PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

The Geology and Gravity Anomalies of the Troodos Massif, Cyprus

I. G. Gass and D. Masson-Smith

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THE GEOLOGY AND GRAVITY ANOMALIES OF THE TROODOS MASSIF, CYPRUS

By I. G. GASS
Department of Geology, The University, Leeds 2

AND D. MASSON-SMITH
Geophysical Division, Overseas Geological Surveys

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CONTENTS

Over Cyprus there is one of the largest recorded gravity anomalies which reaches a maximum of over +250 mgal. This paper records the main geological features of the island, investigates the source of the gravity anomaly and correlates both lines of evidence in support of an hypothesis on the evolution and structure of the area.

The topography of Cyprus, which lies in the north-eastern Mediterranean, is dominated by two east–west mountain ranges separated by the low-lying central plain of Mesaoria. The northern, Kyrenia range is part of the southernmost arc of the Tauro-Dinaric Alps, whilst the southern Troodos range is an igneous massif composed of basic and ultrabasic rocks of plutonic and extrusive character. The Troodos rocks fall logically into three main units: (*a*) the Sheeted Intrusive Complex; (*b*) the Troodos Plutonic Complex; and (*c*) the Troodos Pillow Lava Series.

The *Sheeted Intrusive Complex* forms the major part of the Troodos massif and is a north–south basic dyke swarm cutting basic lavas. The dykes range in thickness from 1 to 15 ft. and form over 90 % of the complex. Abundant evidence is available to substantiate the intrusive nature of this dyke complex. Its unique concentration and regularity is attributed to repeated intrusion coupled with intense erosion. The north–south orientation of the intrusives is thought to be due to the east–west tensional stress that was dominant throughout the evolution of the massif.

The central part of the massif is occupied by the *Troodos Plutonic Complex*, a layered ultrabasic complex of batholithic dimensions in which the rock types range from central dunites and

peridotites outwards through melagabbros and olivine-gabbros to gabbros and granophyres. Field, mineralogical and geophysical data indicate that the parent material was of peridotitic composition. Although gabbros are, by far, the most abundant rocks exposed, it is considered that these represent but a minor percentage of a vast mass of underlying, high-density, ultrabasic material. Differentiation of the ultrabasic parent material is thought to have resulted in the gradual upward and outward change from central dunites and peridotite through melagabbros and olivine-gabbros to overlying gabbros and granophyres.

Forming an incomplete ring around the Sheeted Intrusive Complex is the *Troodos Pillow Lava Series*, a very thick sequence of pillow lavas and their related intrusives. Although divided into two units on the presence of a partial unconformity, and petrographic differences, the general basaltic nature of the series persists throughout. The series shows an increase in basicity with decreasing age, the main rock type in the lower unit being basalt, whilst olivine-basalts predominate in the upper division. There is evidence that this series has resulted from the partial fusion of a rock of peridotitic composition and that the relationship between age and basicity is due to the progressively more complete fusion of the parent material.

Serpentines of post-Lower Triassic age and considered to be the initial phase in the igneous activity of the Alpine orogeny are also present in Cyprus, where they appear to have been emplaced as a serpentine 'magma'.

Cyprus is covered by a strong positive gravity anomaly mainly between 100 and 250 mgal. The axis of the anomaly lies over the Troodos massif, runs parallel to the Kyrenia range and extends from Pomos in the west, eastwards to Famagusta; superimposed upon the main anomaly are smaller local anomalies. The gravity field falls off all round Cyprus to less than 100 mgal; no other gravity anomalies of this size have, so far, been found in the eastern Mediterranean. The high-density rocks, which appear to have produced this large anomaly, have the form of a rectangular, near-surface, subhorizontal slice, which measures 120 miles east–west by 70 miles north–south and whose centre is displaced about 20 miles to the north-west of the centre of Cyprus. This high-density mass must be at least 7 miles thick under Mount Olympus, whilst at Pomos a thickness of over 20 miles is estimated; its maximum elevation is at Mount Olympus where the dunites and peridotites of the Troodos Plutonic Complex crop out. A correlation between the ultrabasic rocks of the Troodos Plutonic Complex and the high-density material causing the main anomaly is well substantiated.

The geological and geophysical evidence suggests that the Troodos massif evolved in pre-Triassic times as an oceanic volcanic pile situated between the then more widely spaced continental masses of Africa and Eurasia. During the Alpine orogeny these continental masses converged, the southern mass underthrusting the Troodos volcanic pile, and parts of the Eurasian hinterland. The underthrusting took place at such a level that not only the volcanic pile but also part of the upper mantle was uplifted above sea level as an undeformed slice. Intense erosion has denuded the volcanic pile almost to its roots. It is thought that the stratiform Troodos Plutonic Complex might represent upper mantle material, partly fused and differentiated to provide the basic volcanic rocks of the Troodos massif.

1. Introduction

Cyprus lies in the north-east corner of the Mediterranean. It is a relatively small island of 3572 square miles, having maximum dimensions of 60 miles north–south and 140 miles east–west. The island is 45 miles south of Turkey; at the nearest point it is only 60 miles west of Syria, and Egypt is some 250 miles to the south.

Topographically, Cyprus can be divided into three main regions; the Kyrenia range to the north, the Troodos mountains in the south and the broad fertile plain of the Mesaoria separating the two upland areas.

The Kyrenia range runs parallel to the north coast at an average distance of 5 miles inland and rises to elevations of over 3000 ft. This range is markedly elongate, rarely more than 3 miles wide, and consists mainly of thrust slices of Permian to Cretaceous limestone. It is flanked by concurrent flysch deposits of Oligocene to Middle Miocene age. The Troodos mountains have the shape of a flattened oval, the long axis being east-south-east and about 60 miles long. The range is between 15 and 20 miles wide, occupying an area of about 900 square miles. Composed of basic and ultrabasic igneous rocks, this massif rises to a maximum height of 6401 ft. on Mount Olympus. Partly surrounding the igneous massif are hills of Cretaceous to Miocene chalks which form a series of low but prominent cuesta. Extending east–west across Cyprus and separating the two upland areas is the Mesaoria plain. This plain is formed of Pliocene, Pleistocene and Recent sediments and only rarely has an elevation of over 1000 ft. South of the Troodos mountains, and especially in the south-west of the island, there is a broad, relatively level expanse of Cretaceous to Miocene calcareous sediments. In places erosion has removed this chalk cover and exposed the structurally complex Triassic to Lower Cretaceous rocks of the Trypa Group.

One of the largest recorded positive gravity anomalies occurs over Cyprus. This anomaly is confined to Cyprus and the adjacent sea areas, the axis of maximum anomaly lying over the Troodos massif. Investigation of the anomaly indicates it to be due to an extensive slab of high-density rock, at least 7 miles thick, which underlies the Cyprus area at shallow depth.

Geological and geophysical evidence suggests that this slab was once part of the upper mantle underlying an oceanic area between the African and Eurasian continents. It will be suggested that when the continental shields approached each other during the Alpine orogeny this slab of mantle was underthrust by the edge of the African shield and thereby raised to its present level in the upper part of the crust.

The geology and the gravity anomalies of Cyprus are discussed separately in the first and second parts of this paper. An independent presentation of these two lines of evidence is necessary as there is limited correlation between superficial geological structure and the major gravity anomalies which all seem to originate from unexposed deep-seated structures. In the third part of the paper the geological and geophysical evidence is brought together in support of our concept of the evolution of the Cyprus structure.

2. General geology of Cyprus

2·1. *Previous geological work*

The first study of Cyprus geology was made by Gaudry (1862) who made a reconnaissance survey of the island which was concerned mainly with the sedimentary rocks. Subsequently, Unger & Kotschy (1865), Russell (1882) and Bergeat (1892) all contributed to the geological knowledge of the island. In 1905, Bellamy produced a geological map of the island and in the same year, with the assistance of Jukes-Browne, he published a book entitled *The Geology of Cyprus*. Bellamy's map remained the basis for all further work until the Geological Survey was initiated in 1950. A major contribution to the understanding of the geology of the Troodos massif and the origin of the cupriferous deposits was made by Cullis & Edge (1922). In 1949, Henson, Browne & McGinty published a paper which described the stratigraphic succession of the sedimentary formations and compared them to the synchronous rocks of the mainland. In addition to this paper Browne & McGinty provided information for a geological map that was printed by 42nd Geological Section, S.A.E.C., G.H.Q., M.E.F., in 1946. The boundaries shown on this map are mainly those of Bellamy, revised in places where Browne & McGinty had undertaken detailed work. Bishopp (1952*a*, *b*) published two short papers on the structure of the Troodos massif.

The Cyprus Geological Survey was inaugurated in 1950 and detailed mapping of the Troodos mountains on the scale of 1:5000 began in 1951. The results of this work are presented in the survey memoirs. Memoir No. 1 was published in 1959, Memoirs Nos. 2 to 5 during 1960.

Much of the information summarized in this paper is taken from the *Cyprus Geological Survey Memoirs* and from the annual reports of that survey for the years 1955–59. The works of Bagnall, Bear, Bishopp, Carr, Ingham and Wilson have been extensively consulted. About half of the Troodos massif has now been surveyed in detail and the remainder has been traversed at wider intervals.

2·2. *Review of the geology and structure of Cyprus*

Cyprus is divisible into five roughly parallel belts. These belts trend approximately east–west and are gently convex to the south. From north to south they are:

(1) The Kyrenia range consisting of a narrow mountainous belt of upthrust slices of limestones of mainly Jurassic age, serpentines, Triassic red shales, radiolarites and basic igneous rocks. Also included in this belt, but topographically not part of the Kyrenia range, are the deformed Miocene flysch deposits of the Kythrea Formation that flank the mountain range. The structure of this calcareous flysch is parallel to that of the main thrust front.

(2) The Mesaoria plain. This broad belt, which separates the tectonically deformed sediments of the Kyrenia range from the contemporaneous undeformed sediments that flank the Troodos igneous massif, is formed of Pliocene, Pleistocene and Recent sediments. These horizontally disposed sediments lie, with marked angular unconformity, on the folded flysch deposits of the Kythrea Formation and also have an unconformable contact with the gently dipping chalks that flank the Troodos mountains. They completely mask the contact between the deformed sediments of northern Cyprus and the undeformed contemporaneous chalks that flank the Troodos massif.

(3) Flanking the Troodos massif to the north are gently dipping calcareous sediments of Upper Cretaceous (Maestrichtian) to Middle Miocene (Tortonian) age.

(4) The Troodos igneous massif: the massif has a threefold, roughly annular structure: (i) the basic and ultrabasic plutonic rocks which occur mainly in the centre of the range; (ii) the Sheeted Intrusive Complex that forms the main part of the range; (iii) the peripheral lavas of the Troodos Pillow Lava Series (see figure 3).

(5) The southern foothill belt consists of gently folded Upper Cretaceous to Middle Miocene calcareous sediments which overlap unconformably on to the igneous rocks of the Troodos range. Large inliers of the tectonically complex Trypa Group are exposed within this belt as are small exposures of schistose rocks that have been described by Bear (in Ingham 1957). These belts can be identified on figure 1.

It is not intended to discuss the sedimentary rocks of Cyprus in detail as this paper is primarily concerned with the deep-seated structure of the igneous rocks which causes the large positive gravity anomalies.

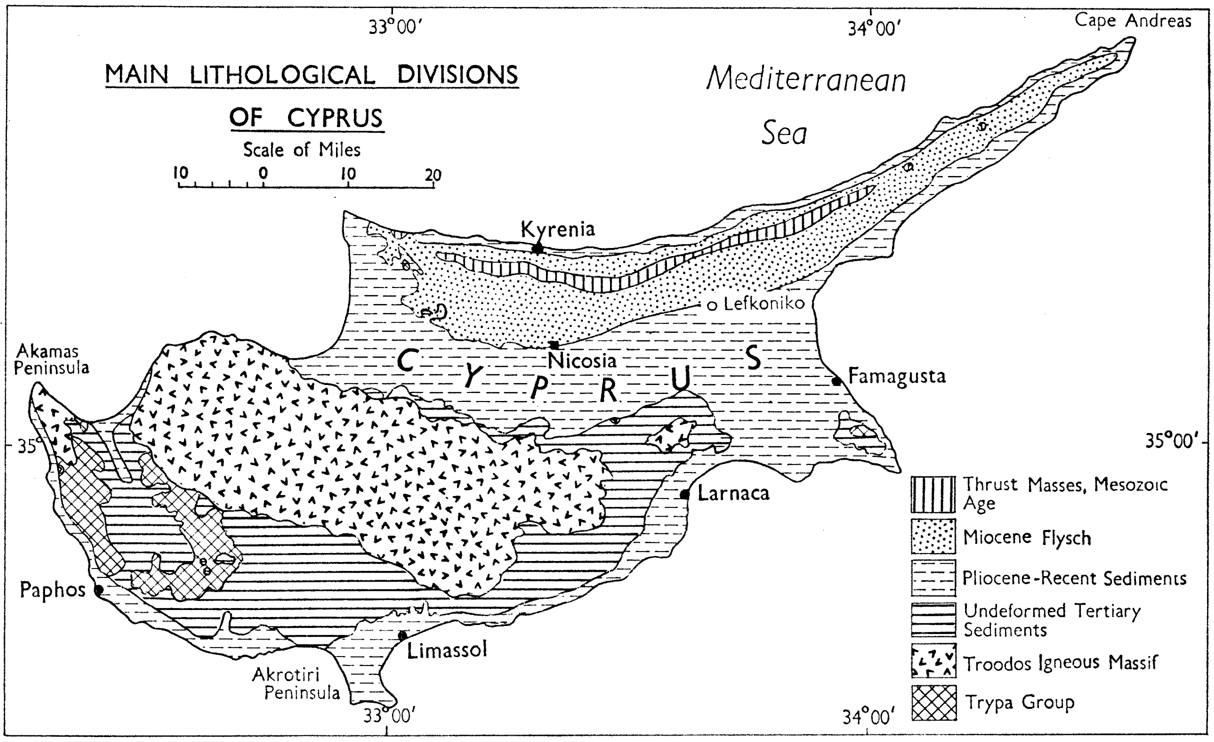

FIGURE 1

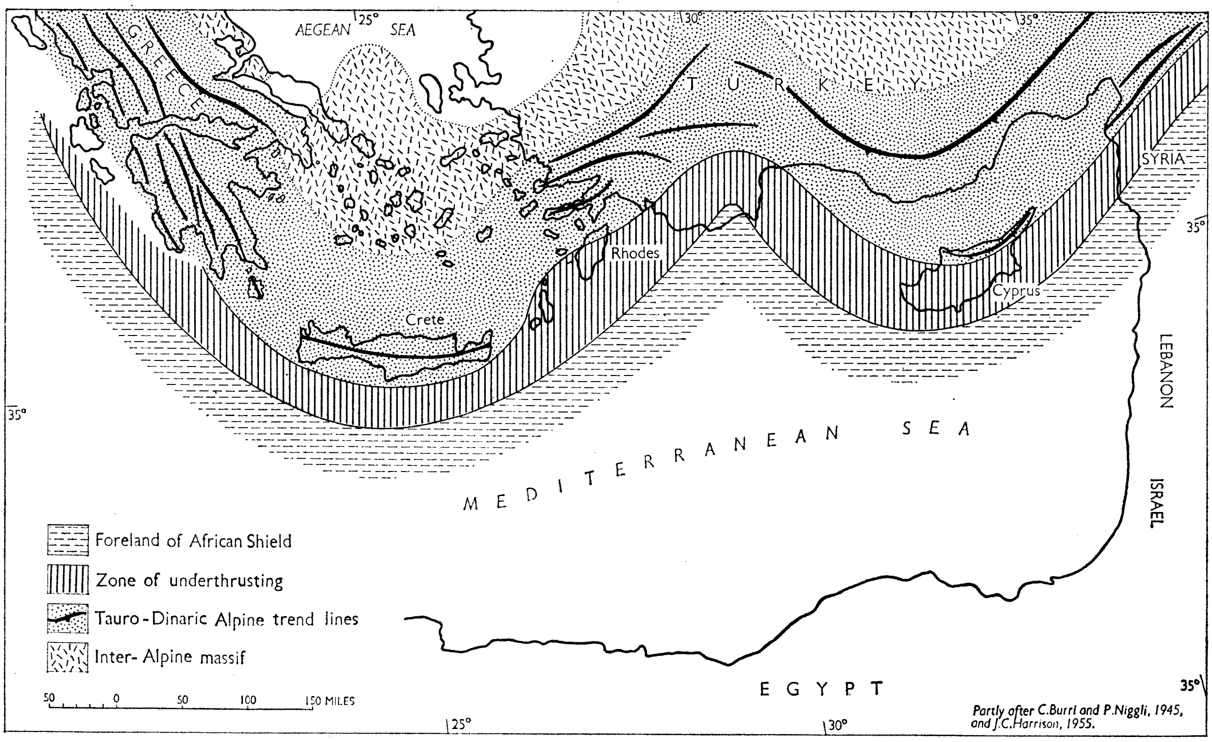

FIGURE 2. Tectonic zones of the eastern Mediterranean.

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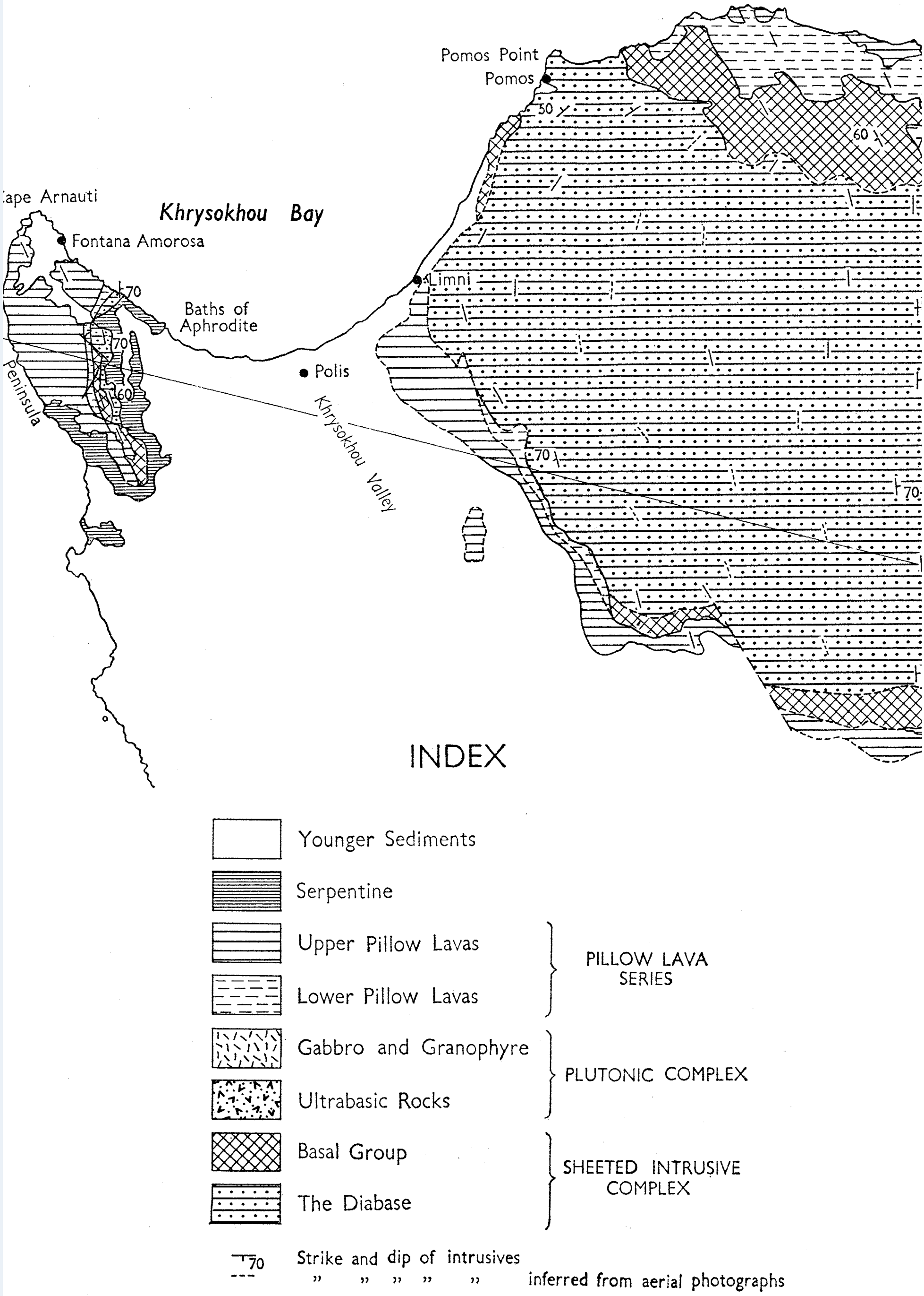
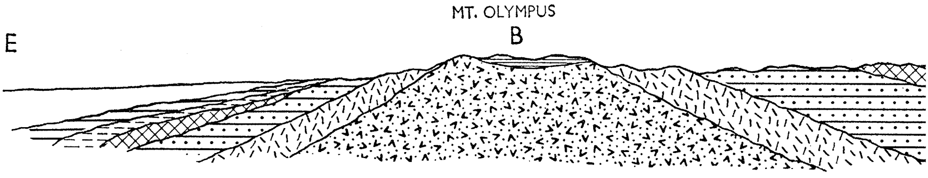
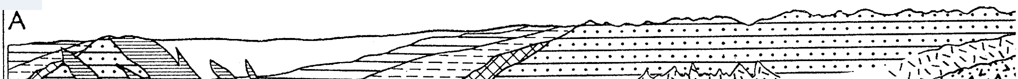




Diagrammatic Section along EBF



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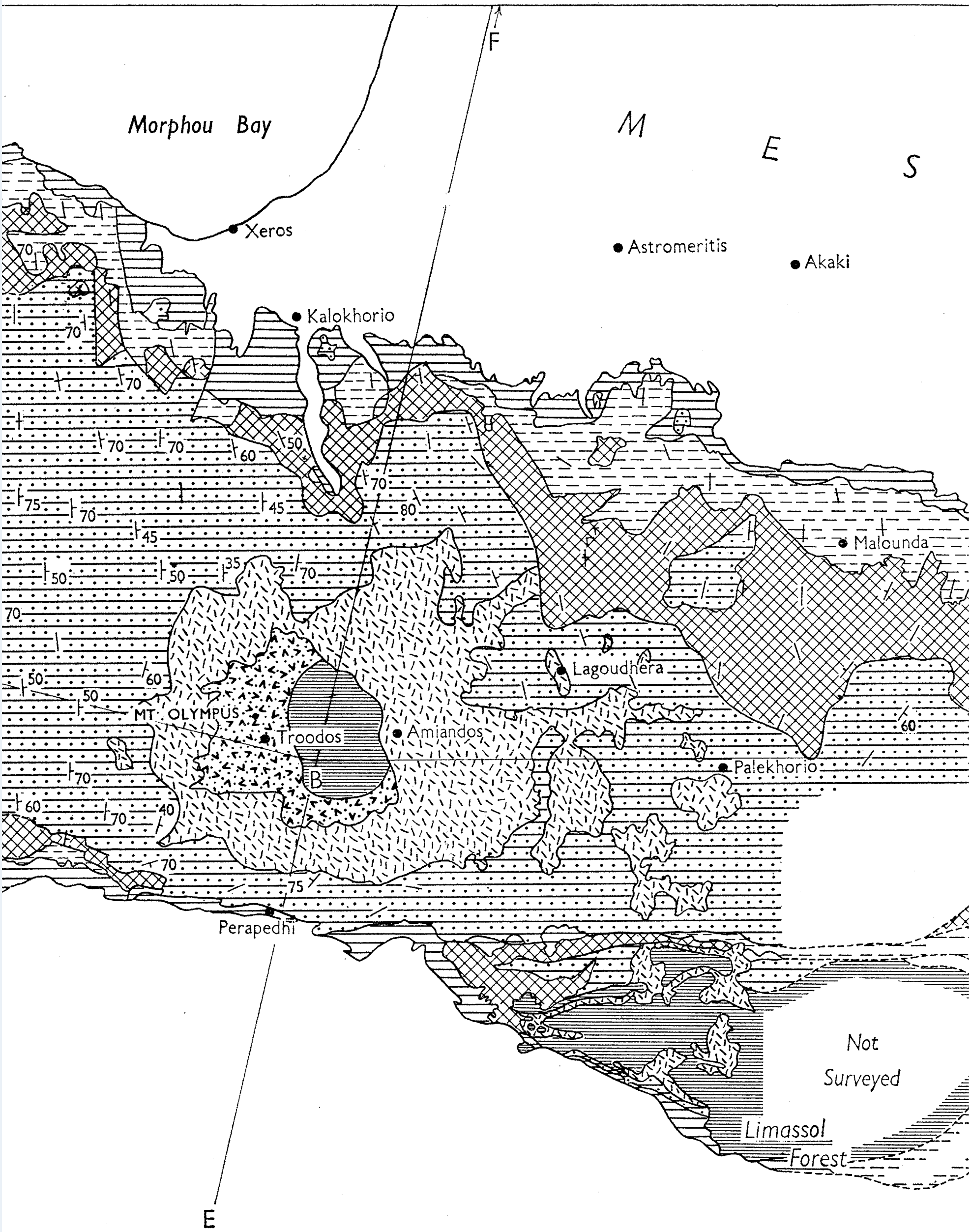

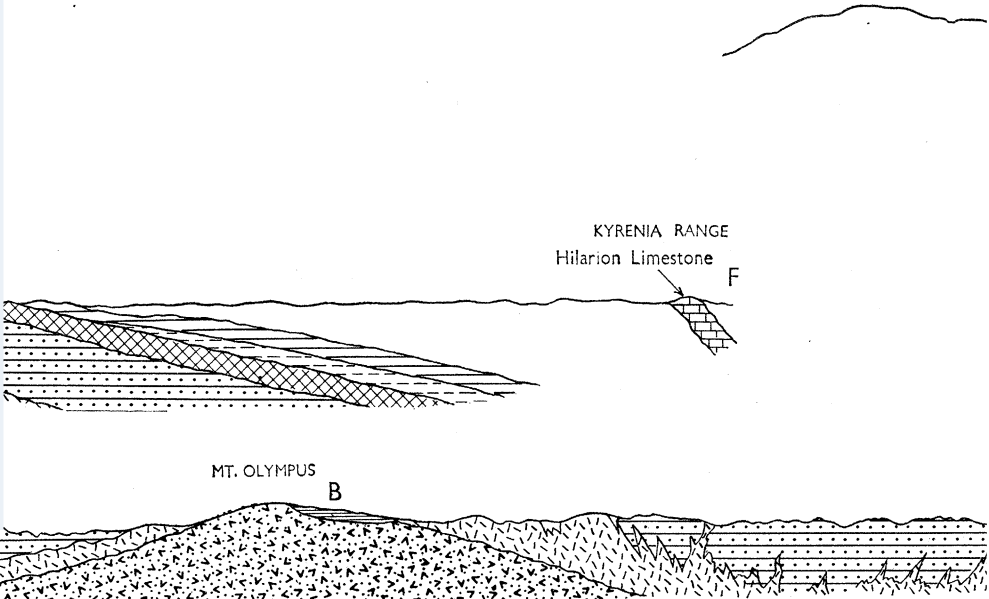

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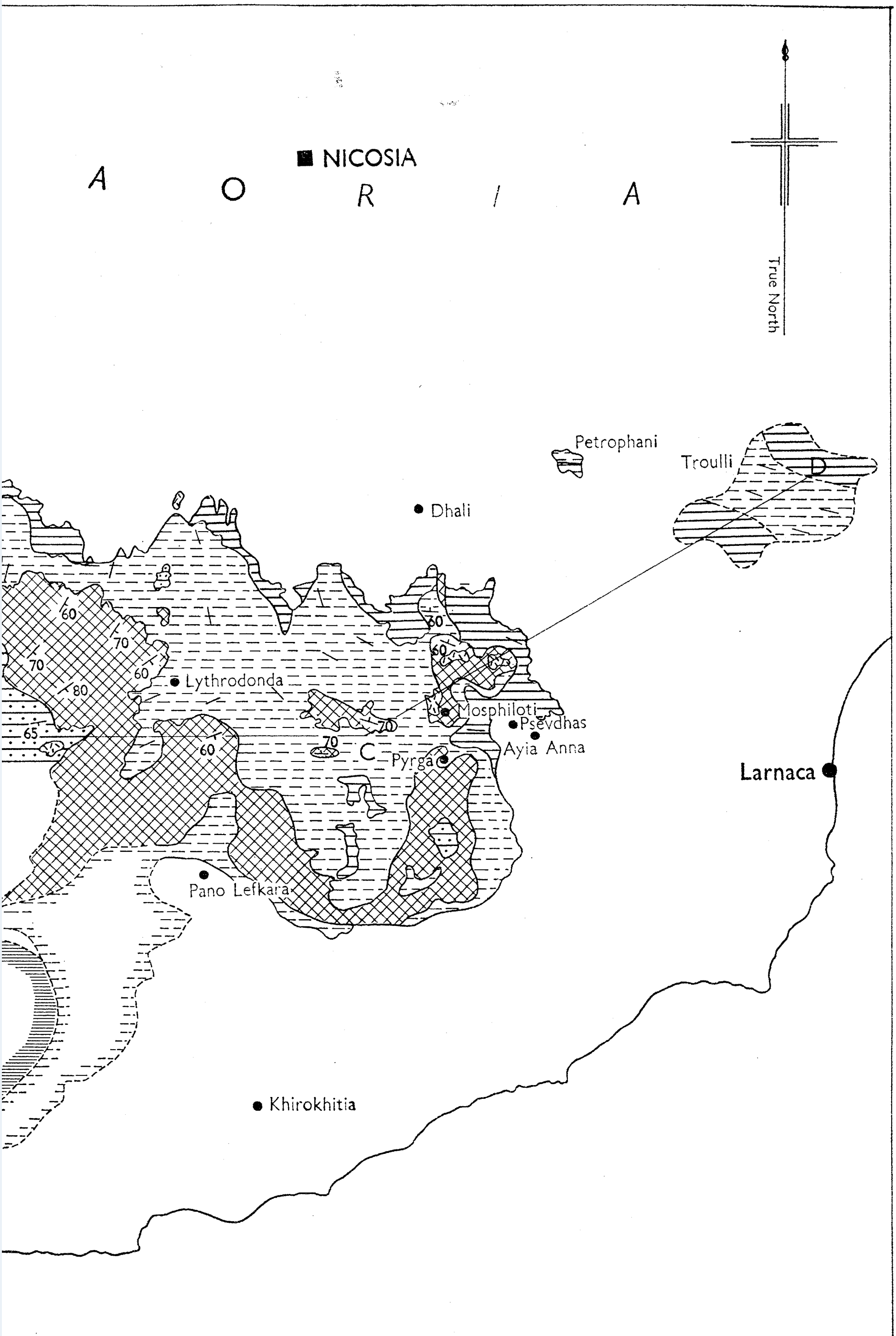
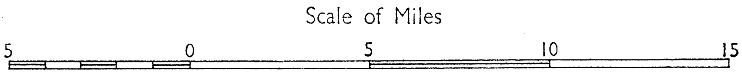
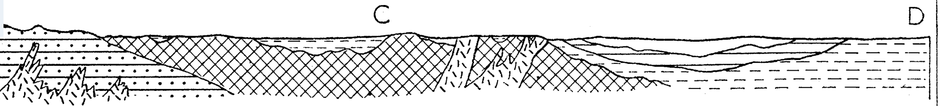


Geological Sketch Map of the TROODOS MASSIF





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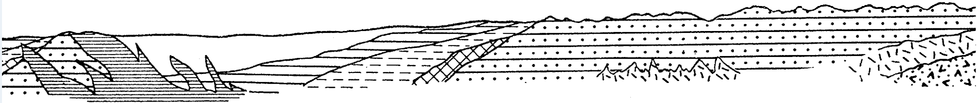
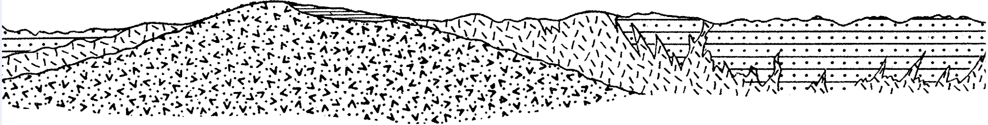
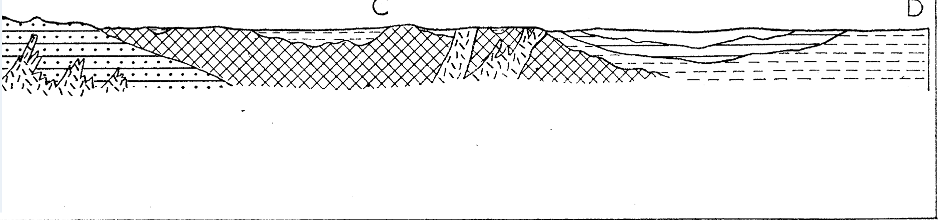




Diagrammatic Section along ABCD

FIGURE 3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For further information concerning the sedimentary rocks of Cyprus the reader is referred to Henson, Brown & McGinty (1949) and to the memoirs and annual reports of the Cyprus Geological Survey.

2·3. *Regional structure*

Excluding the tectonically complicated, and as yet mainly unsurveyed Trypa Group, the overall structural picture of Cyprus is relatively simple. Broadly speaking, the Troodos Igneous Massif has acted as a rigid block against which the rocks of northern Cyprus were compressed by strong southerly alpine earth movements. Henson *et al.* (1949) summarizing the general picture say (p. 4): 'An outer (southerly) fold arc of the Iranides (Arni, loc. cit.) "(Tauro-Dinaric arc)" passes east-and-west through northern Cyprus, forming the spectacular Kyrenia range which is thrust to the south, while the remainder of the island belongs in our opinion, to the foreland'. The position of the Kyrenia range within the Tauro-Dinaric arc is depicted in figure 2.

We envisage the arcuate form of the Kyrenia range and of the concurrent flysch sediments of the Kythrea Formation as due to compaction against the irregular, roughly concave, northern boundary of the Troodos massif. The underground extent of the igneous rocks is directly reflected in the structural character of the overlying sediments. Peripheral to the massif, calcareous sediments mainly of Eocene to Miocene age are undeformed and only slightly inclined, whereas contemporaneous strata in the Kyrenia range have undergone extensive deformation. Both the Pliocene–Recent sediments of the Mesaoria plain and the extensive erosion surface formed during the Upper Miocene (Pontian) marine regression are undeformed, indicating that the latest orogenic movement in Cyprus was during the Lower and Middle Miocene.

3. The geology of the Troodos massif

The Troodos massif, including the inliers of the Akamas Peninsula and Troulli, has an east–west extent of almost 80 miles and from north to south a maximum width of just over 20 miles. The main massif extends from Khrysokhou Bay in the west to Ayia Anna in the east. The inlier of Troodos rocks on the Akamas Peninsula, only recently recognized, is separated from the main massif by a belt of sediments 8 miles wide occupying a north–south, graben-like structure which follows the line of the Khrysokhou valley. Troulli, in the east, is also an inlier of Troodos rocks. Here, a roughly circular outcrop of igneous rocks occupies an area of 16 square miles and is bounded by sediments of the Lapithos Group. There are numerous other minor erosional inliers of the Troodos massif, the largest of which is in the Dhali Area near the village of Petrophani.

The central part of the massif, mainly occupied by the Sheeted Intrusive Complex, is of high relief culminating in Mount Olympus (Chonistra) at 6401 ft. Remnants of an older, Upper Miocene, land surface are present at this highest point but elsewhere a deeply incised, rejuvenated, radial drainage pattern is present. The relief of the peripheral pillow lavas is gentle, planation of this region having taken place during the late Pleistocene.

The threefold subdivision of the massif into the Sheeted Intrusive Complex, the central plutonic complex and the peripheral pillow lavas has long been recognized. Owing to the rather confusing nomenclature used to define and subdivide these units it is proposed in this paper to standardize the terminology. Given below are the original terms used together with those proposed by the writers:

Cullis & Edge (1922)	Ingham (1955)	Wilson (1959)	Bear (1960)	present work
				Serpentine
			post-Upper Pillow Lava Intrusives	post-Upper Pillow Lava Intrusives
Pillow Lava	Upper Pillow Lava, Lower Pillow Lava, Basal Pillow Lava (Pillow Lava Series)	Pillow Lava Series. Upper and Lower divisions	Troodos Pillow Lava Series. Upper and Lower divisions	Troodos Pillow Lava Series. Upper and Lower divisions
Serpentine	Plutonic rocks of Troodos (including serpentine)	Troodos Igneous Complex (including serpentine)	Troodos Plutonic Complex	Troodos Plutonic Complex
Diabase	Diabase	Diabase and Basal Group	Diabase and Basal Group	Troodos Sheeted Intrusive Complex

(1) The Troodos Sheeted Intrusive Complex. This complex consists of a swarm of steeply dipping, altered basic dykes with a dominant north–south trend. These near vertical intrusives form over 90 % of the complex and are only separated by thin screens of basic lavas. In good exposures, the parallelism of the individual units gives rise to a marked sheeted aspect. The host rock in the lower part of the complex is structureless basic lava whilst in the upper portion the host is pillow lava. Initially, this complex was treated as two units, the Diabase with structureless host rock and the overlying Basal Group with pillow lava host rock. As there is no difference in the intrusives, the rocks of the Sheeted Intrusive Complex will be described under two headings: (i) the intrusives; and (ii) the host rocks. As the subdivision of this complex into the Diabase and the Basal Group has been made on survey maps already published this subdivision is retained on figure 3.

(2) The Troodos Plutonic Complex. It is proposed to retain this term originally suggested by Bear (1960). The only alteration made is that the serpentine will be discussed separately as the writers believe it belongs to a younger period of igneous activity. Rocks of the Plutonic Complex range from dunites, through peridotites and olivine-gabbros to granophyres and quartz-albite-porphyries. The more basic members of the complex are found in the central part of the massif and the increasing acidity is roughly proportional to the distance of emplacement from the centre of the massif.

(3) The Troodos Pillow Lava Series. Originally known as the Pillow Lava Series, the prefix 'Troodos' has been added to distinguish between the pillow lavas peripheral to the Troodos massif and those occurring within the Trypa Group. The Troodos Pillow Lavas are thought to lie with marked unconformity on both the Sheeted Intrusive Complex and the Plutonic Complex. Throughout the series, which must be several thousands of feet thick, the extrusives all show well-formed pillow structure. The division into upper and lower divisions, proposed by Wilson (1959) and Carr & Bear (1960), is founded on both field and petrological evidence. Bear (1960) has shown that the lower division is quartz-rich whilst the upper unit is mainly of olivine-bearing rocks. Examination of these rocks, and especially of the chemical analyses, suggests that the whole series is of basaltic composition but that the lower division is less basic than the upper unit.

(4) The post-Upper Pillow Lava Intrusives. Intrusive rocks, petrologically dissimilar to those of the Pillow Lava Series were initially recognized by Bear in the Akaki–Lythrodonda Area. The same group was subsequently identified by one of us (I.G.G.) in the Dhali Area and the Akamas Peninsula.

Table 1

unit		main rock types	remarks
Serpentine (age: post-Lower Triassic)		bastite-serpentine, serpentinized dunite	occurs within the ultrabasic outcrop at Troodos and as large masses in the Akamas Peninsula where field evidence suggests emplacement as a mobile serpentine 'magma'
post-Upper Pillow Lava Intrusives		hypersthene-gabbro, quartz-dolerite	large dykes and small gabbro bosses cutting the pillow lavas but petrologically unlike rocks of that series
Troodos Pillow Lava Series	Upper Pillow Lavas and related intrusives	basalt, olivine-basalt, mugearite, limburgite, picrite-basalt, ultrabasic lava; dolerite, peridotite	mainly pillow lavas, few intrusives. Dominantly basaltic series, olivine common; calcite and analcite common as secondary minerals
	...partial unconformity...		
	Lower Pillow Lavas and related intrusives	andesitic-basalt, quartz-andesitic-basalt, keratophyre; quartz-micro-dolerite, quartz-microgabbro	abundant intrusives as dykes and sills, ratio of intrusive to extrusive rock approx. 1:1. Rocks present are mainly andesitic-basalts. Secondary silica abundant, celadonite common
	...strong unconformity, low-grade metamorphism of older rocks... Emplacement of the Troodos Plutonic Complex		
Troodos Plutonic Complex		dunite, peridotite, pyroxenite, olivine-gabbro, gabbro, granophyre and quartz-albite-porphyry	differentiates of the Troodos Plutonic Complex magma
Sheeted Intrusive Complex		*Intrusives.* Albite, quartz and epidote-diabase, saussauritized dolerite *Host rock.* Pillow lava mainly keratophyre and epidosite. Structureless basic lava altered to epidosite	near vertical, sheeted aspect to the intrusives; structureless lavas are host material in the lower part, pillow lavas in the upper part of the complex. Ratio of intrusives to extrusives 10:1. Affected by low-grade metamorphism. Albite characteristic, pyroxene mainly altered to actinolite and chlorite. Quartz, epidote and albite common as secondary minerals

Feeder dykes for the Pillow Lava Series, identified on their petrological characteristics, occur within the Sheeted Intrusive Complex and probably form as much as 40% of that complex. Similarly, basic and often olivine-bearing intrusive rocks that are related to the Upper Pillow Lavas are found within the Lower Pillow Lava outcrop.

(5) Serpentine. Recent work by Gass (in Ingham 1959) in the Akamas Peninsula shows that serpentine has been emplaced into all members of the Troodos igneous suite as well as into the overlying sediments of the Trypa Group. This activity can therefore be dated as post-Lower Triassic in age whilst, in all probability, all the preceding igneous activity is pre-Triassic.

The main rock types present in the Troodos massif are shown in table 1.

3·1. *The Troodos Sheeted Intrusive Complex*

Rocks of the Troodos Sheeted Intrusive Complex are well exposed in the deeply incised river courses that drain the massif. Here, it can be seen that they form a huge series of near vertical, basic dykes. The dykes are so closely packed that they present, in good exposures, a striking sheeted structure. The original host rock, which is of similar basic composition to the dykes, has been so disseminated by the intrusives that it now forms less than 10 % of the complex.

Forming the greater part of the massif, this complex extends from Limni in the west to Psevdhas in the east with a narrow, north–south strip occurring within the Akamas inlier. Although the main trend of the intrusives is north–south, there are limited areas where the strike diverges markedly from this direction. Dips are usually greater than 60°; see figure 3.

Dykes of the complex belong to at least three phases of igneous activity: (i) those genetically related to the complex; (ii) feeder dykes for the Pillow Lava Series and (iii) post-Upper Pillow Lava Intrusives. Although the feeders for the Pillow Lava Series and the post-Upper Pillow Lava Intrusives maintain a regular attitude whilst within the outcrop of the Sheeted Complex, differences between the three types can be determined on microscopic examination.

Two types of host rock are present, structureless basic lava and basic pillow lavas. The structureless lava is the host material in the lower levels of the complex whilst the pillow lavas are the host in the upper portion. Owing to the subsequent differential uplift, the lower portions of the complex are now exposed in the centre of the massif whereas the upper part occupies the periphery of the complex.

(1) **Alteration of the rocks of the Sheeted Intrusive Complex.** All rocks of this complex, both intrusive and extrusive, have been altered to a greater or lesser extent by saussuritization and attendant uralitization. This alteration has produced a series of rocks that range, with increasing alteration, from keratophyres to epidosites. The processes envisaged in the formation of this suite of rocks are shown diagrammatically below:

BASIC ROCK
calcic plagioclase (andesine–labradorite) —— (saussuritization) → albite, calcite, epidote, clinozoisite, quartz
diopside —— (uralitization) → actinolite, chlorite

This alteration is probably associated with the emplacement of the Troodos Plutonic Complex as it does not affect the feeder dykes of the Pillow Lava Series and the post-Upper Pillow Lava Intrusives within the outcrop of the Sheeted Intrusive Complex.

Detailed field work mainly by Wilson (1959), Bear (1960) has shown that alteration of both the intrusives and extrusive host rock of the complex increases towards the centre of the massif. Bearing in mind the domed structure of the massif this process is regarded as increasing alteration related to depth.

(2) **The intrusives.** The vertical sheeted aspect of the multiple dykes is generally prominent and individual intrusives vary in width from 1 to 15 ft. The sheeting is emphasized

by the joint pattern in which the major joints are concurrent with, and the subsidiary joints normal to, the walls of the intrusives. Chilled margins are common, although in rare cases they are noticeably absent. The absence of chill phenomena has been explained by Bear (1960) who suggests that the temperature of the host rock was sufficiently raised by previous intrusions to ensure the uniform cooling throughout any subsequent intrusive mass.

Both Wilson (1959) and Bear (1960) have subdivided the intrusives on the proportion of the minerals present; albite, quartz and epidote-diabase are the main types. More altered rocks have been termed epidosites.

(3) **The host rock.** As mentioned earlier, the host rock of the Troodos Sheeted Intrusive Complex can be divided into two; Bear (1960) recognized an upper pillow lava unit and a lower zone of structureless lavas.

The structureless host rock consists of iron-stained weathered rocks that are usually extensively altered by mineralization and subsequent oxidation. They occur as narrow sheets varying in width from 1 to 15 ft. and can be distinguished from the intrusives in good exposures where the latter is usually a hard, tough, dense rock of grey colour and medium grain size. These rocks would be best termed epidosites, they are thought to have been originally lavas of basaltic composition.

The pillow lava host rocks range from keratophyres to epidosites. The notable characteristics of all types are the universal presence of the soda plagioclase albite, the partial or complete alteration of the original pyroxene to actinolite or chlorite, the presence of epidote, zoisite, chlorite, calcite and quartz, and the presence of sphene as an accessory.

In many cases these pillow lava screens have been brecciated either by fault movement or by the emplacement of the adjacent intrusives. Flow breccias of the same composition are common.

(4) **Evolution and origin.** One of the main problems in the geology of Cyprus is the structure of the Sheeted Intrusive Complex. Although now thought to be an intrusive complex, Bishopp (1952*b*) using the term 'Folded Diabase' in the same sense that the writers use 'The Sheeted Intrusive Complex' states:

'The Diabase is everywhere sheeted or banded, the bands varying from a few centimetres to a metre or so in thickness. It is thus seen to be regionally and intensely folded along axes which do not deviate much from the north–south direction...There are at least two major anticlines and synclines; dips are generally of the order of 60 degrees, and much of the folding seems to be isoclinal. This fine-textured rock is believed to represent a series of basic lava flows which have suffered orogenic movement, have undergone low-grade metamorphism to epidiorite and have been intruded *lit-par-lit* by numbers of basic dykes.'

Recently some of Bishopp's critical exposures were examined in detail. Although in all cases a reversal of dip was noted, examination of the axial area proved the existence of a fault or fault zone along these lines. In these cases the structures, viewed broadly, simulate anticlinal or synclinal disposition but the variation in all these cases can be explained by hinge faulting along the strike of the intrusives. Bear (1960, pp. 85–88) discusses this problem in more detail.

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Factors indicating an intrusive origin for the Sheeted Complex are:

(1) The horizontal, or near horizontal, disposition of the host rock.

(2) The abundance of chilled margins on units marginal to host rock and also on adjacent sheets. In this context the units showing chilled margins far out-number those without this phenomenon.

(3) The scarcity of vesicles which would have been relatively common in lavas. In addition, such vesicles as are present are elongated parallel to the parent body and not at right angles to the sheeting as would be expected in a series of folded lava flows.

(4) There is noticeable absence of volcanic features such as slaggy tops, flow structures and lateritic boles.

(5) Characteristic features of fold tectonics such as cleavage and schistosity which would be expected in a series of highly folded competent lava flows are lacking. Away from fracture zones the rocks of the complex are massive, showing rectangular jointing, the main joint pattern being parallel to and the secondary jointing normal to the sheeting.

Individually, these factors could not be taken as conclusive evidence of an intrusive origin, but together it is thought that the case of the intrusive hypothesis is well substantiated.

The Sheeted Intrusive Complex is therefore a north–south dyke swarm of unusual density in which the intrusives display a striking parallelism. Although repeated intrusive activity could be solely responsible for the unusual dyke density there is evidence that the massif has suffered erosion during and since its evolution. It is suggested therefore that this intense erosion has exposed the lower levels of the Troodos volcanic area where intrusive material might be expected to predominate over extrusive rocks.

Regarding the north–south alinement of the dyke swarm, it has been suggested by Bear (1960) and Gass (1960) that the controlling feature might have been the structure of an underlying basement. The writers would suggest that the main factor in the alinement of this complex was an east–west tensional stress that was operative throughout the evolution of the massif and that, since the area was probably oceanic, no sialic basement was present under Troodos at that time.

It is interesting to note that the only area, known to the writers, where there is a dyke swarm of similar type and density is within a rift zone on Oahu, Hawaii (Stearns & Vaksvik, 1935).

3·2. *The Troodos Plutonic Complex*

The Troodos Plutonic Complex is composed of a suite of rocks ranging in composition from dunite and peridotite through pyroxenite and olivine-gabbro to gabbro, granophyre and quartz-albite-porphyry; all are thought to be differentiates of one parent magma. Members of this suite were emplaced into the Sheeted Intrusive Complex and so far none has been found intruded into rocks of the Pillow Lava Series. It is thought, with some confidence, that the emplacement of this complex took place after the formation of the Sheeted Intrusive Complex and prior to the extrusion of the Troodos Pillow Lava Series.

No floor to the plutonic complex has been identified and there are no compositional changes towards the margins. Gabbro dykes and veins are relatively common within the

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peridotite outcrop and inclusions of peridotite are abundant in the gabbros. Textures throughout the complex are dominantly allotriomorphic although many of the peridotites display poikolitic associations of olivine in pyroxene. Gneissose textures are absent in the main exposure at Troodos. Chromite deposits on Troodos, associated with dunite host rocks, occur as a series of steeply dipping pod-like lenses which are extremely irregular in shape and size.

The largest outcrop of these rocks occupies a roughly oval area of about 200 square miles in the highest part of the mountains. Details of this area are taken from Wilson (1959). A further large area of basic and ultrabasic rocks belonging to this complex is known to crop out in the Limassol forest about 15 miles south-east of the Troodos occurrence. This area, which is being examined by L. M. Bear at present, is extremely complex and has a marked east–west trend. Throughout the rest of the massif small isolated bosses of gabbro, granophyre and porphyry belonging to this suite have been identified. A broad regional pattern can be discerned in that the rocks become more acidic proportionally with increasing distance from the Troodos ultrabasic outcrop.

(1) **Ultrabasic rocks.** Well exposed on the summit of Troodos the ultrabasic rocks of the Troodos Plutonic Complex are amongst the freshest of the massif. Wilson (1959) divides these rocks into four main groups; the dunite, the enstatite-olivinite, the harzburgite–wehrlite group and the peridotite–pyroxenite group. The main features of these rocks, taken from Wilson, are listed below together with more recent information.

Dunite. The main occurrence of the dunite is near the summit of the Troodos range where it forms a semi-circular outcrop on the north, west and southern flanks of Mount Olympus. Never more than a mile wide, the outcrop, although discontinuous, extends for a distance of over 12 miles. The olivine in the Troodos dunite is rarely completely fresh, between 15 and 75 % being altered to antigorite. The olivine is optically positive and has an average composition of Fa_8. The grain size is markedly uniform although the original crystal form has been modified and rounded by the partial alteration to serpentine. The only other constituents are isolated crystals of chromite and magnetite which never form more than 5 % of the rock.

Enstatite-olivinite. This rock and its serpentinized equivalent bastite-serpentine are the most common ultrabasic types in the Troodos Area where they occupy the highest ground around Mount Olympus. In the west of this area enstatite-olivinite is the prevalent rock type whilst bastite-serpentine occurs in a smash zone in the east of the region. There is no marked contact between the two types; alteration gradually increases from west to east.

Field evidence indicates that part at least of the serpentinization took place after emplacement as the most intense alteration is confined to the smash zone mentioned above and to areas adjacent to fracture planes. Bastite-serpentine also occupies large areas in the Limassol forest.

Under the microscope it can be seen that the enstatite-olivinite has an allotriomorphic texture and contains between 70 and 85 % olivine or serpentine, the remainder consisting of enstatite or bastite. The degree of serpentinization varies considerably; in fresh specimens only 15 % of the olivine is altered and the outlines of the original crystals are still identifiable. In the bastite-serpentine the alteration is complete although relics of the original rock fabric are occasionally still identifiable.

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Harzburgite–wehrlite group. On Mount Olympus this group crops out around the western and southern margins of the ultrabasic outcrop and forms a transitional zone between the dunite and the surrounding gabbro. The group consists mainly of two rock types; harzburgite occupying the more central area and wehrlite which is more common near the gabbro. Wilson (1959), has not attempted to subdivide the group in the field as there are no distinctive macroscopic differences. Further, he suggests that the group is transitional although no lherzolites, containing both orthorhombic and monoclinic pyroxene, have been identified.

Between 60 and 70 % of the harzburgite is formed of olivine, up to 50 % of which may be altered to serpentine. With the exception of the alteration products the rest of the rock is formed of enstatite around which tremolite is commonly developed. The olivine occurs as poikolitic inclusions within the enstatite host and in most cases the inclusions are more abundant than the host material. In the wehrlites the same proportion of olivine is present but in this case the host material is mainly diopside and the olivine is more altered to serpentine. Enstatite is often present in the wehrlite as a minor constituent.

Peridotite–pyroxenite group. Rocks of this group generally form small isolated outcrops within the harzburgite–wehrlite group. They are medium to coarse grained, the crystal size ranging from 5 to 10 mm. Within this group the rocks range from diopside-peridotite and enstatite-diopside-peridotite (bielenite) to enstatite-diopside-pyroxenite (marchite). Diopside-peridotite is the main rock type; others present could be suitably described as local facies. Monomineralic varieties are present though very localized in their occurrence. No subdivision of the group has been made in the field as macroscopic differences are not readily identifiable. In the diopside-peridotite the texture is allotriomorphic with clinopyroxene forming between 60 and 80 % of the rock. Enstatite accounts for 5 % of the rock; the remainder, apart from iron oxides and alteration products, is forsteritic olivine.

Mineral banding. Developed locally within many of the ultrabasic rocks and the adjacent olivine and pyroxene-gabbros is a vertical or near-vertical mineral banding with a north–south orientation. On examination, this banding can be seen to be a vein-like segregation of olivine and pyroxene. Individual bands vary in width from half an inch to a foot and owing to the lack of continuous outcrops are not traceable over distances greater than 10 ft. The banding is prominent near the contacts of the various rock types especially where the enstatite-olivinite and the pyroxenite-peridotite group are in contact with the dunite. In the femic-rich bands the olivines or pyroxenes are orientated with their long axes in the direction of the banding. It was noted by Wilson (1959) that this north–south banding was in alinement with the chromite veins in the dunite and also with the dunite inclusions in the serpentine.

The writers are in agreement with Wilson (1959) that this banding is due to a process that was subsequent to the segregation of the major rock types by magmatic differentiation.

Wilson (1959) citing the work of Bowen & Tuttle (1949) suggests that this banding is due to the passage along fissures, of water vapour under-saturated in SiO_2, which changed the pyroxene into olivine. Owing to the large areas over which this vertical banding is displayed it is thought more probable that it is due to the admixture of quasi-solid magmas during a period of reactivation; large-scale flowage under stress bringing about the intimate vertical association of rocks of dunitic, peridotitic and gabbroic composition.

(2) **The gabbros.** The main gabbro outcrop surrounds the ultrabasic rocks of Troodos and extends eastwards, flanking the watershed, as far as the village of Palekhorio. These rocks are also a common type in the plutonic complex of the Limassol forest. Elsewhere in the massif small dykes and bosses of gabbro have been emplaced into the Sheeted Intrusive Complex.

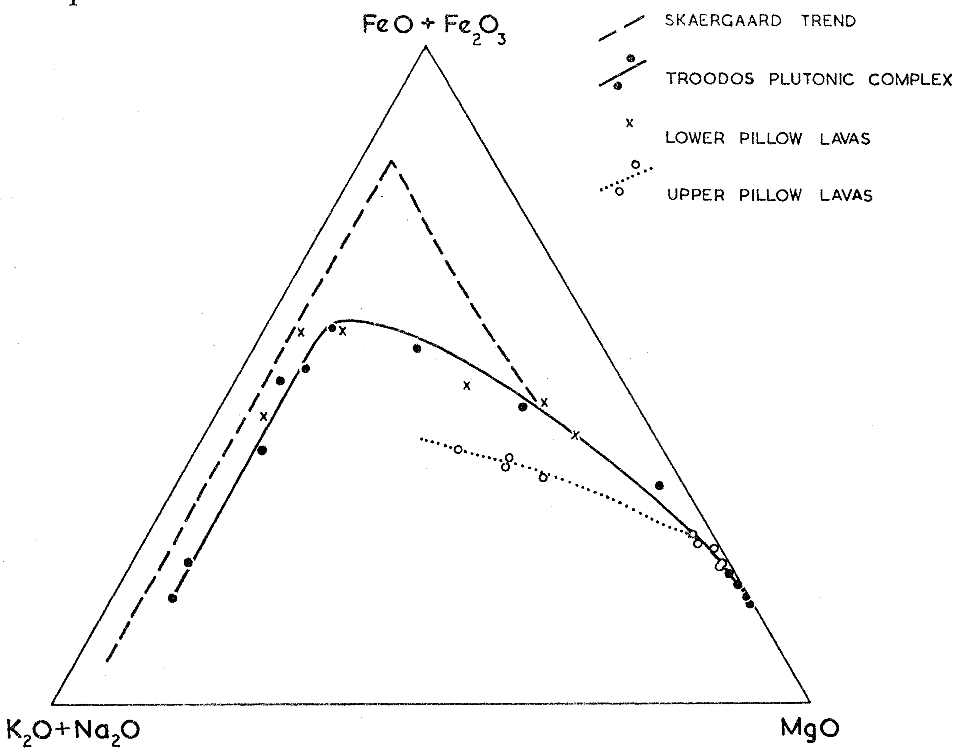

FIGURE 4

For descriptive purposes the gabbros have been divided into melagabbro, olivine-gabbro, gabbro and uralite-gabbro. No clear boundaries can be drawn between these types in the main outcrop. Wilson (1959) suggests that there is 'a gradual transition outwards and upwards from the ultrabasic rocks through the more basic gabbros to the uralite-gabbro'. Further variations in the gabbro and uralite-gabbros have been noted; these are due mainly to a variation in the proportion of minerals present and to the degree of alteration that has taken place. Small masses of microgabbro belonging to this complex have been emplaced into the more marginal areas of the Sheeted Intrusive Complex.

(3) **The granophyre and quartz-albite-porphyries.** Granophyre is a field term used for light coloured, quartzo-feldspathic rocks which are acid differentiates of the Troodos Plutonic Complex and are found as concurrent dykes and irregular bosses near the margins of the Sheeted Intrusive Complex.

Composition. For comparative purposes, the differentiation trends of the Troodos Plutonic Complex is plotted on a *FMA* (iron, magnesia, alkali) diagram together with the curve for the Pillow Lava Series and that given by Wager & Deer (1939) for the Skaergaard Complex (figure 4).

(4) **Evolution and origin.** The distribution of the outcrops of this complex indicates that these plutonic rocks belong to a differentiated ultrabasic mass of batholithic dimensions. Differentiation has led to the upward and outward change from dunites and peridotite through melagabbro, olivine-gabbro and gabbro to granophyre and quartz-albite-porphyry; the ultrabasic rocks, which form the core of the outcrop, are surrounded by roughly annular outcrops of increasingly acid members of the complex. Where exposed, the boundary between these rock types is gradational. The central dunite has probably been remobilized for it interfingers extensively with the adjacent rocks.

Field, mineralogical and geophysical evidence, discussed below, all suggest that the parent magma of the Troodos Plutonic Complex was ultrabasic in character and derived from the earth's peridotite mantle.

Wilson (1959) makes it clear that field evidence in the Xeros–Troodos Area suggests that the proportion of dunite and peridotite to the more acid differentiates increases markedly with depth. It is suggested that the gabbros and more acid differentiates, although abundant at the surface, are, in fact, minor differentiates of a vast mass of ultrabasic material underlying Troodos.

Mineralogically, the main indication suggesting an ultrabasic rather than a basaltic parent is given by the composition of the olivine. It has been proposed (Vogt, 1921) that the composition of this mineral is indicative of the origin of the parent material. In this connexion, Ross, Foster & Myers (1954) have demonstrated that olivines from dunites and olivine-rich inclusions in basalts range in composition from Fa_8 to Fa_{11} (calculated from analyses). These figures are confirmed by Drever & Johnston (1958) who obtained values ranging from Fa_7 to Fa_{12} for the olivines of the minor picrite inclusions of the Hebrides which, they suggest, were derived by selective fusion from pre-existing ultrabasic rock.

On the other hand, Muir & Tilley (1957) have shown that the olivines from the picrite basalts of Kilauea, which are demonstrably differentiates from a basaltic melt, range from Fa_{14} to Fa_{20}. MacDonald (1949) obtained similar values for olivine differentiating from a basaltic melt. These works would suggest that olivines of composition Fa_7 to Fa_{12} are derived from an ultrabasic source, whilst olivines richer in the fayalite molecule originate from a melt of basaltic composition.

Unfortunately, there are comparatively few estimations on the composition of the olivines of the Troodos Plutonic Complex. It has, however, been determined that the olivine in the dunite has a composition of Fa_7 to Fa_8. Similar values were obtained from the peridotites whilst the olivines in the gabbro range from Fa_8 to Fa_{12}. This information is taken from the examination of isolated specimens and can serve as no more than a general indication, but the limited evidence available suggests that the olivines of this complex originated from an ultrabasic source.

The statement that olivines within the range Fa_7 to Fa_{12} were derived from an ultrabasic source needs some qualification for few authors believe that this necessitates primary crystallization from a melt of peridotitic composition. Drever & Johnston (1958) propose that the highly forsteritic olivines of the picrite intrusions of the Hebrides are relict crystals derived from the partial refusion of a pre-existing ultrabasic mass. Similarly, Ross *et al.* (1954) would support the derivation of highly magnesian olivine from the earth's peridotite shell as already crystalline material. Evidence will be put forward later to support the theory that the Troodos Plutonic Complex magma was a crystal mush with olivine carried in an interstitial fluid fraction of basaltic composition.

Turning now to the geophysical evidence, it has already been indicated, and will be shown later, that over Cyprus there is a large positive gravity anomaly caused by the presence under the area of a large mass of high-density material. This mass of high-density rocks belongs in all probability to the Troodos Plutonic Complex, is confined to the area of Cyprus and has a slab-like space-form at least 7 miles thick. In view of the vast volume of high-density rocks of peridotitic affinity present it is suggested that the source of this material is the earth's mantle.

In composition and in the nature of the main rock types present, the Troodos Plutonic Complex is closely comparable to Thayer's (1960) 'Alpine' type complexes. This comparison, it is suggested, indicates a similar parent magma but not necessarily a similar tectonic setting. Many workers, Hess (1938, 1955), Noble & Taylor (1960) and others, have proposed that ultrabasic complexes within orogenic zones derive their parent magma from the upper mantle and that stresses present in these zones enable this ultrabasic material to be injected into the earth's crust. There is no evidence that orogenic forces were present during the evolution of the Troodos massif but all evidence points to the upper mantle as the source of the Troodos Plutonic Complex magma.

3·3. *The Troodos Pillow Lava Series*

Rocks of the Troodos Pillow Lava Series almost entirely surround the Sheeted Intrusive Complex. It can be seen (figure 3) that the width of the pillow lava outcrop varies; on the north-west coast, near Pomos Point, rocks of the Sheeted Intrusive Complex come down to the sea and in places on the southern flank of Troodos, chalks of the Lapithos Group rest directly on the sheeted intrusives of this complex. The widest outcrop of the Pillow Lava Series is in the east of the massif, in the Dhali and Pano Lefkara–Larnaca Areas, where it is between 5 and 8 miles wide.

Tuffs and interbedded umberiferous sediments have been located within this mainly volcanic series. Unfortunately, in no case have any recognizable organic remains been identified so definite evidence for the age of this igneous episode is, as yet, unavailable. Henson *et al.* (1949) postulate a Cretaceous age for the igneous activity, mainly on comparison with other ophiolitic areas in the Middle East. Dubertret (1953) originally agreed with this age and compared the Troodos igneous area to similar rocks in Syria and Southern Turkey. Subsequently, Dubertret, after a brief visit to Cyprus, agreed that the Syrian volcanics were equivalent to those of the Trypa Group and that no volcanic activity comparable to the Troodos massif was to be found on the mainland. Wilson (1959) suggests that the volcanic episode is much older and quotes as evidence the Jurassic radiolarites in the Perapedhi Formation and the fact that Triassic rocks of the Mamonia Complex seem to overlie the Troodos volcanics. Further reconnaissance work (Gass, in Ingham *et al.* 1959) in the Akamas Peninsula supports the view that the Troodos rocks are older than the Triassic Mamonia Complex.

The contact of the Sheeted Intrusive Complex with the overlying Pillow Lava Series is faulted in many places, but in others it is, most probably, an unconformity marking a period of erosion. A definite change in topography indicates the approximate boundary between the sheeted intrusives and the peripheral pillow lavas. Whilst the pillow lavas have a low relief, the harder indurated rocks of the Sheeted Complex form higher, more rugged ground.

Wilson (1959) and Carr & Bear (1960) initially divided the Pillow Lava Series into an upper and lower division on the presence of an unconformity and petrological differences in the rock types of the two units. Although the petrological differences apply throughout the whole outcrop the unconformity is only partial; elsewhere, there is a gradual transition from the Lower Pillow Lavas into the extrusives of the upper unit.

There are too many unknown factors to enable the thickness of the pillow lavas to be estimated with any degree of accuracy. Generally, about 200 ft. of Upper Pillow Lavas are present. At the eastern and western extremities of the massif the succession thickens; about 600 ft. of lavas are exposed in the Dhali Area whilst there is an estimated thickness of about 1000 ft. in the Akamas Peninsula. The lower division, which generally has a much wider outcrop, is probably between 2000 and 3000 ft. thick. It must be emphasized that these figures are no more than a tentative estimate.

Bear (In Ingham *et al.* 1958) lists the most striking petrological differences between the two divisions, these are tabulated in table 2.

TABLE 2

	Upper Pillow Lavas	Lower Pillow Lavas
quartz	usually absent	quartz, opal and chalcedony are abundant
plagioclase	labradorite; mainly skeletal crystals	labradorite–andesine range. More sodic varieties occur
pyroxene	diopside–augite, fresh and unaltered	diopside: marginally altered to celadonite
olivine	commonly present, often replaced by calcite pseudomorphs	absent
celadonite	absent	common
calcite	common in ground mass	rarely present
zeolites	mainly analcite	mainly natrolite and thomsonite, analcite rare
amygdales	mainly calcite, some analcite	quartz, opal, chalcedony, calcite, celadonite, natrolite and thomsonite

(1) **Lower Pillow Lavas.** Extrusive and intrusive rocks are present in roughly equal proportions in this division of the series, extrusive rocks consisting mainly of pillow lavas with very subordinate flow and vent agglomerates. Intrusive rocks include dykes, sills and irregular intrusive masses.

Intrusives that are genetically related to the Lower Pillow Lavas form between 40 and 50 % of the rocks of this unit. Many of the dykes reached the surface and are therefore feeders for the pillow lavas. Others can be seen to fade out vertically and still others transform into sills or into small irregular masses before reaching the surface. Dykes vary in width from 6 in. to 15 ft.; although usually simple, multiple dykes are by no means rare. The outcrop of an individual intrusive can only be traced for a short distance, but plotting of units brings out a marked regional trend.

The rocks of this division are grey to grey-green in colour and are mainly of basaltic composition. Difficulties have been encountered in defining these rocks as the majority have a silica content of about 52 %, are too low in alumina, and too high in magnesia to fall within the true andesite range. Mineralogically, the feldspars are about An_{50} in composition, the phenocrysts usually being labradorite but the ground mass microlites, estimated as An_{45} to An_{50}, were difficult to determine due to alteration and the minute size of the crystals. Diopside is the main femic mineral and usually forms less than 40 % of the rock.

Finally, there are those rocks which, although outcropping within this division, are petrologically similar to the Upper Pillow Lavas and are thought to be feeders for that unit. These rocks are usually olivine-bearing and occur both as dykes and small irregular bosses.

(2) **The Upper Pillow Lavas** occupy a narrow discontinuous belt of country peripheral to the Troodos massif. The widest outcrop is found in the Akamas Peninsula where about 3 miles of Upper Pillow Lavas are exposed between the Sheeted Intrusive Complex and the west coast of the island.

In many areas, such as the Akamas Peninsula, olivine-basalts are the dominant rock type, forming over 90 % of the rock outcrop. In other regions olivine-free basalts are common. In the Dhali Area basalts and olivine-basalts are present in roughly equal proportions whilst in the Akaki–Lythrodonda Area, Bear (1960) is of the opinion that mugearites form a considerable proportion of the rocks present. More basic lavas, ranging in composition from ankaramites through limburgites and picrite-basalts to lavas of peridotitic composition, have been found in the Akaki–Lythrodonda and Dhali Areas where they form less than 5 % of the succession but are easily recognizable owing to their distinctive appearance. Elsewhere, only isolated occurrences of these more basic rocks have been found.

Throughout this division the lavas have a pillowed structure, individual flows are difficult to isolate except in river sections where the contact is usually marked by a thin horizon of pink calcite. The pillows vary in size from large mattress-shaped masses of up to 25 ft. in length and 10 ft. in height to small spherical structures less than 1 ft. in diameter. Chilled margins to the pillows are prominent in places, the black glassy selvedge standing out against the pale grey altered lava. Flow breccias, in which relics of pillow structure are visible, are quite common. In these occurrences the original pillowed structure has been almost obliterated by the later brecciation.

Thin current-bedded tuff horizons occur near the top of the pillow lava succession in the Dhali Area. This occurrence of a pyroclastic horizon indicates that at the time some volcanic vents were above sea level, whilst the current-bedding suggests a shallow water environment.

Intrusive rocks are not common in the Upper Pillow Lavas. The dykes tend to be larger and more continuous than their counterparts in the lower unit but the trend of the dykes in the upper division coincides with that of the intrusives in the lower unit. The dyke density varies considerably; for the main part intrusives form less than 5 % of the outcrop, but in certain areas, especially where the contact between the two divisions of the series is transitional, they can form up to 15 % of the rocks present.

(3) **Evolution and origin**. A long period of erosion separates the extrusion of the Troodos Pillow Lava Series, which is dated with some confidence as pre-Triassic, and the

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overlying sediments which are mainly of Upper Cretaceous or Lower Tertiary age. It seems probable that the lavas of this series originally overlay much of the Sheeted Intrusive Complex, for feeder dykes, identified on their petrology, occur well within the present outcrop of the complex. If this were the case, then the subsequent uplift of the area and the resultant erosion must have removed much of the series that covered the more central areas of the massif.

The relative rarity of vent and flow breccias indicates that, for the main part, the evolution of the Pillow Lava Series was characterized by the gentle effusion of basaltic lava in a subaqueous environment. Treating the Pillow Lava Series as a whole and regarding the partial unconformity between the two units as a relatively localized phenomenon, it would appear that the series becomes progressively more basic with decreasing age. The general relationship is well shown in figure 5, a silica/oxide variation diagram of the Pillow Lava Series on which the time relationship of the various rock types has been superimposed. It must be stressed that this is a generalized picture and that it does not necessarily hold true for isolated exposures.

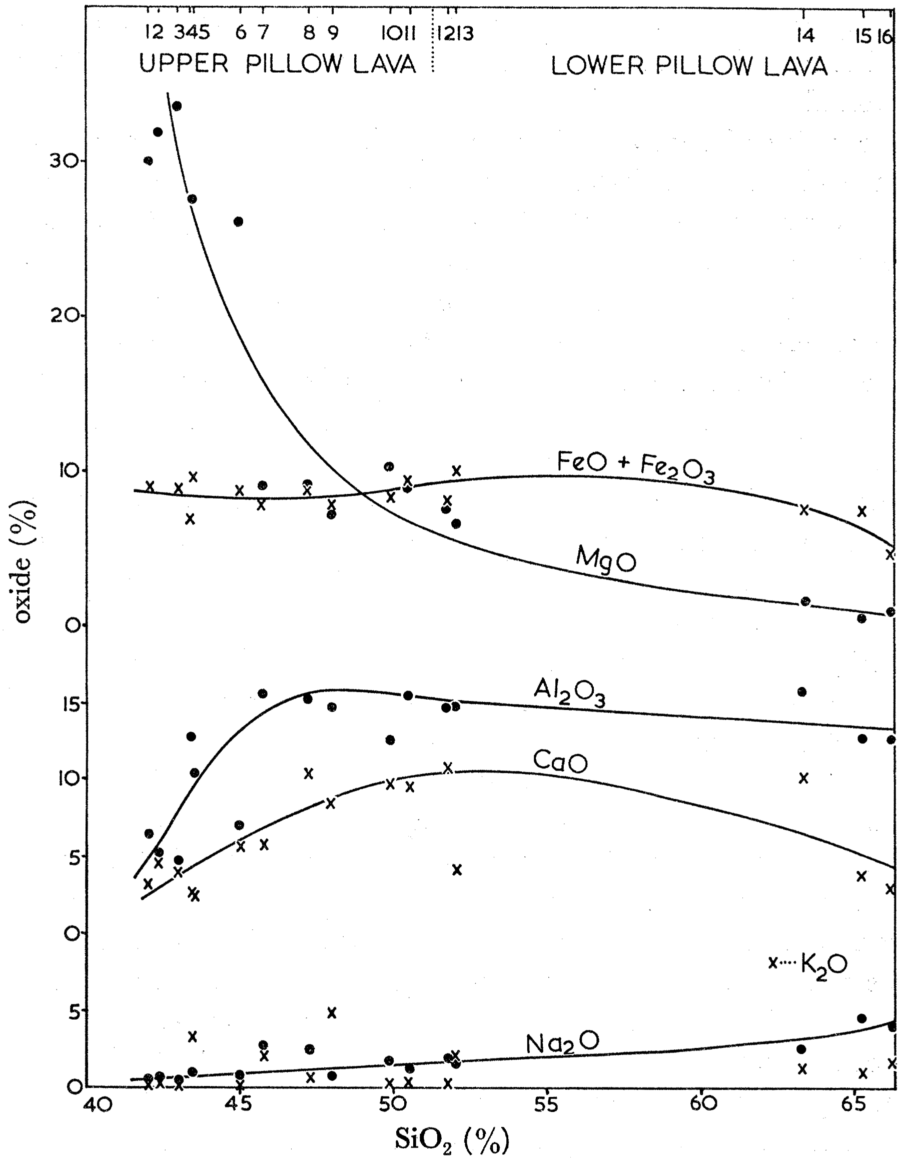
The olivines of the Pillow Lava Series are all highly magnesian and analyses of this mineral from the ultrabasic pillow lavas show a composition of $Fa_{8\cdot0}$, $Fa_{8\cdot1}$ and $Fa_{10\cdot0}$ (Gass 1958). Unfortunately, almost invariably, the olivines in the olivine-basalts and the picrite-basalts, have been replaced by pseudomorphs of calcite. It is also relevant that the series as a whole is highly magnesian and also relatively rich in CaO.

Although the evidence available is not as comprehensive as would be desired, it does point to the conclusion that the rocks of the Pillow Lava Series were derived from an ultrabasic source, similar in all respects, to the parent of the Troodos Plutonic Complex. Accepting this conclusion, difficulties arise in explaining the extrusion of material derived from the peridotite mantle. It will be demonstrated that under Troodos a layer of mantle material was probably near the surface. The writers would therefore suggest that melting within the mantle provided the magma for the Pillow Lava Series. Envisaging the mantle at this level as composed of olivine crystals in a solid matrix of basaltic composition, it is thought that as melting took place the first fluid material available for extrusion would be of basaltic composition. Then, as fusion became more complete the liquid fraction would become increasingly basic. This would possibly explain why the first lavas extruded, those of the lower division, are of relatively acid composition whilst the later extrusives become progressively more basic.

Although envisaging a long period of igneous inactivity between the emplacement of the Plutonic Complex and the extrusion of the Pillow Lava Series it is recognized that the two rock suites are very similar; this is demonstrated in the *FMA* diagram (figure 4). The similarity in composition of the rock types of the Troodos Plutonic Complex and those of the Pillow Lava Series and the fact that the Troodos igneous field is limited to Cyprus and the adjacent areas strongly suggests that the Pillow Lava Series, although younger, was derived from the same source as the Troodos Plutonic Complex.

3·4. *Post-Upper Pillow Lava Intrusives*

Little can be said regarding these intrusives other than that they occur as dykes and small bosses cutting the Upper Pillow Lavas and are petrologically unlike the rocks of that series. Often these rocks contain hypersthene, a mineral rarely found in the Pillow Lava Series, and are of noritic composition.



1. Peridotite boss	5. Limburgite	9. Mugearite	13. Glassy andesitic basalt
2. Ultrabasic pillow lava	6. Picrite-basalt	10. Limburgite	14. Andesitic basalt
3. Ultrabasic pillow lava	7. Basalt	11. Basalt	15. Dacite glass
4. Olivine-basalt	8. Dolerite dyke	12. Augitite	16. Dacite glass

FIGURE 5. Analyses Recorded in the *Annual Reports of the Cyprus Geological Survey* for the years 1955–60.

Although petrologically unlike rocks of the Pillow Lava Series the post-Upper Pillow Lava Intrusives are compositionally similar to the extrusives of the Upper Pillow Lavas. It would seem probable therefore that although they belong to a later phase in the evolution of the Troodos massif they were derived from a similar source to the Upper Pillow Lavas. Lack of sufficient evidence, especially in the form of chemical analyses, precluded more detailed discussion of the evolution of these rocks.

3·5. *The Serpentine*

Serpentine occurs in three main areas; the summit of Troodos, the Limassol forest and in the Akamas Peninsula. Other serpentine masses have been found within the Trypa Group and these will be considered with the Akamas occurrences as they are genetically similar. At Troodos, to the east of Troodos village, part of the enstatite-olivinite outcrop, which forms a roughly circular mass 3 miles in diameter, has been intensely sheared and brecciated and the original rock replaced by bastite-serpentine. This is, most probably, due to the hydration, *in situ*, of the original enstatite-olivinite. The only description of the Limassol forest serpentine so far available is by Bear (in Ingham, 1959) who states: 'there are two serpentines...one is derived from the serpentinization of ultrabasic rocks *in situ*, and a later intrusive mass perhaps correlatable with the Mamonia serpentine'. In the Akamas Peninsula, three major serpentine masses have been located. Here, the rock is mainly bastite-serpentine and it has been emplaced into all members of the Troodos igneous suite as well as into the overlying Trypa Group rocks. It can be simply inferred from the field evidence that the serpentines in this area, which are the youngest phase of igneous activity, are of post-Lower Triassic age.

Wilson (1959) suggested that the Troodos serpentines were formed from ultrabasic rocks, the alteration being brought about by low-temperature water vapours which acted upon the olivine and enstatite to form bastite and serpentine. Further, this author suggested that the serpentinization was confined to the smash zone and that the passage of water vapour in this zone was facilitated by the smashed nature of the ground.

The Akamas serpentines and those occurring within the Trypa Group differ from the Troodos mass in that they do not display the same intense shearing and brecciation. In the Akamas Peninsula, the serpentine can be seen to be emplaced into the Sheeted Intrusive Complex; the contact, although sharp, is brecciated in places and dips to the east at 40°. Isolated pods of the host rock are found within the serpentine near the contact, and offshoots of serpentine penetrate into the Sheeted Intrusive Complex for about 250 yd. The eastern margin of this mass is well exposed below the Baths of Aphrodite where a complex melange of schists, radiolarites, volcanic breccias and crystalline limestones of the Trypa Group seem to have been caught up as xenolithic masses along the margin of the serpentine. No thermal metamorphism is evident on any of the contacts.

The field relations of the Akamas serpentines suggest that they were emplaced as a plastic, low-temperature, serpentine 'magma' and that little or no alteration took place after emplacement.

It was proposed earlier that the peridotite mantle was probably the source of the Troodos Plutonic Complex and the Pillow Lava Series. It is therefore not difficult to envisage the serpentines as being derived as crystal mushes from the same ultrabasic source.

4. Gravity anomalies in Cyprus

4·1. *Earlier surveys*

The first gravity measurements in Cyprus were made by Mace in 1939. Mace established thirteen stations in the island using the Cambridge pendulums. These showed strong positive Bouguer anomalies everywhere, reaching a maximum of 200 mgal at Kalokhorio.

A gravimeter survey of the eastern half of the island was carried out for the Iraq Petroleum Company (I.P.C.) in 1946 and covered much of the area of sedimentary rocks south of the Kyrenia range. In 1958 Overseas Geological Surveys (O.G.S.) extended the gravimeter survey to most of the remainder of the island with the exception of the Akamas Peninsula and Cape Andreas. This latter survey was fairly detailed, stations being occupied at intervals of roughly one mile along all the motorable roads. As the Cyprus road system is good there were no large gaps in the gravity coverage except the areas just mentioned. The areas covered are shown on figure 6. Most of the pendulum stations were reoccupied with the O.G.S. gravimeter. Discrepancies did not often exceed 3 mgal and the larger discrepancies may be partly due to incorrect relocation of the original stations.

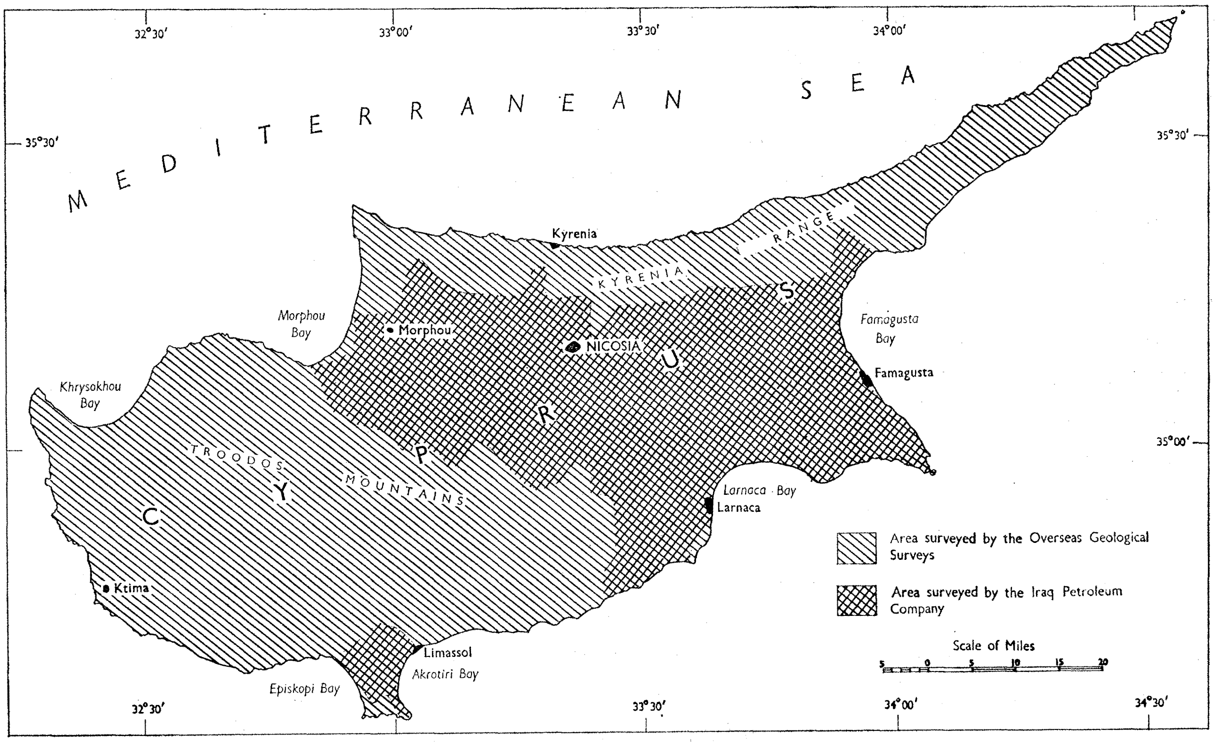


FIGURE 6

Discrepancies between the O.G.S. and I.P.C. surveys were rather larger than would be expected between modern gravimeter surveys. This again may be due to uncertainty in the position of the I.P.C. stations. However, the combined surveys are adequate in accuracy and detail to delineate the larger gravity anomalies reliably.

4·2. *Measurements at sea*

Submarine pendulum measurements were made in the eastern Mediterranean in 1950, and the results were published by Cooper, Harrison & Willmore in 1952. The submarine stations were 50 to 100 miles apart, but give a useful idea of the behaviour of the gravity field in the sea areas around Cyprus. In 1955, J. C. Harrison published an interpretation of the gravity anomalies in the eastern Mediterranean including the Cyprus anomaly. Some of Harrison's conclusions are restated below: those concerning the regional anomalies are not altered by the additional information provided by the O.G.S. gravimeter survey of western Cyprus, but the new information enables more definite conclusions to be drawn about the source of the positive anomalies beneath Cyprus itself.

4·3. *O.G.S. survey*

Worden gravimeter No. 144 was used for the O.G.S. survey. This instrument was calibrated (*a*) by the makers, (*b*) at the Geological Survey calibration base at Macclesfield, England, in October 1957, (*c*) against Mace's pendulum stations in Cyprus, February to June 1958, and (*d*) at the Rome calibration base of the Osservatorio Geofisico, Trieste, in July 1958:

calibration	gravity (gal)	small dial factor (mgal per division)
(*a*)	—	0·11251
(*b*)	981·2	0·11250 ± 0·00003
(*c*)	979·9 to 979·6	0·11249 ± ?
(*d*)	980·2	0·11244 ± 0·00007

TABLE 3

pendulum station		Mace (gal)	Overseas Geological Surveys (gal)	difference (mgal)
no. 1	Apostolos Andreas	979·874	979·8700	4·0
no. 2	Myrtou	·837	·8364	0·6
no. 3	Lefkoniko	·878	·8731	4·9
no. 4	Nicosia	·850	·85010	−0·1
no. 5	Kalokhorio	·922	·9208	1·2
no. 6	Platania	·657	·6562	0·8
no. 7	Larnaca	·861	·8596	1·4
no. 8	Agros	·715	·7147	0·3
no. 9	Troodhitissa	·613	·60943	3·6
no. 10	Stroumbi	·730	—	—
no. 11	Lefkara	·755	·7526	2·4
no. 12	Trimiklini	·750	·74829	1·7
no. 13	Limassol	·808	·80994	−1·9

Calibration (*c*) against the pendulum stations in Cyprus is the least reliable due to the errors of the pendulum measurements (see table 3). Calibrations (*a*), (*b*) and (*d*) are not significantly different and the factor 0·11250 is therefore adopted for the Cyprus survey. Mace's pendulum station No. 4 at Nicosia was used as gravity datum for both the O.G.S. and I.P.C. surveys. This station was first connected to Cambridge by Mace and later by I.P.C. to the Lebanon pendulum station. Both connexions give a value of 979·8501 gal. (Mace's published value was 979·847 based on Cambridge 981·265, but the Cambridge value has recently been revised to 981·2681.) Mace's other pendulum values are compared with the O.G.S. gravimeter values in table 3. There is little doubt that the discrepancies are largely due to errors in the pendulum values.

Three stations of the 1946 I.P.C. gravity survey were re-occupied during the 1958 O.G.S. survey. These are (*a*) Mace's pendulum station 4, Nicosia, (*b*) and (*c*) stations N 1146 and N 1270, respectively:

station	gravity difference (mgal) O.G.S.	gravity difference (mgal) I.P.C.
(*a*) to (*b*)	5·90	7·02
(*b*) to (*c*)	33·35	33·63

The agreement of these connexions is poor. These stations are accurately connected to the O.G.S. base network and the calibration factor of the O.G.S. instrument is probably accurate to 0·05 %. There is no doubt that station (*a*) was correctly relocated and the O.G.S. altimeter heights agree with the I.P.C. heights for stations (*b*) and (*c*) to 1 m, but it is possible that the discrepancies are due to mislocation of the latter stations as these were near mileposts which may have been moved as a result of road improvements.

(1) **Values of gravity at base stations.** The area of the survey (south and west of the line Xeros–Malounda–Khirokhitia) has been covered with a net of forty-three base stations. These stations are accurately connected on the small dial of the gravimeter by the leap-frog method (12123234...). The error in the connexions will usually be 0·01 or 0·02 mgal

TABLE 4

station number	location	gravity (gal)	station number	location	gravity (gal)
	Nicosia	979·85010*	271	Yerakies	979·72775
1	Limassol	·81435	288	Kannaviou	·82319
2	Erimi	·79601	292*a*	Kouklia	·77288
6	Evdhimou	·78444	324	Ktima	·81460
10	Khalassa	·78834	325	Timi	·77728
13	Trimiklini	·74590	333	Tsadha	·73540
14	Souni	·75006	334	Stroumbi	·75200
15	Mallia	·72481	344	Polis	·86620
16	Mandria	·70214	350*b*	Xeros	·95562
35	Kato Amiandos	·68844	351	Kato Pyrgos	·98568
40	Pano Amiandos	·63787	392	Chakistra	·78731
42	Pano Amiandos	·56309	431	Skoulli	·83451
53	Troodos	·54424	561	Yiallia	·94183
58	Platres	·61633	593	Xerarkaka	·86761
100	Pedhoulas	·65905	599	Malounda	·84705
109	Parekklisha	·80461	601	Nikos	·79297
111	Parekklisha	·75948	603	Moutoullas	·70267
116	Odhou	·72809	624	Kato Koutraphas	·90322
124	Melini	·78204	625	Mitsero	·86389
130	Khirokhitia	·80611	626	Ayia Marina	·85048
157	Kakopetria	·71690	631	Nikitari	·88493
229	Pharmakas	·71572	N 1146	Menoyia	·84419†
232	Gourri	·79389	N 1270	Limassol	·81084†

* Adopted datum. † I.P.C. station.

and probably not more than 0·03 mgal. The base station network is shown with closing errors in figure 7. The stations were sited as far as possible on the metalled roads so that the connexions could be made quickly. Progress on the unsurfaced mountain roads was generally so slow that it would have been uneconomical in view of the shortage of time to travel over them more than once and they could only be used for secondary stations. This has resulted in the loops of base stations containing a larger number of connexions than is desirable. The closing errors have been removed by adjusting individual connexions by amounts varying between 0·01 to 0·03 mgal according to the reliability of the drift curves. After these adjustments the error in the gravity difference between any of the base stations and the gravity datum at Nicosia will probably not be more than 0·1 mgal. The uncertainty in the calibration factor of the gravimeter is about 0·05 % so that there may be an additional error of 0·2 mgal in the larger gravity differences.

The values at the base stations are shown in table 4.

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(2) **Location of stations.** The forty-three base stations were sited where possible close to such permanent or semi-permanent objects as signposts and mileposts. Location diagrams have been prepared for the base stations so that they can be re-occupied to an accuracy of one foot. Most of the secondary stations were sited on mountain roads where there are

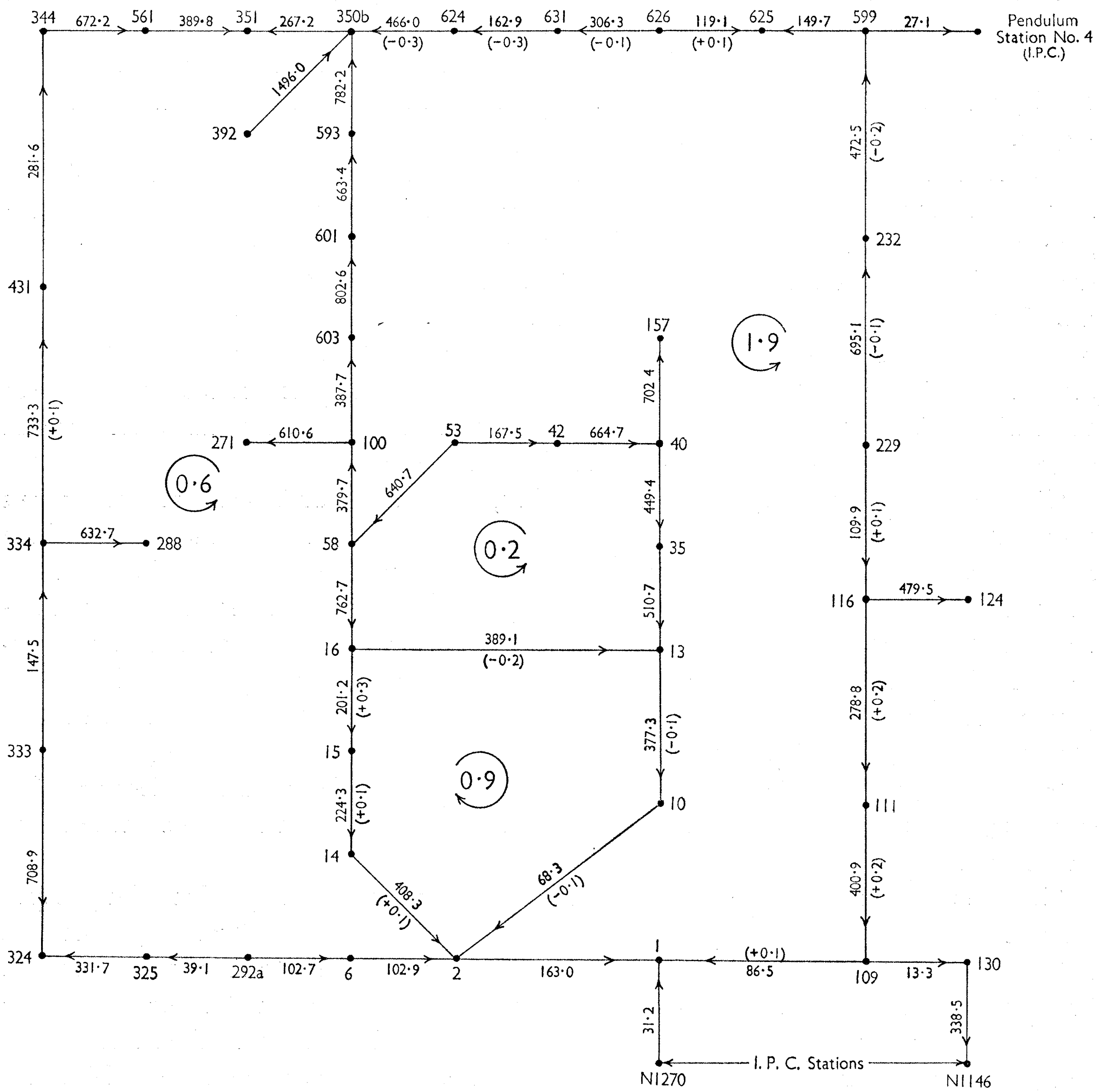

FIGURE 7. Overseas Geological Surveys gravity survey of Cyprus. Base station network.
Gravity differences in small dial units of Worden gravimeter No. 144. Closing errors shown in small dial units of Worden gravimeter No. 144. The adjustments made to remove closing errors are shown in brackets. Calibration factor 0·11251 mgal/unit

generally no mileposts or other prominent objects in positions suitable for gravity stations. To ensure that the stations were properly spaced on these roads the stations were placed at topographic features such as spurs, gullies, cols, etc., which would be easily identified on the map. Some of these stations could be relocated fairly accurately (e.g. where the

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road crosses a culvert or stream) but the position of others will be uncertain to several hundred feet. Although a more accurate location of stations is desirable the uncertainty in position should not be large enough to detract from the reliability of the Bouguer isogal map.

The positions of the O.G.S. stations are marked on the Bouguer isogal map, figure 8, pp. 448–9. The positions of the I.P.C. stations are not shown. These were at intervals of roughly half a mile along the roads in the Mesaoria.

(3) **Reduction of observed gravity to Bouguer anomalies.** In addition to observed gravity (*a*), the height (*b*), latitude (*c*) and topographic attraction (*d*) at each station are required to compute Bouguer anomalies.

(*a*) About half the secondary stations were read on the small range dial of the instrument with proper drift control at base stations at two-hourly intervals. Relative to the base stations the error is probably about 0·05 mgal and the total error (i.e. relative to the datum in Nicosia) probably not more than 0·15 mgal. At the rest of the secondary stations the large range dial was used to avoid the necessity of frequent resetting due to the rapid height changes encountered in the interior of the island. The error at these stations was difficult to estimate but was undoubtedly larger than when the small range dial alone was used. The drift of the instrument was quite small when the large range dial was in use and suggests that relative errors were not more than 0·2 mgal.

(*b*) There are very few trigonometrical points, spot heights or bench marks in Cyprus sufficiently close to the roads to be of use as a gravity station, so that the heights of the stations were taken with altimeters. The altimeters were checked at sea level and at accessible triangulation points. As readings were repeated at gravity base stations a pressure drift curve could be built up for each day and some allowance made for atmospheric changes. Corrections were made for temperature and pressure departures (observed at R.A.F. meteorological stations) from the standard atmospheres for which the altimeters are calibrated. The probable error in height is 10 ft.

(*c*) The position of some of the stations is uncertain to several hundred feet or a few seconds of latitude, but the resulting error in normal gravity will be trivial compared with errors from other sources.

(*d*) The terrain in most of the area surveyed is very rugged and produces large topographic attractions. The only contoured maps available are the 1/50 000 military maps with 100 ft. contour intervals. These are adequate for computing the attraction of topographic irregularities distant from the station but are not sufficiently detailed for the nearer terrain. The expected error in the topographic correction is 0·5 mgal in the mountainous areas. The corrections were calculated with the aid of an electronic computer which cuts down the labour involved and improves accuracy. The computer considers terrain out to 128 km from the station. The effect of more distant terrain is quite easily estimated and a second correction was applied for it.

Taking into account all the above factors the error in the Bouguer anomaly is expected to be 0·9 mgal. This error is large by normal standards, but is only 1 % or less of the total fluctuation of the anomalies which throw most light on the structure of Cyprus. Errors of measurement should not, therefore, detract from the reliability of the interpretation of the large anomalies. Fluctuations of a few mgal, however, should not be regarded as significant, particularly where they depend on a single secondary station.

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The density used in the Bouguer reduction is 2·4 g/cm³ for the Tertiary and Recent sediments, and 2·7 g/cm³ for the older rocks both sedimentary and igneous. The density assumed for the sediments is probably representative, and in any case the elevation of the sedimentary areas rarely exceeds 500 ft. except in the Kyrenia Hills. Systematic error in the Bouguer reduction to sea level should not therefore exceed 1 mgal over the sediments. The density of 2·7 g/cm³ assumed for the igneous rocks is about correct for the Sheeted Intrusive Complex and the Pillow Lava Series, but is 0·2 g/cm³ too low for the ultrabasic rocks and gabbros which are exposed around Mount Olympus. These latter rocks probably dip away from the margin of their outcrop at a low angle, so that the mean density of the rock between ground level, which is 4000 to 6000 ft. in this area, and sea level is rather higher than 2·7 g/cm³. The reduction of the Bouguer anomaly to sea level may therefore be 5 to 10 mgal too positive within 10 to 15 miles of Mount Olympus. A strong negative anomaly overlies the area near Mount Olympus which will be shown later to be due to rocks of density about 2·6 g/cm³ occurring at a shallow depth. This mass deficiency will offset the mass surplus due to the exposed ultrabasic rocks.

4·4. *Regional anomalies*

The regional Bouguer anomaly map (figure 8) shows that Cyprus is covered with strong positive anomaly mainly between 100 and 250 mgal. The axis of maximum anomaly runs parallel to the Kyrenia range extending from Pomos Point in the west, eastwards to Famagusta. The anomaly falls off all round Cyprus to less than 100 mgal but has risen again to over 100 mgal some 100 miles to the south. Elsewhere in the eastern Mediterranean the anomaly does not differ much from zero. The dense rocks which appear to have produced this large anomaly are shown later to have the form of a near-surface horizontal slab which is approximately delineated by the 100 mgal contour. This has an oblong shape, 120 miles east–west, 70 miles north–south, and displaced about 20 miles to the north-west of the centre of Cyprus.

The Bouguer anomalies at the submarine stations have been corrected for the mass deficiency of the sea. If this had not been done, that is if free air anomalies had been used at sea and Bouguer anomalies on land, Cyprus would have been surrounded by a belt of negative anomaly and the region of large positive anomaly 100 miles to the south would have disappeared. A similar picture is presented by the isostatic anomaly map (Airy–Heiskanen System: $T = 30$ km), figure 9, p. 450. Harrison (1955, figure 9, p. 297), shows that the negative isostatic anomaly to the west of Cyprus is the eastern margin of a more extensive zone of generally negative anomaly. The isostatic correction is negative everywhere in the area since it is largely marine, but becomes less negative over Cyprus. The effect of the correction is to increase the total fluctuation of anomaly to nearer 300 mgal, without much altering the shape of the contours over the island. The isostatic corrections assume that the mantle surface is warped upwards or downwards to provide local compensation for topography above and below sea level. No account is taken of the large mass surplus in the crust which produces the positive gravity anomaly or of the possibility that the compensation is regional rather than local. An approximate calculation shows that the net anomaly due to the mass surplus and its compensation, if local, would be about −30 mgal at a distance of 40 miles from the edge of the mass surplus. The actual mean negative anomaly at this distance from the edge of the mass surplus is rather larger. This suggests that the compensation is regional rather than local. The mean isostatic anomaly within 50-mile belts round the centre of Cyprus is:

0– 50 miles	140 mgal
50–100 miles	3 mgal
100–150 miles	−8 mgal
150–200 miles	−18 mgal

The mean anomaly for the whole area within 200 miles of Cyprus is −1 mgal. This suggests that the mean density of the crust is average in the whole area and that the mass surplus causing the large positive anomaly which is about 100 miles across is compensated by mass deficiencies spread out over a larger area. The estimates of mean isostatic anomaly are probably not very reliable over the sea areas because of the large distances between the submarine stations so that no firm quantitative conclusion can be drawn about the nature or extent of isostatic compensation. The existence of a central zone of strong positive mean isostatic anomaly surrounded by a belt of negative anomaly is quite consistent with the idea of an encrustal mass surplus beneath Cyprus compensated by a more widespread mantle down-warping, but it should be realized that the negative belts could equally well be caused by shallower mass deficiencies, say submarine sedimentary troughs, and that there may be no appreciable down-warping of the mantle surface. Further geophysical evidence is necessary to decide this point.

4·5. *The Cyprus anomalies*

The Bouguer anomaly map, figure 8, shows a broad belt of east–west positive anomaly which occupies the whole island. This is the main anomaly on which are superimposed smaller local anomalies. The isostatic reduction alters the level of anomaly everywhere but does not alter the shape or relative amplitude of the local anomalies. Its effect on the main anomaly is to steepen the gradients on the flanks slightly without altering the shape or trend of the contours. In view of the uncertainty as to the nature of the isostatic compensation, and the small effect the isostatic correction has on the Bouguer anomaly over the island, it seems preferable to interpret the Bouguer anomalies rather than the isostatic anomalies while making allowance for possible crustal thickening where necessary.

Before interpretation, the main and local anomalies must be separated. The axis of the main anomaly obviously follows the belt of igneous rocks forming the Troodos massif, but there is no obvious correlation of the local anomalies with superficial geology. The process of separating the anomalies must therefore be largely subjective. It is assumed that the contours of the main anomaly would have a steady east–west trend in the absence of the local anomalies. The main anomaly contours have been extrapolated across the local anomalies and the isolated local anomaly is taken as the difference between the hypothetical main anomaly contours and the observed anomaly. Listed below are the more important local anomalies.

Olympus anomaly. There is a sharp negative anomaly of 120 mgal amplitude centred on the ultrabasic outcrop. Hereafter, the term 'Olympus anomaly' will be used to designate this sharp negative anomaly over the summit of Mount Olympus.

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Pomos anomaly. A positive anomaly of 60 mgal amplitude centred on Pomos. The southern part of this anomaly is circular. The northerly extent is not defined since it is over the sea.

Polis anomaly. A negative anomaly of about 80 mgal amplitude at Polis. Southwards, towards Paphos, the anomaly broadens and its amplitude decreases.

Nicosia anomaly. A north–south elongated negative anomaly of 50 mgal amplitude cutting right across the main anomaly.

Lefkoniko anomaly. A north–south elongated negative anomaly of 60 mgal amplitude.

Kyrenia anomaly. A narrow band of positive anomaly, seldom exceeding 10 mgal amplitude, follows the Kyrenia range along the north coast.

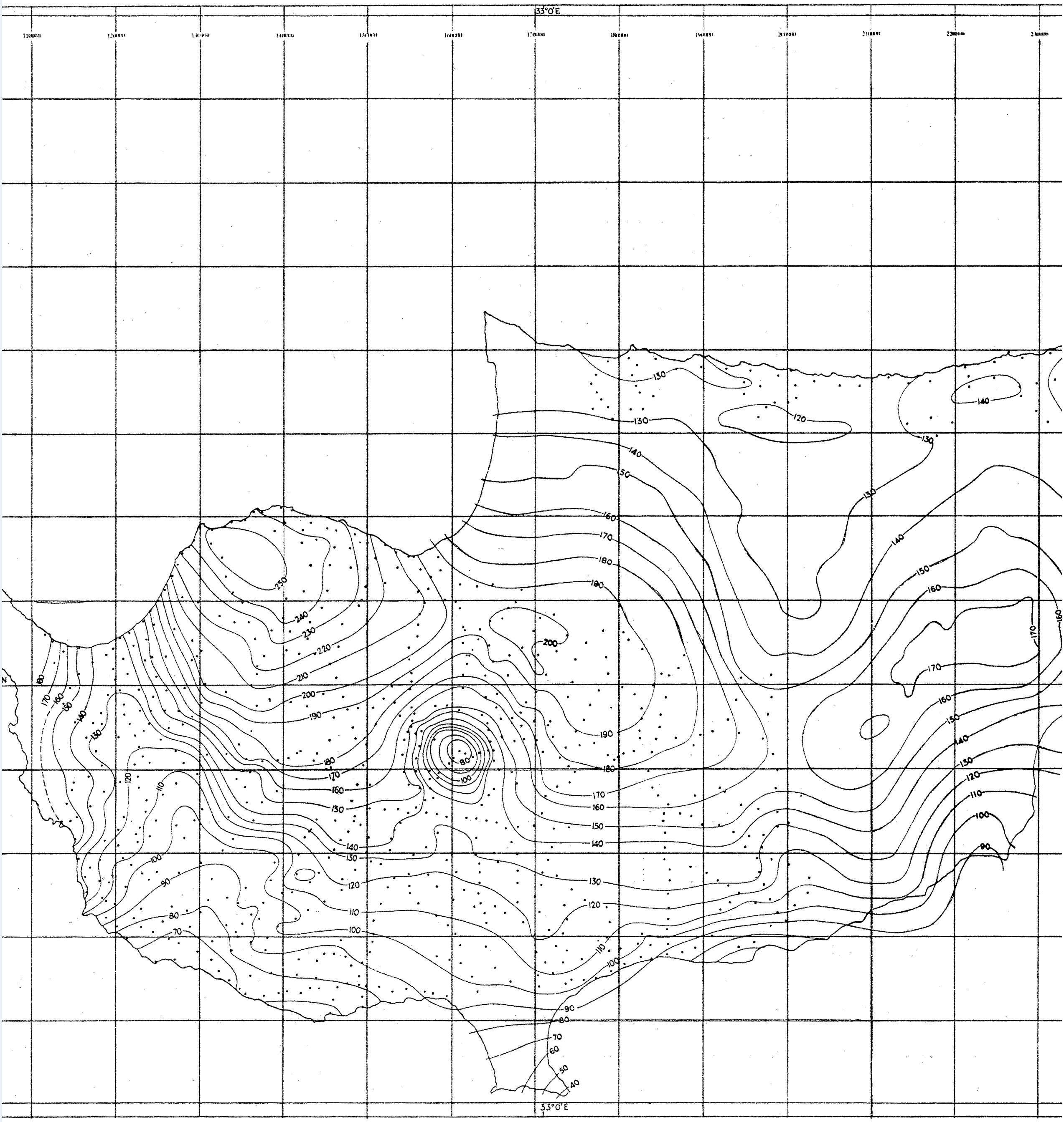
There are many other minor fluctuations of anomaly, some of which appear to be correlated with geological structure, but with a probable error of about 1 mgal in the Bouguer anomaly it is not considered that these are worth serious examination in this paper which is primarily concerned with the major structure of Cyprus. It is proposed to discuss the main and local anomalies separately.

Density contrasts. A combination of density measurements on samples and calculated normative densities suggests that the following effective bulk densities should be ascribed to the main rock types:

	g/cm^3
ultrabasic rocks (at depth)	3·30
ultrabasic rocks (near surface)	2·90
Sheeted Intrusive Complex	2·77
Pillow Lava Series	2·65
serpentine (fully hydrated)	2·55
sediments (at depth)	2·40

This list is a simplification in that rocks which occur in considerable bulk but with slightly different densities have been grouped together, but there is too little information about the density of Cyprus rocks to justify more detailed subdivision.

The Bouguer anomaly is the difference between observed gravity and the value expected on the model earth which most closely represents the actual earth. In continental areas the mean density at the surface is about 2·7 g/cm^3. Non-zero Bouguer anomalies are therefore expected over areas where there are appreciable departures from this density in the near-surface rocks. The present-day environment of Cyprus is intermediate between typical continental and oceanic conditions in that the surrounding Mediterranean is between 500 to 1000 fathoms deep. In this situation isostatic theory requires the upper layer to be sialic with a density of about 2·7 g/cm^3 but to be thinner than under typical continental areas. On this basis the Sheeted Intrusive Complex and the Pillow Lava Series which make up most of the Troodos massif at the surface would not produce large anomalies unless present in very great thicknesses. The Sheeted Intrusive Complex, for instance, could not produce the main positive anomaly unless it was 45 miles thick. It is assumed in the following interpretation that the ultrabasic rocks of the Troodos Plutonic Complex produce positive anomalies (particularly the unaltered ultrabasic rocks at depth) and that the sediments and fully hydrated serpentine produce negative anomalies, whilst the Sheeted Intrusive Complex and Pillow Lava Series produce no anomaly. It is probable that these latter groups do produce anomalies of a few mgal but these will be sufficiently small in comparison with the major anomalies to be neglected.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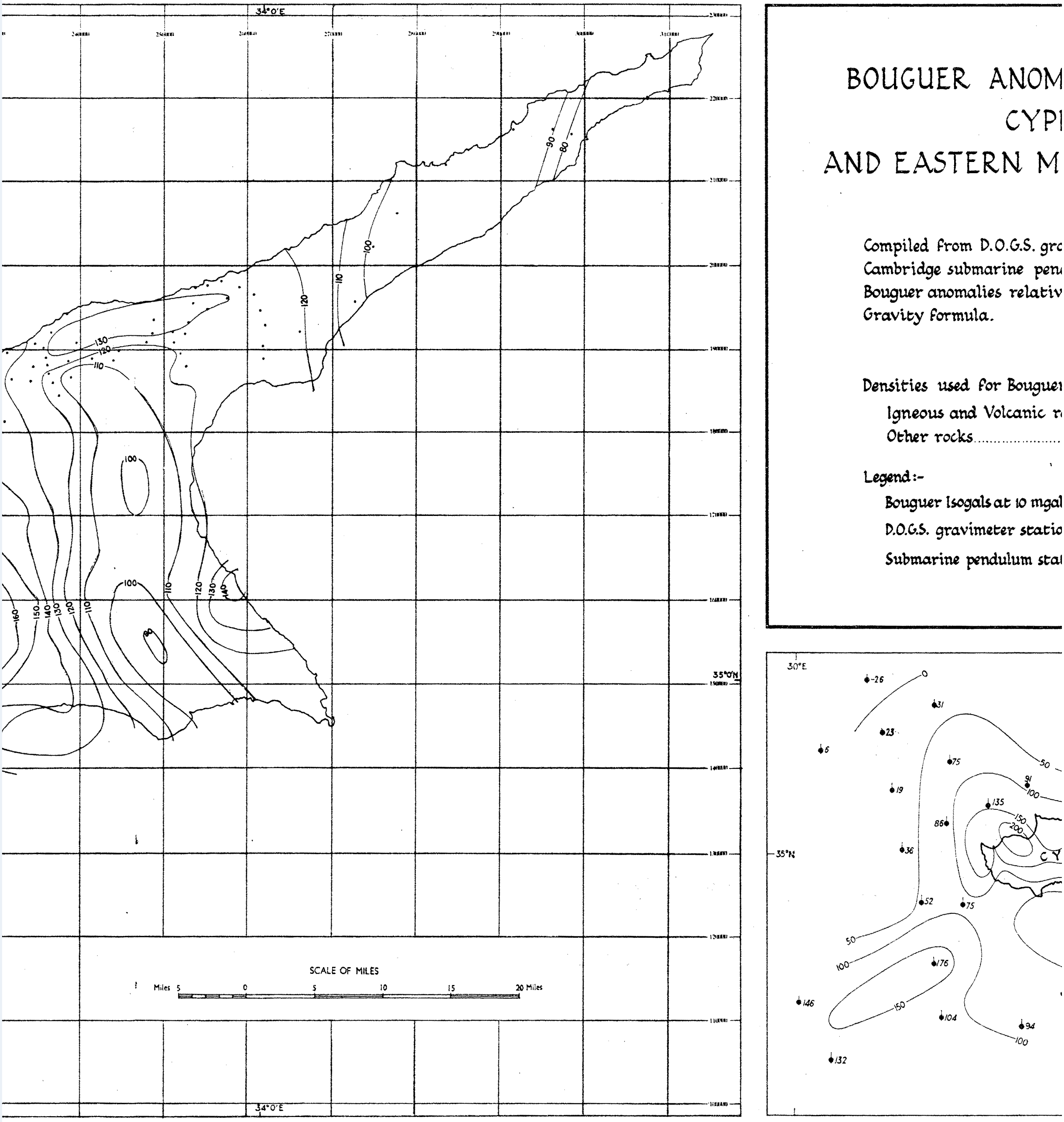

FIGURE 8

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ANOMALY MAP OF
CYPRUS
RN MEDITERRANEAN

D.O.G.S. gravimeter surveys and
ıarine pendulum measurements.
ies relative to the International
ı.

ʼor Bouguer reductions:-
Volcanic rocks.......2·7 g/cm^3
.................................2·4 g/cm^3

ls at 10 mgal intervals ——110——
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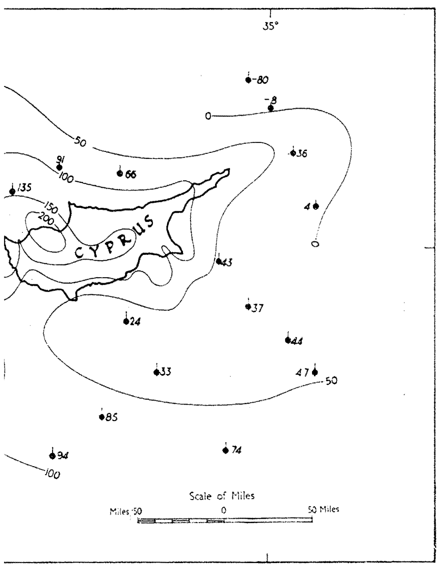
ndulum stations....................

December 1959



(1) **Olympus anomaly.** A circular negative anomaly of 120 mgal occurs over the ultrabasic outcrop around Mount Olympus and is almost concentric with it. The anomaly is well defined in an east–west direction and there are sufficient stations to the north and south to show that it is closed and circular. The surface density in this area is about 2·9 g/cm³. The eastern portion of the outcrop is serpentinized but its density, although locally reduced to below 2·9 g/cm³, would not be low enough to cause an anomaly of this size. Further,

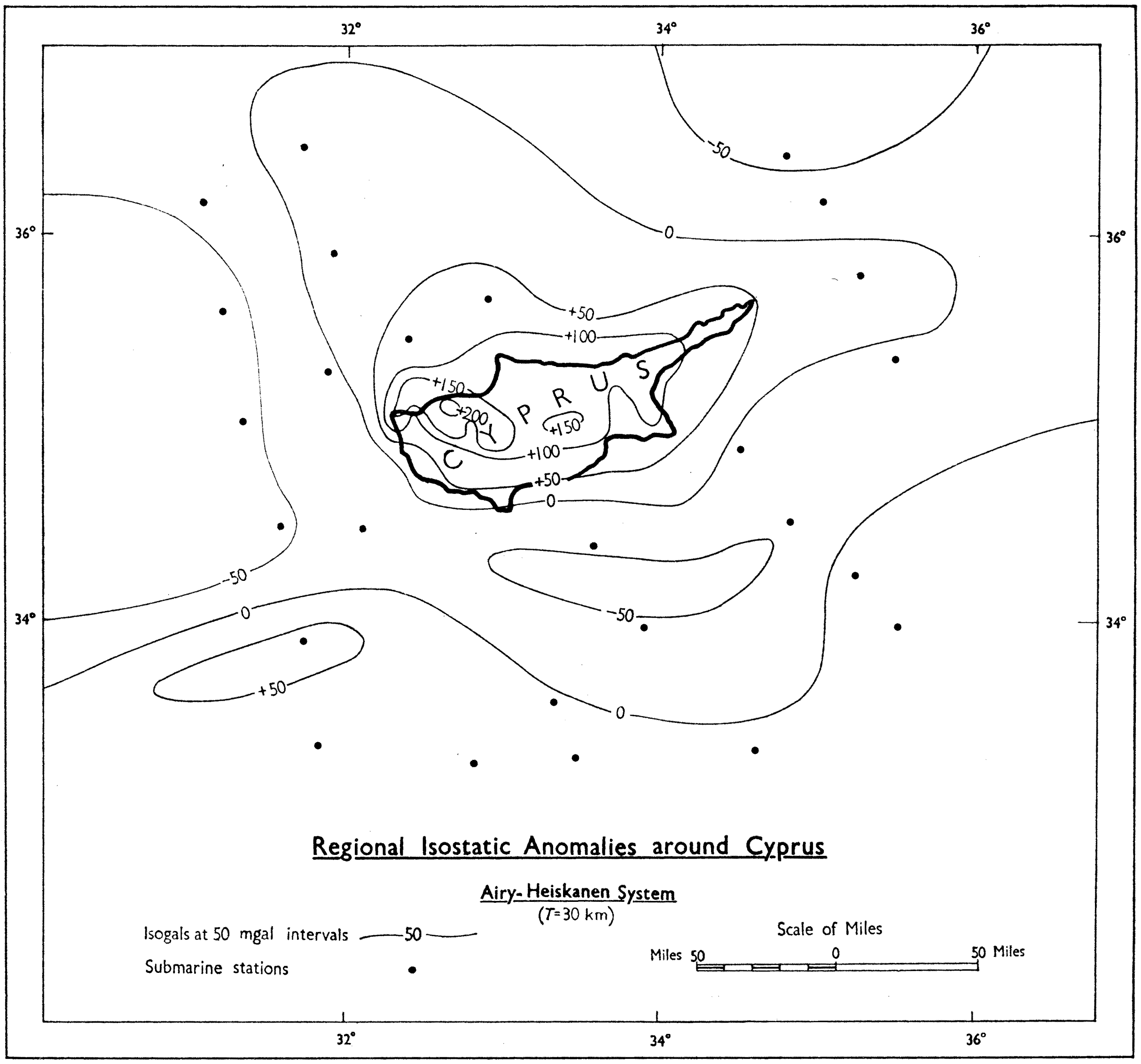


FIGURE 9

this surface serpentine has been shown (Wilson 1959) to be the serpentinization of enstatite-olivinite *in situ* and would, premsuably, not extend to any great depth. In any case, as the anomaly is correlated with the whole ultrabasic outcrop rather than the serpentinized portion, it appears that the low-density material causing the negative anomaly does not reach the surface.

Smith (1960) has recently published some formulae for finding limits to the depth of the upper and lower surfaces and the density contrast of structures producing gravity anomalies. Using these tests the following limits are derived for the Olympus anomaly:

the density contrast is at least 0·68 g/cm^3,
the upper surface is less than 1·8 miles deep,
the lower surface is more than 0·75 miles deep.

Figure 10 shows comparisons of the observed anomaly (*A*) with cylindrical model anomalies. The first comparison (*B*) is with two vertical cylinders 6 miles in diameter (the

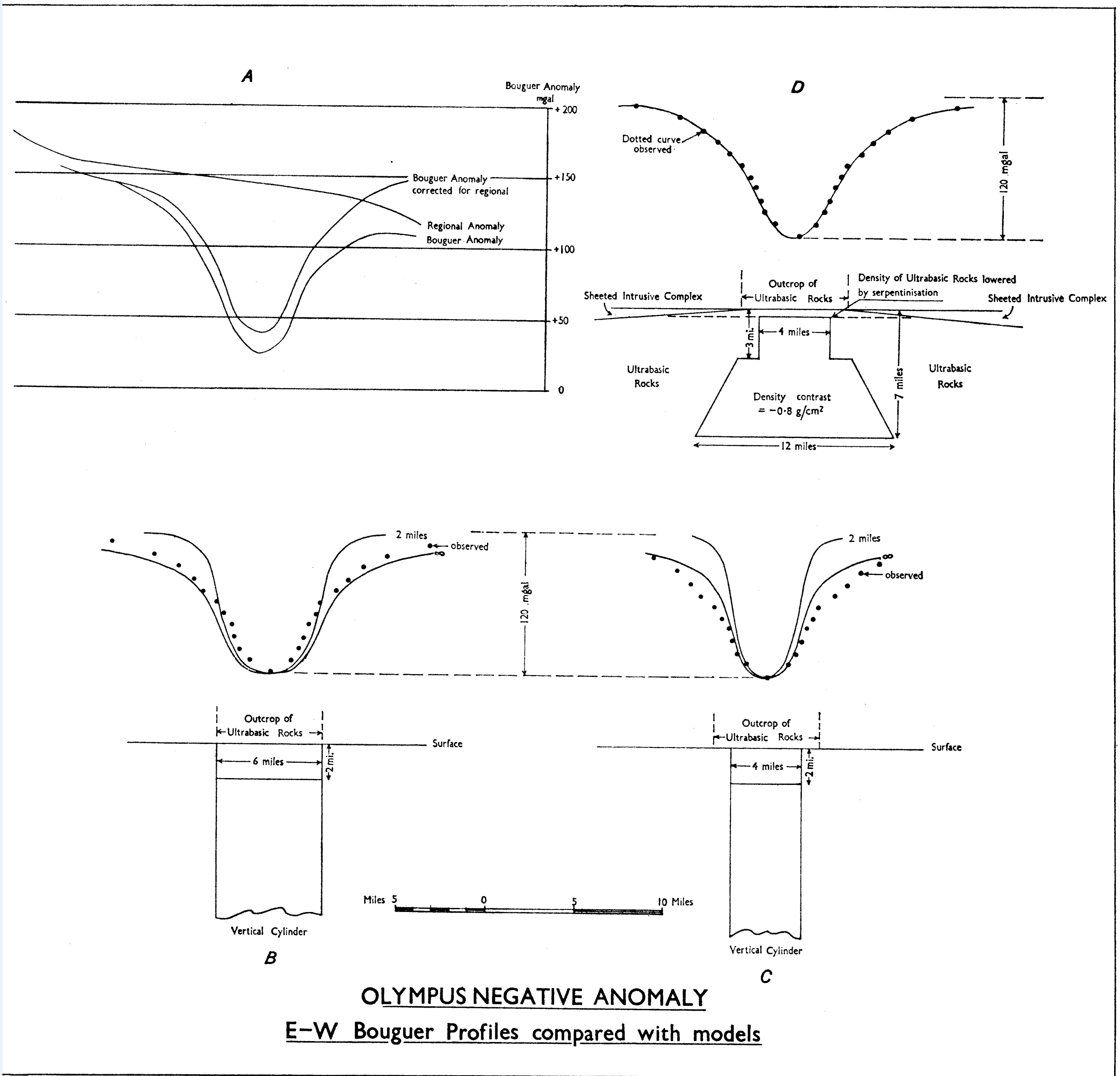


FIGURE 10

diameter of the ultrabasic outcrop), one 2 miles deep and the other of infinite depth. The model curves are solid and the observed curve dotted. The density contrasts of the models are adjusted to give the observed total fluctuation of anomaly. Cylinders of intermediate depths will have anomaly curves enclosed within the two extreme cases. It is clear that the diameter of the model must be reduced, since the bowl of the model anomaly is wider than the observed anomaly for cylinders 6 miles in diameter. The second diagram (C) shows a good fit at the bowl with cylinders 4 miles in diameter, but the flanks of the observed anomaly are more negative even than the flanks of the cylinder of infinite depth. This shows that the model should be 4 miles in diameter near the surface but that its diameter should increase with depth. A fairly good fit is produced by the conical model (D). This has its top surface at a depth of 2000 ft. and is 4 miles in diameter at the top. The walls plunge vertically for 3 miles; thereafter, the diameter increases steadily to 12 miles at a depth of 7 miles. The density contrast required to produce the observed anomaly is 0·85 g/cm³.

In the interpretation of the Olympus anomaly no account has been taken of the variation of the anomaly with elevation. The elevation at the centre of the anomaly is 5500 ft. and at the flanks about 4000 ft. If the anomaly had been measured on a plane 4000 ft. above sea level the flanks of the anomaly would not be much affected but the centre would become 10 mgal more negative. Furthermore, the rocks overlying the centre of the Olympus mass deficiency have a density of 2·9 g/cm³ while the Bouguer reduction was made for a density of 2·7 g/cm³. When allowance is made for this the central part of the anomaly will become more negative by a further 4 mgal. Since only the central part of the anomaly is affected the conical model of figure 10 can be made to reproduce this revised anomaly by bringing its top surface 1500 ft. nearer the plane of measurement without altering the density contrast. This places the top surface at an elevation of 3500 ft. above sea level. The fields due to the conical and cylindrical models were calculated by zone chart methods. At the centre and edge the field can be calculated precisely in terms of tabulated functions. This serves as a check on the approximate method.

Since the dimensions and density contrast of the model differ only slightly from the limits given above, the model must resemble the actual mass deficiency quite closely.

In this geological setting the low-density material producing the anomaly would, almost certainly, be igneous. As the low-density material extends to a depth of some 7 miles it seems most unlikely that it is either of glassy texture or highly vesicular. It is improbable that a magma could be cooled quickly enough at depths of 7 miles to form a glass; furthermore, the density difference between crystalline and glassy ultrabasic rock is about 0·5 g/cm³ which is hardly large enough to explain the anomaly. If some 25 % of the volume of an ultrabasic rock were occupied by vesicles the necessary density contrast would be obtained. This porosity is exceptional even for extrusive rocks and would be most unlikely at depth. Only two rocks would appear to meet both density and petrological limitations, these are granite and serpentine.

Granite. It is shown later that the ultrabasic rocks have probably been underthrust by the northern margin of the African continental shield. A source of granite magma is therefore available beneath Olympus. Bott (1956) considers that granite bosses with a space-form similar to the Troodos low-density mass are emplaced by stoping of overlying higher

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density rock into liquid granite magma below. The space-form of the body causing the Olympus mass deficiency and the gravity anomaly are very similar to those observed over exposed granite bosses emplaced by stoping. The density of granite is at least 2·65 g/cm³ which means that the rock surrounding the boss would have a density of at least 3·3 g/cm³ and probably nearer 3·5 g/cm³, while the normative density suggested by the surface samples of the ultrabasic rocks at Olympus is 3·3 g/cm³. The density of naturally occurring ultrabasic rocks does not usually exceed 3·3 g/cm³ but their constituent minerals can have densities as high as 4·4 g/cm³.

Serpentine. Fully hydrated naturally occurring serpentine has a density of 2·5 g/cm³. This gives the ultrabasic rocks a probable density of 3·35 g/cm³ and a minimum density of 3·18 g/cm³, which agrees better with the density of the exposed plutonic rocks. On this basis serpentine is more probable than granite but its mode of emplacement is more difficult to understand. Modern views on the formation of serpentine (reviewed in Turner & Verhoogen 1960) are that it is either intruded in the semi-solid state or results from the hydration of ultrabasic rocks *in situ*. Experiments on artificial serpentine (Bowen & Tuttle 1949; Yoder 1952) show that it becomes unstable at the temperatures and pressures existing at the base of the crust but that it can probably exist to a depth of 7 miles. As there is no geological evidence of bulk expansion of the boss beneath Olympus, serpentinization of the ultrabasic rocks *in situ* by percolating water therefore seems unlikely. Forcible intrusion in a solid state seems more probable, but it is then difficult to reconcile the cylindrical space-form of the boss with the pronounced north–south elongation of the exposed intrusives in the Sheeted Intrusive Complex.

On the evidence available granite and serpentine seem to be most probable as the source of the Olympus anomaly, although neither is entirely satisfactory, granite having too high a density and the space-form being unlike known serpentine masses. An air-borne magnetometer traverse across Olympus would help to decide this point as granite should have a negligible susceptibility compared with the surrounding ultrabasic rocks, while serpentine should have a comparable or even larger susceptibility.

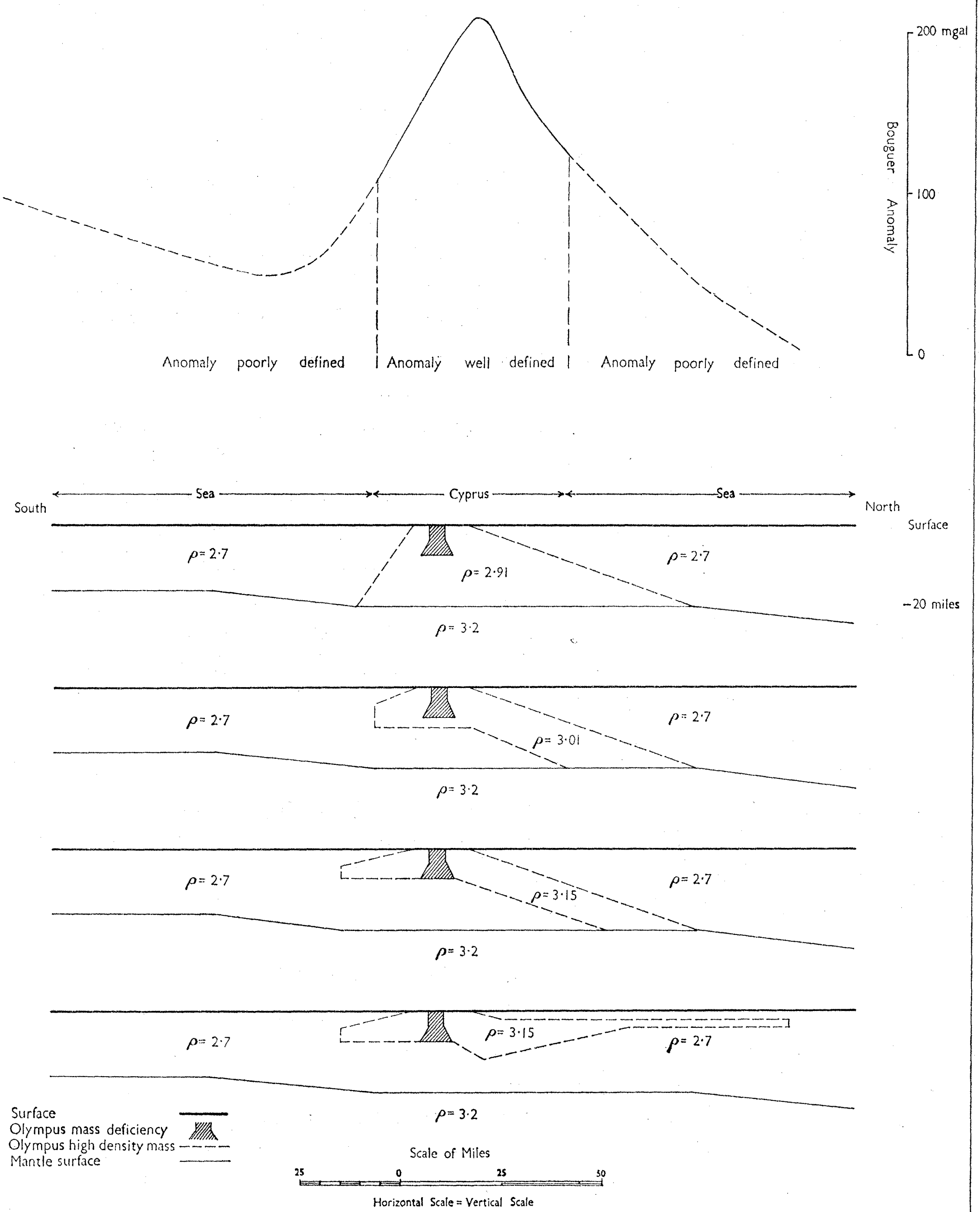
The Olympus negative anomaly could not have been anticipated on geological grounds. Geological evidence shows that the ultrabasic rocks dip under the Sheeted Intrusive Complex at a low angle. This would produce a small positive anomaly, which may be superimposed on the larger negative anomaly. Apart from the negative anomaly there are no abrupt changes of anomaly in the vicinity of Mount Olympus. This suggests that the roof of the ultrabasic mass continues to dip at a low angle to the east and west of Troodos, with rather larger dips to the north and south.

The concentricity of the Olympus ultrabasic outcrop and the negative anomaly can hardly be fortuitous. Olympus is an elevated area. This could be due to local isostatic uplift in response to the mass deficiency producing the negative anomaly. This mechanism has been proposed by Bott (1956) to account for doming of granites. The mass deficiency is equivalent to 100 cubic miles of rock of density 2·9 g/cm³. This potential uplift greatly exceeds the actual local topographic uplift which is about 5 cubic miles above the 4000 ft. contour in the vicinity of Mount Olympus. Since the actual present-day topographic uplift is only 5 % of the potential uplift in terms of mass deficiency at depth it seems quite probable that uplift in the Olympus area has been continuous since the emplacement of the mass deficiency and that this, combined with surface erosion, has resulted in the more extensive denudation of the ultrabasic rocks at Olympus than at any other point in the extensive area underlain by the ultrabasic material.

(2) **The main positive anomaly.** The Olympus negative anomaly has already shown that between $\frac{1}{2}$ mile and 7 miles of the surface there must be a layer of material of density greater than 3·2 g/cm³. The high-density layer may be thicker than $6\frac{1}{2}$ miles but is certainly not thinner. The upper half of the typical continental crust is mainly granitic to dioritic with a density of about 2·7 g/cm³. If covered with sediments or metasediments the mean density of this layer may be lower or higher than 2·7 g/cm³, but this figure seems a fairly good mean value to take for the purposes of the following argument. An extensive layer of this high-density material replacing normal continental crust would cause a positive Bouguer anomaly of at least 240 mgal if $6\frac{1}{2}$ miles thick.

Figure 11 shows four models which could produce a north–south section of the main anomaly through Troodos. These models are two-dimensional—i.e. they are assumed to have infinite extent in an east–west direction whereas the high-density layer probably does not actually extend more than 60 miles to the east and west of the section. The assumption of infinite extent does not, however, affect the computed field by more than a few percent. Since the actual length in an east–west direction is uncertain there is no point in making allowance for the truncation. In these models the crust is assumed to have normal thickness under Cyprus but too thin to the south so as to provide isostatic compensation for the sea. Ideally, geological corrections should be applied for the sediments to the north and south of the Troodos massif. Unfortunately, the thickness of these sediments is not accurately known so that no reliable correction can be applied. These corrections are not likely to be more than a few tens of mgal and if applied would have the effect of reducing the gradients at the flanks of the anomaly but would have little effect on the central part. The effect on the models would be to reduce the dip of the upper face to the north and south without greatly affecting their thickness in the centre. In each model the density contrast relative to a normal crustal density of 2·7 g/cm³ is adjusted to give the observed total anomaly fluctuation. The only acceptable models are the third and fourth which have a density of 3·15 g/cm³ required by the Olympus negative anomaly. The density of the first two models is much too low. Even the density of 3·15 g/cm³ which corresponds to the thinnest possible slab of ultrabasic rock at Olympus is too low. This suggests that the main positive anomaly is superimposed on a regional negative anomaly. If the regional negative anomaly was 50 mgal the effective fluctuation of the positive anomaly would be increased from 200 to 250 mgal at Olympus. This would enable the ultrabasic density in models 3 and 4 to be raised to 3·4 g/cm³ without much alteration of shape which is more consistent with the environment demanded by the Olympus negative anomaly. Such a regional negative anomaly could be caused by crustal thickening of $1\frac{1}{2}$ miles beneath Cyprus brought about by isostatic adjustment to the excess load of the ultrabasic slab or to tectonic processes.

The cross-sections of the ultrabasic slab suggested by the models should be treated with some reserve. Harrison (1955) has assembled some seismic and borehole evidence for down-faulting along the northern and southern flanks of the Troodos massif which may amount to several thousand feet. There is evidence that the Recent sediments south of



Models which can reproduce a N–S profile of the Bouguer Anomaly through Troodos after removing the field due to the Olympus negative anomaly model.

FIGURE 11

Nicosia were protected from contortion during the Alpine orogeny by a stable block at shallow depth, while the sediments to the north are highly contorted. This implies a fairly sudden change in depth of the ultrabasic slab. Although the cross-sections may not represent the ultrabasic slab accurately in detail they should give a fairly reliable idea of its general distribution.

The limits imposed by the Olympus negative anomaly show that the slab is close to the surface round Olympus and at least 7 miles thick. The steady east–west trend of the isogals suggests that this thickness and depth is maintained for considerable distances to the east and west of Olympus. The appearance of the main anomaly after removing the Olympus and Pomos anomalies is shown in figure 12. There is little doubt that the upper surface of the slab and possibly the lower surface dip to the north and south. The actual rate of thinning would be determined by the nature of the flanks of the anomaly, but this is not possible as no reliable geological correction can be made for the sediments, and there is no reliable gravity information for the sea areas.

As the anomaly is not well defined over the margins of the slab its areal extent cannot be delineated with any certainty. With a near-surface structure of this type the 'half total fluctuation' anomaly should occur over the margins. If the slab is steep sided this will roughly correspond to the 100 mgal contour. On this basis the southern margin runs east and west through Paphos and a point 12 miles south of Larnaca. This position for the southern margin seems to be supported by the results of an aerial magnetic traverse discussed below. The eastern margin runs about 10 miles east of Famagusta and the western margin about 20 miles west of the Akamas Peninsula. To the north the anomaly gradients are smaller and if the structure resembles model 3 of figure 11 the 100 mgal contour will not mark the margin of the ultrabasic slab.

Surprise has been expressed in the past that Cyprus is an area of present-day uplift, when the large mass surplus indicated by the gravity anomaly would lead one to expect isostatic down-warping. This could be explained if tectonic processes had originally produced crustal thickening in excess of that required to balance the surplus mass of the ultrabasic slab. Subsequent adjustment would tend to produce uplift which could continue until balance had been reached.

Kennedy (1960) has proposed an alternative theory of isostasy in which the classical chemical boundary between the crust and mantle is replaced by a basalt–eclogite phase boundary. The difficult concept of the lateral transport of large quantities of material involved in classical isostatic adjustment caused by pressure changes only is replaced by a vertical migration of the phase boundary in response to both pressure and temperature changes. Changes in the temperature gradient in the crust, after the establishment of isostatic equilibrium, could cause uplift without appreciable change in surface loading. A rise in temperature at the phase boundary causes it to move downwards so that the material above expands and causes uplift at the surface. The emplacement of this large mass of high-density rocks near the surface would have considerably disturbed the temperature gradients in the crust and could have caused the Cyprus uplift without excess crustal thickening.

The gravity results suggest a minimum thickness of 7 miles for the ultrabasic slab, and a minimum crustal thickening of $1\frac{1}{2}$ miles beneath Cyprus. A thicker slab with greater

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crustal thickening would still be consistent with the gravity evidence provided the net Bouguer anomaly remained 200 mgal at Olympus.

To provide local isostatic compensation for the ultrabasic slab the crust would have to be thickened by at least 7 miles beneath Cyprus. It has already been pointed out that the

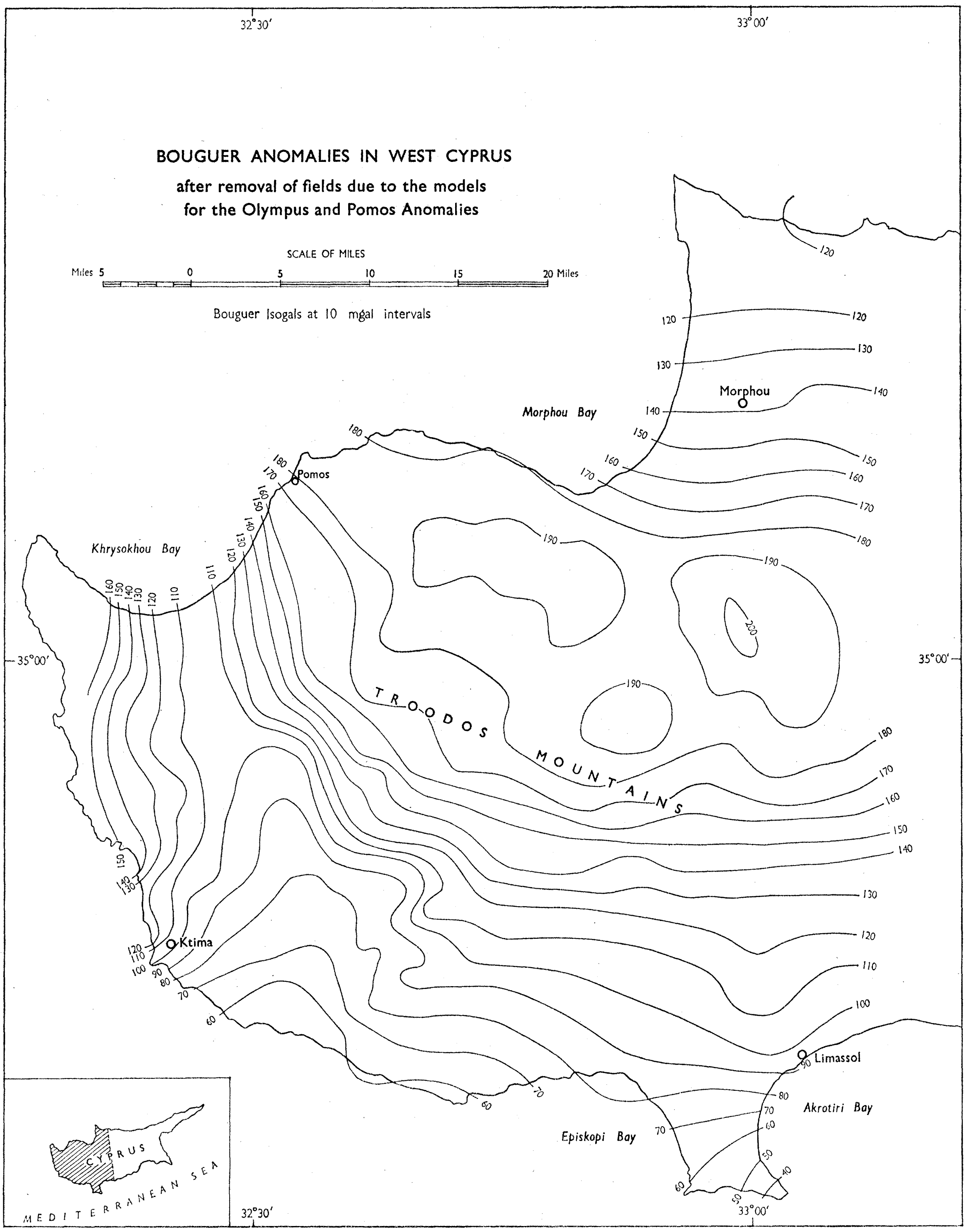

FIGURE 12

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mean isostatic anomaly within 200 miles of Cyprus is nearly zero, which suggests that the mass surplus of the slab is balanced in the area as a whole by crustal thickening. Within 150 and 200 miles of Cyprus the mean anomaly is −18 mgal, which could be produced by crustal thickening of $\frac{1}{2}$ mile. There is some evidence therefore that in addition to changes of thickness required to balance topography there is a basin-like thickening more than 400 miles in diameter centred on Cyprus which balances the ultrabasic slab regionally but not locally.

(3) **Pomos anomaly.** Centred on Pomos there is a broad positive anomaly of 60 mgal which appears to be superimposed on the axis of the main positive anomaly. Although only the southern half of the anomaly lies over the land a submarine station to the north suggests that the Pomos anomaly cannot be much elongated out to sea, and probably has a roughly circular outline. Smith's test applied to this anomaly gives the following limits:

the density contrast is greater than 0·18 g/cm^3,
the depth to the top surface is less than 4·8 miles, density contrast = 0·18 g/cm^3,
the depth to the top surface is less than 18 miles, density contrast = 0·50 g/cm^3.

These limits do not lead to any definite conclusions about the body producing the Pomos anomaly. They show that the mass surplus can be at any depth in the crust. Since geological evidence shows that the upper surface of the ultrabasic layer is deeper at Pomos than at Troodos, the anomaly could be caused either by a local increase in the density of the ultrabasic slab or by a downward protuberance of its lower surface. In the first case the density of the ultrabasic slab would have to be increased from 3·4 to 3·8 g/cm^3 within 6 miles of Pomos. In the second case the downward protuberance could be a cylinder with upper surface at a depth of 7 miles, a lower surface at 22 miles and diameter 16 miles. Its density would be 3·2 g/cm^3 relative to sialic material of density 2·7 g/cm^3. As the crustal density is probably higher than 2·7 g/cm^3 at these depths the density of the cylinder may be nearer 3·3 or 3·4 g/cm^3. It is not possible to discriminate with certainty between these alternatives but the low-gravity gradients on the flanks of the anomaly suggest that the deeper structure is more probable and the existence of a large volume of rock of density 3·8 g/cm^3 seems unlikely.

(4) **Polis anomaly.** The Polis anomaly seems to be a north–south elongated negative anomaly superimposed across the main positive anomaly. Although the Akamas Peninsula was not completely surveyed the anomaly rises to 180 mgal west of Polis and from the steepness of the gradients there appears every likelihood that it will continue to rise westwards to the general maximum value of the main positive anomaly, that is about 200 mgal. Farther south, the contours turn sharply west before reaching the coast. This suggests that the Polis anomaly does not mark the end of the main anomaly and is only an interruption in its steady east–west trend. The following limits are obtained for this anomaly:

the density contrast is greater than 0·25 g/cm^3,
the upper surface is shallower than 3300 ft. for density contrast 0·25 g/cm^3,
the upper surface is shallower than 8000 ft. for density contrast 0·5 g/cm^3,
the upper surface is shallower than 14000 ft. for density contrast 0·8 g/cm^3.

These results show that the mass deficiency must be above or within the ultrabasic slab but not below it.

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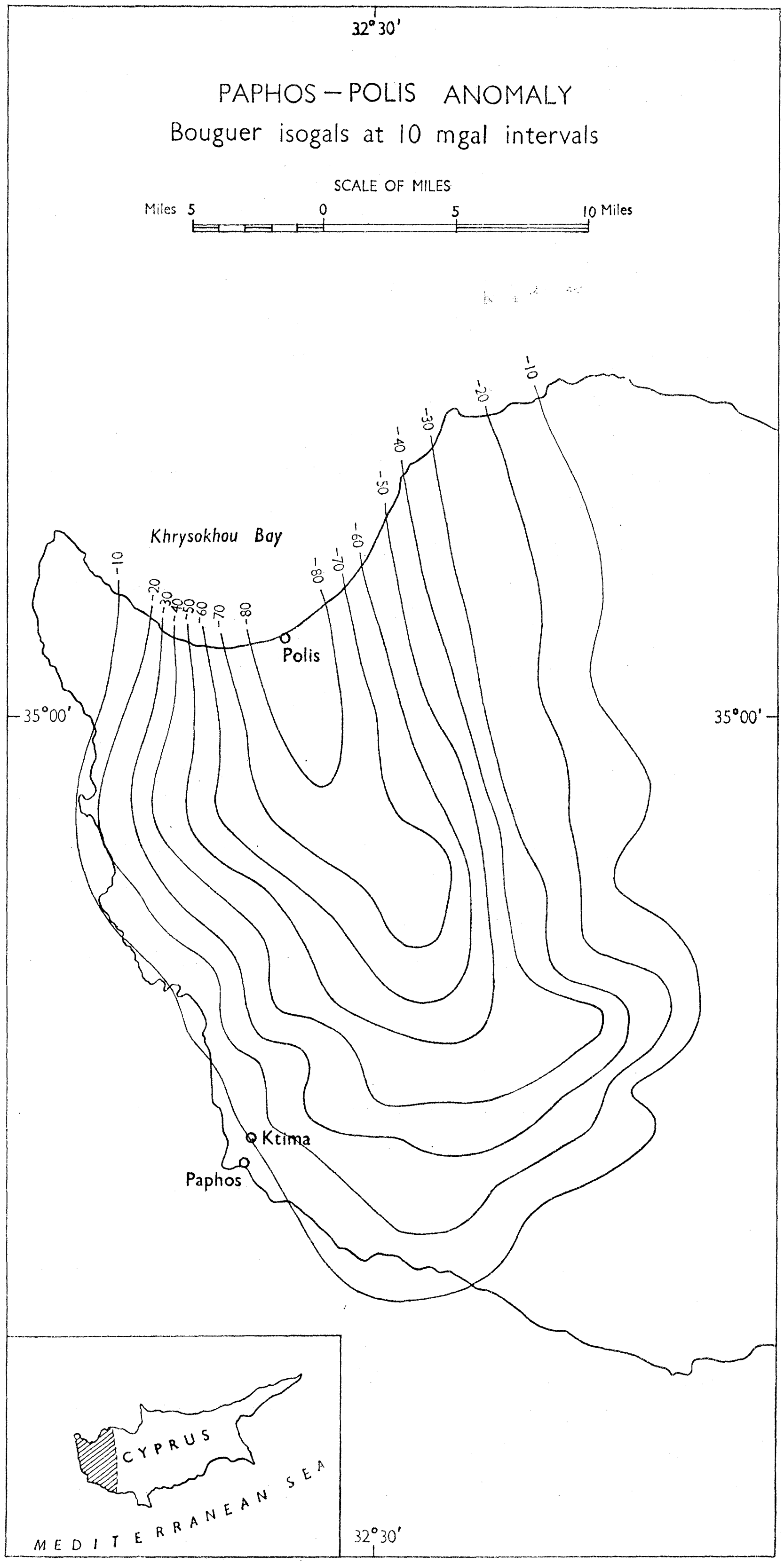

FIGURE 13

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Most of the area of this anomaly is covered by Recent, low-density sediments, and it is possible that the negative anomaly may be partly due to a trough of these sediments. At Polis the outcrop of the sediments is displaced to the west of the anomaly, and here the anomaly could not be produced by the sediments alone. Further, there are no sudden changes of the gravity gradient over the margins of the sediments; this would suggest that the sedimentary trough is not steep sided and may not therefore acquire great thickness at the centre. As there is no obvious geological correlation it seems more probable that the structure responsible for this anomaly is a deep-seated low-density mass which is not exposed. Exposed injected serpentine in the Akamas Peninsula appears to have come from the direction of this low-density mass, in that the eastern contact dips to the east at about 40°. Serpentine masses, which field evidence suggests were emplaced as a fully hydrolized serpentine magma, are abundant in this area.

As in the case of the Olympus anomaly the identity of the low-density mass is not certain although it is probably granite or serpentine. The exposures of serpentine and the north–south trend of the anomaly might suggest that in this case serpentine is more probable, but a granite magma emplaced in this position could have provided the necessary heat and water for the serpentinization of near-surface ultrabasic rocks. Here, again, the uncertainty could probably be resolved by a magnetic traverse.

(5) **The Nicosia and Lefkoniko anomalies.** These anomalies are similar in character to the Polis anomaly although of smaller amplitude. The areas are covered with Recent sediments and there is no geological evidence as to the nature of the mass deficiency. Seismic work in the Mesaoria indicates thicknesses of sediments of the order of 10000 ft. but does not confirm local increases in thickness which might explain the Nicosia and Lefkoniko anomalies. The situation is similar to the Polis anomaly in which the anomalies could be explained equally well by a thickening of surface sediments, or a more deep-seated low-density mass cutting through the ultrabasic slab. The absence of any surface indication of movement along the lines where these bounding fault planes would cut the surface strengthens the case for a more deep-seated structure.

4·6. *Other geophysical evidence*

In 1960, Agocs flew an airborne magnetometer across Cyprus en route from Beirut to Farnborough. The flight crossed Cyprus from Larnaca on the south coast to Kyrenia on the north coast. Agocs (1960) reports that between Beirut and a point 12 miles south of Larnaca the magnetic anomalies were broad and of low amplitude. Between this point and the north coast the anomalies were much sharper with an amplitude of 200 gamma. Beyond the north coast the anomalies became broader and of low amplitude. Over Troulli, where the ultrabasics are probably near-surface, the amplitude increased to 800 gamma for a few miles. As the sea is several thousand feet deep to the north and south of Cyprus a reduction in amplitude of magnetic anomalies would be expected in these areas. The sudden change in amplitude 12 miles south of Larnaca where the sea is about 500 fathoms deep may be significant and mark a transition from sialic crust to ultrabasic material.

Apart from the gravity results and this single magnetic traverse there is no other geophysical evidence as to the nature of the ultrabasic mass. Seismic depth probes have established considerable thicknesses of Recent sediments in places, but no reflector or refractor which could be positively identified as ultrabasic has, so far, been found.

5. Conclusions

Any interpretation regarding the origin of the Troodos massif must depend on the following conditions:

(*a*) The pre-Triassic age of the massif.

(*b*) That rocks which characterize Troodos are limited to Cyprus.

(*c*) That east–west tensional stress was dominant throughout the evolution of the massif.

(*d*) The ultrabasic nature of the Troodos igneous suite and the possibility that the parent magma was derived from the earth's peridotite mantle.

(*e*) The virtual absence of sediments within the massif.

(*f*) The fact that most extrusives display pillow lava structure.

(*g*) The probable presence of sialic crust under Cyprus.

We consider that these conditions are best satisfied by the following course of events.

(1) In pre-Triassic times the continents of Eurasia and Africa were more widely separated than at present, the area now occupied by the eastern Mediterranean was oceanic and the sialic crust was thin or absent.

(2) Vulcanicity commenced on the ocean floor in the Palaeozoic and started to form a volcanic pile, perhaps 50 miles in diameter, at the site of Troodos.

(3) East–west tensional stress, which caused the invading material to follow a north–south alinement, continued until the east–west dimensions of the volcanic pile had increased to 150 miles.

(4) During the Alpine orogeny the continental shields started to move together and the African shield underthrust the pile (and also the Eurasian shield elsewhere) raising it above sea level as an undeformed thrust block. Intrusions of serpentine or granite occurred during this period, but the main vulcanicity ceased as the pile was separated from its roots.

(5) The underthrusting by the African shield beneath the pile depressed the mantle surface by an amount greater than that necessary for isostatic compensation of the excess load of the pile. Isostatic uplift has therefore continued since the cessation of the Alpine orogeny and has resulted in the denudation of the pile almost to its roots. The Troodos massif now represents the remnants of the pile.

The evidence supporting this hypothesis, which has already been discussed in detail, is summarized below.

Space-form. The gravity evidence shows that Cyprus is underlain at a shallow depth by an extensive layer of high-density material. The thickness of this layer is at least 7 miles and its density at least 3·3 g/cm^3. Evidence of regional isostatic equilibrium shows that this high-density layer is separated from the mantle proper by a layer of lower density. The upper layer is identified as ultrabasic by its density and petrography; on its density, the intermediate layer is probably sialic. These results are quite definite. The gravity field also gives an approximate margin for the ultrabasic layer, showing that it underlies the whole of Cyprus with small extensions under the sea. The space-form of the ultrabasic layer is clearly not batholithic in the sense that it occupies the whole thickness of the crust, although the connexion to the mantle beneath Pomos might suggest a laccolith form. The shape of the ultrabasic layer is quite consistent with its emplacement in the solid state as a thrust block but does not preclude its emplacement as a laccolith by forcible intrusion of molten or plastic magma. Evidence against this latter alternative depends on structural and petrological considerations.

Structure. It is evident that east–west tensional stress was dominant throughout the evolution of the massif. This is manifest in the north–south structure of the Sheeted Intrusive Complex a dyke swarm of unusual density and regularity, the north–south banding in the Troodos Plutonic Complex and the fact that the dykes in the Pillow Lava Series are generally alined about a north–south direction. Furthermore, the elongated north–south anomalies at Polis, Nicosia and Lefkoniko could represent tension structures now infilled with low-density material.

Although east–west tensional stress during the evolution of the massif is well substantiated no direct evidence is available regarding the earth movements causing this stress. It has already been noted that north–south compression was the dominant stress in the eastern Mediterranean during the Alpine orogeny and that during this period the continental masses of Africa and Eurasia moved towards each other. It seems possible that such a movement was in progress during the evolution of the massif and that the north–south compression could have resulted in a complimentary east–west tension. This mechanism is only in the nature of a speculation and it may be that the stresses operative during the evolution of the massif are in no way related to an Alpine stress pattern.

It should be emphasized that there is no evidence that the Troodos massif has been deformed since it evolved. Furthermore, there are no radial or peripheral structures which might have been expected if the Troodos Plutonic Complex magma had been forcibly injected as a molten or plastic magma from a central feeder.

Petrology. Petrological evidence leaves little doubt that the high-density layer, causing the main gravity anomaly, can be correlated with the Troodos Plutonic Complex. Rock types of this complex range from dunites and peridotites through olivine-gabbros to gabbros and granophyres. On density evidence it would seem that the high-density layer is dominantly of peridotitic composition. It is suggested that the roughly stratiform Troodos Plutonic Complex was formed by the differentiation, *in situ*, of peridotitic material. The petrogenesis of the Sheeted Intrusive Complex is masked by subsequent alteration but the Troodos Pillow Lava Series would seem to have originated by the selective fusion of peridotitic material similar to that of the Troodos Plutonic Complex.

5·1. *Emplacement of the massif*

Only two modes of emplacement of the Troodos massif seem possible; either peridotitic material was injected along a profound fracture in the sialic crust, or the crust moved into its present position after the evolution of the massif.

Emplacement by injection. Considering the first alternative, that the Troodos igneous material was injected through a fracture in the sialic crust, numerous difficulties are encountered. The emplacement of the magma would have had to occur during pre-Triassic times when such evidence as is available suggests that the African and Eurasian shields were more widely separated than at present. It would also be difficult to explain why the Troodos igneous suite is limited to Cyprus when a profound fracture of this nature would undoubtedly extend east and west far beyond the Cyprus area. The fact that only north–south fracturing occurred during the evolution of the massif cannot readily be explained by emplacement along an east–west fracture of the type envisaged. If the suggestion that the Troodos magma was derived from the mantle is correct, we must conclude that it travelled through 20 miles or more of sialic crust. Such a concept is difficult to envisage especially as there are no petrological signs that the sialic crust has been assimilated by the invading ultrabasic material. Finally, other than the very occasional layers of manganiferous shales, there are no sediments within the Troodos massif, and all lava flows exhibit well-formed pillow structures; both these facts would suggest a subaqueous environment throughout the evolution of the massif.

So, whilst meeting the geophysical demand for sialic crust under Cyprus, the injection of the Troodos material, in either fluid or semi-solid state, through a profound fracture in the earth's sialic crust would not appear to satisfy the condition imposed by the known facts. The alternative hypothesis is that the sialic material now under Cyprus moved into its present position subsequent to the evolution of the Troodos massif.

Emplacement by thrusting. Argand (1922) and Blumenthal (1941) indicate that during the Alpine orogeny the Eurasian hinterland was moving south in relation to the African shield and that during this movement the African shield underthrust the Eurasian massif in places. It seems likely therefore that prior to the Alpine movements the non-continental area between Africa and Eurasia, especially in the eastern Mediterranean, was wider than at the present time. Indeed, it is suggested that there existed in this area between the continental masses an oceanic region where the sialic crust was not present.

Bishopp (1952*b*) recognizing that the Troodos massif might represent an immense volcanic pile which developed in an oceanic environment, states:

'It has occurred to the writer and has been suggested in discussion that there is a resemblance between the basic igneous province which seems to have been a feature of the eastern Mediterranean region in pre-Tertiary and perhaps pre-Cretaceous time, and certain regions in the Pacific which are largely occupied by sima without an appreciable sialic cover. If so, the Cypriot province may represent an early uplift of the sima in the gap between the sialic blocks of Eurasia and Indo-Africa which now converge upon it.'

The writers are in general agreement with Bishopp for they postulate that if Troodos evolved as a volcanic pile in an oceanic setting between the continental masses of Africa and Eurasia this would explain the major features of the massif. It is proposed that the Troodos massif evolved in pre-Triassic times as a volcanic pile in an oceanic environment. The fact that the rocks of the Troodos igneous suite are limited to the Cyprus area could be explained as indicating the size of the original volcanic area. Furthermore, the east–west tensional stress dominant throughout the evolution of the massif would allow complete or partial fusion of the upper mantle in the Cyprus area and thus provide the basic to ultrabasic material of which the massif is composed. As the mantle would be relatively near the surface in an oceanic area of the type envisaged, the ultrabasic nature of the parent magma might be expected. Finally, an oceanic setting would account for the pillowed nature of the extrusives and also for the rarity of sedimentary deposits.

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Having postulated that Troodos evolved as a volcanic pile in an oceanic setting, a mechanism must now be sought to explain the presence of sialic crust under Cyprus. Argand (1922) and Blumenthal (1941) provide evidence that during the Alpine orogeny the two continental masses of Africa and Eurasia converged one upon the other to such an extent that the African foreland underthrust the Eurasian hinterland. It is suggested that during these earth movements the African foreland also underthrust the Troodos massif. Stages in this movement are diagrammatically represented in figure 14.

J. C. Harrison, although not accepting Bishopp's concept that Troodos may represent an oceanic volcanic mass (discussion following Bishopp 1952*b*) does, in a later paper (Harrison 1955), detail geophysical evidence why sialic crust must exist under Cyprus. In the same paper, Harrison, in explaining the arcuate zone of negative gravity anomaly in the eastern Mediterranean, suggests that the sialic crust fractured in an east–west direction, that this fracture plane was inclined to the north at 45°, the southern section of the crust being pushed under the northern section (Harrison 1955, figure 15, p. 305). The mechanism proposed by Harrison is essentially similar to that envisaged by the authors with the exception that we would postulate movement of two separate masses rather than the fracturing of a sialic crust of uniform thickness.

Uplift. The underthrusting of continental Eurasia by the African shield could result in crustal thickening in the zone of underthrusting. This would explain the regional negative gravity anomaly in the eastern Mediterranean and also why eustatic uplift is dominant in the area as a result of isostatic adjustment. The emplacement of a mass of low-density material under Troodos, as indicated by the Olympus anomaly, may have further disturbed the isostatic balance and initiated differential uplift centred on Troodos. Factual evidence of the differential uplift of Troodos is provided by de Vaumas (1960) who has shown that the summit of Troodos has been elevated by some 10000 ft. since Cretaceous times whilst the rest of the island has risen only some 2000 ft.

Proceeding *pari passu* with the elevation would be resultant erosion, most intense where elevation was the greatest. It is not surprising therefore that the plutonic rocks are to be found cropping out at the summit of Troodos immediately over the area where a low-density mass is situated.

Resemblance to volcanic pile. It is suggested therefore that the Troodos massif with a core of layered ultramafic plutonic rocks surrounded by dyke swarm of remarkable density and unconformably overlain by basic pillow lavas, represents a pre-Triassic volcanic pile in which exceptional uplift and resultant deep erosion have revealed the plutonic core and the deeper part of the volcanic mass where intrusive material predominates markedly over extrusive rocks. The extrusive pillow lavas, which once possibly covered the whole of the mass, are now preserved as a relatively thin perifery where uplift and erosion have not been so intense.

Finally, it is tentatively suggested that the exposed rocks of the Troodos massif might indicate the structure and nature of other oceanic volcanic masses and that the exposed rocks of the Troodos Plutonic Complex might in fact represent the sub-Mohorovicic peridotitic material which has been partially fused to provide the volcanic material of the Sheeted Intrusive Complex and the Pillow Lava Series. It is also thought that whilst this sub-Mohorovicic material was either partially or completely fused it was mobile enough to

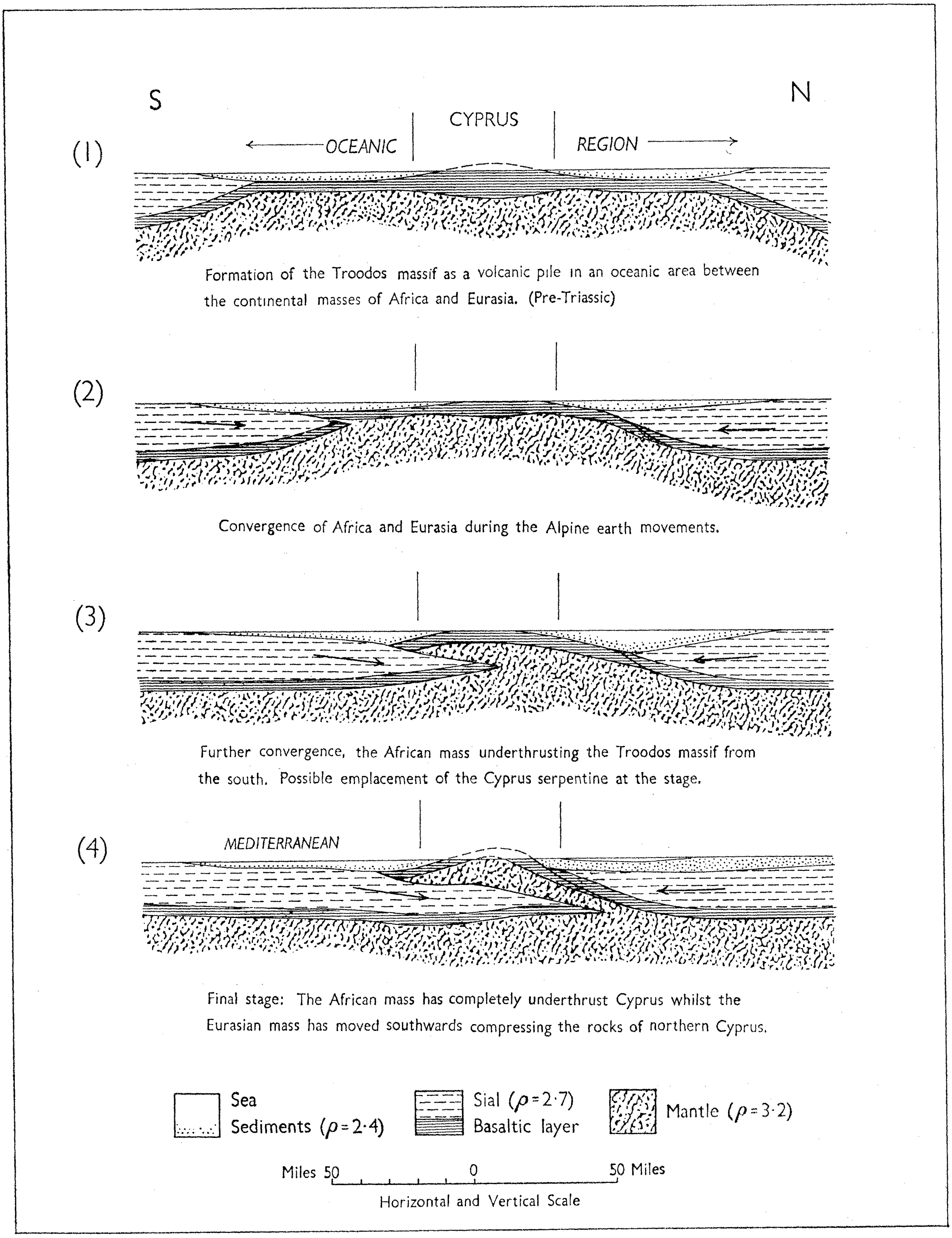

FIGURE 14. Diagrammatic representation of stages in the underthrusting of Cyprus by the African shield.

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allow differentiation into a roughly stratiform complex ranging in composition from dunites and peridotites to overlying gabbros and granophyres.

We are indebted to several Cyprus Government Departments for their assistance with the gravity survey in March to June, 1958. The Water Development Department provided transport for the survey; Survey Department staff on several occasions accompanied the gravity party to locate pendulum and trigonometric stations; the Meteorological Department provided data for the reduction of altimeter readings; the Geological Survey carried out preliminary work on the gravity maps.

In London, the gravity data was reduced and compiled by members of the Overseas Geological Surveys (O.G.S.) staff. In particular, we would like to thank Mr L. Z. Makowiecki of O.G.S. for his help with the collection and reduction of field data.

A large part of eastern Cyprus was surveyed for the Iraq Petroleum Co. in 1948; we are grateful to Iraq Petroleum Company for allowing us to incorporate their gravity map with our own.

We also wish to acknowledge our gratitude to many persons who have given us assistance in the preparation of this paper; noteable amongst these are: Professor W. Q. Kennedy, F.R.S., Drs M. H. P. Bott, T. N. Clifford, P. G. Harris, and former members of the Cyprus Geological Survey whose published works have been consulted extensively.

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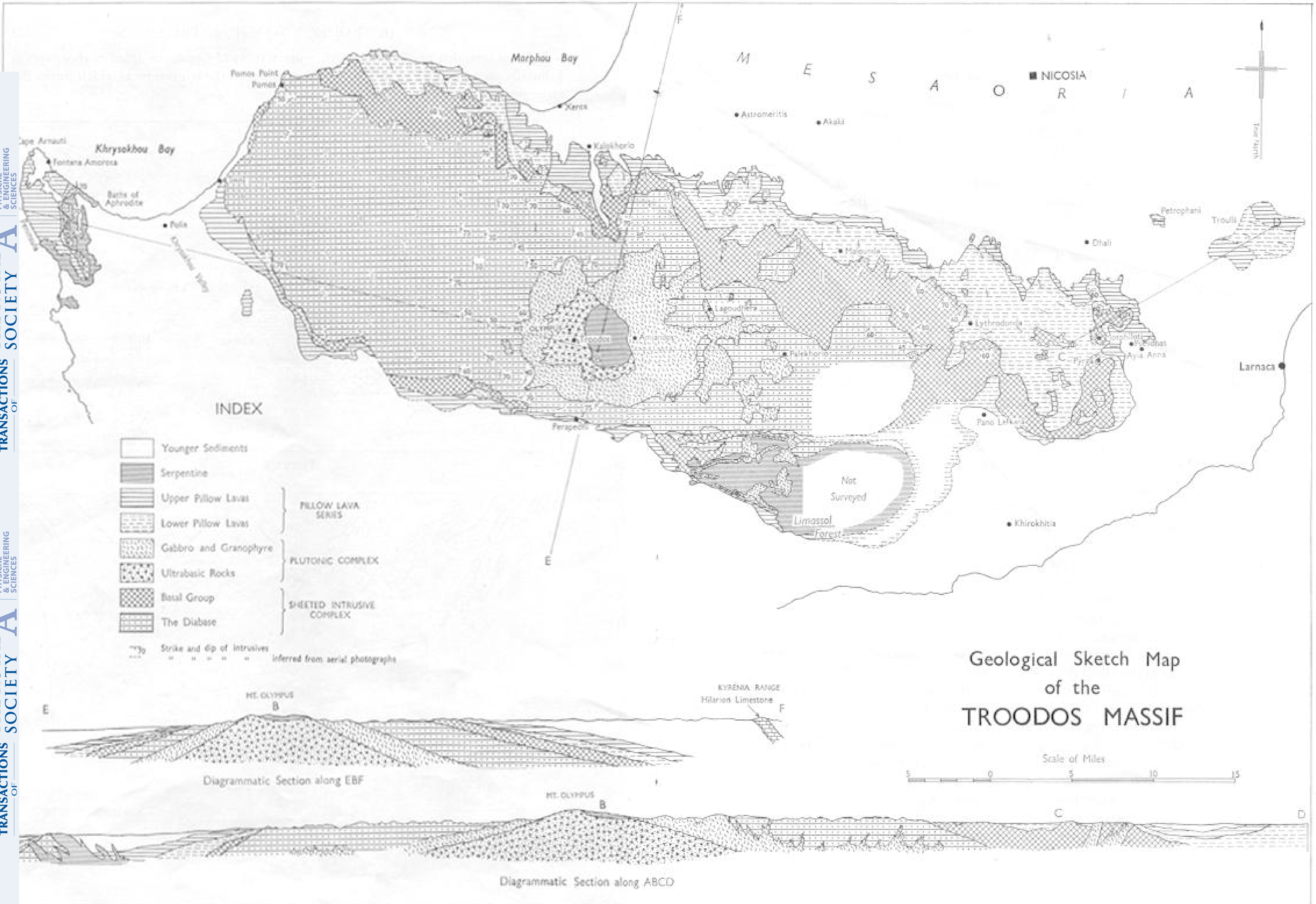

Figure 3

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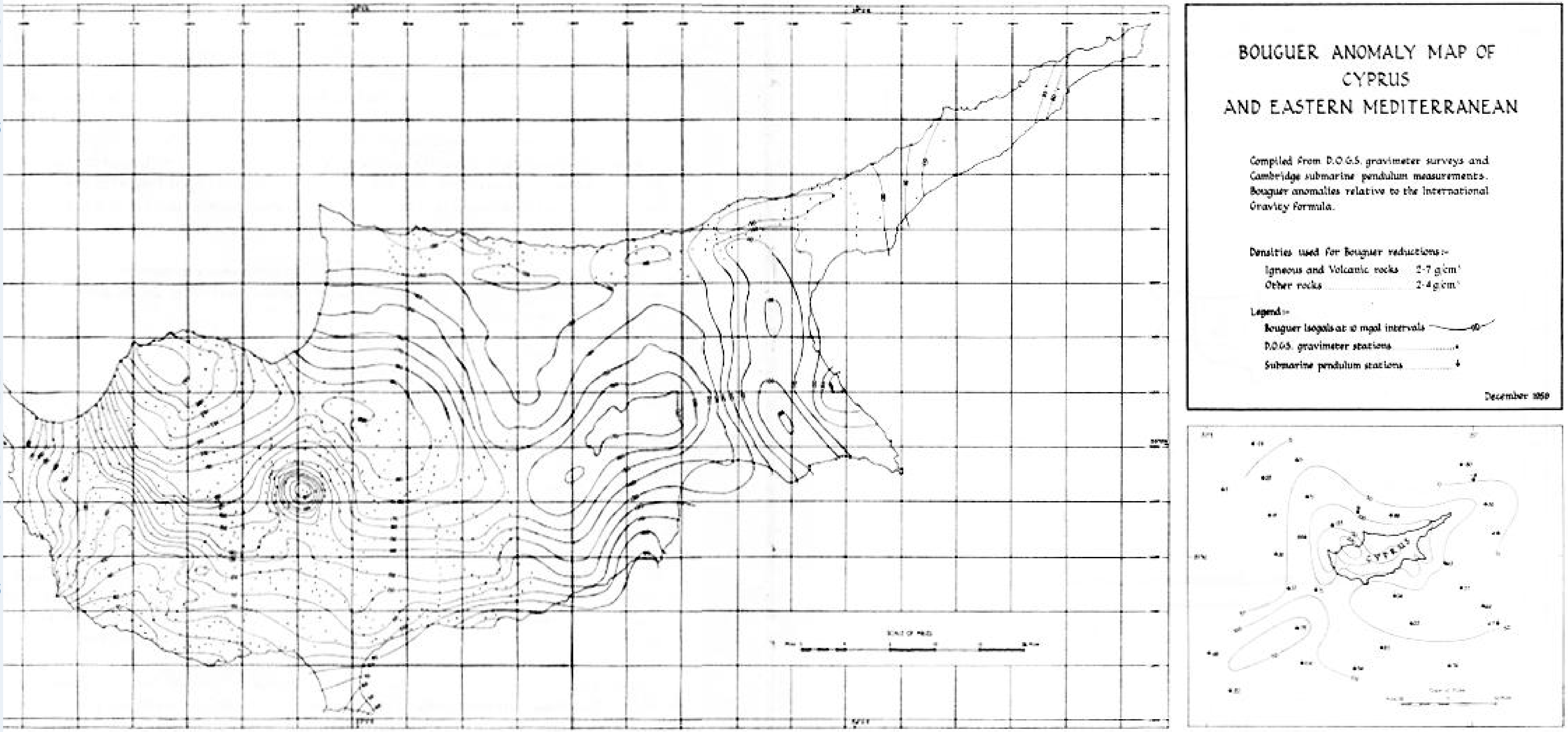
BOUGUER ANOMALY MAP OF
CYPRUS
AND EASTERN MEDITERRANEAN
Compiled from D.O.G.S. gravimeter surveys and Cambridge submarine pendulum measurements. Bouguer anomalies relative to the International Gravity formula.
Densities used for Bouguer reductions:-
Igneous and Volcanic rocks 2·7 g/cm³
Other rocks 2·4 g/cm³
Legend:-
Bouguer Isogals at 10 mgal intervals
D.O.G.S. gravimeter stations
Submarine pendulum stations
December
CYPRUS

Figure 8