Geology of the Lucea Inlier,
Western Jamaica

Abstract of
a thesis presented to the Faculty
of the State University of New York
at Albany
in partial fulfillment of the requirements
for the degree of
Master of Science

School of Arts & Sciences
Department of Geological Sciences

Jack Grippi

ABSTRACT

The Lucea Inlier exposes a Santonian to Campanian 4 km + thick sequence of shale-siltstone, resedimented volcaniclastics, lenses of shallow-water limestone, micritic limestone, pebbly mudstone and sandy pebble to boulder conglomerate. Clastics were deposited by a variety of gravity flow mechanisms. Petrographically sandstones are lithic or feldspathic arenites and contain only very small amounts of detrital quartz. Structurally the inlier is characterized by simple, open, east-west trending folds. A spaced, vertical axial-planar cleavage is developed in shales and fine siltstones. Two major east-west trending left-lateral fault zones, the Fat Hog Quarter and Maryland faults, cut the inlier into three blocks, northern, central and southern. The basal part of the sequence has been subjected to a prehnite-pumpellyite metamorphism.

The rocks of the Lucea Inlier are interpreted to represent a shelf to basin sequence within an upper slope basin of a Cretaceous intra-oceanic arc trench system. Detritus shed from the arc was funneled down submarine canyons feeding a submarine fan complex. Between canyon heads, shoal areas fringing volcanic islands locally accumulated bio-clastic, reef-type limestone.

The geology of the northern Caribbean plate boundary records a complex array of Cretaceous to Eocene arc-trench systems that ha~~ve~~s been modified by Cenozoic left-lateral slip along the Oriente and Swan transforms.

1 4450

Ridge related north-south lineated topography of the Cayman Trough suggests that a minimum of 720 km of left-lateral movement has occurred between the North American and Caribbean plate since approximately Oligocene times. Presently active northwest, northeast and east-west trending structures within Jamaica are interpreted as being of compressional, extensional and strike-slip origin, respectively, and are thought to be related to Recent left-lateral slip along the northern Caribbean plate boundary.

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Acknowledgments

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(located in back pocket)

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Note to the Reader

For names of geographic locations refer to Plate 1. Plate 2 is a detailed geologic map of the east central part of the inlier and lacks geographic names.

Roads are informally named throughout the text. A road is named after the villages at each end of the length of road in question, e.g., Cascade-Patty Hill Road. For the location of a specific locality the Jamaican grid reference is used, and this appears on the maps. It is used liberally throughout the text in hope that it will be helpful to geologists using this thesis in the field.

All descriptions of lithologic units within the text pertain only to the area mapped by this study, unless otherwise specified.

CHAPTER I

INTRODUCTION

Geologic Setting

The island of Jamaica is an emergent segment of the Nicaraguan Rise and is considered part of the Greater Antilles (Fig. 1). Rocks in Jamaica record Cretaceous convergent processes at the northern margin of the Caribbean plate (Fig. 2). Volcanic arc, underlying granodiorite-tonalite pluton, upper slope basin and subduction complex (with blueschist and serpentinite) environments are all represented in the Cretaceous (Barremian(?)-Maastrichtian) inliers of the island. From Paleocene to Mid-Eocene times, Jamaican geology was dominated by the development of northwest-southeast trending troughs (Wagwater and Montpelier-Newmarket troughs). From Mid Eocene to Mid Miocene Jamaica and the Nicaraguan Rise were subsiding shelf regions that accumulated shallow water limestones. Since Mid Miocene times Jamaica has been tectonically active due to its impingment with the Oriente Transform (see Fig. 2).

The Lucea Inlier is one of 28 Cretaceous inliers on the island of Jamaica (Fig. 3). It is one of the larger inliers and has an approximate outcrop area of 90 sq. km. An approximately 4 km + thick sequence, consisting, in order of abundance, siltstone and shale, volcanic rich sandstone, conglomerate, lenticular bodies of reefoidal limestone and minor breccia make up most of the section and represent a shelf to basin sequence. Although minor in volume, the following rock types also occur: mafic sills, micritic limestones and *in situ*

Figure 1 - Index map to major geographic and geologic features of the Caribbean region.

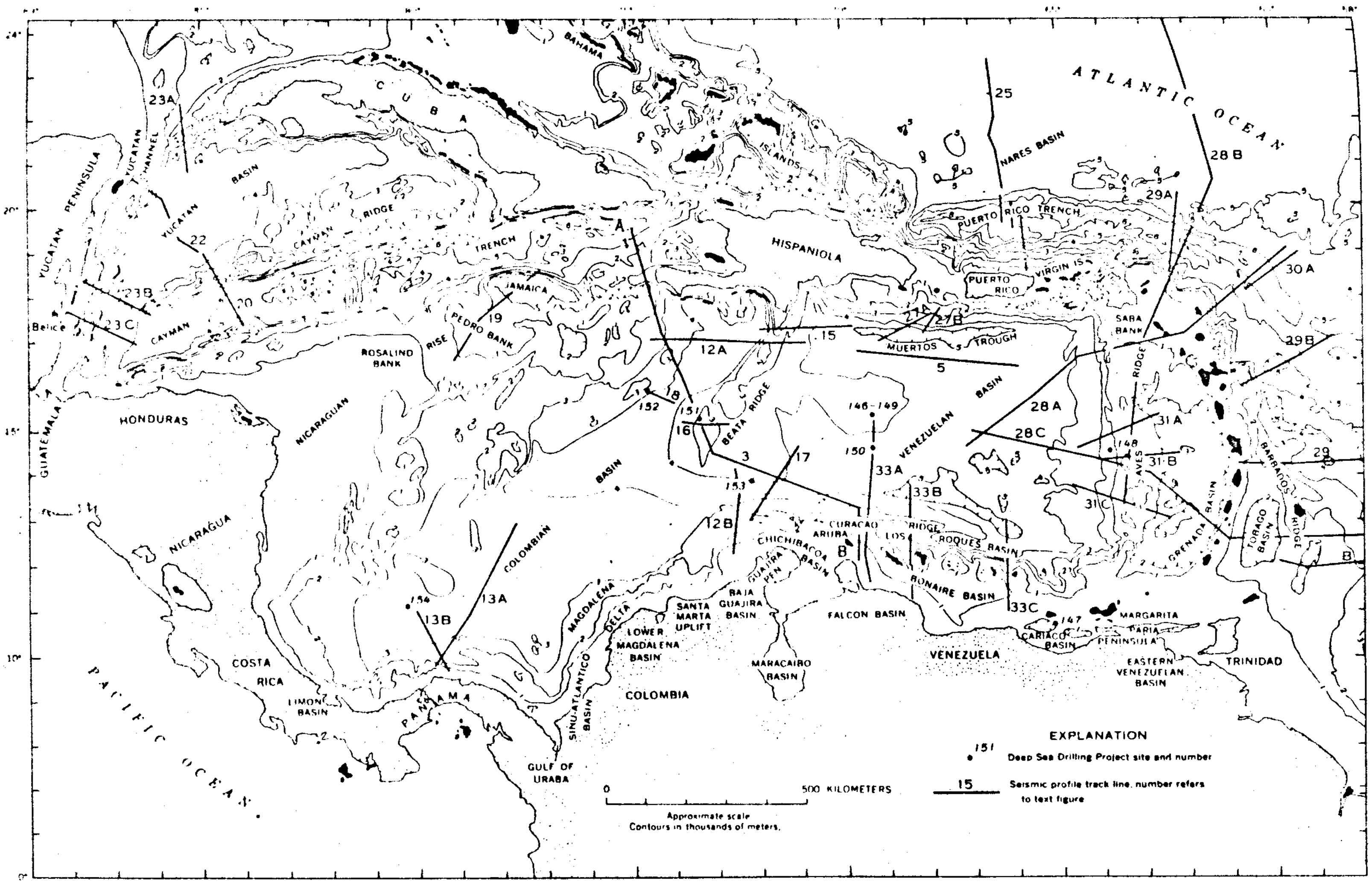

ATLANTIC OCEAN
PACIFIC OCEAN
CUBA
BAHAMA
ISLANDS
HISPANIOLA
PUERTO RICO
VIRGIN IS
PUERTO RICO TRENCH
NARES BASIN
JAMAICA
PEDRO BANK
RISE
ROSALIND BANK
CAYMAN RIDGE
CAYMAN TRENCH
CAYMAN
YUCATAN BASIN
YUCATAN CHANNEL
YUCATAN PENINSULA
Belice
GUATEMALA
HONDURAS
NICARAGUA
NICARAGUAN
COSTA RICA
LIMON BASIN
PANAMA
COLOMBIAN BASIN
GULF OF URABA
SINU-ATLANTICO BASIN
MAGDALENA DELTA
LOWER MAGDALENA BASIN
SANTA MARTA UPLIFT
COLOMBIA
BAJA GUAJIRA BASIN
GUAJIRA PEN
MARACAIBO BASIN
CHICHIRIBACOA BASIN
ARUBA
CURACAO
LOS ROQUES BASIN
RIDGE
BONAIRE BASIN
FALCON BASIN
VENEZUELA
VENEZUELAN BASIN
BEATA RIDGE
MUERTOS TROUGH
SABA BANK
AVES RIDGE
GRENADA BASIN
TOBAGO BASIN
BARBADOS RIDGE
MARGARITA
PARIA PENINSULA
CARIACO BASIN
EASTERN VENEZUELAN BASIN
TRINIDAD
23A
22
23B
23C
20
19
A
12A
15
16
18
151
152
153
3
17
12B
13A
13B
154
146-149
150
33A
33B
33C
147
5
27A
27B
25
28A
28B
28C
29
29A
29B
30A
31A
31B
31C
148
B
EXPLANATION
151 Deep Sea Drilling Project site and number
15 Seismic profile track line, number refers to text figure
0 500 KILOMETERS
Approximate scale
Contours in thousands of meters.

Figure 2 - Present-day plate boundaries and major structural features of the Caribbean region. Bold arrows indicate the relative motion of neighboring plates with respect to the Caribbean. Slightly modified after Jordan (1975)

- offshore regions experiencing compression

- subduction zones

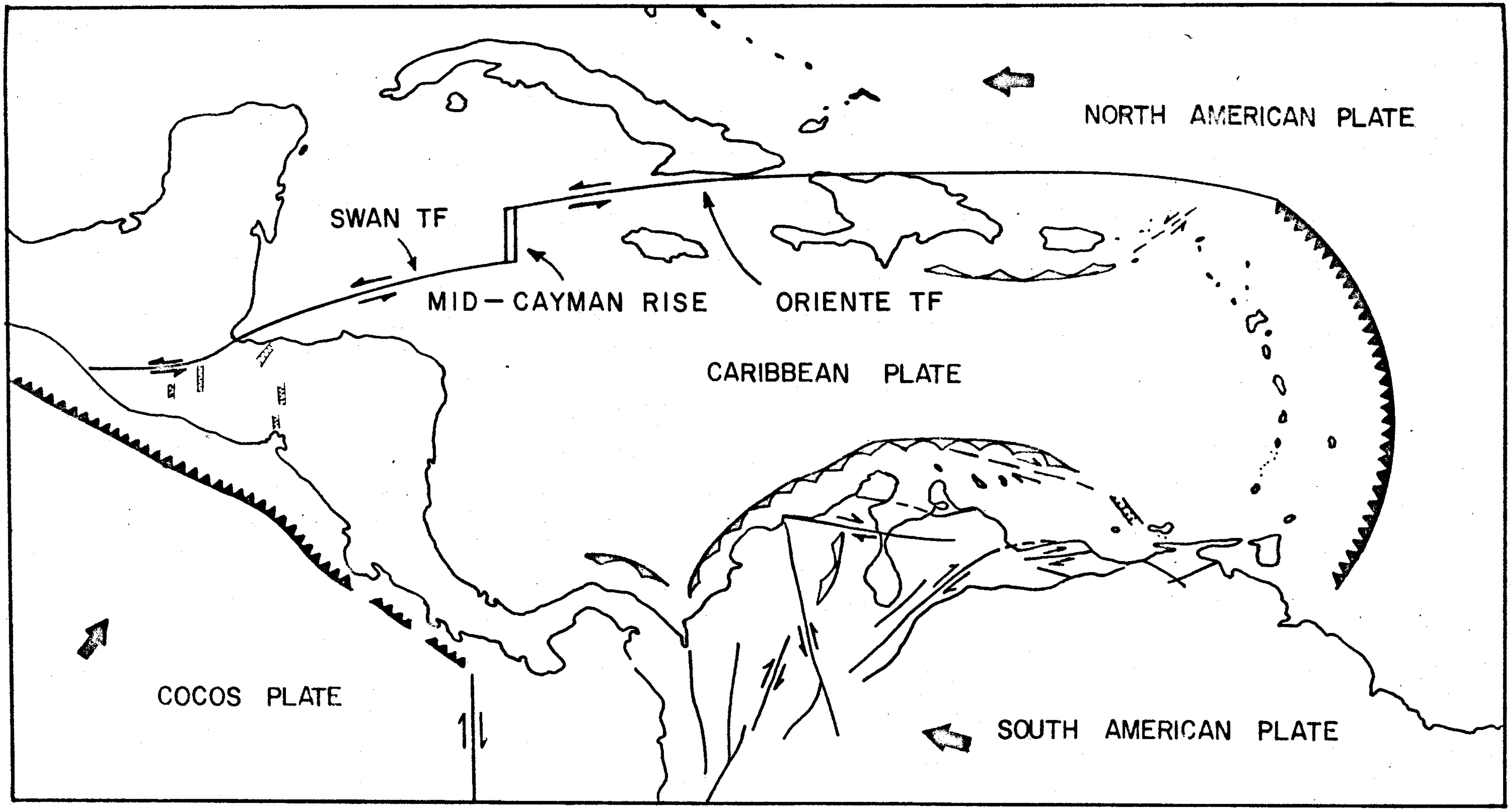
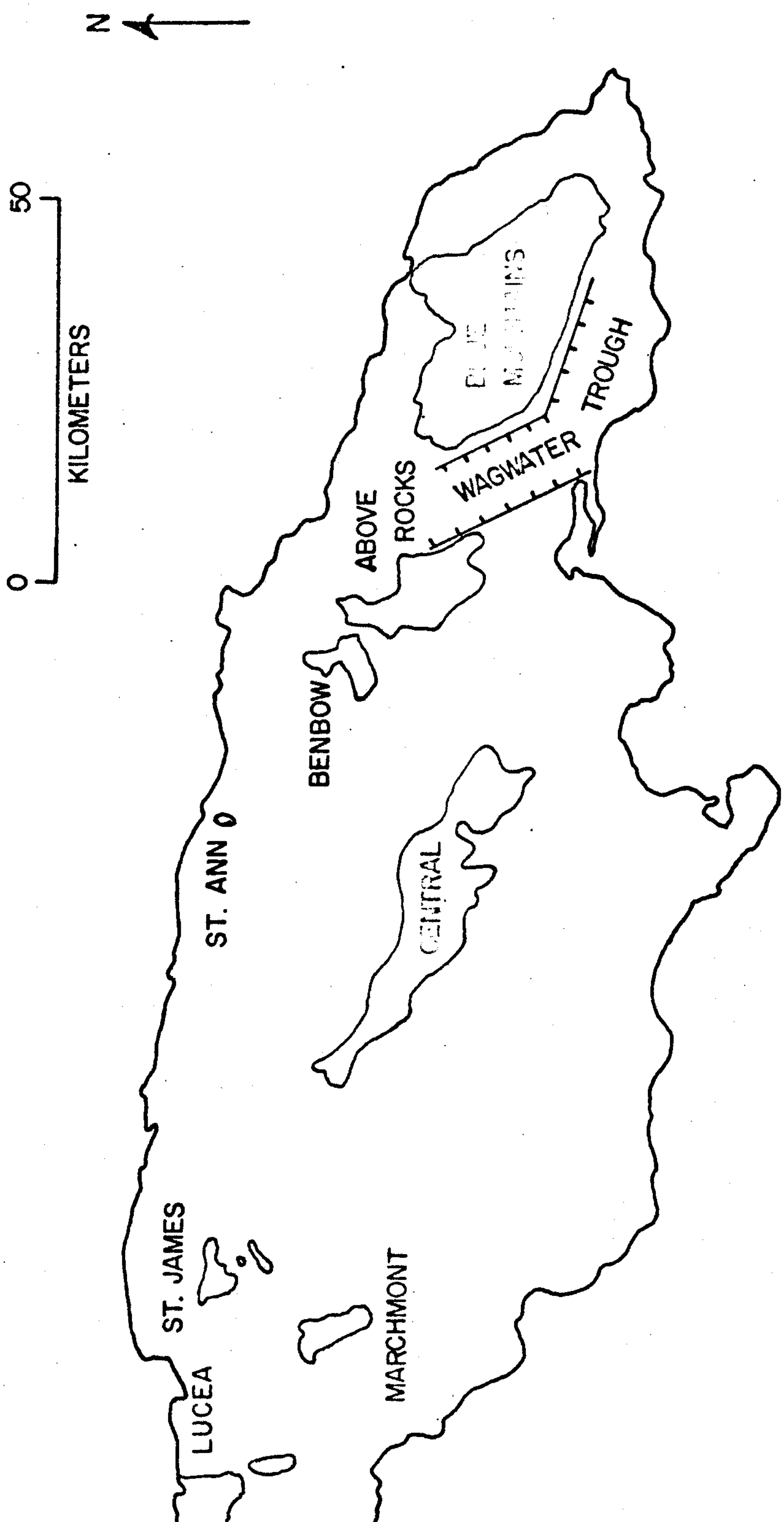

NORTH AMERICAN PLATE
SWAN TF
MID—CAYMAN RISE
ORIENTE TF
CARIBBEAN PLATE
COCOS PLATE
SOUTH AMERICAN PLATE

Figure 3 - Index map to the major Cretaceous inliers of Jamaica. For a complete listing, see the Geologic Map of Jamaica (1977).


N
0
50
KILOMETERS
ST. ANN
BENBOW
ABOVE
ROCKS
WAGWATER
TROUGH
CENTRAL
ST. JAMES
LUCEA
MARCHMONT

volcanic rocks. Most of the sediments have been redeposited by a variety of sediment gravity flow mechanisms (Middleton and Hampton, 1973). Fossil inoceramids and rudists occur approximately midway up the section and are of Late Santonian to Early Campanian in age. This Upper Cretaceous sequence is unconformably overlain by Eocene to Miocene shallow water limestones of the Yellow and White Limestone Groups.

Structurally, the Cretaceous sequence is simply folded about east-west trending axes. A spaced vertical axial planar cleavage is developed in shales and fine siltstones. Two major east-west strike slip faults separate the inlier into three structural blocks.

Metamorphic assemblages within mafic sills and volcaniclastic rocks suggest that the lower part of the section was metamorphosed to the prehnite-pumpellyite facies.

General Setting

Recent uplift of the island and heavy rainfall cause most slopes to be precipitous and exposures to be surprisingly good within the Lucea Inlier. Elevations within the area range from sea level along the north coast, to 1810' in the southeast. Most of the inlier is low lying (< 600') steep, sloped terrane with a very dense dendritic drainage pattern. High ground in excess of 800' is generally restricted to the southeastern end of the inlier. For most of the inlier the lithology generally has a strong influence on topography. High ground around Cascade (1670-5447) and to the northwest is formed by the massive volcanic conglomerate of the Tom Spring Formation.

Sections of conglomerate and/or thick sandstone beds (> 1 m) generally form steep, rough topography (Claremont, 1565-558 and Cash Hill, 1630-5370 region) while shales and intercalated sandstones form open valleys and low-lying topography. Topography is structurally controlled in the southeast where dip of beds are generally in excess of 40°.

No major road transects the inlier except for the "North Coast Road"; therefore, roadcut exposures are generally small but locally continuous. The dispersed distribution of population has created an extensive network of footpaths, and primary and secondary roads (see Plates 1 and 2). Footpaths generally allow easy access to stream sections. The quality and extent of exposure is quite variable throughout the inlier and is generally poor within sections dominated by shale.

Rainfall is generally restricted to late afternoon showers of short duration; therefore, most days spent in the area were devoted to mapping.

Previous Studies

Previous studies of the Lucea Inlier were generally of a reconnaissance nature and were primarily of a paleontologic or economic interest. Early studies (Chubb in Zans, et al., 1962) subdivided the inlier into the Hanover Shales and Clifton Limestone. The Clifton Limestone was assigned a Lower Campanian age on the basis of rudists and corals. Later work by geologists of the Jamaican Geological Survey was compiled by J.H. Bateson and presented as the Lucea Sheet (sheet 2) of the Jamaican 1;50,000 provisional geologic map series. On the basis of sandstone content Bateson subdivided the Cretaceous sediments into two lithologic units (Kha and Khs), though no formal names were proposed. Also several bodies of agglomeratic rocks (Ki) and Clifton Limestone (Khl) were defined. Structurally, Bateson interpreted the Cretaceous sequence to represent the nose and northern limb of a southwesterly plunging anticline that has its northern limb folded.

Purpose

The Caribbean region exclusive of the Lesser Antilles and Central America south of Honduras, is a complex array of Jurassic to Eocene island arc systems. Continued tectonism due to Cenozoic strike-slip motion along the northern and southern margins of the Caribbean (Fig. 2) and slight convergence across the South American-Caribbean plate boundary has enabled geologists to study the well exposed Cretaceous sequences (Burke, et al. 1978). From the 1930's to 1970 Hess and colleagues of Princeton studied the geology and tectonics of the region primarily in view of the tectogene hypothesis. Since, geologists powered with plate tectonic theory have proposed a variety of contradictory models to explain the tectonic evolution of the Caribbean region. The apparent lack of identifiable magnetic anomalies (see Christofferson, 1973; 1976 for another interpretation) within the Caribbean Sea has severely hindered geologists from constructing viable plate tectonic models for the geologic evolution of the Caribbean region. However, Ladd (1976) used magnetic lineations within the Atlantic Ocean to interpret the relative motion of the South American and North American plates for the time interval 127 m.y. to the present. This enabled him to interpret the geologic evolution of the Caribbean region in terms of the relative motion between the North American and South

American plates. Studies similar in principle, have been made for the relative motion of the African and Eurasian plates (Dewey, et al., 1973). These studies have limitations because they only explain the relative motion between the major plates and cannot consider the relative motion of potentially intervening plates. Detailed on-land studies using the principles set out by Dewey and Bird (1970); and Dewey (1975), for identifying and understanding the evolution of plate boundaries, in conjunction with Ladd's study will be the most effective in developing plate tectonic models for the Caribbean.

A major objective of this study was to determine the polarity of subduction with respect to Jamaica during the Cretaceous, from an analysis of paleoslope indicators within the Lucea Inlier of western Jamaica and comparison with present day island arc models (Karig, 1974; Dickinson, 1977). The spatial relationship of other sub-island arc environments, present on the island were also considered in this study. It soon became apparent that paleoslope indicators in themselves for a restricted region of an ancient island arc complex are not significant parameters for understanding the polarity of subduction. Paleocurrent data for island arc sedimentary basins can be expected to be variable and to represent transport into fore-arc basins, intra-arc basins and back-arc regions.

Meaningful tectonic models of the Caribbean have to consider Cenozoic strike-slip displacements across the northern and southern Caribbean plate boundaries. This displacement is analysed for the northern boundary.

Methods of Investigation

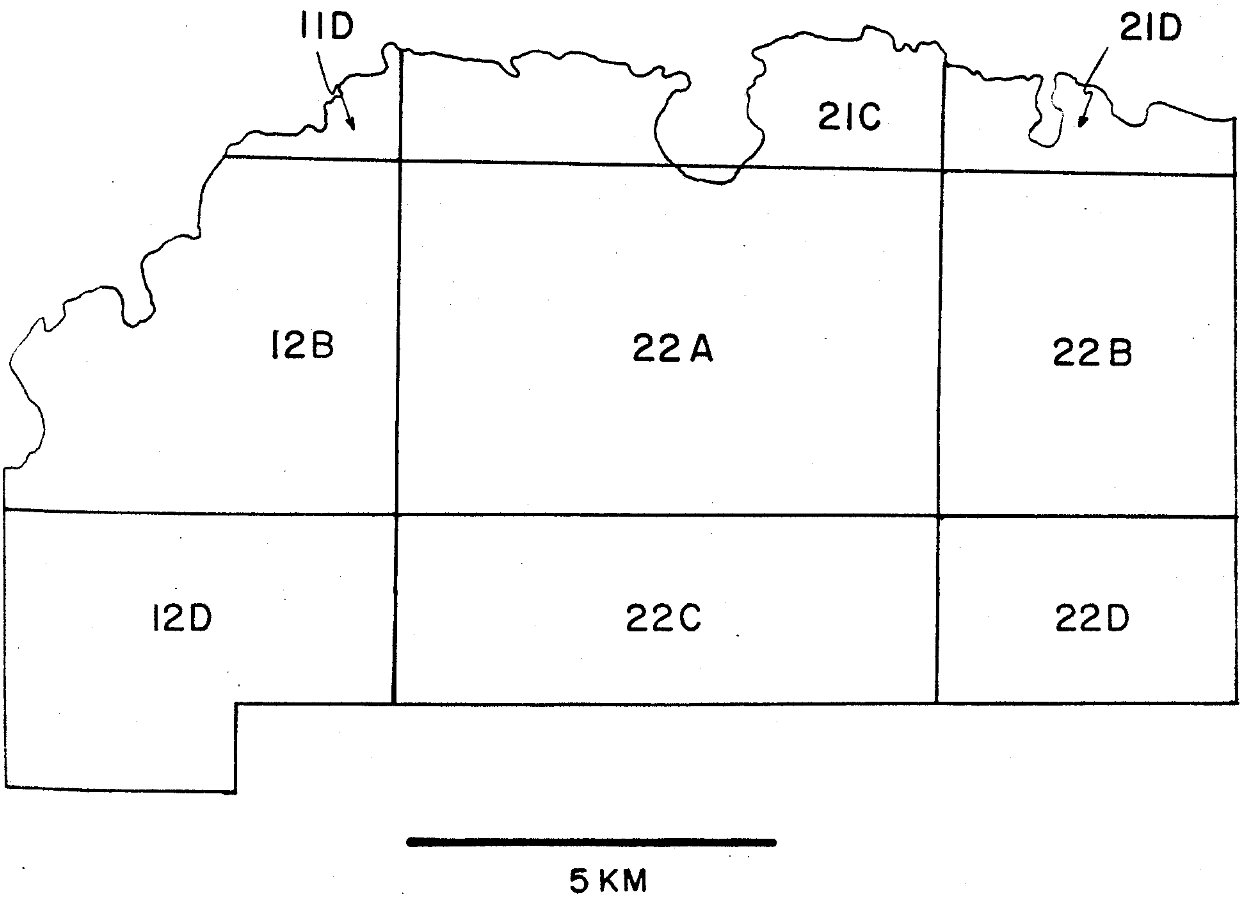
The area mapped extends over an area of approximately 70 sq. km. and covers Cretaceous sequences that crop out on topographic sheets (1:12,500 scale): 21C, 21D, 22A, 22B, 22C and 22D of the Jamaican Survey (Fig. 4). A period of three months were spent in the field mapping. The above area covers approximately 80 percent of the Lucea Inlier, however, Plate 1 is a geologic map of the whole inlier and is based on data collected in this study and by geologists of the Jamaican Geological Survey. The area of the latter is contained within topographic sheets 12B, 12D and 11D (see Fig. 4).

Mapping was done on accurate 1:12,500 topographic maps (contoured at 25' and 50' intervals) of the Jamaican Survey. All road sections were mapped by pacing. Because of the quality of the topographical maps and accuracy of pacing the location of stations are thought to be accurate to ± 20 m along roads and locally less accurate along stream sections. Aerial photographs at a scale of approximately 1:25,000 (Spartan Series, 1968) were used in the lab to check and make structural interpretations.

Dense vegetation and the deep level of weathering has restricted exposures to roadcuts, excavations near dwellings, footpaths and to stream beds. Exposures along primary roads were surprisingly good; though the degree of weathering varied greatly. Most of the roads within the study area were mapped. However, more extensive coverage was carried out in the eastern part of the area (Plate 2), where many stream sections and footpaths were traversed. The eastern end was mapped in much greater detail for its geologic interest, and because previous geologic investigations of the Lucea Inlier were largely

Figure 4 - Index map to the 1:12,500 topographic maps covering the Lucea Inlier and adjacent Cenozoic limestone.


N
50 KM


11D
21D
21C
12B
22A
22B
12D
22C
22D
5 KM

restricted to its western end.

Most of the boundary between the Cenozoic limestone and Cretaceous sequence is based on an interpretation of aerial photographs by Dr. Kevin Burke and is controlled by data on unpublished maps of Jamaican Geological Survey. This data was kindly made available to us through the efforts of Neville McFarlane.

CHAPTER II

STRATIGRAPHY

Introduction

Earlier work within the Lucea Inlier was of a reconnaissance nature and/or was primarily fossil oriented or of economic interest. The inlier has been subdivided into the Clifton Limestone and the overlying Hanover Shale (Wright, 1974). However, because this present study is more extensive many more lithostratigraphic units have been defined. Rock names used in this study are proposed as formal lithostratigraphic units but are subject to approval by the Jamaican Geological Survey.

All thicknesses given are estimated from the map by methods described by Compton (1962, p. 240). Only the Patty Hill Formation is continuously exposed and therefore was measured directly.

The gross lithologic and sedimentological characteristics of each unit are described below in ascending order.

BIRCHS HILL FORMATION

The Birchs Hill Formation is a 820 m + to 500 m + thick sequence of predominantly medium to thickly bedded massive volcanic lithic and feldspathic sandstones with minor intercalated siltstones. A mappable horizon of interbedded thin sandstone, shale and micritic limestone occurs near the base of the Birchs Hill Formation, and is here defined as the Castle Hyde Member of the Birchs Hill Formation. The upper and lower sandstone units of the formation are separated by the Castle Hyde Member and will be informally referred to as the lower and upper

Birchs Hill Formation. The type section of the Birchs Hill Formation is exposed along the Cash Hill-Dunalva road and within road cuts southeast of Cash Hill. The Birchs Hill Formation crops out in the southeastern corner of the mapped area, within the core of the westward plunging Cash Hill Anticline and is the lowermost lithostratigraphic unit defined in this study. Cretaceous sediments that crop out farther to the east within the Lucea Inlier are poorly exposed and too little known to permit their formal designation in this study. They consist mainly of shales and interbedded sandstones and local conglomerate.

The exposed thickness of the lower Birchs Hill Formation is 50 m and 160 m thick in the Castle Hyde (1732-5404) and Cash Hill (1632-5370) areas, respectively. It is generally composed of 0.3 to $\geq$ 4 m thick, medium to coarse sandstone beds with granule to cobble-sized conglomerates. Green, red and dark grey feldspar phenocrystic volcanic rocks form clasts in the conglomerates (Fig. 5). Thinner sandstone beds and dark-grey siltstones are intercalated but are minor. Sandstones are generally thicker and coarser to the southwest. On fresh surfaces the units have a light bluish-grey color (see Fig. 5) and weather from a dark blue-grey to a buff-tan on deeply weathered surfaces. This contrast in color is well displayed by generally fresh exposures west of Castle Hyde and deeply weathered ones east of Cash Hill.

Sedimentary structures of the kinds associated with turbidity currents are generally scarce. Ta and Tb intervals of the Bouma sequence were observed but are very minor, and channeling is locally developed. Amalgamation (Walker, 1967) is suspected for the thicker sandstone beds. Beds are generally massive and do not appear to have any internal structures. A finely layered (1642-5365) horizon of graded silt and shale occurs between massive sandstone beds east of Cash Hill. This horizon shows penecontemporaneous brittle and ductile deformational features probably induced during deposition of the overlying thick sandstone bed (Fig. 6).

The basal contact of the lower Birchs Hill Formation was not defined in the Cash Hill area but is fault bounded west of Castle Hyde (1729-5405). Its contact with the overlying Castle Hyde Member is gradational, over a 20 m interval, in the southwest and fairly abrupt to the northeast.

Castle Hyde Member

The Castle Hyde Member is defined from exposures west of Castle Hyde and within the Cash Hill area. The exposed thickness is 160 m and 210 m in the two areas, respectively. In the Cash Hill area the unit is separated from the overlying thick sandstones by a fault. Where not faulted the upper and lower contacts are clearly defined by a fairly abrupt to transitional relationship with thick massive blue-grey sandstones of the Birchs Hill Formation.

This member is composed of thinly (4 cm) bedded, laminated and non-laminated fine sandstones and intercalated shale and siltstone (Fig. 7). Local 25 cm thick sandstone beds also occur. West of Castle Hyde the upper 40 m consist of 3 to 40 cm thick beds of micritic limestone. On fresh surfaces the micrite is light-grey and weathers to a cream buff-brown. The limestone has a choncoidal fracture and is a clink stone. This horizon of micrite is lenticular or has been faulted out so that it does not crop out to the southwest. Since

Figure 5 – Typical texture and coloring of a massive pebbly sandstone of the Birchs Hill Formation. Note subrounded to subangular clasts of plagioclase phenocrystic mafic volcanic rocks.

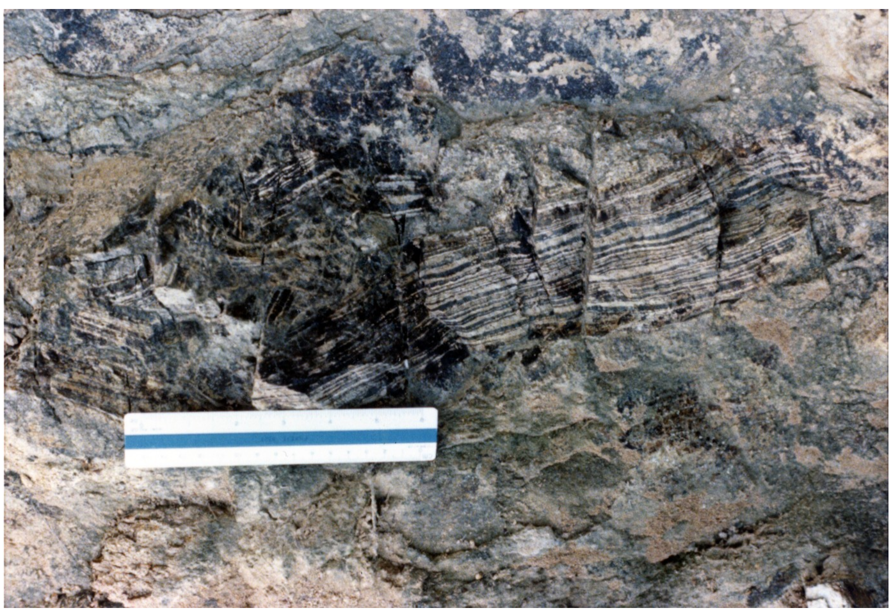


Figure 6 – Penecontemporaneously disrupted finely layered horizon between massive sandstone beds of the Birchs Hill Formation.

it represents a period of low detrital input and probably has a significant lateral extent; it would then seem more reasonable that it has been faulted out.

The upper Birchs Hill Formation is similar to the lower Birchs Hill Formation in outcrop character, color, sedimentary structures and lithology. This upper horizon is well exposed along the Cash Hill-Dunalva road and the Castle Hyde-Patty Hill road. It is approximately 450 m thick in the southwest and 280 m thick to the northeast. The reference section for this interval is along the Cash Hill-Dunalva road.

Within this section there is a gross fining and thinning upward sequence. The lower 2/3 of this section consists of 0.7 to > 5 m thick massive, generally ungraded, medium to coarse sandstones with minor pebbly sandstones and intercalated thinly bedded (1-20 cm) fine sandstones and siltstones (Fig. 8). The upper 1/3 of the sequence is generally more thinly bedded, although thick ($\geq$ 1 m) sandstone beds occur sporadically.

Generally, beds in excess of 1 m are ungraded to slightly graded and are massive except for local faint laminae (Fig. 9); discontinuous silty shale lenses separating thick sandstone beds suggest amalgamation. Intraformational dark-grey silty shale rip ups, rare sole markings, small scale flame structures and complex relationships between sandstone beds and underlying shale (Fig. 10) suggest deposition by turbidity currents. These features are generally restricted to beds < 1 m in thickness. Locally, extraformational medium-grey calcarenite clasts make up to 25% of the rip ups. Rip ups generally occur low in the Ta interval, though one bed was observed to have

Figure 7 – Deeply weathered outcrop of the Castle Hyde Member. Note thinly bedded character.

Figure 6 – Typical outcrop character of the Birchs Hill Formation. Medium to coarse sandstone beds with little intervening shale-siltstone. Hammer as scale, center.

Figure 9 – Laminated medium to coarse sandstone bed of the Birchs Hill Formation. Note wispy character in lower left-hand corner of photograph and size sorting. Bands of coarser sand that are bounded by laminations may represent discrete pulses of sediment deposition

Figure 10 – Complex interfingering of turbidity current emplaced sand and underlying shale. Note development of a thin (~5cm) parallel laminated interval at top and bottom of photograph. Scale is 16 cm.

them in the upper portion with a thin Tc interval above it. Thinner (~ 20 cm) sandstone beds generally show good development of Tab and Tb intervals.

To the northeast the upper Birchs Hill Formation is generally more thinly bedded and more pelitic. However, thick (up to 3 m +) beds of coarse sandstone to pebbly conglomerate occur within the section. Generally beds here are distinctly more laminated then beds to the southwest. Thick massive beds of volcanic breccia were observed intact at several localities (1649-5393 and 1687-5405). To the southwest only one loose boulder of volcanic breccia was observed along the Cash Hill-Dunalva road.

PATTY HILL FORMATION

The Patty Hill Formation is a ± 40 m thick sequence of thinly bedded ($\lesssim$ 10 cm) shale, and minor sandstone and limey beds that conformably overlies the Birchs Hill Formation. In outcrop, it is generally deeply weathered to a rusty, buff-brown (Fig. 11). However, 0.8 km west of Patty Hill (1632-5400) it is relatively fresh and continuously exposed in a road cut; this exposure has been chosen as the type locality. The outcrop is cut by a few small faults. Here the formation consists of 2-10 cm thick beds of shale, and minor fine to coarse sandstone and calcareous beds. Horizons of ribbon-banded brown weathering shale and light grey calcareous bands occur within this sequence (Fig. 12). Sandstone and limestone locally make up to 30% of the section. Overall the formation shows a well-developed thickening and coarsing upward sequence. The contact with the

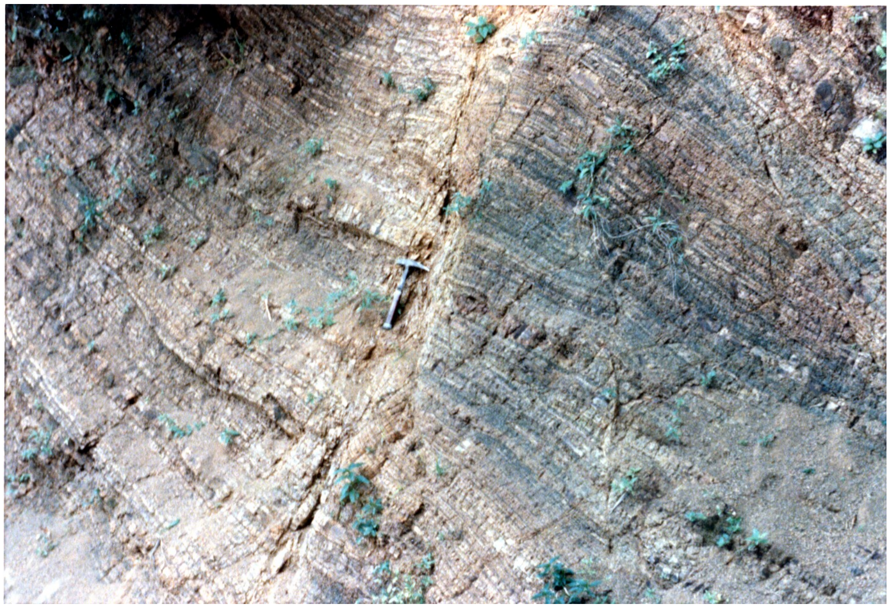


Figure 11 – Deeply weathered finely banded horizon within the Patty Hill Formation. It has a similar appearance to horizons within the Castle Hyde Member (see Figure 7).

Figure 12 – Close-up of a relatively fresh exposure of the fine ribbon banding of the Patty Hill Formation.

overlying Tobolski Formation is conformable and marked by the first appearance of a 1 m thick coarse, poorly sorted sandstone bed with abundant rip ups. Above this bed the sequence is 90% medium to coarse sandstone.

The Patty Hill Formation is a distinctive lithological unit that has been traced to the south and east. To the south horizons of red and green shale are locally developed within it (1591-5366 and 1608-5342).

TOBOLSKI FORMATION

The Tobolski Formation is a 200 to 450 m thick sequence of predominantly sandy pebble to boulder conglomerate with local horizons of interbedded sandstone and shale (see Plate 2B). Up to 4 m thick pebbly mudstones have been observed. Clasts are generally well rounded and all are volcanic. The Tobolski Formation is not well exposed in the study area and is defined by deeply weathered road cut exposures (1584-6360) and by small exposures and float within footpaths approximately 1 km south of Dunalva. The latter area is taken as the type locality. The formation is more or less restricted to the core and southern limb of the Cash Hill Anticline. It is thought to be represented in the northern limb, but here the Tobolski Formation is much thinner, although whether this thinning is due to east-west faulting, erosion or non-deposition is as yet undetermined.

ASKENISH FORMATION

The Askenish Formation is the uppermost definable lithostratigraphic unit exposed in the Cash Hill Anticline. It is predominantly a medium to thinly bedded silty shale which is bluish-grey on fresh surfaces and weathers to a rust-red to brown. Thin (2-7 cm) beds of medium to coarse sandstone and 1 to 70 cm thick sandy pebble to boulder conglomerate and pebbly mudstone beds occur sporadically but increase in quantity upsection. The exposed thickness ranges from 180 m in the southern limb to 230 m + in the nose of the Cash Hill Anticline, south of Askenish. To the northwest it is not exposed and is thought to be unconformably overlain by the Pioneer Formation. In the Askenish area the Askenish Formation is unconformably overlain by rocks of the Cenozoic Yellow and White Limestone Groups, therefore its original thickness is not known. Contacts with the underlying Tobolski Formation are not well exposed but appear to be gradational (1572-5460).

The following definition of the Askenish Formation is based on exposures in the nose of the Cash Hill Anticline, south and east of Askenish and is taken as the type area.

The lower 80 m of the formation averages 90% silty shales with thin (6-7 cm) calcsiltite beds and elnticular 3 to 5 cm coarse sandstone beds. Branching arthropod networks (Thalassinoides identified by E. Kauffman) occur on the tops of sandstone beds. Within grey shales three elongate fossilized objects were observed to have preferred orientation of 090°, 090° and 072° (1566-5376). Above the lower 80 m a variety of relatively thin pebbly mudstones to sandy pebble to cobble conglomerates begin to occur and increase in frequency farther up in the section (Fig. 13). Sand may make up to 40% of the section but silty shale is still the dominant lithology. Clasts within pebbly mudstones are dominantly limestone or an admixture of limestone and purple volcanic rocks. Matrix makes up to 90% of the beds, but some exposures show lateral heterogeneities in the matrix to clast ratio.

Thin irregularly bedded and sharp-bottomed sandy pebble conglomerates commonly occur intercalated with thinly bedded shales (Fig. 14). Most conglomerate beds are structureless though some show normal grading while two examples of inversely graded pebbly mudstone and conglomerate have been observed.

The inoceramid Platyceramus cycloides cycloides (Wegner) (identified by E. Kauffman) of Santonian to Lower Campanian age was found in thickly bedded buff-brown weathering shales (1525-5380) of the Askenish Formation.

The contact of the Askenish Formation with the overlying Eocene Yellow Limestone was observed at two localities (1572-5326, 1544-5368); relief within the contact zone was 1 to 2 m and approximately 10 m, respectively (Fig. 15). This contact is probably an angular unconformity but could not be unequivocably shown to be so in this region.

PIONEER FORMATION

The Pioneer crops out over a large area (in excess of 6 sq km) to the north, east and west of Cascade and consists predominantly of sandstone and conglomerate, with lesser shale. Bedding consistently strikes north-northwesterly and has a gentle (0°-25°) westerly dip. Undiagnostic and complex facies patterns, and inadequate outcrop information have made it impossible to subdivide the Pioneer Formation. The thickness of it is also poorly understood, but an estimate of 200-400 m seems reasonable. The formation is generally deeply weathered to hues of rust-red to tan making exposures generally poor along road sections; however, good outcrops can be observed along stream sections. The formation is named after a "Pioneer Farm" that

Figure 13 – Shales and interbedded thin sandy pebble conglomerate and coarse sandstone beds of the Ashkenish Formation.

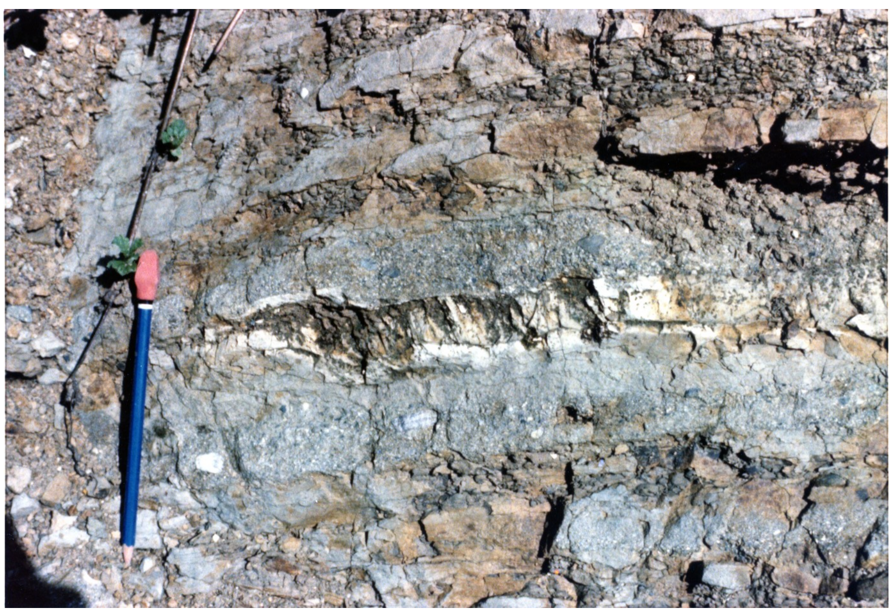


Figure 14 – Close-up of common thinly bedded pebble-granule conglomerate within a predominantly shale sequence, Ashkenish Formation.

Figure 15 – Contact of shales of the Ashkenish Formation and unconformably overlying Yellow Limestone of Eocene age.

is located 1 km northeast of Cascade. Exposures along the Cascade-Pondside road and immediately northeast and southeast of Cascade are considered to be the type section.

Along the Cascade-Pondside road the Pioneer Formation consists of thick (1-4 m +) massive, coarse sandstones and sandy pebble to boulder conglomerate (up to 5 m + thick) beds with minor intercalated parallel laminated (see Fig. 16), locally channeled, sandstone (2-20 cm+) and shale beds. Breccia beds containing clasts of grey, fine-grained and minor red and vesicular grey volcanic rocks were observed low in the section (Fig. 17). Clasts (1-20 cm) within conglomerates are generally well to subrounded mafic volcanic rocks with minor limestone and chert. Conglomerates are typically unorganized and have a sandy matrix in which clasts may be either matrix or clast supported.

North of Cascade and east of the Tom Spring Formation the Pioneer Formation is well exposed within stream sections of the Maggotty River basin. These exposures probably represent higher stratigraphic levels within the Pioneer Formation. The unit is still characterized by thick (1-8 m) massive coarse sandstone beds and sandy pebble to boulder conglomerate beds, but 5-80 cm thick turbiditic sandstone beds and intercalated shales are more abundant within this section. Channeling and amalgamation were commonly observed. Channeling of conglomerates was observed at the outcrop level (see Fig. 18) and is inferred on the map scale on the order of tenths of a kilometer. A spectacular 20 m high waterfall outcrop of 1 to 8 m thick sandstones can be seen in the Maggotty River approximately 2 km north of Cascade (1655-5508). Most clasts within conglomerates are mafic volcanics with minor limestone and medium-cream colored grey chert (Fig. 19).

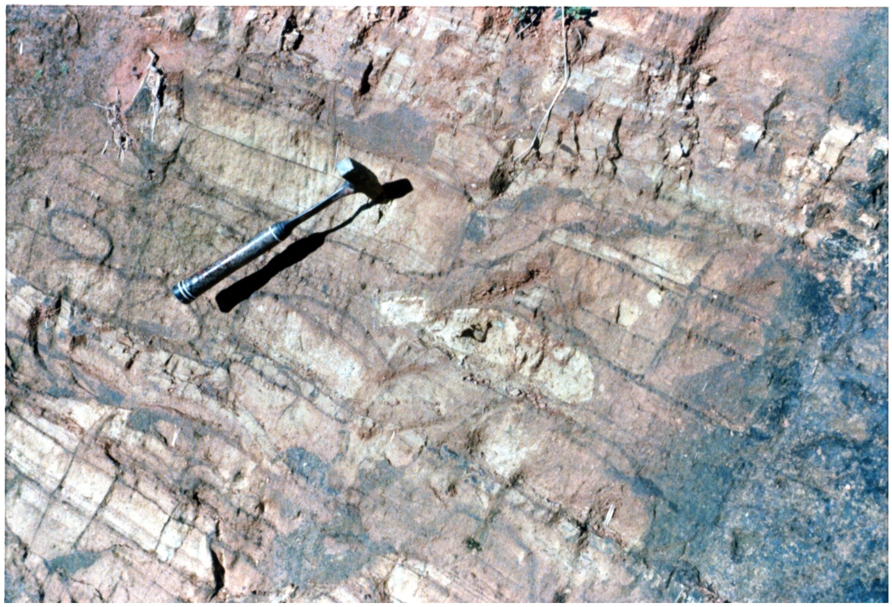


Figure 16 – Deeply weathered exposure of laminated medium grained sandstones of the Pioneer Formation. Note concretion, right of hammer, and general lack of shale.

Figure 17 – Texture of volcanic breccia near the base of the Pioneer Formation.

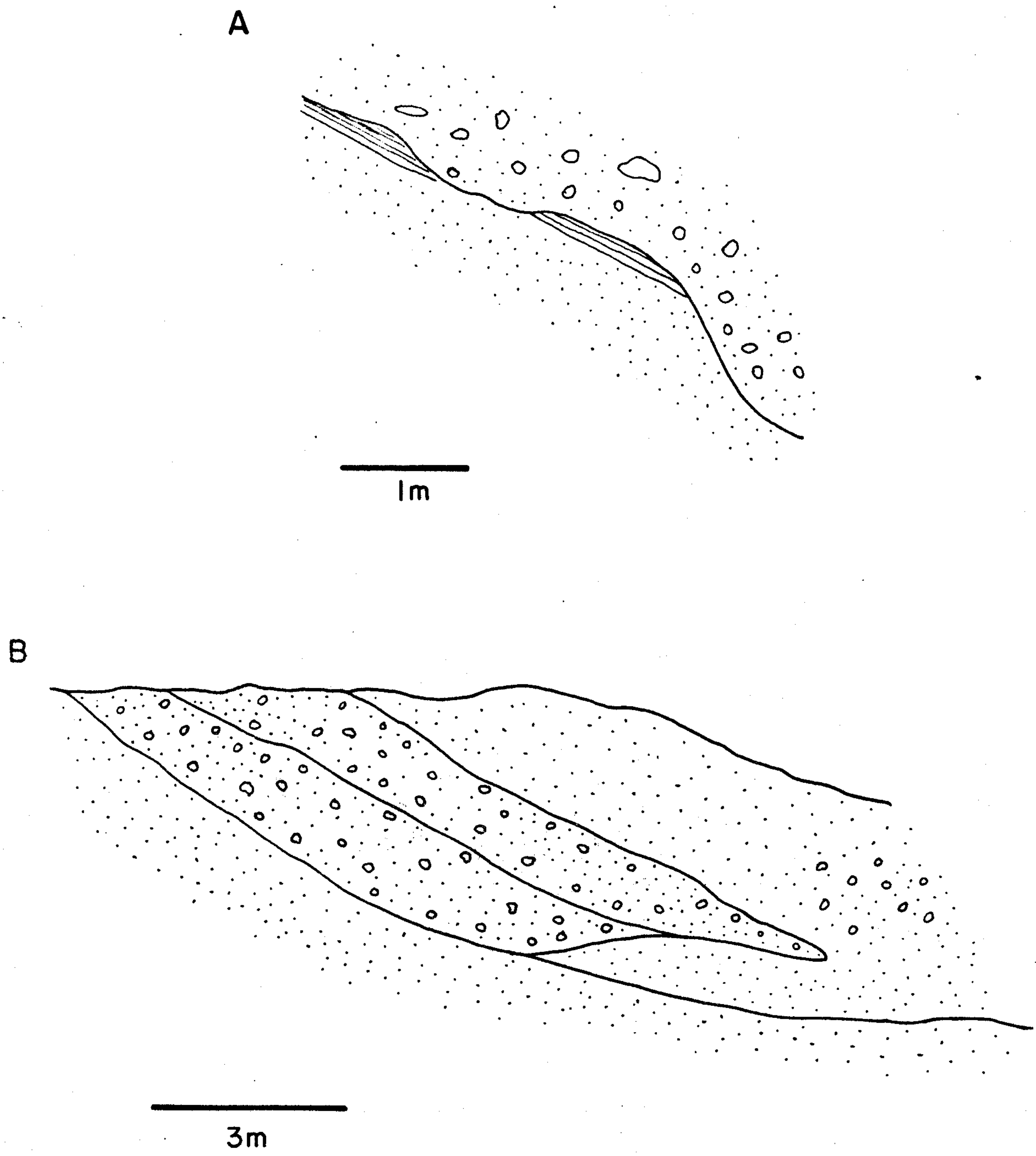


Figure 18 - Outcrop sketches of channelized sandy pebble to boulder conglomerates. A. underlying sandstone bed has the Ta and b interval developed. B. note heterogenous distribution of pebbles in the overlying sandstone bed.

Figure 19 – Close-up of subrounded to rounded volcanic clasts within a sandy matrix. Clast supported conglomerate of the Pioneer Formation.

Thinner sandstone beds show good T_b and T_a intervals of the Bouma sequence. It is relatively common to observe rip up clasts concentrated within the upper half of sandstone beds around 1 m thick. This may be an expression of the contrast in density between mud rip ups and incoming sands.

A more pelitic and thinly bedded facies of the Pioneer Formation crops out west of Cascade within the eastern headwaters of the Lucea East River (1622-5438). Sandstone beds are generally well laminated. Thick (1 m+) massive sandstone beds occur in a few places within the section. Locally, sandstone beds contain small ($\lesssim$ 2 cm) irregular nodules of pyrite. These localities are close to the overlying Tom Spring Formation. This is discussed further below (see p. 48).

The upper and lower contacts of the Pioneer Formation are obscured by the Maryland Fault zone (p.95) and by complex facies relationships between the Tom Spring Formation and the Georgia Complex. The upper contact, with the Mount Peace Formation, is discussed below (see p. 54). In the east a sliver of the Pioneer Formation is thought to crop out south of the Maryland Fault (1700-5426, see Plate 2). However, its contact with the underlying units of the Cash Hill Anticline, is not well understood. Plate 2 shows it to overly the Tobolski and Patty Hill Formations. The Askenish Formation, which overlies the Tobolski Formation to the west, is not observed and the Tobolski Formation thins from 400 m in the west to < 100 m in the east. This data plus the apparently sharp structural discordance between the Pioneer and underlying units of the Cash Hill Anticline suggest the existence of an angular unconformity. Also lending support to this idea is a small outcrop of coarse sandstone on the crest of Birchs Hill. Bedding within it is similar in attitude to the Pioneer Formation nearby and this outcrop may represent an erosional remnant.

The above relationships may also be explained by omission of the Askenish Formation due to strike-slip faulting. The thinning of the Tobolski Formation can also be explained by a facies change. The conglomerates of the Tobolski Formation in the east probably represent a submarine channel deposit and should be expected to laterally thin out.

TOM SPRING FORMATION

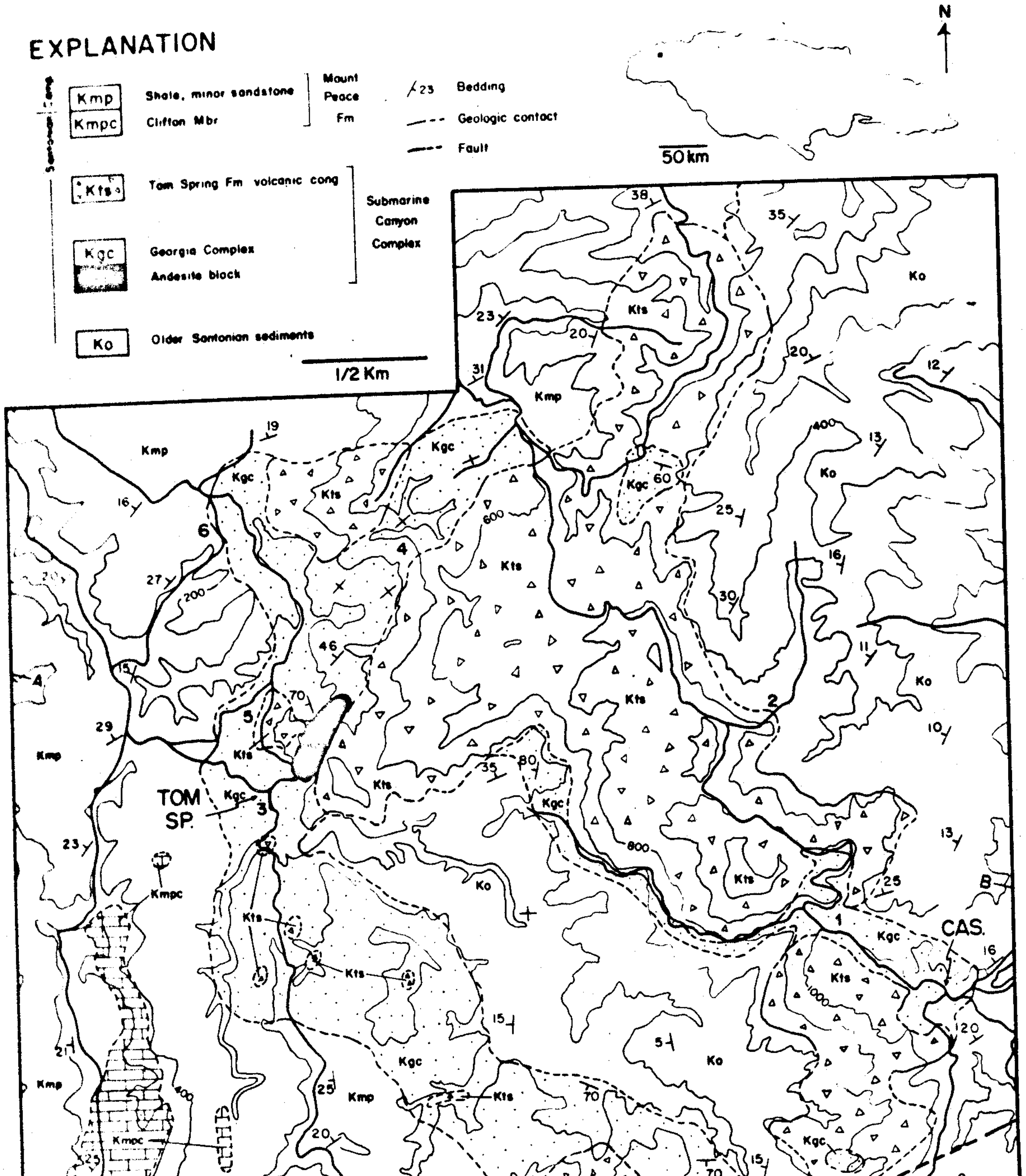
The Tom Spring Formation is a massive, unbedded, irregular 100 m+ thick body of andesitic conglomerate. Non-vesicular volcanic rocks ranging from $\leq$ 4 cm to 1.5 m across are encased in a matrix of mafic sand. The matrix is always the dominant phase, generally making up 90-95% of the rock, but locally it forms as little as 60% (Fig. 20). Many of the volcanic rock fragments have irregular shapes which suggest that they have not been transported for any appreciable distance within a fluvial or marine environment. Outlines of some of the volcanic blocks show signs of resorption (Fig. 21). The type lithology can be seen along the Cascade (1667-5446)-Jericho (1576-5558) road and around Tom Spring (1565-5480). The Tom Spring Formation has a very complicated map pattern and relationship with surrounding and underlying sediments (Fig. 22). However, it appears to form an elongate lens with a gross north-northwesterly trend, and a gentle northwesterly dip is apparent (see Plates 2 and 2B). This trend locally cuts across the regional strike and the dip of adjacent beds.

In outcrop the Tom Spring Formation is extremely massive and unbedded (Fig. 23). Indications of layering in the Tom Spring

Figure 20 – Outcrop character of Tom Spring Formation within the Lucea East River, approximately 0.25 km south of Tom Spring (1565-5480). Note the range in size (<= 4cm-1.5m) of volcanic rock fragments and their irregular shapes. White veins cutting exposure are filled with barite and carbonaceous material. Note hammer as scale.

Figure 21 – Resorbed, undulating contact between volcanic rock (lower half of photograph) and mafic matrix of the Tom Spring Formation.

Figure 22 - Geologic map of the Tom Spring Formation and adjacent units. Refer to Plate 2 for greater detail.


EXPLANATION
Kmp
Shale, minor sandstone
Kmpc
Clifton Mbr
Mount Peace Fm
Kts
Tom Spring Fm volcanic cong
Kgc
Georgia Complex
Andesite block
Submarine Canyon Complex
Ko
Older Santonian sediments
23 Bedding
Geologic contact
Fault
1/2 Km
50 km
N
TOM SP.
CAS.
Kmp
Kgc
Kts
Ko
Kmpc

Formation have been observed at only two localities (1665-5447 and 1584-5512). Both were discontinuous in character while one (1584-5512) showed fine discontinuous laminae with 3 cm graded intervals. It is possible that the Tom Spring Formation is bedded on a scale larger than the outcrop scale. Color is variable and ranges from hues of blue-purple to red weathering to pale hues of light purple. Red tones are due to hematitic staining which occurs in a patchy pattern. Veins filled with barite and carbonaceous matter cut some exposures of the Tom Spring Formation (e.g., 1566-5475).

Because the Tom Spring Formation contrasts in layering, lithology and outcrop pattern with the adjacent units, it is generally associated with a characteristic topography and soil. Topographically, it is expressed by high ground and steep gradients. Its contact with the underlying and laterally adjacent Georgia Complex is generally defined by an abrupt decrease in slope (see Fig. 24 and Plate 2). Soil developed on the Tom Spring Formation is characteristically pale purple in color and contrasts with rust red-brown soil of adjacent units. This contrast in soil color proved to be a useful tool in mapping.

Petrographically, the matrix of the Tom Spring Formation consists of volcanic lithic fragments that are set in a matrix of altered plagioclase and pyroxene crystals. Chlorite extensively replaces pyroxene, hornblende and plagioclase. Epidote is developed but is very minor.

Based on the detrital character and geologic relations of the Tom Spring Formation it has been interpreted as a resedimented volcanic conglomerate that filled and disrupted the walls and floor

Figure 23 – Weathered outcrop of the massive Tom Spring Formation. This type of spheroidal weathering is locally developed.

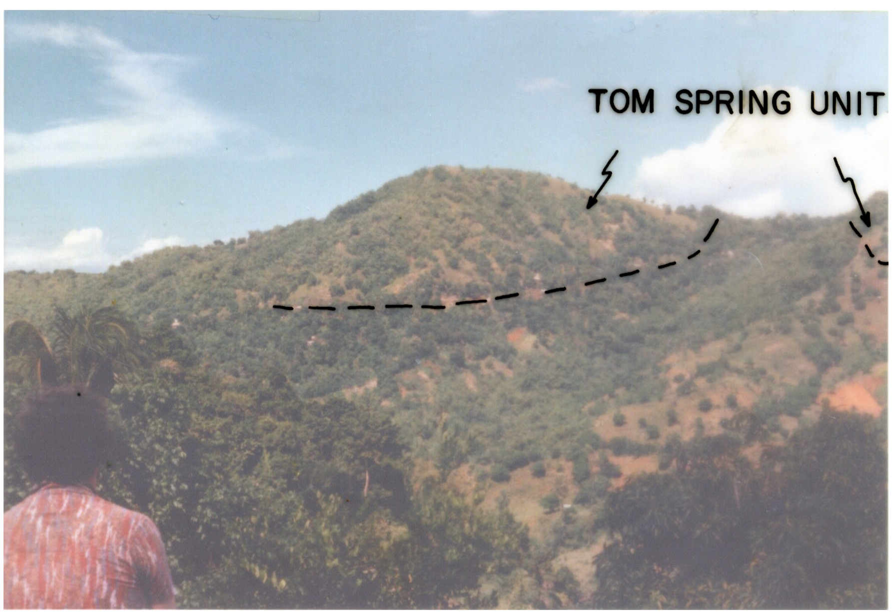


Figure 24 – View looking north from a locality (16031-4508) along the Patty Hill (1665-5407) to Maryland (1588-5412) road. Hill is northwest of Cascade and is formed by the massive Tom Spring Formation. Note contrast in slope, type and color of vegetation. Overlay outlines approximate contact of the Tom Spring Formation and underlying Georgia Complex.

of a submarine canyon. Further evidence to substantiate this interpretation is presented below.

GEORGIA COMPLEX

The Georgia Complex is a disturbed to chaotically disrupted sequence of shale, volcaniclastic sandstone, calcarenite and minor conglomerate that crops out in the east-central part of the mapped area (Plates 1 and 2). It has a very complex three-dimensional relationship with (Fig. 22, p. 37) the overlying, interfingering and cross cutting Tom Spring Formation. The Georgia Complex is defined by its generally complex structure and spatial proximity to the Tom Spring Formation; therefore, it is not easily accommodated as a "normal" lithostratigraphic unit but is here defined as a complex.

The Complex is lithologically variable, ranging from 90-100% dark olive green to dark silty shale (1673-5447) with interbedded calcarenite and conglomerate, to an interbedded sequence of 0.1-> 3 m fine sandstone to pebbly conglomerate beds and intercalated shales. Thinner sandstone beds generally display grading, parallel laminations and cross bedding while thicker beds are generally massive and ungraded.

The deformation of the Georgia Complex is thought to be penecontemporaneous and to be related to the catastrophic emplacement of the overlying Tom Spring volcanic conglomerate. However, some of the structural complexity of the Georgia Complex may be attributed to primary sedimentary features associated with development of the submarine canyon before the Tom Spring was emplaced, and others to later tectonic processes. The complex array of processes active within submarine canyons have been documented by Stanley and Unrug (1973); Whitaker (1977, 1974) and Andrews and Hurley (1978).

Detailed descriptions of the contact relations and structural variability of the Georgia Complex are given below.

Plate 2 shows the Georgia Complex to border the Tom Spring Formation on most sides. It has only been mapped where it has been seen in outcrop. However, if the Georgia Complex is related to the emplacement of the Tom Spring Formation, one would expect it to occur on all sides of the Tom Spring Formation. However, one locality (1644-5591) shows sediments surrounding the Tom Spring Formation to be undisturbed while another (1662-5476) has a very narrow zone (~ 40 m) of disrupted beds.

Contact Relations

The contacts between the Georgia Complex and the overlying Tom Spring Formation are variable. These relationships are best exposed in the Cascade area and to the northwest in single outcrops, and are discussed below.

Two exposures show shales of the Georgia Complex in sharp lithologic contact with the overlying massive volcanic conglomerate of the Tom Spring Formation (1635-5456 and 1627-5459, Fig. 25). East of these exposures the Tom Spring Formation is channelled > 10 m into shales of the Georgia Complex. North of Cascade a large exposure (1665-5447) displays a gradational lithologic contact between the Tom Spring volcanic conglomerate and shales of the Georgia Complex. Figure 26 is a sketch of this outcrop. Large blocks of light tan-weathering sandstone beds have been folded and attenuated to form discrete fragments that are encased in a coarse andesitic matrix. Discontinuous

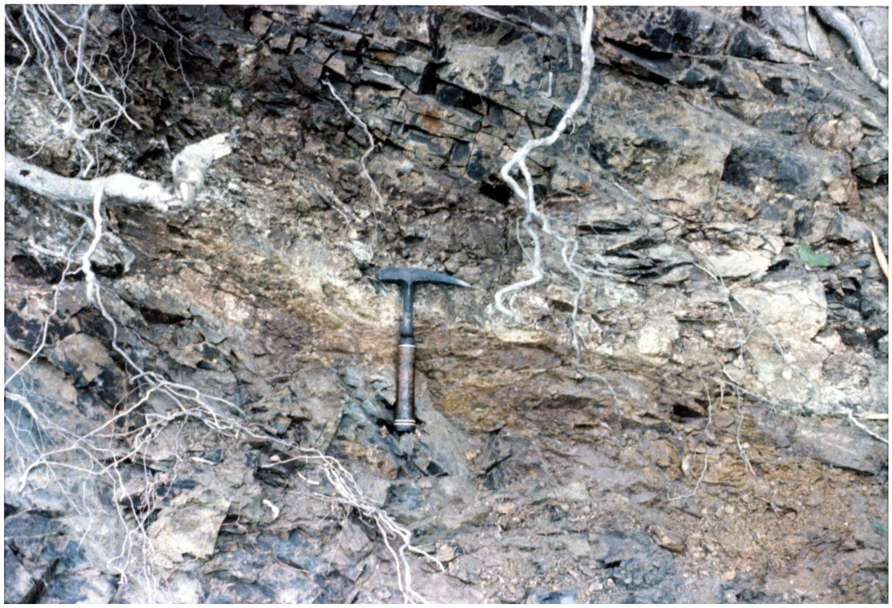


Figure 25 – Relatively sharp lithologic contact of Tom Spring Formation and underlying unbedded shale-siltstone of the Georgia Complex. Hammer rests on contact. Coloring of the Georgia Complex along the contact gives the appearance that it has been hydrothermally altered.

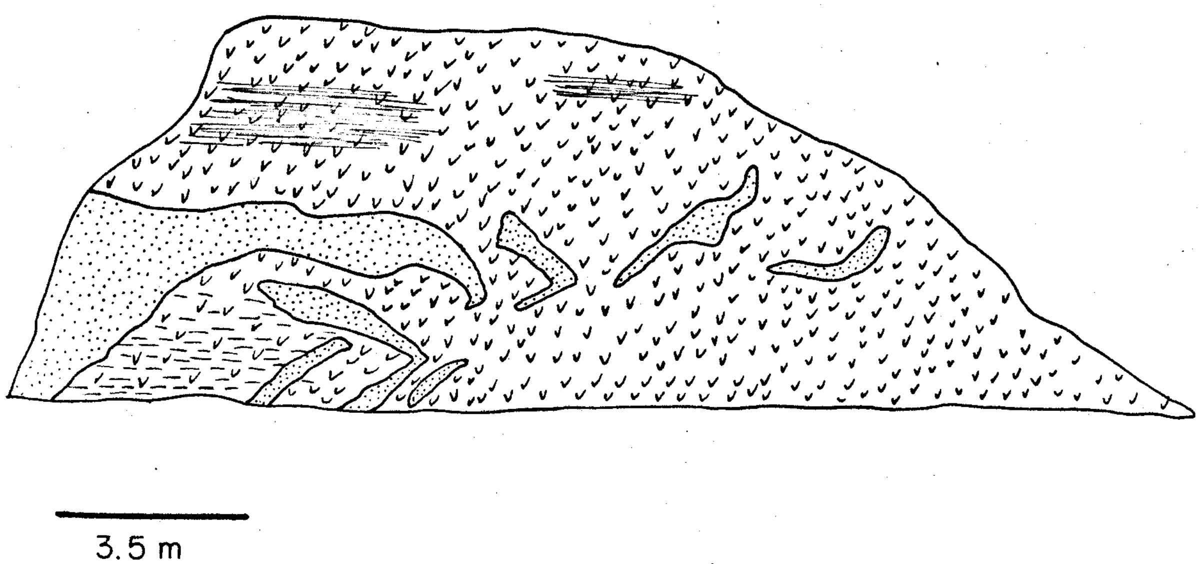


Figure 26 – Sketch of outcrop showing the contact zone between Tom Spring Formation (v pattern) and underlying Georgia Complex. Note admixture of shale and volcaniclastic matrix of Tom Spring Formation. Also note faint layering and disrupted beds of sandstone.

layering is locally developed in the andesitic debris. In the lower right hand corner of the sketch the andesitic matrix grades into a poorly sorted mixture of coarse andesitic debris and shale. Similar gradational lithological contacts have been observed to the north (1645-5556 and 1609-5529). Gradational contacts such as these record local intimate mixing of the incoming volcanic debris of the Tom Spring Formation and sediments that previously lined the walls and floor of the submarine canyon in which it was deposited.

Near vertical contacts have been mapped to the north (1600-5525) between the Tom Spring Formation and a steeply dipping ($\gtrsim 60°$) coherent block of the Georgia Complex. Continuous exposures showing this relationship have not been observed but are inferred from map patterns.

A transition from the locally developed chaotic structure of the Georgia Complex to the coherent adjacent and underlying units is observable at several localities. It can be shown that at these localities lithologies similar to those within the Georgia Complex make up the laterally adjacent and underlying units. This relationship is well developed northeast of Cascade (1659-5642) and east-southeast of Tom Spring (1573-5474).

The structure and lithologic character of three regions of the Georgia Complex are discussed below.

Region 1

The relationship of the Georgia Complex to the overlying and cross-cutting Tom Spring Formation is shown exceedingly well in road cuts in the Cascade area and to the north-northwest. Here the Georgia

Complex is predominantly an olive green to grey shale with locally intercalated calcarenite beds, and minor conglomerate and pebbly mudstone. Intervals of 95-100% shales have a well-developed spaced cleavage and are generally unbedded, though where bedding is visible its attitude is chaotic and inconsistent with that of adjacent units that have not been affected by the emplacement of the Tom Spring Formation. Once intercalated beds of calcarenite, conglomerate and shale now show characteristics of a broken formation in which bedded blocks of various size are encased in a cleaved shale to pebbly mudstone (1649-5462). Where well exposed, bedded blocks are generally aligned parallel to the foliation of the shale. Figure 27 is a series of photographs that show this relation well.

Region 2

A large 035° trending block of the Georgia Complex crops out at the northwestern end of the canyon complex. This mass of the Georgia Complex is well exposed in the Lucea East River and its west-flowing tributaries. It measures approximately 0.4 km by 1.2 km and separates a large body of the Tom Spring Formation from the main outcrop area. This block of Georgia Complex is characterized by northeast trending strikes. Dips are generally in excess of 60° and vary in direction, but locally are as gentle as 7° in the southwest. Within sections of thick sandstone beds strike and dip orientations are relatively consistent. However, more thinly-bedded and pelitic intervals show greater variability in attitude, and continuity in bedding is locally lost.

At the southwestern end of this mass is a 0.3 km long 350° trending

Figure 27 – Outcrop of Georgia Complex, northwest of Cascade. Coherently bedded blocks of turbiditic calcarenite and intercalated shales are encased in disrupted sandstones, shales, boulders of volcanic rocks and minor limestone clasts.

ridge (1571-5500). An exposure, at the southern end, of conglomerate, sandstone and shale with an attitude of 345°, 80°W suggests that this ridge is a structurally-controlled strike range. North and northwest of this ridge the units of the Georgia Complex have a general dip to the northwest and appear to grade laterally into the Mount Peace Formation.

The descriptions above show that rigid body rotation of bedded blocks has occurred within the Georgia Complex. This type of deformation is recorded on both a map and outcrop scale in this region. The abrupt change from a gross northeast trending block to a north-northwest trending one suggests rigid body rotation on a map scale. Within stream sections it can be shown that chaotically disrupted shaly horizons separate more coherent blocks.

This type of deformation is analogous to subaqueous slumping and sliding (Dott, 1963); in which rigid to semi-rigid masses move downslope, with little or no internal deformation, along discrete bounding shear planes. These shear planes would be equivalent to disrupted horizons that separate more coherent blocks. The deformation of the Georgia Complex is not thought to have been driven solely by gravity but also be the rapid emplacement of the Tom Spring Formation. However, primary mass movements, prior to the emplacement of the Tom Spring Formation, associated with the submarine canyon cannot be ruled out as a contributing factor to the deformation of the Georgia Complex. Post-depositional movements for resedimented tephra are also to be expected because of its ability to absorb water (Parsons, 1969).

Region 3

The change in structure from the chaotic Georgia Complex to the gently west dipping Pioneer Formation is well exposed southeast of Tom

Spring (1573-5474) in a westward flowing tributary of the Lucea East River. Here the Georgia Complex consists of a well bedded sequence of 5 to 50 cm thick sandstone beds and intercalated shales and minor mass flow mudstones showing complex folding. Geometry and orientation of folds are not consistent and are typically discontinuous in form, as fold limbs are cross-cut by limbs of other folds. Figure 28 is a sketch of this complex folding. Large scale necking and separation of sandstone beds and infilling of extensional regions with shale was observed (Fig. 29). Horizons of shale and encased dismembered thin sandstone beds within folds may represent primary mass flow events or "plastic" deformation associated with water rich shale horizons, during this penecontemporaneous deformational event.

Somewhat similar structures are developed approximately 1 km downstream in the Lucea East River. Here the Georgia Complex is a sequence of generally thick .1 to >3 m sandstone beds with minor shale and conglomerate. Folds are not as extensively developed and are generally broad open structures. This contrast is probably due to primary differences in rheological properties, e.g. viscosity contrast between beds. Several folds with vertical fold axes were observed. Some of the coarse, massive sandstone beds have irregular nodules of pyrite developed within them. These nodules are similar to those observed within the Pioneer Formation. Both localities of occurrence are proximal to the Tom Spring Formation and may have a genetic relationship.

MOUNT PEACE FORMATION

The Mount Peace Formation is a 500 m ± thick sequence of

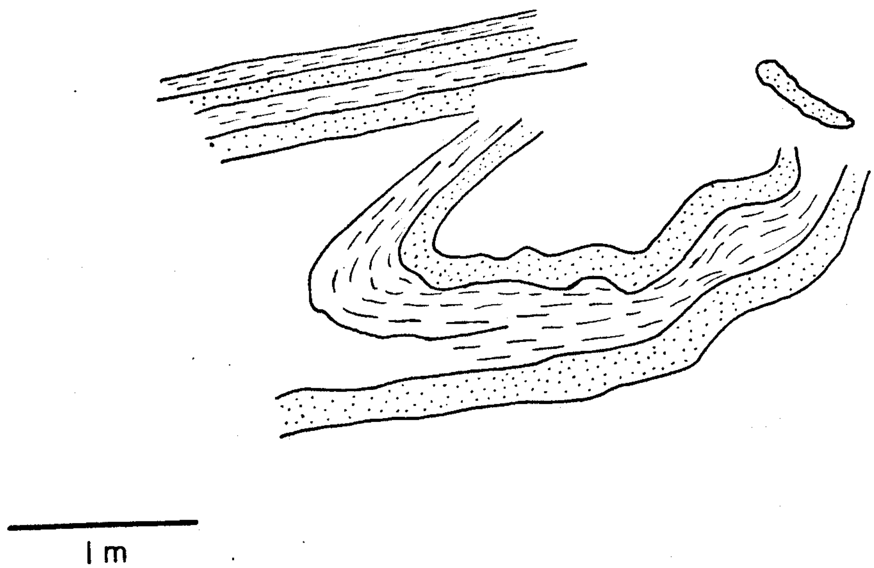


Figure 28 - Outcrop sketch of complex penecontemporaneous folds of the Georgia Complex.

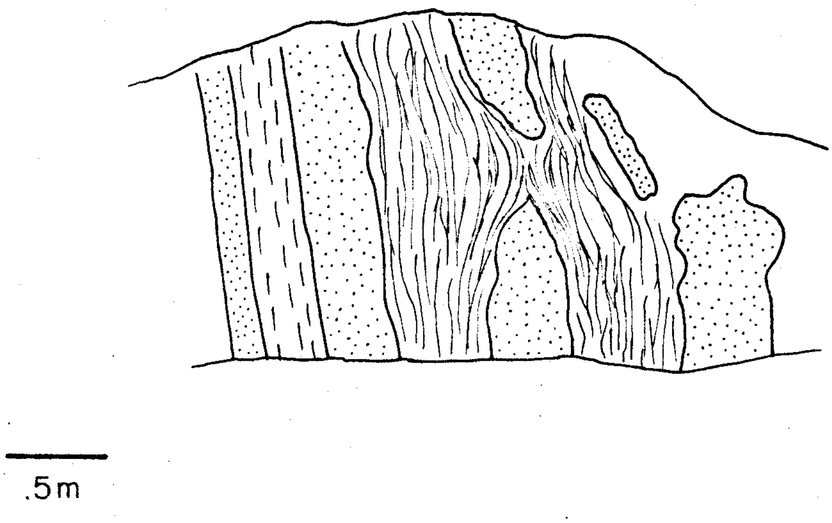


Figure 29 - Outcrop sketch of large penecontemporaneous boudins within the Georgia Complex. Surrounding shale is non-laminated and an anastomosing foliation is developed.

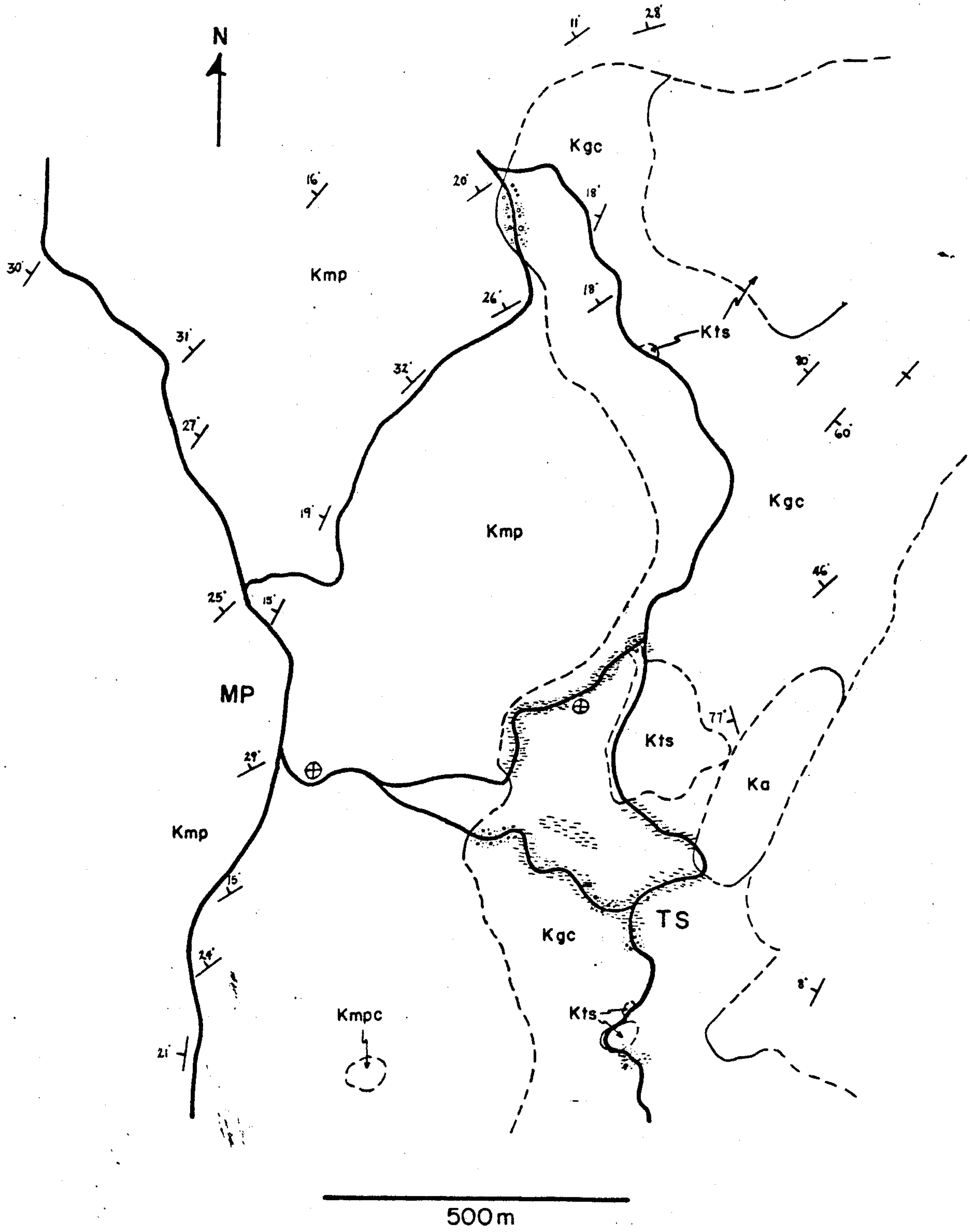
predominantly thinly bedded shales with minor thin ($\leq$ 20 cm) sandstone beds and several lenticular bodies of reefoidal limestone and associated calcarenite beds. The main body of limestone was referred to as the Clifton Limestone by Zans et al. (1962). This study defines limestone bodies as the Clifton Member of the Mount Peace Formation. The Clifton Member will be discussed in more detail below.

The type locality of the Mount Peace Formation is in and around the Mount Peace area (1543-5491) where the unit is semi-continuously exposed in road cuts. Here it generally forms low rolling topography and consists of 95-100% thinly bedded shales that have a well developed spaced cleavage (Fig. 30). Pencils up to 4 cm long have been measured 3 km north of Maryland. Locally, up to 20% of the section consists of fine sandstone beds with Tbc intervals of the Bouma sequence. More commonly sandstone beds are 1 to 2.5 cm thick and only make up a few percent of the section. Locally calcareous concretions and thin (6 cm) limestone beds have been observed. Dispersed streaks of plagioclase crystals and layers with high concentrations were observed north of Mount Peace (1546-5502). Perhaps these local concentrations of plagioclase crystals represent air-fall crystal tuffs. Fresh exposures of the Mount Peace Formation north of Maryland, show it to be dary grey in color which weathers from a brown to rust red.

The lateral transition from the Mount Peace Formation to the Georgia Complex is well displayed in road cut exposures north of Maryland and east-northeast of Mount Peace. Figure 31 is a detailed geologic map of the Mount Peace area. The east-west trending Mount Peace-Tom Spring road shows a fairly abrupt change in facies, from well bedded shales in the east to subangular conglomerates, thin pebbly

Figure 30 – Thinly bedded, cleaved shales of the Mount Peace Formation.

Figure 31 - Detailed geologic map of the Mount Peace (MP) and Tom Spring (TS) area. See text for discussion. Kmp - Mount Peace Formation, Kmpc - Clifton limestone, Kgc Georgia Complex, Kts - Tom Spring Formation and Ka Andesite block.


N
Kgc
Kmp
Kts
Kmp
Kgc
MP
Kts
Ka
Kmp
TS
Kgc
Kts
Kmpc
500m

sandstone beds and intercalated shales, cobbles to bedded blocks (> 40 cm) of dark grey bioclastic limestone (equivalent to the Clifton Member?) within shales and pebbly mudstones of the Georgia Complex. Blocks of angular conglomerate containing fine blue-grey sandstone clasts are set in an unbedded shale matrix. These blue-grey sandstone clasts may be equivalent to sandstones of the Birchs Hill Formation. North of Mount Peace along the northeast trending road sandstone bed thickness (2 - 10 cm) and percent sandstone (0-70%) gradually increases approaching the Geogia Complex. Thin (2-10 cm) sandstone beds show good Ta and Tbc structures and are generally well sorted. These beds may represent overbank deposits associated with the canyon. Once in the Georgia Complex along the north-south segment of this raod the units are characterized by mass flow deposits of sandy cobble conglomerates of medium grey limestone (maximum clast size - 15 cm), and poorly rounded pebble to boulder mudstones.

To the northeast a poorly exposed interbedded sequence of sandstone and shale with local conglomerate and massive speroidal weathering volcaniclastic sandstone beds make up the Mount Peace Formation

The Mount Peace Formation lies both conformably above the submarine canyon complex and is laterally equivalent to it. Its contact with the overlying Harvey River Formation is not continuously exposed but appears to be gradational over a 10 m interval, at the type locality, north of Mount Peace. However, to the northeast its upper contact is not well defined due to a lack of exposure, and loss? of the lithologically distinct basal horizon of the overlying Claremont Formation; which is correlated with the Harvey River Formation. The lower contact with the underlying Pioneer Formation is not well understood due to east-west faulting east of Maryland and to the cross-cutting canyon complex. Therefore, the mapped contact between these two formations is at best approximate.

Clifton Member

The Clifton Member is a bioclastic limestone of variable thickness (generally less than 100 m), and is largely composed of rudists and corals. It is generally believed to be of reef origin (Zans, et al., 1962 and Coates, 1977) and forms lenticular bodies that crop out in the Clifton (1537-5442) and Maryland (1575-5425) areas. Horizons of calcareous shales and intercalated 10-60 cm thick calcarenite beds occur around the edges of the major lenses of the Clifton limestone.

Chubb (in Zans, et al., 1962) had assigned a Lower Campanian age to the Clifton limestone. Kauffman (1966) later collected Inoceramus balticus cf. subsp. kunimiensis and I. mulleri from horizons stratigraphically above the Clifton limestone and suggest a minimum very Early Campanian age for the Clifton limestone. Since then I have collected the Inoceramid Cordiceramus sp. cf. C. Mulleri (Petrascheck) (identified by E. Kauffman) of Upper Santonian and Santonian-Campanian passage beds, from the uppermost horizons of the main Clifton limestone body (1526-5424). Fossils collected from the overlying Harvey River and Claremont Formation (Harvey River is stratigraphically equivalent to the Claremont Formation) give a Lower Campanian to Upper Santonian age. It seems that the Clifton limestone is Upper Santonian to lowest Campanian in age.

The map accompanying this report shows for the first time an isolated(?) body of Clifton limestone capping a hill west-southwest

(1564-5393) of Maryland. Other small lenses of the Clifton limestone shown on the map, along the west side of the Lucea River valley, were not visited during this study and are taken from unpublished maps of the Jamaican Geological Survey.

Approximately .4 km west of Maryland an east-flowing tributary of the Lucea East River (1568-5395, where road crosses stream) exposes a continuous section of the Clifton limestone. The section consists of massive, thick-bedded bioclastic limestone except for a 5 m thick horizon of oomicrite that occurs near the top of the sequence. More resistant, subrounded, boulders (.7 m max.) of calcirudite are enbedded within it (Fig. 32). This relationship suggests that processes of resedimentation were operative for this horizon of the Clifton limestone. To my best knowledge no one has ever reported the occurrence of oolitic limestone within the Clifton limestone. Also near the base of the exposed section a dark-grey, fine-grained andesitic lava (flow?) occurs and rests on a limestone conglomerate. Stream boulders of similar volcanic lithology occur above this lens of Clifton and suggest that they have been derived from lower units of the Mount Peace or Tobolski Formation.

CLAREMONT FORMATION

The Claremont Formation crops out in the northeastern corner of the mapped area, east of Lucea Harbour. The unit has a gross east-west strike and a gentle to moderate (10°-30°) northerly dip. In the west the Claremont Formation is a sequence of channelized conglomeratic sandstone and minor shale. Farther to the east, near Mosquito Cove, it is dominated by shale and minor thin-bedded sandstone. The nature of the contact between these two facies is not well known, but several traverses along strike suggest a gradational to fairly abrupt contact. This contact also appears to migrate in a northwesterly direction at higher stratigraphic intervals. The type sections of the conglomerate and shale facies are well exposed along the Jericho-Paradise road and north of Elgin Town (1504-5620), as well as along the southern and western sides of Mosquito Cove, respectively. The Claremont Formation is approximately 700 ±200 m thick.

The basal 100 ±25 m of the Claremont Formation is lithologically distinct from the rest of the formation and is well exposed in road and stream sections around Dundee (1537-5554). The section consists of poorly bedded calcareous shales and siltstones that locally contain concretionary-type masses of calcarenite (Fig. 33). Minor sandy, pebble to boulder conglomerate and massive 1 m + coarse sandstone beds were observed at one locality (1567-5562). Fossiliferous horizons containing the inoceramids Platyeramus rhomboides (Sietz) subsp. rhomboides and Endocostea balticus (Bohm) aff. E.B. toyajoanus (Nago and Matsumoto) (identification by E. Kauffman) of Upper Santonian to Lower Campanian age were found within a stream bed (1550-5562) north of Dundee.

The eastern extension of this basal lithology was not observed. Therefore the contact of the Claremont Formation and underlying Mount Peace Formation in the east is approximate. The western conglomeratic and shaly eastern sections of the Claremont Formation are discussed separately below.

The conglomeratic section is characterized by rugged topography

Figure 32 – Massive oomicrite horizon within the Clifton Limestone. Subrounded more resistant boulders of calcirudite are also present. Hammer, bottom-center, rests on a calcirudite clast.

Figure 33 – Poorly bedded calcareous shale and siltstone of the basal horizon of the Claremont Formation. Note concretionary-type mass, left of hammer.

with short, generally dry, steep walled north draining valleys. Exposures across strike along the Jericho-Paradise road and north of Elgin Town (1500-5620) show a continuous fining and thinning upward sequence of conglomerates and coarse, lithic sandstones to sandstones and intercalated shales and minor conglomerate. The lower 600 ±100 m of this section consists of .5-3 m + channelized well-subrounded pebble to boulder conglomerates (Fig. 34). Matrix of the conglomerate is invariably a medium to coarse grained sand. Only a few clasts of bioclastic limestone were observed and all others were a variety of mafic volcanic rocks. Large (2 by .5 m) blocks of coarse, bedded sandstone embedded in conglomerates have been observed (1572-5572). The contact between conglomerate and encased sandstone blocks are very sharp. Locally, grading is developed in the conglomerates. Some $\lesssim$ 1 m coarse sandstone beds are faintly laminated and have pebble-boulder size clasts floating in a sandy matrix (Fig. 35). Similar facies are exposed to the west along the Elgin Town (1504-5620) - Claremont road. An interesting thin-bedded, 3-5 m thick, horizon of sandstone and shale was observed at two localities (1524-5618, 1540-5624, approximately .3 km apart). The sandstone beds are fine to medium grained and range in thickness from 1 to 15 cm but average 4 cm. Sandstone beds have sharp upper and lower contacts and undulate on the order of $\lesssim$ 1 cm (Fig. 36). Upper surfaces of sandstone beds show worm burrows. This horizon of thinly bedded sediments has a very complicated and locally interfingering, disrupted contact with adjacent conglomerates and thick sandstone beds. Amalgamation of thick massive sandstone beds is common but is particularly well displayed at one locality (1524-5616) west of Claremont (see Fig. 37).

Figure 34 – Channelled, well rounded, unorganized, pebble to boulder conglomerate of the Claremont Formation. Note sharp unabrasive, basal contact of upper conglomerate bed.

Figure 35 – Massive, coarse sandstone bed of the Claremont Formation. Note "free floating" well rounded volcanic clast adjacent to hammer.

Figure 36 – Thinly bedded fine to medium grained sandstone and intercalated shale of the Claremont Formation.

Figure 37 – Photographs are of the same stratigraphic interval and are separated by ~2m. Note contrast in number of beds within each photograph.

Large exposures to the west, north of Elgin Town, expose generally more thinly (2-70 cm) bedded channeled and lensing turbiditic medium to coarse sandstone beds and intervening shales. Uncommon thick (1 m) pebbly sandstones occur. Figure 37A shows the general outcrop character of this facies in this case showing complex penecontemporaneous slump folds.

Approximately a kilometer or less to the east a sequence of greenish and reddish shales and interbedded sandstone (1-45 cm thick, but averaging 2-5 cm thick) and minor conglomerate is laterally equivalent to the conglomeratic-sandstone section described above. Topography within this shaly section is noticeably less rugged. Shale and siltstone generally makes up to 95-100 percent of the section, but, locally, near the base, thin, (4 cm average) graded, parallel laminated and cross-stratified sandstone beds makes up to 70% of the section. Mass flow deposits were observed near the base and midway (up to 6 m thick, 1650-5615) up the section. Mafic volcanic clasts, bedded blocks of underlying reddish shales (1609-5584) and calcarenite beds were observed within a generally, poorly sorted volcaniclastic matrix. A poorly exposed 2 m thick sandy boulder conglomerate bed was observed within a shale sequence (1658-5613). Many of the thin sandstones around Mosquito Cove have sharp upper and lower contacts and show evidence of channeling.

HARVEY RIVER FORMATION

The Harvey River Formation, an approximately 250 m thick unit of interbedded turbiditic fine to coarse sandstone and greenish-grey shale (Fig. 38). It conformably overlies the Mount Peace Formation

Figure 37A – Synsedimentary slumping within the more thinly bedded horizons of the Claremont Formation.

Figure 38 – Outcrop of medium bedded classical Bouma type turbidites and intercalated shale of the Harvey River Formation.

and has a fairly abrupt contact with overlying basal shales of the Middlesex Formation (1490-5475). The Harvey River Formation crops out as a north-south trending band that is fault bounded on the north (see Plate 1) by the Fat Hog Quarter Fault.

The basal 75-100 m is a distinct lithology of poorly bedded hard, olive-grey calcsiltite and interbedded coarse tuffaceous sandstone beds (10-30 cm). This lithology is well exposed north of Mount Peace (1526-5517) and can be seen in small exposures east of Harvey River. This relatively resistant lithology is expressed by a north-south trending strike ridge. Above this basal lithology a fossil of the Inoceramid _Endocostea balticus marki_ (Giers) (identified by E. Kauffman), of Upper Santonian to lowest Campanian in age, was found (1524-5536). At a similar stratigraphic level, but farther south, the Inoceramid _Plt-yeeremus_ cycloides n. subsp. cf. _P. ahsenensis_ Seitz (identified by E. Kauffman) of Upper Santonian to Lower Campanian in age, was found in a loose boulder (1525-5465). This is the first report of these two forms in the Caribbean. These inoceramids occur within a sequence of calcareous shales and intercalated calcarenite beds. It could not be conclusively ascertained from either of these fossil localities whether the fossils were _in situ_. However, an equivalent stratigraphic horizon within the Claremont Formation contains _in situ_ inoceramids of a similar age. This would then suggest that the basal portion of the Harvey River Formation was deposited in shallow water, possibly a shelf environment.

Sandstone beds average approximately 5 to 10 cm in thickness but range up to 40 to 70 cm in thickness. However, massive, coarse sand to pebbly conglomerate beds from 1 to 2 m+ thick beds that occur sporadically throughout the upper two-thirds of the section. East of Harvey River several of these thick beds are dominantly composed of angular volcanic lithic and crystal (plagioclase and clinopyroxene) fragments. Convolute and parallel laminated (Tc?) are extensively developed in the thinner sandstone beds (5-20 cm). Grading is locally developed. Lensing of sandstone beds in excess of 30 cm was observed in outcrops that had significant lateral extent. The percentage of sandstone varies from 90% to 10% locally, but averages approximately 40-50% for the entire section. Thin (4-8 cm) limestone and calcarenite beds are minor but occur throughout the section. Shales with randomly dispersed sand lentils were observed and are similar to features recorded within the Dias Formation (see p.73).

Mass movement deposits are not extensively developed but have been observed at three localities and range from slumping of reddish-grey to purplish silty shale beds (1525-5521) to unbedded shale with discontinuous sandstone lenses.

The Harvey River Formation is thought to be a lateral facies of the Claremont Formation. However, their lateral continuity is obscured by sinistral slip along the Fat Hog Quarter Fault. Similarities in lithology, bedding characteristics and fauna of the basal portions of these formations suggests that they are stratigraphically equivalent.

MIDDLESEX FORMATION

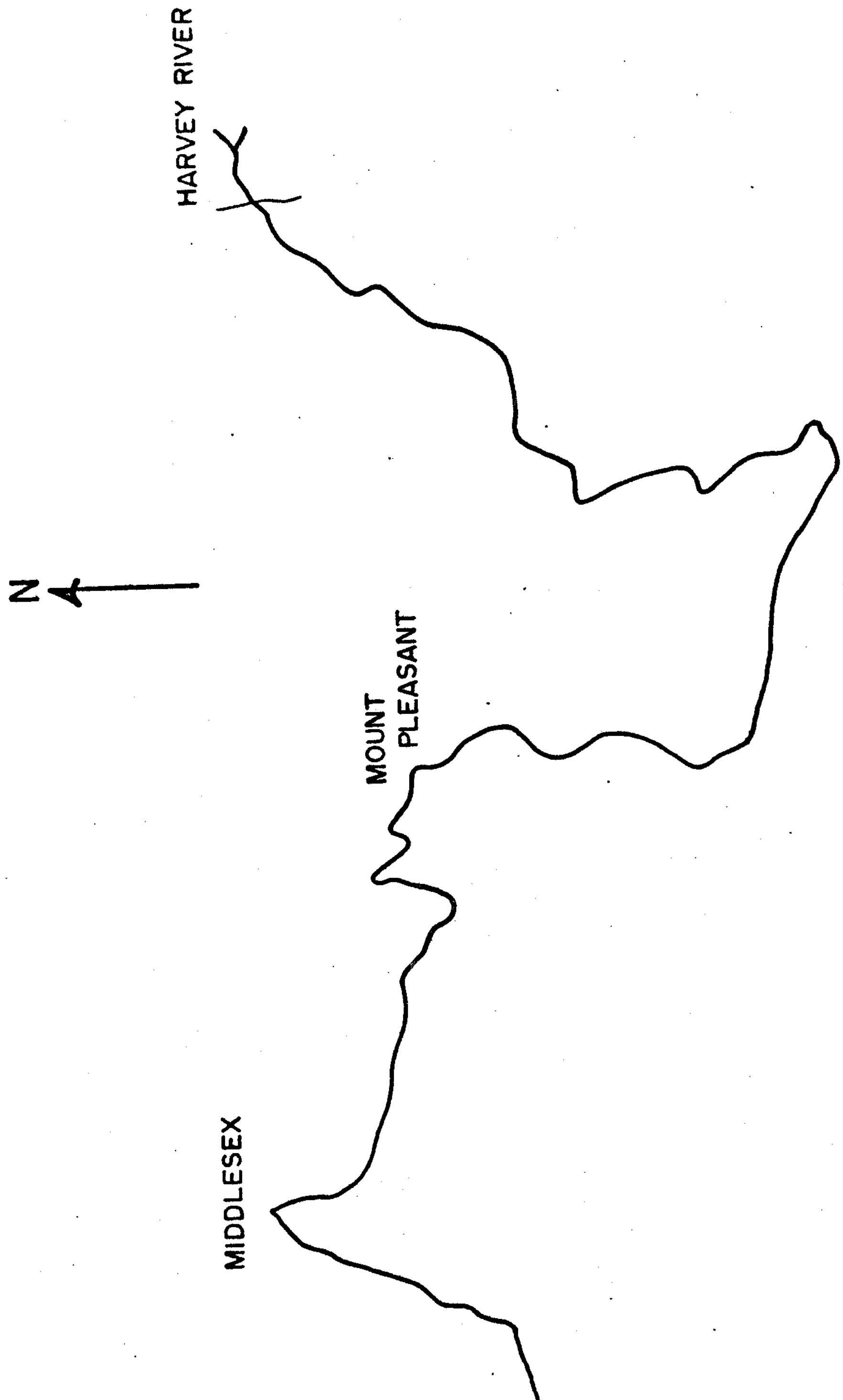
The Middlesex Formation is an approximately 1200 ±100 m thick sequence of interbedded sandstone and shale. Sandstone commonly makes up to 50% of the unit but locally is non-existent. A variety of sediment gravity flow deposits also occur throughout the section. The formation crops out as a north-south trending band that is bounded on the north by the Fat Hog Quarter Fault zone and south by the Maryland Fault zone (see Plate 1). Bedding has a relatively consistent north-northwesterly strike and gentle (5°-25°) westerly dip.

The formation is well to continuously exposed in road cuts from Dias eastward to Harvey River (Fig. 39). The section exposed in these roadcuts is taken as the type section of the Middlesex Formation. Descriptions that follow are primarily based on this section.

The basal 170 ±20 m of the Middlesex Formation is composed of 80-100% thinly bedded, greenish-grey weathering shales and siltstone, and interbedded sandstone ($\lesssim$ 9 cm thick) and limestone beds. Parallel laminated siltstone and graded sandstone beds are common. Locally, some 2 cm thick sandstone beds have plagioclase crystals concentrated in their lower 2 mm of the bed. Horizons of red and interbedded red and green shales were observed northwest (1476-5492) and southwest (1472-5424) of Harvey River. At the top of this basal lithology, a horizon of slumped shale is well exposed southwest of Harvey River (1472-1460) localized zones of flowage were observed. This basal pelitic interval is distinct from the more thickly bedded sandstone and shale sequence of the underlying Harvey River Formation.

Above this basal lithology is a 410 ±25 m thick sequence of interbedded sandstone (averaging ~50%) and shale and a variety of channelized slide conglomerates. Sandstone beds are invariably less than 30 cm thick and are locally cross-stratified and parallel laminated. Couplets (2-8 cm thick) of fine to coarse sandstone and overlying silty shale are common to this and the overlying interval

Figure 39 - Road section along which the type section of the Middlesex Formation is exposed. See Plate 1 for location.


HARVEY RIVER
N
MOUNT
PLEASANT
MIDDLESEX
DIAS

of the Middlesex Formation. Within these couplets the shale-sand contact is generally very sharp (Fig. 40). Conglomerates are usually characterized by sub to well rounded clasts of limestone, volcanic rocks, rip ups of reddish-brown pelite and bedded blocks of underlying and laterally equivalent sediment (Fig. 41). The conglomerates are unorganized and have a coarse sand to pelitic matrix.

The above interval is separated from the rest of the overlying Middlesex Formation by a fault zone (1420-5456) that trends 000° and has a 65° easterly dip. The amount and/or type of offset along this fault is not known; therefore, the extent to which it repeats or omits units within this section is not known. Above this fault zone there is an approximately 130 m thick sequence of thinly bedded grey shale and minor (1-10%) 1 to 5 cm thick fine to medium sandstone beds. Small rip up clasts, channeling, grading and parallel laminations are developed within couplets of sandstone and shale (see Fig. 40).

In sharp contact with the above described pelitic interval is the uppermost 325 ±25 m of the Middlesex Formation which consists of an interbedded sequence of sandstone (averaging 50-60%) and grey-green shale and a variety of mass flow deposits. Sandstone beds are generally regularly bedded and range in thickness from 2 to 20 cm (Fig. 42), though they become less regularly bedded near the top of the section. Rare, massive, coarse 1 m thick sandstone beds are also developed near the top of the section. Sandstone beds are commonly channeled, graded and parallel laminated. Mass flow deposits range from slumped shale and intercalated sandstone sequences (1396-5481) to sandy conglomerates. Mafic volcanic rocks, blocks of underlying and adjacent bedded sediments and minor limestone are the dominant clast types. Figure 43 is

Figure 40 – Thin bedded sandstone and shale of the Middlesex Formation. Note sharp upper and lower contacts of the fine sandstone bed. Also note thin, discontinuous laminae of sandstone immediately below thicker sandstone bed.

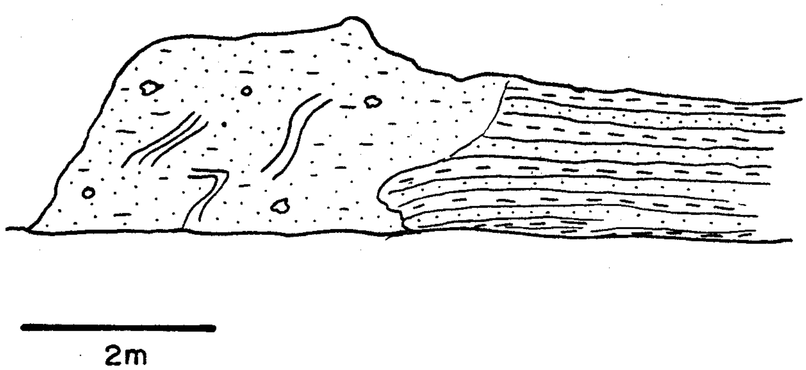


Figure 41 – Outcrop sketch of channeled mass flow deposit and laterally equivalent sandstone and shale of the Middlesex Formation.

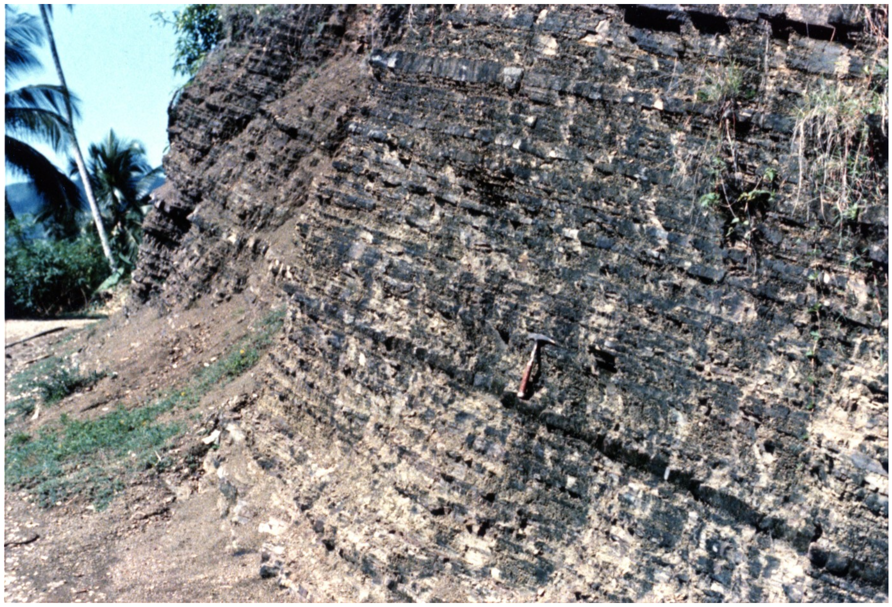


Figure 42 – Well bedded sandstone and shale of the upper horizons of the Middlesex Formation. Note slight thickening and thinning of individual sandstone beds.

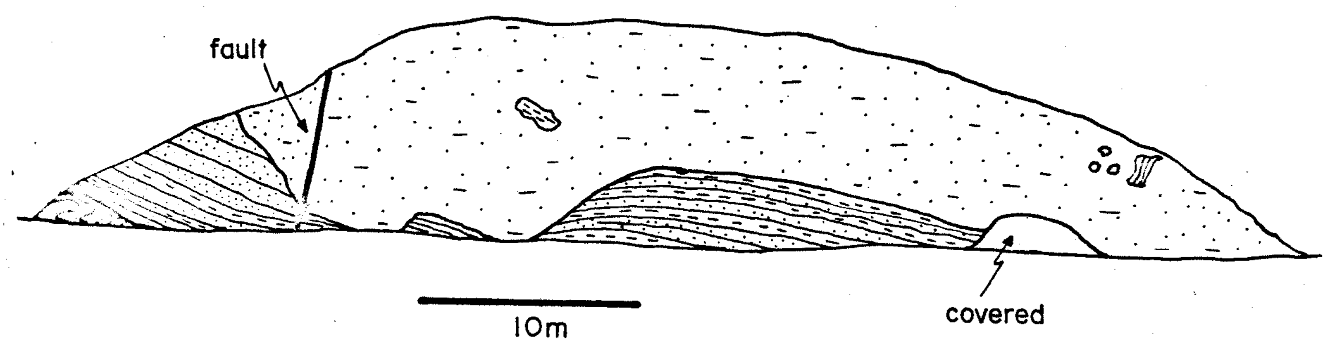


Figure 43 – Outcrop sketch of large channeled mass flow deposit of the Middlesex Formation. Note that the mass flow deposit contains few conglomeratic clasts and is dominantly made up of a heterogeneous mixture of clay and sand sized grains.

an outcrop sketch of a well exposed mass flow deposit and underlying bedded sediments.

The upper contact of the Middlesex Formation with the overlying Dias Formation is marked by a fairly abrupt change to 95 to 100% deeply weathered shales of the Dias Formation.

DIAS FORMATION

The Dias Formation is a 370 ±20 m thick, thinly bedded sequence of greenish-grey shale and siltstone, with interbedded, generally lensing sandstone beds. Minor conglomerate and slumping appear to be restricted to the northern end of the outcrop area. The formation crops out in a north-south trending band in the western part of the field area. Topographically, it is expressed by low-lying, steep-sided topography. Exposure is generally poor (particularly south and west of Dias) and outcrops are often deeply weathered to variegated hues of rust-brown (e.g., around the Kingsvale area, Fig. 44).

The shale and siltstone component generally make up 90 to 100% of the section. However, west of Cacoon (1330-5490) the proportion, thickness and coarseness of sandstone beds increase. Sandstone locally makes up to 100% of the section, beds are up to a meter in thickness and their grain size is that of medium to pebbly sandstone.

Characteristically, the sandstones within the Dias Formation are lensoid to lenticular in nature; however, thin, evenly bedded, graded sandstones do occur. Medium to coarse sandstone may occur as irregular discontinuous 0.5-2 cm to 6-8 cm thick beds. Fine sand laminae-lenses, on the order of 0.5-1 cm long and 1 mm thick were commonly observed west of Kingsvale. Some of these lenses are composed of weathered ?plagioclase crystals.

Conglomerates have been observed at several localities within the Dias Formation and appear to be restricted to the northern segment of the outcrop area. An approximately 3 m thick structureless, angular to rounded sandy boulder (max. clast size-35 cm) conglomerate with less cobbles and pebbles is well exposed (1385-5480) northwest of Greenland. Sub to well rounded volcanic rocks make up to 95% of the clasts while roughly 5% are angular to rounded bioclastic limestone. A few bedded clots of sandy mudstone were also observed. The sandy matrix makes up to 60% of the conglomerate. Thinner conglomerate beds occur both above and below. Farther up in the section (1344-5550) north of Richmond, several conglomerate beds were observed. Here a sub- to well-rounded sandy cobble (4 cm) conglomerate bed with shale clots has a sharp non-abraded contact with underlying shales (Fig. 45). Most clasts are grey and red fine-grained volcanic rocks though some have phenocrysts of plagioclase. Conglomerates also occur immediately downsection though they are poorly exposed and involved in east-west faulting associated with the Fat Hog Quarter Fault zone (see p. 98).

Northwest of Greenland a distinct 70 m horizon of massively bedded terra cotta and minor green colored shale occurs stratigraphically below a slumped horizon of shale and intercalated thin (2-15 cm) sandstone beds. Bedding within the slumped zone is generally preserved, however, locally, cohesion and bedding are lost. Figure 46 is a sketch of a possible recumbent anticline. Overturning is to the west, 250°, several boulders of volcanic rocks were also observed

Figure 44 – Deeply weathered shales of the Dias Formation.

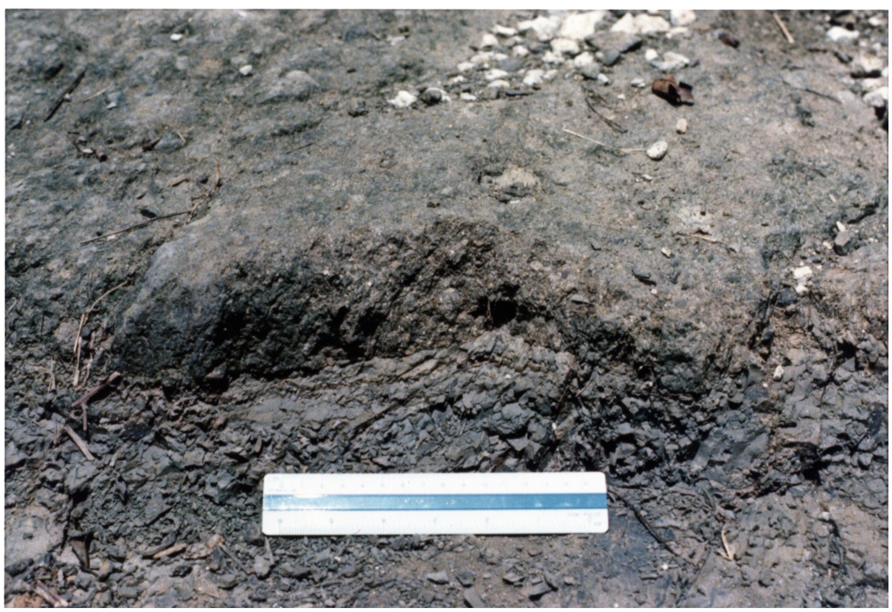


Figure 45 – Close-up of unabraded, sharp bottomed mass flow deposit and underlying shales of the Dias Formation.

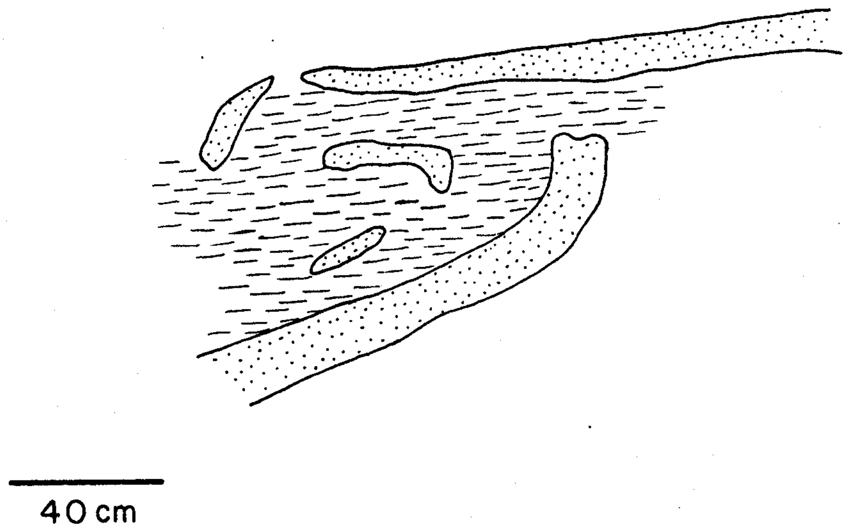


Figure 46 - Outcrop sketch of a synsedimentary (?) recumbent antiform within the Dias Formation.

within the slumped shales.

The upper contact of the Dias Formation is not defined in this study.

HAUGHTON COURT FORMATION

The Haughton Court Formation crops out in the northwest corner of the mapped area, west, southwest and south of Lucea Harbour. Structurally, it defines an east-west trending anticline. The formation is separated from the Dias, Middlesex and Harvey River Formations to the south by the Fat Hog Quarter Fault zone and from the Claremont Formation, to the east, by north-south trending faults. The relationship between the Haughton Court and Claremont Formation across Lucea Harbour is not understood. The stratigraphy, thickness and lithostratigraphic positioning of the Haughton Court Formation with respect to neighbouring lithounits is not clear due to extensive faulting and hence no type section is defined.

The Haughton Court Formation is dominantly shale, shale and interbedded thin sandstone and minor, though locally important, conglomerate. The structure and therefore lithostratigraphy are apparently coherent, southwest of Lucea, within the southern limb of the anticline. Figure 47 is a stratigraphic column for this section. A similar succession from the conglomerate horizon (see Fig. 47) is observed south (1518-5567) of Kew (1507-5595). Near the base of this section, within the Lances River (1375-5567), a fresh ~8 cm thick bed of graded plagioclase crystals was observed (Fig. 48). Petrographically, the bed is composed of lath-shaped plagioclase crystals (95%) and volcanic and quartz lithic fragments. Carbonate matrix is restricted to upper 1 cm

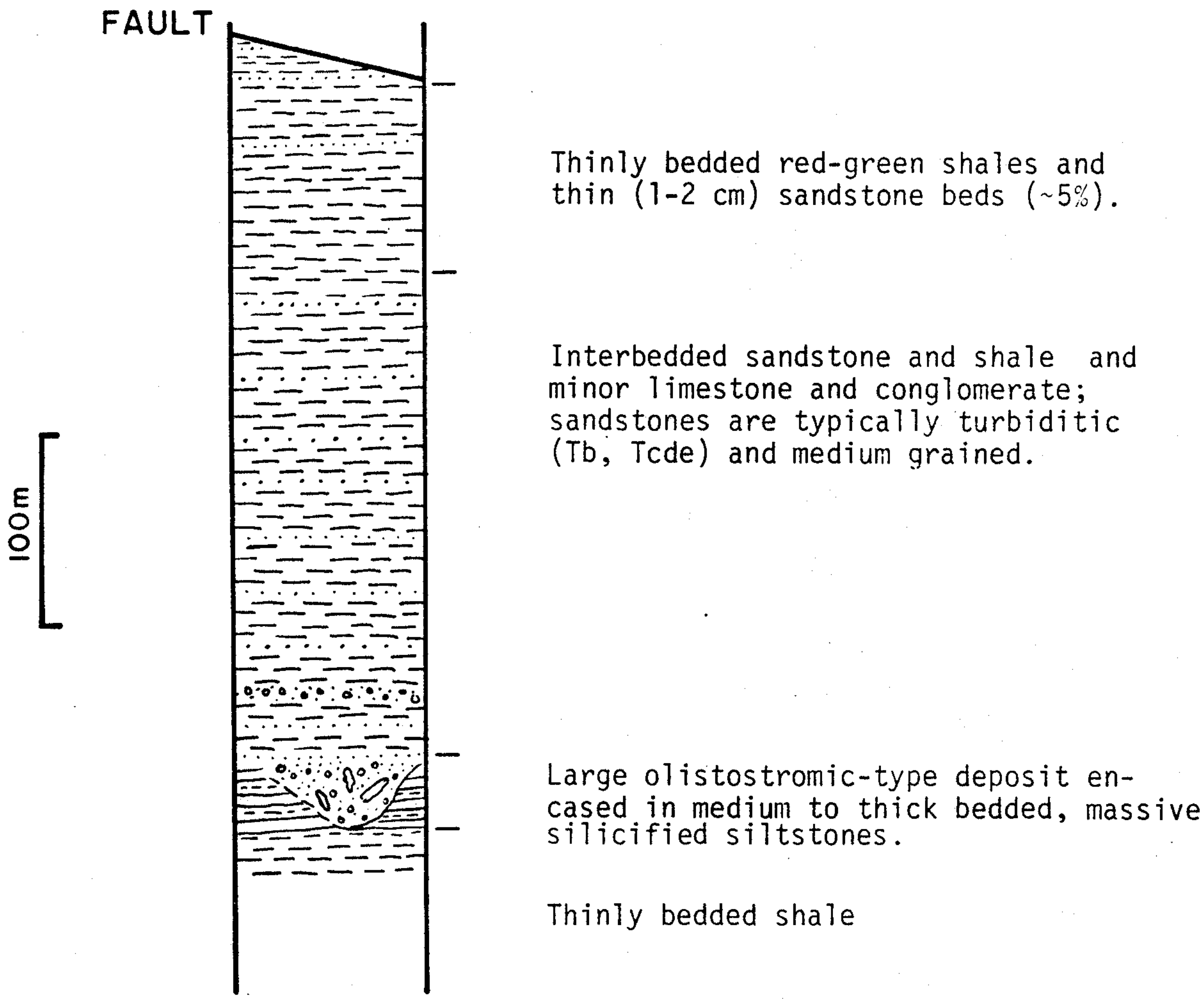


Figure 47 - Stratigraphic column of the south-central area of the Haughton Court Formation.

of the bed, making up to 40% of the rock. This and other localities with concentrations of plagioclase crystals attest to the predominance of plagioclase within island arc volcanics.

The thick (45 m +) slide conglomerate of Figure 47 is well exposed in a 20 to 30 m high outcrop, 1.3 km (1434-5568) south of Lucea. It consists of variable-sized bedded blocks of shale and minor sandstone and conglomerate. Clasts of bioclastic limestone (25%) and mafic volcanic rocks (75%) and rip-ups of reddish silty shale are set in a poorly sorted coarse tuffaceous matrix of fine unresolvable detritus, corroded volcanic lithic and bioclastic limestone fragments, altered plagioclase crystals and minor clinopyroxene (Fig. 49). From outcrop relations it can be shown that this olistostrome type deposit is channeled into adjacent shales and calcareous silty sandstone beds. On the basis of lithologic succession this olistostrome is observed at three localities to the east (1518-5567, 1450-5564 and 1454-5563) in which the latter two are offset .2 km left-laterally by a north-south trending fault (see Plate 1). Other slide conglomerates within the Lucea Harbour area have similar characteristics as described above, though are generally not as well exposed. Curiously, in the vicinity of Lucea Harbour many of the conglomerates have a ?hydrothermally indurated character. Many of these conglomeratic horizons are invariably overlain by classical Bouma type turbidites and intercalated shales.

Bateson (1974) defined several bodies of agglomeratic rocks within the area that is here defined as the Haughton Court Formation. These bodies were visited and are interpreted as slide conglomerates that were deposited within a marine environment. Many exposures appear to have been hydrothermally altered possibly giving the appearance of

Figure 48 – Close-up of a resedimented graded, plagioclase, crystal tuff.

Figure 49 – Close-up of poorly sorted matrix and clasts of a 45m.+ thick slide conglomerate of the Haughton Court Formation. Note indentation of reddish silty shale rip ups by clasts.

in situ volcanic rock.

The northern limb of the anticline, defined by the Haughton Court Formation, consists predominantly of shale and lesser amounts of interbedded fine to coarse sandstone and local conglomerate. Sandstone beds are invariably thin, generally less than 20 cm. Conglomeratic horizons were observed in the western end of the area and typically they do not have a hydrothermally indurated appearance.

STRATIGRAPHIC CORRELATION OF THE LUCEA INLIER WITH OTHER CRETACEOUS INLIERS OF JAMAICA

Several fossil localities of inoceramids and rudists date the Lucea Inlier as Upper Santonian to Lower Campanian in age. Fossils have generally been retrieved from the middle of the 4 km + thick sequence; therefore, the Upper Santonian to Lower Campanian age interval is a minimum age span. In terms of age, units within the Lucea Inlier are correlatable with the neighbouring Sunderland Inlier and the more distant St. Ann and Central Inliers. Other neighbouring inliers, e.g. Marchmont, Jurusalem Mountain and Green Island are demonstrably younger in age, Upper Campanian to Maastrichthian age (Zans, et al., 1962; Coates, 1977). The Grange Inlier lies immediately southeast of the Lucea Inlier, though its age is not known.

Central Inlier

The Peters Hill Formation, of Late Santonian to Early Campanian age (Kauffman, 1966), is a siltstone-mudstone sequence with subordinate sandstone and limestone. Volcanics and volcaniclastics of the Authors Seat and Bull Head Formations underly and overly the Peters Hill

Formation, respectively (Robinson and Lewis, 1970).

St. Ann Inlier

On the basis of inoceramid bivalves the Inoceramus Shales (Chubb, in Zans, et al., 1962) of the St. Ann Inlier are of Santonian to Lower Campanian age; however, planktonic foraminifera of Late Coniacian to Lower Santonian age have also been collected from the Inoceramus Shales (Esker, 1968, in Giles, 1977).

Sunderland Inlier

East of the Lucea Inlier, the Sunderland Inlier of the parish of St. James exposes a Santonian to Maastrichthian sequence. Inoceramids of Santonian to Lower Campanian age have been recovered from the Sunderland Shale (Kauffman, 1977). The basal Johns Hall Conglomerate underlies the Sunderland Shales. Reconnaissance work by the author and K. Burke suggests that clasts, averaging 30 cm within the Johns Hall Conglomerate, are significantly larger than ones observed within the Lucea Inlier. The matrix of conglomerates was locally observed to be composed of 80 to 90% plagioclase. The Johns Hall Conglomerate may represent a similar influx of conglomerate as does the Pioneer Formation of the Lucea Inlier. However, in outcrop the Johns Hall Conglomerate appears similar to ?hydrothermally indurated conglomerates of the Haughton Court Formation.

The Upper Santonian to Lower Campanian horizon of the Lucea, Central, St. Ann and Sunderland Inliers is dominated by pelite. This stratigraphic position may represent a period of reduced vulcanicity or it may just indicate the lithologic character of fossiliferous horizons. Conglomeratic horizons below and/or above the pelitic interval are common in all of the inliers discussed except the St. Ann. These conglomeratic intervals probably represent volcanically active periods.

CHAPTER III

STRUCTURE

Introduction

The Lucea Inlier consists of a simply folded sequence of Upper Cretaceous volcaniclastic rocks. Folds have a general east-west trend. High angle faulting has affected both the Cretaceous and unconformably overlying Cenozoic sequences. The Inlier is dominated by major east-west trending fault zones that separate the mapped area into three blocks. From north to south the blocks will be referred to as the northern, central and southern blocks. The northern block is separated from the central by the Fat Hog Quarter Fault zone, and the Maryland Fault zone separates the central block from the southern. It is not clear at present whether each block is continuously fault bounded (see Plates 1 and 2). However, it is believed that the stratigraphy of each block can be tied into that of the neighboring block.

The structure of the southern block is dominated by the relatively tight, westerly plunging Cash Hill Anticline. North of the Maryland Fault zone the style of folding is very different. It is dominated by a broad, open westerly-plunging anticline with a well developed axial planar spaced cleavage. A small, doubly plunging, east-west trending anticline and tight, easterly closing syncline, to the south, occupy the northwestern end of the northern block. Mesoscopic folds related to the major folding affecting the Cretaceous units were observed at only one locality (1573-5474). Other folds of tectonic origin have been observed, but these are thought to be related to compression along faults.

The deformational event that formed the east-west trending folds within the Cretaceous units is thought to be of Late Cretaceous to Paleocene or older Eocene age as it affects all Cretaceous rocks in the inlier and none of the Cenozoic rocks.

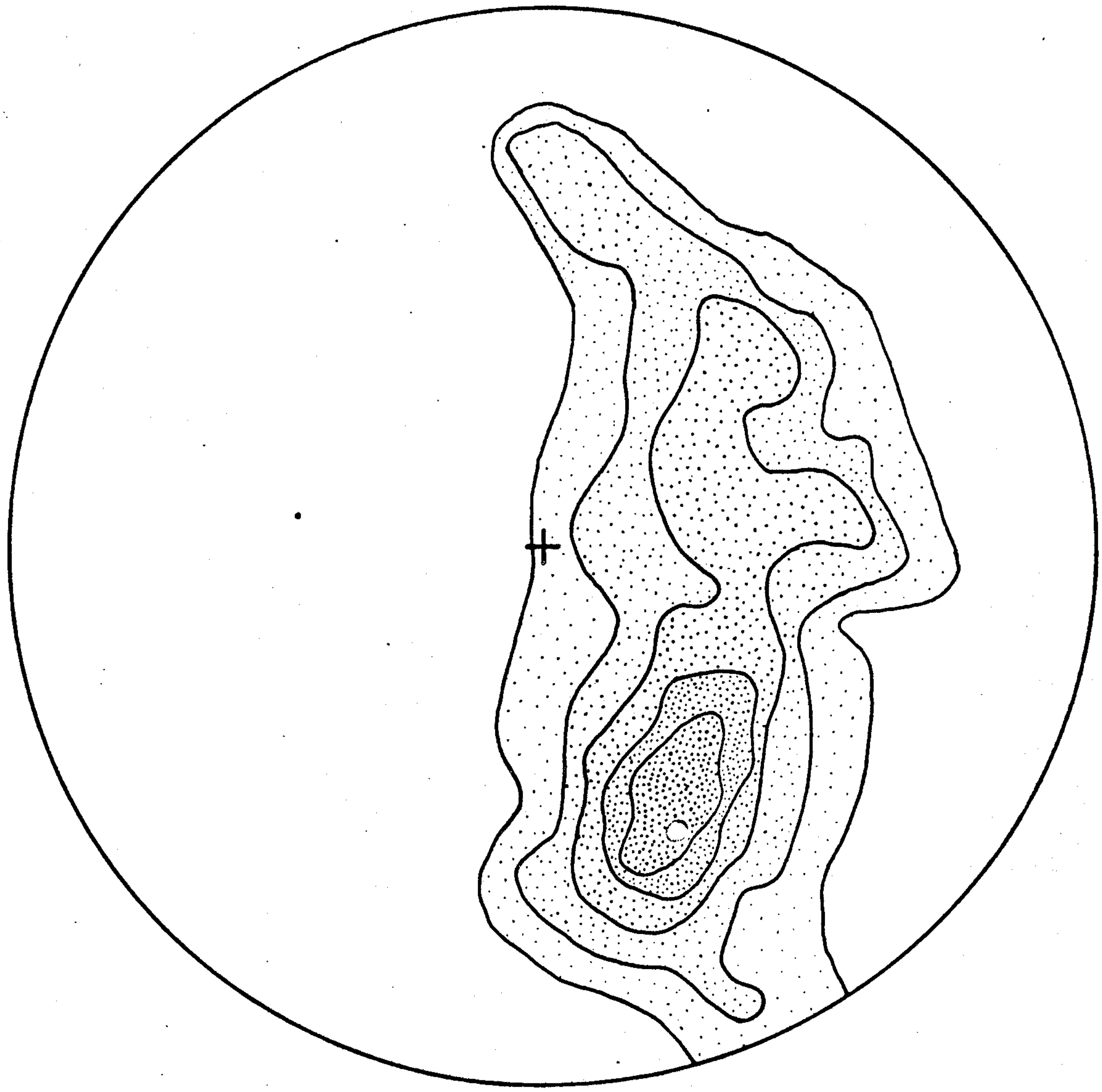
FOLDING

The fold pattern is relatively simple with axes trending approximately east-west. The Inlier is sliced into three major blocks of which the southernmost one has a fold style distinctly different from that of the central and northern blocks.

The Cash Hill Anticline of the southern block is a relatively tight westerly plunging anticline. Figure 50 is a plot of poles to bedding. The Cash Hill Anticline is one of the few, and perhaps the only, structure that is simply related to topography within the Lucea Inlier. Examination of Plate 1 reveals that its structure and lithologic succession control the spatial distribution of ridges and valleys. The style of folding north of the Maryland Fault zone, within the central and northern blocks, is generally much more open. Only the core region and northern limb of a large westerly plunging anticline are presently exposed. This first order interpretation of the structure is similar to Bateson's (1974) conclusions. Only a portion of the structural data for the northern limb, east of Lucea Harbour, is represented on Plate 1 because of problems with reproduction. Data that was omitted is essentially similar to the data that is included on Plate 1. The northwestern corner of this simple anticline is complicated by a small, doubly plunging, parasitic, anticline. This

Figure 50- Lower hemisphere. Stereographic projection of poles to bedding for the Cash Hill Anticline.

116 MEASUREMENTS



2 5 10 15 20 25

POINTS PER 1% AREA

structure crops out west of Lucea Harbour. The structural transition from this small anticline to the simple east-west trending limb east of Lucea Harbour may explain the location of the near circular Lucea Harbour (Fig. 51). Lucea Harbour is interpreted as a structural depression. This structural control was probably the impetus for location and form of the Lucea Harbour but has since been modified by erosion, deposition and reef construction. Also shown in Figure 51 is a tight faulted syncline south of the doubly plunging anticline. A close examination of structural data on Plate 1 will support this interpretation.

Thin ($\leq$ 1 cm) calcite veins parallel to bedding have been observed to be folded (1644-5613). Axial traces of folds have a range of 280°-310°, (6 measured). Slight overturning of folds is toward the north-northeast (Fig. 52).

CLEAVAGE

A very consistent (Fig. 53) axial planar cleavage is associated with the sequence of Cretaceous sediments north of the Maryland Fault zone. It is a spaced vertical cleavage that strikes east-west and is restricted to shales and fine siltstones (Fig. 54). The spacing interval of the cleavage is dependent upon grain size. Pencils, formed by the intersection of cleavage and bedding are up to 4 cm long (1572-5424). Weathered and cleaved shales locally crumble into very fine needles ($\leq$ 5 mm). No microstructural studies have been made on the cleavage.

No cleavage was observed to the south within the Cash Hill Anticline. It is possible that units within the Cash Hill Anticline behaved mechanically in a different manner during the deformation and therefore never developed an axial planar cleavage. The Cash Hill

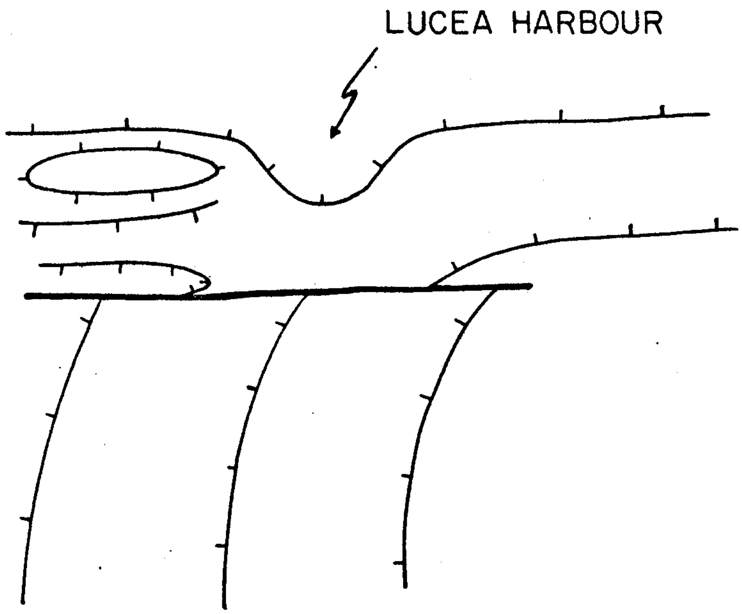
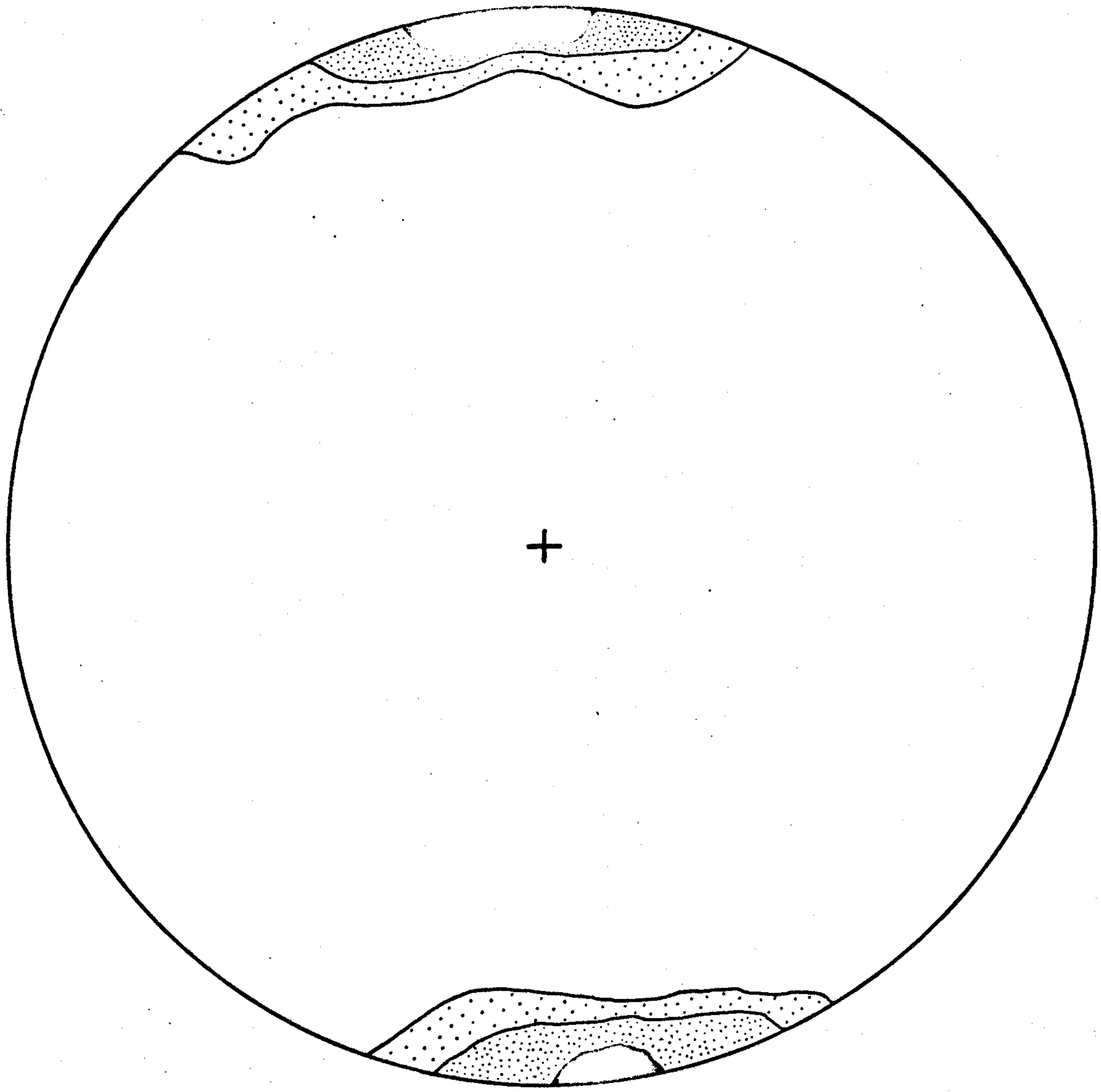


Figure 51 – Sketch, lines with ticks represent strike and dip orientations of the northwest corner of the mapped area. Broad line represents the Fat Hog Quarter Fault zone. Note the structural setting of the Lucea Harbour. See text for description.

Figure 52 – Folded calcite veins within a poorly bedded shale sequence.

Figure 53 - Lower hemisphere stereographic projection of poles to cleavage for the central and northern blocks.

POLES TO CLEAVAGE 53 MEASUREMENTS



9 15 21

POINTS PER 1% AREA

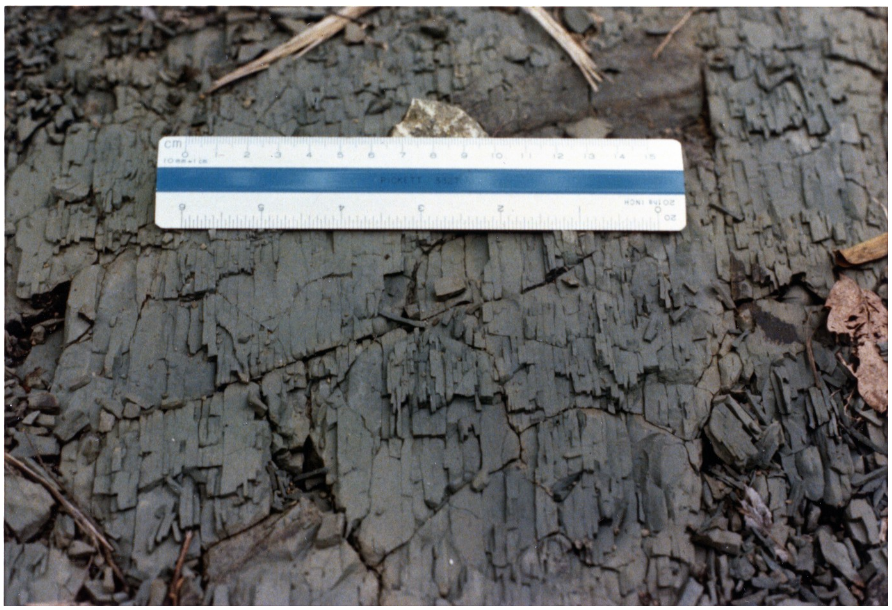


Figure 54 – Spaced, east-west trending cleavage in shales.

Anticline is a relatively tight fold and is strikingly dissimilar to the more open style folding to the north (see Plate 1). This also lends support to the idea that these two regions have behaved in a mechanically different manner. This contrast in folding style can be explained by postulating that units north of the Maryland Fault zone suffered internal deformation (formation of cleavage) while units south of it underwent less internal deformation and therefore were buckled into much tighter folds.

The regional spaced cleavage within coherently bedded units north of the Maryland Fault Zone, is significantly different from the one within the Georgia Complex. The cleavage of the Georgia Complex is more penetrative and locally has a discontinuous-phacoidal character. Two localities had polished, curved, cleavage surfaces (1620-5399) and (1563-5483). The significance of this contrast in cleavage is not well understood, but the penecontemporaneous deformation of the Georgia Complex and/or its proximity to the massive Tom Spring Formation may have been significant in developing this contrast in cleavage.

FAULTING

Fault zones on Plates 1 and 2 are not generally defined by simple, single fault traces but are shown as zones with a braided, splayed and sigmoidal pattern. This style is based on local field evidence and where evidence is lacking is stylized to be comparable to the kind of faulting mapped by such workers as Tchalenko and Ambraseys (1970); Tchalenko and Berberian (1975) in Iran and the Benbow Inlier in Jamaica by Burke et al. (1968).

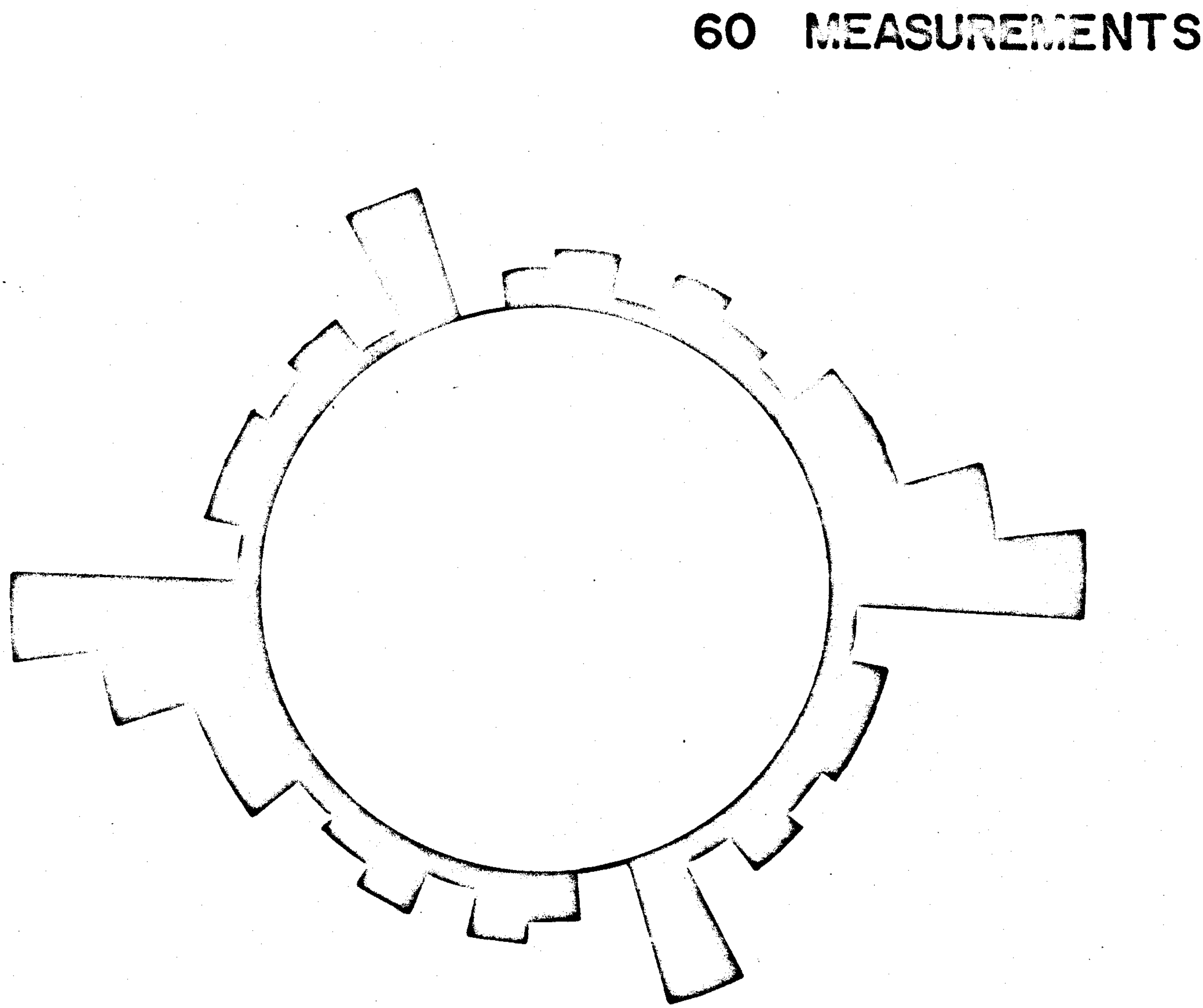


Figure 55 - Rose diagram of faults with dips in excess of 70° that were observed in outcrop.

Within the mapped area faults with dips in excess of 70° have strikes that are very similar to the rest of the island of Jamaica (see Fig. 55). The two maxima obtained for the Lucea Inlier are similar to the two maxima (335° and 275°) defined by Horsfield (1974) for the whole of Jamaica.

The significance and age of faulting are discussed below (p.165).

Maryland Fault Zone

The Maryland Fault is a major east-west trending fault zone that separates the central block from the southern block (see Plates 1 and 2). North of the fault the units have a gentle (0°-30°) dip to the west-northwest. To the south, units have moderate to steep (30°-70°) dips that define the Cash Hill Anticline.

In the Maryland area (1588-5412) exposure is generally good and the fault pattern at this locality is well displayed. West of Maryland the fault is well exposed in a westerly flowing tributary of the Lucea East River (1559-5409). The fault zone is defined by an abrupt change from a lens of the Clifton limestone, to the south, to strongly cleaved calcareous shales, calcarenite beds (5-10 cm thick) and equant masses of calcareous ?concretions. The cleavage is distinctly more closely spaced within the fault zone than in the adjacent sediments. The few calcarenite beds are aligned parallel to the cleavage and have an attitude of 070°-090°, 70°-90°. A second spaced cleavage with a spacing interval of .5 cm is locally developed within this zone (Fig. 56), although its orientation appears to be quite variable. Its significance is not well understood but may be an expression of the history of displacement along this segment of the fault. The possibility that this second spaced cleavage is bedding is rejected because its orientation is discordant to the orientation of the calcarenite beds. This splay of the fault zone is approximately 30 m wide. Units south of this zone are not as pervasively cleaved but have been disrupted by a series of faults.

The Maryland fault zone continues farther to the west and is inferred from abrupt changes in the attitude of bedding (Askenish area) and/or lithological changes and from an interpretation of aerial photographs. The continuation of the fault zone through the Kingsvale area is poorly defined. The lack of distinctive lithologies and deep level of weathering makes its identification difficult. The eastern extension of the Maryland fault zone between Maryland and Patty Hill is not well understood due to inadequate outcrop information and the complex lithologic facies of the submarine canyon complex. Northeast of Patty Hill weathered exposures are limited to footpaths and stream beds, but erratic attitudes of bedding within footpath exposures suggest the proximity of a fault zone. The fault zone is also well expressed in aerial photographs. Farther east the fault zone crosses the Cascade-Pondside road (1724-5441) and is expressed by an abrupt change in the strike and dip of beds. North of the fault zone thick, coarse sandstone and conglomerate beds predominate, while to the south shale and thin, 3 cm thick sandstone beds are exposed.

If one assumes that the lenses of Clifton limestone contained within the fault zone, west of Maryland, were once closely associated with the main Clifton limestone, north of the fault zone; then a left-lateral offset of .5 km can be postulated for this segment of the fault.

Fat Hog Quarter Fault Zone

The Fat Hog Quarter Fault zone is a $\leq$ 1 km wide zone of east-west trending faults and subsidiary north-northwest trending faults. It separates the northern block from the central block. East of Johnson Town (1450-5588) the units north of the south bounding fault have a consistent east-west strike and southerly dip; south of the fault zone the units dip westerly and represent the nose of a large anticline.

The fault zone is named after the village of Fat Hog Quarter. Here the south bounding fault is well exposed on both sides of a 20-25 m long road cut (1315-5557). Figure 57 is a sketch of the west face of this road cut. The east face has a quite different character. The very complex, disrupted kink type folding (see Fig. 58) of an interbedded sequence of turbiditic sandstones and shales is in contrast to gently dipping thinly bedded shale to the north. The exposure is dissected by an anastomosing network of high angle faults (see Fig. 56). Small thrusts were observed in the outcrop and adjacent ones. Structures similar to these described above are well known within the San Andreas Fault Zone south of Palmdale, though several times larger.

Farther east the fault is observed in outcrops north of Richmond (1350-5556) and is expressed as a 2 m wide shear within a sandy cobble to boulder conglomerate. There is also a sharp increase in the proportion of sand as the fault zone is neared. Near Dundee, 4.5 km to the east, the fault is inferred to offset the contact of the Mount Peace Formation and overlying Claremont Formation in the north and the Mount Peace Formation and overlying Harvey River Formation to the south. The Harvey River and Claremont Formations are thought to be stratigraphically equivalent (see p. 66). Between Dundee and Richmond the location

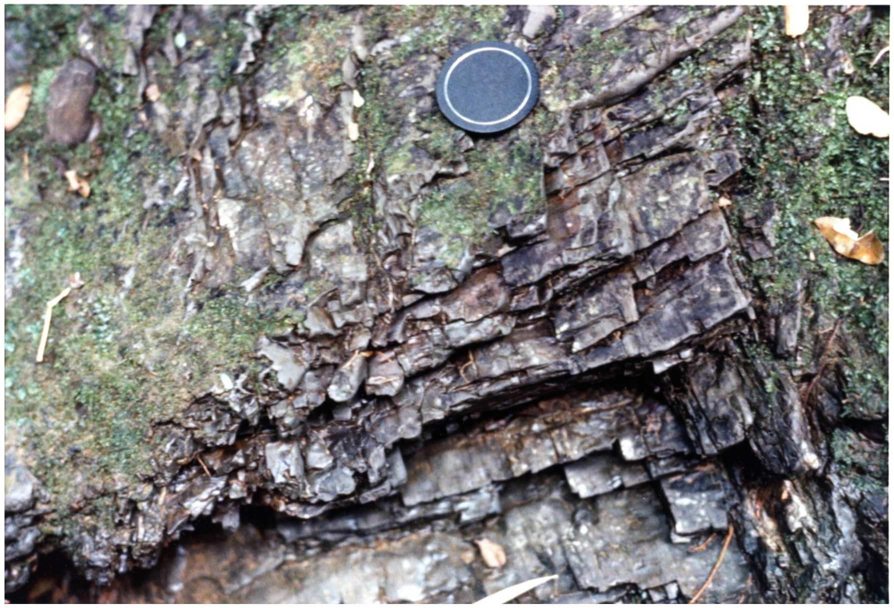


Figure 56 – Deformed, unbedded calcareous shales within the Maryland Fault zone. "Second" spaced cleavage dips off to lower left in photograph.

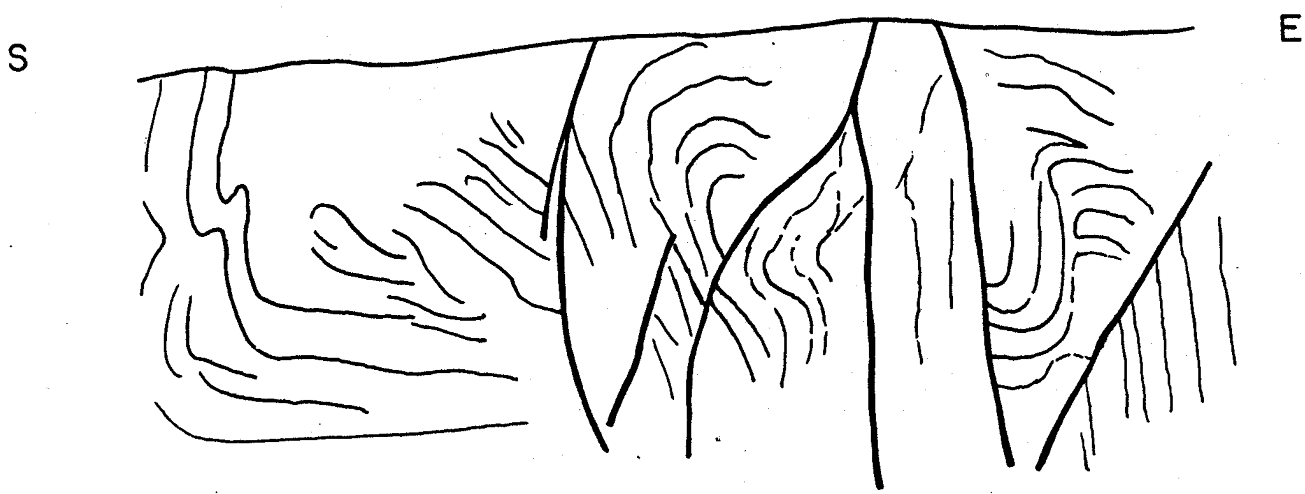


Figure 57 – Sketch of outcrop displaying the style of deformation within the Fat Hog Quarter Fault zone. Note kink type folding and anastomosing fault pattern.

Figure 58 – Photograph of large kink style fold. Western end of Figure 57.

of the fault is based on abrupt changes in attitude of beds and the occurrence of disrupted zones that have been observed in outcrops along numerous north-south trending roads.

The western extension of the Fault is defined from aerial photographs and offsets of the White Limestone Group.

East-west trending faults south and east of the Lucea Harbour are defined by abrupt changes in strike and dip of beds and lithology, and by an interpretation of aerial photographs.

The approximately .3 km left-lateral offset of the contact between the Mount Peace Formation and overlying Claremont and Harvey River Formations (the Claremont and Harvey River Formation have been shown to be stratigraphically equivalent, see p.66) is the only way to estimate the amount of displacement within the Fat Hog Quarter Fault zone. However, a possible continuation of it to the east offsets the White Limestone Group approximately .2 km in a left-lateral sense.

No information is available for determining the amount of offset along NW trending faults.

CHAPTER IV

PETROGRAPHY

Intrusives

Luecogabbroic to dioritic sills were observed at several localities within the southern block. Primary mineralogies are hard to make out due to alteration and metamorphism. Plagioclase is by far the most abundant mineral (60% +) and is generally extensively altered to white mica, prehnite, epidote-zoisite and rare (?)actinolite or (?)other amphibole occurring as fine, high relief needles. Generally equant crystals of poorly cleaved pigeonite occur interstitially as aggregates. Larger crystals of pyroxene show an ophitic texture. Primary(?) hornblende and biotite occur in trace quantities. Metamorphic spherules of prehnite and crystal aggregates of epidote-zoisite are well developed. Chlorite is relatively widespread and is commonly replacing biotite and pyroxene. Pumpellyite was also observed as small interstitial crystals pleochroic in pale green.

Volcanic Clasts of the Tom Spring Formation

Volcanic clasts within the Tom Spring volcanic conglomerate are invariably mafic and range from basalt to andesite. In outcrop most clasts are non-vesicular and in thin section are holocrystalline to hypocrystalline. Plagioclase is invariably the most abundant mineral and may make up to 90% of the rock, and occurs interstitially and as phenocrysts. Phenocrysts of hornblende, pyroxene and olivine occur, though they generally together make up less than 15% of the rock. Olivine is commonly replaced by chlorite so that primary olivine is rarely seen. Pyroxenes are usually pigeonitic though augite may be present. Hornblende crystals are commonly rimmed by magnetite which is common for andesites (Hatch, et al., 1972). Most samples are extensively chloritized.

Arenites

The character of the arenites within the Lucea Inlier varies, but they are generally moderately to poorly sorted and subrounded to subangular. Petrographically they are immature sediments and are composed of volcanic lithic fragments and crystal fragments and can be classified as lithic and felspathic arenites (Pettijohn, et al., 1972). Matrix is quite variable and ranges from being non-existent to 40-50% carbonate. Extensive recrystallization and alteration make it difficult to recognize a detrital matrix similar to that reported from "true" greywackes (Pettijohn, et al., 1972). Normal and oscillatorially zoned plagioclases are extensively altered and recrystallized so that An contents could not be readily determined. Plagioclase is the most abundant framework grain and generally makes up 30-60% of the arenites. It is commonly replaced by chlorite, zeolites, calcite and white mica. Crystal fragments of augite, pigeonite and hornblende also occur but together generally do not make up more than 15-20% of the rock. Biotite is occasionally seen in trace quantities. Kuno (Miyashiro, 1974a, p. 328) originally equated volcanic rocks whose groundmass pyroxenes are augite, ferroaugite and pigeonite with the tholeiitic series. The occurrence of pigeonite within clastic and volcanic rocks of the Lucea Inlier may indicate tholeiitic volcanic sources. Although on the basis of geochemistry Miyashiro (1974a) has shown that Kuno's mineralogical classification is not always true. Quartz occurs as clear, lucid fragments and polycrystalline, undulose fragments and are thought to represent, respectively, volcanic and plutonic sources. Quartz-bearing specimens commonly contain <5% quartz and rarely exceed ~10%. Volcanic lithic fragments (VLF) are generally more abundant in coarser arenites and may make up to 80% of the rock. VLF are typically of volcanic flows. Detrital epidote occurs as equant crystals that are pleochroic in light yellow. It commonly makes up only a few percent of the rock. Metamorphic epidote is locally developed.

METAMORPHISM

Mineral assemblages of chlorite, epidote, prehnite, pumpellyite and white mica within mafic sills of the southern block indicate a low-grade prehnite-pumpellyite metamorphism (Miyashiro, 1974b) and the extensive development of zeolites and epidote, within arenites, suggests that much of the section has been subjected to low-grade metamorphism. The lack of <u>in situ</u> volcanics and plutons or geologic evidence indicative of a subduction complex (see Chapter VI) suggests that the Lucea Inlier represents an arc-related sedimentary basin. The metamorphic grade is attributed to deep burial and concomitant high geothermal gradient associated with the nearby volcanic arc. The exposed section within the Lucea Inlier is a minimum of 4.5 km thick. Overlying uppermost Cretaceous to (?)Paleocene sediments, if deposited, have been eroded; therefore, a minimum depth of burial for the basal section of the Lucea Inlier of 6 km does not seem unreasonable. Burial metamorphism of the basal part of the [illegible] arc basins (Great Valley Sequence) of California has been reported by Dickinson, et al. (1969). Depth of burial was probably in excess of 10 km for the Great Valley Sequence. Back-arc and arc regions have geothermal gradients that are considerably greater than fore-arc regions (Uyeda, 1977).

The possibly high geothermal gradient of the Lucea Inlier may also lend support to the interpretation that the Lucea Inlier represents an intra-arc to back arc basin (see Chapter VI).

CHAPTER V

ENVIRONMENT OF DEPOSITION

Introduction

Recent work on the morphology and sedimentological characteristics of submarine fan complexes and their feeder channels has given geologists models by which to interpret turbidite facies associations (Walker, 1970). Grossly analogous models for submarine fans were developed simultaneously by workers in the Apennines (Mutti and Ricci Lucchi, 1972) and marine geologists working off the coast of California (Normark, 1970).

Initial reconnaissance and later more detailed field work within the Lucea Inlier had shown that the facies characteristics and their distribution were comparable to the turbidite facies association of fan models. Below I briefly review the characteristics of submarine fan models and then apply them to the stratigraphic succession within the Lucea Inlier. The submarine canyon complex identified is discussed in detail at the end of this chapter.

A submarine fan complex is a fan-shaped wedge of sediment that accumulates at the base of a submarine slope. The sediment wedge is fed by a feeder channel(s) (Walker, 1978) or a canyon or canyons which are commonly incised into a shelf ot slope sequence. Sediment is funneled down the feeder channel by a variety of sediment gravity flow mechanisms. Downfan, the feeder channel divides into a myriad of distributary channels that eventually lose their topographic expression as they merge with the basin plain (see Fig. 59). The distributary channels deliver sediment to a series of depositional lobes. Most research on interpreting fan complexes from the rock record has assumed a prograding wedge of sediment. Obviously this is too simplistic an

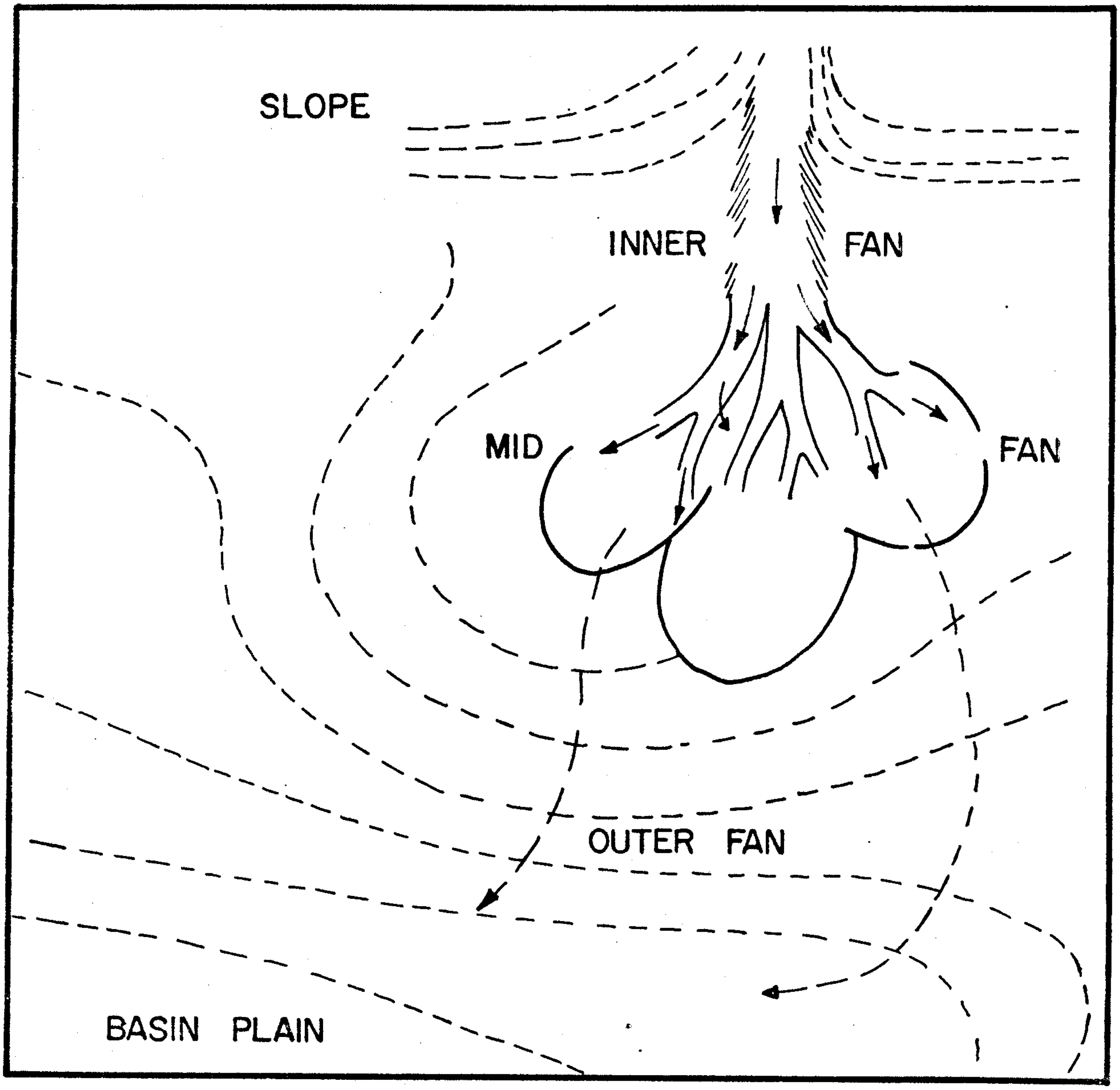


Figure 59 - Simplified sketch of submarine fan model (redrawn from Walker and Mutti, 1973).

approach but appreciation of the variables involved has been expressed by Walker (1978) and Anonymous (1978, Penrose Conference participants).

The terminology used by students of recent and ancient submarine fans is variable and has been reviewed by Ingersoll (1978). In essence submarine fans can be subdivided into inner, mid and outer fan facies association (see Fig. 59) (Ricci Lucchi, 1975).

Inner Fan

The inner fan facies association is commonly characterized by a conglomeratic filled feeder channel or canyon that is encased in shales of shelf to slope environment (Piper, et al., 1976). The canyon fill is characterized by a variety of mass flow type deposits. Overbank levee-type deposits may be present along the margins and document over-spilling of the canyon walls.

Mid Fan

The association of the mid fan region is characterized by a series of distributary channels that head into the main feeder channel. Channels commonly are filled with thick, massive, coarse sandstone beds that are commonly amalgamated. Intervening interchannel regions are characterized by overbank levee-type deposits near channels and grade laterally into interchannel shale. Commonly, channels are abandoned due to avulsion or meandering.

Outer Fan

The outer fan is characterized by more distal classical Bouma type turbidites and intervening hemipelagic deposits (Mutti, 1977). Sandstone beds are generally regularly bedded.

Stratigraphic successions of fan facies associations may be described as progradational or retrogradational suites (Ricci Lucchi, 1975) that correspond, respectively, to offlapping and onlapping relations within the basin. Rate of subsidence (relative change in sea-level), rate of sediment influx and availability of sediment are the major variables that will determine apparent onlapping and offlapping relationships. See Normark (1974) for a discussion of variables affecting submarine fan characteristics. Changes in sediment input may reflect actual changes in source regions or may be an expression of channel avulsion of a small scale, e.g. distributary channel or large scale, e.g. main feeder channel (see Normark, 1970a). It is apparent that the actual cause for a given retrogradational or progradational suite may not be known, unless detailed data can show otherwise, therefore such suites are best described in a purely descriptive way. This approach is followed in this study.

Below I attempt to discuss the succession of facies exposed within the Lucea Inlier in terms of submarine fan facies association. The stratigraphic relationship between the southern and central block is not well, and they are therefore discussed separately.

Southern Block

In a gross way the stratigraphic sequence exposed within the Cash Hill Anticline represents a mid-inner fan to an inner fan to (?)shelf environment. Thick, coarse, commonly amalgamated sandstone beds and local conglomerate of the lower Birchs Hill Formation represent channel fill deposits of the mid to inner fan environment and are analogous to mid fan channel deposits reported by Ingersoll (1978, Fig. 4c, d).

The overlying interval of thinly bedded shale and intercalated sandstone and micrite of the Castle Hyde Member indicates a reduction in clastic sediment input into this sector of the basin. The upper Birchs Hill Formation records a resumption of coarse clastic sedimentation and is thought to represent a retrogradational mid-fan to outer fan succession. The basal 2/3 of the section is dominated by thick, massive, mass flow deposits to Tab turbidites (see Fig. 8, p. 21). Sandstone beds are commonly amalgamated though intervening intervals of pelite are locally preserved. This facies is interpreted as representing a mid-fan environment. The upper 1/3 of the sequence is progressively more thinly bedded and in it classical Bouma-type turbidites are more abundant. Although massive, thick ($\gtrsim$ 1 m) beds are still present. The upper Birchs Hill Formation has been shown to thin to the northeast (see Plate 2A) where beds are more thinly bedded and pelitic and locally well laminated. These units to the northeast may represent channel margin and overbank deposits (Mutti and Ricci Lucchi, 1975). An interval of thinly bedded to ribbon banded shale, fine sandstone and limestone separates the underlying mid fan facies association from the overlying inner fan channel deposits of the Tobolski Formation. The channel fill is composed of pebble to boulder conglomerate, pebbly mudstones, massive sandstones and local pelitic intervals. A progressive thinning and fining of the channel fill from the southern limb of the Cash Hill Anticline to the core region probably marks the northwestern edge of the channel. Overlying thinly bedded, locally red, shales and thin channeled conglomerates of the Askenish Formation may represent a (?)slope to (?)shelf environment. In situ Inoceramids within massive shales may indicate a shelf environment. Though, no slump features thought to be indicative of slope environments (Ricci Lucchi, 1975;

Piper, et al. 1976) were observed.

Central Block

The lowest lithounit of the central block is the Pioneer Formation and is a sequence of thickly bedded, channeled conglomerate to massive sandstone. This facies probably represents an inner fan to mid fan channel deposit. Conglomerates are typically channeled and unorganized. Large scale traction structures and imbricated pebbles as reported by Winn and Dott (1978) and Walker (1975a) have not been observed. More thinly bedded and more pelitic horizons are developed to the northeast and southwest. This relationship indicates that the trend of sediment transport and location of the main channel were respectively to the northwest and east of Cascade. Sparse paleocurrent supports a south to north transport direction (see p. 111).

The overlying Mount Peace Formation represents a shelf environment and for this reason the Pioneer and Mount Peace succession is considered to represent a retrogradational phase. The Mount Peace Formation and associated submarine canyon complex are discussed separately below.

The shelf and submarine canyon environment of the Mount Peace Formation and submarine canyon complex are overlain by basal shelf sediments of the Claremont and Harvey River Formation. An overlying mega, fining and thinning upward sequence of channeled conglomerate mass flow to turbiditic sandstones and intercalated shales of the Claremont Formation represent a retrogradational suite of an inner to mid-outer fan environment. The conglomerates typically have a sandy matrix and are unorganized and are similar to channel deposits reported by Carter and Lindqvist (1977) and Stanley and Unrug (1972). Mass flow, massive, coarse sandstone beds locally contain isolated, floating clasts (see Fig. 35 p.60). Similar features have been reported by Carter and Lindqvist (1977). Apparently allochthonous blocks of thinly (~4 cm) bedded medium to fine sandstone beds and intercalated shales are complexly associated with mass flow channel deposits (see p. 59). These thinly bedded horizons are thought to represent overbank deposits that may have accumulated outside the channel or on terraces within the channel and have subsequently become incorporated into the channel fill. Widening or migration of the channel due to periodic incision and/or slumping of the channel walls may cause blocks of levee deposits to be incorporated into channel fill deposits. Similar features have been reported within the Great Valley Sequence of California by Ingersoll (1978, Fig. 8a). The channel fill is encased in thin (5-15 cm) Tbc turbidites and intercalated shales and interchannel shale deposits that are developed to the east and south (Harvey River Formation). Occasional channeled slide conglomerates and massive sandstone beds are developed to the south and may represent distributary channel deposits. Higher intervals within the Claremont Formation, north of Claremont, probably represent a mid to outer fan environment that is characterized by thin to thick (2-70 cm), locally, channeled fine to coarse turbiditic sandstone and intercalated shale. Thick disorganized conglomerates occur locally (1508-5661) but their stratigraphic position is not understood due to faulting. These higher stratigraphic intervals of the Claremont Formation are laterally equivalent to thinly bedded shales and local irregularly bedded fine sandstone. This facies probably represents overbank to interchannel deposits of the mid to outer fan environment.

The overlying succession of the Meddlesex Formation and Dias Formations is not clearly understood in terms of fan environments. Though on the basis of stratigraphic arguments the basal shales of the Middlesex Formation may be lateral equivalents to the upper horizons of the Claremont Formation and therefore represent mid to outer fan interchannel shale deposits. The occasional horizons of channeled mass flow slide conglomerates and massive sandstone beds of the Middlesex Formation may represent the distal portions of major mass movement events of the inner fan. Encasing shales and intercalated turbiditic sandstones may illustrate interchannel to outer fan deposition. The succession of the Dias Formation may be interpreted in a similar way. Though, there is a marked increase in the quantity of shale within the Dias Formation.

No attempt is made to describe the section exposed within the northern block, west of the Lucea Harbour, in terms of submarine fan facies association. Extensive faulting has complicated the original succession of facies.

Paleocurrent Data

From a total of 1000 localities visited only 27 paleocurrent measurements were obtained of which 6 are vectorial. Figure 60 is a rose diagram of this data. The paleocurrent data shows a fan pattern about a north-south axis with a south to north transport direction. Current features are typically small scaled and do not have great variability in form and were limited to flute, groove and brush casts (Potter and Pettijohn, 1977). Paleocurrent structures are best developed within classic Bouma type turbidites.

The fan-shaped pattern of sediment transport is what one would expect and does observe (Ricci Lucchi, 1975) for submarine fan settings.

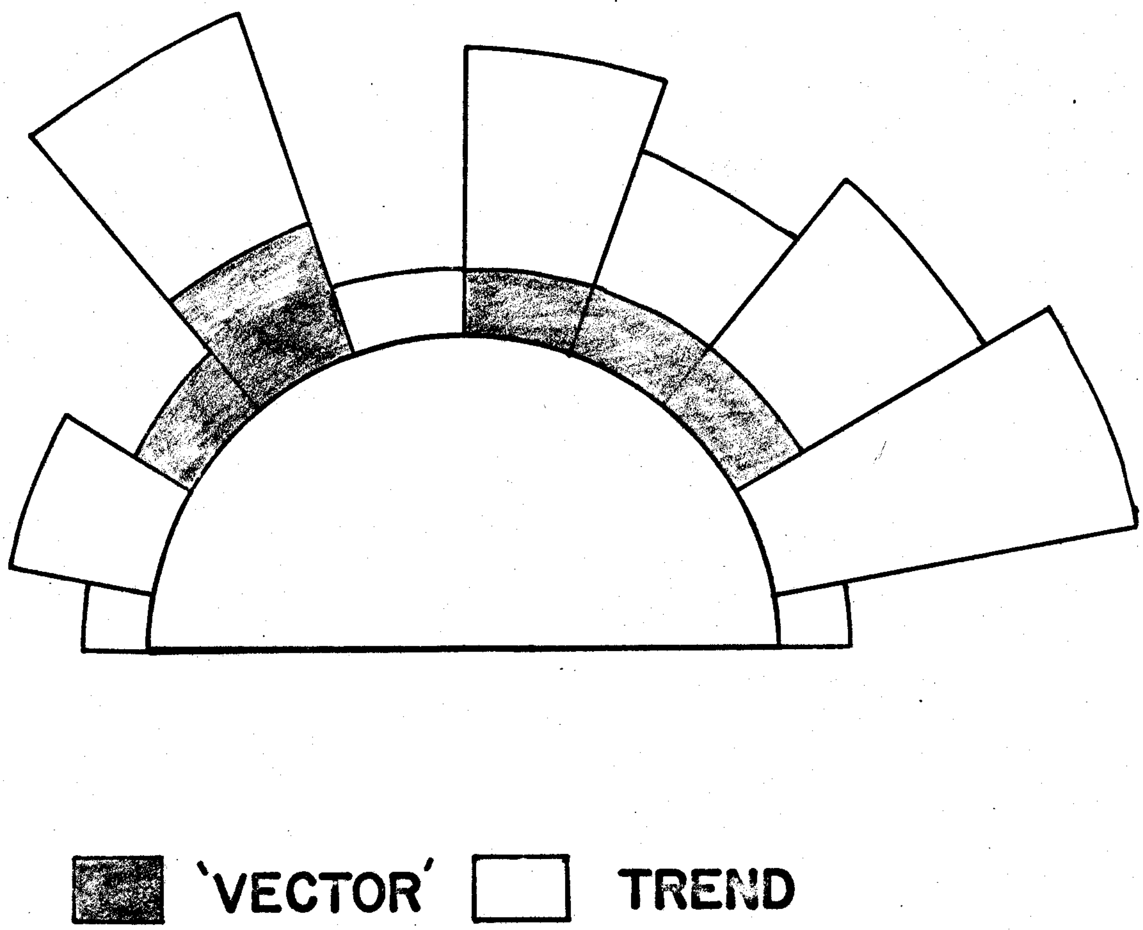


Figure 60 - Plot of 27 paleocurrent measurements for the Lucea Inlier.

SUBMARINE CANYON COMPLEX

Reconnaissance maps of the Lucea Inlier show parts of the Tom Spring Formation as an andesite intrusive (north of Cascade) and others as agglomerate (Tom Spring area). Both of these units were shown as relatively small bodies (Ki and Ka) on the 1:50,000 geologic map of the Lucea Inlier (Bateson, 1974). In this study I have mapped both as part of the Tom Spring Formation and have mapped out its distribution in much greater detail than earlier geologists found practicable. It is argued here that the Tom Spring Formation is neither an andesitic intrusive nor an agglomerate but, a catastrophically emplaced volcanic conglomerate that deformed the walls and floor (Georgia Complex) of a submarine canyon.

North of Tom Spring (1571-5549) a 335° trending andesite dike was delineated on the 1:50,000 geologic map of the Lucea Inlier (Bateson, 1974). I have reinterpreted this 'dike' to be an allochthonous block of an andesitic flow that was transported in the Tom Spring Formation. The earlier 335° trend was reasonably based on topography, however, my mapping has shown this trend to be topographic but not structurally controlled.

Below is a summary of evidence on which my interpretation of the Tom Spring Formation and the Georgia Complex as canyon-related is based.

Tom Spring Formation

Gross spatial relationships show the Tom Spring Formation to have the form of a roughly tabular body that on an outcrop scale is locally channeled into the underlying Georgia Complex and also cuts across the regional structure. The Tom Spring Formation and the associated underlying Georgia Complex are channeled into an extended stratigraphic thickness of the Mount Peace Formation and probably the Pioneer Formation. Figures 22 and 31 (pages 37 and 52) are interpreted as indicating that while the Mount Peace Formation was accumulating, probably in a shelf to slope environment, a northwest-southeasterly trending canyon cut across this shelf. The lack of slump features within the shales of the Mount Peace Formation may suggest a shelf environment for its deposition (Piper, et al., 1976 and Ricci Lucchi, 1975). This canyon was then filled by the catastrophic emplacement of the Tom Spring Formation and the Georgia Complex. Assuming that the emplacement of the Tom Spring Formation was a single event, then an estimate can be made about the relief within the canyon. This was done by estimating the stratigraphic interval cut by the canyon complex from Plate 2. A relief of 450 m + is estimated.

Recent and ancient submarine canyons generally have coarser facies lining their floors compared with laterally adjacent shelf and slope shales (Stanley, 1974; Whitaker, 1974; Normark, 1974; Mutti and Ricci Lucchi, 1975).

Similar facies relationships are seen between shales of the Mount Peace Formation and coarse sandstones, slide conglomerates and pebbly mudstones of the Georgia Complex (Fig. 31 p. 52).

Gradational lithologic contacts between the Tom Spring Formation and adjacent shales and sandstones of the Georgia Complex suggest that the two lithounits became intimately mixed during emplacement of the latter.

Petrographically the matrix of the Tom Spring Formation has a detrital character in which a variety of volcanic lithic fragments are set in a coarse matrix of altered plagioclase crystals and minor clinopyroxene and hornblende. Textures suggesting an igneous origin have only been observed in a possible extension of the Tom Spring Formation to the south (1618-5418).

It appears that geologists working in the mapped area have used the term agglomerate in a very general way. MacDonald (1972, p. 129) defines agglomerate as "Masses of tephra containing a large proportion of bombs which generally form either in the throat of the volcano itself or on the flanks close to the vent." No bombs have been observed in the Tom Spring Formation and for this reason the term agglomerate is inappropriate. Also, the Tom Spring Formation is intimately associated with marine sediments. The occurrence of subaerial volcanics 25 km to the south, within drill holes (Meyerhoff and Krieg, 1977), suggests that volcanic islands lay to the south. Sparse paleocurrent data also shows a south to north transport. These observations tend to rule out the possibility of the Tom Spring Formation being an agglomerate.

The possibility of the Tom Spring Formation being an andesitic intrusive is disclaimed because of its textural character, contact relations and complex geometric form.

Georgia Complex

Exlcuding the Cash Hill Anticline, south of the Maryland Fault zone, the Lucea Inlier consists of a broad, open, westward-plunging anticline with gentle to moderate dips (10°-40°). A spaced vertical cleavage is locally well developed in fine siltstones and shales. No mesoscopic folds of tectonic origin have been mapped. The structural character of the Georgia Complex is anomalous as compared with the rest of the mapped area in that: 1) dips range from horizontal to overturned; 2) complex folding is common; 3) locally chaotic structure is developed and 4) extended intervals of unbedded pelite occur within the unit. Cleavage, although coincident with the regional cleavage, is more penetrative in character. The significance of this is not apparent. The marked contrast in structure between the Georgia Complex and surrounding sediments is observable at three localities (1573-5474, 1659-5462 and 1662-5476) where equivalent lithologies are deformed and undeformed.

The deformation of the Georgia Complex is believed to be of penecontemporaneous origin because of its chaotic and anomalous structure compared with the rest of the inlier.

It might be hypothesized that the deformation of the Georgia Complex is of tectonic origin. The contact zone between the massive Tom Spring Formation and adjacent bedded sediments might have been a high strain zone during deformation. Strain would be high in this region due to the marked rheological contrast between the two units. If this were the case one would then expect the foliation in the Georgia Complex to parallel its contact with the overlying Tom Spring Formation. This is not the case, foliation in the Georgia Complex is coincident with the regional spaced cleavage. The only contrast is that the cleavage of the Georgia Complex is more penetrative and locally has a phacoidal character.

It would seem reasonable that some of the deformation of the Georgia Complex is of tectonic origin. However, the relative importance of tectonic and sedimentary structures is not well understood. Structures of tectonic origin have been observed southeast of Tom Spring (1573-5474). Here gentle open folds with horizontal axes trending east-west and an axial planar cleavage are coincident in orientation with the regional structure. However, all other structures observed within the Georgia Complex have no relation in style or orientation to the regional structure of the inlier. Deformational features of tectonic origin are considered of only minor importance within the Georgia Complex.

Size and Map Pattern

At the present level of erosion the submarine canyon complex is 4.5 km in length, 4 km wide at its northern end tapering to 2 km in the south. These dimensions plus the estimated canyon relief of 450 m + are comparable to typical small submarine canyons such as those reported by Whitaker (1974) and Shepard (1973).

Two significant features revealed by mapping the lithologies in the submarine canyon complex are: 1) the apparent absence of the Georgia Complex on the eastern side of the Tom Spring Formation and 2) the Tom Spring Formation's near horizontal contact with underlying units along its eastern side. These relationships can be explained in this way: originally the Tom Spring Formation was bounded on both its east and west sides by the Georgia Complex, as is seen for its west side (see Fig. 22, p. 37). During the folding of the Cretaceous sequence the submarine canyon complex was tilted to the west so that, at the present level of erosion, the canyon wall facies has been preserved only on the westerly side. While the eastern wall has been eroded and only the base of the canyon fill is observable (Fig. 61).

In conclusion, it is envisaged that prior to the deposition of the Tom Spring Formation and development of the Georgia Complex a north to northwesterly trending submarine canyon cut across the shelf and (?)slope of a volcanic island arc that lay to the south (Fig. 62). Shales and intercalated thin sandstones and lenticular reefoidal limestones of the Mount Peace Formation were accumulating on the shelf.

Coral and rudist frameworks in the Clifton limestone (Coates, 1977) and in situ inoceramids stratigraphically above the Mount Peace Formation (1566-5527, 1525-5465 and 1544-5565 - fossil localities) suggest shallow water conditions during the deposition of the Mount Peace Formation. A complete reconstruction of the facies distribution for the canyon walls and floor, pre-Tom Spring emplacement, is not possible, although abrupt facies changes have been documented within the Mount Peace Formation (see Figure 31, p.52). Also, thin (2-10 cm) Tbc well sorted sandstone beds are developed near the submarine canyon fill, within the Mount Peace Formation, and may represent overbank levee deposits.

The Tom Spring Formation is a resedimented mass of tephra that was either erupted during a major volcanic event and immediately transported into a marine envirnoment, or a mass of volcanic debris that had accumulated near the head of the submarine canyon and was later moved downslope perhaps after an earthquake or major storm. Whatever the source of the Tom Spring Formation, it was funnelled down a canyon as a massive body deforming the walls and floor of the canyon as it moved (Fig. 63). The apparent lack of stratification in the Tom Spring Formation suggests that it was emplaced en-masse. The actual

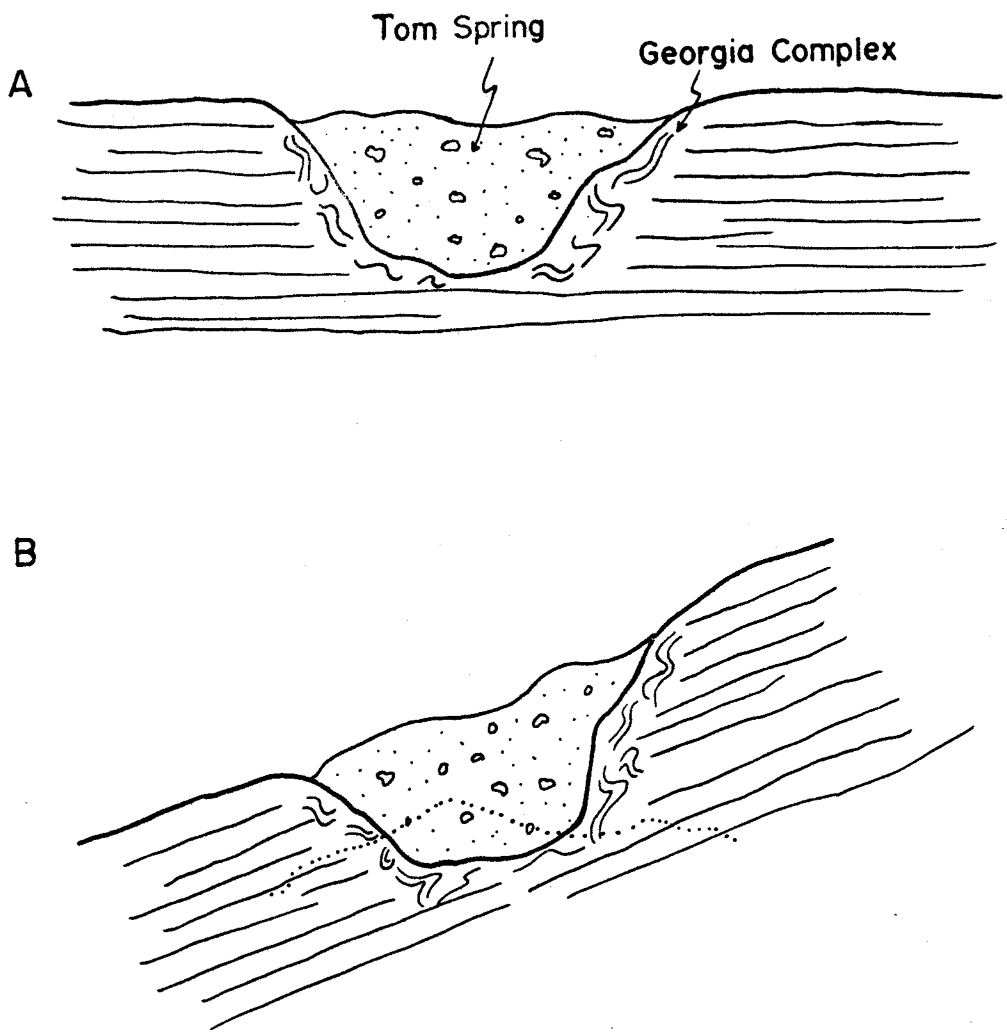


Figure 61 - A) Pre-folding relationship of submarine canyon complex and adjacent undeformed sediments. B) Folding of sequence and approximate position of present erosion level.

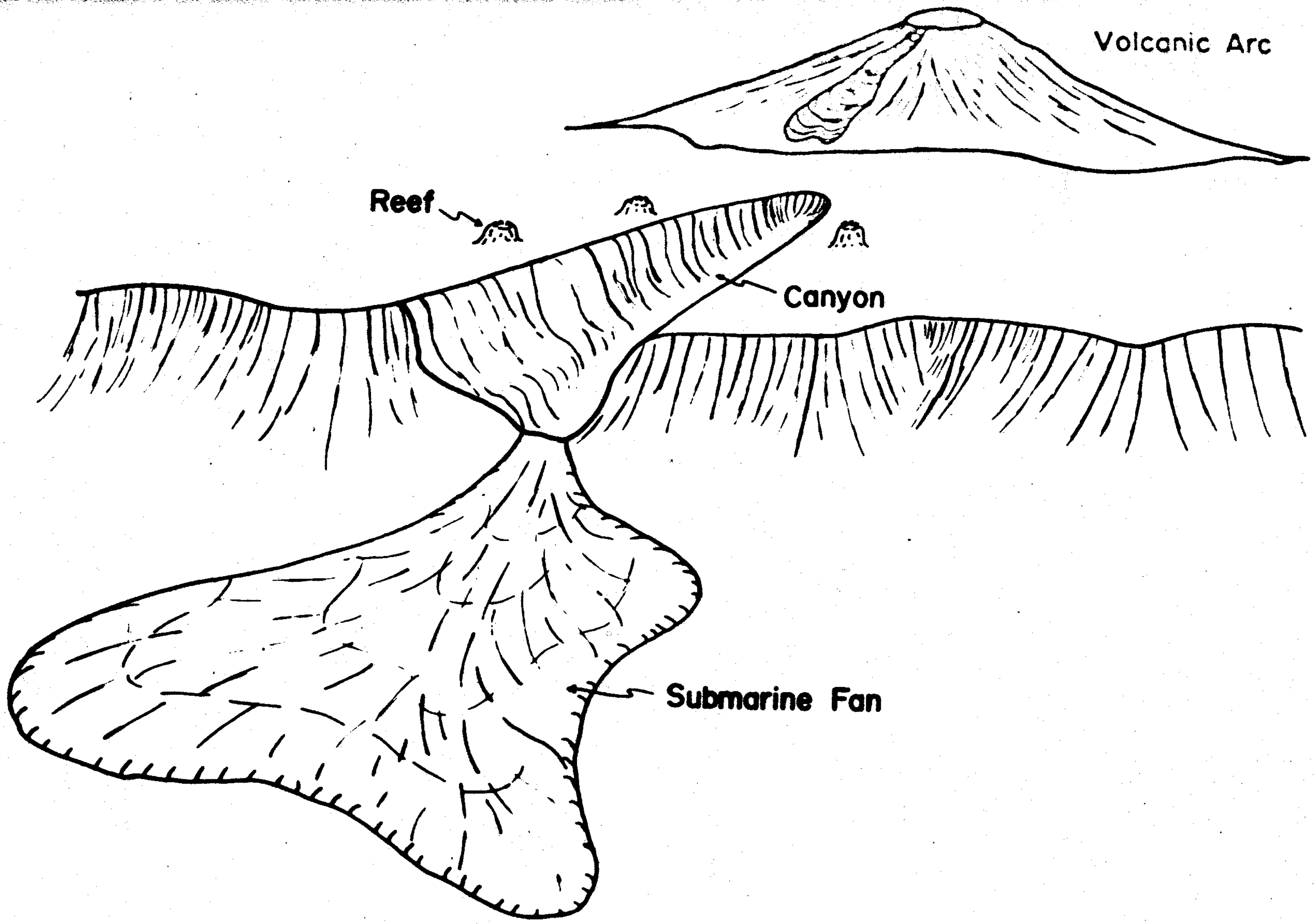


Figure 62 - Generalized paleogeographic reconstruction of the Lucea Inlier region prior to the emplacement of the Tom Spring volcanic conglomerate.

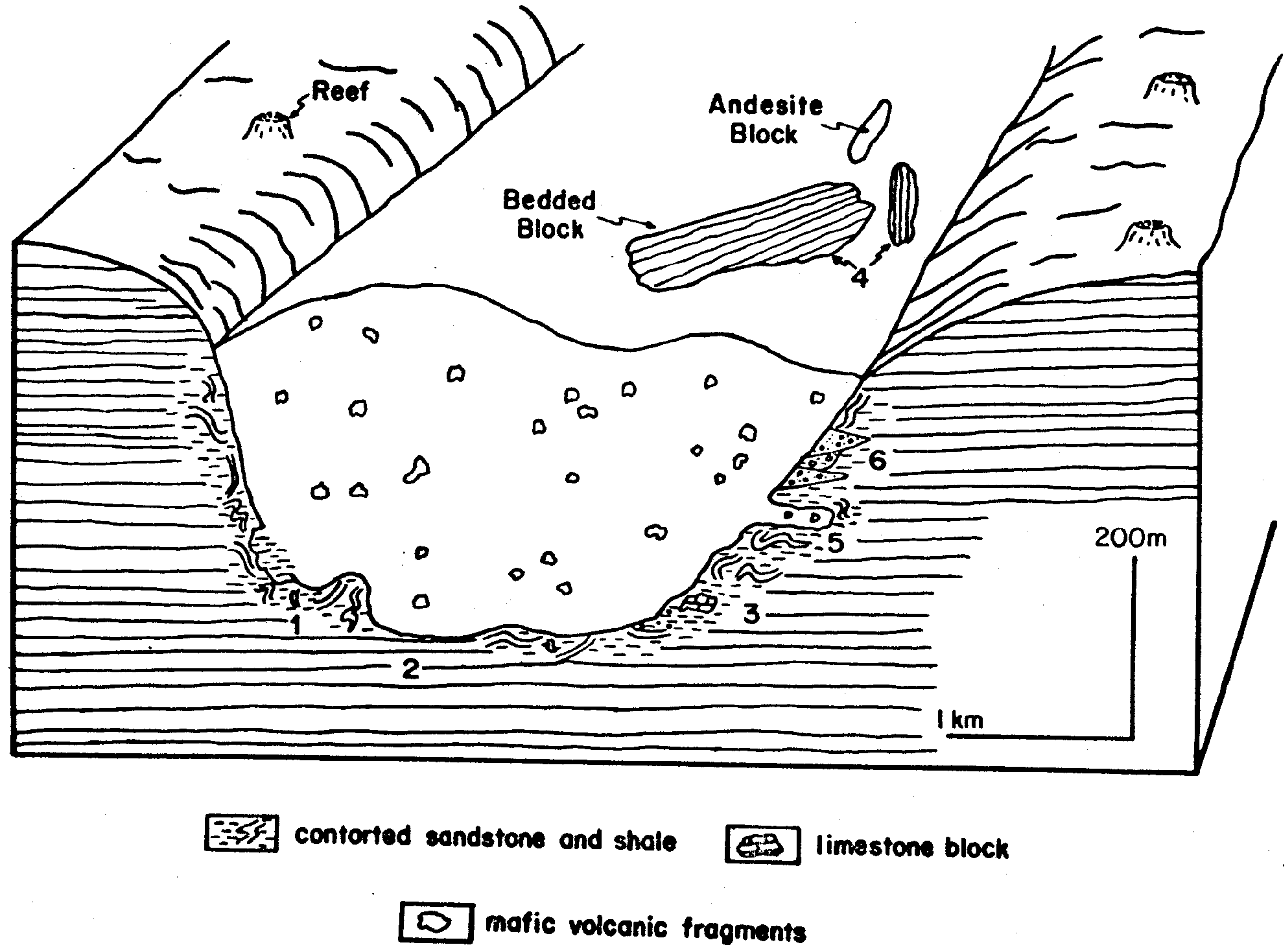


Figure 63 - Block diagram illustrating the general structure of the submarine canyon complex. Canyon fill is the Tom Spring Formation with encased mafic volcanic fragments. Note contrast in structure of canyon walls (Georgia Complex) and coherently bedded adjacent units. Numbers correspond to localities on Figure 22. Scale is approximate.

mechanics of transport is not well understood, but may be analogous to a debris flow (Middleton and Hampton, 1973). However, it has been suggested by Parsons (1969) that post-depositional slumping of re-sedimented tephra may obliterate previously developed stratification. This mechanism would need to be sufficiently effective to obliterate all signs of bedding except for that in the two localized areas, it seems more likely that the Tom Spring was deposited structurelesss.

The varied occurrence of modern and ancient submarine canyons cutting into shelves and slopes has been reviewed by Whitaker (1974; 1977) and Shepard (1973) and documented by Shepard, et al. (1969); Shepard and Emery (1970); Piper (1970); Piper, et al. (1978); Carter and Lindquist (1977). Canyons cutting into the shelf and slope environments of ancient island arcs have not been extensively documented in the geologic literature. However, bathymetric maps of the Aleutian island arc (Scholl, et al., 1974). Indonesian archipelago (Hamilton, 1974) and the West Indies arc (Case and Holcombe, 1976) show that submarine canyons cut into shelf and slope facies are common features of arcs and as such preservation within the rock record of fossil island arcs seems not unlikely. It may also be argued that the near sea level location of submarine canyon heads causes them to be easily eroded.

Recent submarine canyons, within the Aleutian island arc, have been reported by Marlow, et al. (1970). Submarine canyons filled with coarse volcaniclastics, that are incised into mudstones of a narrow shelf, have been documented by Ingersoll (1975) from the Upper Jurassic of the Great Valley sequence in California. The dimensions and characteristics of these canyons have not been described.

CHAPTER VI

TECTONIC SETTING

The occurrence of blueschist metamorphics, serpentinite, granodiorite-tonalitic plutons and overlying calc-alkaline volcanic sequences strongly suggest that the Cretaceous geology of Jamaica records an episode of plate convergence. The Lucea Inlier represents a segment of the Jamaican Cretaceous arc-trench system, and is here interpreted in the light of present-day island arc models.

The structural framework and basin evolution of active island arc systems have been reviewed by Karig and Sharman, 1975; Dickinson, 1974). Part of the objective of this study is to understand what sub-environment of an island arc does the Lucea Inlier represent, e.g. subduction complex, lower slope basin (Moore and Karig, 1976), fore arc and upper slope basin, intra-arc and backarc basin or the volcanic edifice proper. The term upper slope basin was defined by Karig (1977, p. 175) as the region that lies between the trench slope break and the frontal arc. Here the term is used to indicate a sedimentary basin that is upslope (arcward) of the trench slope break. This includes back and intra-arc basins. Upper slope basins are characterized by sedimentation without significant deformation.

The section exposed within the Lucea Inlier consists of a 4 km + thick sequence of resedimented volcaniclastics. The occurrence of lenses of reefoidal limestone (Clifton) and immediately overlying *in situ* inoceramids indicate a shelf environment for this part of the section. There is no evidence to indicate a beach or subaerial environment; therefore, a shelf to basin (depth unknown) setting seems most reasonable for the Lucea Inlier. There is no direct evidence for the existence of in situ volcanic rocks, to suggest a volcanic arc, or the lithologic heterogeniety and complex structure commonly associated with subduction complexes. Furthermore, deep water sediments (e.g. radiolarian cherts and pelagic limestone) that are commonly associated with subduction complexes are not developed within the Lucea Inlier. Lithologically the volcaniclastics of the Lucea Inlier lack significant amounts of quartz. Quartz-bearing specimens commonly contain <5% quartz and rarely exceed ~10%. The lack of significant quantities of quartz within the Lucea Inlier and other parts of Jamaica clearly show that Jamaica is an intra-oceanic arc that was not in reach of continental detritus. South to north paleocurrent transport and bore hole data (Meyerhoff and Krieg, 1977), to the southeast, indicate the existence of a volcanic arc to the southeast. Also, the tapering in width of the submarine canyon complex, from 4 km in the north to 2 km to the south suggests a south to north transport. On the basis of lithology and sedimentological setting the Lucea Inlier appears to represent an upper slope basin of an intra-oceanic Cretaceous island arc system. This interpretation assumes a north-dipping Cretaceous subduction zone for Jamaica (see p.133 for discussion).

The simple, open structural character of the Lucea Inlier also seems analogous to present-day upper slope basins of active island arc systems. Reflection profiles across upper slope basins are characterized by near horizontal, relatively undeformed sediments (Coulburn and Moberly, 1977; Karig, 1977; Hussong, et al., 1976; Scholl, et al., 1975; Karig and Sharman, 1975; Seely, et al., 1974; Silver, 1972). Large wavelength, low amplitude folds and high angle faults are commonly observed. However, low angle thrusts with vergence towards the volcanic arc have been observed, within the Indonesian region, on the arc side of the trench slope break (Hamilton, personal commun., 1977). Karig (1977, Fig. 2) has also suggested the presence of thrusts within the fore arc basin of the Middle America arc trench system. Synsedimentary thrusts as noted by Karig and Hamilton have not been observed within the Lucea Inlier. Upper slope basins that have been defined within the rock record are typically mildly deformed packages of volcaniclastic rocks (Dickinson, 1976, 1970; Mitchell, 1970); though, under conditions of suturing or extensive strike-slip faulting upper slope basin sequences may become complexly deformed.

The Lucea Inlier probably represents an upper slope basin that was situated within or behind the Cretaceous volcanic arc of Jamaica. A recent analogue of the Upper Cretaceous geologic setting of the Lucea Inlier may be 'summit basins' (Scholl, et al., 1975) that are perched on top of the Aleutian ridge. Morphologically they are roughly rectangular fault bounded, sedimentary basins that contain up to 4 km of sediment. Water depths in the basins range from 1 to 2 km. Reflection profiling across these basins indicate mild warping and high angle faulting.

Scholl, et al. (1975) have suggested that the summit basins are extensional features associated with a major late Cenozoic volcanic event. They also suggested that they may be extensional structures due to right-lateral oblique convergence. However, the orientations of the basins with respect to the trend of right-lateral slip (Jacob, et al., 1977) is not consistent with models of Wilcox, et al. (1973). DeLong and Fox (1978) (DeLong, personal commun., 1978) have since suggested that they are related to the subduction of the Kula ridge under the Aleutian island arc. Subduction of the Kula Ridge is thought to have occurred in Eocene to Oligocene times. It seems apparent that more definitive models, for their formation, will have to wait til more data is available on their age of formation.

CHAPTER VII

GEOLOGY OF THE CARIBBEAN REGION

Introduction

I here briefly review Caribbean geology in plate tectonic terms. No attempt is made to describe the geologic evolution of each sector in detail (see Nairn and Stehli (1975) and Mattson (1977) for reviews of Caribbean geology). The objective of this review is to attempt to understand the evolution of plate boundaries within the Caribbean from Jurassic to Eocene times. It has to be kept in mind that major Cenozoic strike-slip motion across the northern (see Chapter XIII) and southern Caribbean plate boundaries has greatly modified the original spatial distribution of geologic units.

Central America

It is generally believed now that the geology of Central America records a Late Cretaceous suturing event (Donnelly, 1977; Perfit and Heezen, 1978). The suture zone presently trends approximately east-west and lies astride the Chixoy-Polochic-Motagua fault zones. Gneisses and schists of the Chuacus metamorphic series and overlying upper Paleozoic sedimentary cover make up the northern continental mass. Here, informally referred to as the Guatamalan block. The southern Chortis block (Dengo and Bohnenberger, 1969) consists of Las Ovejas gneisses and schists and overlying phyllites of (?)Paleozoic age. The Cretaceous evolution of each margin is somewhat obscure. But a review of the Cretaceous geology by Wilson (1974) suggests that by Mid-Cretaceous times a broad shelf accumulating carbonate and locally evaporites extended from Mexico to central Honduras. By early Late Cretaceous times a northward facing arc and associated subduction complex developed along the northern side of the Chortis block. Eclogites, ultramafics, blueschist metamorphics and oceanic sediments of the El Tambor (Donnelly, 1977) represent this subduction complex. Continued southward subduction caused the El Tambor to be obducted onto the south facing shelf sequence of the Guatamalan block. As this nappe advanced in Late Cretaceous times it isostatically depressed the shelf and formed the Sepur trough. The trough accumulated turbiditic arenites, hemipelagic shales and reworked shallow-water limestone clasts that occur within a pelitic matrix (Wilson, 1974). The development of the Timor Trough of the Banda arc region is thought to be an analogous feature (Hamilton, 1977).

This plate tectonic interpretation is probably too simplistic (Donnelly, personal commun., 1978) but is presented as a possible working model.

Bay Islands

The Cretaceous geology of the Bay Islands (McBirney and Bass, 1969) strongly resembles that of the Motaqua Suture zone (Donnelly, 1977) farther west. This correlation has also been made by Wilson (1974). The islands lie on the northwest margin of the Nicaraguan Rise. Geologic maps published by McBirney and Bass (1969, Figures 1 and 2) show the islands to consist of mélanges in which a northern band of schists and gneisses contains variably sized lenses of serpentinite and hornblendite. South of this band and in fault contact is a mildly metamorphosed sequence of pelite, chert, marble and quartzose conglomerate. Small plutons of albite granite occur within both zones. McBirney and Bass (1969, p. 237) have described the southern band thus "Faulting and folding are so extensive that meaningful stratigraphic sections cannot be measured and there is no way of estimating the total or relative thickness of the units." The Bay Islands are interpreted in this thesis as being a continuation of the El Tambor tectonofacies.

Nicaraguan Rise

The Nicaraguan Rise (Fig. 2) is a broad 1000 km long easterly trending swell that extends from Honduras and Nicaragua eastward through Jamaica to the western end of Hispaniola. It is bounded on the north by the Cayman Trough and on the south by the Columbian Basin. Bathymetrically the rise is clearly defined by the 1000 fathom isobath. It is 650 km wide in the west and tapers to less than 70 km in the east. Approximately 25% of it is above the 100 fathom isobath of which most lies close to Central America.

Seismic refraction studies (reviewed in Edgar, et al., 1971) have shown the Nicaraguan Rise to consist of a 5.2-5.5 km/sec, 2.5-5 km thick crustal layer overlying a 6.2-6.7 km/sec, 11-16 km thick layer. Crustal thicknesses down to moho (~8.1 km/sec) are 19-22 km. This crustal thickness is significantly less than the average 35-40 km for continental regions and is more comparable to island arcs (see Figure 6 of Miyashiro, 1974a).

Drill hole data on the Nicaraguan Rise is reviewed by Arden

(1975) and indicates a Late Cretaceous-Paleocene volcano-plutonic event. Subsequent erosion and subsidence is indicated by widespread Mid-Eocene carbonates. Between a Paleocene to Eocene period of graben formation and infilling with clastics seems apparent. This generalized model for the geologic evolution of the rise closely parallels that of Jamaica (see p.131).

Extensive dredging of metamorphic, plutonic, volcanic, sedimentary and carbonate rock units by Perfit and Heezen (1978), along the northern slopes of the Nicaraguan Rise also indicate a Cretaceous period of volcano-plutonic activity and subsequent carbonate accumulation since the Eocene.

The data summarized above indicates that much of the Nicaraguan Rise represents a Cretaceous to Paleocene intra-oceanic island arc system. There is no indication or reason to believe that significant portions of the rise are underlain by continental crust. Although, Paleozoic basement exposed in eastern Central America probably extends some distance eastward and underlies the western end of the Nicaraguan Rise (see Fig. 68, p.153). Present-day analogues to this Cretaceous transition from the continental crust of Central America to an intra-oceanic arc may be analogous to the transition from the Alaskan Peninsula to the Aleutian Islands and that from the Sunda arc to Banda arc (Hamilton, 1977).

The geologic evolution of the Nicaraguan Rise indicates that once vulcanism and associated magmatism ended in Early Cenozoic times it subsided as a unit. This evolution is thought to be analogous to that of remnant arcs (Karig, 1970; 1972) and farther east a similar history has been documented for the Aves Swell (Fox and Heezen, 1975).

It can be postulated that island arcs with protracted histories of magmatism "ride" near or above sea level, like continental crust, while island arcs with shorter histories of magmatism (or generation of granodiorite) are not bouyant enough to maintain themselves above sea level and hence subside isostatically.

Cayman Ridge

The Cayman Ridge (Fig. 2) is an approximately 800 km long, narrow submarine ridge that separates the Cayman Trough from the Yucatan Basin to the north. The ridge parallels the trend of the Cayman Trough and is a western continuation of the Sierra Maestra of Cuba were once part of the Nicaraguan Rise and have since been rifted away (Bowin, 1968; Perfit and Heezen, 1978). An Eocene age has been suggested for this event. This age seems reasonable and it is here suggested that the initiation of the Mid-Cayman Spreading Center (see Fig. 66) caused the Cayman Ridge to be obliquely rifted away from the Nicaraguan Rise. This may indicate an Eocene age for the inception of the Mid-Cayman Spreading Center (see Chapter VIII for further discussion).

Jamaica

Recent models for the Cretaceous geologic evolution of Jamaica have generally invoked island arc models (but see Meyerhoff and Krieg, 1977 for an alternate interpretation). Features typical of island arc systems, e.g. subduction complex and associated blueschist metamorphics and ultramafics, calc-alkaline volcanics and arc related sedimentary basins have all been reported by recent workers (Draper, et al., 1976;

Roobol, 1976; Donnelly and Rodgers, 1978; Grippi and Burke, 1978). Thus far the oldest dated sediments on Jamaica are (?)Barremian-Albian limestones of the Benbow Inlier (Burke, et al., 1968). Below this horizon of Lower Cretaceous limestone is a 1 km sequence of volcanics and volcaniclastics. There is no direct indication that Jurassic or older rocks occur on Jamaica. There is also no indication that continentally derived sediments or continental crust underlies Jamaica. Hence Jamaica is interpreted to represent an intra-oceanic arc. The Cretaceous volcanic edifice of Jamaica is thought to be represented by in situ Mid to Late Cretaceous subaerial and submarine volcaniclastics and associated shelf limestone lenses that are exposed in the Benbow (Burke, et al., 1968), Central (Coates, 1968; Roobol, 1976) and Blue Mountain (Krijnen and Lee Chin, 1978) inliers. Relatively small granodioritic plutons of Late Cretaceous age (Chubb and Burke, 1963; Lewis, et al., 1973) within the Central, Benbow, Above Rocks and Blue Mountain inliers are interpreted as representing cores of volcanic centers. Metamorphic rocks of the southern and southwestern end of the Blue Mountain Inlier and within the Lazeretto Inlier (Chubb, 1954), to the southwest may represent fragments of a subduction complex (Mount Hibernia schist) and (?)regional metamorphics associated with the volcanic arc (Westphalia schist) (Draper, et al., 1976). Mineral assemblages containing crossite, riebeckite and stilpnomelane within the Mount Hibernia schist suggest that it was formed within a subduction zone (Miyashiro, 1974b). Mineral assemblages indicative of the epidote-amphibolite facies are developed within the Westphalia schist and may indicate a high T/P regime typical of island arc edifices of paired metamorphic belts (Miyashiro, 1961) as suggested by Draper, et al. (1976). However, metamorphic rocks of greenschist and epidote-amphibolite grade are developed within the classical high P/T Sanbaqawa metamorphic belt (Miyashiro, 1974b). Presently, the metamorphics of Jamaica are spatially associated with granodioritic plutons, though contacts between units are typically fault bounded or unexposed (Draper, personal commun., 1977). The extensive Cretaceous to Recent faulting has apparently modified the original spatial relationship between the subduction complex and volcanic arc (Dickinson, 1973), though a complex geometry of subduction zones and associated volcanic arcs may easily explain the present proximity of volcanic arc granodiorites and blueschists. Recent work by Hamilton (1977) and Silver and Moore (Fig. 2, 1978) within the Indonesian region and theoretical work by Dewey (1975) clearly indicate the complexities that may be expected for convergent boundaries.

Due to limited exposure and continued Cenozoic tectonism the polarity of Cretaceous subduction for Jamaica is not easily decipherable. A southerly dipping subduction zone has been suggested by several recent workers (Perfit and Heezen, 1978; Krijnen and Lee Chin, 1978; Draper, et al., 1976; Horsfield and Roobol, 1974). Though others have suggested a northerly dipping subduction zone (Grippi and Burke, 1978; Burke, et.al., 1978). A north-dipping subduction zone is preferred here because of the southerly location of blueschist metamorphics in Jamaica and their apparent continuity with obducted slivers of ophiolite in parts of the Southern Peninsula of Haiti (Maurrasse, et al., 1977). It is also possible that the Nicaraguan Rise and Jamaica represent multiple arc-trench systems that have been sutured together. If a north-dipping subduction zone is accepted for the eastern half of the Nicaraguan Rise then a transform of unknown orientation would have to connect it with the southerly dipping subduction zone of Central America.

The Paleocene is not represented by fossiliferous sediments on the island of Jamaica and may represent a period of widespread emergence and erosion. Though the lack of a Paleocene section may be due to limited resolution of paleontological data. A major (?)Paleocene to Lower Eocene rifting event developed two northwesterly-trending grabens in Jamaica, the Wagwater, of eastern Jamaica and the Montpelier-Newmarket trough in the west. The Paleocene to Eocene section of the Wagwater Trough is well exposed and hence has received much attention (Green, 1977; 1974; Holliday, 1971; Jackson and Smith, 1978). The rift contains a 5 km thick, folded and faulted volcanic and clastic sequence. Clastics dominate the section and consists of a conglomeratic facies (Wagwater) and a laterally equivalent and overlying turbiditic sandstone and shale facies (Richmond). Clasts within conglomerates for the lower 2/3 of the section are of volcanic derivation, though granodioritic clasts are present in the upper part of the sequence. This relationship records the unroofing of the (?)defunct volcanic arc centers to the west and (?)east. Pillowed basaltic-spilitic volcanic rocks (Halberstadt volcanics) occur near the base of the section while more silicic dacite-quartz keratophyres occur as massive flows and volcaniclastics (Newcastle Volcanics) midway in the section. Geochemistry of the Newcastle and Halberstadt volcanics suggest that they are calc-alkaline and slightly alkaline in character, respectively (Jackson and Smith, 1978). Burke (in Burke and Fox, 1977) has suggested that the Wagwater Trough represents a small inter-arc basin (Karig, 1970) in which the basal Halberstadt Volcanics represent an extensional, rifting phase while the overlying Newcastle Volcanics represent vulcanism associated with the closure of this small basin. This model requires significant amounts of folding and thrusting of the troughs fill (Burke and Robinson, 1963) and thus differs from the present models of Green (1977) which suggest only vertical movements and associated high angle faulting. Reconnaissance work by Burke and Grippi (1977) and geologic relationships suggests that the model presented by Burke is viable but will have to wait until more data is available on the structural relationship between key lithologies.

The post-rifting evolution of Jamaica is one of submergence and accumulation of shallow-water carbonates from Mid-Eocene to Mid-Miocene times. This submergent event was of regional extent for the Nicaraguan Rise (see p.130 for discussion).

Cuba

The Early Cretaceous to Eocene geological evolution of Cuba records the consumption of oceanic crust between it and the Bahama Platform, along a southward-dipping subduction zone. Extensive tracts of ophiolitic mélange along the north coast clearly marks the suture between Cuba and the Bahama Platform (see Fig. 2 of Kozary, 1968). Geologists working in Cuba have usually subdivided the island into linear belts (see Meyerhoff and Hatten, 1967 for review). Within central Cuba three major linear belts can be depicted, from north to south, a carbonate platform and south-facing scarp facies, deep water pelagic realm and an arc-trench association (see Pardo, 1975, for details). These three tectofacies have been juxtaposed by consumption of oceanic crust throughout the Cretaceous and to extensive imbrication and telescoping during the suturing of the Bahama Platform with Cuba, during Early to Mid-Eocene times. This interpretation is based on data published by Kozary (1968); Wassall (1957); Meyerhoff and Hatten (1967); Pardo (1975).

Though it was shown as early as 1957 by Wassall that volcanics, serpentine and metamorphics were thrust over a sequence of marine limestone section in the Santa Clara region of central Cuba. It has been suggested by J. Grippi (unpublished manuscript) and Gealey (1977) that the extensive tracts of ophiolitic mélange along the north coast of Cuba may represent the basement of the fore arc region of Cretaceous Cuba that was intensely imbricated and obducted during suturing.

The presence of Oxfordian and older continentally-drived shelf and (?)rise (Pszczółkowski, 1977) sediments (San Cayetano Oxfordian. Formation) and high T/P grade quartz-rich metamorphics of unknown age suggests that parts of Cuba are underlain by slivers of continental crust. Associated with the high T/P metamorphics are metamorphics with mineral assemblages indicative of high P/T conditions (Boiteau, et.al, 1972). This relationship suggests that Cuba was also affected by an older (?)Jurassic episode of convergence. Presently Cuba is surrounded by oceanic realms and defunct intra-oceanic arcs. It is apparent then that slivers of continental crust within Cuba were derived from Central America or South America during a major Jurassic rifting event (Burke and White, 1978). Volcanics and extensional structures within Jurassic sequences of western Cuba (Piotrowski, 1977) may be an expression of this Jurassic rifting event. H. Yarborough (personal commun., 1977) has suggested that Cuba was derived from the Yucatan Peninsula, though it appears that the exact derivation of continental slivers within Cuba may have to wait until more data is available on their character.

Hispaniola

The island of Hispaniola is probably the least known geologically of the Greater Antilles. The most recent review of the Cretaceous geology of Hispaniola was made by Bowin (1975). Most recent views on the plate tectonic evolution of Hispaniola have suggested a trench-trench-trench triple junction for the present site of Hispaniola (Burke, et al., 1978). This interpretation seems reasonable and may be the explanation for the complex geology of Hispaniola. An east-west trending blueschist metamorphic belt along the north coast of Hispaniola and a high T/P metamorphic belt, 80 km to the south, indicate a south-dipping subduction zone for northern Hispaniola (Nagle, 1974). Geologic age relationships suggest that subduction occurred from Cretaceous to Eocene times. The occurrence of obducted slivers of Caribbean oceanic crust and overlying Upper Cretaceous sediments within parts of the Southern Peninsula of Haiti have been reported by Maurrasse, et al. (1977) are thought to be related to a north-dipping subduction zone.

Puerto Rico

Cretaceous to Mid-Eocene volcaniclastics and underlying grano-dioritic plutons of Puerto Rico indicate island arc vulcanism for this period (Mattson, 1972; Glover, 1971; Donnelly, et al., 1971; Cox, et al., 1977). Radiolarian chert, serpentinized peridotite, pillow lavas and tholeiitic amphibolite (Lee and Mattson, 1974) of the Bermeja Complex (Mattson, 1960) of southwestern Puerto Rico probably represents a dismembered ophiolite assemblage that formed within a subduction zone. Volcaniclastics are complexly intercalated with the Bermeja Complex. Contacts are assumed to be thrusts, due to a northward dipping subduction zone that consumed Caribbean oceanic crust (Burke, et al., 1978; see Mattson, 1973 for an alternate interpretation). Upper Jurassic ages of radiolarian chert (Mattson and Pessagno, 1974) within the Bermeja Complex indicate at least a Jurassic age for consumed oceanic crust. Obviously this single age does not place an upper or lower limit on the age of oceanic crust consumed or for the duration of subduction.

Extensive (?)gravity driven north vergent imbricate thrusts within the Upper Cretaceous volcaniclastics and limestone of south-central Puerto Rico have been reported by Glover (1971). In the framework of a northward-dipping subduction zone for Puerto Rico these thrusts may be equivalent to arc vergent thrusts of fore-arc basins as reported by Hamilton (1977).

The Puerto Rican trench to the north is probably an early Tertiary feature (Monroe, 1968) that was initiated and is presently maintained by left-lateral strike-slip motion across the northeastern North American-Caribbean plate boundary.

South America

The Jurassic to Eocene geologic evolution of the Caribbean

Mountain system of Venezuela has been reviewed and interpreted in terms of plate tectonics by Maresch (1974). Maresch (1974) suggested a north-facing Atlantic-type margin and offshore south-facing intra-oceanic island arc system for the Jurassic to Lower Cretaceous. A mid Cretaceous flip in subduction and continued subduction caused for this offshore arc to collide with the continental margin of South America. This suturing event along a southward-dipping subduction zone caused for the Villa de Cura (blueschist and greenschist metavolcanics-subduction complex) to be obducted onto the continental margin of South America. Continued compression caused for a new south-dipping subduction zone to form outboard of the present Netherland Lesser Antilles. Calc-alkaline igneous activity of Late Senonian to Early Tertiary age are evidence of this Late Cretaceous event (Beets and MacGillavry, 1977).

The emplacement of the Villa de Cura Complex along north-dipping thrusts planes and a concomitant southward-dipping subduction zone seems mechanically unsound to the author and does not appear to be consistent with recent models for other orogenic belts. Geologic relationships from the Oman Mountains (Gealey, 1977), Bay of Islands, Newfoundland (Nelson and Casey, in press), Cuba (see p.135), British Caledonides (Mitchell, 1978) and the Timor region (Hamilton, 1977; Carter, et al., 1976) suggest that fragments of island arc that have been obducted onto a continental margin can best be explained by a subduction zone dipping away from the continent.

Caribbean Sea

An analysis of magnetic anomalies within the North and South

Atlantic by Pitman and Talwani (1972) and Ladd (1976, Fig. 2A) has shown that Caribbean ocean floor was formed in Jurassic to Cretaceous times. This concept has already been expressed by Burke, et al. (1978). Within the Caribbean, rifting of South America away from North America is indicated by evaporites along the north coast of South America and the Northern Range of Trinidad (Burke and White, 1978). Mafic vulcanism and high level intrusives of Mid-Upper Jurassic age within Cuba are thought to be associated with . . . "Lower? Middle Jurassic deep tension fractures of rift character that formed at the southern margin of Cuba." (Piotrowski, 1976, p. 233). These features within Cuba may also be an expression of widespread rifting between North and South America.

Exclusive of the Yucatan Basin the crustal character of the Caribbean Basins have been extensively studied by many institutions, particularly by Lamont-Doherty Geological Observatory (see Fox and Heezen, 1975; Case, 1975, for review). The true crustal character of the Caribbean basins is still not well understood but results of Leg 15 of the DSDP have prompted fresh new ideas (Edgar, et al., 1973; Fox and Heezen, 1975). In essence the Caribbean is now considered to be anomalously shallow, thick (~15 km, see Edgar, et al., 1971) oceanic crust that was subjected to a Late Cretaceous (80-75 mybp) basaltic "sill" event (see Houtz and Ludwig, 1977 for alternate interpretation). With this interpretation in mind, Burke, et al. (1978) have suggested that the Columbian and Venezuelan basins of the Caribbean (see Fig. 2) are hard-to-subduct remnants left after a major Jurassic to Eocene convergent event within the Caribbean. They have also suggested that the Ontong Java and Manihiki plateaus of the Pacific Ocean may be analogous thickened and bouyant "oceanic" crust.

The meager seismic refraction data of the Yucatan Basin indicate that it has a crustal velocity structure and thickness that is analogous to "normal" oceanic crust (Ewing, et al., 1960 in Fox and Heezen, 1975).

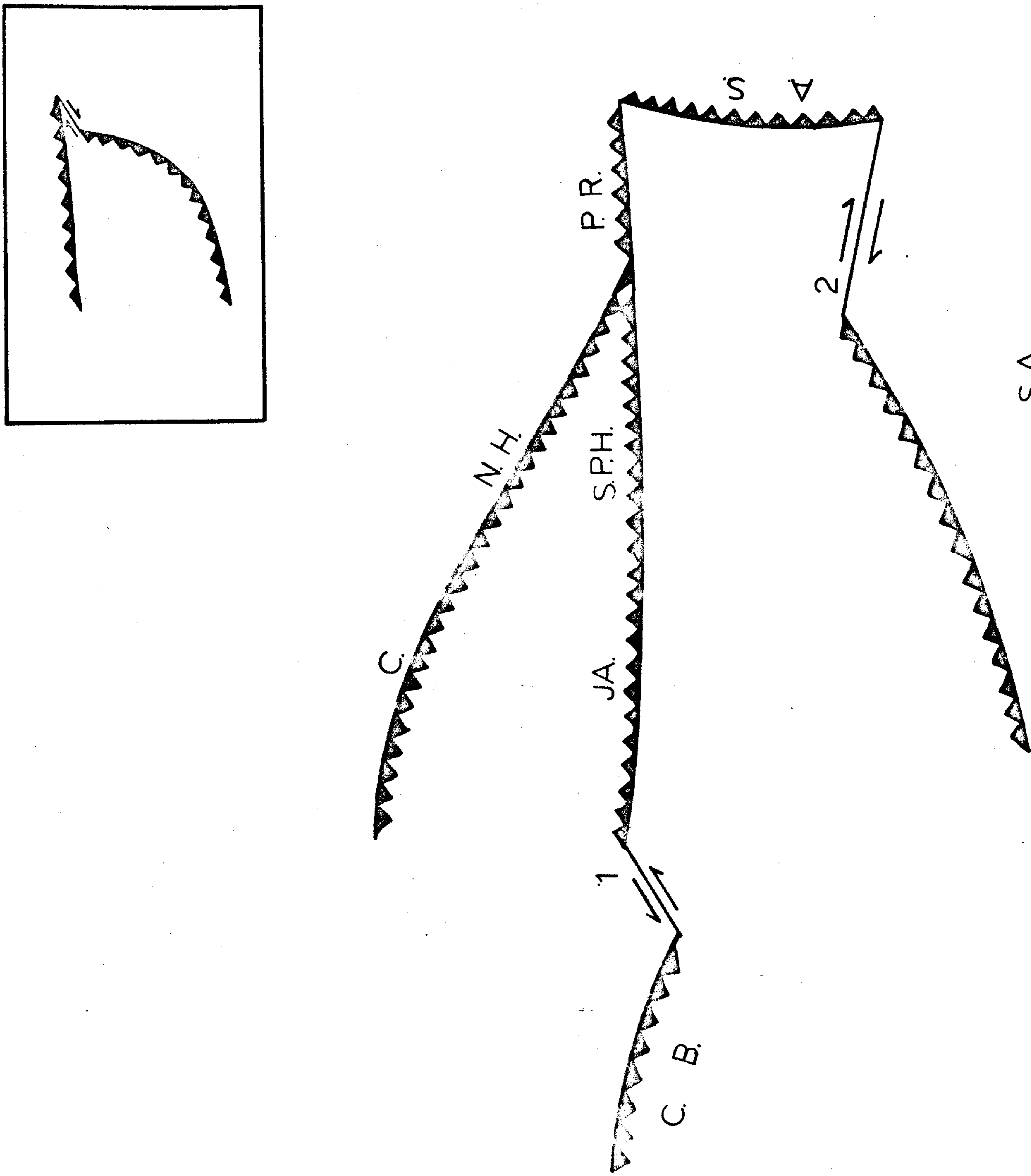
For a discussion of the Cayman Trough, please refer to Chapter IX.

Conclusion

Widespread Jurassic to Eocene vulcanism throughout the Caribbean probably occurred along a complex network of consuming plate boundaries. Figure 64 shows a possible geometry of Mid-Late Cretaceous subduction zones for the Caribbean region. The trench-trench-trench triple junction within the Greater Antilles has been suggested before (Burke, et al., 1978) and is followed here. The present Aves swell is probably the site of a Late Cretaceous arc (Fox and Heezen, 1975), though available data does not indicate whether Caribbean or Atlantic oceanic crust was consumed at this arc. Here, I have suggested an east-dipping subduction zone. A north-dipping subduction is postulated for South America in order to explain the position of the Villa de Cera complex. The plate junction between the Aves swell and subduction zones to the north and south is one alternative and is in part dependent upon the polarity of subduction zone believes for the Aves swell. An alternative geometry is shown in the inset. This alternative model suggests that the Aves swell arc-trench system was continuous with the one off South America. Progressive subduction would cause for an arc-continent collision to sweep diachronously from west to east across the northern margin of South America. The Aves swell would then represent an unsutured segment of this once larger arc system. A similar model was proposed by Maresch (1976, Fig. 2). A DSDP hole along the eastern and western flank of the Aves swell may clarify the polarity of Cretaceous subduction (Biju-Duval, et al., 1978). The eastern and western extension of plate boundaries within Figure 64 are not known.

Figure 64 - Possible geometry of Mid to Late Cretaceous subduction zones for the Caribbean region. Base map is a pre-Oligocene restoration of the Caribbean plate (see Chapter XIII, Fig. 66). Teeth are on the upper plate. Transforms 1 and 2 progressively shorten with time. See text for discussion.

C - Cuba, N. H. - Northern Hispaniola, P. R. - Puerto Rico, C. B. - Chortis block, JA. - Jamaica, S.P.H. - Southern Peninsula Haiti, A. S. - Aves Swell, S. A. - South America.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CHAPTER VIII

CAYMAN TROUGH AND THE CARIBBEAN

The Cayman Trough is now accepted as the boundary between the North American and Caribbean plates (Molnar and Sykes, 1969). The importance of it to the Cenozoic tectonics of the Caribbean region has been stressed by many geologists (Hess, 1938; Bucher, 1947; Hess and Maxwell, 1953; Eardley, 1954; Wilson, 1966; Bowin, 1968). Hess (1933, in Molnar and Sykes, 1969; 1938) was the first to suggest a Late Miocene to Recent left-lateral offset along it. Hess and Maxwell (1953) later postulated a total displacement of 1,100 km across it. More recent estimates on the total amount and rate of displacement have ranged from 10's-1000 km and 4-60 mm/yr, respectively (Molnar and Sykes, 1969; Pinet, 1972; Malfait and Dinkelman, 1972; Uchupi, 1973; Holcombe, et al., 1973; Jordan, 1975; Perfit and Heezen, 1978).

Recent work on the seismicity (Molnar and Sykes, 1969), crustal structure (Bowin, 1968; Edgar, et al., 1971) topography and sediment distribution (Holcombe, et al., 1973) and petrology and geochemistry of dredged rocks from the Cayman Trough (Perfit, 1977; Perfit and Heezen, 1978) has convincingly shown the Cayman Trough to contain a central 110 km long north-south oriented spreading center with very long northern and southern bounding transforms, Oriente and Swan, respectively (see Fig. 2). Seismic refraction and dredging within the Cayman Trough have shown it to be floored by normal oceanic crust (~ 6 km thick). Seismic refraction studies (Ewing, et al., 1960, in Perfit and Heezen, 1978; Arden, 1975) on the north-bounding Cayman Ridge and south-bounding Nicaraquan Rise have shown them to have intermediate crustal thicknesses (19-22 km). Extensive dredging by Perfit and Heezen (1978) of metamorphic, plutonic, volcanic, sedimentary and carbonate rock units along the inward facing scarps of the Cayman Ridge and Nicaraquan Rise suggest that their Cretaceous geologic development is similar to the rest of the Greater Antilles, and record convergence.

A recent analysis of the magnetic anomalies within the Cayman Trough, by Macdonald and Holcombe (1978), has quantified the rate of displacement across the Cayman Trough. They calculated a full spreading rate of 20 mm ± 2 mm/yr for the interval 0-2.4 mybp and 40 mm ± 2 mm/yr for 2.4-6.0 mybp. However, magnetic anomalies back to 8.3 mybp (anomaly 4') were identified on the west flank. Extrapolation of the 40 mm/yr rate by Macdonald and Holcombe (op. cit.) yields an age of 30 mybp for the initiation of the Mid-Cayman Spreading Center (MCSC). It should be stressed that this approximate age, if correct, is only a minimum age for the initiation of large strike-slip motion along the Cayman Trough.

The Cayman Trough is dominated by ridge related north-south lineated topography (Holcombe, et al., 1973); however, the eastern extremity of the Trough contains northeast-southwest trending ridges (see Case and Holcombe, 1976). These northeast-trending structures may represent primary extensional structures associated with sinistral slip across the Cayman Trough, prior to the inception of the MCSC (Fig. 65). Analogous extensional structures, as interpreted by the author, are developed immediately west of Israel (Mart, et al., 1978) and are related to slip along the Dead Sea transform fault. Similar structures were discovered during a near-bottom study of the Quebrada transform fault (Lonsdale, 1978). The MCSC may have nucleated along

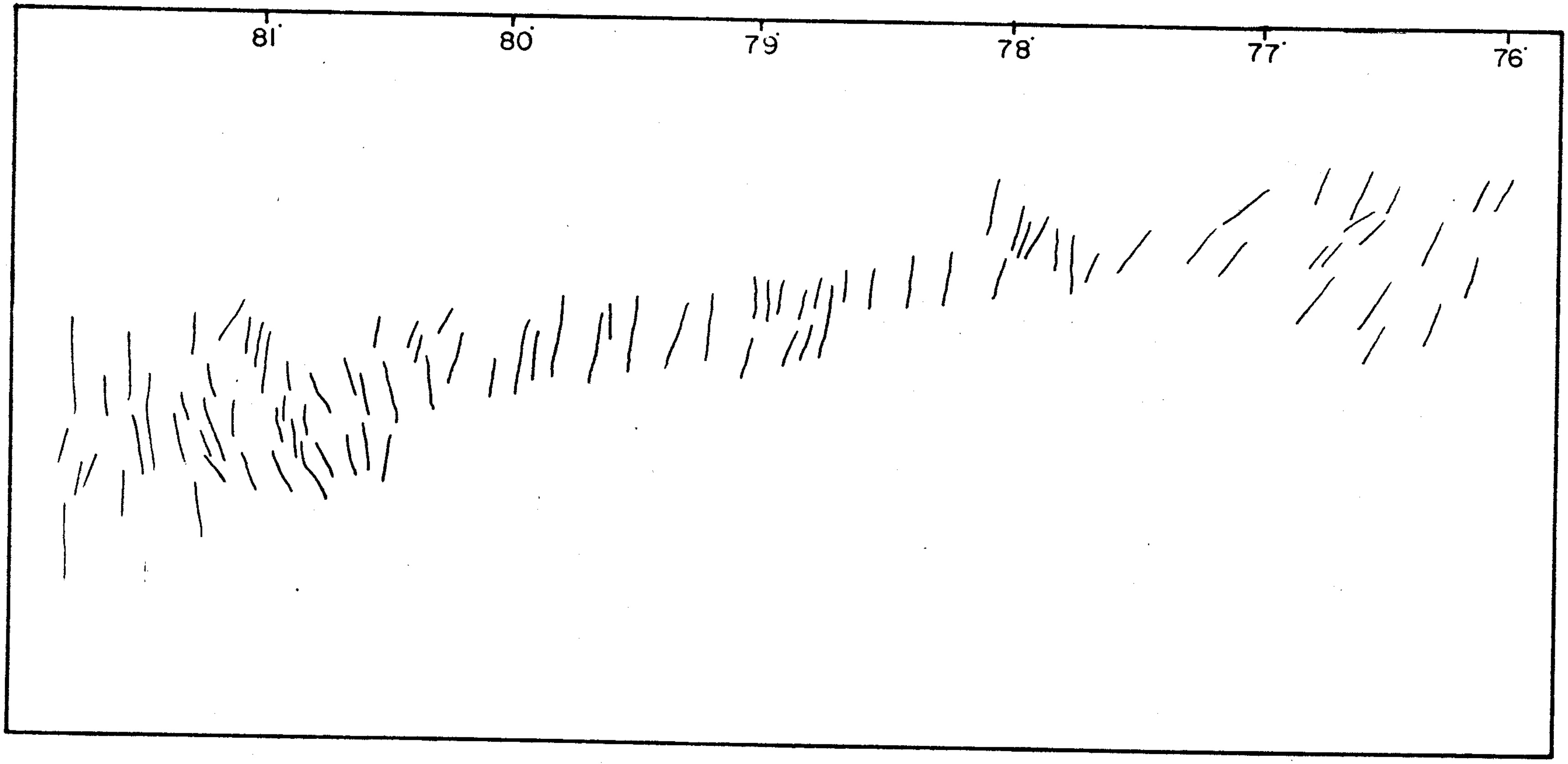


Figure 65 - Structural lineaments within the eastern end of the Cayman Trough. Note abrupt change in trend of lineaments from north-south to a northeast-southwesterly trend. See text for interpretation. Data from Case and Holcombe (1976).

such extensional structures that were actively being rotated counter-clockwise into a north-south orientation (Tchalenko, 1970; Wilcox, et al., 1973). Continued extension may cause sufficient thinning and necking of the lithosphere so that a small ridge link connecting long transforms comes into existence. Alternatively, these high relief northeast-southwest trending ridges have been interpreted by Macdonald and Holcombe (1978) to possibly represent an earlier opening period of slower, episodic and/or of a different orientation.

In order to understand the Cretaceous tectonic evolution of the Greater Antilles region the post-Cretaceous displacements along the Cayman Trough have to be removed. This has been done by simply post-ulating that all north-south lineated topography within the Cayman Trough is newly accreted oceanic crust. This north-south lineated topography extends from 85°10'W to 77°40'W and represents a total dis-placement of approximately 720 km between the Caribbean and North American plates. This may be a minimum value because strike-slip motion across the Cayman Trough could have begun prior to the incep-tion of the MCSC. This simple analysis assumes that both plates have been torsionally rigid during this rotation. Burke, et al. (1978) have recently suggested that the Caribbean plate east of the Beata Ridge, has been experiencing internal deformation from the Early Oli-gocene to Early Miocene (38-9 mybp). If this were the case then some strain associated with slip along the Cayman Trough would be taken up by internally deforming the Caribbean plate. This would tend to lessen the displacement along the Cayman Trough for this time interval.

The instantaneous pole of rotation, at 33.83°S, 70.48°W, drived by Minster and Jordan, in press (in Macdonald and Holcombe, 1978), for the Caribbean and North American plates was used to restore the two plates to their positions at the time of inception of the MCSC. This rotation is a simple finite rotation (Dewey, 1975). The MCSC was held fixed in performing the rotation on a large Wulff stereonet (75 cm in diameter) (Fig. 66).

In making this rotation the North American-Caribbean plate boundary had to be defined for the finite period in question. This boundary is apparently well defined within the Cayman Trough, however, its eastern and western extensions through the Greater Antilles and Central America are not so unambiguous. However, our knowledge of transform faults intersecting continental crust tells us that they are wide, diffuse zones of deformation, unlike sharp, intra-oceanic plate boundaries. At present there is no evidence (e.g., kinks within transform segments) to suggest that there has been a change in the pole of rotation for the Caribbean-North American plate system (P.J. Fox, personal commun., 1978). Therefore, the eastern and western paleo-traces of the Oriente and Swan transform, for this finite rotation, should closely approximate their present trends.

The geological implications for the 720 km of left-lateral displacement is considered below for Central America and the Greater Antilles.

The paleo-trace of the Swan transform through Central America probably passed, in a complex way, through the Motagua suture zone; similar to the present Cuilco-Chixoy-Polochic and Motagua fault zones of Central America (see Fig. 67).

The Motagua suture zone (see p. 127) of Maastrichtian to Paleocene age mandates that terranes north and south of it had to have been in

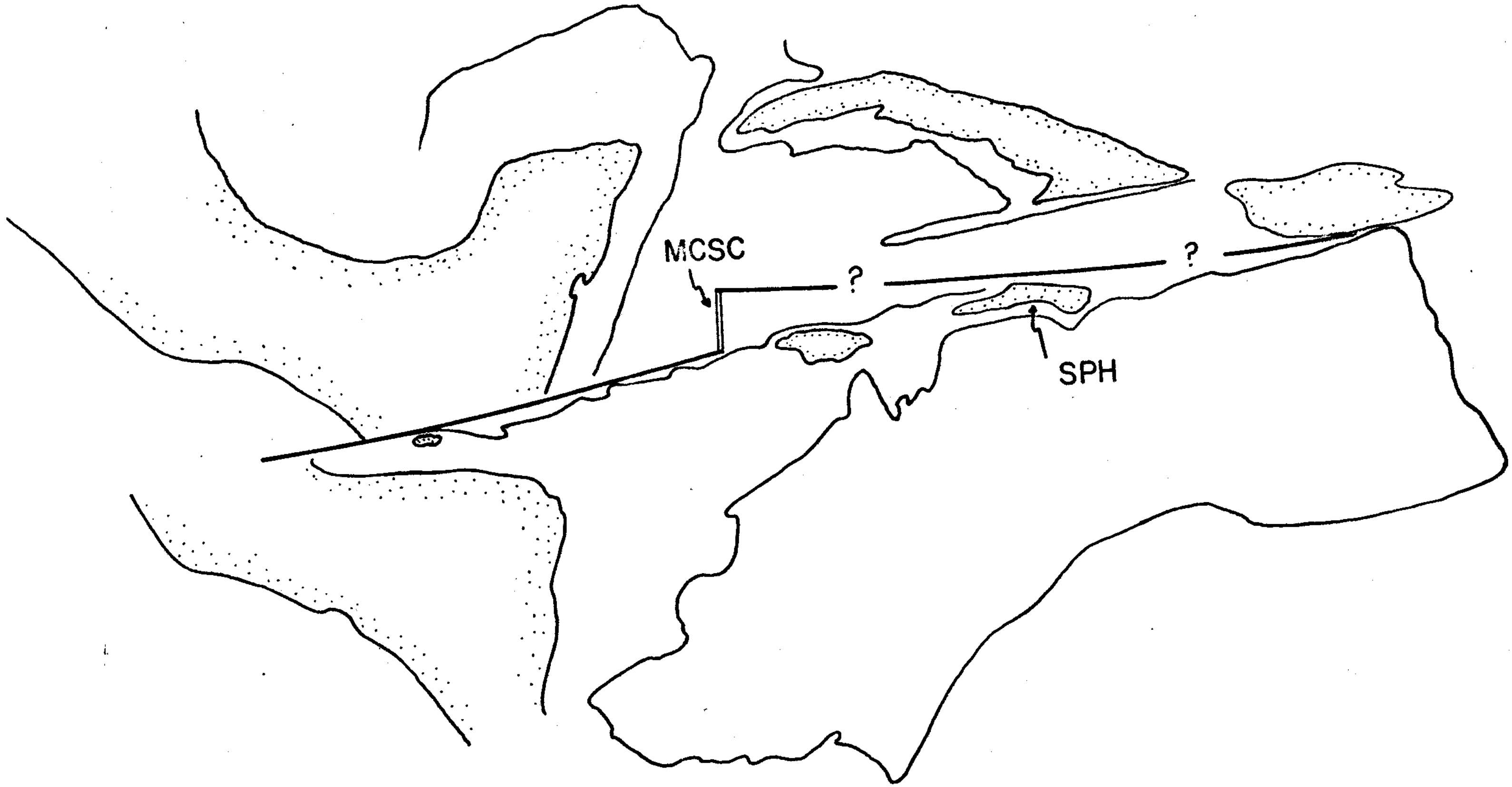


Figure 66 - Oligocene reconstruction of the northern Caribbean plate boundary. Stippling outlines present land masses. 1000 m isobath is outlined. Double ruled line represents the MCSC and bold lines represent east-west trending transforms. Eastern extension of transform through the Greater Antilles is greatly simplified. See text for discussion.

contact prior to the left-lateral offset associated with the Cayman Trough. However, the original length and modification the suture zone has suffered since its formation is not known. Figure 67 is a simplified map of Central America displaying the present distribution of suture-related lithologies and Paleozoic or older basement. The plate tectonic interpretation of the Cayman Trough presented stipulates that the northern "Guatemala" block has moved approximately 720 km eastward relative to the southern Chortis block. This amount of displacement has occurred since Eocene-Oligocene times. Figure 68 shows the Oligocene positioning of these blocks and outlines the extent of possible Paleozoic basement (see p.130). At first this 720 km of left-lateral displacement seems unreconcilable with the geology of Central America (compare Figures 67, 68). A major problem with this reconstruction is that the present-day emergent portion of the Chortis block is not in contact with the Guatemalan block. This seems totally unreasonable because suture zone related geology (ophioltic rocks, eclogite and blueschist metamorphics) dominates this region in its present position. A possible solution to this problem is to postulate that the original Late Cretaceous-Paleocene suture zone extended from Point A to B on Figure 68; and since has been lengthened and modified by wrench tectonics (G. White, personal commun., 1978). Intra-continental transforms typically have very complicated braided and sigmoidal traces. They are generally very wide zones (hundreds of kilometers) of deformation. During the proposed minmum 720 km of left-lateral displacement the Cuilco-Chixoy-Polochic-Motagua fault zones could easily have repeated and/or transported structural blocks along its length. In well studied orogenic belts it is clear that large stike-slip faults, closely

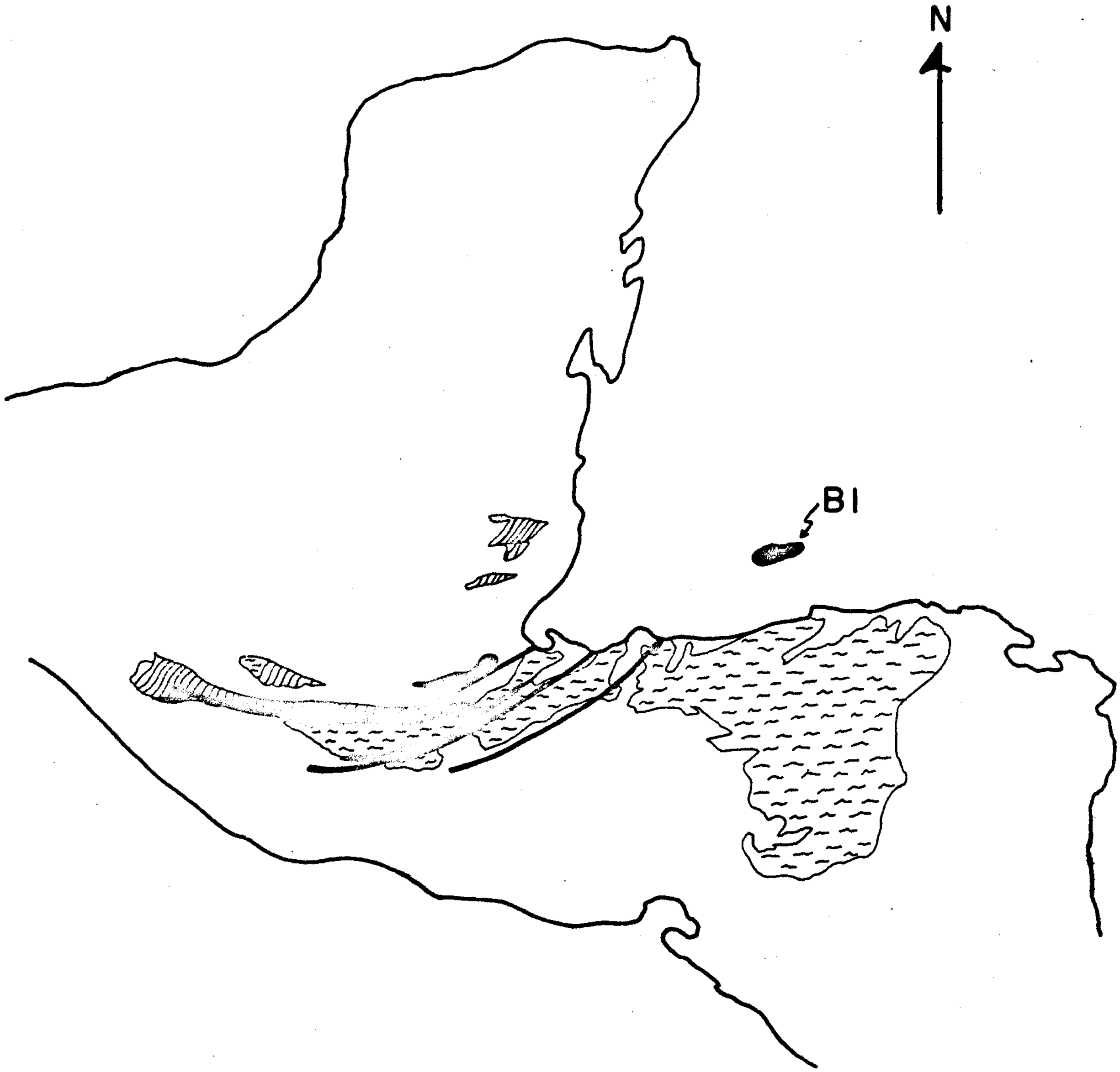


Figure 67 - Simplified geologic map of Central America showing the distribution of Paleozoic basement and Maastrichtian-Paleocene suture related serpentinite, in black

Paleozoic or older basement, Paleozoic sedimentary cover. Bold lines represent major fault traces (from Dengo, 1973). BI - Bay Islands.

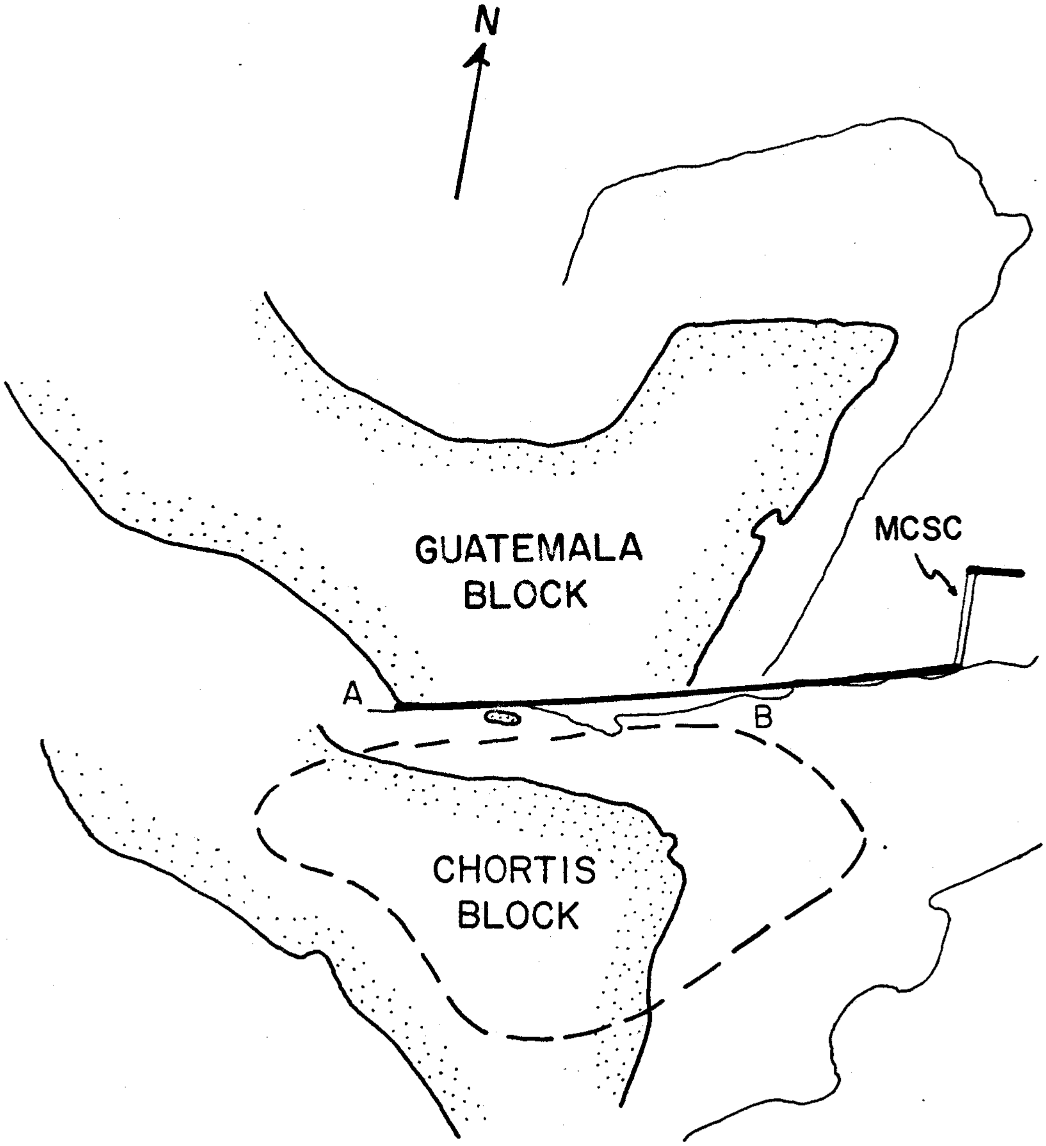


Figure 68 - Oligocene reconstruction of the northwestern Caribbean-North American plate boundary . Stippling outlines present land masses. 1000 m isobath is outlined. Dashed line outlines possible extent of continental crust forming the Chortis block. A-B may represent the original length of the Maastrichtian-Paleocene Motaqua suture that welded the Chortis and Guatemalan blocks. See text for discussion.

paralleling their lengths, have repeated and transported tectonic belts for great distances, e.g., California and New Zealand. This is only one possible way to explain the apparent discrepancies of Figure

The eastern extension of the paleo-Oriente transform probably passed through Hispaniola as a wide zone of deformation. The complex Cenozoic geology of Hispaniola attests to this interpretation. The apparent continuity of the subduction complex of eastern Cuba (Kozary, 1967) with the one exposed along the north coast of Hispaniola (Nagle, 1974) suggests that not much relative displacement has occurred between Cuba and Hispaniola. Both subduction complexes are of Cretaceous to Mid-Eocene in age. Much of the 720 km of left-lateral displacement probably occurred farther to the south. Figure shows this offset to have passed through the Enriquillo graben of Hispaniola. This interpretation is very similar to what Hess and Maxwell (1953) had postulated more than 20 years ago. However, they inferred an approximate offset of 1100 km. It would seem more reasonable to postulate that displacement was not taken up by a single fault plane but by a wide zone in which the locus of displacement shifts with time.

In conclusion, it is envisaged that during the Early Tertiary-Eocene times a major reorganization of plate geometry occurred within the Caribbean region. A change from convergent to strike-slip tectonics along the northern and southern margins of the Caribbean region has been suggested by earlier workers (Malfait and Dinkelman, 1972; Ladd, 1976; Burke, et al., 1978). On a worldwide basis Rona and Richardson (1978) have also suggested that the Early Tertiary was a time of major reorganization of plate motions. The timing of reorganization for the Caribbean is likely to have been diachronous. The switching of vulcanism from the northern and southern Caribbean margins in the Eocene to its eastern and western margins lends strong support to this interpretation, although dredging of basaltic to dacitic volcanic rocks and granitic plutons of Late Cretaceous to Paleocene age (Fox and Heezen, 1975) along the Aves ridge suggests that Upper Cretaceous vulcanism occurred along the eastern margin of the Caribbean. Tertiary strike-slip motion along the northern Caribbean margin was nucleated near the contact between oceanic crust of the Yucatan Basin and Cretaceous arc terrains of the Nicaraguan Rise. This region is a zone of rheologic contrast and weakness, and therefore, a likely region to take up the relative motion of the Caribbean and North American plates. The eastern extension of this fault probably passed through southern Hispaniola. Secondary structures associated with transforms intersecting continental crust (Crowell, 1974) dominated the late ?Eocene to recent geologic evolution of Hispaniola. At this time the island was fragmented into west-northwest to east-southeast trending grabens that were filled with carbonates, salt and conglomerate, sandstone and shale that were shed from intervening cordilleras (Khudoley and Meyerhoff, 1971). The western extension of the Swan transform probably passed through the Motagua suture zone (Donnelly, 1977) and connected up with a subduction zone along the west coast of Central America, in a geometry similar to that of the present day. Continued strike-slip motion along the proto Cayman Trough caused the MCSC to nucleate along an extensional structure, approximately 30 mybp (Macdonald and Holcombe, 1978) and connect with two long east-west trending transform faults. This gross geometry has persisted to the Recent. However, small fluctuations in the pole of rotation have probably caused transforms to migrate or to develop compression and/or extensional segments along their length.

CHAPTER IX

NEOTECTONICS OF JAMAICA

Since approximately Mid to Upper Miocene times Jamaica has been an emergent landmass. Presently it is characterized by high rugged topography (Blue Mountain Peak, 2134 m), uplifted Quaternary reef terraces (Horsfield, 1975), incised rivers, positive Bouguer anomalies, local alkaline vulcanism, seismicity and steeply inclined Pliocene beds. All this data indicates that Jamaica at present is tectonically active. Positive Bouguer anomalies in excess of 100 mgl (Bowin, 1976; Horsfield, 1974) suggest that most of Jamaica is undercompensated and that its topography is being dynamically maintained (Fig. 69). The present tectonism of Jamaica and its emergence as a landmass, approximately 12 mybp (Wadge and Draper, 1977) is thought to be related to strike-slip motion along the Oriente transform (see Fig. 24, p. 4). Work by Horsfield (1975) on raised Quaternary reef terraces in Jamaica, Cuba and Hispaniola has shown that uplifted regions are spatially associated with the Oriente transform.

The outline of Jamaica is defined by northwest-southeast and east-west trending faults which are prevalent on-land (see Fig. 69). In a study of on-land faults Horsfield (1974) suggested that major east-west faults were sinistral while northwest-southeast ones were dextral or normal. A similar fault pattern has been observed by Green (1977) within the Wagwater Trough, though he also noted a northeast-southwest trend. Draper (1977) suggested that post early Eocene movements along the northwesterly trending Yallahs and Blue Mountain faults were sinistral with northeasterly-southwesterly oriented compression for

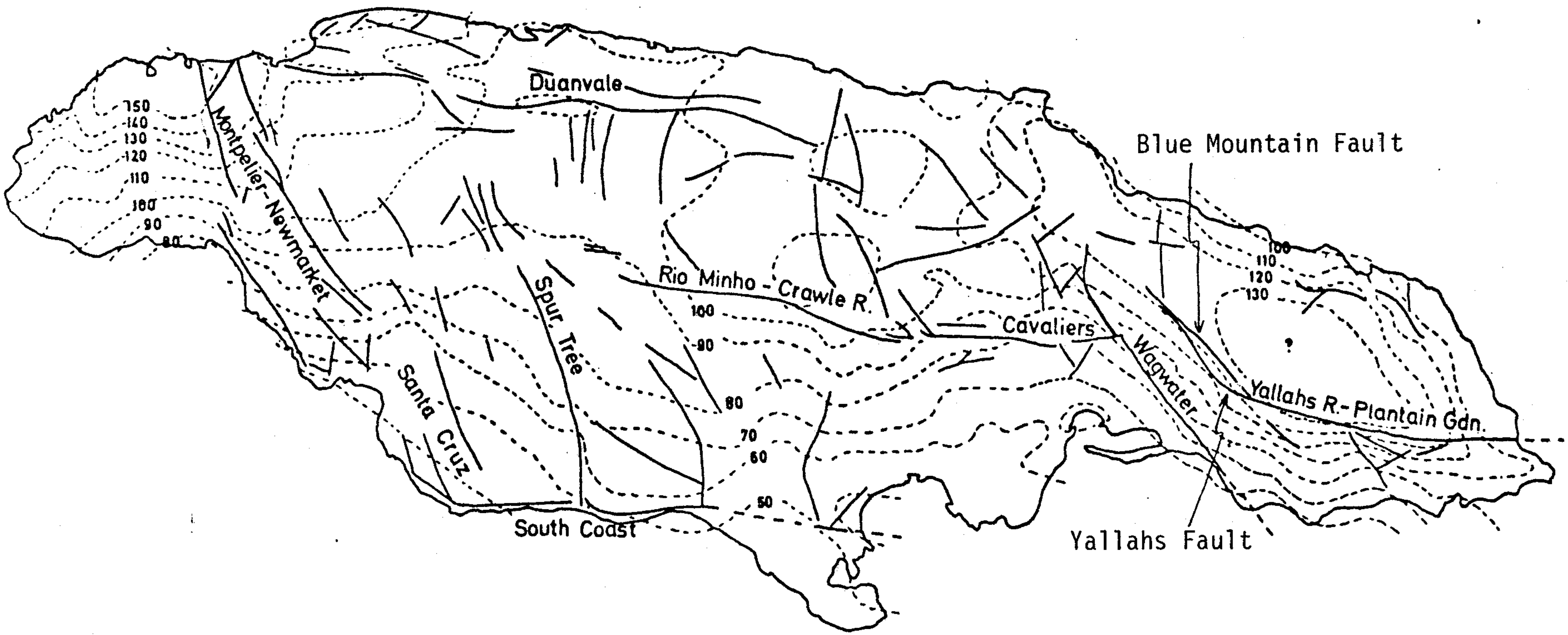


Figure 69 - Fault and Bouguer gravity anomaly map of Jamaica. From Andrew (1969, in Horsfield, 1974)

the intervening block (see Fig.69).

An interpretation of the fault pattern has been recently presented by Burke, et al. (1978) and is discussed below. Their interpretation predicts that east-west trending faults are sinistral while northwest-southeast and northeast-southwest trending structures are compressional and extensional, respectively. Evidence to support this interpretation is presented below.

Northwest-trending structures representing northeast-southwest oriented compression are widespread throughout Jamaica. Figure 70 shows the relationship between major faults and high ground within Jamaica. It is apparent that high ground lies northeast of northwest trending faults. This relationship is particularly well developed in the south-west corner of Jamaica where a series of steep westerly facing cuestas offset Late Miocene bauxite deposits as much as 800 m. Topographically the cuestas have the appearance of block fault mountains, but are here interpreted as high ground related to compression (see Fig.71). In the east, the northwesterly trending Yallahs-Blue Mountain fault zone is a restraining bend as defined by Crowell (1974) which links up the Plantain Garden Fault, in the east with the Rio Minho-Crawle River-Cavalier fault zone to the west (see Fig.69). The Blue Mountains lie northeast of this bend (Fig. 70) and are probably maintained due to northeast-southwest oriented compression. The Transverse Ranges and associated restraining bend of the San Andreas Fault system is an analogous feature (Crowell, 1974). Northwest lineated topography is also developed within the Wagwater Trough (Green, 1977, Figure 9) and the southern Blue Mountains region (Wadge and Eva, 1977). Fold axes with northwesterly-southeasterly trends are developed in the Tertiary

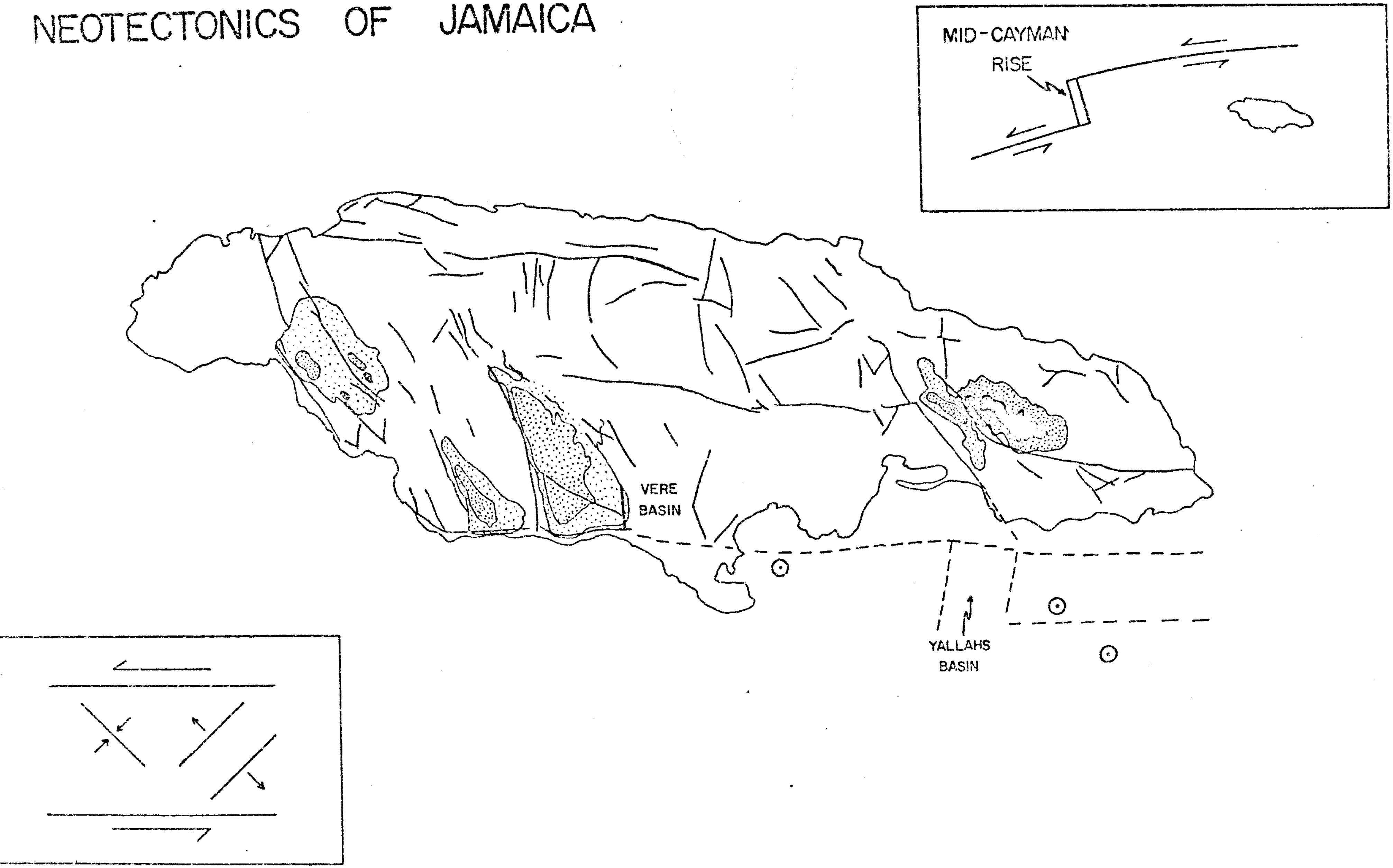
Figure 70 - Lines represent faults and stippled areas high ground. Northwest trending faults are suggested to be compressional. The graben of Vere and the Yallahs Basin lie southwest of bends in the fault pattern.

The fault pattern is interpreted as related to imperfection of transform motion as shown in the inset. Movement on east-west faults in Jamaica is predicted to be largely strike-slip movement.

Fault pattern from Horsfield (1974) and Burke (1967).

⊙ Locations of epicenters

NEOTECTONICS OF JAMAICA



cover of southwestern Jamaica (Wadge and Draper, 1977) and within the Eocene Richmond beds in the vicinity of the Yallahs-Blue Mountain fault zone (Draper, 1977). Northeast-southwest oriented compression has been suggested by Wadge and Draper (1978) for the deformation of the southern Blue Mountains region. Immediately northwest of Kingston Harbour, folding has affected Mid Eocene to Pliocene beds of Dallas and Long Mountain (see geologic map of Jamaica, 1977 and Green, 1977, Figures 2 and 5). These folds have trends similar to 310° and locally have southwesterly overturned limbs. The Yallahs basin (Burke, 1967) located southeast of Kingston (see Fig. 70), is interpreted to be a small, 1.3 km deep, pull-apart basin. The basin contains >.5 km of sediment and is fed by gravel deltas of the Yallahs and Hope Rivers, which drain rugged topography to the north. The lack of an appreciable volume of sediment within the basin and its 1.3 km depth may indicate that it is a relatively young feature. Sedimentation rates of 3 km or less per million years for similar sedimentary basins along the San Andreas fault (Yeates, 1978) may indicate that the floor of the Yallahs basin has subsided with great speed such that sedimentation has not kept pace with subsidence. However, this interpretation assumes the availability of sediment. The subaerial Vere basin to the west may also be an extensional structure (Fig. 70). Robinson and Cambray (1971, in Horsfield, 1974) identified a series of submarine northeast-southwest trending fault bounded troughs and ridges. These structures may also represent extension northwest-southeast orientation.

It is apparent that this structural interpretation of Jamaica does not apply to all regions within Jamaica, e.g. major compressional structures within the Tertiary cover along the north coast trend east-west (Wadge and Draper, 1977). Also directly south of Jamaica, geophysical data on the Pedro Bank (Meyerhoff and Krieg, 1977) shows that its structure is different from Jamaica. This suggests that it is not being affected by slip along the Oriente transform (see Arden, 1969, for discussion). It is felt by the author that domains of structural contrast are to be expected.

Within the Greater Antilles, regions of great relief are spatially associated with the eastern extension of the Oriente transform. Hence, the neotectonic interpretation of Jamaica is based upon the hypothesis that wrench tectonics along the Oriente is dominating the present tectonics of Jamaica. The orientation of fractures and compressional structures on Jamaica are similar to structures predicted by theory for left-lateral strike-slip fault regimes. Experimental (Tchalenko, 1970; and Courtillot, et al., 1974) and field data (Tchalenko and Ambraseys, 1970; and Tchalenko and Berberian, 1975) show similar orientations of secondary structures.

Secondary structures associated with strike-slip regimes are: conjugate riedel shears, P-shears, folds and extensional structures. Figure 72 shows their approximate orientation within a pure strike-slip regime. The relative importance and order of development or destruction of these secondary structures will depend upon the properties of the material, primary structural inhomogeneities, strain rate and the pressure and temperature conditions during displacement. Also, convergence or divergence across the trace of the fault will have an affect on the evolution of secondary structures.

Lucea Inlier

The pattern of faults with dips in excess of 70° for the Lucea

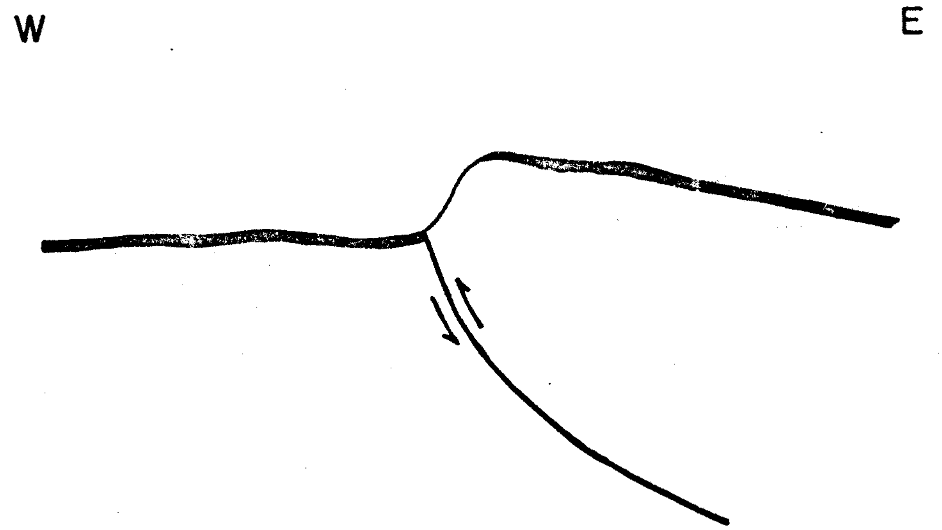


Figure 71 - Thrust fault interpretation of northwest trending cuestas of southwestern Jamaica. Note offset of Late Miocene bauxite(in black), Santa Cruz Mountains. Not to scale.

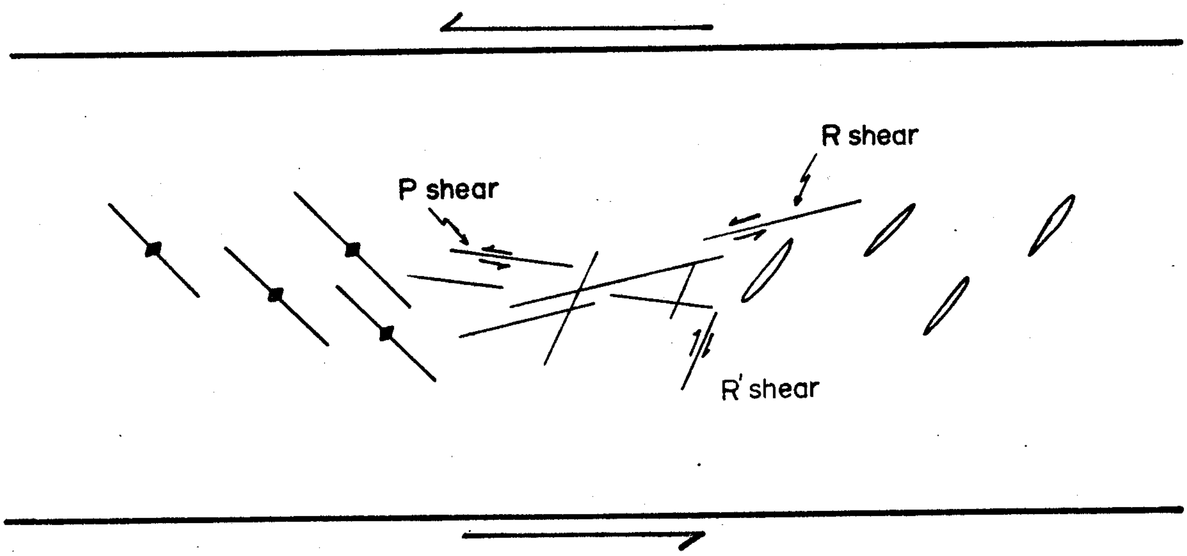


Figure 72 - Orientation of secondary stuctures associated with strike-slip fault zones (compiled from Tchalenko, 1970 and Wilcox et al. 1973).

Inlier is similar to the fault pattern for the rest of Jamaica (compare Figures 55 and 70, pages 160 and 94). East-west trending faults locally show a left-lateral offset and are interpreted as sinistral faults related to the Oriente transform. The type of offset on northwest trending faults is not obvious. Their steep (≥ 70°) inclinations may suggest that they are not thrust related as predicted by Burke, et al. (1978). However, several thrusts with the following orientations: 300°, 20°NE, 300°, 40°NE and 300°, 33°SE. were observed in outcrop. The presence of a northeast-southwest trending normal fault system within the Lucea Inlier is not apparent except for one superb locality. A large 80 m long and 10 m high exposure (1650-5615) within Mosquito Cove is sliced up by a series of normal faults that have an attitude of 035°, 40°SE (Fig. 73). Normal faults are spaced at 1 m intervals and have an average displacement of 30 cm. A simple calculation gives an extension of approximately 25% or 18 m. The lower valley of a river immediately east of this outcrop trends 040° and may be controlled by this extension. These trends are predicted by models of strike-slip fault regimes and add strength to the model presented by Burke, et al. (1978).

Age of Faulting

It is hard to understand the age and history of faulting within Jamaica because many of the faults have probably had a complex history of repeated displacement (Horsfield, 1974).

The existence of Cretaceous faulting is hard to demonstrate for Jamaica due to the overprinting of Cenozoic events. However, Cretaceous left-lateral faulting and associated thinning of the Guinea Corn

Figure 73 – Northeast trending extensional structures within silty shales of the Claremont Formation, Mosquito Cove. See text for description.

Formation along the Rio Minho-Crawle River fault zone has been suggested by Robinson and Lewis (1970). Recent island arc systems commonly show major strike-slip faults paralleling their lengths (Fitch, 1972). It then seems reasonable that Cretaceous faulting should be expected for Jamaica.

Early Tertiary faulting is more apparent for Jamaica. Dramatic facies changes across the Dunvale (Wright, 1968) and other major faults defining the Wagwater and Montpelier-Newmarket Troughs (Green, 1977; Wadge and Eva, 1977; Eva, personal commun., 1977) suggest that faulting was active during the Early Tertiary.

The neotectonic model presented here implicitly implies that displacements are occurring along the present-day fracture pattern of Jamaica. The recent seismicity (Tomblin, 1976) (see Fig. 70) and topography imply on-going faulting within Jamaica.

In conclusion, the model presented here and by Burke, et al. (1978) for the neotectonics of Jamaica is intended to be a working model in which to view new on-land and offshore data. In light of this model much more work is needed on the seismicity, fracture and offset patterns within recent deposits and raised Quaternary reefs on Jamaica and seismic reflection work in offshore areas is necessary. Programs including the latter two are being initiated by K. Burke and P. Goreau, respectively.

CHAPTER X

CONCLUSIONS

Sedimentological, structural, petrographic and paleontological data together suggest that the Lucea Inlier represents an Upper Cretaceous upper slope basin of an intra-oceanic island arc system. The exposed section represents a shelf to basin sequence in which submarine canyons cutting into shelf deposits fed a submarine fan complex. Sedimentary structures suggest that clastics were deposited by a variety of sediment gravity flows. South to north and east-west paleocurrent data suggest that granodioritic plutons and associated subaerial volcanics of Upper Cretaceous age exposed and located within drill holes, to the east and southeast, were source terranes for clastics of the Lucea Inlier. If one assumes a north dipping Cretaceous subduction zone for Jamaica then the Lucea Inlier would represent an intra-arc to back-arc basin. The simple, open structural character of the inlier is analogous to the structure of modern and ancient upper slope basins.

The Caribbean oceanic crust was probably created at a ridge during the Jurassic to Cretaceous period, though a Late Cretaceous (80-75 mybp) basaltic sill event thickened a large portion of the Caribbean oceanic crust. From Jurassic to Eocene times unmodified oceanic crust was consumed by a complex geometry of subduction zones. A series of Late Cretaceous to Eocene, diachronous, suturing events within the Caribbean region eg. Motagua suture of Central America; Cuba and northern Hispaniola with the Bahama Platform; arc-continent of South America (Maresch, 1974); modified, thickened, Caribbean oceanic crust and subduction zones (Burke et al. 1978) and a concomittant change in plate motions established the present-day plate geometry of the Caribbean region. Original Late

Cretaceous geometries of arc-trench systems have been modified by a minimum of 720 km of Cenozoic left-lateral slip along the northern Caribbean plate boundary.

Since Mid-Miocene times this left-lateral slip has dominated the tectonics of Jamaica and much of the northern Caribbean plate boundary. Presently active northwest, northeast and east-west trending structures within Jamaica are interpreted as being of compressional, extensional and strike-slip origin, respectively, and are thought to be related to left-lateral slip along the active Oriente transform to the north.

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