sample velocities. Numbers written along the paths are depths beneath the sediments base (by km).



Source: Spudich and Orcutt (1980).

THE VOLCANIC CHAIN IN NORTHWESTERN SOUTH AMERICA

By applying these concepts to the Northwestern Pacific region in South America, we see that at a distance of 270 km and parallel to the trench the range of volcanoes for the Central Cordillera is lining up. Also the Benioff zone dipping for this region lies between 30 degrees (Ramirez and Acosta, 1982) and 35 degrees Pennington (1979). But the depth of the subducted plate beneath the volcanoes would be between 160 and 190 km (Van Hissenhoven, 1988) which is between the required ranges for generation of andesites (figure 11).

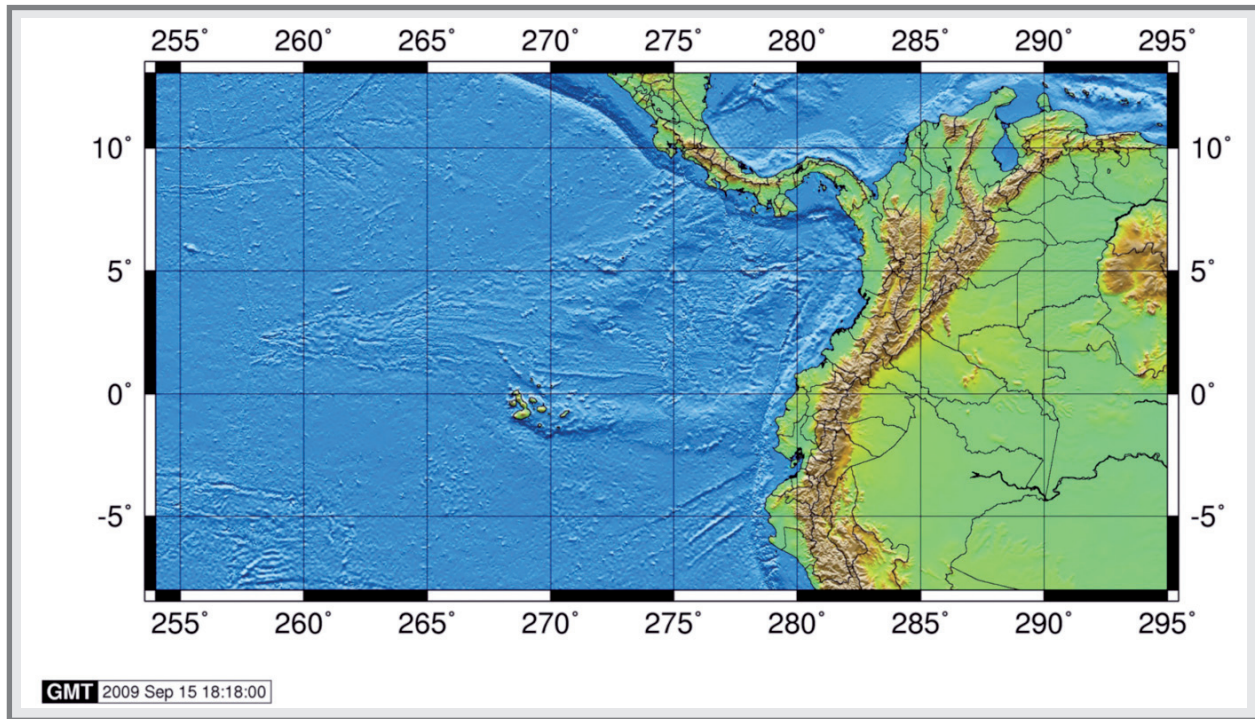
Volcanic activity along the Andes is usually divided in 3 sections –Northern 5° N-2° S, Central 16° S-2° S, and Southern 31° S-52° S–. Central and Southern sections are very well documented and studied due probably to their well-defined tectonic and geological settings. The Northern part is complex but data is very scarce.

Only the Northern region will be considered here. Two distinct Cordilleras may be observed in Ecuador, but enter the Southern part of Colombia as a single range and form the Massif of the Andes around Pasto (figure 17). From this point the range is divided in three Cordilleras which are increasingly separated

by valleys as they go northern. Each Cordillera is different by origin, geology, and age. Active volcanism in Colombia coincides with the Central

Cordillera where great activity takes place during the Tertiary period (Ramirez, 1975. See figure 18).

Figure 17. Colombia showing three Cordilleras and the analyzed area.



Source: Author's.

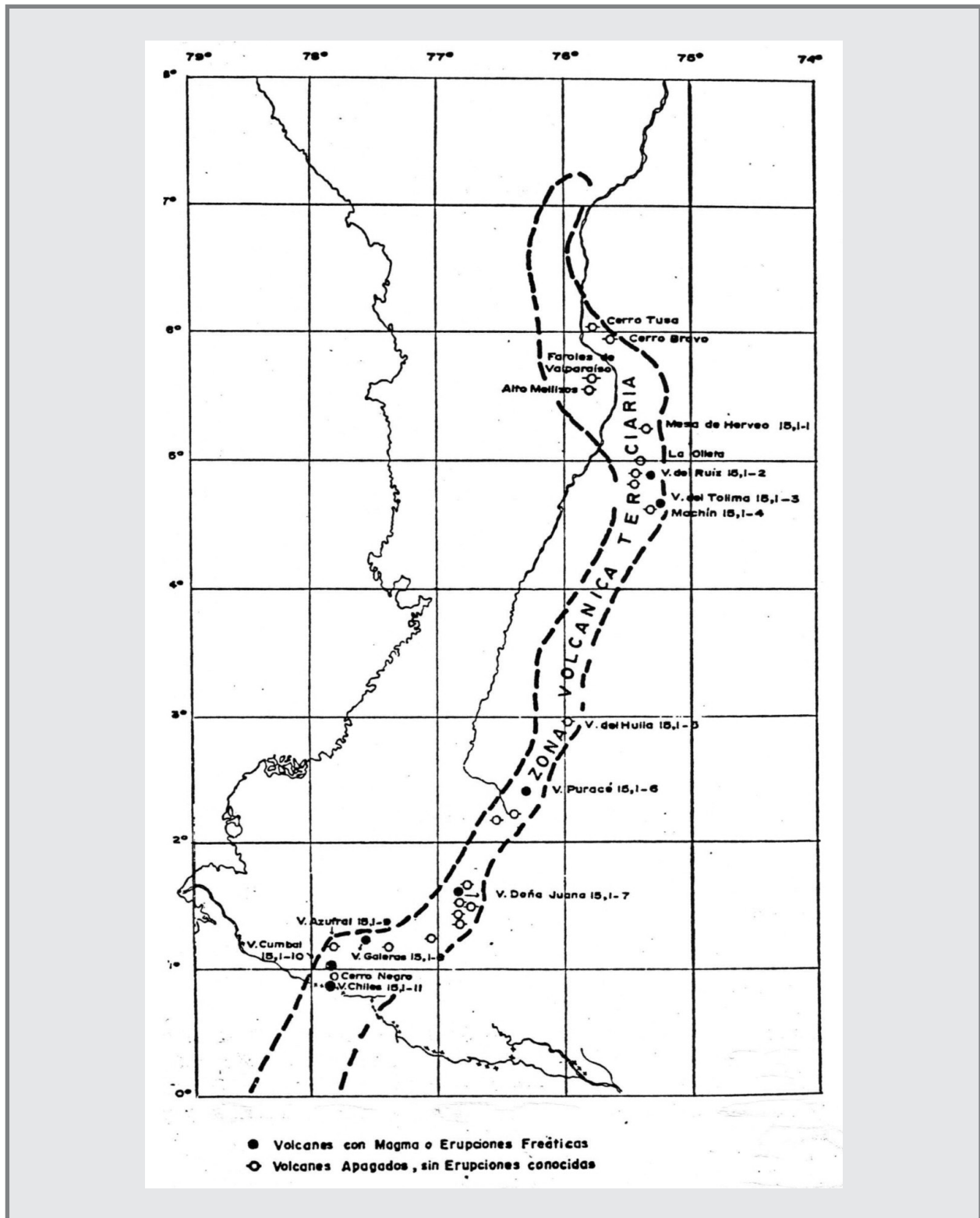
A basic igneous complex is defined along the Pacific coast of Colombia and Ecuador for a group of mafic volcanic and intrusive rocks. These rocks are also found on the Western slopes of the Western Cordillera, showing the first magmatism of the Andes. They are submarine flows on top of coarse-grained gabbro-harzburgite (Ecuador). Intrusions of diorite-tonalite rocks show differentiation of the same basaltic magma. Goossens (1973) finds a younger age for lavas in Ecuador, concluding that volcanism in the coastal plains was active until early Eocene age. Their composition in general is $Al_2O_3 < 17\%$, $TiO_2 < 1.7$, $K_2O < 4\%$.

In this northern region strato volcanoes have –as their foundation– mesozoic and cenozoic rocks. The Western Cordillera is separated from the Central by

the main fault of Romeral which runs north-southern. So the eastern side of the Central Cordillera has a continental feature but towards the western, the Western Cordillera has an oceanic composition.

Case et al. (1971) see the transition between both cordilleras at the eastern edge of the high gravity positive anomalies measured along the Western Cordillera (figure 19). Ophiolites are clearly observed along the Western Cordillera, both in Colombia and Ecuador. The discussion whether a segment of pure cretaceous ocean crust due to a westward jump of the original trench of the subduction zone in the lower tertiary was uplifted after a new trench and subduction began at the present location (Van Andel et al. 1971)

Figure 18. A volcanic zone.



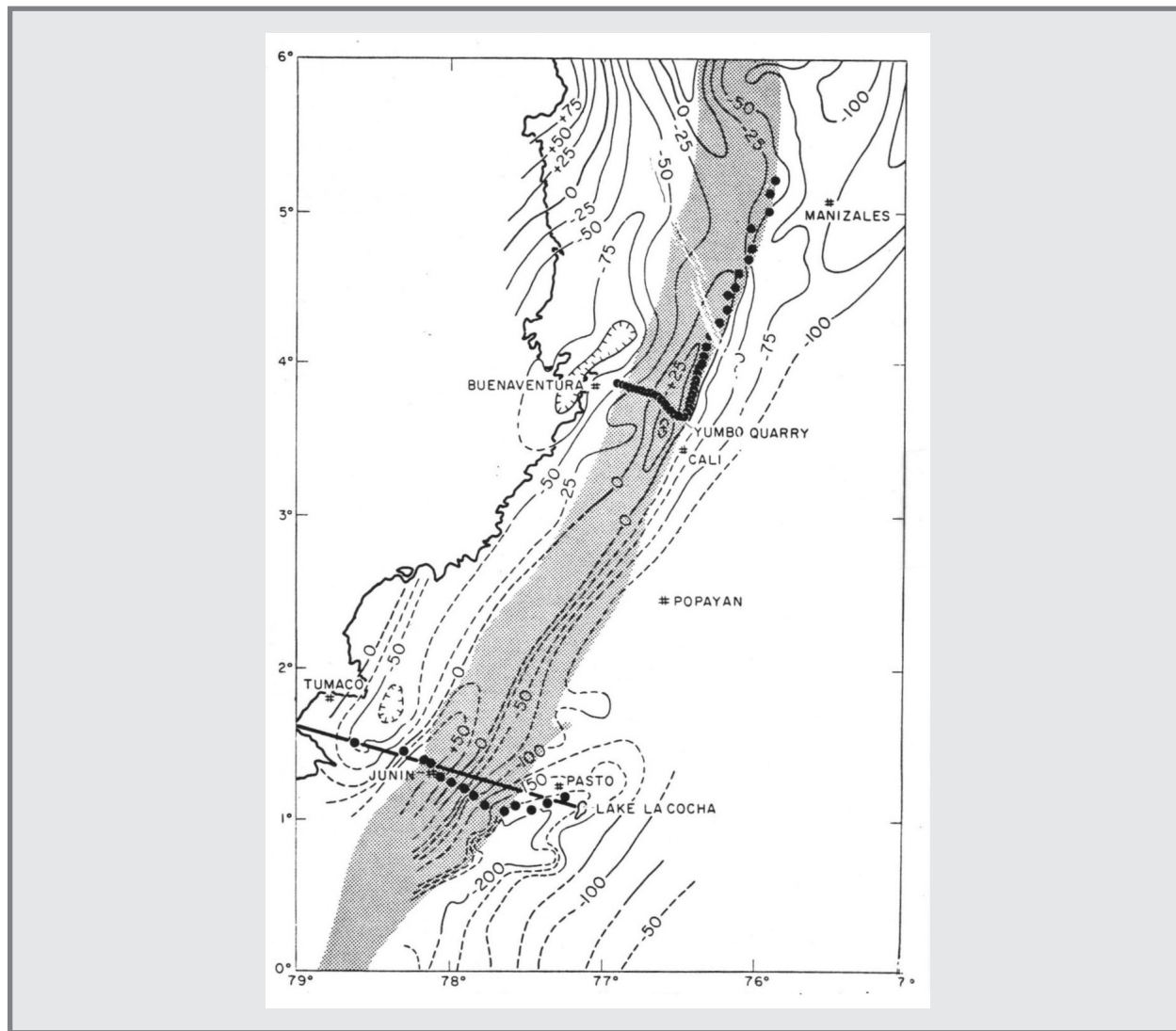
Source: Ramírez, 1975.

Mooney et al. (1979) think it is a fragment of ocean crust overlaying continental material. Flueh et al. (1982) accept Toussaint and Restrepo's (1976) explanation regarding Western Cordillera is an obducted piece of ocean crust over continental crust, over the paleo-margin, or over remnants of a derelict mesozoic subduction area which stepped and jumped westwards.

Today's picture shows from the trench landwards (figure 20), an oceanic crust with a 7.5 - 10.0 km

thick layer of sediments overlaying it, and extending to the lowland plains 80 km; then the ophiolitic Western Cordillera crops up. A sudden transition – from oceanic to continental material– is reached 100 km further East, where the Central Cordillera has its range of cretaceous and quaternary volcanoes sitting on plutonic rocks formed from previous eruptions of Tertiary age. Hall and Wood (1985) say the volcanic range belongs to the Quaternary period.

Figure 19. A Shaded Zone: Western Cordillera and Gravity values.

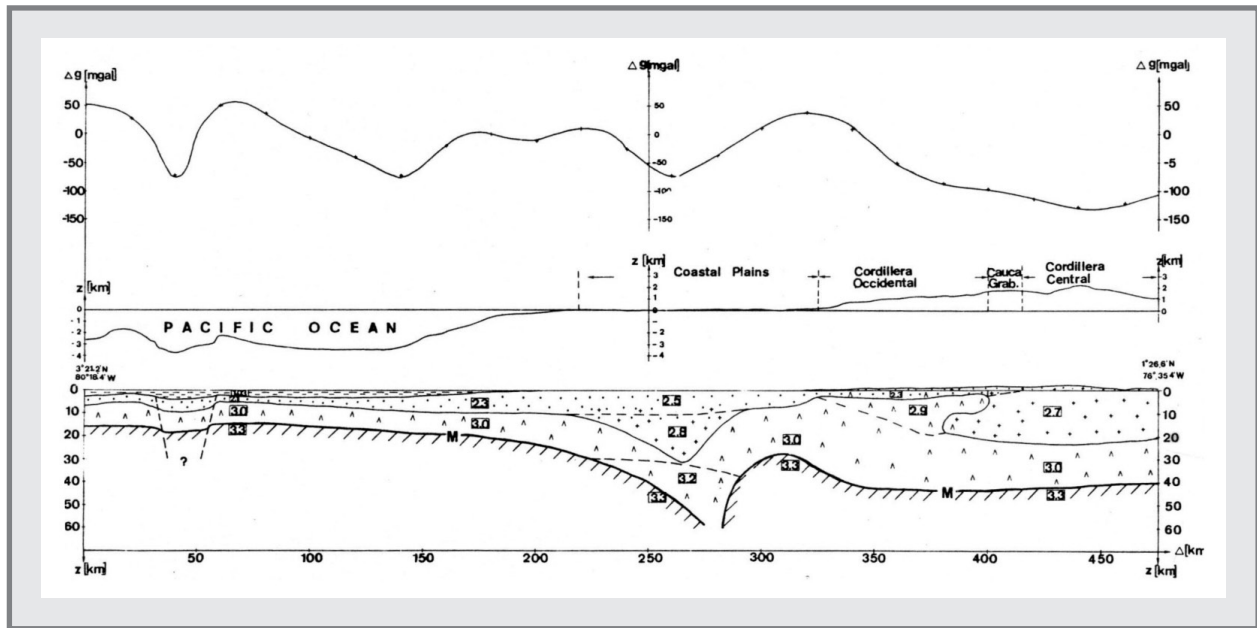


Source: Ramírez and Aldrich (1977).

The Colombian volcanoes have been divided by Ramirez (1975) in five groups. Most of them lay on the Central Cordillera. As the Eastern Cordillera gets closer to southern side and steps to western side,

a row of volcanoes is formed on it. The location and petrology of the most important Colombian volcanoes is given in table 1.

Figure 20. A crustal cross section. Ocean to Continent at Lat. 1° N. Gravity values on top of the figure.



Source: Ramirez and Aldrich (1977).

Table 1. Location and petrology of the most important Colombian volcanoes.

Name	Height	Lat	Lon	Petrology
Ruiz	5480 m	4 33'N	75 22'W 2-	pyroxene or olivine clinopyroxene andesites (58-62% SiO)
Tolima	5525 m	4 39'N	75 22'W	andesites, dacites
Huila	5750 m	2 35 ' N	75 59 ' W	andesite, dacite, augite.
Purace	4700m	2 22 ' N	76 23 ' W	dark andes, lavas, tuffs.
Doña Juana	4250 m	1 31'N	76 56' W .	andesitic rocks with hornblende & biotite
Galeras	4262 m	1 10'N	77 18'W	andesites
Azufra	4070 m	1 05'N	77 41'W	andesites
Mayasquer	4470 m	0 48'N	77 51'W	andesites

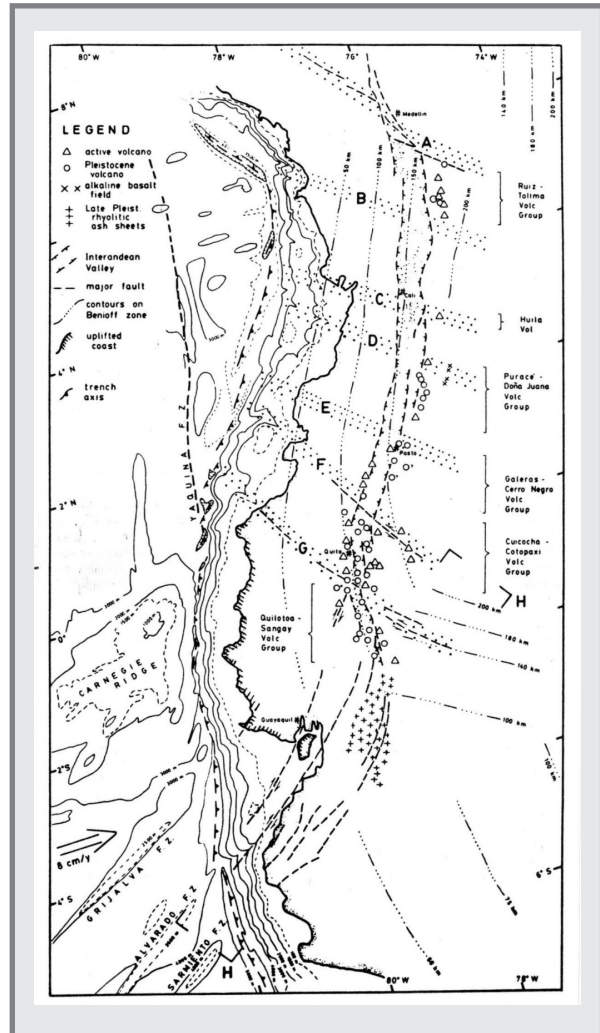
Source: Ramírez, 1975.

Above the largest active volcanoes but in total there are 30 including those which are semi-active or dormant. Figure 21 shows the map including locations of these volcanoes –Ecuador volcanoes can also be seen in the southern section. Pichler et al. (1976) and Thorpe et al. (1982) have basaltic andesites as the characteristic rocks for this Northern section. Magmas are formed here by partial melting and fractional crystallization within the mantle, with little contribution from the crust (Thorpe et al., 1982). Kroonenberg et al. (1982) describe a newly discovered young cinder cone field of alkaline basalts 30 km east from Purace.

CONCLUSION

The Eastern Equatorial Pacific is subject to spreading mechanisms of the ocean floor leading to subduction. A hotspot is observed to be the cause of the Galapagos Islands as well as other aseismic features on the segmented plates being subducted. A portion of the ocean crust has been obducted along the Colombia-and-Ecuador coast resulting in the Western Cordillera. Ophiolite studies and rock sample measurements show agreement with marine refraction data. A synthetic seismic model contributes to know the oceanic crustal structure. Volcanic rocks in the northern part of the Andes are correlated with the dipping of the Benioff zone showing agreement between the andesitic volcanism observed and subduction actually taking place in this complex area of northwestern South America.

Figure 21. Schematic diagram of proposed segment boundaries (A to H) for Northern Colombia-and-Ecuador Andes.



Source: Data from Lonsdale (1978), Lonsdale and Klitgord (1978), Barazangi and Isacks (1976, 1979), and Pennington (1981) [from Hall and Wood, 1985].

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Figures 4 and 17 have been drawn using GMT (Wessel and Smith, 1991, 1998), the topography

and bathymetry come from University of Alaska (Lindquist et al., 2004.)

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