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The Structure of the Crust and Upper Mantle in the Region of Barbados and the Lesser Antilles

Graham K. Westbrook

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Summary

The results of marine geophysical surveys across the Lesser Antilles in 1971 and 1972 by Durham University and the Royal Navy as part of the CICAR Project have revealed that the arc front sediment complex is nearly 20 km thick beneath the Barbados Ridge, where the igneous crust of the Atlantic is subducted beneath the Caribbean Plate. It appears that the sediment complex has grown away from the island arc, engulfing any bathymetric trench that was originally present. The Barbados Ridge is underlain by metamorphosed sediments and has been uplifted 4 or 5 km since the Pliocene. The crust beneath the Lesser Antilles island arc is about 35 km thick, and the crustal segments either side of the arc differ from each other in their crustal structure. The whole arc complex shows a change in character along the arc at Lat. 14° N. A positive gravity anomaly of 40 mgal computed to be the theoretical anomaly caused by the subducted lithosphere beneath the Lesser Antilles, is compatible with the interpretation of the crustal structure. The Lesser Antilles are an example of a maturely developed island arc complex.

1. Introduction

The Lesser Antilles arc of calc-alkaline volcanic islands form the eastern margin of the Caribbean (Fig. 1). The arc dates from the Eocene, but in the northern part of the Lesser Antilles an inner arc of islands formed during the Pliocene, and it is this arc which is presently active (Martin-Kaye 1969). In the south the newer volcanic material was superimposed on the older volcanics. Miocene limestones cap the islands of the older outer arc. La Desirade, which lies east of Guadeloupe, is anomalous, with Jurassic igneous rocks, and it may be an uplifted segment of ocean floor (Mattinson, Fink & Hopson 1973).

A zone of seismicity dips westward beneath the Lesser Antilles to a depth of 200 km with an average dip of about 40°. Calculations of the subduction rate from the seismicity, by the writer and Molnar & Sykes (1969) using the method of Brune (1968), both give a value of 0·5 cm/yr.

The island of Barbados lies 160 km east of the Lesser Antilles and is the highest point on the Barbados Ridge which is a ridge of sedimentary rock that runs parallel to the island arc. On Barbados oceanic sediments of Middle Eocene to Miocene age (The Oceanic & Bissex Hill Formations) overlie a tectonically disturbed sequence of Eocene flysch (The Scotland Formation) shown by exploration boreholes to be greater than 4·5 km thick (Baadsgaard 1960). The axis of the Barbados Ridge is coincident with a strong negative Bouguer and isostatic gravity anomaly (Andrew,

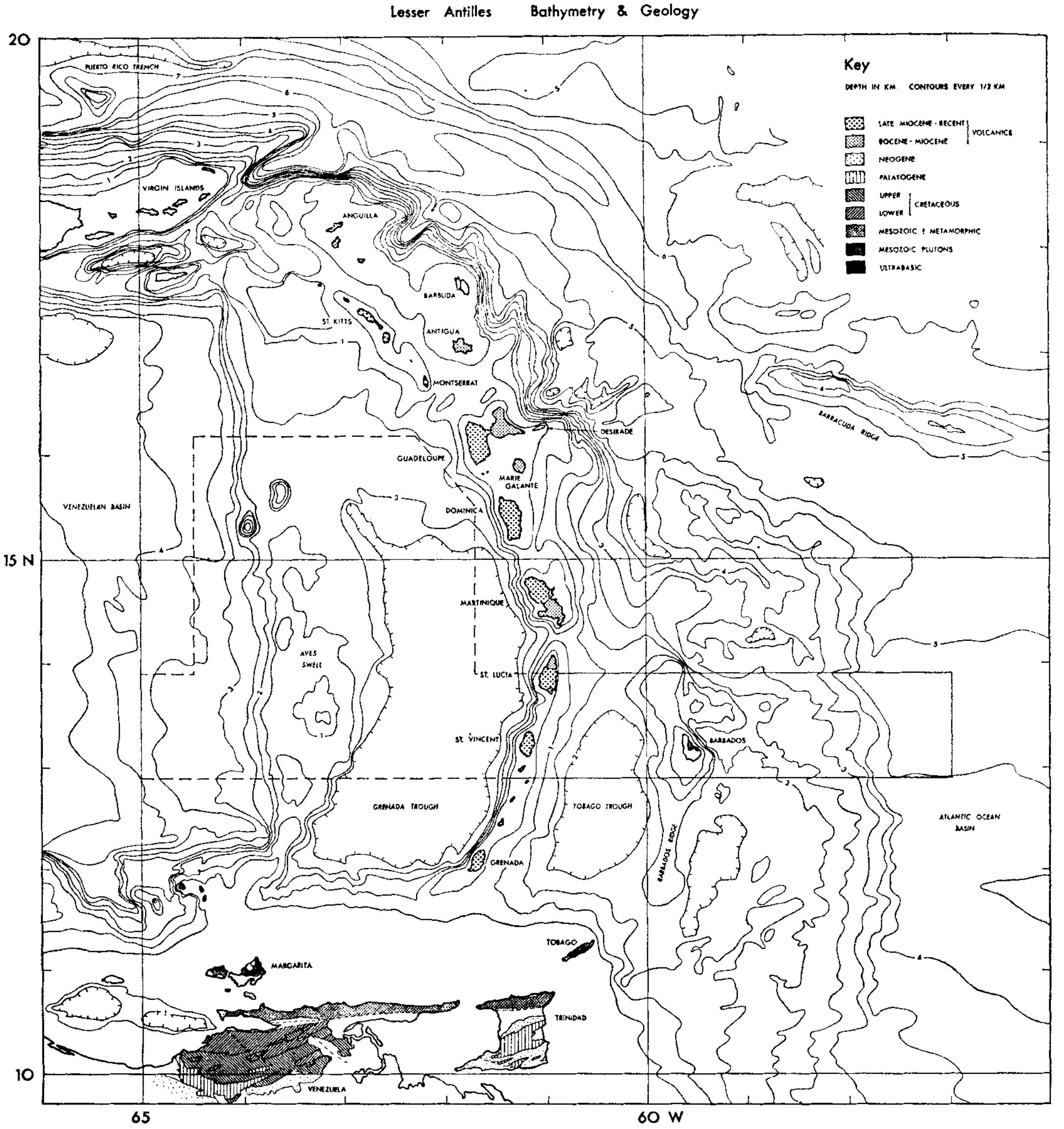
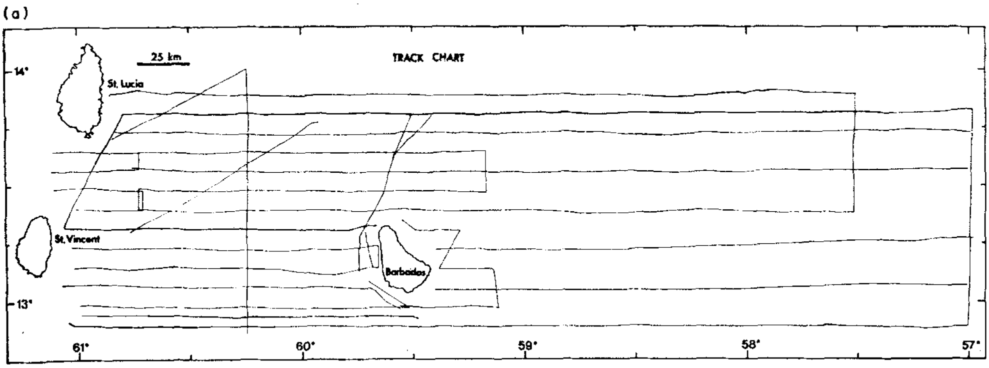
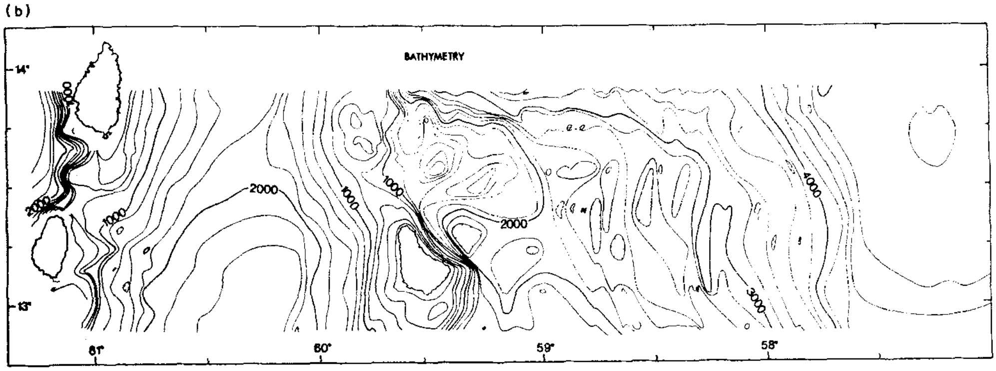


FIG. 1. Bathymetry and geology of the Lesser Antilles, eastern Caribbean. The areas surveyed by Durham University and the Royal Navy in 1971 and 1972 are outlined. Detailed maps of the area with solid outline are given in this paper.

Masson Smith & Robson 1970) and the eastern limit of the seismicity of the Benioff zone. The ridge is a major element of a 300 km wide region of uplifted and disturbed sediment that flanks the eastern side of the Lesser Antilles. There is no ocean trench opposite most of the arc. Between the Barbados Ridge and the arc is the Tobago Trough containing undisturbed sediment.

West of the arc lie the Grenada Trough, a flat-bottomed basin containing a 4 km thickness of sediment, and the Aves Ridge, or Swell, a submarine ridge which has many features that suggest it is an old island arc (Kearey 1974). Geophysical maps





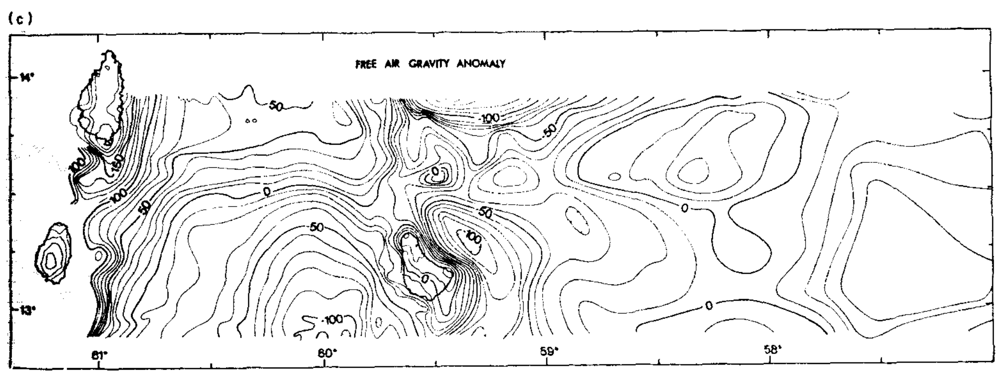


FIG. 2(a). Track chart showing lines along which bathymetric, gravity and magnetic data were measured. Heavy lines indicate that seismic reflection data was also collected.

FIG. 2(b). Bathymetry. Depths are in metres obtained for a sounding velocity of 1500 m s^{-1}. Contours are at 200-m intervals.

FIG. 2(c). Free-air gravity anomaly, in milligals. Contours every 10 mgal.

of the area bounded by latitudes 10° N and 17° N and longitudes 57° W and 65° W may be found in Kearey, Peter & Westbrook (1975).

Maps of the ship's tracks bathymetry, free-air gravity anomaly, Bouguer gravity anomalies and total field magnetic anomaly measured in the survey area are shown in Figs 2 and 3. Navigation was mainly by the Decca LAMBDA system with control from satellite navigator. In calculating the Bouguer gravity anomalies a two-dimen-

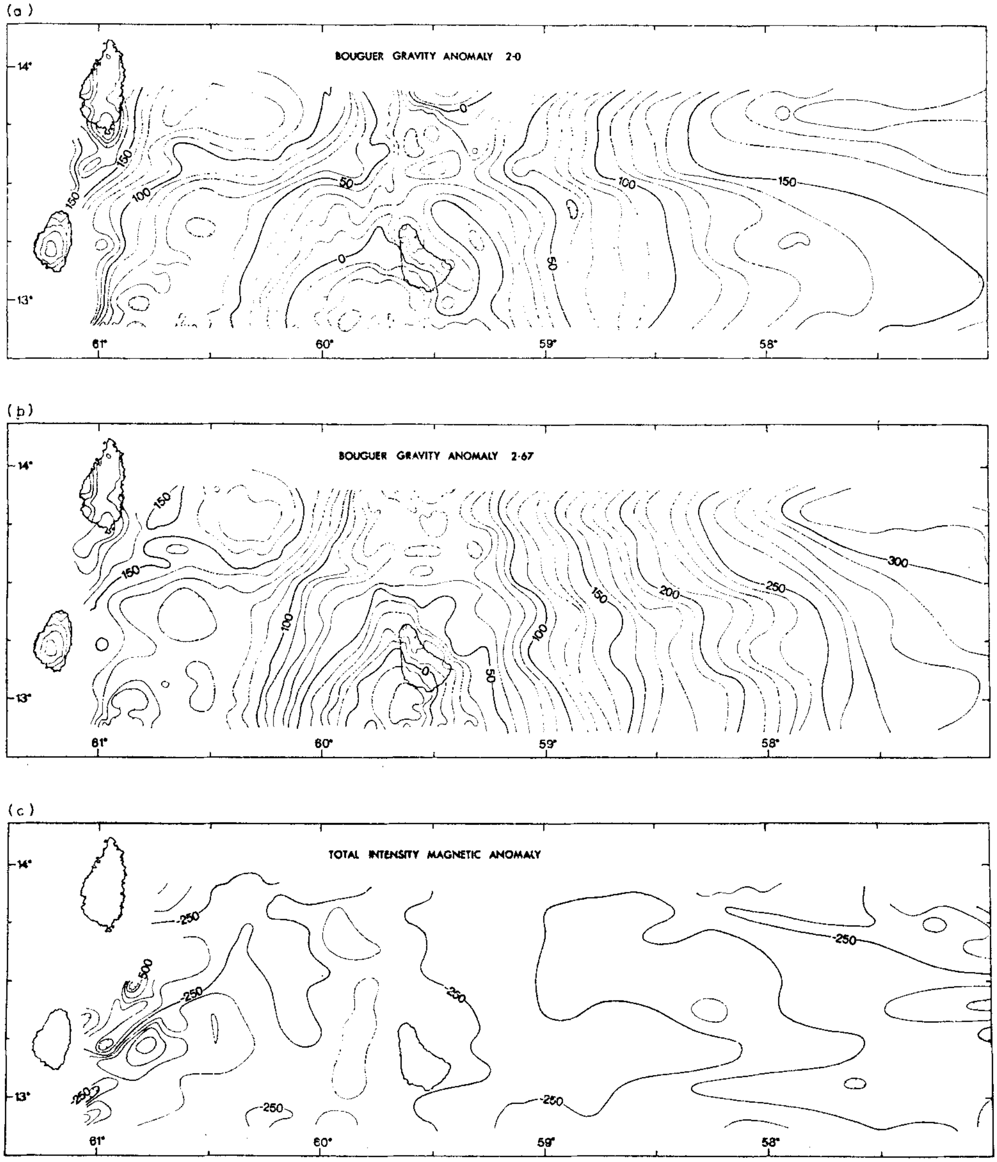


FIG. 3(a). Bouguer gravity anomaly. Correction for a density of $2 \cdot 0$ g cm^{-3} with the two-dimensional effect of topography along the ship's track incorporated. Contours every 10 mgal.

FIG. 3(b). Bouguer gravity anomaly correction for a density of $2 \cdot 67$ g cm^{-3} with the two-dimensional effect of topography along the ship's track incorporated. Contours every 10 mgal.

FIG. 3(c). Total intensity magnetic anomaly in gamma. Contours every 50 gamma. The IGRF and corrections for ionospheric disturbances have been removed. Between St Lucia and St Vincent there are anomalies of -500 to 30 gamma of too short a wavelength to be able to contour them at the track spacing of the survey.

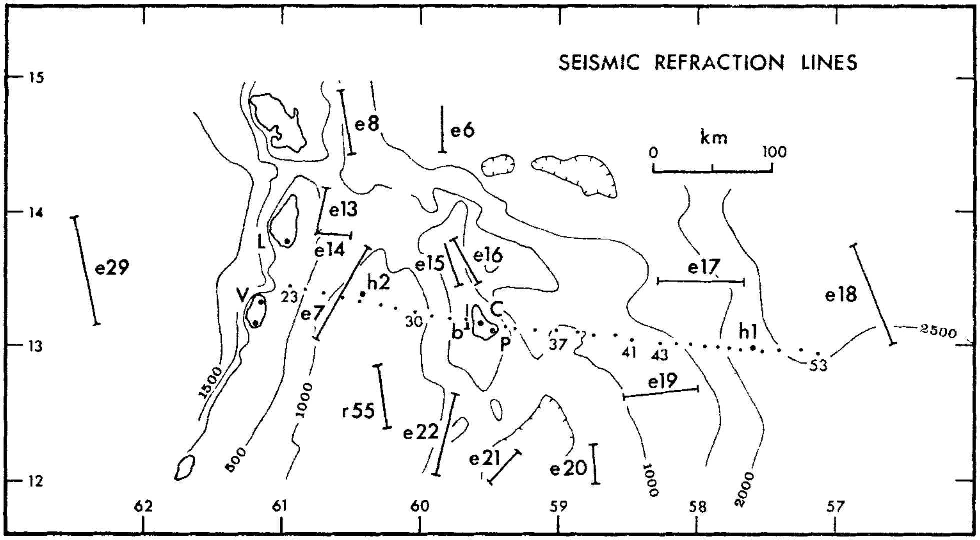


FIG. 4. Summary Chart of seismic refraction work. Small dots indicate shot points for LASP Project. Large dots are receiving stations: C, Coles Cave; P, St Phillips; L, St Lucia; V, St Vincent; h1 & h2, hydrophone stations occupied by Discoverer, Refraction profiles of other workers: e, Ewing *et al.* (1957); b, Worzel & Ewing (1948); r, Edgar *et al.* (1971). Contours in fathoms from Hess (1966).

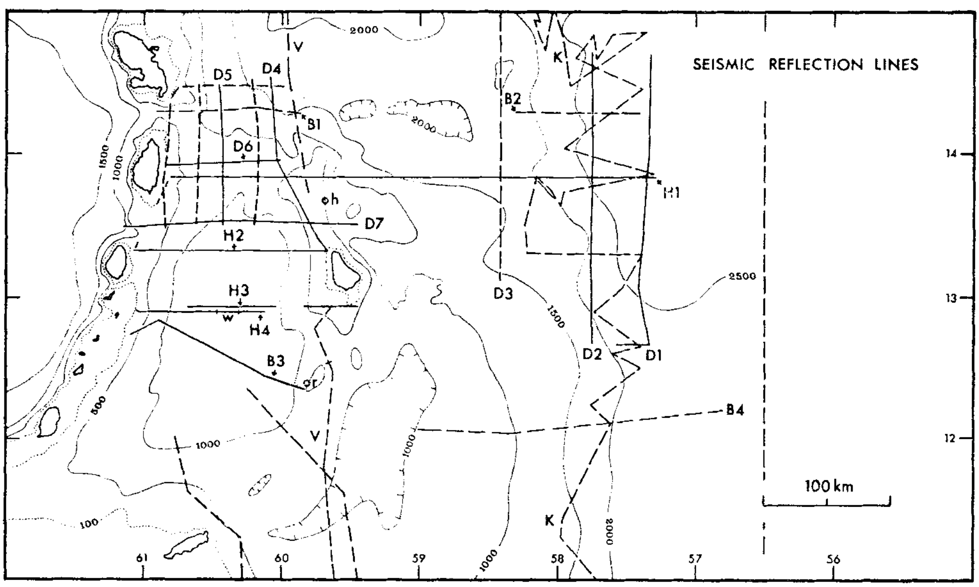


FIG. 5. Summary chart of seismic reflection profiles. H, HMS *Hecla* (this work); W, wide angle reflection run; D, Discoverer. NOAA; B, Bunce *et al.* (1971); V, Collette *et al.* (1969); K, Kane (Lowrie & Escowitz 1969); Bassinger *et al.* (1971); o h position of dredge sample taken by Hurley (1966); o r position of Miocene core taken by Ramsay (1968); Bathymetry in fathoms is from Hess (1966).

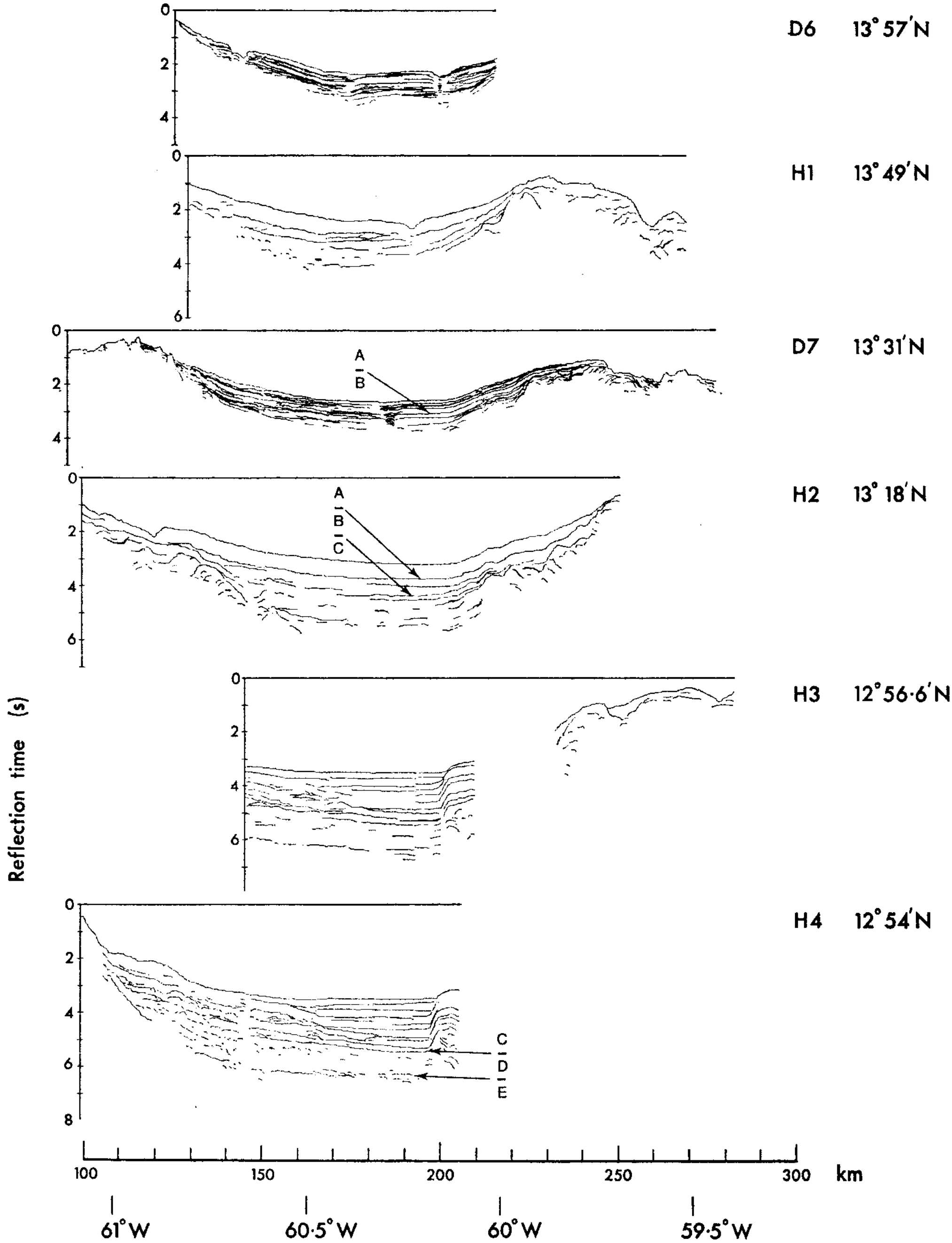


FIG. 6. Line drawing of reflection profiles across the Tobago Trough. Horizontal scale in kilometres east of Long. 62° W.

sional correction was applied for the effect of seabed relief along the ship's track. The IGRF (Epoch 1965) and a correction for ionospheric effects were removed from the measured total magnetic field to give the magnetic anomaly. Fig. 4 shows the positions of shots and receivers for part of LASP (Lesser Antilles Seismic Project) and the published refraction lines of other workers. Reversed profiles were obtained between h2 and St Vincent and Barbados. Arrivals were received at h1 from shots 37 to 52. The line was not reversed, but the dip of the refractors was obtained by using the velocity data from the lines of Ewing *et al.* (1957). Details of the profiles are given in Appendix 1.

2. The sedimentary layer

Seismic reflection profiles across the area (Fig. 5) have provided most information on the upper structure of the sediments. Seismic velocities in the sediment layers are known from seismic refraction (Ewing *et al.* 1957; Edgar, Ewing & Hennion 1971), and some new wide angle reflection data (Appendix 2). The area is divided into four provinces: the Tobago Trough, the Barbados Ridge, the Barbados Slope and the Atlantic Ocean Floor.

The Tobago Trough

This terminates against the South American continental shelf in the south. A gentle rise east of St Lucia forms its northern margin. Reflection profiles (Fig. 6) show regular reflecting horizons in the centre of the trough which become more disturbed at the margins where they bow upward. By combining the seismic refraction data with wide angle reflection (Appendix 2) and the reflection profiles the following generalized description of the sedimentary sequence was obtained.

Layer	Velocity km s^{-1}	Thickness km
A	1·55	0·4
B	1·9	0·5
C	2·2–2·6	0·8
D	2·9–3·4	2·0–3·0
E	3·8–4·3	2·0–5·0

Layers A and B have prominent regular bedding clearly visible on profiles D6 and D7 (Fig. 6). On the volcanic arc side of the trough the reflectors in Layer A have greater continuity than those in Layer B but are less reflective.

In the north of the trough Layers A and B conform in shape to the bathymetry, and thin towards the margins of the trough. Further south (Fig. 7) flat-lying reflectors in the centre of the trough unconformably overlie a series of reflectors which dip gently eastward (Bunce *et al.* 1971). The underlying sequence bends upward at the margins of the trough suggesting correlation with layers A and B further north. The flat-overlying sequence (referred to as layer A*) extends as far north as profile H3, and is prominent in the southeast of the trough (Bassinger, Harbison & Weeks 1971).

Layer C is the lowest layer in which continuous reflectors are observed. Apart from its higher velocity it resembles Layer B. The layering of Layers A*, A, B and C suggests that they contain turbidites. A few cores taken by Keller *et al.* (1972) from

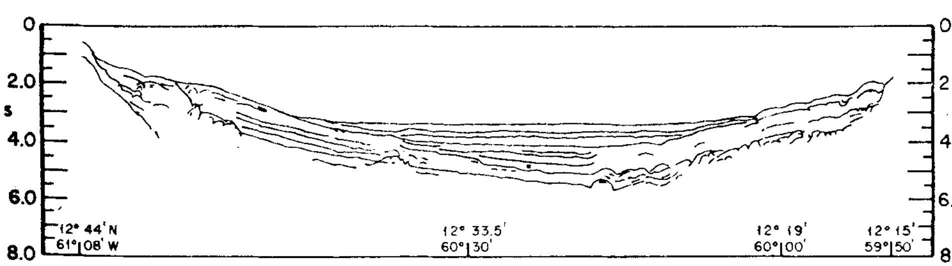


FIG. 7. Reflection profile of Bunce *et al.* (1971) across Tobago Trough.

the uppermost sediments in the trough showed graded beds. On the western side of the trough the sand content was high, up to 75 per cent, derived predominantly from the island arc. Sediments from the centre of the trough were more homogeneous containing more clay and silt size material much of which appeared to have come from the Amazon and Orinoco. This change in sediment type probably accounts for some of the variation of the appearance of reflecting horizons across the trough.

Layer D does not contain any continuous reflectors. There is often a marked discontinuity between it and overlying layers. This is seen on the western flank of the Barbados Ridge where the top of D is irregular and appears to truncate discontinuous reflectors within D, many of which dip to the east (Fig. 6).

The top of Layer E is the deepest reflector seen on profiles H3 and H4. A velocity of $4 \cdot 0\ \mathrm{km\ s^{-1}}$ was determined for it on refraction line e7. The top of E is not seen in the east of the trough, but in the west it curves gently upward to within $0 \cdot 7$ km of the seabed and is overlapped by later sediment.

The sediment structure of the western part of the trough is much less uniform than the centre, with many slump structures. On profiles D6 and H2 canyons can be seen to cut into the uppermost layer; that on D6 leads northeastward from St Lucia and that on H2 leads southeastward. H2 shows a subsurface rise containing disturbed reflectors on the western flank of the trough. Associated with this rise is a pronounced magnetic anomaly and it seems likely that the rise is made up from volcanically produced material.

The trough's basinal form is partly the result of deposition against the western side of the trough of material from the island arc, but is due principally to the uplifted Barbados Ridge forming the eastern margin of the trough. Reflectors in the top sediments near the crest of the Barbados Ridge intersect the seabed at a gentle angle indicating that it is now an erosional surface. The upper sedimentary layers (A, B & C) show pronounced warping on the ridge flank that appears to be caused by differential vertical movement along their base presumably resulting from faulting. An angular unconformity forms the top of layer D. The material below was disturbed and eroded to give an irregular surface before the overlying layers were deposited. The density contrast at this undulating surface produces several short wavelength gravity anomalies of low amplitude (<15 mgal). On profiles H3 and H4 the edge of the disturbed flank region is marked by a sharp upturn of the flat lying sediments of the centre of the trough. Two-dimensional migration performed on the reflection record shows that the upper layers are not faulted, but the strong warping may result from faulting in more competent rock below. Some relative uplift of the island arc may be indicated by the eastward dip of the reflectors beneath Layer A* in the centre of the trough.

The Barbados Ridge

On the Barbados Ridge stratified sediment of varying thickness overlies an irregular surface, beneath which lies material with a disturbed internal structure that often shows no coherent reflectors. The density contrast at this surface produces small gravity anomalies, as on the western flank of the ridge. Seismic refraction (Ewing *et al.* 1957) shows a top layer of low velocity ($1 \cdot 7\ \mathrm{km\ s^{-1}}$) and varying thickness ($0 \cdot 2$–$1 \cdot 1$ km) which conforms well with Layers A and B of the Tobago Trough when traced up onto the ridge crest. On the ridge north of Barbados the upper layers are underlain successively by layers of velocity $2 \cdot 2$–$2 \cdot 75\ \mathrm{km\ s^{-1}}$, $3 \cdot 2$–$3 \cdot 4\ \mathrm{km\ s^{-1}}$, and $4 \cdot 0$–$4 \cdot 3\ \mathrm{km\ s^{-1}}$ (Ewing *et al.* 1957). Hurley (1966) dredged samples of flysch similar to the Scotland Formation of Barbados from a point on the ridge north of Barbados (h on Fig. 5). The seabed there is underlain by material showing discontinuous and distorted reflectors. A nearby refraction line indicates that its seismic velocity is

$2 \cdot 2$ km s^{-1}. From this evidence it would appear that the Eocene flysch lies below layers A and B north of Barbados.

Just south of Barbados, Hurley (1966) dredged limestones containing manganese nodules which he thought likely to be a facies of the Upper Oceanic or Bissex Hill Formations of Barbados. A nearby reflection profile (H3) shows Layer D coming to the seabed near the ridge crest. This suggests that south of Barbados the Oceanic and Bissex Hill Formations form the base to the upper stratified sediments (Layers A & B). Refraction velocities obtained by Ewing *et al.* (1957) for the rock beneath the upper layer in this area were $3 \cdot 97$ km s^{-1} and $3 \cdot 81$ km s^{-1} which are quite normal values for lithified limestone. The thickness of this layer on the ridge crest is $2 \cdot 56$–$2 \cdot 82$ km which is thicker than the maximum $1 \cdot 5$ km thickness of oceanic sediments seen on Barbados (Baadsgaard 1960). It is probable that the boundary between the oceanics and the Scotland Formation is not observed by seismic refraction.

The deepest refractors in the ridge south of Barbados detected by Ewing *et al.* (1957) gave velocities of $4 \cdot 9$ and $5 \cdot 3$ km s^{-1} at depths of $4 \cdot 4$–$10 \cdot 4$ km. The gravity anomaly and seismic results from LASP preclude this refractor from being the top of oceanic igneous crust and it seems most likely that it is metamorphosed sediment. The refractor was not detected in the region of Barbados and obviously metamorphosed rocks have not been reported from the deepest borehole, of $4 \cdot 6$ km depth (Baadsgaard 1960).

On the Barbados Ridge the upper stratified sediment layers drape over rises in the lower disturbed sediment giving on a reflection record the appearance of occasional anticlines and synclines; a feature that can be seen along the ridge as well as across it. At the southern end of the ridge Bassinger *et al.* (1971) have tried to trace these structures along the ridge and into the continental shelf. However only the crest of the ridge and a shallow trough east of it can be shown clearly to cross more than one of their profiles. The largest folds in the Scotland Formation of Barbados have a wavelength of about 2 km and would not show their shape clearly on the available reflection records, but would appear as rather irregular and probably intermittent reflectors. The Oceanic Formation is gently folded and forms a periclinical structure on Barbados (Daviess 1971). It is structures of this type that are shown by the reflection profiles along the ridge, they seem to be formed by differential vertical movement of blocks of material below the upper layers.

The Atlantic Ocean Floor

To the east of the Lesser Antilles, at the base of the slope from the Barbados Ridge, reflection profiles over the ocean floor show comparatively flat lying reflectors to about 2-s penetration below the seabed. The uppermost half-second shows considerable stratification. Apart from these upper layers there are two main reflectors, which are seen at $0 \cdot 9$ and $2 \cdot 0$ s at the eastern end of profile H1 (Fig. 8). The deepest reflector is approximately at the depth to basement of $2 \cdot 24$ km below the seabed found at the north end of refraction line e18 (Fig. 4). Profile H1 runs just south of the crest of an easterly trending basement ridge which can be seen on profiles D1 and D2 (Fig. 9). The basement deepens southward by at least $1 \cdot 8$ s, and it also dips slightly to the west (a depth increase of $0 \cdot 3$ s in 20 km, which is a dip of about 1°). This is compatible with basement depths of $4 \cdot 05$ km obtained at hydrophone station h1, and $3 \cdot 1$ km obtained at the southern end of e18. These depths to basement are to a refractor of seismic velocity $6 \cdot 6$–$6 \cdot 8$ km s^{-1} (Oceanic Layer 3). Layer 2 is almost certainly present, but the thickness of overlying sediment prevents first arrivals from it being observed, and consequently the depth to igneous crust will be about $0 \cdot 25$ km shallower than the basement depths given above.

The reflector at $0 \cdot 9$ s has an undulating surface which is not as rough as the basement, and gets deeper to the south, where by Lat. 12° N it has a depth of $2 \cdot 0$ s

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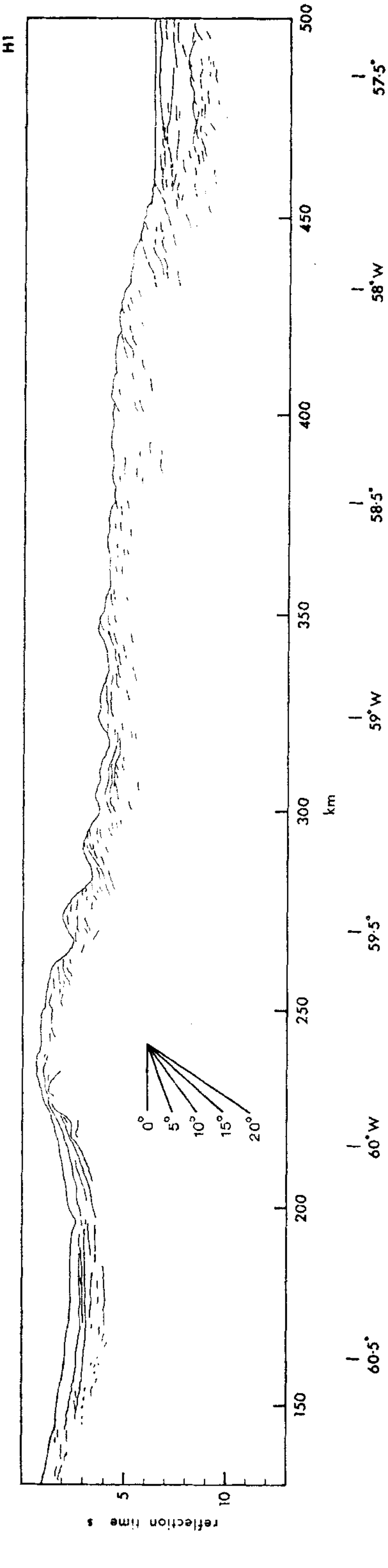

FIG. 8. Line drawing of reflection profile H1. Rose diagram indicates inclination of reflectors in material of velocity 2·0 km s^{-1}.

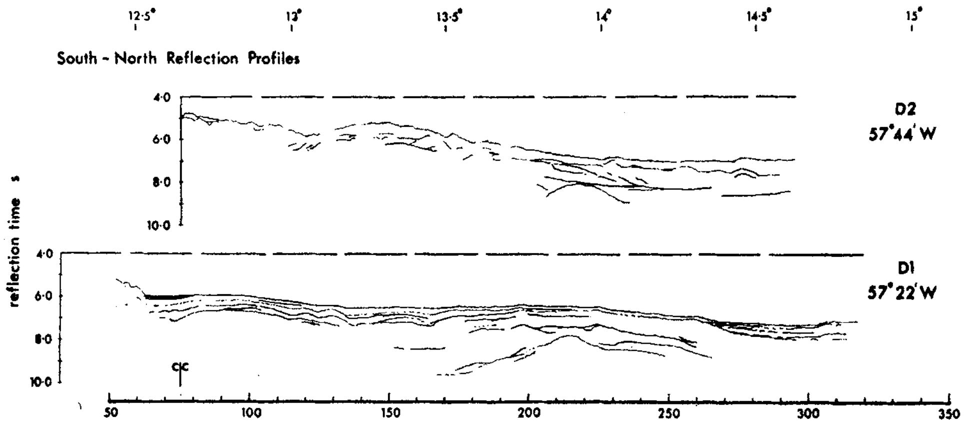

FIG. 9. South–north reflection profiles D1 and D2 at edge of Barbados Ridge Complex.

below the seabed, shown on profile B4 of Bunce *et al.* (1971). The basement is not seen south of Lat. 13° N on any of the reflection profiles until one goes as far east as Long. 54° 30′ W (Peter & Westbrook 1974).

By extrapolation along southerly running reflection profiles (Peter, Schubert & Westbrook 1974) from the area of JOIDES drill site 27 (Lat. 15° 51·39′ N, Long. 56° 52·76′ W) it can be shown that the upper stratified sediments are of Pliocene to Recent age. The youngest sedimentary feature is a small wedge-shaped basin 0·2 s thick at the base of the Barbados Slope, which is well shown on profile D1 west of the course change.

The Barbados Slope

This region, between the crest of the Barbados Ridge and the level ocean bed of the Atlantic, has an irregular topography of many minor ridges and troughs with an overall trend parallel to the Barbados Ridge. Much minor relief is shown on north–south lines as well. Disturbed reflectors appear not far beneath the seabed. Their reflectance is great, but their coherence is low, and the material might be described as acoustically opaque.

Unconformably overlying the disturbed reflecting material is an acoustically transparent layer of greatly variable thickness (0·5 s average) but which thickens overall from east to west. The layer is virtually absent from much of the easternmost part of the slope and in the west fills minor troughs and is internally stratified. The profiles of Bassinger *et al.* (1971) show the layer with a thickness of about 1 s filling a shallow basin east of the Barbados Ridge.

Reflection profiles show that the upper sediment layers beneath the Atlantic ocean floor are uplifted and dislocated at the very beginning of the slope. Some of the deeper reflectors in the sediment extend several kilometres beneath the foot of the slope before they become disrupted or are otherwise obscured. Profile H1 (Fig. 8) shows that although the lower reflectors are irregular they have a predominantly westerly dip of about 5° which gives a similar appearance to reflection profiles across the sediment pile before the Java arc (Beck 1972). Other reflection profiles do not show this dip on reflectors, indicating only a rather chaotic disturbance (Bunce *et al.* 1971; Peter *et al.* 1974).

The formation of the sediment pile of the Barbados Slope is considered below (p. 28). The initial uplift and deformation of the sediment takes place in the first 40 or 50 km of the slope.

Correlation between seismic layers and lithologies

This can be accomplished best where a well-defined reflector on a reflection profile can be traced to a point where the lithology is known from sampling, as in the case of the top of Layer D of the Tobago Trough and the Miocene limestone dredged from south of Barbados (Hurley 1966). Most of the information, however, has been obtained from seismic refraction related to reflection profiles. Identification of a sedimentary layer from its seismic velocity presents problems because the velocity is dependent on depth of burial and degree of lithification. Layers which can be shown to be continuous on reflection have different velocities at different depths. Also the detection of layers of similar velocity on several refraction lines can lead to the extrapolation of a single layer between the lines which cannot be shown to be a single unit on a reflection profile. East of the Barbados Ridge refraction lines have detected a layer with velocities in the range $2 \cdot 4$–$2 \cdot 6$ km s^{-1} and thicknesses from $2 \cdot 6$–$6 \cdot 2$ km. This layer appears at the crest of the Barbados Ridge where Eocene flysch outcrops and in the Atlantic is only $0 \cdot 25$ km below the seabed where sediments are probably of Pliocene age. This seismic 'layer' therefore is diachronous and must be principally a result of compaction.

Higgins (1959) has compared seismic velocity data from boreholes in Trinidad with the seismic refraction results of Ewing *et al.* (1957). He suggested that the $2 \cdot 2$–$2 \cdot 6$ km s^{-1} layer (C in the Tobago Trough) corresponds to Plio–Miocene rocks, that the $2 \cdot 9$–$3 \cdot 43$ km s^{-1} layer (D) corresponds to Oligocene–Eocene rocks, and that the $4 \cdot 9$–$5 \cdot 3$ km s^{-1} layer corresponds to the Cretaceous metamorphics of the North Range of Trinidad which are mostly phyllites, quartzites and limestones of lower greenschist facies metamorphism (Potter 1968).

Higgins' correlation fits fairly well the ages of horizons deduced from the reflection profiles except that layer D contains Miocene rocks. The top surface of D is erosional and consequently rocks of different age are exposed at different places. Layers A, B and C are clearly late Miocene to Pleistocene age, and A* and possibly some of A must be Quaternary. The layer covering the disturbed sediments of the Barbados Slope is probably diachronous at its bottom surface, younging towards the east, but there appears to be a high proportion of fairly recent sediment in it.

3. Structures producing magnetic anomalies

In the region surveyed there are two areas where magnetic anomalies related to the crustal structure are observed (Fig. 3(c)). In the vicinity of the island arc there are many anomalies of short wavelength and high amplitude (up to 500 gamma peak to peak), and along the west side of the Barbados Ridge runs a positive anomaly with an amplitude of about 70 gamma and wavelength of 90 km.

Barbados Ridge Anomaly

The magnetic anomalies measured along east–west track lines in the vicinity of the Barbados Ridge are shown in Fig. 10. They show a long wavelength anomaly with a positive peak over the western side of the ridge on which are situated shorter wavelength anomalies, particularly north of Barbados where positive anomalies with a wavelength of about 20–30 km occur.

There are several possible causes of the anomaly. The first of these is the variation in structure of the igneous basement. Interpretation of the gravity and seismic data shows that the igneous basement is depressed beneath the Barbados Ridge to a depth of nearly 20 km, and therefore could only produce the long wavelength part of the anomaly. The possible magnetic effect of the basement structure was modelled two dimensionally using as a starting model the shape of the basement surface, as deter-

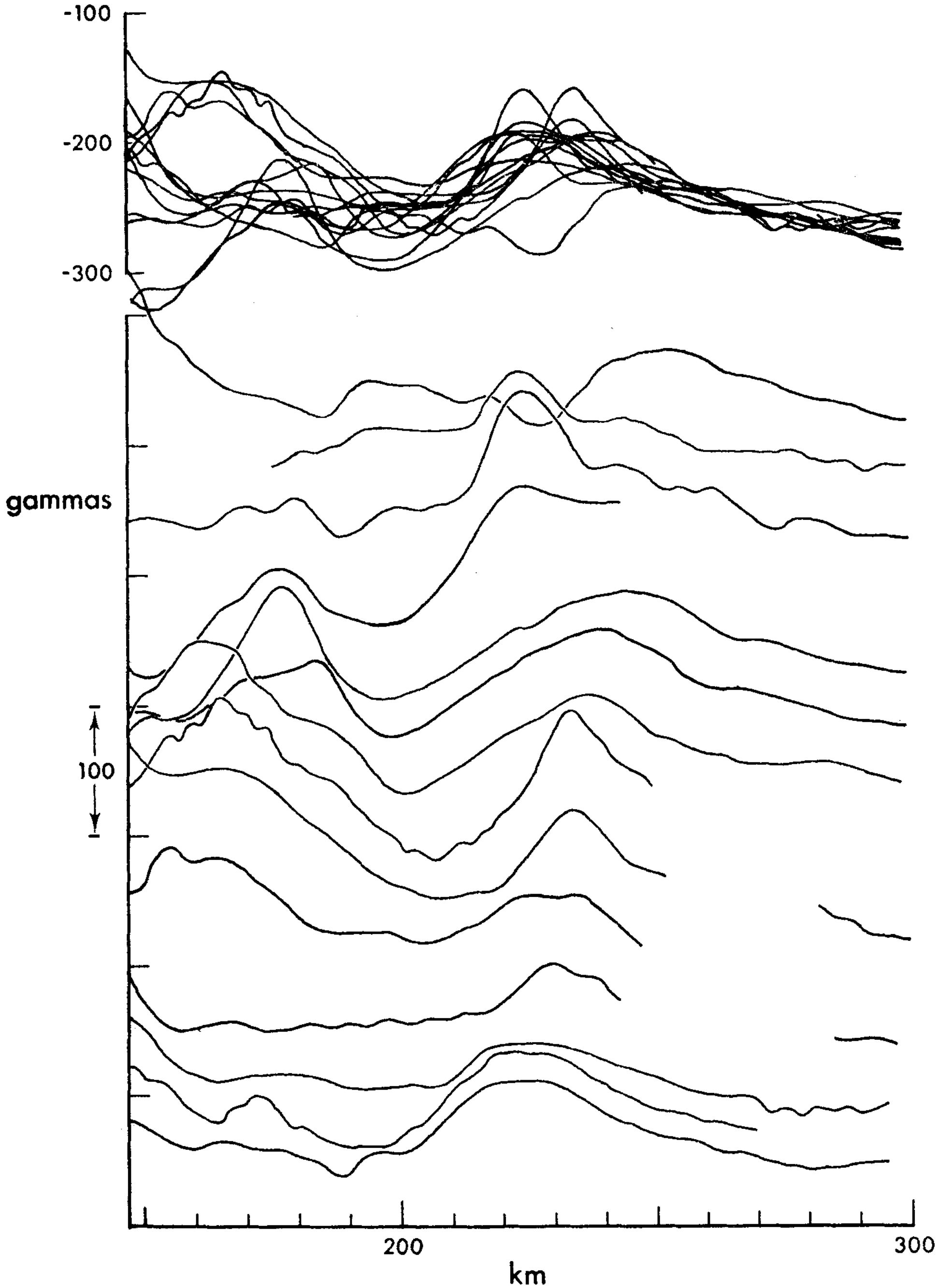

FIG. 10. Magnetic anomaly profiles across the Barbados Ridge and Tobago Trough. Upper part shows the anomalies superimposed. Lower part shows anomalies spaced proportional to line spacing. Horizontal scale shows kilometres east of Long. 62° W.

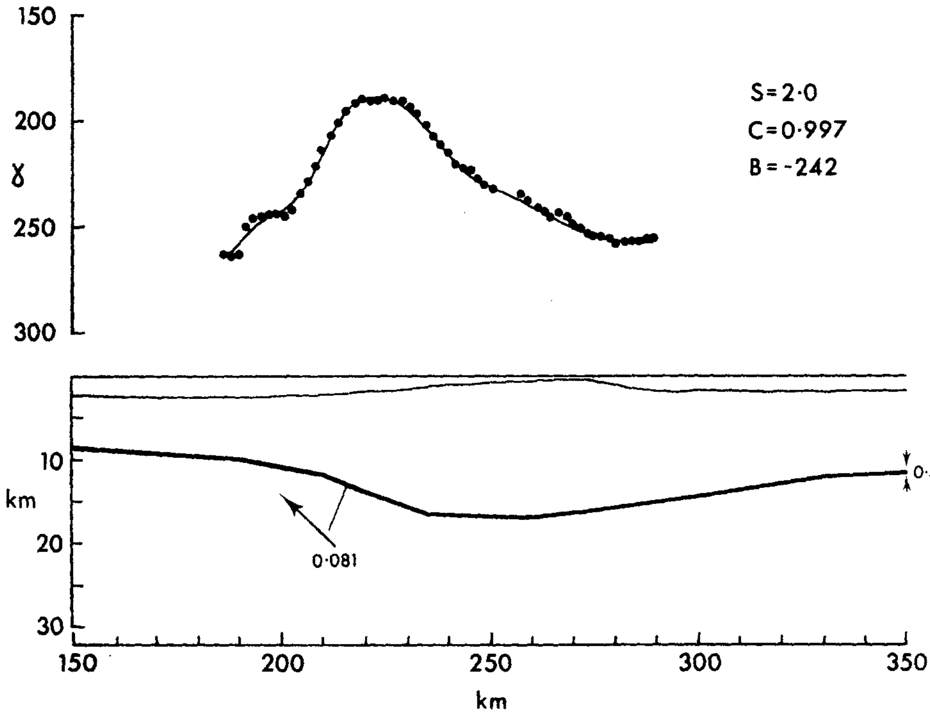

FIG. 11. Models of the magnetic basement to produce the magnetic anomalies across the Barbados Ridge at Lat. 12° 54′ N. The model is a magnetized layer 0·4 km thick representing oceanic layer 2. C, Correlation coefficient between observed and computed anomalies; S, Standard error; B, Background value. Dots are the observed values. The continuous curve is the calculated anomaly. Horizontal scale in km east of Long. 62° W. Numbers indicate magnitude of magnetization. Arrows show the direction of magnetization. Note vertical exaggeration of 2 : 1.

mined from the gravity and seismic data, as the top surface of a magnetised layer 0·4 km thick. The work of Talwani, Windisch & Langseth (1971) on the Reykjanes Ridge and Vine & Moores (1972) on the Troodos Massif of Cyprus has shown that the thickness of the highly magnetized top layer of igneous ocean crust is in the range of 0·4–1·0 km. The magnitude and dip of the magnetization vector and the positions of the body-points defining the model were varied using a non-linear optimization routine until a best fit between calculated and observed anomalies was obtained. The body-points had constraints put on them to stop them deviating too strongly from seismically determined structure. The model obtained for the profile at 12° 54′ N is shown in Fig. 11. The magnetization value of 0·081 is rather high, but increasing the thickness of the layer to 1·0 km reduces the value by more than half. This model represents one possible variant of a series of magnetized basement configurations in which the magnetization may vary laterally, possibly changing at the point of subduction, and serves only to show that it is possible for the magnetized igneous basement to produce the long wavelength part of the anomaly over the Barbados Ridge. The low angles of magnetization obtained are similar to those obtained from Cretaceous rocks around the Caribbean. The presence of shorter wavelength anomalies north and west of Barbados indicates that the basement cannot be the only source contributing to the anomaly. There are clearly higher level sources within the Barbados Ridge. There appear to be two geological possibilities; firstly small detached pieces

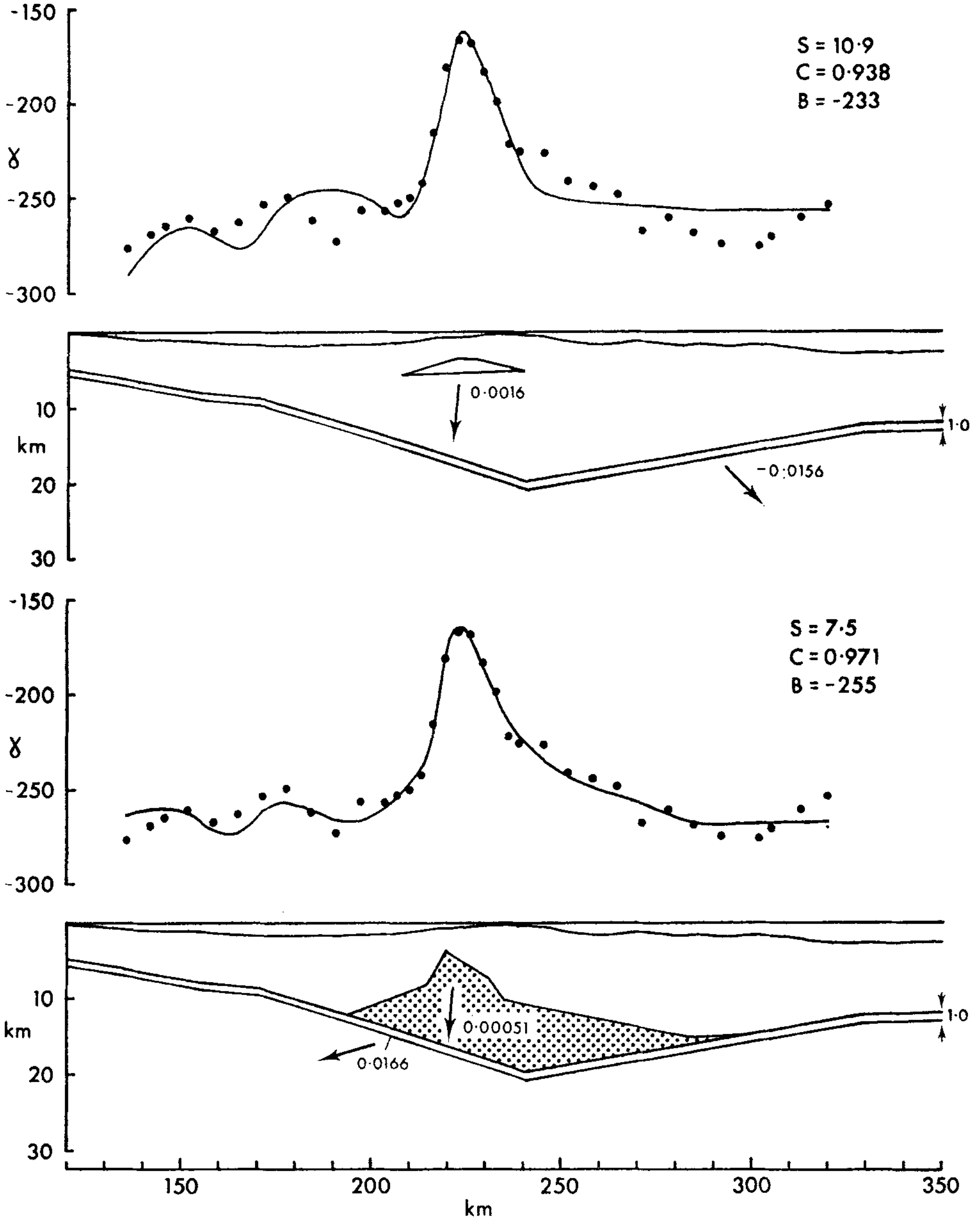

FIG. 12. Magnetic models for the magnetic profile across the Barbados Ridge at Lat. 13° 44′ N. Magnetic basement is represented by a 1 km thick magnetized layer. (a) The short wavelength anomaly is produced by a small discrete body, which may be interpreted as detached sliver of oceanic igneous crust. (b) Above the basement is a large body of low magnetization which produces the short wavelength anomaly, and may be interpreted as low grade metamorphosed sediment. For legend see Fig. 11.

of oceanic crust within the sediments, or secondly a rise composed of metamorphosed sediments of pelitic or semi-pelitic type in which at the chlorite grade of the greenschist metamorphic facies magnetite will be the stable iron oxide (Turner 1968). The Scotland Formation of Barbados has quite a high iron oxide content, and a depth of burial of 10–15 km could have produced sufficient metamorphism.

The two possibilities were modelled using the magnetic anomaly observed on the line crossing the ridge at 13° 44′ N. In each model the magnetic basement was included as a layer 1·0 km thick with its top surface fixed at top surface of the basement layer of the gravity and seismic model. The dip and magnitude of the magnetization vector of this layer in the plane of the profile was allowed to vary. Above the basement the two different structures shown in Fig. 12 were modelled. The magnetization of these structures was assumed to be induced. Fig. 12(a) shows the anomaly produced by a body 2 km thick 30 km wide with a magnetization of 0·0016 emu cm^{-3}. A fit may be obtained with a body 20 km wide, but in either case the body is rather thin for its width. The length of the body estimated from the length of the anomaly along strike is about 25 km. If the body does represent a piece of oceanic crust then it may be difficult to explain how it reached such a high level without breaking up more obviously than could be allowed for by the model. The presence of more, but slightly smaller bodies of a similar type must be invoked to account for the many other short wavelength fluctuations present, which cannot be explained as ionospheric effects.

Fig. 12(b) shows a model of a ridge of low grade metamorphic rocks with a magnetization of 0·00051 emu cm^{-3} which rises to a depth of 3·5 km below sea level. This produces a rather better fit to the anomaly than the single discrete body of the previous model. It is a more plausible model, although it implies a tectonic uplift of material which has been metamorphosed by 6–10 km. The top of the model must be within about 4·5 km of the surface for a good fit to be obtained. The model is not strongly dependent on the magnetization of the basement. A good fit may also be obtained with a model in which the basement magnetization is assumed to be induced (Westbrook 1974). The hypothesis of a ridge of low grade metamorphosed sediment appears to explain the magnetic anomalies in a more satisfactory manner than detached pieces of igneous basement, but this latter hypothesis cannot be excluded.

Anomalies over the Island Arc

The magnetic anomalies over the island arc are complicated, and indicate the presence of many magnetic bodies. Although the island arc is a linear magnetic body of a sort, it is quite clear that many of the magnetic bodies of which it is composed are not linear or planar in shape, but are comparatively irregular. In between St Lucia and St Vincent it was not possible to contour the magnetic anomalies at the line spacing of the survey because of their short wavelength and great variability.

To obtain some idea of the possible form of the magnetic bodies a two-dimensional model was computed for a profile across the arc south of St Vincent. The top surfaces of the bodies in this model were constrained to conform with the shape of the basement derived from gravity and seismic interpretation. Parameters were optimized non-linearly as in the other models.

The intensities and dips of the magnetization vector obtained for the model (Fig. 13) are close to what might be expected from measurements made on lavas from St Vincent by Khan (1968) who obtained a mean susceptibility of 0·0016, and a mean remanent magnetization of 0·0026 emu cm^{-3}. The directions of magnetization were near that of the present Earth's field.

The model does not quite match some of the short wavelength parts of the anomaly which must result from near surface changes in magnetization or structure. The model has a more complicated structure on its eastern side than its western side. Its

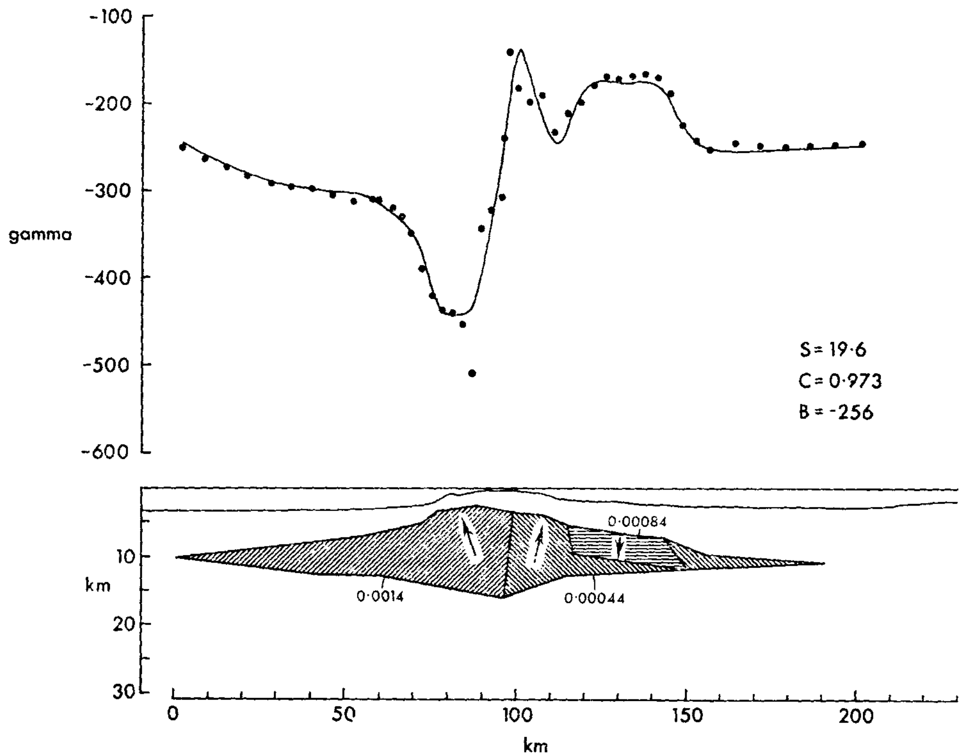

FIG. 13. A model of the magnetic bodies causing the magnetic anomalies across the arc at Lat. 13° 4′ N. For legend see Fig. 11.

shape is close to that expected from the interpretation gravity and seismic data for the upper of the two main layers making up the crust of the arc.

Another method used in the interpretation of the magnetic anomalies over the island arc was the derivation of pseudo-gravity anomalies from the magnetics using a two-dimensional magnetics to gravity transformation (Bott & Ingles 1972). Although the usefulness of the method is severely limited by the three-dimensional nature of the anomalies comparison of the pseudo-gravity anomalies with the measured gravity anomalies gives some indication of whether the magnetic anomalies are primarily due to changes in magnetization, in which case there should be little correlation between the true and pseudo-gravity anomalies, or to changes in the structure of the basement where some correlation between the pseudo-and true gravity anomalies is expected. The results indicate that the magnetic anomalies of the western side of the arc are primarily related to the structure of the igneous basement, but that the anomalies of the centre and eastern side of the arc are caused mainly by changes in magnetization which are presumably the result of a long history of eruptive igneous and intrusive igneous activity.

4. Overall crustal structure

The crustal structure was interpreted using two dimensional modelling to fit the gravity anomaly, constraining the positions of the layers to conform to the seismic measurements. The lines chosen for interpretation are crossed diagonally by the shot line of LASP and run along or close to the seismic reflection profiles across the

Tobago Trough. The anomaly used for interpretation was a Bouguer anomaly corrected for the two-dimensional effect of seabed topography along the profile with a density of $2 \cdot 1 \text{ g cm}^{-3}$. The density of $2 \cdot 1 \text{ g cm}^{-3}$ is that corresponding on the empirical curve of Ludwig, Nafe & Drake (1971) to a seismic velocity of $2 \cdot 5 \text{ km s}^{-1}$ which is the average velocity of the material that provides most of the relief across the area and is also that estimated to be the best Bouguer correction density using Nettleton's method. The gravitational effect of low velocity (average $1 \cdot 7 \text{ km s}^{-1}$) sediment layers in the Tobago Trough was removed from the anomaly prior to interpretation using the shape of the layer defined by seismic reflection and density of $1 \cdot 8 \text{ g cm}^{-3}$ (Horn, Horn & Delach 1968).

The densities used in the model were derived from the seismic velocities measured in the region using the empirical curve of Ludwig *et al.* (1971) and are shown below:

Layer	Velocity km s^{-1}	Density g cm^{-3}
Water	1·5	1·04
Unconsolidated sediment	1·7	1·8
Semi-consolidated sediment	2·5	2·1
Consolidated sediment	4·0	2·5
Igneous crust	6·8	2·9
Upper mantle	8·1	3·3

Oceanic Layer 2 was omitted from the model, because its small and relatively constant thickness over large distances will give it little anomalous effect on the gravity at the depth at which it is present. In the region of the island arc the seismic velocities obtained in the main igneous crustal layer were about $6 \cdot 2 \text{ km s}^{-1}$ and a density of $2 \cdot 8 \text{ g cm}^{-3}$ was used for the model there.

Figs 14 and 15 show the models derived for the lines at 13° 4′ N and 13° 24′ N. A major feature of these models is the depression of the igneous basement below the Barbados Ridge and the rise of consolidated sediment within the ridge. Forty-five kilometres south of Barbados a layer with a seismic velocity of $4 \cdot 9 \text{ km s}^{-1}$ was detected at a depth of 4·7 km (Ewing *et al.* 1957). Inclusion of this layer in its estimated position produces little change in the structure of the model. The Moho is depressed slightly west of Barbados and the top surface of the consolidated sediment is depressed in part as compared with models in which the additional layer was not included. If all other parameters are held constant the addition of the extra layer can depress the top surface of the basement by up to 2 km.

The rise of higher density sediment in the Barbados Ridge, which is up to 5 km above material of the same density either side of the ridge, fits the geological and seismic reflection evidence for the uplift of the ridge, and supports the idea that a ridge of metamorphosed sediment is the cause of the magnetic anomaly associated with the Barbados Ridge.

The thickness of sediments beneath the Barbados Ridge accounts well for the 2·0 s delay of earthquake arrivals at Barbados relative to stations on the island arc (Tomblin, private communication; see also Barr & Robson 1963).

The deepest part of depression of the igneous crust below the Barbados Ridge lies approximately at the eastern limit of seismicity associated with the Benioff zone beneath the Lesser Antilles and is presumably the position of subduction of the igneous crust of the Atlantic Plate beneath the igneous crust of the Caribbean.

The crustal models in the region of the island arc must inevitably be an over-simplification of the true structure. Masson Smith & Andrew (1965) found great variation in the densities measured on the islands. The higher density rocks collected by them had densities of about $2 \cdot 8 \text{ g cm}^{-3}$ which is the likely density for the main crustal layer indicated by seismic refraction velocities. The results of LASP in the

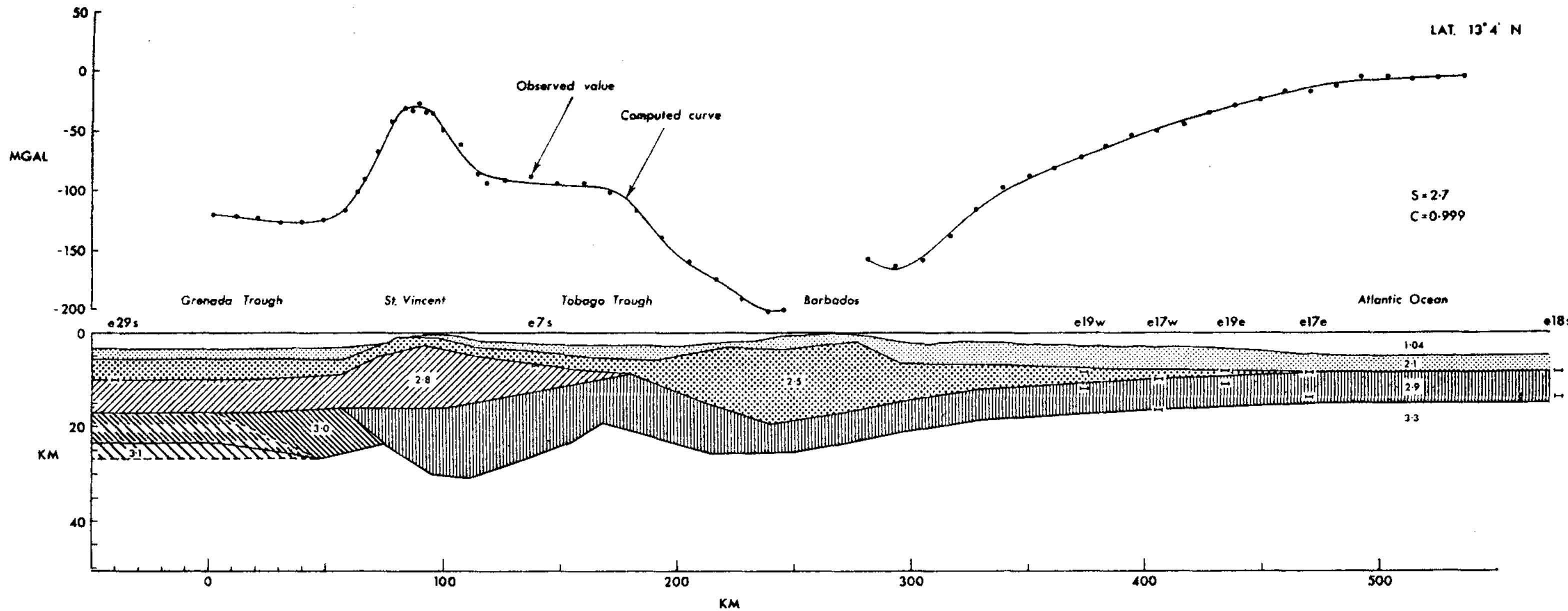

FIG. 14. Model of crustal structure across the region at Lat. 13° 4′ N derived from gravity and seismic data. Dots are observed values of the Bouguer gravity anomaly (correction density 2·1 g cm^{-3}). The continuous curve is the anomaly calculated from the model. Bodies of different density are shown with different ornament, with the values indicated. The short bars below letters and figures such as e17w indicate the depths of layers determined by the seismic refraction lines of Ewing *et al.* (1957) see Fig. 4. C is the correlation coefficient between the observed and computed gravity anomalies. S is the standard error. Horizontal scale is kilometres east of Long. 62° W. Vertical exaggeration 2 : 1.

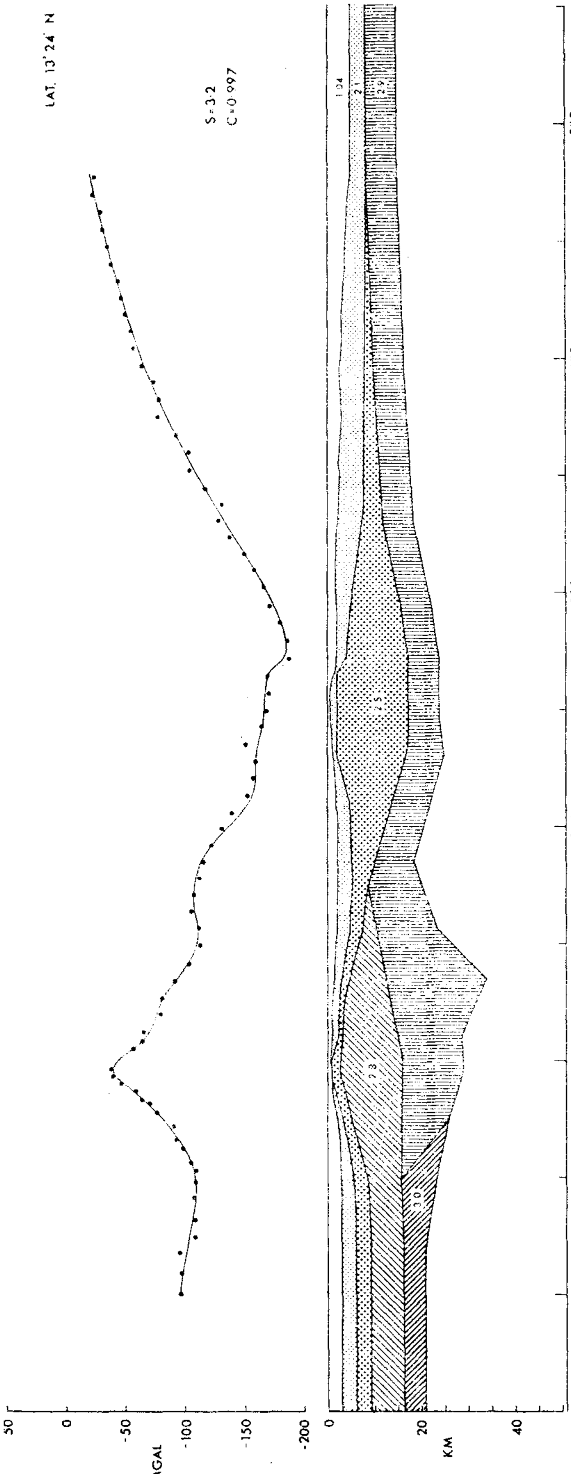

Fig. 15. Model of crustal structure across the region at Lat. 13° 24′ N, derived from gravity and seismic data. For legend see Fig. 14.

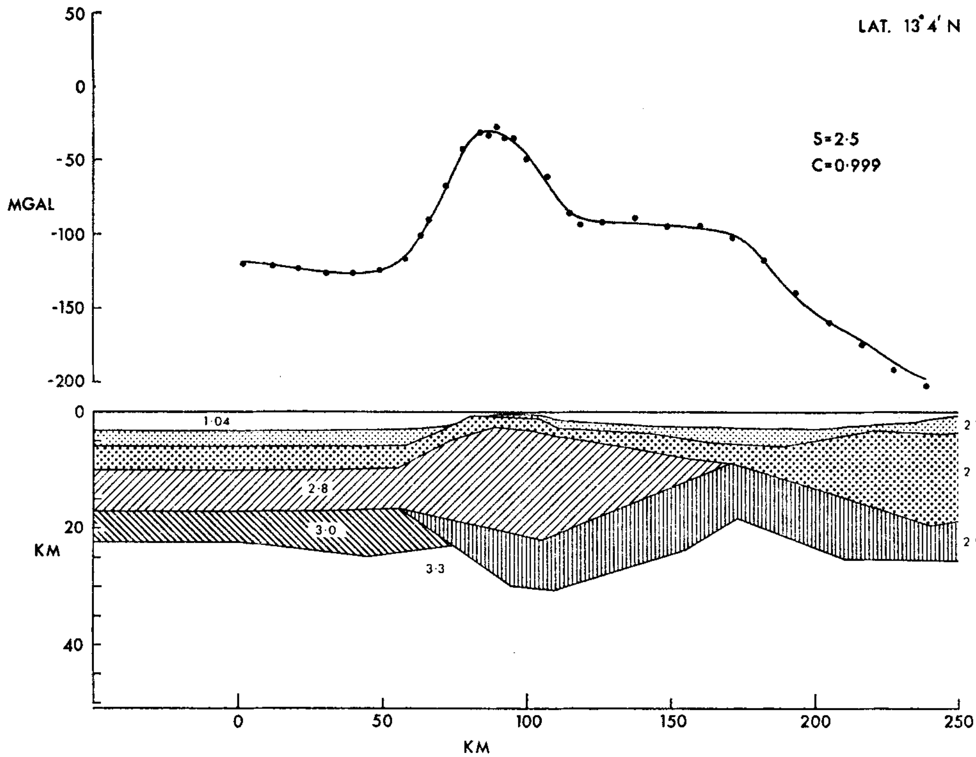

FIG. 16. Variation of the density model for the island arc at Lat. 13° 4′ N, representing island arc material built on old ocean crust which is depressed beneath arc.

region of the arc show that the minimum depth to the top of the $2 \cdot 8\,\mathrm{g\,cm^{-3}}$ ($6 \cdot 2\,\mathrm{km\,s^{-1}}$) layer was between $2 \cdot 5$ and $3 \cdot 25$ km for the two lines across the arc used for modelling. They also show that this layer is underlain by another layer of seismic velocity $7 \cdot 0\,\mathrm{km\,s^{-1}}$ at an average depth of 15 km (C. Boynton 1974). The Moho beneath the arc was not seen seismically, but the gravity modelling indicates that if it is present as definite discontinuity then its depth must be between 30 and 35 km. It is likely that the Moho is deeper under the islands than between them. What the interface between the upper and lower crustal layers represents is open to some interpretation. It seems that the $2 \cdot 8$ layer is a result of the creation of the island arc and probably consists mainly of plutonic rocks that have been intruded into the original basaltic pile. The $2 \cdot 9\,\mathrm{g\,cm^{-3}}$ layer may be the main crustal layer of the oceanic crust upon which the island arc was formed and the interface between the two layers is close to the old seafloor (Fig. 16). It is, however, quite likely that beneath the island arc the former ocean crust is strongly intruded by the material forming the island arc, and much of the lower layer could be made up of igneous cumulates. There is evidence from xenoliths in rocks at the surface that probably quite substantial quantities of these rocks exist within the arc (Lewis 1973; Wills 1974).

Over the arc and on its flanks there are short wavelength anomalies that are not satisfied by the models and must be caused by minor variation in the structure and density of the upper part of the crust. In the islands there are many rocks with densities between $2 \cdot 5$ and $2 \cdot 8$.

In the west of the Grenada Trough a layer of seismic velocity $7 \cdot 4\,\mathrm{km\,s^{-1}}$ is the lowest layer detected (Ewing *et al.* 1957) and probably has a density of about

$3 \cdot 1$ g cm^{-3}. How this layer joins the island arc is problematical, but is probable that there is some gradational change of structure and density, so a density of $3 \cdot 0$ g cm^{-3} was used for the layer on the eastern side of the trough. Consequently in some models the layer is thinner in the west than it should be. Fig. 14 shows the effect of giving the western part a density of $3 \cdot 1$ g cm^{-3}.

The Tobago Trough is shown clearly by the models to be a consequence of the formation of the Barbados Ridge. The depth to basement increases from west to east across the trough. Beneath the centre of the trough the Moho is shallower than beneath the island arc or the Barbados Ridge. This rise in the Moho is the result of the relative depression of the Moho beneath the island arc and the Barbados Ridge.

The crustal structure is different either side of the arc. The Tobago Trough is underlain by only one lower crustal layer whereas the Grenada Trough is underlain by two. This may imply a different mode of formation for these two crustal segments.

Easterly trending structures

Between St Lucia and the northern end of the Barbados Ridge lies a broad saddle-like feature which forms the northern margin of the Tobago Trough. The upper sedimentary layers drape gently over the feature in the north–south direction (Bassinger & Keller 1972), and the depth to a refractor of velocity $5 \cdot 35$ km s^{-1} near the centre of the saddle is $6 \cdot 2$ km (Ewing *et al.* 1957). Over the feature is a positive Bouguer anomaly with an amplitude of about 75 mgal (correction density $2 \cdot 1$ g cm^{-3}) with respect to the centre of the Tobago Trough. Two-dimensional gravity modelling shows that a rise in the basement of 3 km will explain the anomaly. The amplitude of the anomaly with respect to the area north of the saddle is only 30 mgal and consequently the relief of the ridge on that side is less and the slope is more gentle. Gravity anomalies of a similar type to this one also occur east of Martinique and Guadeloupe (Kearey *et al.* 1975).

East of the Barbados Ridge at approximately Lat. 14° N is an easterly trending positive Bouguer anomaly which can be seen clearly at Long. 57° W where it is not obscured by the strong gradient of the anomaly caused by the crustal depression beneath the Barbados Ridge. The amplitude of this anomaly is 30 mgal with respect to the region to the south and 50 mgal with respect to the region to the north. The peak of the anomaly lies over a rise in the basement which appears on two northerly running reflection profiles at Long. 57° 22′ W and 57° 44′ W (Fig. 9). Two-dimensional gravity-modelling shows that the rise is underlain by a small root and that the sediments on the crest of the rise are less than half the thickness of those further south.

The Bouguer anomaly map shows that this rise extends westward, apparently as far as the Barbados Ridge. At the same latitude the elevation of the Barbados Ridge Complex begins to decrease rapidly northwards, with the ridge itself disappearing as a prominent feature. The basement ridge may have influenced the early development of the Barbados Ridge Complex by acting as a barrier to sediment coming from the South American continent as turbidity flows. When the increased thickness of sediment south of the ridge was piled up to form the Barbados Ridge it achieved a greater elevation than the region north of the ridge. Bunce *et al.* (1971) have noted that east of the Lesser Antilles there is a general thickening of sediment towards the south, no doubt related to distance from sediment source.

5. The gravity effect of subducted lithosphere

The subducted lithosphere beneath the Lesser Antilles could be expected to cause a gravity anomaly contributing to the observed anomaly. Calculations of the likely gravity anomaly were made using published data on thermal models and phase relationships. The thermal model of subduction at a rate of 1 cm yr^{-1} with all heat sources included that has been computed by Toksöz, Minear & Julian (1971) was used, as this model is closest to the observed situation. The model uses the oceanic geothermal gradient of MacDonald (1965) and a value for shear strain heating that is greater than that of some other models (e.g. Griggs 1972). Turcotte & Schubert (1973), however, agree with Toksöz *et al.* on the amount of shear strain heating, but find that the minimum temperature in the descending lithosphere obtained by Toksöz *et al.* is higher than they would expect. The thermal model is a refinement of the earlier models of Minear & Toksöz (1970) which were also criticized for the high temperatures produced (Hanks & Whitcomb 1971; McKenzie 1971). It seems possible that the temperatures in the model are a little too high, leading to lower densities and a larger negative component in the gravity anomaly.

The maximum depth of the Benioff zone beneath the Lesser Antilles is 200 km. The subducted plate in the model reaches a depth of nearly 800 km. McKenzie (1969) and Griggs (1972) have put forward criteria relating the length of the seismic zone to the temperature anomalies within the subducted slab. According to these two criteria the maximum depth of the seismic zone given by the model would be 200 km and 120 km respectively. The length of lithosphere subducted beneath the Lesser Antilles, therefore, may be greater than that indicated by the Benioff zone, but it does not seem likely that the subducted lithosphere would exist as a distinct and separate entity very far below the end of the Benioff zone. The subducted lithosphere is unlikely to be as deep as 700 km because there has not been sufficient time for this to take place at a rate of 1 cm yr^{-1}. The maximum time available is about 80 My, and the history of the arc suggests that 40 My is the time span during which subduction has occurred. The former estimate gives a maximum depth to the subducted slab of 500 km and the latter gives 270 km. Increasing the rate of subduction to give a deeper slab would also increase the depth of the Benioff zone. Two models were considered; one reaching a depth of 200 km (the maximum depth of the Benioff zone); the other reaching a depth of 500 km. In both models the temperature field below the maximum depth of the slab was ignored. Toksöz *et al.* showed that the temperature field in the region already penetrated by the slab was not very time dependent and so the errors introduced by truncating a model down to 800 km depth at 200 km and 500 km are not large.

Gravity anomaly caused by thermal expansion

The relative mass excess or deficiency of the disturbed mantle was derived from the average temperature anomaly over a square cross section and represented as the mass per unit length of a line of infinite length passing through the centre of the square. The total anomaly was found by summing the effect of all the squares, which were spaced such that the depth to the top of a square was always greater than its width. The normal density distribution with depth, and the coefficient of thermal expansion were taken from Press (1970) and Birch (1952) respectively.

The gravity anomaly caused by phase changes in the mantle

The 350–400 km pyroxene–garnet and olivine–spinel B.Mg_2SiO_4 transitions only affect the 500 km deep model. The total density change is about 7 per cent (Ringwood 1970) giving a density contrast of 0·24 g cm^{-3}. The effect of the thermal anomaly of the model is to depress the transition below its normal depth giving a negative gravity effect.

The main factor influencing density changes in the uppermost part of the mantle is the relative proportion of orthopyroxene to garnet, which is dependent on the percentage of Al_2O_3 in the orthopyroxene (Green & Ringwood 1970). This percentage increases with increasing temperature, raises the proportion of orthopyroxene and consequently decreases the density. The lower temperature inside the subducted

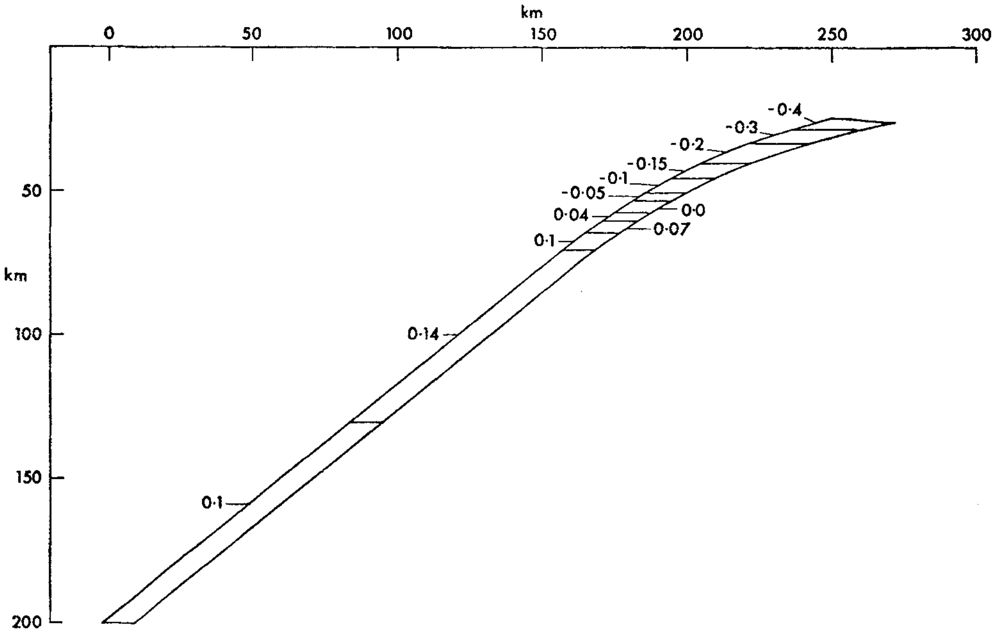

FIG. 17. A density contrast model of the crustal layer of subducted lithosphere. Contrasts in g cm^{-3}. Horizontal scale is in km east of Long. 62° W.

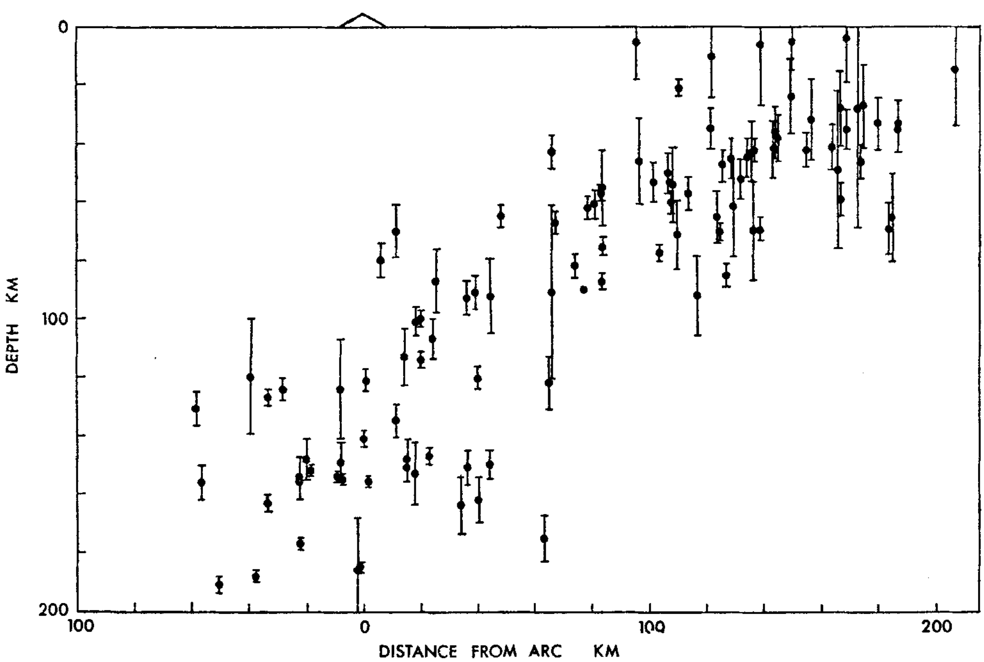

FIG. 18. Cross-sectional plot of earthquake hypocentres in the region of the Lesser Antilles during the period 1964–1970; determined by the ISC Edinburgh. Hypocentres are plotted radially from the arc and error bars on the depth are shown. (Taken from an unpublished compilation by J. Tomblin.)

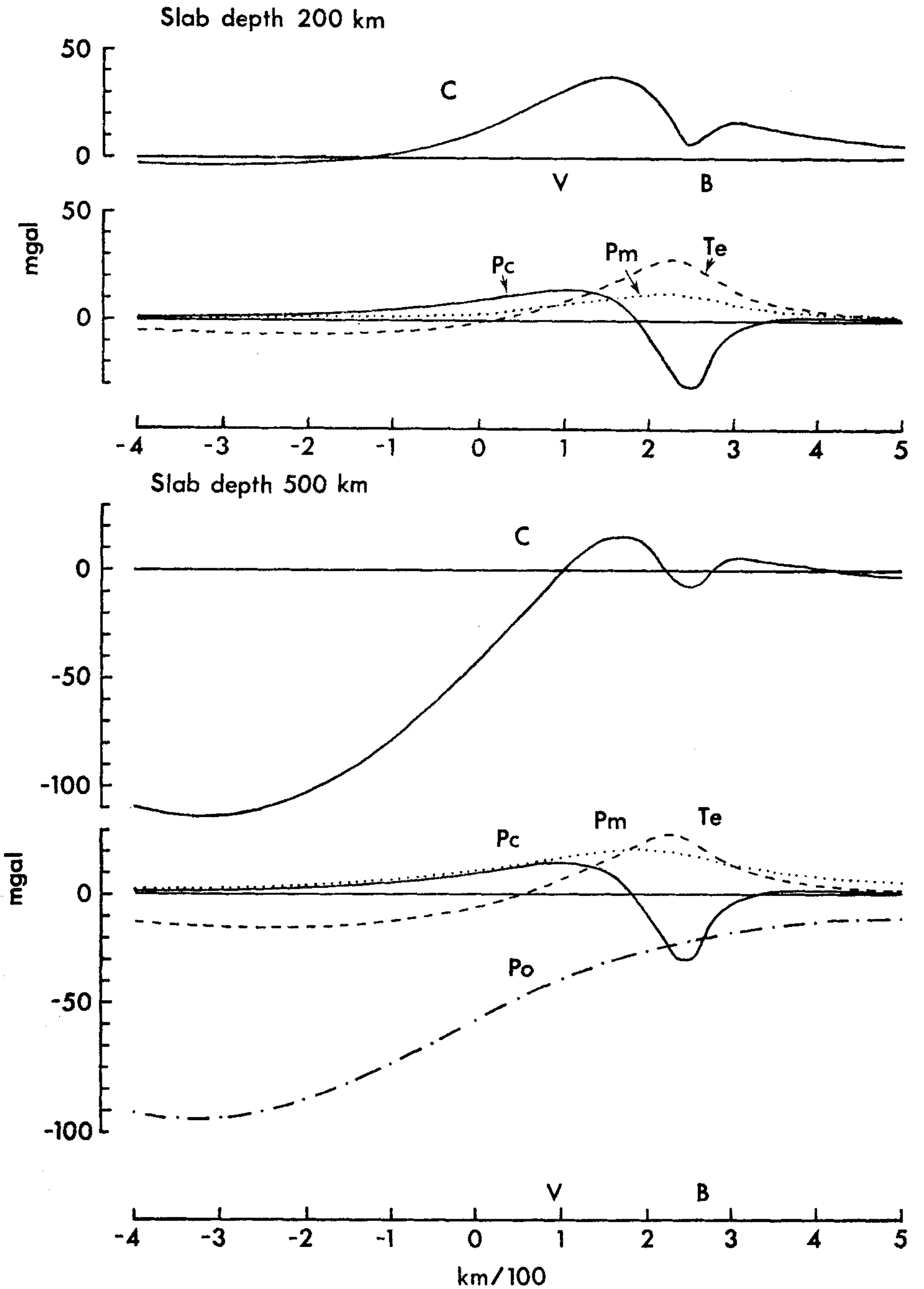

FIG. 19. Theoretical gravity anomalies caused by subduction of the lithosphere beneath the Lesser Antilles to a depth 200 km (upper diagram) and 500 km (lower diagram) at a rate of 1 cm yr^{-1}. Pc, the anomaly from phase changes and initial contrasts in the subducted crust. Pm, the anomaly from phase changes in the mantle. Te, the anomaly from thermal expansion and contraction. Po, the anomaly from phase changes in the olivine–spinel transition at 350 km depth (500 km model only). C, the resultant anomaly from the combined effect of the components given above. V, the position of St Vincent. B, the position of Barbados. Horizontal scale in km east of Long. 62° W.

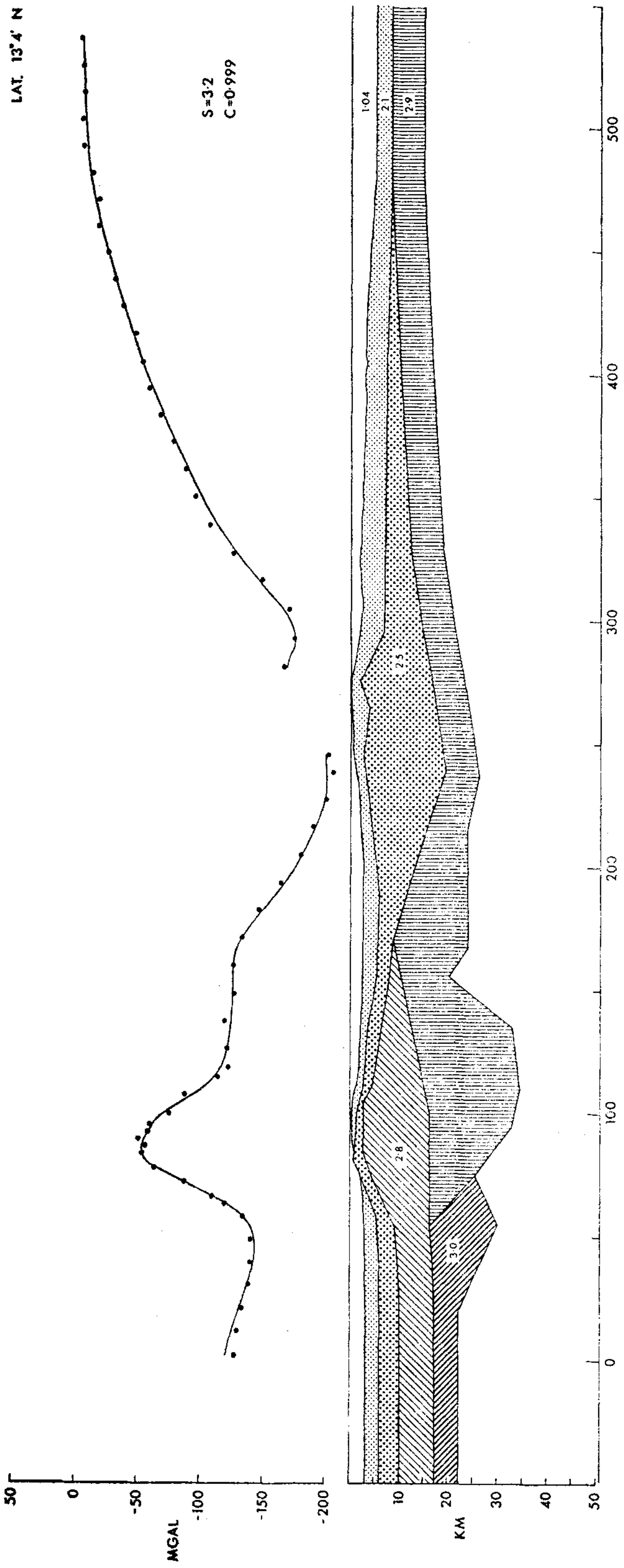

FIG. 20. Crustal model of the arc at Lat. 13° 4' N, derived using a Bouguer anomaly (correction density 2·1 g cm^{-3}) from which the gravity anomaly caused by lithosphere subducted to 200 km has been removed. Compare with Fig. 14.

lithosphere increases the proportion of garnet and gives it a higher density than the surrounding mantle. The densities of the phases used to calculate the gravity anomalies were taken from Haigh (1973a). The Pyrolite III model of Green & Ringwood (1970) is assumed for the mantle.

The gravity anomaly caused by phase changes and initial density contrasts in the subducted oceanic crust

The transformation of gabbro to eclogite has been shown by Ringwood & Green (1966) to have a considerable importance in subduction, because of the large increase of the density of subducted crust that occurs. Essene, Hensen & Green (1970) have shown that basaltic material under hydrous conditions passes through a phase of garnet amphibolite before finally reaching eclogite at a pressure equivalent to a depth of 70 km. They concluded that the amphibolite containing plagioclase feldspar was stable to depths of about 30 km whereafter the transformation from amphibolite to eclogite was a gradual reaction among amphibole, pyroxene and garnet.

A simple model of the density contrast between the subducted crust and the mantle was made using the data of Essene *et al.* (1970) for the crust and Green & Ringwood (1970) for the mantle (Fig. 17). The shape of the upper surface of the subducted plate was deduced from the distribution of earthquake hypocentres (Fig. 18), and the top of the model was fixed at the base of the crust derived from modelling (Fig. 14).

The computed gravity anomalies

The gravity anomalies produced by the two models are shown in Fig. 19. In the 500 km model the 350–400 km phase transitions produce a large negative anomaly of long wavelength. An anomaly of this size and position is not observed in the eastern Caribbean, and the 500 km model must be discarded. The thermal model may be incorrect or the subducted lithosphere does not reach a depth of 500 km. Also as the temperatures in the subducted slab below 300 km indicated by the thermal model are above those of normal mantle it is likely that the subducted lithosphere could not exist as a discrete entity below that depth.

The negative part of the anomaly produced by the 200 km model is very small and the anomaly reaches a maximum positive value of nearly 40 mgal. To gauge its possible effect on the determination of the crustal structure a crustal model was computed using a Bouguer gravity anomaly from which the gravity anomaly produced by the subduction model had been subtracted. The resulting model is shown in Fig. 20. Comparison with the model in Fig. 14 shows that the Moho is deeper beneath the centre and western side of the Tobago Trough. This is more consistent with time term analysis of LASP results across the trough, which gave Moho depths greater than those originally indicated by the gravity modelling (Fig. 14), although the variation in depth was similar (Boynton 1974). The minimum depth of the Moho beneath the Tobago Trough on the adjusted gravity model is 24 km and 23 km from the seismic results. Although not conclusive, the improvement in the consistency of the crustal model with the seismic data suggests that the gravity anomaly computed for the effect of subducted lithosphere could be fairly close in shape and amplitude to the actual anomaly resulting from subduction of the lithosphere beneath the Lesser Antilles.

6. Discussion

The volcanic arc

Although the gravity models indicate a crustal root 35 km thick beneath the island arc, the Moho has not been located seismically and consequently the true nature of the base of the arc remains uncertain.

The southern part of the volcanic arc shows little indication of migration of the volcanic front away from the subduction zone unless the slight asymmetry of the root implies westward growth. The more complicated pattern of magnetic anomalies on the eastern side of the arc may indicate a longer history of igneous intrusion and extrusion there than on the western side.

The difference in the structure of the crust either side of the Lesser Antilles may have some important genetic implications. The crust beneath most of the Tobago Trough with its single ($6 \cdot 9$ km s^{-1}) main crustal layer is more like that of the Atlantic than the two layer crust of the Grenada Trough ($6 \cdot 3$ and $7 \cdot 4$ km s^{-1}) which is similar to that of the Caribbean. If the island arc was simply built on a pre-existing Caribbean plate why should the structure differ either side of it? If the Grenada Trough was formed by some back-arc spreading process, does that imply a similar origin for the rest of the Caribbean? (Formed behind the Greater Antilles, perhaps.) It appears from the oceanic magnetic anomaly pattern east of the Lesser Antilles (Peter *et al.* 1973) and the early history of the opening of the Atlantic (Le Pichon & Fox 1971) that the age of the oceanic crust in the region of the Lesser Antilles before the arc was formed was Jurassic. The age of La Desirade is possibly further evidence of this (Mattinson *et al.* 1973). The oldest age obtained for the main Caribbean basin, however, is Coniacian (Edgar, Saunders *et al.* 1973). Although there has been speculation that the basalts sampled in Leg XV of the Deep Sea Drilling Project are not from oceanic basement (e.g. Donnelly 1973) it is quite probable that the crust west of the arc is younger than the crust between the arc and the subduction zone.

The easterly trending ridges east of St Lucia appear to have some significance in connection with variations of structure and geochemistry that occur along the arc. Between St Lucia and Guadeloupe a series of Bouguer gravity anomaly maxima lie 60 km east of the arc, extending southward from La Desirade the line of bathymetric highs that form a dissected shelf outside the northern outer arc (Kearey *et al.* 1975). No similar anomalies occur south of St Lucia. The petrology and chemistry of the volcanic rocks show marked differences north and south of St Lucia. Grenada, the Grenadines and St Vincent are more undersaturated than the islands north of St Lucia, falling in the field of high alumina basalt whereas the northern islands fall in the tholeiite field (Wills 1974). Also the southern islands show higher Sr 87/86 ratios than the northern islands (Stipp & Nagle 1974). One possible cause of these variations may have been differing rates of subduction resulting from movement on a transform fault running into the subduction zone just north of Lat. 14° N. There are several easterly trending ridges and troughs in the oceanic crust east of the Lesser Antilles (Peter & Westbrook 1974), but there are no indications of present-day activity on a transform fault, from the seismicity.

Arc front sediment complex

The great thickness of sediment beneath the Barbados Ridge and its disturbed nature, on Barbados and in the slope region to the east of it, implies tectonic emplacement. Considered in relation to the distribution of earthquakes and volcanoes in the Lesser Antilles, the most plausible explanation is that the sediments have been piled up as a result of the subduction of the Atlantic Plate beneath the Caribbean Plate (Chase & Bunce 1969; Westbrook, Bott & Peacock 1973). Seismic reflection profiles show the uppermost sediments of the ocean floor to be uplifted and typically tilted westward at the end of the slope of sediment. Absence of any clearly defined internal layered structure probably indicates deformation. The lower reflectors extend westward for some distance beneath the pile before they become involved in the deformation. These observations suggest that the sediment is stripped away from the subducting plate on successively deeper planes of decollement until basal decollement finally occurs between the sediment and the igneous crust (Westbrook *et al.* 1973). This

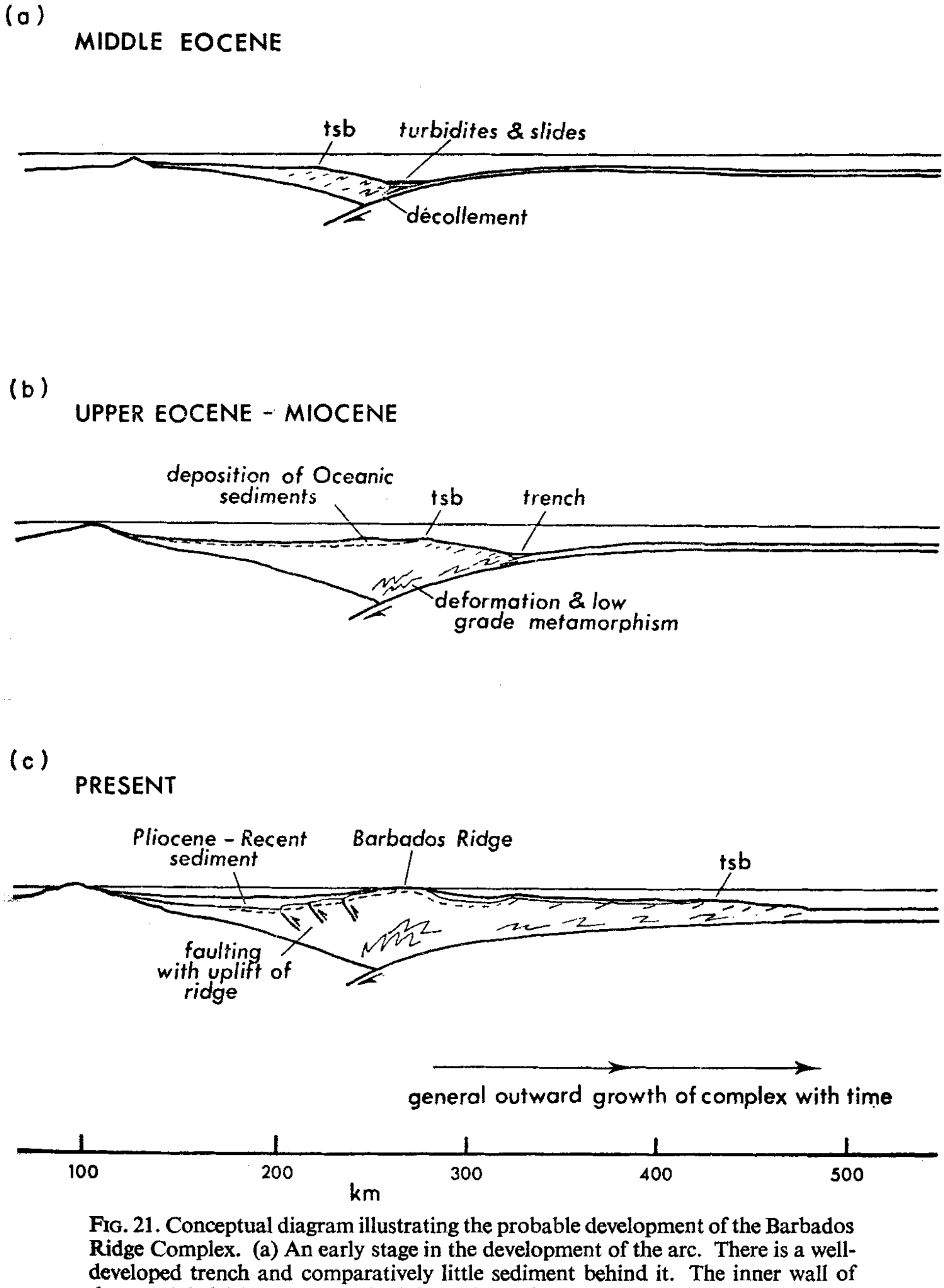

FIG. 21. Conceptual diagram illustrating the probable development of the Barbados Ridge Complex. (a) An early stage in the development of the arc. There is a well-developed trench and comparatively little sediment behind it. The inner wall of the trench is fairly steep. Much of the sediment incorporated in the sediment wedge (Scotland Formation) may have reached the trench by gravity sliding from the south triggered by seismic activity during the initiation of subduction. (b) With further development of the system the trench has migrated forwards and a sedimentary ridge is beginning to develop behind the trench slope break. The Oceanic Formation is laid down in the inactive region on and behind the sedimentary ridge. It is possible that during much of the time that the Oceanic Formation was deposited subduction may have ceased or continued at a very slow rate. (c) Present stage of development. The trench has disappeared and a sedimentary ridge is well developed. The increased resistance of the sediment pile to subduction produced by its increased width has apparently resulted in the uplift of the Barbados Ridge with accompanying faulting. It is also possible that during the evolution of the arc system the separation of the point of subduction and the island arc has increased by not more than 40 km. tsb, trench slope break. Horizontal scale in km east of Long. 62° W. Vertical exaggeration 2 : 1.

situation is similar in many respects to that in the Front Ranges of the Rockies, where long slices of sedimentary rock are successively thrust over a gently dipping crystalline basement (Bally, Gordy & Stewart 1966). As more sediment is added, the leading edge of the sediment pile must migrate away from the subduction zone, unless sediment is carried down into the mantle. The reason for the absence of a trench opposite the Lesser Antilles is that the leading edge of the sediment pile has migrated completely out of the crustal depression caused by subduction of the lithosphere (Fig. 21). The shape of the igneous crust beneath the sediment pile suggests that the trench had vanished after the sediment pile had migrated over 120 km from the position of subduction.

How does the Lesser Antilles arc compare with other arcs? The following features are common to most arc front systems. Behind the trench lies the inner wall or slope which terminates at a ridge or marked break in slope ('trench slope break' *cf.* Dickinson 1973). Between the ridge or trench slope break and the island arc lies a basin or platform containing undeformed sediment. The negative Bouguer gravity anomaly usually reaches a minimum over the sedimentary ridge or the trench slope break, which is also commonly at the outer limit of Benioff zone seismicity (Sykes 1966). The width of the slope is generally about 50 km. In some young arcs such as the New Hebrides a clear trench slope break is often absent (Karig & Mammerickx 1972). The slope east of the Barbados Ridge is on average 200 km wide. The evidence provided by gravity anomalies indicates that a sedimentary ridge is situated over the deepest depression of the crystalline crust, at the plate boundary (e.g. Grow 1973). The principal cause of the negative gravity anomaly is the depression of the crystalline crust, although subduction of crustal material may contribute. From the scheme outlined above, the width of the sediment slope would be expected to depend on age, rate of subduction, proximity of sediment source and the thickness of sediment on the adjacent ocean floor (Fig. 21). The slope in the New Hebrides (A young arc far from continental sediment sources) is only 20 km wide (Hayes & Ewing 1970; Karig & Mammerickx 1972) whereas the slope of the Java arc is 100 km wide on average. At the northern end of the Java arc, the Andaman and Nicobar islands occupy a position analogous to that of the Barbados Ridge and the width of the slope is 130 km (Weeks, Harbison & Peter 1967). It seems that the wider the slope, the better developed are the sediment ridge and the basin behind it. Arcs which have pronounced outer sedimentary ridges are old and close to continental sediment sources. Simple calculations assuming a constant rate of sedimentation show that the amount of sediment indicated by the gravity and seismic models to be in the Barbados Ridge complex could have been accumulated at a constant subduction rate of $0 \cdot 5 \text{ cm yr}^{-1}$ in 55 My. The situation is complicated by an apparent increase with age of the distance between the volcanic arc and the trench slope break, noted by Dickinson (1973). He suggests that this may occur by prograde migration of the subduction zone, or retrograde migration of the volcanic front, or both. Prograde migration of the subduction zone, in shortening the distance between plate boundary and trench, would tend to oppose the lengthening of the slope by the sediment accretion process outlined above.

The trench slope break and the sedimentary ridge have tended to be treated as synonymous, but it is possible that they are different features which occupy the same position in arcs which are not very fully developed. The inner wall of outer arc trenches has an average angle of slope of about 5° and the first major break of slope occurs 30–50 km from the trench. Profiles across the Aleutian arc (Grow 1973) show a major break of slope sometimes developed as a ridge 30 km from the trench. Fifty kilometres from the trench lies another higher ridge behind which lies a basin of undisturbed sediment. It is over the upper ridge that the Bouguer gravity anomaly reaches a minimum, and this upper ridge may be a less well-developed equivalent of the Barbados Ridge. The initial break of slope on the sediments of the Barbados

Ridge Complex is 50–60 km from the edge of the sediment pile, with an average angle of slope down to the ocean floor of 2·5°. A similar break of slope is present seaward of the Nicobar Islands in the Java arc, lying 50 km from the edge of the sediment pile with a slope down to the ocean floor of about 3° (Weeks *et al.* 1967). It is this initial break of slope which is the trench slope break and marks the maximum height to which sediment is raised during initial accretion. The sedimentary ridge is situated over the point of subduction and is a product of further deformation of the sediment pile. Until the leading edge of the sediment pile has migrated about 50 km from the position of subduction beneath crystalline crust, the sedimentary ridge will not appear as a separate feature. The sedimentary ridge as a distinct and separate feature is therefore a feature of arcs which have accumulated a large amount of sediment and in which the outward building of the sediment pile has been more important than outward migration of the subduction zone itself.

The crustal models (Figs 14 and 15) show that the mass of sediment in the Barbados Ridge Complex overlying the subducting Atlantic Plate is about twice that overlying the Caribbean Plate opposing it. Why does slippage continue between the Atlantic Plate and the overlying sediment, rather than the whole pile being pushed westward? The controlling factors are the slope of the basement beneath the sediment on the Caribbean Plate up which the sediment would have to be pushed, and the fluid pressure in the sediments, which lowers the resistance to shear stress on the thrust plane (Hubbert & Rubey 1959). It can be shown that (neglecting internal distortion of the sediment pile) if the coefficient of friction on the thrust is 0·7 (and it could easily be less) the fluid pressure in the sediments must exceed 0·8 × the lithostatic pressure for the thrust to operate as it appears to be doing. It is probable that at the depths considered the fluid pressure would be equal to the lithostatic pressure. By increasing the length of the thrust plane and the mass of sediment overlying it, the growth of the sediment pile must increase the stresses acting on the centre of the pile; the result of which is probably further deformation, faulting and increase of thickness causing uplift. This is presumably the main cause of the uplift of the Barbados Ridge.

A 4- to 5-km uplift of the ridge is indicated by the upwarp of horizons in the Tobago Trough and the present position above sea level of the Oceanic sediments, similar sediments to which were found at the base of JOIDES hole 27 in the floor of the Atlantic east of the Lesser Antilles (Bader *et al.* 1970). Most of this uplift has occurred during Pliocene to Recent times. The timing of this may be because during the period of deposition of Miocene limestones on the older volcanic arc there was a lull in subduction which was followed by the present period of activity which created the new inner arc and substantially increased the size of the sediment pile. The rate of uplift of Barbados over the past 120 000 yr has been between 0·3 and 0·45 km My^{-1} (Mesolella *et al.* 1969). Variations in isostatic response to subduction may also produce uplift and subsidence, the effect of which will be superimposed on the long-term tectonic uplift.

The trend of fold axes in the Scotland Formation of Barbados is east–north-east instead of northerly direction expected from a simple subduction model. This may be the result of oblique convergence between the Caribbean and the Atlantic at the time of deformation or subsequent rotation. Alternatively the deformation of the Scotland Formation may be related to its emplacement as gravity slides shortly after the initiation of subduction (Herrera & Spence 1964; Westbrook 1974); since when it has been unaffected by deformation produced directly by subduction, except for vertical movement. The minor deformation in the overlying Oceanic Formation implies that the zone of accretion and active deformation had migrated away from Barbados by Upper Eocene time.

Talwani (1971) and Watts & Talwani (1974) have examined the gravity high associated with a rise in the seabed on the seaward side of trenches. They have shown that this is probably caused by bending of the lithosphere as it is subducted. The

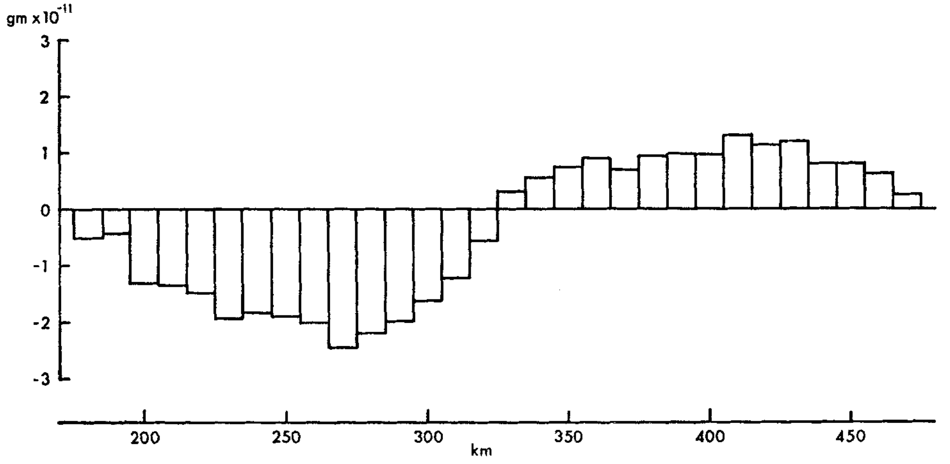

FIG. 22. Histogram of mass anomalies across the Barbados Ridge Complex, calculated from the crustal structure models. Anomaly given as mass per unit thickness section.

separation between the gravity low over the landward wall of the trench and the outer high is typically 150 km (Talwani 1971). The outer high does not appear clearly opposite the Lesser Antilles. However, the mass anomalies across the sediment pile (Fig. 22), derived by finding the relative mass excess or deficiency of 10 km wide columns with respect to the Atlantic from the gravity and seismic models, show a strong negative anomaly in the region of the Barbados Ridge and a positive anomaly 150 km east of it. These features, therefore, which are common to other island arc trenches, are present east of the Lesser Antilles, but are not obviously expressed because of the load of sediments covering them.

7. Conclusions

The island arc probably has a crustal root of about 35 km thickness with an internal discontinuity at a depth of 15 km. The Moho is depressed beneath the island arc and the Barbados Ridge, but rises to a depth of 24 km beneath the Tobago Trough. The Tobago Trough itself was formed as a consequence of the creation and uplift of the Barbados Ridge.

The volcanic arc and the Barbados Ridge Complex both show a change in structural character at 14° N, where an easterly trending ridge occurs in the basement.

A positive gravity anomaly of 40 mgal amplitude calculated as the possible effect of subducted lithosphere is compatible with the interpretation of the crustal structure.

The Barbados Ridge Complex is a pile of sediment that is deformed and to some extent metamorphosed. It has a maximum thickness approaching 20 km beneath the Barbados Ridge, where the igneous crust of the Atlantic Plate is subducted beneath the leading edge of igneous crust of the Caribbean Plate. The sediment pile appears to have accreted as a consequence of subduction, and to have grown with time away from the island arc, losing its trench in the process. The oldest sediments metamorphosed at the base of the pile are possibly late Jurassic. Since the end of the Miocene, Barbados has been uplifted about 4–5 km. The southern part of the Barbados Ridge Complex is one of the most developed examples of this type of feature in the world. The development of a sedimentary ridge is itself characteristic of a mature island arc. The structure of the arc and the processes thought to be active in its creation are illustrated in Fig. 23.

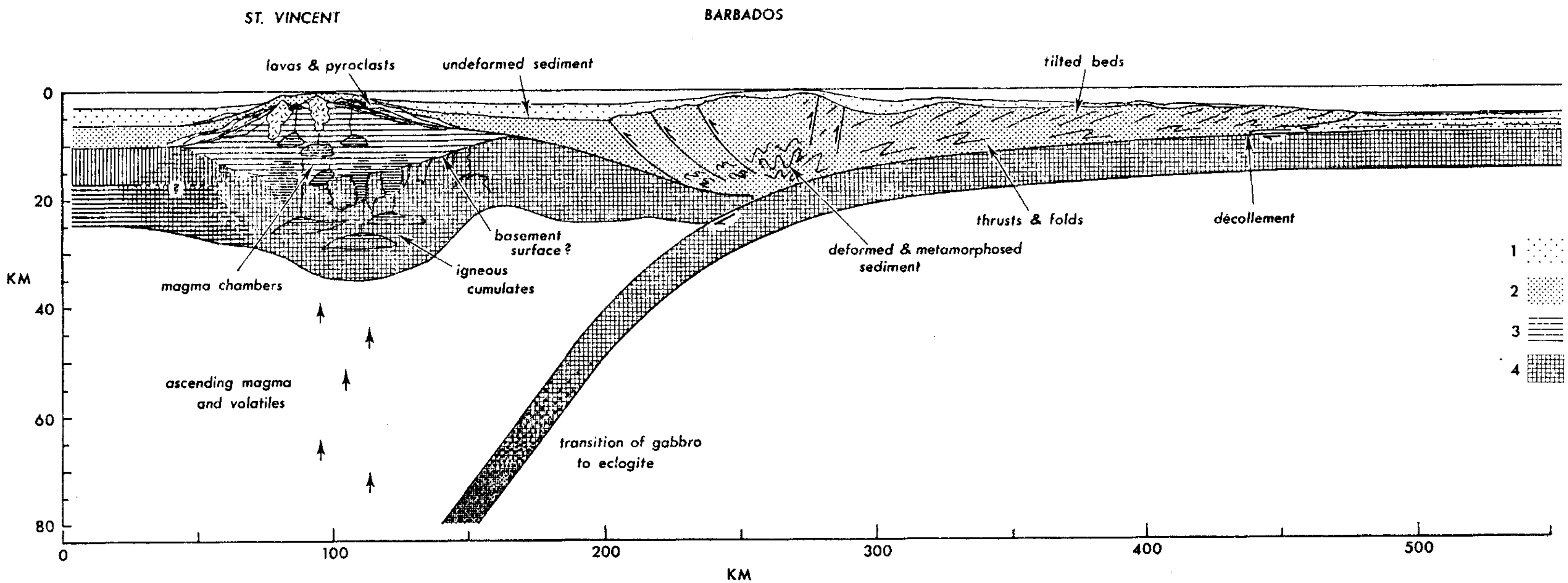

FIG. 23. Diagrammatic cross-section of the Lesser Antilles island arc illustrating the structure and the processes acting on it. 1. Undeformed sediment. 2. Deformed and/or consolidated sediment. 3. Igneous crust produced by the volcanic arc. 4. Main oceanic crustal layer and lower crust of arc. Vertical exaggeration 2 :1.

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Department of Geology,
The University,
Keele,
Staffs. ST5 5BG.

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Appendix 1

Seismic refraction results

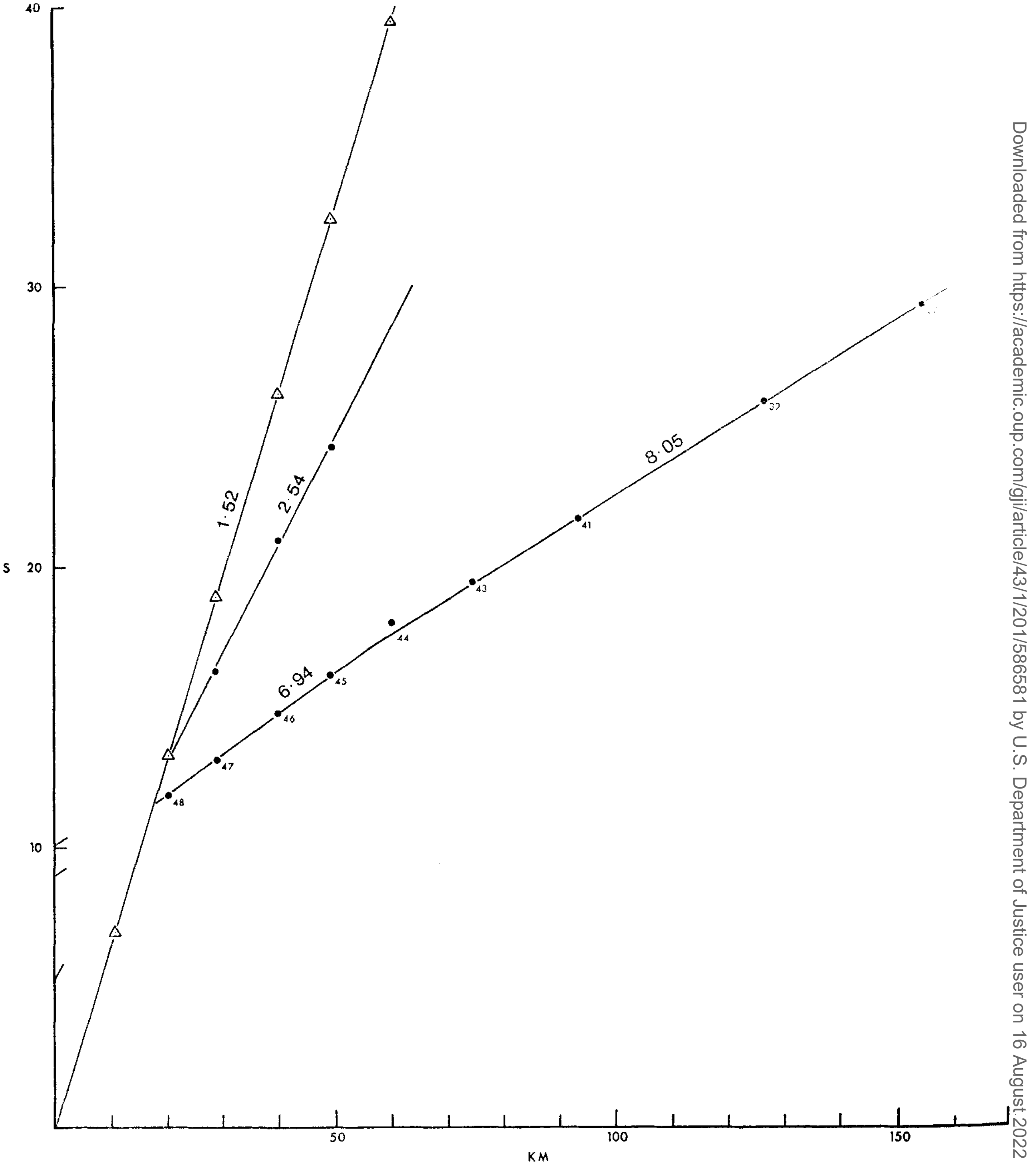
For locations see Fig. 4

*Hydrophone station at position h*1

Uncorrected apparent velocities: 2·54, 6·94 and 8·05 km s^{-1}. After correction for the variation in depth of water the following results were obtained:

Velocity km s^{-1}	Intercept s	Standard error of fit s
2·45±0·09	5·02±0·66	0·22
6·61±0·06	9·04±0·05	0·03
7·79±0·19	10·19±0·37	0·20

This line was not reversed by the stations on Barbados. Nearby reversed profiles by Ewing *et al.* (1957) gave velocities of 6·64–6·77 and 8·09–8·32 km s^{-1} for the



Hydrophone 1

FIG. A1

main crustal layer and upper mantle respectively. The following table shows combinations of the true velocities in the range measured by Ewing *et al.* and the angles of dip that will give the apparent velocities observed at h1.

Dip of refractor	0°	1°	2°	2·5°	(Velocity of overlying layer of varying thickness)
Velocity km s^{-1}	7·79	8·02		8·33	4·0
,,	6·61	6·67	6·94		4·0
,,	6·61	6·92	7·26		2·44

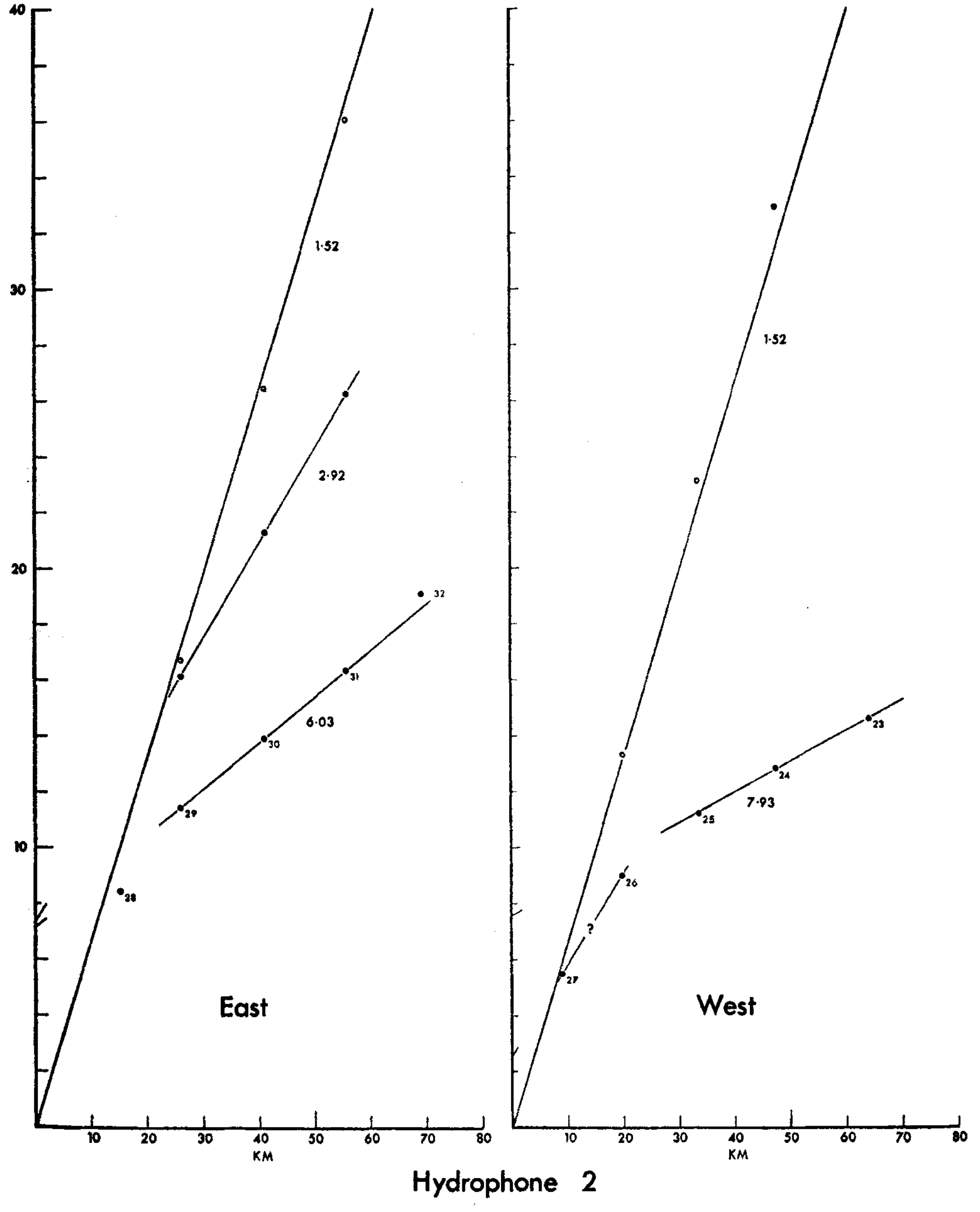
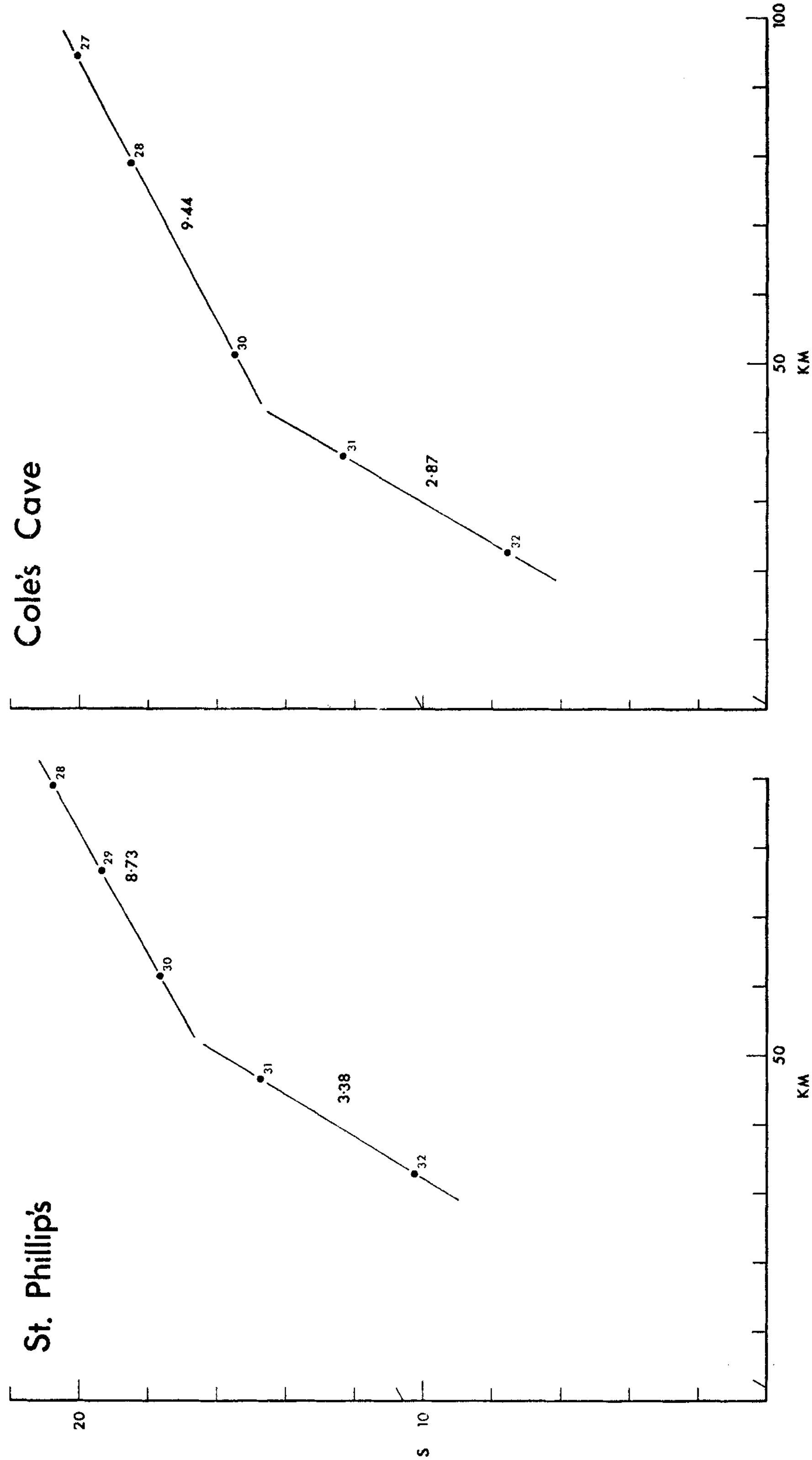


Fig. A2



St. Phillip's
Cole's Cave
27
28
29
30
31
32
8·73
3·38
9·44
2·87
50
100
KM
20
10
S

Fig. A3

From the data of Ewing *et al.* (1957) it is clear that the dip of the igneous crust beneath the shot points leads to a variation in of thickness of the $4 \cdot 0$ km s^{-1} layer. This layer is probably not present beneath h1, however. The following velocity layering beneath h1 was calculated.

	Velocity km s^{-1}	Thickness km	Depth km
Water	1·5	4·46	=depth below lowest h/phone
Sediment (assumed)	1·7	0·38	4·46
Refractor 1	2·45	3·68	4·84
Refractor 2	6·73	6·57	8·52
Refractor 3	8·02		15·1

Hydrophone station at position h2–Tobago Trough

Uncorrected apparent velocities east of h2: 2·92 and 6·03 km s^{-1}
Uncorrected apparent velocity west of h2: 7·93 km s^{-1}.
After correction for variation in water depth using an average sediment velocity of 3·0 km s^{-1} the following results were obtained:

	Velocity km s^{-1}	Intercept s
East of h2	$5 \cdot 76 \pm 0 \cdot 08$	$6 \cdot 81 \pm 0 \cdot 10$
West of h2	$7 \cdot 22 \pm 0 \cdot 13$	$6 \cdot 70 \pm 0 \cdot 12$

Barbados–St Phillips and Coles Cave

Uncorrected apparent velocity at St Phillips: 8·73 s^{-1}.
Uncorrected apparent velocity at Coles Cave: 9·44 km s^{-1}.
After correction for variation in water depth using a sediment velocity of 3·0 km s^{-1}:

	Velocity km s^{-1}	Intercept s
St Phillips	$8 \cdot 98 \pm 0 \cdot 3$	$10 \cdot 05 \pm 0 \cdot 29$
Coles Cave	$9 \cdot 54 \pm 0 \cdot 05$	$9 \cdot 41 \pm 0 \cdot 04$

With the apparent velocities measured at the stations on St Vincent and St Lucia (Boynton 1974), reversed profiles were obtained across the western and eastern halves of the Tobago Trough.

West Tobago Trough	h2	to	St Vincent	&	St Lucia
Apparent velocity	7·22		5·59		5·64
Intercept	6·7		1·86		1·77

Velocity in overlying layer	True velocities and dips of refractor			
	velocity	dip	velocity	dip
3·0	6·27	3·95° E	6·31	3·80° E
3·5	6·26	4·90° E	6·305	4·70° E
4·0	6·25	6·02° E	6·30	5·77° E

East Tobago Trough	h2	to	St Phillips	&	Coles Cave
Apparent velocity	5·76		8·98		9·54
Intercept	6·81		10·05		9·41

Velocity in overlying layer	True velocities and dips of refractor velocity	dip	velocity	dip
3·0	6·98	5·93° E	7·14	6·34° E
3·5	6·96	7·76° E	7·12	8·13° E
4·0	6·94	8·78° E	7·10	9·42° E

Using the refraction data of Ewing *et al.* (1957) for the upper layers in the Tobago Trough the following velocity layering was obtained for the structure beneath h2.

Velocity layering beneath h2

Velocity km s^{-1}	Thickness km	Depth km
1·5	2·12	water
1·7	0·95	2·12
2·5 or 3·0	1·75 or 2·10	3·07
4·0	4·15 or 4·27	4·82 or 5·17
6·3 or 7·0		8·97 or 9·44

Appendix 2

Wide angle reflection profile in the Tobago Trough

For location see Fig. 5.

Layer velocities and thicknesses

Velocity km s^{-1}	Thickness km	Dip	Standard error of fit s
1·50	2·52	0·0°	0·012
1·54	0·19	0·0°	0·019
1·55	0·22	0·0°	0·019
1·90	0·48	1·1° E	0·013
2·49	0·47	1·0° E	0·009
2·51	0·34	0·0°	0·022

Details of derivation are given in Westbrook (1973).