THE ORIGIN AND IMPLICATION OF THE STEEP GRAVITY GRADIENT IN THE VICINITY OF THE MARTIC ZONE, SOUTHEASTERN PENNSYLVANIA

A Thesis Submitted to the

Temple University Graduate Board in Partial Fulfillment

of the Requirements for the Degree

Master of Arts

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by

Taeyoung S. Song

December 1987

Dr. Mary Louise Hill
Thesis Advisor

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A Thesis Submitted in Partial Fulfillment
of the Requirements for the Degree of
Master of Arts in Geology,
College of Arts and Sciences,
Temple University
Philadelphia, Pennsylvania

December 1987

Dr. Mary Louise Hill	Date
Dr. George H. Myer	Date
Dr. Peter W. Goodwin	Date

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Abstract

A long, narrow, straight belt of closely spaced Bouguer gravity contours and lithologic boundaries coincides with the Martic shear zone, southeastern Pennsylvania. The most likely cause of the steep gravity gradient is sought by checking the influence of the density of surface lithologies, mass distribution associated with the topography, and the isostacy model. 15-20 *km* difference in crustal thickness across the Martic Zone is calculated using Sharma's data and maximum depth equation. Based on a shear zone geometry, gravity anomaly parallelism, and crustal thickness contrast, the Martic Zone is proposed to be the western boundary of the Piedmont terrane. The Martic Zone is compared with other areas such as the San Andreas Fault and the Alpine Fault to evaluate possible plate boundary features. This tectonic interpretation of the Martic Zone may contribute to a new view on the central Appalachian orogenic belt.

Dedication

To God, my friend

Dedication

To God, my friend

Acknowledgements

I would like to express my deepest appreciation to my advisor, professor Mary Louise Hill, for her suggestions, inspirations, highly constructive criticism on my work and of course her thorough review of my thesis. This thesis would not have been completed without her considerate help.

Old and new members of the structural group including D. Valentino and D. Goldblum deserve my thanks for their friendly discussion on this thesis. I would like to extent my thanks to the staff and students of the Temple University Geology Department. Especially, I owe Drs. G. H. Myers and P. W. Goodwin a great deal of thanks for their support and encouragement.

Finally, I would like to express my sincere thanks to my husband, Iickho, for his patience, love, and help in getting this manuscript by a laser printer.

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Chapter 1. Introduction

1.1 Objectives

The main objective of this study is to determine the most likely cause of a steep gradient in the Bouguer gravity fields observed in the vicinity of the Martic Line, southeastern Pennsylvania. This gradient occurs as a long, narrow, and very straight belt of closely spaced gravity contours separating negative gravity anomalies to the north and positive to the south. This belt coincides with the Martic Zone (Figure 1).

The second objective is to consider the implication of the steep gradient in gravity, given its likely cause.

The results of this study yield new information about the structure of the Martic Line. This information is significant to the tectonic history of the Appalachian orogenic belt.

1.2 Previous Work

Previous studies of the area focused on the age of the Wissahickon Schist (Mackin, 1939), structure of the Martic Line (Wise, 1970), and metamorphic petrology of the Pennsylvania Piedmont (Crawford and Crawford, 1980). The Martic Line was interpreted as a normal lithologic contact between carbonates (Mackin, 1939) and phyllite without regard for differences in age, metamorphic history, and structure. The dominant explanation is a low angle thrust (Keppie, 1970) directed northwestward. In a recent geophysical study, Eisner (1986) agreed with the interpretation of thrusting, and suggested that there are several imbricate overthrusts associated with the Martic Line. Recent work by geologists at Temple University (Myer *et al.*, 1985; Hill, 1987) introduced the idea of a shear zone

Figure 1a. Bouguer gravity fields of the study area. Solid lines represent the gravity contours. The gravity contour interval is 5 *mgals* (gravity map compiled by Simpson and Godson, 1981).

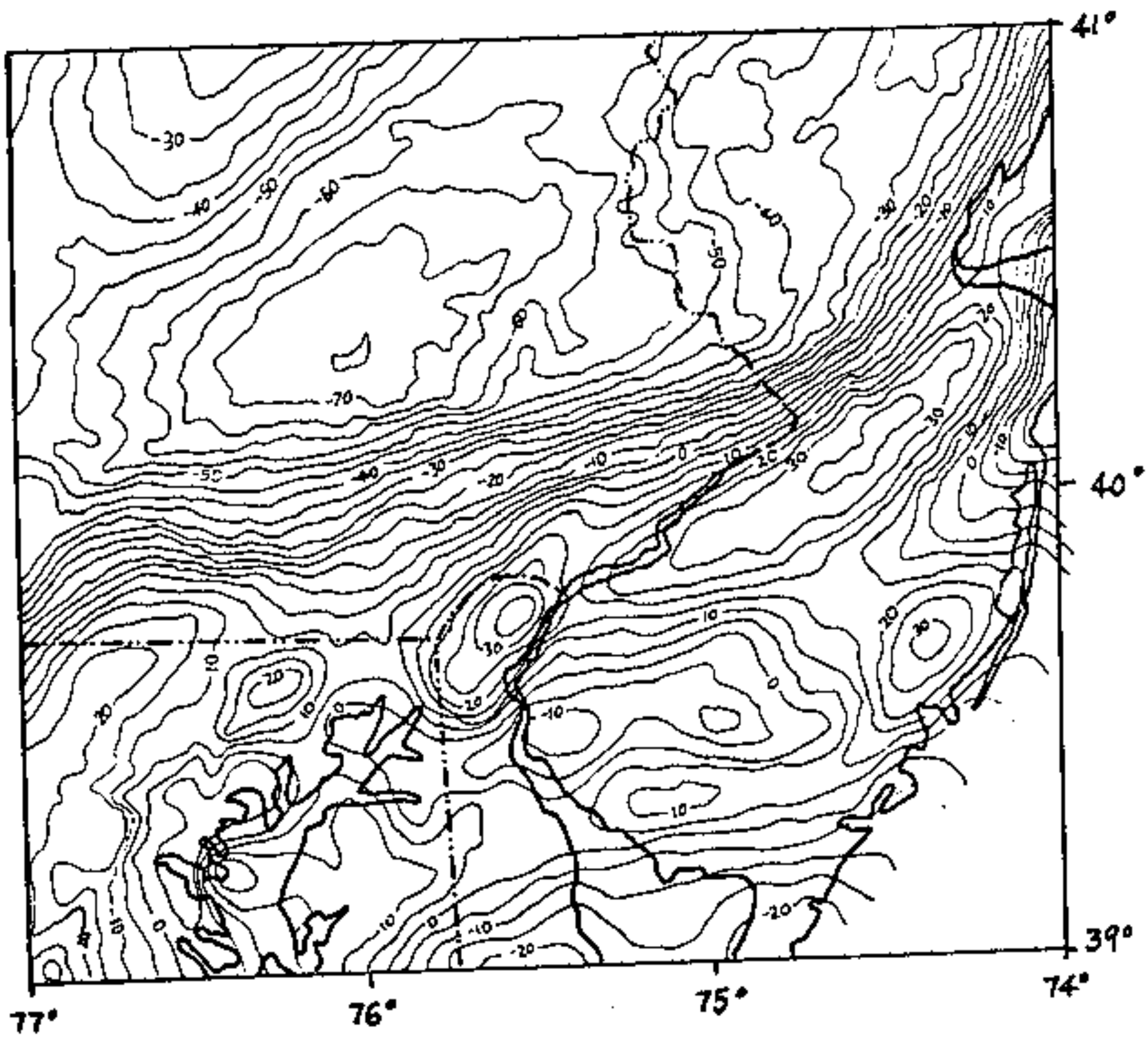
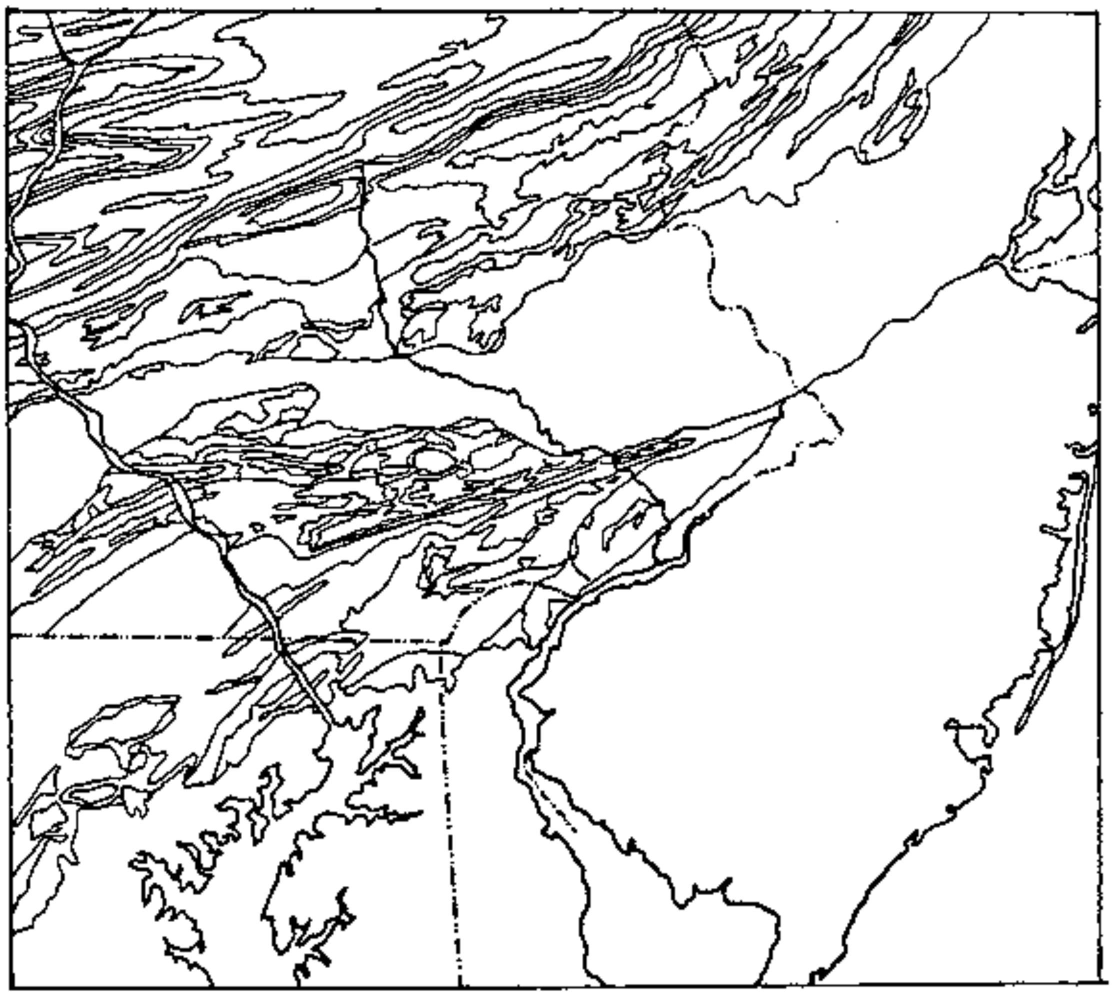


Figure 1b. Geology of the study area. Solid lines represent lithologic contacts. Dash-dot lines are state boundaries. In order to compare the gravity pattern, legend is not shown (Geologic Map of Pennsylvania).



coincident with the Martic Line.

Various interpretations of the Appalachian Gravity Parallelism (AGP) have yielded different views on Appalachian tectonics. Although the basic idea of ancient plate collision has been accepted, Thomas (1983) suggested that the positive-negative pair in AGP originated from a steep-angle suture zone. Cook (1984) analyzed the COCORP data and challenged with a diverging interpretation based on a thin skinned-tectonic model.

In mapping the east coast of North America as a mosaic of accreted allochthonous terranes (Williams, 1978), Williams and Hatcher (1983) interpreted the Martic Line as the western boundary of the Piedmont terrane.

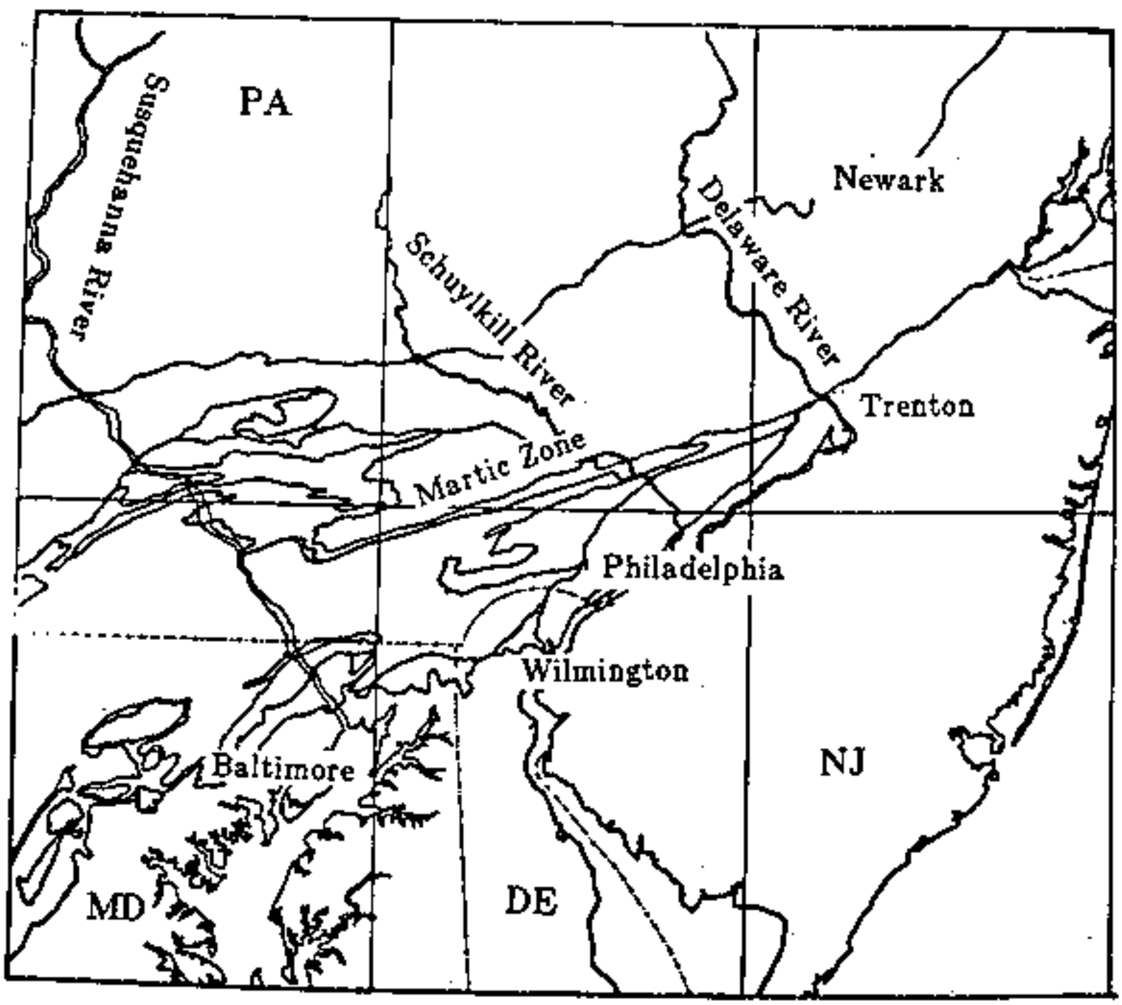
1.3 Study Area and Methods

This thesis considers the area of $39°-41°$ N latitude, $74°-77°$ W longitude and includes parts of four states: Pennsylvania, New Jersey, Maryland, and Delaware (Figure 2).

The Martic Line seems to be unique with respect to geography, geology, and gravity anomalies. No other place in the Appalachian orogenic belt shows such a strong coincidence of surface geology, gravity patterns, and structural features. The geology, structure, and gravity patterns of the southern Appalachians are rarely coincide on the surface.

Gravity profiles were drawn based on "Colored gravity anomaly and terrain maps of the east central United States" (Simpson and Godson, 1981). Individual N-S gravity profiles were made at $7' \, 30''$ longitude intervals to compare with lithologic and topographic profiles. The density of each rock unit was measured by Jolly-balance. Then the rocks which show no larger

Figure 2. Local geography of the study area.



than 0.2 g/cm^3 variation in density were grouped together regardless of age or lithology. Topography of the study area was traced with 500 *ft* contour intervals. The effects of the contrast of rock density across the Martic Line, variations in topographic relief, and differences in crustal thickness are analyzed as possible causes of the steep gravity gradient.

Chapter 2. Geologic Setting

2.1 The Martic Zone

The Martic line has been recognized by local geologists in the Pennsylvania Piedmont area since the early 1900s. Figure 2 shows the local geography of the Pennsylvania Piedmont area. On both sides of the Triassic lowland section the topography shows valleys and uplands. South of and adjacent to the Triassic lowland is the Conestoga valley section. South of this is the Piedmont upland section. In the northern part of the area is the Great Valley and Appalachian mountain section. Compared to the northern area, the southern Piedmont upland is relatively low in elevation. The Schuylkill and Susquehanna Rivers flow in a southeasterly direction and between these two rivers are the parallel Blue Mountain and Second Mountain folds. Chester Valley is a long straight valley underlain by Conestoga Limestone. The Martic Line was defined as the contact between Wissahickon Schist to the south and Conestoga Limestone to the north. The map trace of the Martic Line is along the southern edge of Chester Valley. The area of south of Chester Valley is underlain by lithologies of the Glenarm lithogroup. The Martic Line is an easily recognized sharp contact between two very different lithologies on the map.

The exact nature of the contact at the Martic Line has been a long-standing controversy. One possibility is that the Martic Line represents a normal stratigraphic contact (Mackin, 1939). However, the Martic Line is widely accepted as being a low angle overthrust fault (Wise, 1970; Keppie, 1970). The Martic controversy is closely related to theories about the age of the Glenarm Series. Those who believe the Glenarm Series to be Precambrian in age favor the overthrust theory as an explanation for the structural superposition of the Wissahickon schist above the younger Conestoga limestone. The controversy is unresolved.

Myer *et al.* (1985) introduced the concept of "Martic Zone", hypothesizing that the Martic Line lies within a major ductile shear zone. This shear zone is more extensive than the Martic Line itself, extending at least from Hanover, Pennsylvania in the southwest to Trenton, New Jersey in the Northeast. It is the Martic Zone that coincides with the steep gradient in gravity anomalies and is likely to be a significant tectonic feature, and not the Martic Line which is simply a transposed contact lying within the shear zone. For the purpose of this study, correlation of geophysical data will be to the Martic Zone rather than to the Martic Line.

2.1.1 Geology of the Martic Zone

The Martic Zone runs from Trenton, New Jersey to Hanover, Pennsylvania (Figure 1b). In the northeast the Martic Zone disappears under the Triassic sediments.

The Central Appalachian belt can be divided into 4 major lithologic groups: Coastal plain, Glenarm series, Triassic sediments, and Paleozoic sediments (Figure 1b). The major strike direction of the central Appalachian belt is generally NE. Almost parallel contacts between the 5 lithogroups is a distinctive feature of the area.

The Coastal plain is located along the Atlantic Ocean coast and is underlain by unlithified sediments. Quaternary (0-1.6 m.y.) age sand and gravel covers the major parts of Delaware and southern New Jersey.

The Glenarm series is a comprehensive lithogroup outcropping south of the Martic Zone. It includes the Setters Quartzite, Peters Creek Schist, Peach Bottom Slate, Cockeysville Marble, and Wissahickon Schist. Maryland and most southeastern Pennsylvania are underlain by Precambrian to Ordovician age metamorphic rocks. These are highly metamorphosed gneiss, schist, quartzite and marble.

The Triassic sedimentary sequence is 30-40 miles wide in northern New Jersey but narrows toward Pennsylvania. This sequence is mostly shale and sandstone intruded by diabase. It cuts the Glenarm series at the western part of the Susquehanna River.

North of the Traissic sediment is underlain by Ordovician to Devonian age sedimentary rock. A wide belt of shale, dolomite, limestone, and sandstone runs the entire length of the folded northern part. Beekmantown group is mostly limestone but includes coarsely crystalline dolomite. Martinsburg Formation of shale and slate and Hamburg sequence are the dominant type of Paleozoic sediments. Table 1 summarizes the lithologies of the study area. This table is adapted from the Pennsylvania geologic map and Eisner (1986) who presents a review of the stratigraphy of the Martic Zone with great detail. Further detail in descriptions of each lithology requires petrologic, mineralogic, and/or geochemical analyses which are beyond the scope of this study.

2.1.2 Previous work on the Martic Line

The Martic Line Problem has been a traditional controversy in Central Appalachian geology. Support for the Martic Overthrust hypothesis has been dominant. Another suggestion is that the line represents a normal lithocontact. Recently, it has been suggested that the line lies within a shear zone.

Table 1. Lithologies of the study area (not in stratigraphic order).

Location	Age	Rocks		
Southeast of the Martic Zone	Precambrian Lower Paleozoic	Glenarm Supergroup	Cockeysville Marble	
Southeast of the Martic Zone	Precambrian Lower Paleozoic	Glenarm Supergroup	Setters Quartzite	
Southeast of the Martic Zone	Precambrian Lower Paleozoic	Glenarm Supergroup	Peters Creek Schist	
Southeast of the Martic Zone	Precambrian Lower Paleozoic	Glenarm Supergroup	Peach Bottom Slate	
Southeast of the Martic Zone	Precambrian Lower Paleozoic	Glenarm Supergroup	Wissahickon Group	Octoraro Phyllite
Southeast of the Martic Zone	Precambrian Lower Paleozoic	Glenarm Supergroup	Wissahickon Group	Marburg Schist
Southeast of the Martic Zone	Precambrian Lower Paleozoic	Glenarm Supergroup	Wissahickon Group	Wakefield Marble
Southeast of the Martic Zone	Precambrian Lower Paleozoic	Glenarm Supergroup	Wissahickon Group	Obligoclase Schist
Southeast of the Martic Zone	Precambrian Lower Paleozoic	Wilmington Complex	Baltimore Gneiss	
Southeast of the Martic Zone	Precambrian Lower Paleozoic	Wilmington Complex	Anorthosite	
Southeast of the Martic Zone	Precambrian Lower Paleozoic	Wilmington Complex		
Northwest of the Martic Zone	Ordovician Devonian	Beekmantown	Rickenbach Dolomite	
Northwest of the Martic Zone	Ordovician Devonian	Beekmantown	Stonehenge Limestone	
Northwest of the Martic Zone	Ordovician Devonian	Beekmantown	Epler Limestone	
Northwest of the Martic Zone	Ordovician Devonian	Martinsburg Shale		
Northwest of the Martic Zone	Ordovician Devonian	Conestoga Limestone		
Northwest of the Martic Zone	Ordovician Devonian	Chickies Quartzite		
Northwest of the Martic Zone	Ordovician Devonian	Harpers and Antietam		
Northwest of the Martic Zone	Ordovician Devonian	Vintage Dolomite		
Northwest of the Martic Zone	Ordovician Devonian	Elbrook Dolomite		
Northwest of the Martic Zone	Ordovician Devonian	Hamburg Sandstone		
Northwest of the Martic Zone	Ordovician Devonian	Triassic Basin		

Comparing three types of pellitic rock (Harpers Phyllite, Octoraro Phyllite, and Wissahickon Schist), Keppie (1970) confirmed the existence of the Martic overthrust. Structural analysis reveals that Harpers Phyllite is distinct from the Octoraro Phyllite and Wissahickon Schist. He concluded that the Octoraro Phyllite and the Wissahickon Schist are Precambrian in age and were subjected to the same deformation history, supporting the existence of the Martic thrust between the Octoraro Phyllite and Conestoga Limestone. Lyttle (1982) supported the hypothesis that the Martic Line is an overthrust by studying the South Valley Hills (SVH) phyllite and a series of elliptical hills within Chester valley. The outcrop pattern of the phyllite allowed him to suggest that SVH phyllites consisted of Cambrian and older rocks like those of the higher Taconic slice in New England, thrust over the Conestoga Limestone (Lyttle, 1982). Along the northern contact of SVH, the phyllite is thrust over the Conestoga Limestone of Cambrian-Ordovician age. He found similarities between the high Taconic slices of Vermont and the SVH phyllite and thought that SVH phyllite might be allochthonous Taconic rocks in the Maryland-Piedmont area. In Duffy and Myer's study (1984), oriented thin-sections of phyllitic rock from six of the elliptically shaped hills and from the carbonate rock showed that both rock types were subjected to the same deformation and metamorphism. However they continued to support the idea of thrusting (Duffy and Myer, 1984). The Martic overthrust theory is related to an idea about the age of Wissahickon Schist. Knopf and Jonas (1949) concluded that the Martic Line must be an overthrust due to the younger age of Ordovician Conestoga limestone which underlies the Precambrian Wissahickon group.

Miller (1935) listed evidence that opposes the overthrust idea. If the Martic Line were an overthrust, one would expect that the low dip of the contact should produce a deeply embayed scarp along the south edge of Chester Valley. However the scarp is usually straight and there is an apparent lack of drag features. Wise (1970) thought the Martic Line to be the transition zone of the geosyncline, containing a significant facies change, overprinted by folds. Also he thought the Martic Line to be the facies boundary (unconformity boundary) due to the parallelism of facies overlap (Wise, 1970). Mackin (1939) also observed this field evidence and then concluded that the contact is a normal stratigraphic conformity. Mackin (1939) proposed in his paper, that the Martic Line is an exposed part of the horizontal overlying formation. The idea of an overthrust should be supported by the evidence of shallowly dipping beds. Rather, what Mackin had found was very steeply dipping beds. He ruled out the hypothesis of overthrust because of field evidence which is: (1) the formation parallelism on both sides of the contact (2) regular straight line contact instead of expected irregular line for overthrust (3) no brecciation (Mackin, 1939).

Geologists at Temple University recently proposed that the Martic Zone is a large scale shear zone (Myer *et al.*, 1985; Hill, 1987). The Octoraro Phyllite and Peters Creek Schist both show similar ductile shear-strain structures. Asymmetric tails, rotated micas, ribbons of quartz, C-surfaces, pressure shadows, undulose extinction and polygonal quartz grains are all indications of shear strain fabrics (Myer *et al.*, 1985). Baker (1987) has studied the Octoraro Phyllite unit petrographically and structurally and found pervasive ductile deformation fabrics. Lineation, asymmetric augen structures, rotated megacrystals, similar folds, and pressure shadows are all evidence of shearing. After analyzing about 100 oriented thin-sections and outcrop scale samples, he concluded that the Martic Zone is a wide and laterally extensive shear zone. Investigation of the high temperature deformation history of the metamorphic rocks, especially in relationship to petrologic information and metamorphic history is providing good data for tectonic models (Hill, 1987). Hill suggested that the importance of transcurrent displacement across vertical shear zones may explain some juxtaposition of contrasting lithologies and terranes with different metamorphic history. Hill also suggests that the Martic Line lies within a major transcurrent shear zone, extending from South Central Pennsylvania possibly as far as Staten Island.

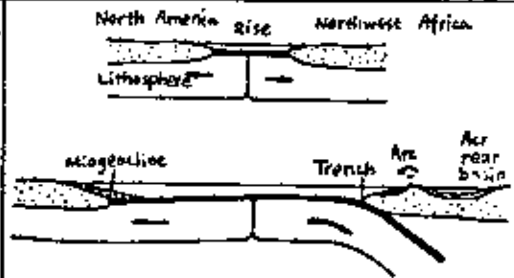
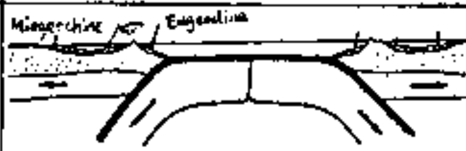
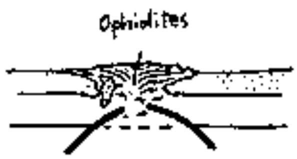
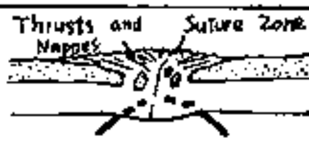
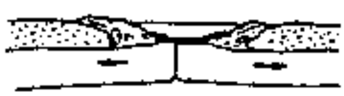
The results of this thesis are consistent with the idea of a shear zone. But as is mentioned earlier, it is not a goal of this thesis to determine whether the Martic Zone is a shear zone or a thrust.

2.1.3 Tectonic view of the Martic Zone

Recent debate about Appalachian tectonic history has been stimulated since Wilson (1966) proposed the model of the reopening of Iapetus ocean. It is generally accepted that the Appalachian orogenic belt was subjected to several mountain-forming events. Table 2 summarizes the tectonic events and their ages. Even though this simplistic evolutionary scheme describes the overall history of the Appalachian belt, the detailed history at any locality is considerably more complex (Condie, 1982). Numerous studies on the southern Appalachian orogeny have been done.

Condie (1982) summarized that the southern Appalachian Mountains can be divided into several tectonic zones: the Valley and Ridge province, the Blue Ridge province, Inner Piedmont, Charlotte and Carolina Slate Belts in order eastward. According to Condie's summary, the metavolcanics and

Table 2. Summary of Appalachian tectonic history by Condie. Schematic cross-sections from North America to Africa illustrating the opening and closing of the Iapetus Ocean and the opening of the Atlantic Ocean.

Age	Cross-Section	Comments
Precambrian		Opening of Iapetus. Accumulation of sediments at passive continental margin
Cambrian		Development of island arc
Ordovician		Activation and collisional orogeny began, recording the Taconic orogeny. Major westward thrusting and klippen formed. Plutonic and metamorphic activity occurred.
Devonian		Africa and Europe collide with the northeastern coast of North America. Thrusting, metamorphism, and magnetism record the Acadian and Caledonian orogeny.
Permian		Compressive forces continued producing large continent-directed thrust sheets and nappes, the Alleghenian orogeny.
Triassic		The opening of the Atlantic Ocean began in the Triassic and is continuing today.

metasediments of the western Blue Ridge province were deposited in the ocean basin between North America and the Blue Ridge-Inner Piedmont province block. Subduction and related volcanism began beneath the Charlotte Belt-Carolina Slate Belt province by 659 m.y. ago and along the western margin of the Blue Ridge-Inner Piedmont province by 500 m.y. ago. The small ocean basin between the Blue Ridge-Inner Piedmont block and North America closed and cratonic sediments derived from America begin to accumulate by 500 m.y. ago. This closure between 450 and 500 m.y. produced the Taconic orogeny. At this time the major thrusting of the Blue Ridge and Piedmont over the North American Plate occurred. Such thrusting must involve decoupling of the lower crust and mantle lithosphere from the upper crust, which alone is involved in the thrust. After this closure sediments were derived primarily from the east. The Acadian orogeny, between 350 and 400 m.y. ago, was characterized by widespread deformation, metamorphism and plutonism and appears to have been triggered by the closing of the ocean basin between the Blue Ridge-Inner Piedmont and Carolina Slate Belt. By this time the Blue Ridge and Inner Piedmont were accreted to the North American Plate. The King Mountain Belt may be a surface remnant of the collisional zone between the Inner Piedmont and Carolina Slate Belt provinces. After the Acadian orogeny, a new convergent boundary developed beneath the Carolina Slate Belt, dipping to the west. The Alleghenian orogeny (250-300 m.y. ago) resulted from the closing of the Iapetus Ocean and the collision of Africa with America. This collision produced extensive plutonism and large-scale overthrusting directed toward both the American and African Plates. The African Plate was again fragmented from the North American Plate in the Triassic as the the Atlantic Basin began to open. The line of rifting is east of the Carolina Slate Belt, buried today by Atlantic

Coast cratonic sediments (Condie, 1982).

Recently, the Brevard Zone in the southern Appalachians has been interpreted as a huge scale shear zone (Vauchez, 1987). The Inner Piedmont and Blue Ridge provinces show a marked dissimilarity in geology on either side of the zone. Detailed kinematic analysis of the mylonites in the Brevard Zone lead him to say that the zone is compatible only with right-lateral strike-slip faulting along a moderately dipping shear zone several *km* wide, causing southwestward displacement of the Piedmont relative to the Blue Ridge province.

2.2 The Bouguer Gravity Anomalies

2.2.1 gravity anomalies

Gravitational force is a universal phenomenon that has been recognized since Newton. All masses on the earth are subject to the earth's gravitational field. Since the earth is a non-uniform rotating sphere, the gravity is not uniform from place to place. Gravity variations in gravity can be caused by the departure of the earth from a spherical form or lateral contrasts in the density of subsurface rocks. Any geologic conditions that result in a horizontal variation in density will cause a horizontal variation in gravity or a "gravity anomaly" (Grant and West, 1965).

The field data of gravity differences between an arbitrary reference point and a series of stations are subject to various extraneous effects which are unrelated to the subsurface geology. That is the reason why gravity measurements need to be corrected. The Bouguer anomaly Δg_B is expressed as

$$\Delta g_B = g_{obs} + C_F - C_B + C_T - g_\phi$$

g_{obs} : the observed gravity

C_F : the free-air correction

C_B : the Bouguer correction

C_T : the terrain correction

g_ϕ : the normal sea-level gravity at a latitude ϕ

Basic knowledge for correction will be discussed in Chapter 4.

These anomalies would be zero if there were no horizontal variations in the density of rocks below sea-level. Gravity is measured by a very sensitive instrument called a gravitimeter. In the c.g.s system the unit for g is cm/s^2 and the practical unit is *miligal* ($mgal$). These units are related by the equation 980 $gal = 9.8 m/s^2$.

Gravity anomalies for a region are usually summarized on gravity anomaly maps. The gravity contour lines are iso-gravity anomalies and the negative sign represents anomalies under the expected reference gravity while positive sign represents anomalies above.

Negative gravity anomalies have been a target for finding salt domes. Salt rises through the heavier overlying strata as a result of differential hydrostatic pressure. Being a lighter substance than the rock which it displaces, salt will diminish the gravity field locally (Grant and West, 1965). Oppositely very dense, heavy rock causes a high positive gravity anomaly.

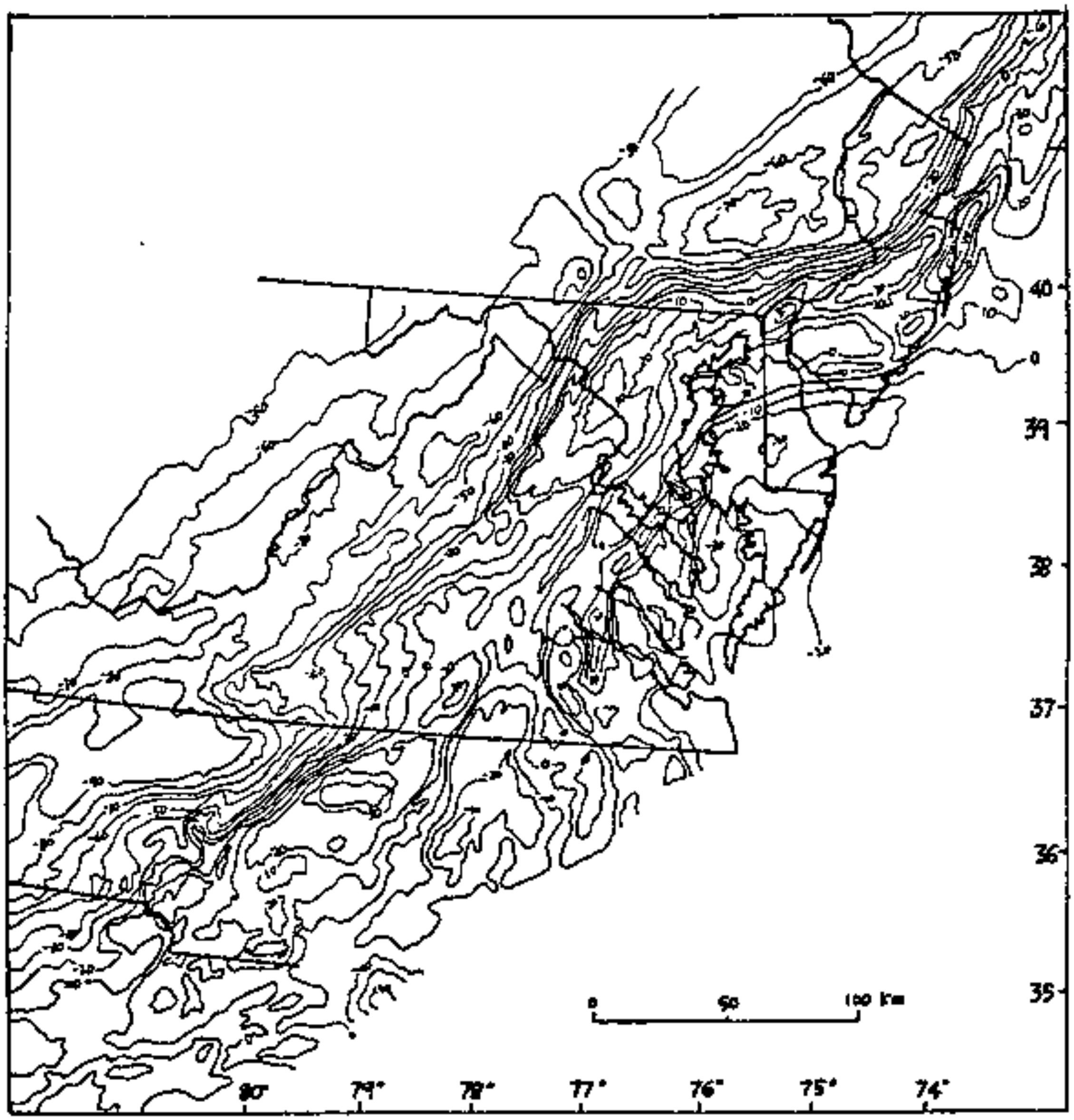
The gravity method has proven to be useful in mapping structural uplifts and depressions. If there are significant density contrasts between thick sections of sedimentary columns, they would yield observable gravity anomalies. If density contrast is found among the materials involved in these displacement, positive gravity effects will be observed over those regions where the heavier material is uplifted, and negative effects where it is depressed and replaced by lighter material (Grant and West, 1965). Bouguer gravity anomalies in most continental areas reflect subsurface rock types and structural discontinuities, whereas in ocean areas they reflect the presence or absence of anomalous upper mantle (Condie, 1982).

2.2.2 Bouguer gravity anomalies of the Martic Zone

There is a long belt of positive-negative paired gravity anomalies that extends through the Central Appalachians. Figure 3 shows closely spaced gravity contours separating negative anomalies to the northwest, positive to the southeast. This gravity pair runs the entire length of the Appalachian orogenic belt with no consistent relationship to surface geology. Gravity anomalies in the study area are shown in Figure 1a. Figure 3 is from "Bouguer gravity anomaly map of the Appalachian orogen" by Haworth *et al.* (1980). The map actually used to draw profiles was "Regional Bouguer anomaly field composed of wavelengths greater than 100 *km*".

The center of negative anomalies is in northwestern Pennsylvania and the center of positive anomalies is in southeastern Pennsylvania in an outcrop area of Baltimore Gneiss which is very dense rock. Considering the density interpretation, we expect to see the lighter rock to the north and heavier rock to the south. Especially, the most negative values should coincide with the lightest rock in this area. In a recent gravity and magnetic study of Lancaster County in the vicinity of Mine Ridge, Eisner (1986) interpreted residual gravity anomalies that are the anomalies remaining after the regional gravity anomalies are subtracted. He filtered the boundary between the crust and the mantle through the removal of regional anomaly trends. Lateral density inhomogeneities originating in the deeper crust or mantle do not bear on the interpretation of the residual anomalies. Unlike residual anomalies, Bouguer anomalies contain no assumptions about lower crustal structure and can be modeled to examine possible non-isostatic configurations of the Moho (Snyder and Barazargi, 1986). This thesis, however, considers the regional gravity anomalies which result from deep density

Figure 3. Bouguer gravity anomalies around the central and southern Appalachians (Bouguer Gravity Anomaly Map of the Appalachian Orogen, Haworth, *et al.*, 1980)



inhomogeneities.

For this study, gravity profiles are made parallel to lines of longitude at 7′ 30″ intervals of longitude. It is usual to draw profiles along the direction perpendicular to the strike of major structures. But in this study, in order to compare profiles of gravity, density, and topography, easily repeatable sections are needed. Longitude is convenient to use for the cross-sections (Figure 4). These cross-sections are approximately, but not exactly, perpendicular to the strike of major structures. Figure 5 is 8 profile-sets at 1° intervals. The scale on the gravity anomaly axis is units of mgal with an arbitrary proportional scale for the grid (0.5 cm).

Almost all of the gravity profiles show the steep gradient between negative and positive anomalies near 40° N. Intervals of 74°–75° W longitude are not complete due to data deficiency at the Atlantic ocean. Gravity cross sections of 75°–76°, 76°–77° clearly show the steep gradient. Local positive highs of gravity show good agreement with dense rock such as gabbroic gneiss. But the light sand of the coastal plain is not a general explanation for the positive gravity anomalies. Figure 6 shows the gradient locations. Gravity gradient belts are defined by the horizontal distance of straight line on a profile (Figure 7). An average distance of the gravity gradient belt is 23 km. 23 km is about one tenth of 221.25 km which is the entire distance of horizontal axis from 41° to 39°. Gradient belt exactly coincides with the Martic Zone.

2.2.3 Gravity interpretation of the Appalachian belt

Recently Eisner (1986) used residual gravity and magnetic data to study the Martic Line. Eisner's geophysical data suggested that the Tucquan

Figure 4 Cross-section lines. Gravity, density, and topographic profiles are made parallel to lines of longitude at 7' 30" intervals.

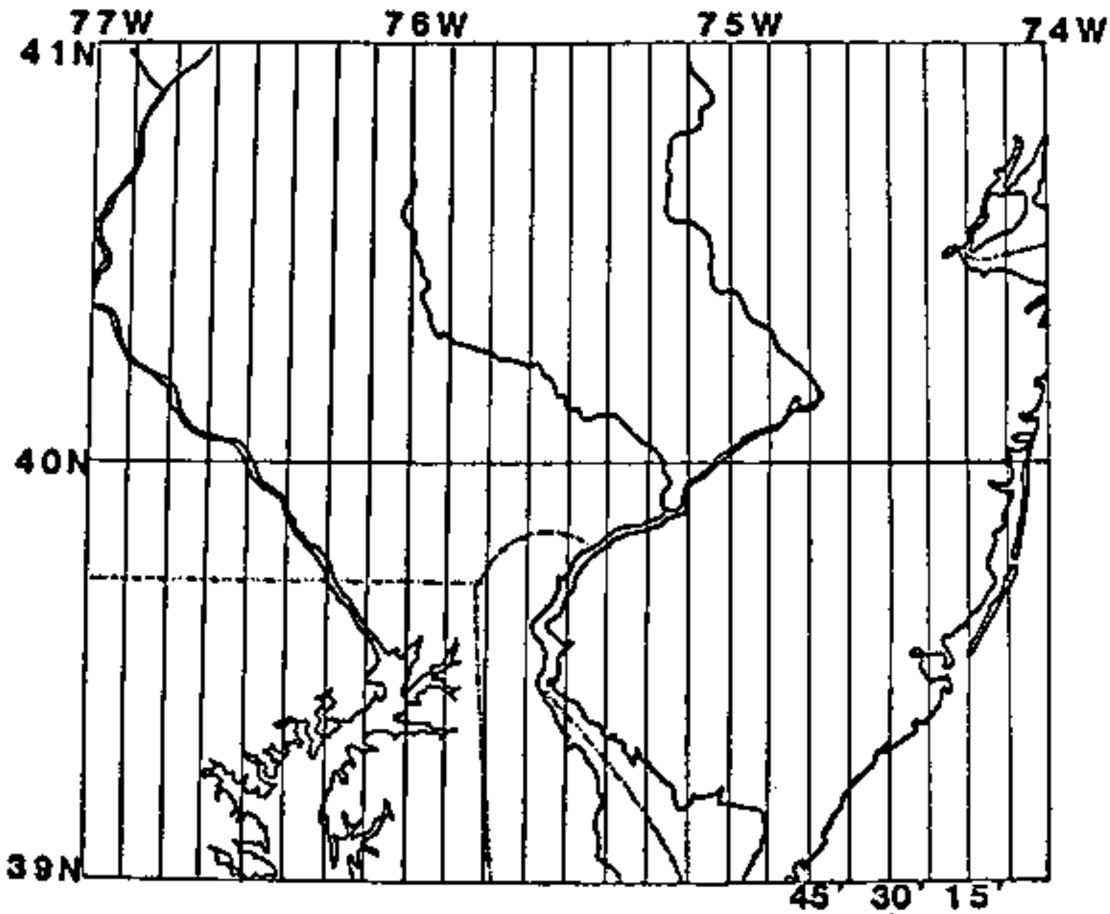


Figure 5a. Gravity profiles of 74°–74° 52′ 30″ W. All curves (74°, 74° 7′ 30″, 74° 15′, 74° 22′ 30″, 74° 30′, 74° 37′ 30″, 74° 45′, 74° 52′ 30″) clearly show the steep gradient around 40° N.

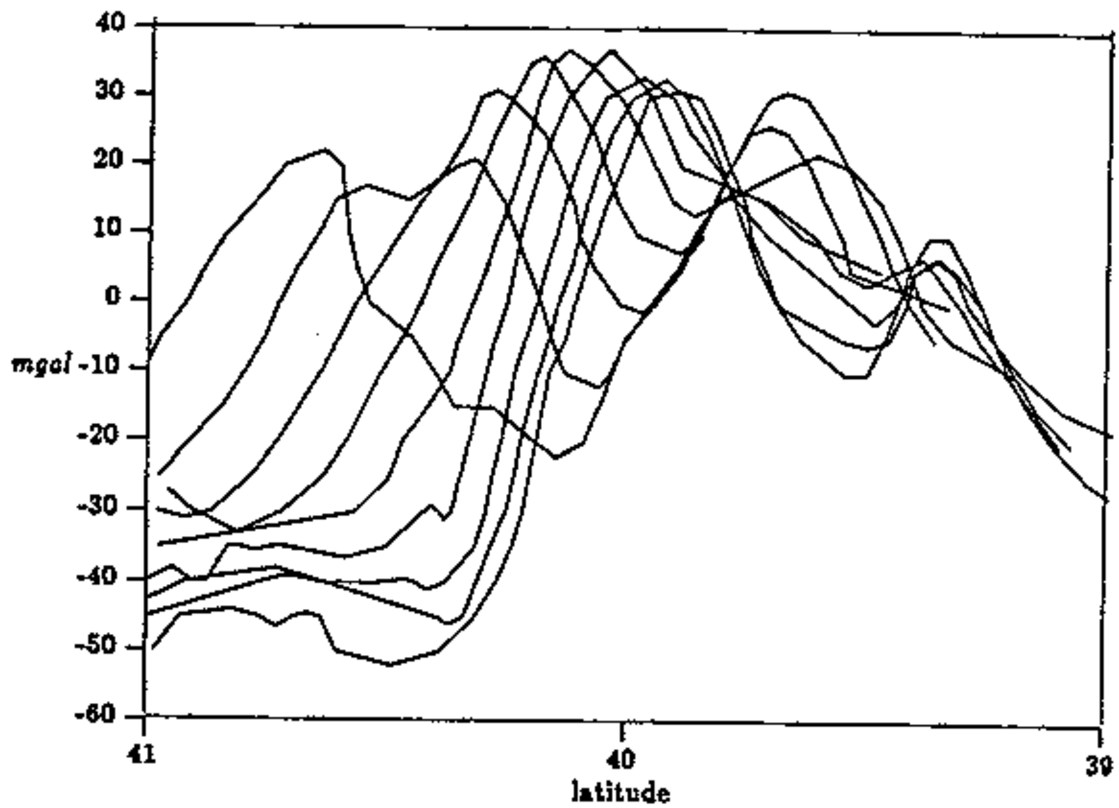


Figure 5b. Gravity profiles of 75°–75° 52′ 30″ W

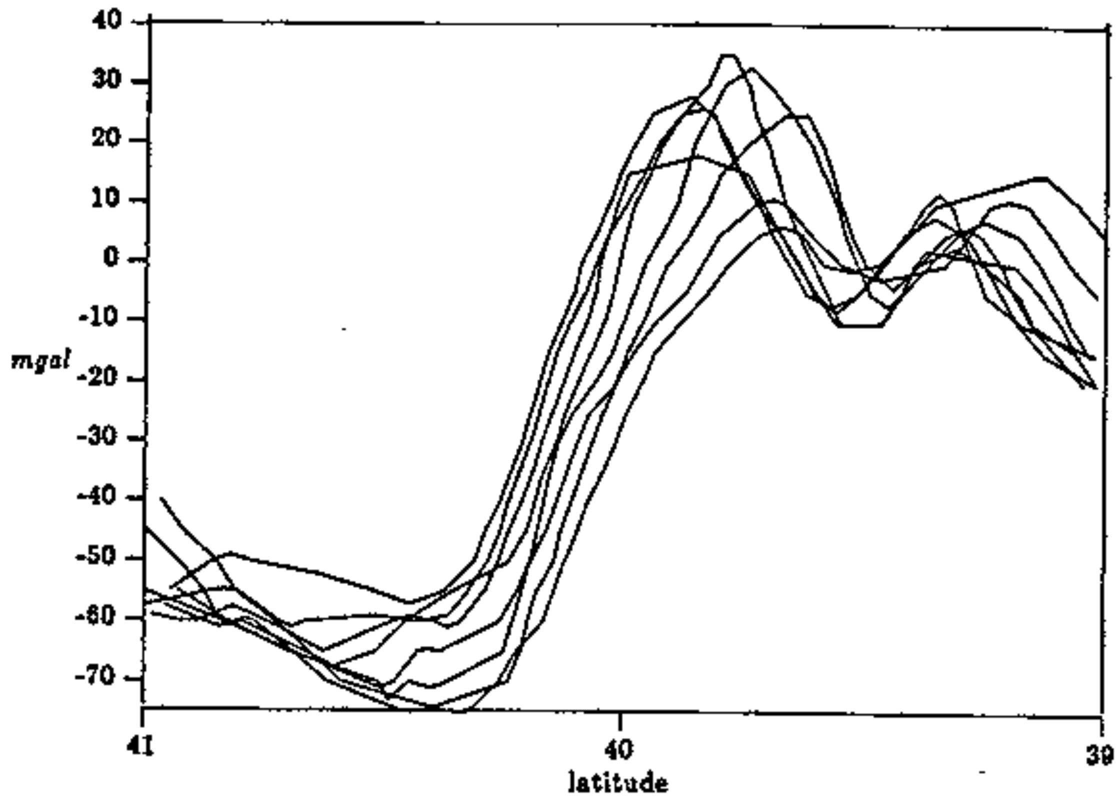


Figure 5c. Gravity profiles of 76°–77° W

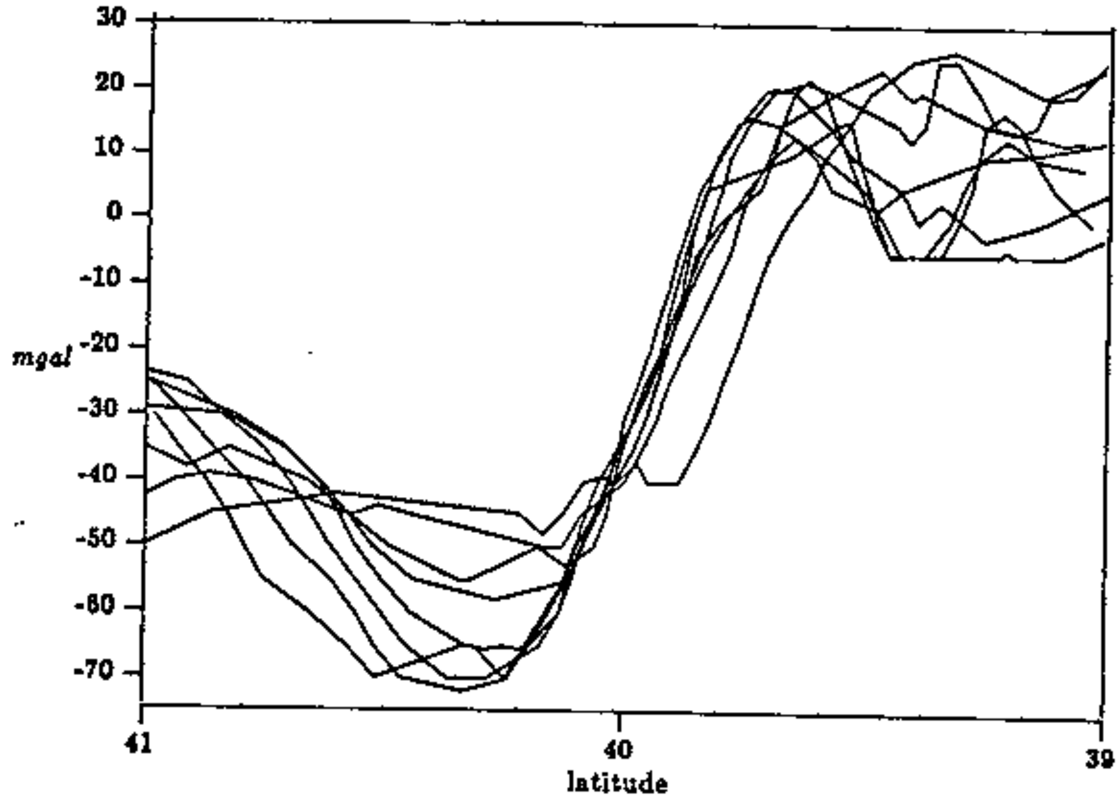


Figure 6. Gravity gradient belt

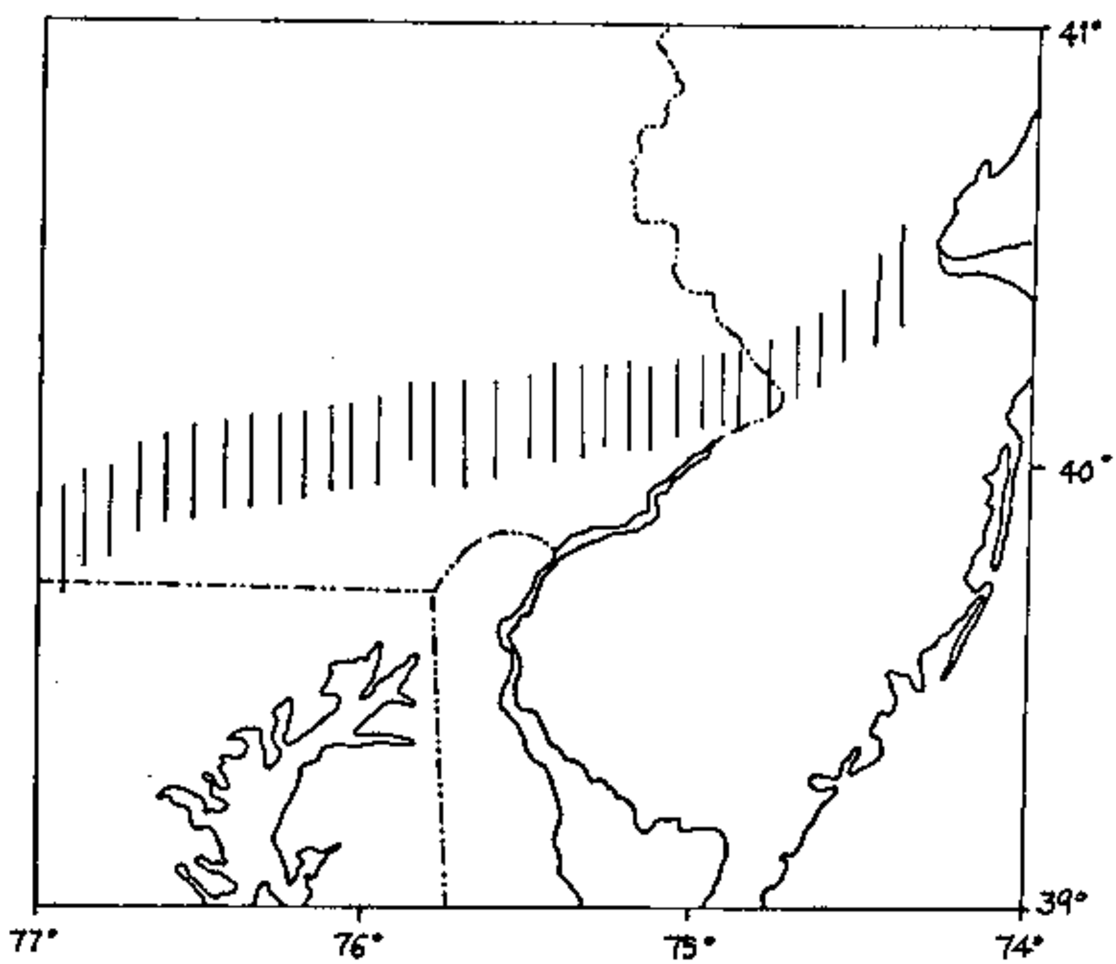
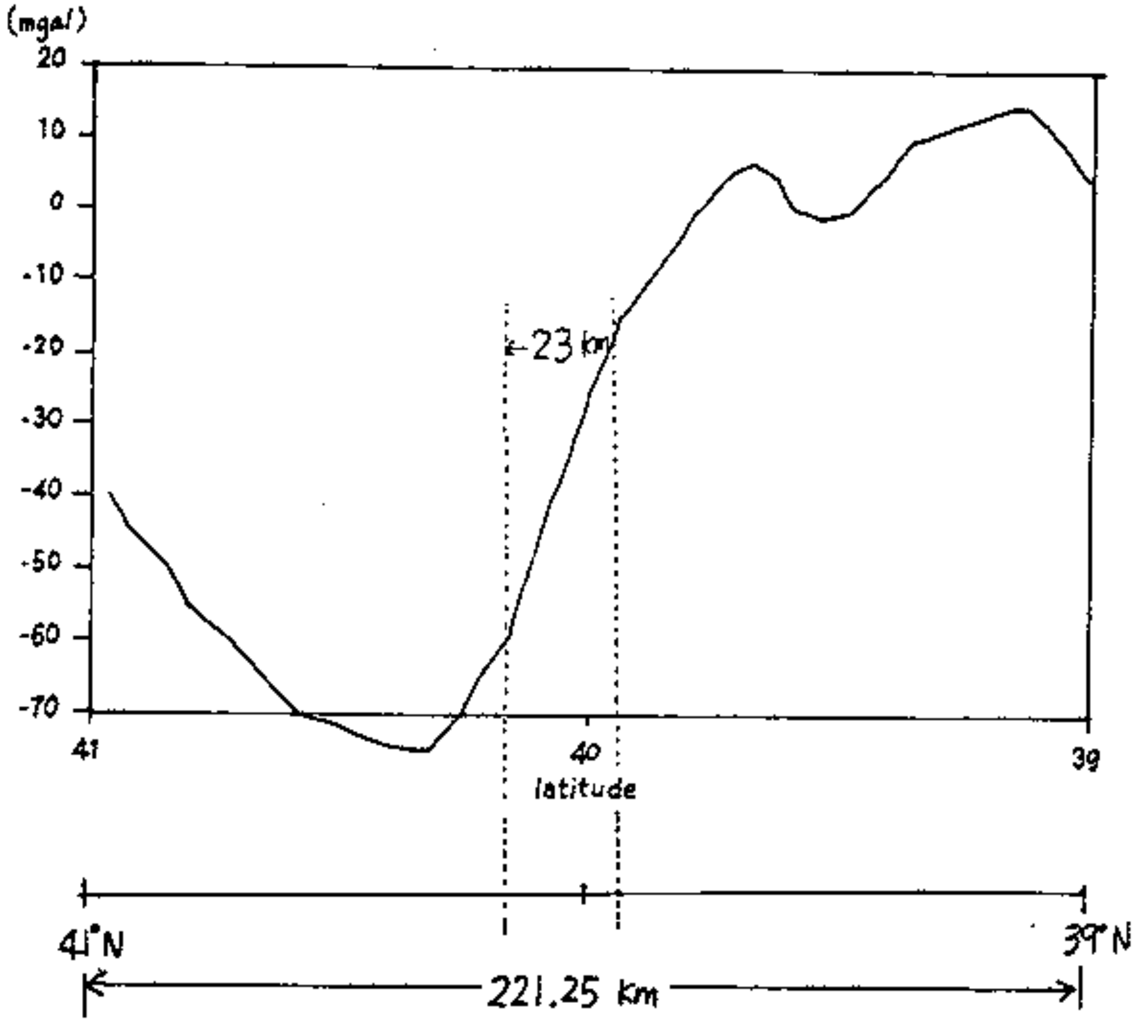


Figure 7. Gravity gradient belt distance. Gradient belt is defined by the horizontal distance of the straight line on a profile.



overthrust, a thrust ramp, coincides with the magnetic Martic Line. Linear bands of Antietam are proposed to be allochthonous klippen stratigraphically equivalent to rocks south of the Martic Line. Eisner concluded that the Mine Ridge allochthon overlies Chilhowee group clastics, either structurally or stratigraphically.

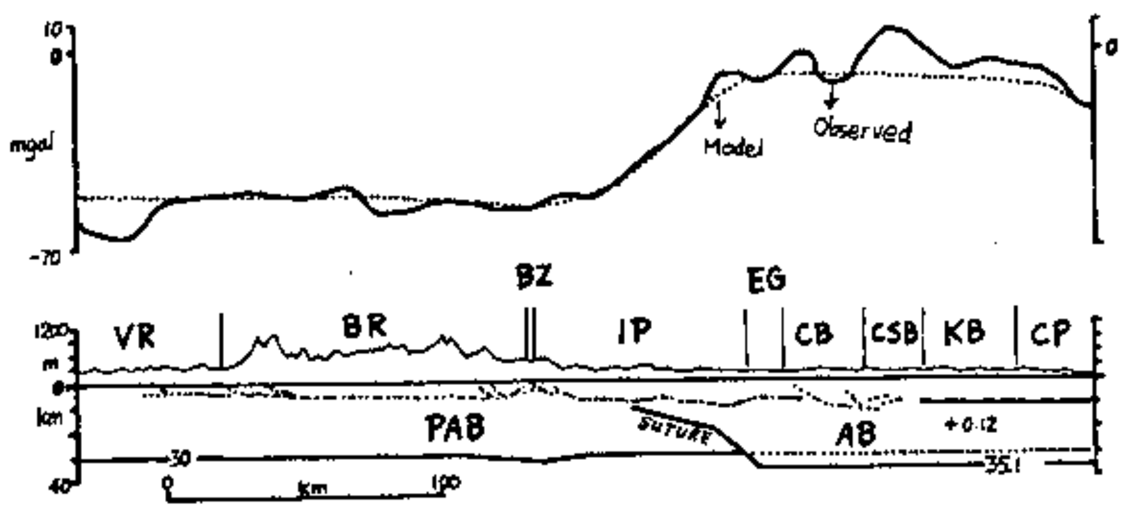
The strong coincidence between the Appalachian orogenic belt and Bouguer gravity anomalies leads to the assumption that the origin of Appalachian Gravity Parallelism (AGP) has something to do with tectonic history. The lack of a consistent relationship between either anomaly and a singular geological feature suggests that the anomalies, in large part, are generated by a large-scale buried feature (Thomas, 1983).

Lately enthusiastic discussion (Thomas, 1983; Cook, 1984; Cook *et al.*, 1979) is going on about AGP. Interpretations of this feature concentrate on two major ideas. The first one is that a crust- and lithosphere-penetrating boundary (suture) formed during Paleozoic plate collision between North America on the west and an island arc on the east, which acts as a "root zone" for Alleghenian detachment to the west. The second is that a relatively undeformed late Precambrian continental margin transition from North American crust in the west to oceanic or "rift-stage" crust on the east was subducted beneath the Inner Piedmont and Charlotte Slate Belt.

Cook *et al.*'s model (1979) supports the displacement of an overthrust. They analyzed the COCORP reflection data and found horizontal layering. It is deduced that the major tectonic feature in the Southern Appalachians is an overthrust that transported crystalline rocks over sediments. Furthermore, much of the crystalline Appalachian is allochthonous and overlies an extensive area of relatively undeformed lower Paleozoic sediments (Cook *et* *al.*, 1979). But how could the deformed thicker (7-15 *km*) allochthonous sheet be thrust over the thin (4-6 *km*) undeformed sediments? That remains unsolved. An interesting thing is that the gravity gradient coincides exactly with the Charlotte Slate Belt where they showed a boundary of contrasting crustal thickness. Cook (1984) carefully collected geophysical data and suggested that interpreting with one geophysical datum may mislead us. For example, COCORP reflection data can be explained as (1) layered metasediments, (2) fault, (3) manifestation of a decollement "root zone". Adding conductivity data and magnetotelluric measurement data supports the theory of the continuity of shallow highly conductive rock from the Valley and Ridge to the Charlotte belt. Cook (1984) tried to look at AGP with a tectonic view and use all the available geophysical data, which is a good approach to regional geology.

It is concluded that the belt of steep gradients of AGP separating regions of negative and positive gravity anomalies is associated with a major collisional suture that separates crusts of contrasting density and thickness (Thomas, 1983). Figure 8 shows the tectonic and gravity model of the southern Appalachian orogeny. The northwestern block is referred to as the "proto-American block", and the denser and thicker southeastern block is called the "accreted block". Thomas (1983) suggested that the proto-American plate subducted southeastward beneath the accreted plate. The thickness of the two blocks was constrained to be approximately 30 *km* for the proto-American block and 35 *km* for the accreted block. The suture dips 45° at depths greater than 10 *km* below the suture. Although Thomas pointed out many good ideas, it is unexpected that lighter, thinner crust would subduct beneath the denser, thicker younger crust. Besides, without the evidence of subduction-related volcanics, it is ambiguous to determine

Figure 8. Thomas' (1983) tectonic model of southern Appalachian with the Bouguer gravity anomalies. AB: Accreted Block, PAB: Proto-American Block, VR: Valley and Ridge, BR: Blue Ridge, BZ: Brevard Zone, IP: Inner Piedmont, EG: Elberton Granite, CB: Challotte Belt, CSB: Carolina Slate Belt. Heavier dotted lines show the master decollement and subsidiary thrust. Accreted block has 0.12 g/cm^3 more dense crustal materials than that of Proto-American Block.



the direction of paleosubduction.

Although these two opinions on AGP are different, both Thomas and cook look at AGP from a tectonic point of view and use all of the available geophysical data. They both agree with the idea of crustal thinning. The major cause of the gravity change is eastward crustal thinning (Cook, 1984).

Chapter 3. Rock Density

3.1 Sampling and Measurement

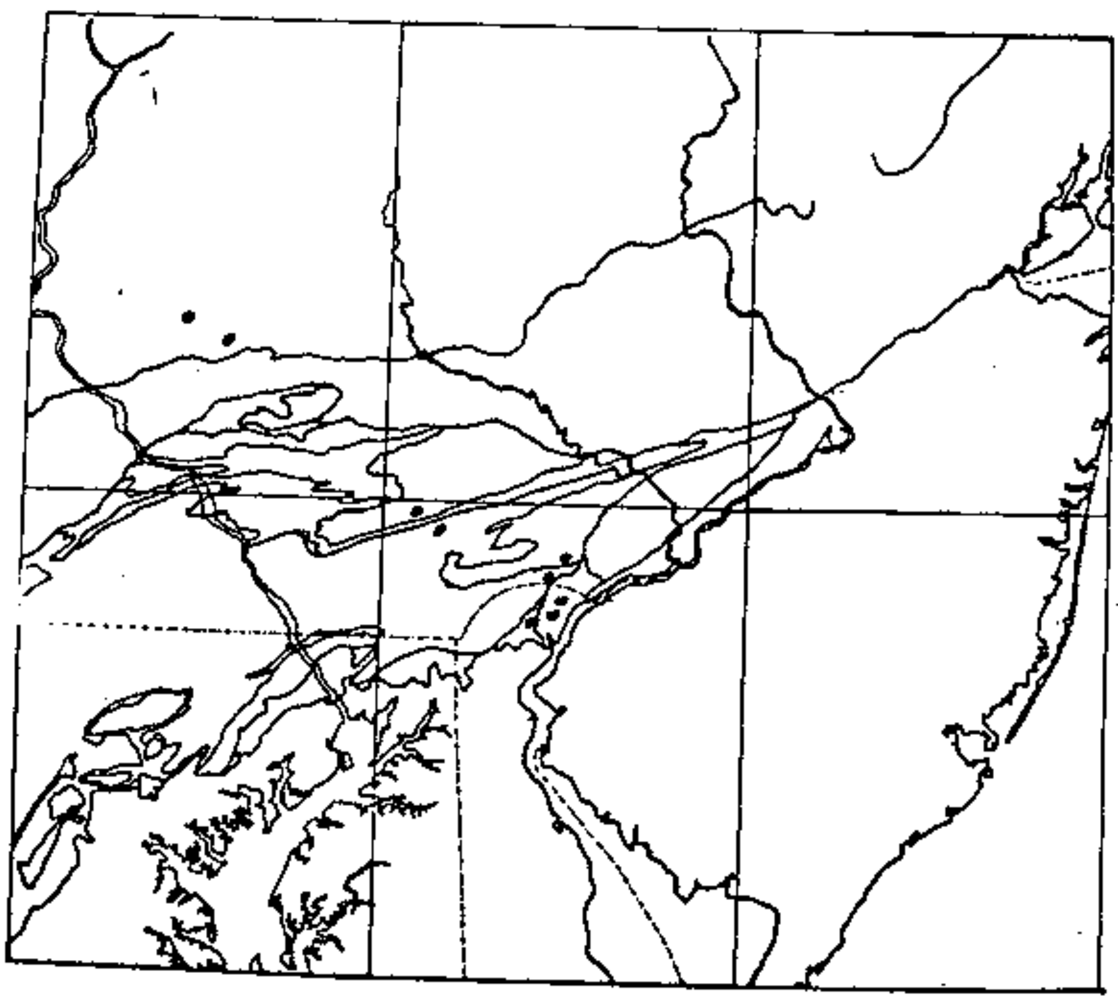
Density of rock is the most significant property to consider in the interpretation of gravity anomalies. Local gravity anomalies have been explained by the distribution of rock of various density. The strength of the gravitational field of a body is in proportion to its density (Grant and West, 1965).

The major properties that affect the rock density are composition, age, texture, and metamorphic grade. Among these factors composition is more important for igneous rock. Mafic igneous rock is more dense. The higher the metamorphic grade is, the denser the rock. Sedimentary rock has a wide range of density depending upon its degree of compaction and age.

The density of any particular lithology may vary widely. For instance, quartzite has the range of 2.3 - 2.8 g/cm^3 density (Judd and Shakoor, 1981). Sandstone has the range of 1.4 - 2.9 g/cm^3 due to a wide range in the composition. The lower values are for tuffaceous sandstone and the higher values for calcareous or argillaceous sandstone. Densities may also vary because some parts of rock contain heavier and/or lighter minerals than other parts. In order to avoid these problems, mean bulk densities for lithologies in the study area are taken from literature data (Judd and Shakoor, 1981). Some samples collected in the field are measured to compare with the literature data. Figure 9 shows the location of outcrops where rock samples were collected. Some of the measured densities are anomalous, but the number of samples measured was too small to attribute significance to these anomalies.

The purpose of the analysis of the surface rock density is to see how

Figure 9. Locations of sample outcrops



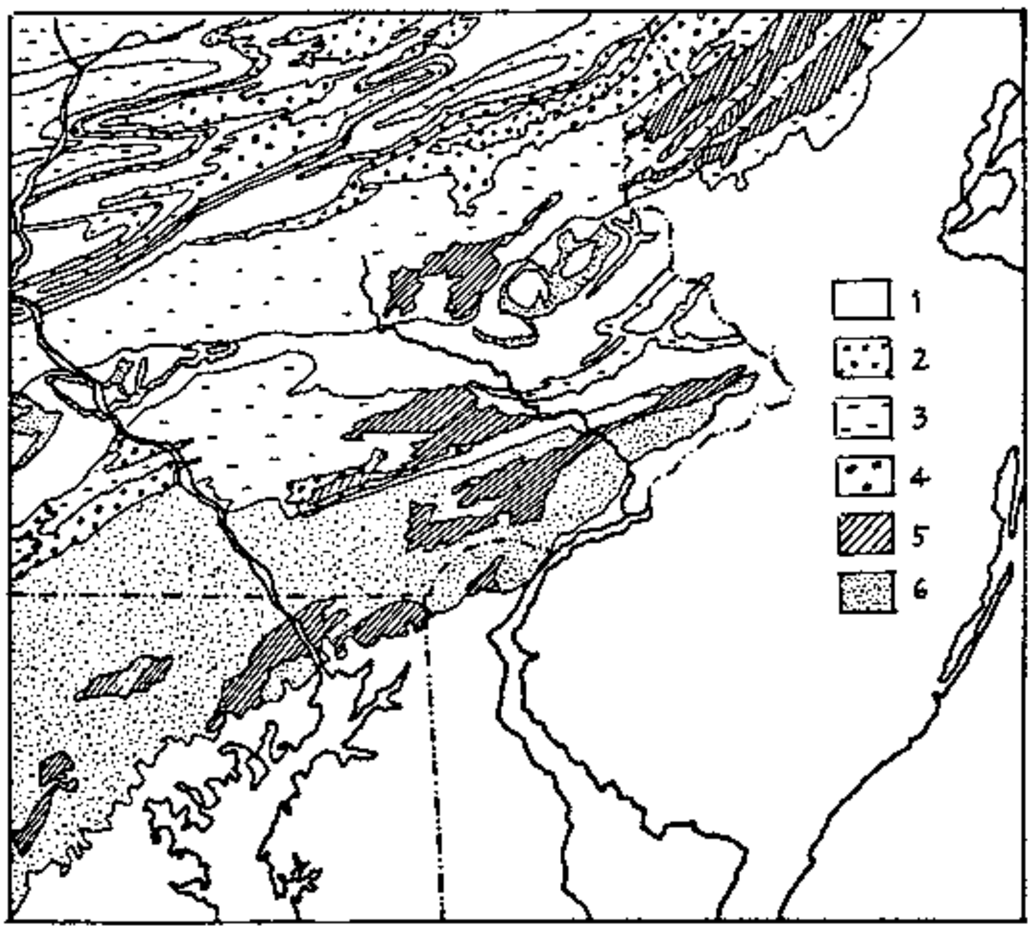
the rock density and the gravity anomalies are related. Any rock units that have the same density value belong to a density group in this study. Six groups of rock density are presented in Table 3. For example, group 4 includes Basalt, Granite, Dolomite, Quartzite, and Slate, which represent all kind of rock types. The map "Rock types of Pennsylvania" (Berg *et al.*, 1984) is used as a basis to map these six density groups (Figure 10).

Collected rock samples were broken into 25 - 50 gram size pieces to be measured by Jolly-Balance. There are two types of density measurements: bulk density and grain density. The measurement made here was bulk density, which is the volume density of specimens of macroscopic size. On the other hand the grain density is the density of the rock-forming minerals in powder form (Grant and West, 1965). Table 3 shows the individual density value for each measured hand sample and average group densities.

Table 3. Density measurement chart (Grant and West, 1965; Judd and Shakoor, 1981)

Group	Density (g/cm^3)	Rocks	Measurement
1	$<$ 2.19	moraine	—
		sand	—
2	2.2-2.39	sandstone	1.9
3	2.4-2.59	limestone	3.7
		dolomite	—
		shale	—
4	2.6-2.69	granite	—
			—
			—
		serpentinite	—
		quartzite	—
		slate	—
5	2.7-2.79	felsic gneiss	3.7
			3.9
			2.75
			2.85
			2.13
			—
		basalt	—
		marble	4.6
			—
			—
6	$>$ 2.8	schist	—
			—
			—
		phyllite	1.46
			1.7
			—
		gabbro	—
		mafic gneiss	4.4
			3.06
			—
			—
			—
			—

Figure 10. Six typical lithogroups in the study area



3.2 Density Profiles

The rock types of Pennsylvania is regrouped according to "density groups". Six different rock groups are shown on the map, and indicated by symbols (Figure 10). The area southeast of the Martic Zone, which contains various ages and grades of schist and phyllite, is united into one huge group. Coastal sand and the Triassic sandstone are grouped into a same density group.

If the areas northwest and southeast of the Martic Zone show generally two different rock types with significant differences in density, then the most reliable reason for the gravity gradient could be a rock density contrast.

As mentioned earlier, light rocks usually show negative gravity anomalies. One would expect to have the lightest rocks to the north of the Martic Zone, because the area northwest of the Martic Zone has the lowest gravity value (Figure 1a). Figure 11 has the same N-S section scale as figure 5, the Bouguer gravity anomaly profiles. Along the north-south section, profiles are made at 7' 30" intervals to compare with the gravity profiles. The following set of profiles match gravity and rock density. Nothing but the longitude axis corresponds to the gravity profiles. Rocks that belong to the density groups 1 through 3 have a density range of 0.2 g/cm^3, while rocks belonging to density groups 4 through 6 have a 0.1 g/cm^3 range. This is because groups 4, 5 and 6 show a great variety in rock type.

Figure 11a shows that the northern area of the gravity gradient zone is underlain by dense rock. Usually the negative gravity anomalies are related to less dense rock. And the southern area of the gravity gradient zone has the least dense rock. In both areas of the gravity gradient zone, low density group rocks are shown in the density profile but are not reflected in the

Figure 11a. Rock density-gravity profiles of 75° W

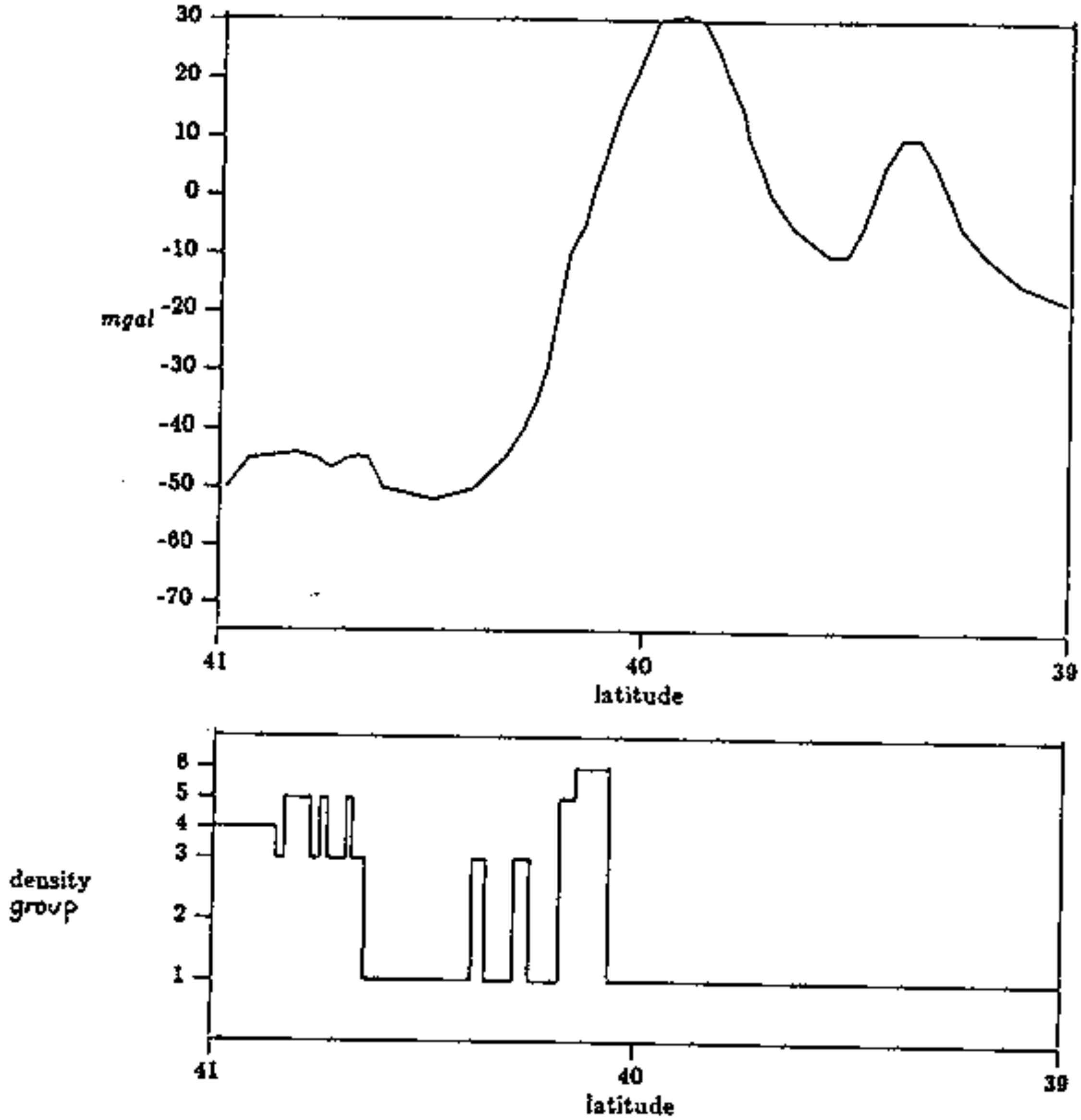


Figure 11b. Rock density-gravity profiles of 76° W

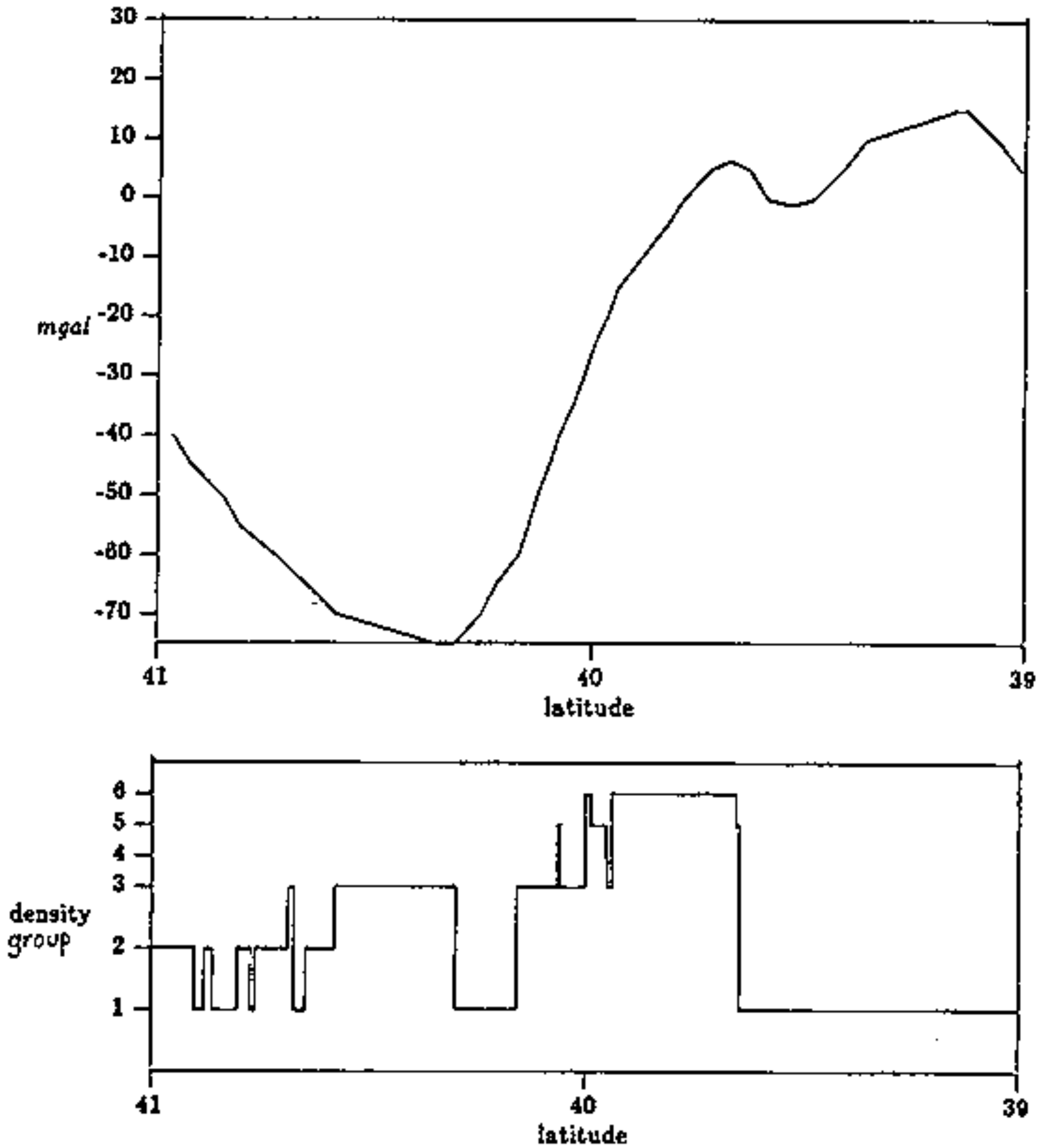
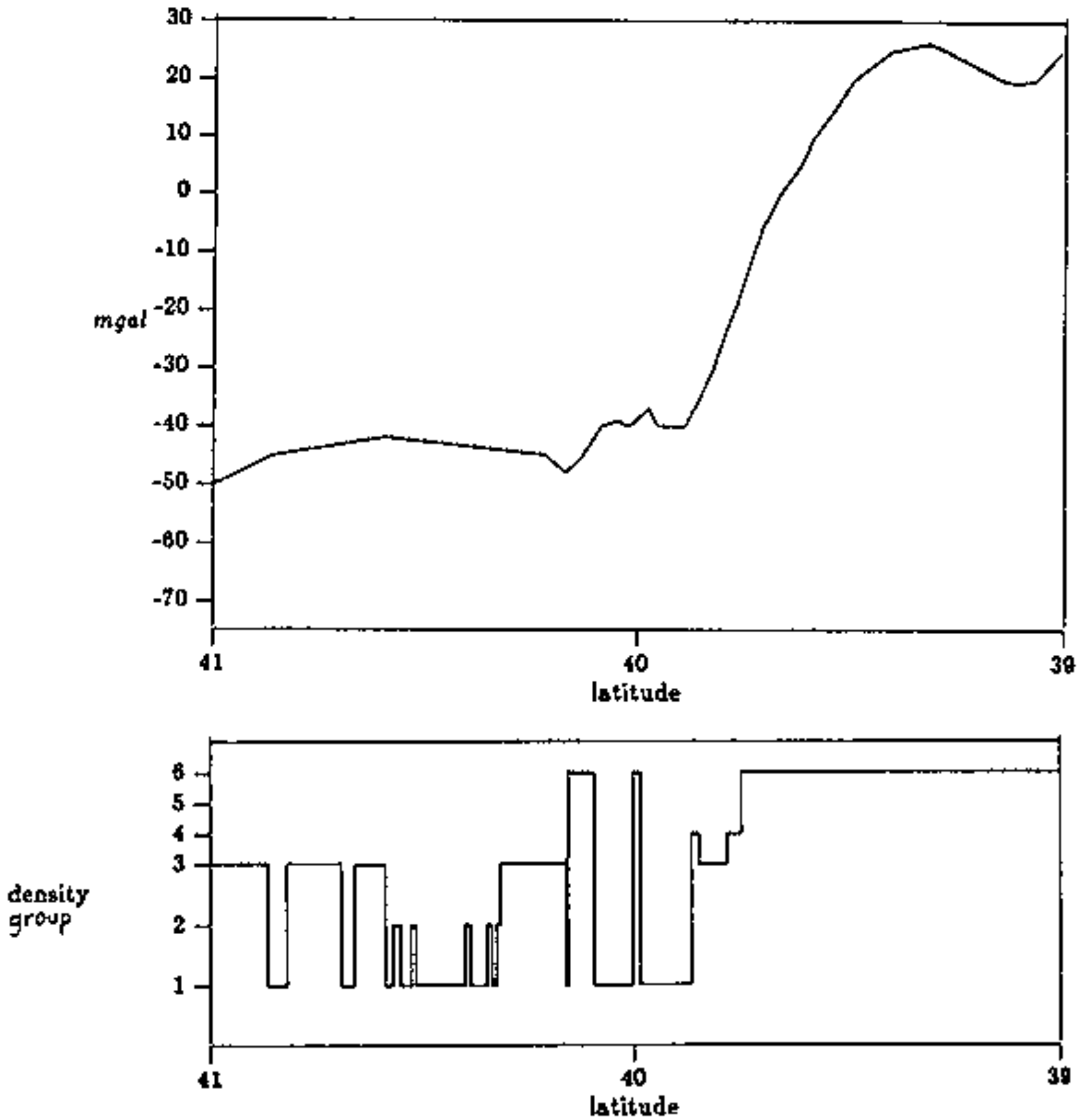


Figure 11c. Rock density-gravity profiles of 77° W



gravity profile. In Figure 11b, a 10 *mgal* gravity peak corresponds to rock of the highest density (schist, gabbro, and mafic gneiss). However, a 20 *mgal* gravity peak has the lowest density (moraine and sand). Furthermore, the gravity profile of Figure 11c shows a distinct change from -50 *mgal* to 30 *mgal* anomalies, while the density profile does not show a corresponding change. Apparently the rock density distribution does not coincide with the gravity anomalies. Density contrast across the Martic Zone can not be said to be the only cause for the gravity gradient.

The effects of rock density on the gravity field have been tested. No strong relation between the positive anomalies and the area northwest of the Martic Zone and negative anomalies and the area southeast of the Martic Zone has been found. Although there are some variations in density, they are insufficient to cause such a sharp gravity gradient at the Martic Zone.

Chapter 4. Topographic Relief

4.1 Influence on Gravity Anomalies

The gravitational effect of mass associated with topography is another important source of surface gravity anomalies. Gravity varies from place to place depending on the latitude, elevation, material and the shape beneath the measuring station. Since topography is considered here, the relation between topography and gravity anomalies will be discussed first.

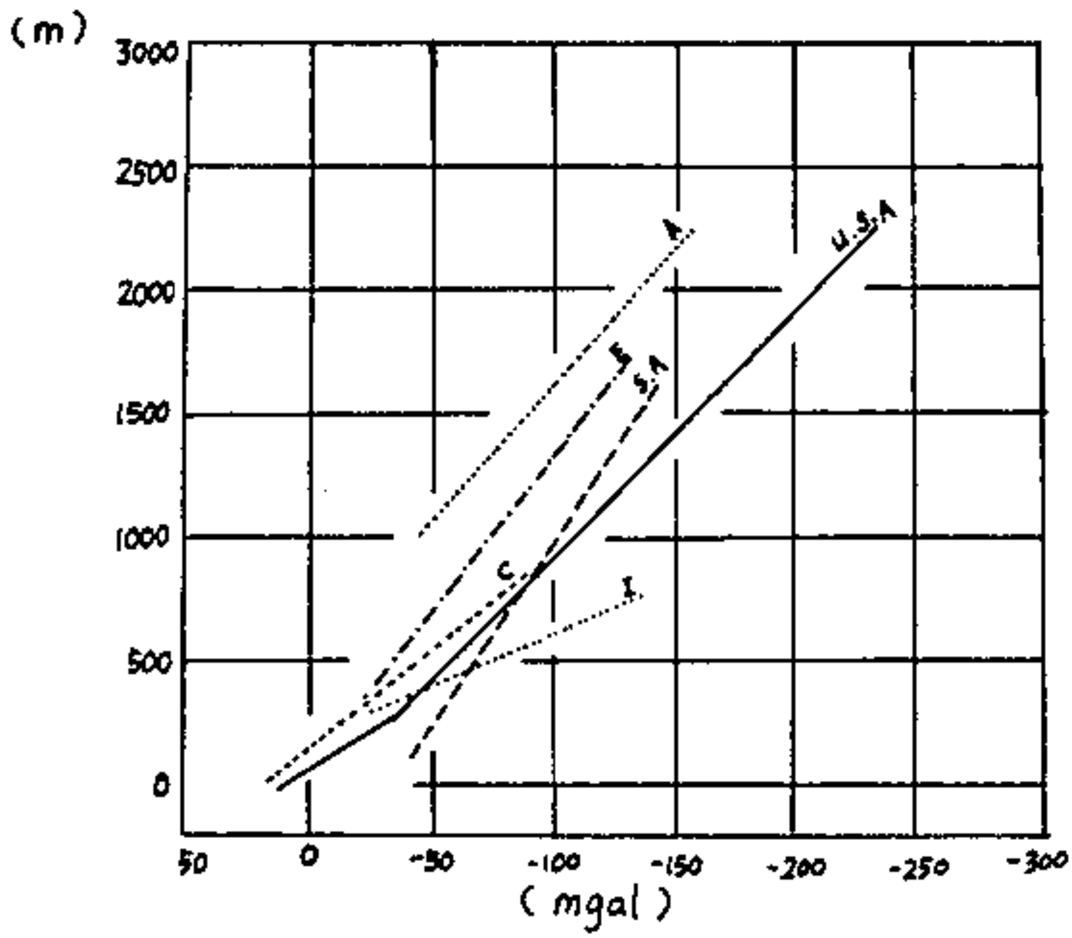
Bouguer gravity anomalies are idealy thought to be free of topographic effects. However, there are still some additional effects of mass distribution. Since higher topography has a deeper crustal root due to isostacy, the topographic effects on gravity can not absolutely be ruled out. Figure 12 shows the proportionality of topography and gravity anomalies. Notice the large relief of topography in this comparison. Greater than 500 km topography shows a significant influence on gravity fields. Although central Appalachia is not considered as a high relief area, how the topography and gravity profiles are related will be considered.

The Bouguer gravity anomaly Δg_B is given by

$$\Delta g_B = g_{obs} - g_\phi + C_F - C_B + C_T$$

where $g_{obs} - g_\phi$ is called the latitude correction. The observed gravity at the point must be corrected for latitude because, as the earth is not a perfectly shaped sphere, the gravity value for each latitude will increase polarward. The reference gravity field g_ϕ is defined by $g_\phi = 9.78031846(1+0.005278895\sin^2\phi+0.000023462\sin^4\phi)$ where ϕ is the latitude of the gravity measuring point. If the gravity measuring station is carried up to elevation h, the gravity value will be affected by the increase in distance

Figure 12. The relation between topography and gravity (after Sharma, 1976). Observed relations of $3° \times 3°$ mean elevations and Bouguer anomalies are proportional. U.S.A: United States, A: Andes, E: Europe, S.A: South Africa, C: Canada, I: India.



from the earth's center. To obtain surface gravity anomalies, the elevation-related value must be subtracted. The elevation correction C_F is described as Δg_h which has a value of $2hg_\phi/r_o$ (where r_o is the radius of the geoid sphere). The higher the station is, the lower the observed gravity will be, according to the inverse square root law of distance from the earth's center. That is why C_F is added when elevation is measured above the geoid. This value $(g_{obs}-g_\phi+C_F)$ is called the free-air gravity anomaly, which is independent of latitude. Even though the effect of latitude and elevation have been removed, the free-air gravity anomalies strongly correlate with local topography. In order to remove the gravitational attraction of the local topography, the Bouguer gravity formula is used. The Bouguer gravity anomaly Δg_B is useful when the topography is not too steep and crustal density is correctly chosen. Let us focus on the reason why the Bouguer gravity anomalies are applied at the short wavelength only (i.e., gentle slope topography). The elevation of topography can be generalized by

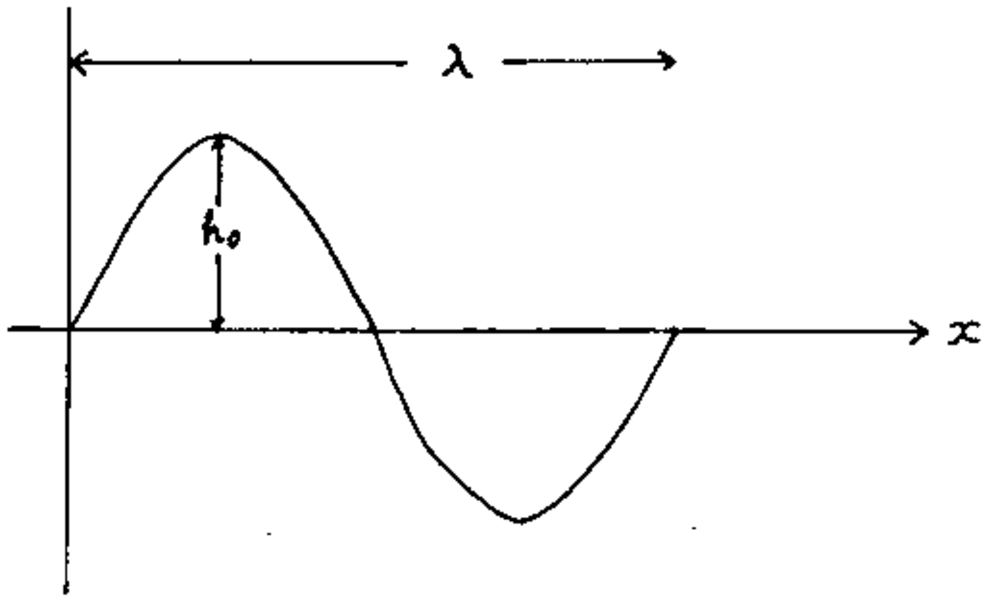
$$h=h_0\sin\frac{2\pi x}{\lambda}$$

where h=topographic height and λ=wavelength (Figure 13). The load on the lithosphere corresponding to the topography given by above formula is

$$q(x)=\rho_c gh=\rho_c gh_0\sin\frac{2\pi x}{\lambda}$$

Since the loading is periodic in x, it is clear that the response (ridge) or deflection (valley) of the lithosphere will also vary sinusoidally in x with the same wavelength as the topography. Short wavelength topography causes virtually no deformation of the lithosphere. The lithosphere is infinitely rigid for loads of this scale. And smooth slope topography is well fitted to the Bouguer gravity anomalies without the mass gravitational effect. But on the

Figure 13. Simplified topographic formula.



other hand, for topography of sufficiently long wavelength, the lithosphere has no rigidity and the topography is fully compensated; that is, it is in hydrostatic equilibrium (Turcotte and Schubert, 1982). It might be said that the Bouguer gravity anomaly is effective in removing the gravitational influence of the local (short wavelength) topography, while not effective in removing the influence of regional (long wavelength) topography. In about 10 *km* of small horizontal scale, mountain or valleys do not influence the density distribution at depth. Regional scale is usually said to be at least 100 *km* wavelength of mountain range.

4.2 Profiles

Four topographic maps are used to trace contour lines: Newark, Harrisburg, Baltimore, Wilmington in 1:250,000 scale, published in 1944, 1957, 1957 and 1927, respectively (USGS topographic map series). Since small variations in elevation are not likely to influence gravity measurements, only 500 *ft*, 1000 *ft* and 1500 *ft* lines are traced (Figure 14).

Along the same N-S longitude sections at $7' \, 30''$ intervals sections as Figure 4, topography was traced. Topographic profiles are drawn to compare with the gravity profiles at the same scale as the gravity profiles but with a vertical exaggeration of 100 times (Figure 15).

No matter what the vertical scale is, to have an influence on gravity, the topographic profile should show a mirror image of the gravity profiles. If there is a mirror image of gravity pattern, the magnitude of influence can be examined. However, as shown in Figure 15, there is no such correlation between the gravity and topographic profiles. The northern part of the area shows generally high elevation. If this elevation difference between 40° N and 39° N is the cause of the steep gravity gradient, one would expect a significant change in gravity anomaly. Figure 15b shows almost same profile pattern of the gravity and the topography at the area of $40^\circ - 41^\circ$, which is the opposite case of what we expect. The highest elevation of 1200-1800 *ft*. area is correlated to the negative 40-50 *mgal* gravity anomalies where is not the least. Figure 15c is a gravity-topography profile at 77° W. This gravity profile has a distinct change from -50 *mgal* to 30 *mgal* at south of 40°. But topographic profile shows limited peak of 1500 *ft* elevation at north of 40°.

If it is assumed that the topographic mass has a constant density of $2.7 \; g/cm^3$, how much topographic mass would we need to cause -70 *mgal* in

Figure 14. Topographic contours. Traced on Newark, Baltimore, Harrisburg, Wilmington Topographic Map. Solid line is for 500 *ft*, dotted line for 1000 *ft*, dash-dot line for 1500 *ft*.

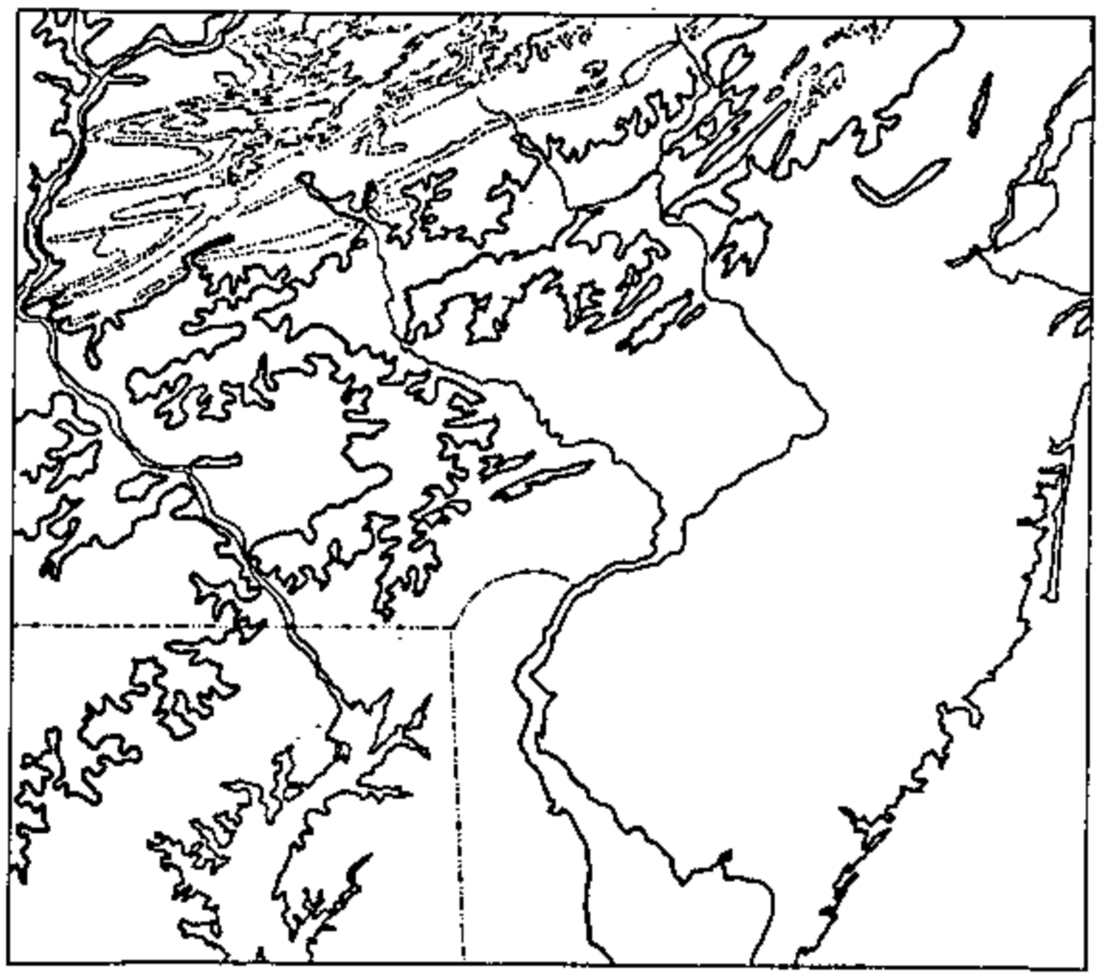


Figure 15a. Profile of topography with gravity profile of 75° W.

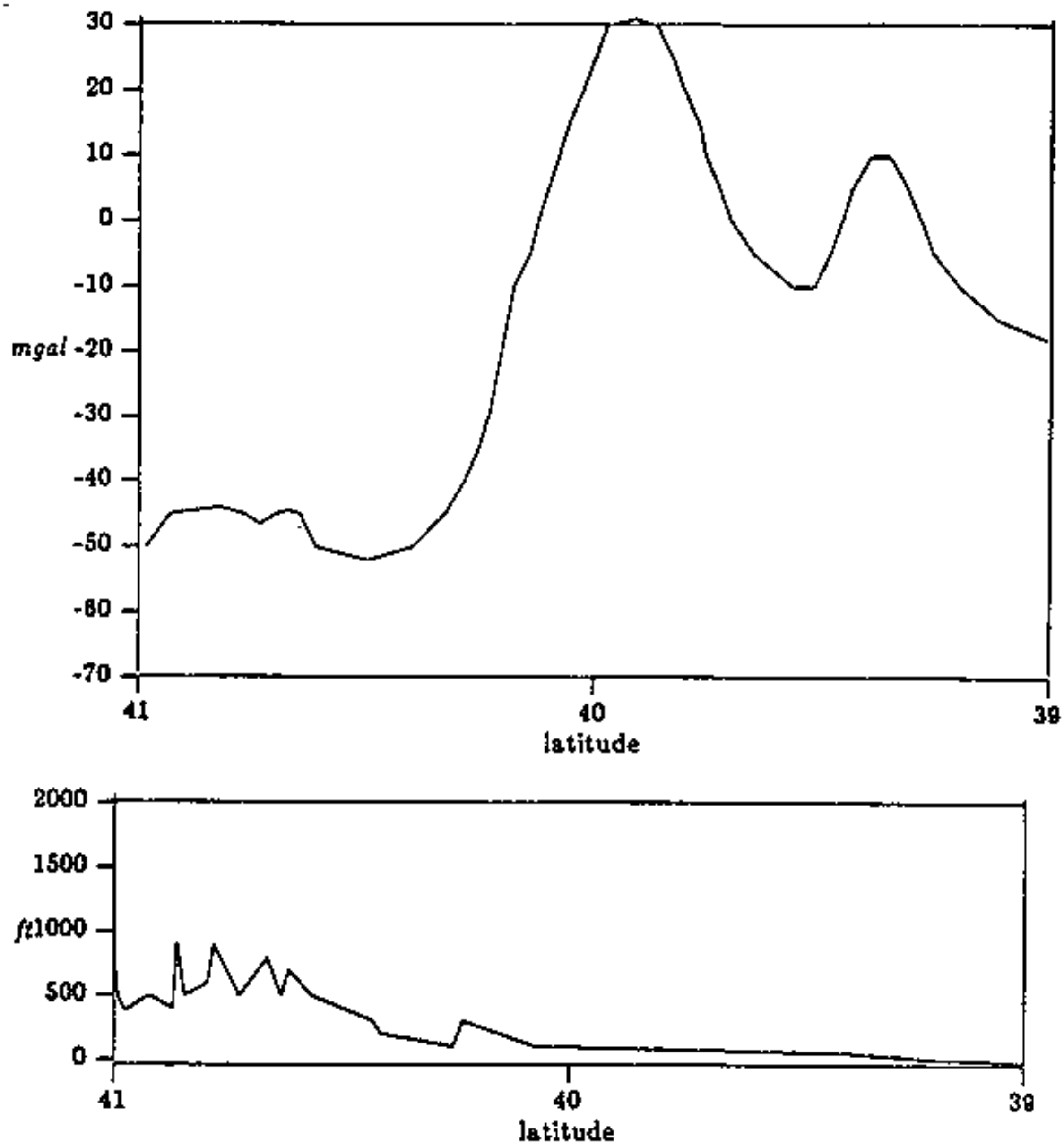


Figure 15b. Profile of topography with gravity profile of 76° W.

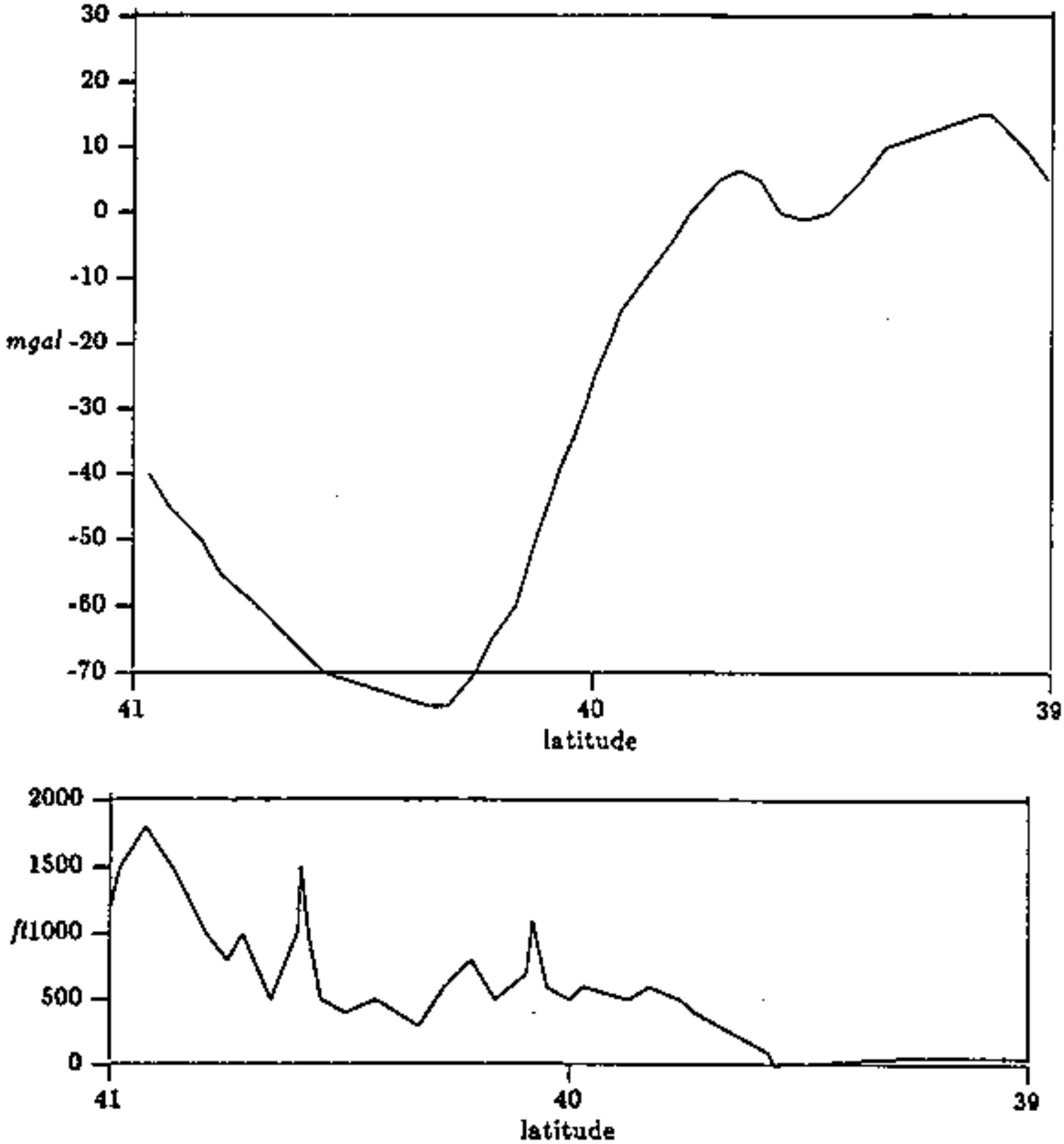
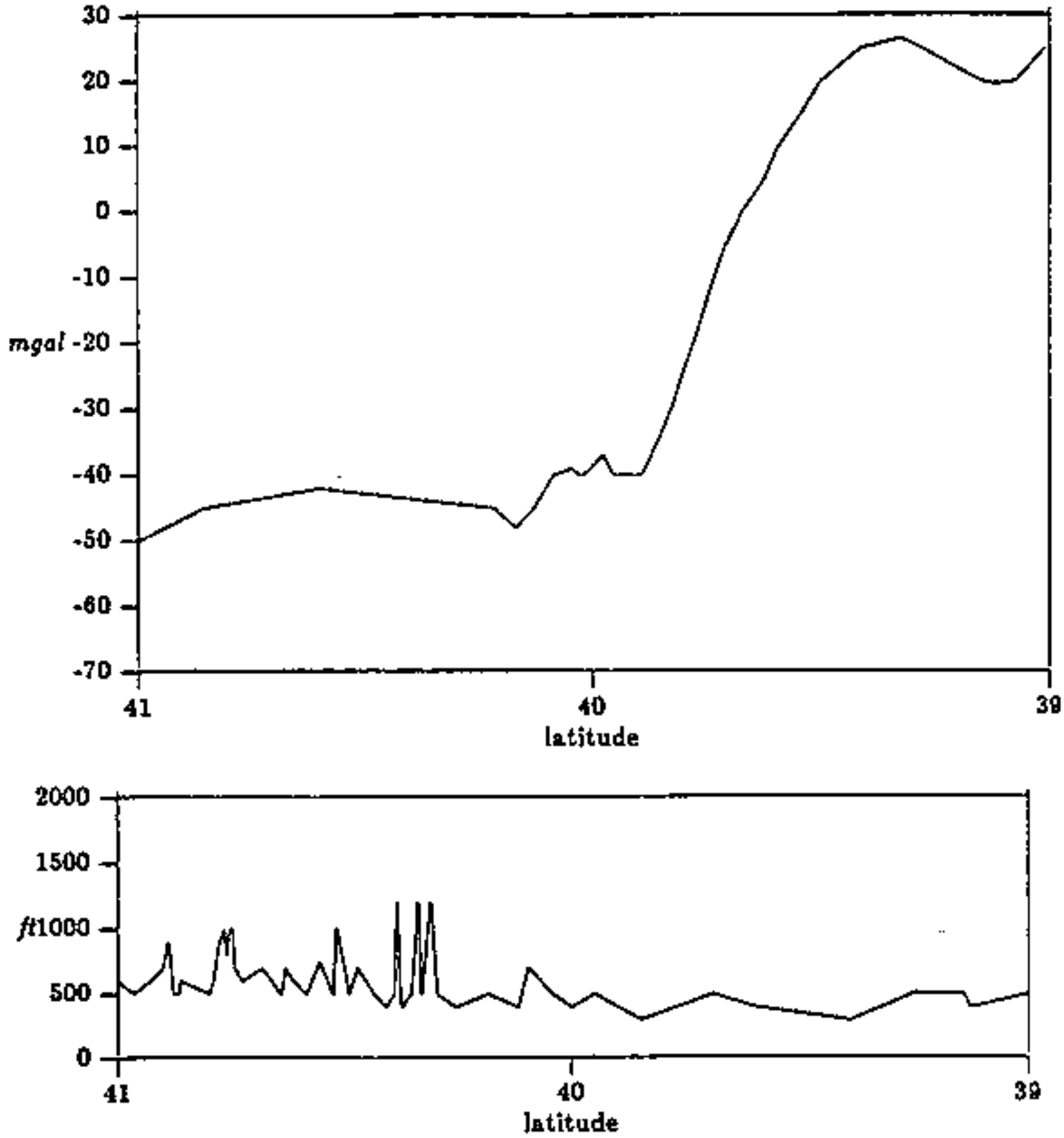


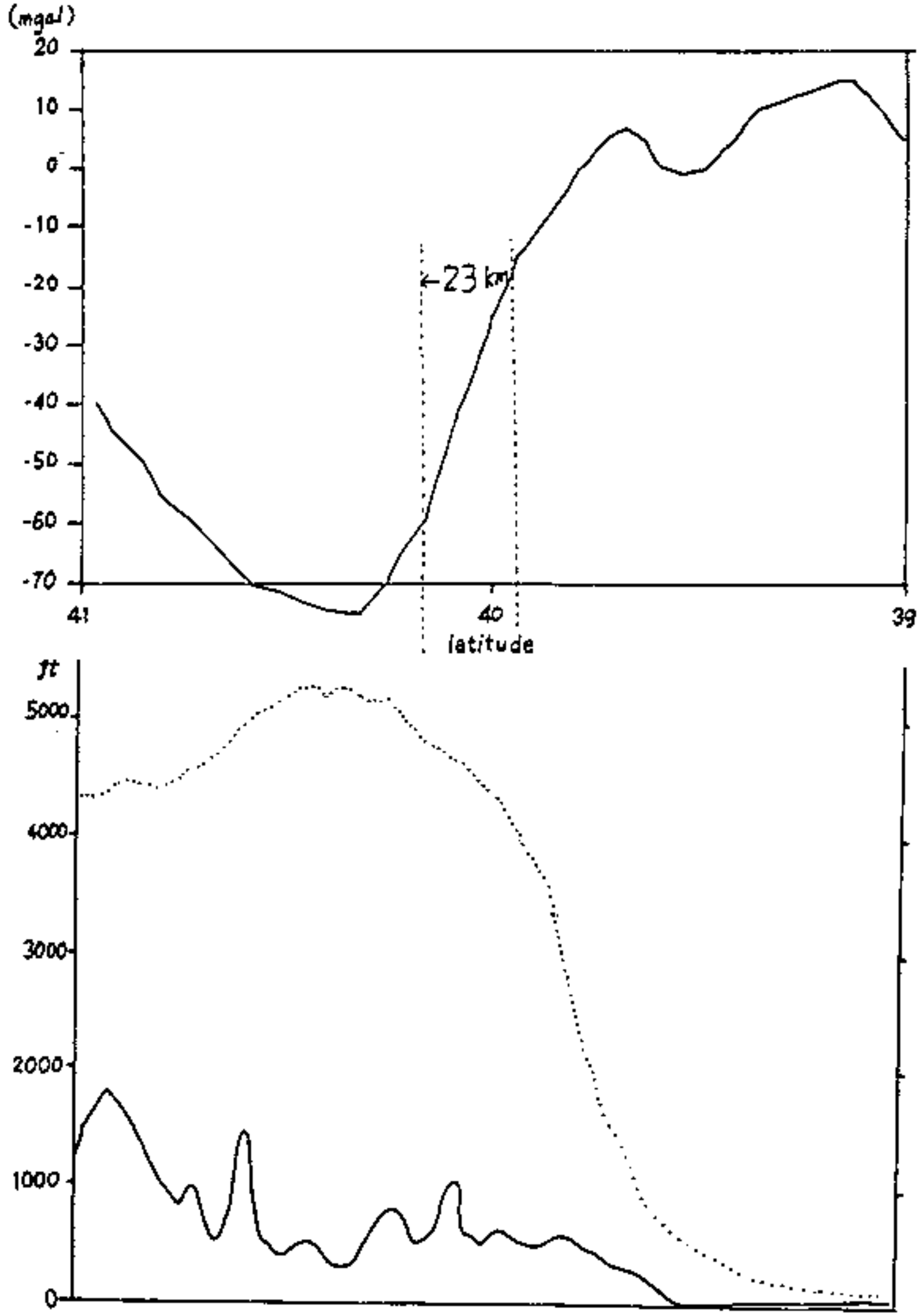
Figure 15c. Profile of topography with gravity profile of 77° W.



gravity in the NW part of the area? From the relation of $g_t = -2\pi G \rho h$, the expected topography is calculated. g_t is the topographic correction and G is the gravitational constant (G=6.67 $\times 10^{-11}$ $m^3/kg \cdot s^2$). An average density of 2.7 $\times 10^3$ kg/m^3 is taken. In order to cause the -70 *mgal* gravity anomaly, the calculated elevation height is 5300 *ft*, which is about 3 times the real topography of today (Figure 16).

Referring the horizontal distance (23 *km*) of gradient belt, the angle of subduction (θ) can be calculated. If the condition that the gradient belt coincides with the crustal thickness contrast zone is assumed, θ is 43° for Sharma's model. A steeper angle of 64° is calculated for the maximum depth equation. Both angles are moderate to steep.

Figure 16. Expected topography. 5300 *ft* elevation is expected to cause -70 *mgal* of gravity anomalies. Solid line is the topography of today. Dotted line is an expected topography.



Chapter 5. Isostacy

5.1 Crustal Thickness Modeling

Both rock density and topographic relief were insufficient to cause the steep gravity gradient. Then what could be the most reasonable cause ? By process of elimination, the observed gravity gradient must be caused by lower crust or mantle structure.

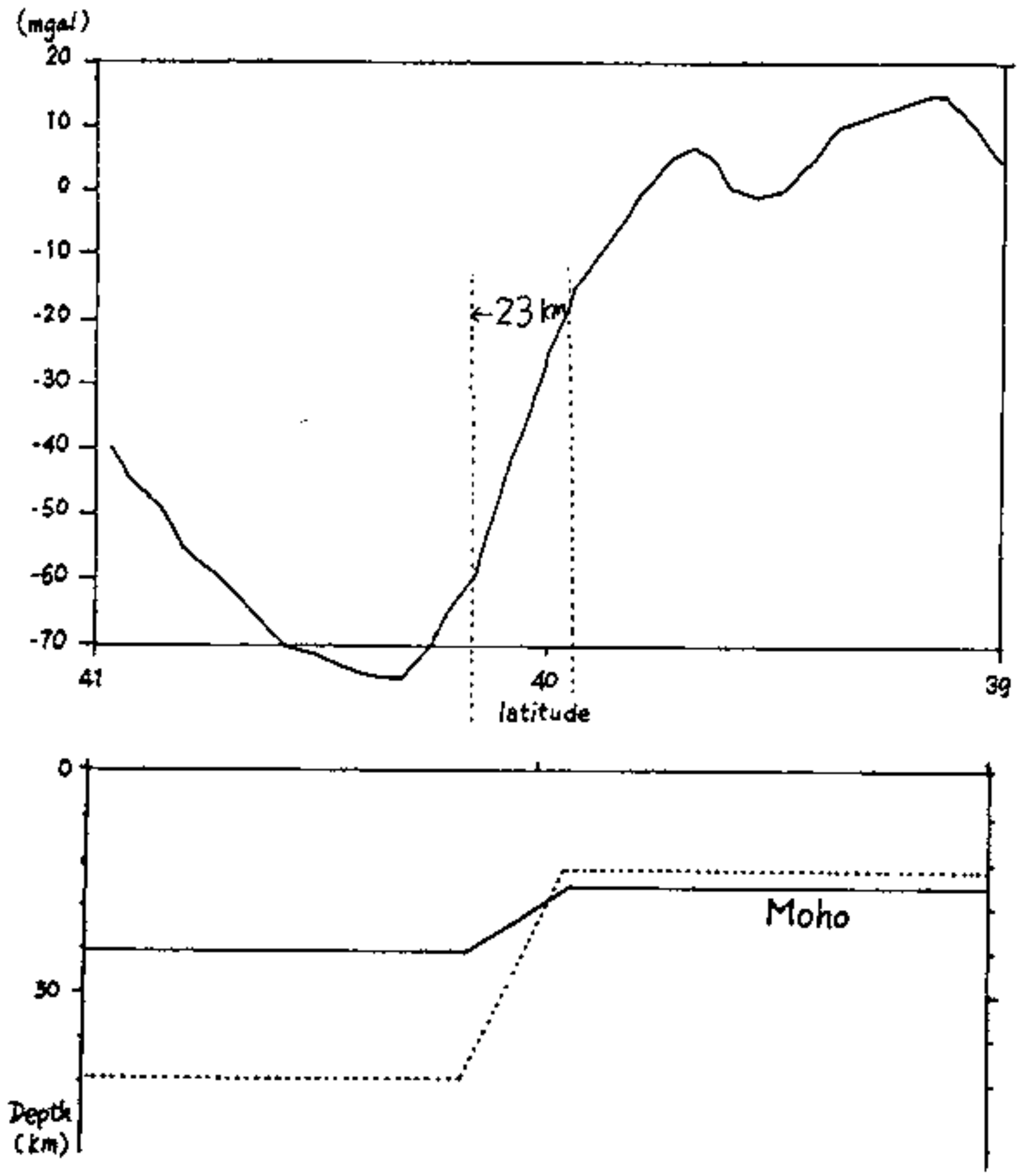
Usually negative values over young orogenic areas reflect thicker crust and/or warmer, less dense upper mantle. Small isostatic anomalies in the crust indicate a close approach to isostatic equilibrium (Condie, 1982).

Following sentences are quoted from Sharma (1976, p. 116):

> *In high areas the Bouguer anomalies are almost always negative, whereas over the ocean area the anomalies are strongly positive. Over land, very near sealevel, the average Bouguer anomaly is close to zero, but for broad areas with large relief these anomalies can reach some hundreds of miligals. The existence of such anomalies can only mean that beneath high areas the rocks are of below-average density and beneath ocean basins they are of above-average density. ...*

Of particular importance is the study of regional variations in gravity and their relationship to visible topography and crustal thickness. The magnitude of the Bouguer anomalies reflects the extent of the compensation in terms of the thickness of the low density crust. Using Sharma's model (1976) crustal thickness can be calculated based on the gravity anomalies (Figure 17). Assuming there are no variations in either density or surface topography, a thickness of 35-40 *km* to the north and 20-25 *km* to the south is calculated. This simplification is justified because density contrast and

Figure 17. Isostacy model. Gravity and crustal thickness model using Sharma's data and maximum depth equation. Dotted line represents maximum depth to Moho and solid line represents Sharma's gravity related crustal thickness. Slope is connected from gravity gradient belt. Average density of $2.7 \times 10^3\ kg/cm^3$ is taken. There is no vertical exaggeration on the crustal model.



topography have been shown to be insufficient to cause the gravity gradient. An abrupt change in crustal thickness is the most likely cause of the steep gravity gradient at the Martic Zone.

The following equation can be used to estimate the maximum depth to the Moho.

$$G_T \leq 0.86 \left| \frac{\Delta g'}{\Delta g} \right| ,$$

where Δg is the regional gravity gradient and Δg is the maximum gravity anomalies. Using this equation, the northern part of the area has a maximum crustal thickness of 68 km, while southern has 22 km. The crustal thickness based on the Sharma's model (35 - 40 km) is less than this calculated maximum depth.

Such a difference in crustal thickness is sufficient to cause the observed gravity gradient.

Chapter 6. Implications

The main objective of this study is to determine the most likely cause for the steep gravity gradient in the vicinity of the Martic Zone. Rock density contrasts or topographic effects are insufficient to explain the gradient. Not only the Martic Zone but the Alpine and San Andreas Faults clearly show a gravity contour parallelism separating negative-positive anomalies. If we accept that the origin of the steep gravity gradient in the vicinity of the Martic Zone is an abrupt change in crustal thickness, as it is for the other two transform fault areas, then we have to answer what that means.

6.1 Suspect Terrane Theory

The suspect terrane theory is summarized by Coney *et al.* (1980). In short, a suspect terrane is a microplate which survives above the subduction zone due to its buoyancy. This piece of allochthonous plate is accreted to the autothonous continent or to other preaccreted allochthonous microplates. Suspect terranes are characterized by internal homogeneity and similarity in stratigraphy, tectonic style, and history (Coney *et al.*, 1980). Discontinuities in stratigraphy at terrane boundaries in the North American Cordillera could not be explained by facies changes or unconformities. Some terranes contain paleomagnetic and faunal records that totally differ from those of cratonal North America, which indicates that large displacements occurred between the terranes themselves or between terranes and stable North America (Coney *et al.*, 1980).

These allochthonous terranes may have originated in the oceans as oceanic plateaus. These oceanic plateaus are anomalously high parts of the sea floor including volcanic arcs or spreading ridges (Ben-Avraham and Nur, 1983). Some terranes originate as continental fragments, separated by rifting or strike-slip displacement.

Boundaries between terranes are faults or shear zones (Condie, 1982). Terranes are suspect when they are of unknown paleogeography with respect to nearby miogeoclines and stable cratons. Their time of accretion can be gleaned from the age of similar cover rocks, the depositional age of detritus shed from one upon another, or similar postdocking intrusive, metamorphic, structural, paleontologic, or paleomagnetic histories (Coney *et al.*, 1980).

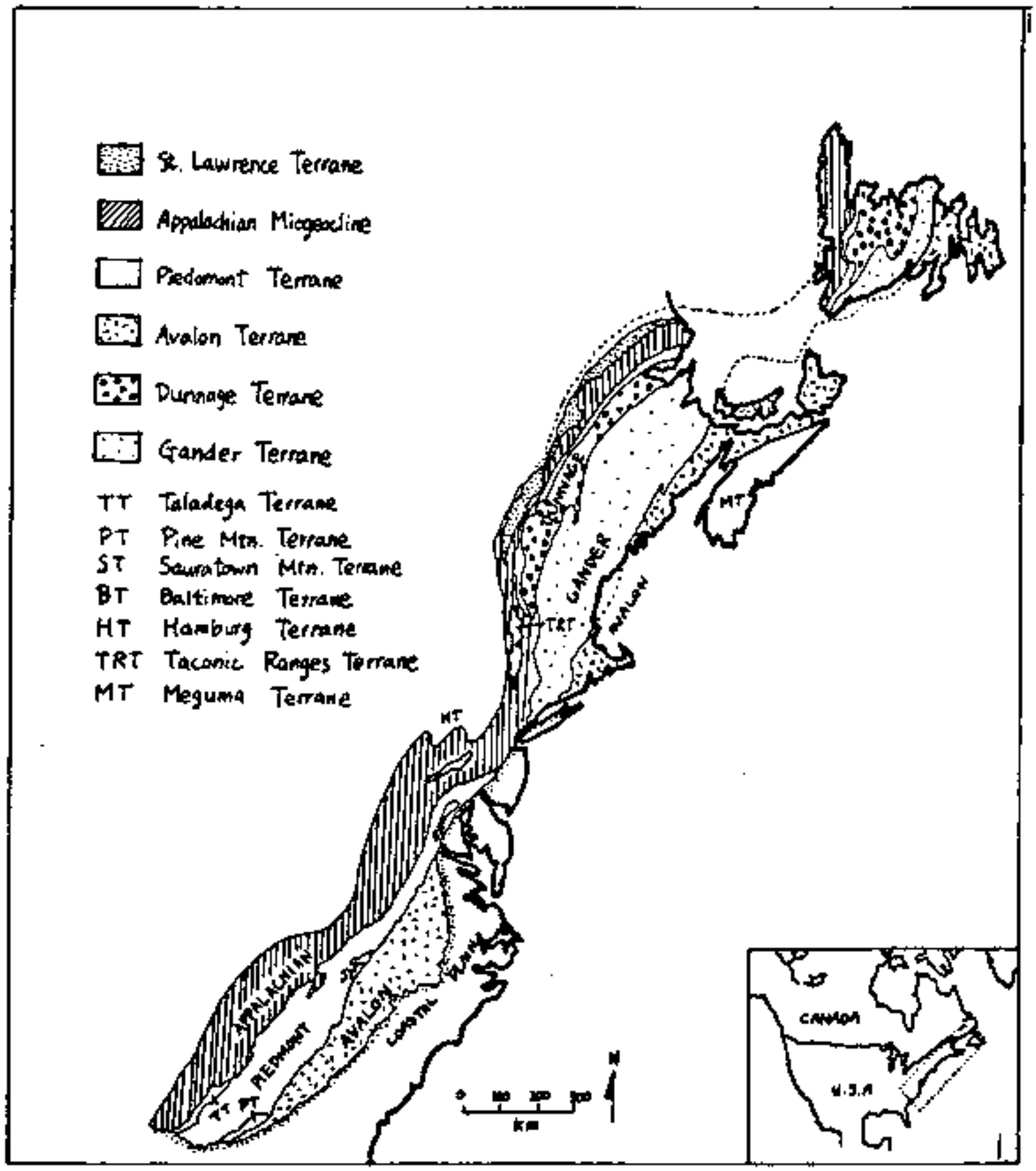
Geological, paleontological and paleomagnetic data from western North America indicate that much of the Cordilleran orogenic belt is a mosaic of foreign terranes (Condie, 1982). Juxtaposition of units of various age, stratigraphy, tectonic style, and fossils in the Cordilleran orogenic belt makes geologic interpretation challenging.

Williams and Hatcher (1983) using the suspect terrane theory tried to explain the significance of ophiolitic vestiges of Iapetus along the North American margin. They have to assume that the terranes now east of the North American miogeocline were not always connected to the North American continent (Williams and Hatcher, 1983). Therefore, the orogen is viewed as a mosaic of accreted terranes, all of unknown paleogeography. Suspect terranes of the Appalachian orogen are named and outlined in Figure 18 (Williams and Hatcher, 1983).

Recent studies (Ben-Avraham, 1983; Coney *et al.*, 1980) have shown the Cordilleran orogenic belt from California to Alaska to be composite, consisting of smaller units of crust, each having its own stratigraphic and tectonic integrity but not related in obvious ways to those next to them.

In a very detailed study of the New England area (Zen, 1983) also developed the idea of suspect terranes which he named "exotic terranes".

Figure 18. The Appalachian suspect terranes (after Williams and Hatcher 1983)



He separately examined the three Paleozoic orogenies (Taconian, Acadian, and Alleghenian) and concluded that there are three types of accreted terranes: (1) Taconian, thrust allochthons directly attributable to subduction-induced collision during the closing of Iapetus Ocean, (2) Acadian, continent-continent collision and possible large concomitant transcurrent displacement, (3) Alleghenian, oblique-slip high angle thrusting, the concomitant formation of a sedimentary basin having no immediately identifiable sediment source, and formation of a microplate collage (Zen, 1983).

Some geologists (*e.g.* Hill, personal communication, 1987) think that almost all orogenic belts are made of accretional terranes. Facing the limits of explanation for the significant topographic relief and the contrast in geologic history, the idea of allochthonous terrane accretion is accepted.

The Martic Zone separates rock of very different lithology and metamorphic history. The juxtaposition of these very different rocks may be explained by the theory of suspect terranes.

6.2 Shear Zone Geometry

Since the boundaries of suspect terranes are thought to be either shear zones or faults, it is important to consider the nature of the shear zone known as the Martic Zone. Shear zones are defined by Ramsay and Huber (1987) as zones of high deformation which are long relative to their width and which are surrounded by rocks showing a lower state of finite strain. He subdivided shear zones into ductile shear zones, brittle shear zones, and brittle-ductile shear zones. The deformation state in ductile shear zone varies continuously wall to wall through the zone, whereas the walls are separated by a discontinuity or fracture surface in brittle shear zones (Ramsay and Huber, 1987). The deformation geometry of ductile shear zones takes several forms. The zone may be parallel sided and the strain profiles determined from strain markers may remain almost constant in different transects through the zone. In constant, shear zones can show walls which converge and diverge and which show deformation profiles at differing localities along the zone which are not the same (Ramsay and Huber, 1987).

When the boundary of a shear zone can be observed, the following features can be used to define shear sense: 1. change in finite strain state, 2. sigmoidal form of schistosity, 3. the orientations of folded and boudinaged component layers, 4. the geometry of en-echelon extension vein array, and 5. fracture opening. However, when the shear zone is very wide or the boundaries obscure, 1. s-c band structure, 2. rotation of porphyroblast, 3. σ-structure, 4. δ-structure, 5. bookshelf sliding, 6. displaced crystals, 7. dynamic recrystallization, and 8. preferred optical orientation can be used to determine the shear sense (Ramsay and Huber, 1987).

The sense of shear can be used to determine the direction of plate movement. In ductile shear zones, simple shearing leads to the progressive approach of the XY plane of the strain ellipsoid towards the shear plane and progressive approach of X, the extension axis, to the shear line. Therefore, large extensions are approximately parallel to the shear line. There is a suggestion (Shackleton and Ries, 1984) that the extension axis X (stretching lineations), rather than the fold axis, is of prime significance in interpreting tectonic features. By tracing the stretching lineations in a plate margin scale shear zone, the direction of the plate motion can be inferred. Shackleton and Ries (1984) found that the regional stretching lineations in the Himalayas and southern Tibet coincide with the known relative motions of the convergent plate, as do those in the Variscan and Caledonian belts of Europe.

6.3 The Martic Zone as A Terrane Boundary

The most likely cause of the steep gravity gradient along the Martic Zone has been considered. Let us remember that the Martic Zone is a shear zone. The suspect terrane theory claims that boundaries between terranes are faults or shear zones with obvious dissimilarity in geologic phenomena across them.

Many geologists working in the central Piedmont area have noticed the difference in rock type, metamorphic history, and stratigraphy across the Martic Zone (Crawford and Crawford, 1980; Hill, 1987). In addition, an abrupt change in crustal thickness, petrofabric analysis, and strong parallelism of positive-negative Bouguer gravity anomalies leads us to say that the Martic Zone might be a terrane boundary.

Williams and Hatcher (1983) mentioned the Martic Line as the possible western boundary of the Piedmont terrane. The following quotation is after William and Hatcher's description of Appalachian suspect terranes (p. 38):

> *Piedmont terrain assemblages are less distinct in the Central Appalachians. The western boundary here is shown as the Martic Line, but it may be that this terrain is discontinuous and consists of a series of disconnected klippen resting upon the miogeocline east of the Martic Line. .... The Martic Line is, however, a significant lithotectonic boundary with contrasting assemblages on either side.*

The interpretation of the Martic Zone as the boundary of the Piedmont terrane is supported by the results of this thesis study. The most likely cause of the steep gravity gradient at the Martic Zone is an abrupt change in crustal thickness. The coincidence of this major crustal boundary with a large scale shear zone indicates that the Martic Zone is likely to be a terrane boundary.

Chapter 7. Comparisons

In this chapter analogues of other areas with gravity parallelism will be compared. The San Andreas Fault, California and the Alpine Fault, New Zealand are chosen because these faults are known to be transform faults, plate boundaries.

Gravity anomalies, rock density of surface lithology and topographic relief of the Martic Zone have been presented. Any analogues which indicate similarity to the Martic Zone are considered to see what they imply.

7.1 The San Andreas Fault

San Andreas Fault has numerous branches which generally show NW-SE strikes. Hanna, Burch and Dibblee (1972) described the regional gravity gradient at the San Andreas Fault as about 2 *mgal* per mile, decreasing toward the northeast. The northeastward decrease of the regional field may be caused by a northeastward thickening of the continental crust or be coupled with a northeastward decrease in average density of the lower continental crust. Their study on the San Andreas Fault considers the relationship of Bouguer gravity data, total intensity aeromagnetic data, and vertical intensity ground magnetic data. Comparing the areas NE and SW of the San Andreas Fault, they carefully seek the cause for gravity variation. Their study is restricted to a $0.5° \times 0.5°$ area. They concluded that the gravity features are quite dissimilar on opposite sides of the faults. The gravity fields of the Cape San Martin map are shown in Figure 19.

In an interesting gravity study on the San Andreas fault, economic geologist Noble (1970) mapped the available data for mining. His Bouguer gravity map and crustal thickness contour map are used here (Figure 20).

Figure 19. Local gravity fields of the Cape San Martio area. Darker solid lines are the branches of the San Andreas Fault. Gravity contour intervals are 5 *mgal*.

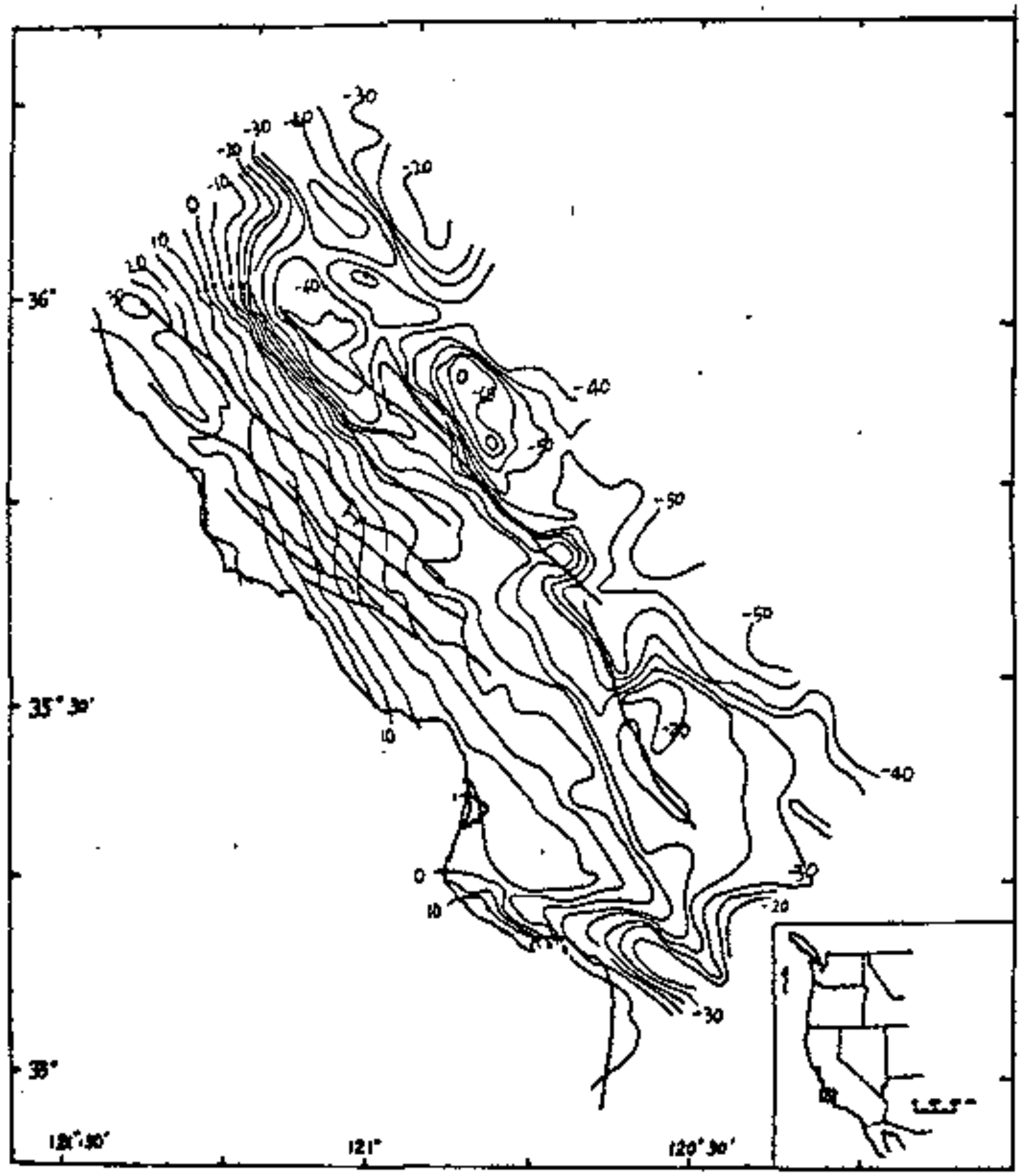
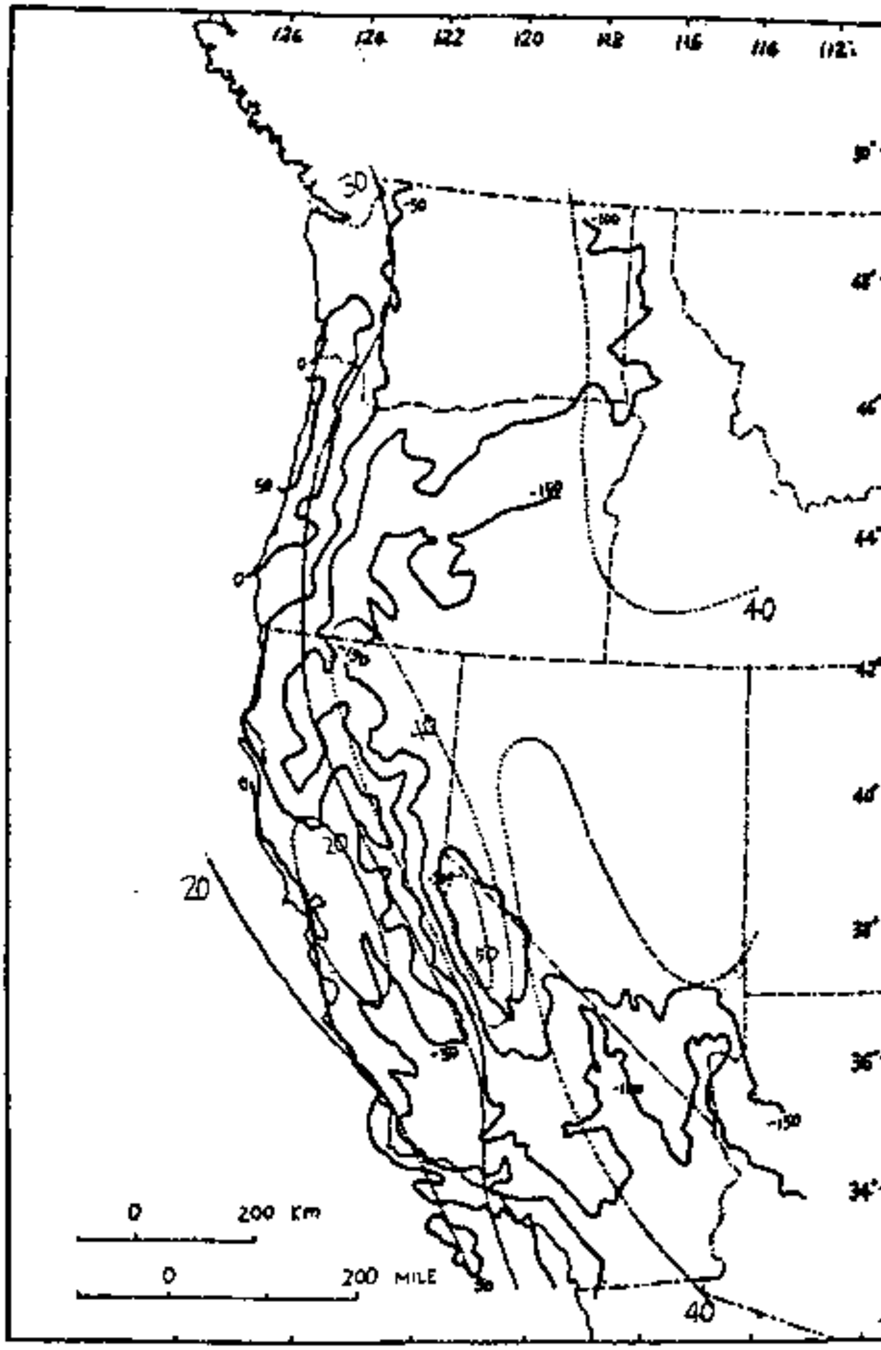


Figure 20. Regional gravity fields of the San Andreas Fault area (Bouguer gravity map of the Northeastern United States and adjacent Canada, Bothner *et al.*, 1980) Dotted lines represent the crustal thickness in *km* unit. Solid lines represent the Bouguer gravity anomalies in *mgal* unit. Good correlation of negative gravity anomaly and the deepest thickness of crust is shown.



The Bouguer gravity anomalies rapidly decrease from 50 *mgal* offshore to -200 *mgal* near the San Andreas Fault.Noble made a crustal depth map for the western coast of North America. Overlapping the Bouguer map with a depth-to-Moho map, Figure 20 shows the gradual change in depth to the Moho in close relation to the Bouguer gravity anomalies.

The San Andreas Fault shows a distinctive gravity anomaly parallelism, crustal thickness change across it, and is a transform plate margin.

7.2 The Alpine Fault

The Pacific - Australian plate boundary, marked by the Alpine Fault in New Zealand has been studied to show the relative motion of the Australian plate during the last 10 m.y. (Allis, 1986). A Bouguer gravity map of the Alpine fault shows a strong positive-negative pair of gravity anomalies (Figure 21). At the bottom left of map, a negative center of -200 *mgal* is located adjacent to a positive gravity anomaly center of 175 *mgal*, showing a regional gradient of 4.5 *mgal/mile* (Figure 21). Two profiles of gravity and crustal thickness are illustrated in Figure 22. Both profiles clearly show an abrupt change in the Bouguer gravity field and crustal thickness. These profiles are drawn for the purpose of Allis' study which concludes that crustal shortening has occurred by the plate movement. The locations of the two profiles are far north from the negative-positive gravity anomaly pair. However, the profiles still show a steep gradient with 10 *km* of crustal thickness change. If a southern profile is drawn, steeper gravity gradient and greater crustal thickness change would be expected. Allis' study shows a good correlation between the gravity, crustal thickness, and plate boundary features.

Figure 21. Bouguer gravity map of the Alpine Fault. The contour intervals are 25 *mgal*. Solid lines represent the Bouguer gravity contours. Dark dotted lines represent the Alpine Fault.

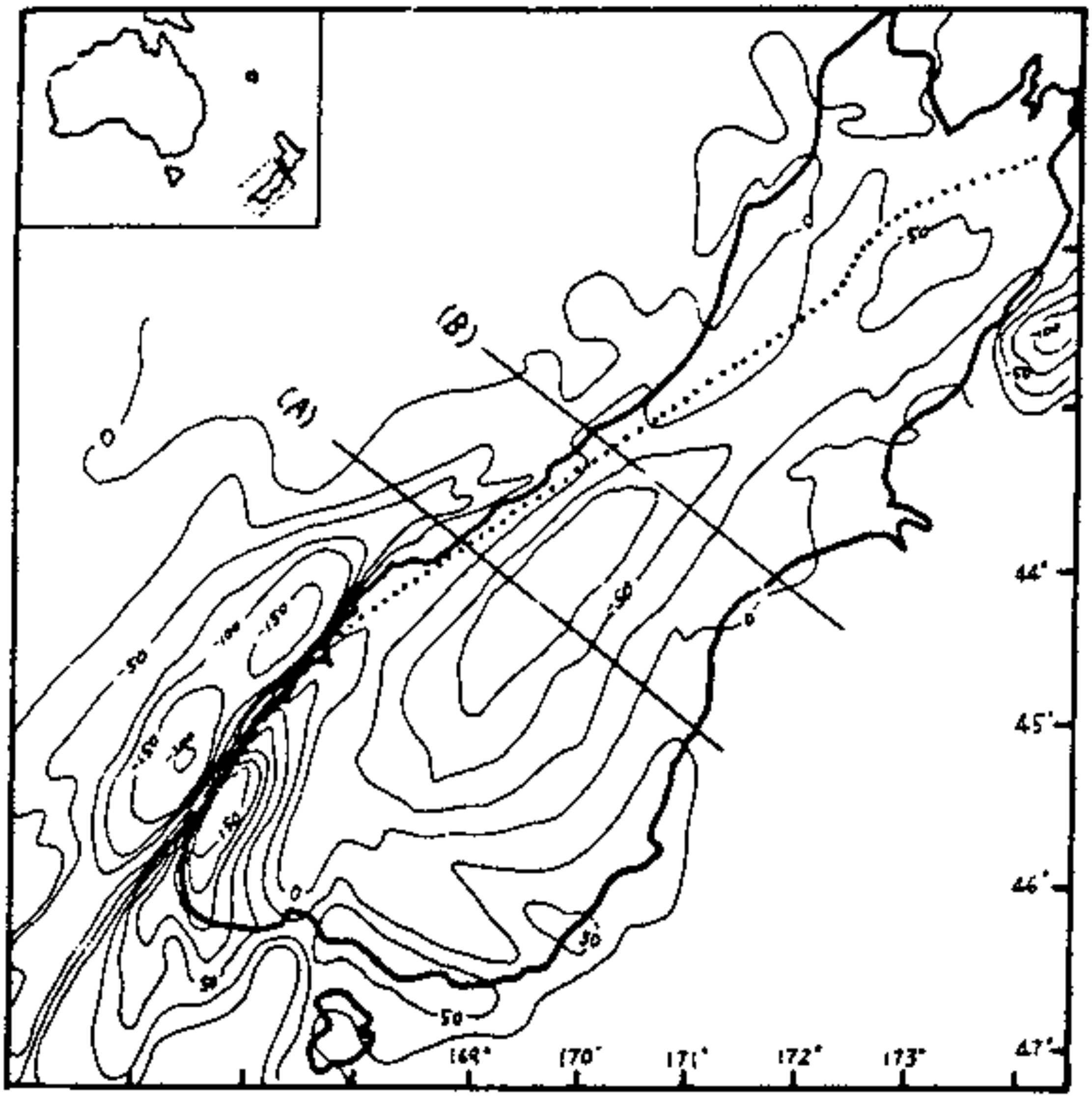
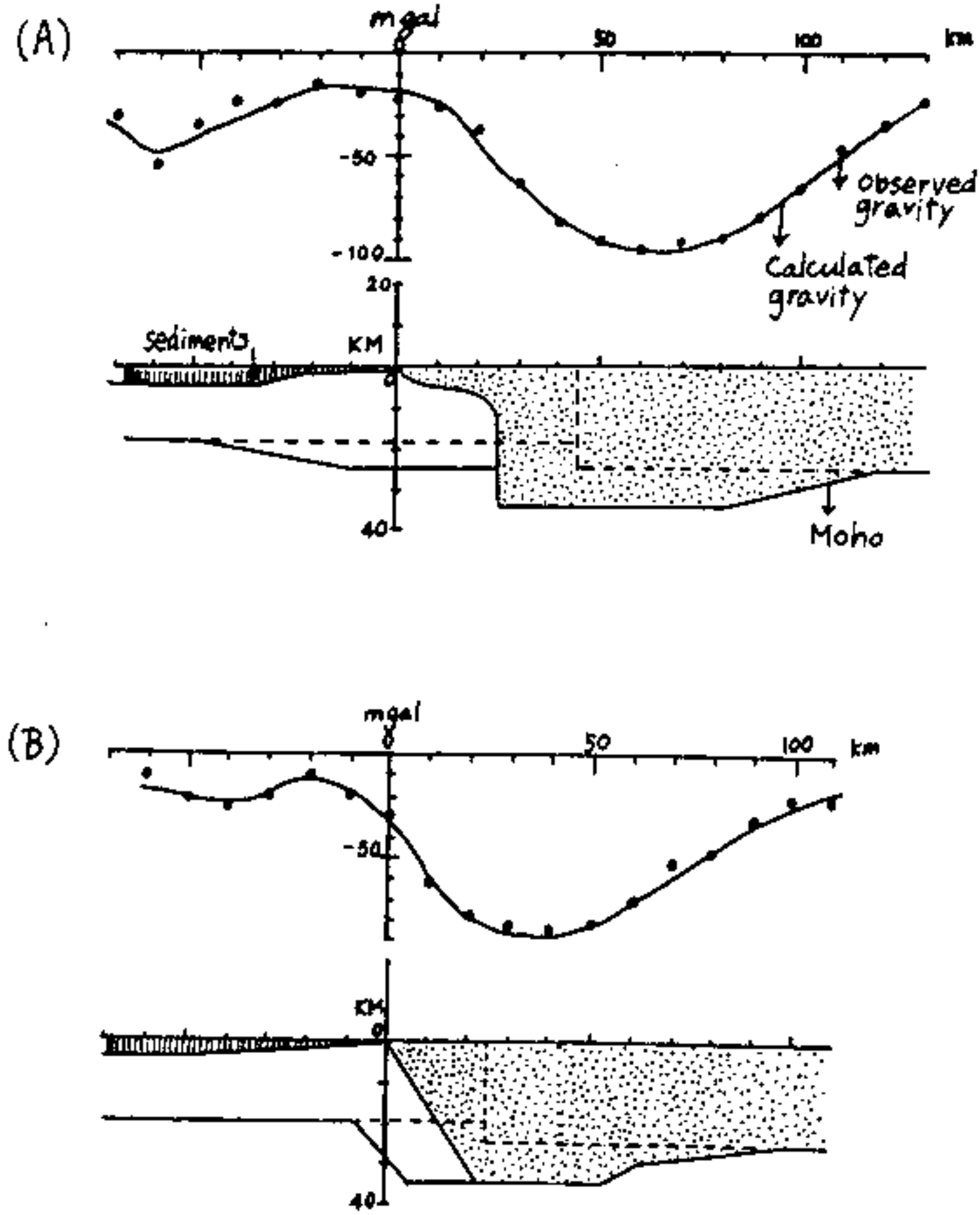


Figure 22. Allis' model of crustal thickness. (A) and (B) are cross-sections of NW to SE line. Localities of (A) and (B) are shown in Figure 21. Notice the abrupt change of crustal thickness. The Alpine fault coincides with the zero line on the profile. Stippled area denotes the Pacific Plate Crust. The dashed lines indicate the Moho and plate boundary prior to the onset of the late Cenozoic shortening.



7.3 Comparison of the Martic Zone to Other Faults

Both the San Andreas Fault in California and the Alpine Fault in New Zealand show distinct gravity parallelism of positive-negative anomaly parallel to the fault-strike direction. The San Andreas Fault shows an anomaly range from 50 *mgal* to -200 *mgal* across it. The offshore data of Free-Air gravity could give more wide range of gravity anomalies. The Alpine Fault has a range from -200 *mgal* to 175 *mgal* including offshore Free-Air gravity data. the Martic Zone shows from -70 *mgal* to 30 *mgal* gradient range. While the close spacing gravity contour belts of both the San Andreas and the Alpine Faults coincide with coast lines, the Martic Zone has the belt on land.

Regional gravity gradient of the San Andreas Fault is 2 *mgal/mile*, and that of the Alpine Fault is 4.5 *mgal/mile*. The regional gravity gradient of the Martic Zone is 2.7 *mgal/mile*. This number is greater than that of the San Andreas Fault. It is because the Martic Zone is very narrow belt with wide range of gravity anomaly. Even though the San Andreas Fault has wider range, fault zone spreads out. So relatively regional gradient will be lower.

There is a study on comparison of these two major active transverse faults (Yeats and Berryman, 1987). Both the South Island of New Zealand and the Transverse Range s of California have a major right-slip fault marking the plate boundary.

The cross-section of the Alpine Fault shows sharp 9 *km* change in the Moho depth at around 25 *km* east of the Alpine Fault, marking $45°$ dip, The Martic Zone has $43° - 63°$ dip of crustal boundary.

Chapter 8. Conclusions

The following are the conclusions of this study regarding the geophysical and geologic features of the Martic Zone.

1. A steep gradient in Bouguer gravity anomaly contours coincides with the Martic Zone.
2. The density of surface lithologies and the mass distribution associated with the topography are insufficient to cause the steep gravity gradient across the Martic Zone.
3. The surface geology and topography are not related to the geophysical feature. Somewhat deeper structure must account for the gravity gradient.
4. An abrupt contrast in crustal thickness (15-20 km) across the Martic Zone is the most likely cause of the steep gravity gradient.
5. The Martic Zone shows shear zone geometry, steep gravity gradient, and crustal thickness contrast like other plate margins such as the San Andreas Fault and the Alpine Fault.
6. Being a significant feature of the Central Appalachian orogeny, the Martic Zone would be interpreted as the western boundary of the Piedmont terrane.
7. Implication of being a terrane boundary may solve many aspects of the regional tectonic problem.

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