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EVALUATION AND TARGETING OF GEOTHERMAL ENERGY RESOURCES IN THE SOUTHEASTERN UNITED STATES

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Final Report

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TABLE OF CONTENTS

Orogenic Studies Laboratory

Regional Geophysics Laboratory

page

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(See Note, page ix)

October 1, 1977 - June 30, 1982

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Note: Only Drs. J.K. Costain, L. Glover, III, and J.A. Speer provided continuous service to this program since its inception. For example, most of the laboratory technicians listed on p. viii were associated with the VIBROSEIS program which did not begin until April 1979. Periods of employment of administrative assistants, bookkeepers, and many other personnel did not overlap. Others were employed on a part-time basis.

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SUMMARY AND OVERVIEW

J. K. Costain and L. Glover, III

The objectives of the geothermal program at VPI&SU have been to develop and apply geological and geophysical targeting procedures for the discovery of low-temperature geothermal resources related to heat-producing granite. The different geologic setting associated with the warm springs in Virginia was discussed by Perry et al. (1979) and is not reviewed in this report.

Optimum sites for geothermal development in the tectonically stable eastern United States are associated with areas of high heat flow derived from crustal igneous rocks containing relatively high concentrations of uranium (U), thorium (Th), and potassium (K). Exploration strategy has been directed toward confirmation of the "radiogenic model" (Costain et al., 1980a). In this model, granite plutons with relatively high concentrations of U, Th, and K relative to the surrounding country rock occur in the crystalline basement concealed beneath the wedge of relatively unconsolidated sediments of the Atlantic Coastal Plain. Such granites are sources of heat produced by the natural radioactive decay of isotopes of U, Th, and K (Birch, 1954). Because the thermal conductivity of the Coastal Plain sediments is roughly half that of crystalline rocks (Ziagos et al., 1976; Perry, 1979), the sediments act as a thermal insulator, and raise geothermal gradients within the sediments. In a sedimentary section above a heat-producing granite, isotherms are warped upward, and anomalously higher temperatures occur at shallower depths.

The storage of commercially exploitable geothermal heat at accessible depths (1-3 km) also requires favorable reservoir conditions in sedimentary rocks overlying heat-producing granite. The geothermal resource is primarily a hydrothermal resource. The yield of water from deep aquifers is therefore an important consideration because water is the medium by which the heat at depth is transferred to the surface. The yield must be sufficient to meet energy requirements. This is partly an economic problem, but it also involves the intrinsic permeability and transmissivity of the aquifers. There is little data concerning the temperature and hydrology of deep aquifers in the sediments beneath the Atlantic Coastal Plain. Good porosities were indicated from a test hole near Norfolk, Va., in the thick Cretaceous sand unit (Brown, 1968) that forms the basal sedimentary unit throughout much of the northern Atlantic Coastal Plain (Brown et al., 1972). Comparison of geophysical logs from deep wells in the Coastal Plain of Maryland, Virginia, and North Carolina suggest that deep aquifers are widespread (Lambiase et al., 1980).

No economic factors related to the use of eastern geothermal resources were considered at VPI&SU; however, our program was closely coordinated with the work sponsored by the Department of Energy at the Applied Physics Laboratory of Johns Hopkins University where economic and engineering factors related to the development of an eastern geothermal resource were being considered (Johns Hopkins University, 1981).

In order to locate optimum sites systematically, a methodology employing geological and geophysical methods of investigation was developed and applied. Heat-producing granites similar to those in the basement beneath the Atlantic Coastal Plain are exposed to the northwest in the Piedmont province. The petrology, radioelement distribution, and bulk chemistry of several of these young (254-420 Ma.) granites was studied in detail (Speer et al., 1980 and this report). These late Paleozoic granitoids in the central and southern Appalachians are the youngest known to date, and are generally coarse-grained granites with alkali feldspar megacrysts. Field relations, mineralogy, chemistry, and geochronology have been summarized by Fullagar and Butler (1979) and Speer et al. (1979). A finer-grained muscovite-bearing phase considerably higher in heat production than the coarse-grained phase commonly intrudes the coarse-grained facies and is often found near the center of the plutons. Because the source of excess heat flow above the background flux is primarily from the radioactive decay of isotopes of U and Th, an understanding of the horizontal and vertical distribution of these heat-producing elements in granites is important for efficient targeting of sites of high heat flow. A summary of the relationship between rock type and heat production is given in this report by Speer and others. The horizontal distribution can be examined from surface sampling; however, the distribution of U and Th to crustal depths can only be inferred from heat flow-heat generation data. Uranium mobility in igenous rocks is sensitive to subsolidus processes, and this problem received high priority. The migration of uranium in granitoid rocks and the petrography of the U-bearing minerals of the Liberty Hill pluton exposed in South Carolina was reported by Speer and others (1981).

A linear relation, $q = q^* + DA$ between surface heat flow, q, and surface heat generation, A, was confirmed for many heat flow sites in granites exposed in the Piedmont of

Virginia, North Carolina, South Carolina, and Georgia. This is a significant result of the program because from this relationship a crustal temperature profile can be computed (Lachenbruch, 1968). The drill sites defining the linear relationship were located along a region parallel to the major structural trend of the Appalachians. Although the interpretation of the physical significance of the parameter D is ambiguous (Birch et al., 1968; Lachenbruch, 1968), limited seismic data (Clark et al., 1978; Cook et al., 1978) suggest that D could represent thickness of pluton (Costain and Glover, 1980). Values of (q,A) that do not lie on an established relationship may therefore represent granites of different thickness beneath which different crustal temperature profiles would presumably be different. Important exceptions to this generalization will be found, however, if U and Th have been mobilized from a relatively uniform distribution within a granite body to a "rind" or thin peripheral zone surrounding the granite. A heat flow determination directly over such a granite may be approximately the same as that over a similar volume of granite with the same total U and Th content uniformly distributed (Green's theorem), but the heat generation value from the same site will not be consistent with an interpretation of D as thickness of pluton. This appears to be true for the Castalia pluton in North Carolina. The gamma log obtained from the Castalia heat flow site as well as determinations of U, Th, and K concentrations in core samples indicated a remarkably uniform distribution of U and Th, quite unlike any of the other results from heat flow sites in granite.

An understanding of the linear relation between heat flow and heat generation remains an important unsolved problem wherever it has been observed on continents, and has important implications for predicting crustal temperatures in areas of potential hot-dry-rock applications. For further comments, see "Heatflow in the Southeast United States", elsewhere in this report.

Following detailed study of selected heat-producing granites exposed in the Piedmont, emphasis shifted to granitoids concealed beneath the thermally insulating sediments of the Atlantic Coastal Plain. Thermal conductivity varies with sediment type. Abrupt lateral and vertical facies changes occur within the Coastal Plain sediments (Brown et al., 1972) making prediction of lithology, thermal conductivity, and temperature at the surface of crystalline basement difficult. Generally, the sediments in the deeper sections of the northern Coastal Plain are non-marine deposits which are more quartz-sand rich than the upper 300 m; this tends to increase thermal conductivity in the deeper part of the section relative to the upper part and thus lowers the geothermal gradient by as much as 15 percent. Thermal conductivity also increases with compaction. Granites beneath such sediments must be located indirectly by means of geophysical data including gravity, heat flow, and reflection seismology. Several granite bodies targeted by VPI&SU using geophysical data were later confirmed by the drill (Chesapeake and Petersburg, Va.; Dort, N.C.; Salley, S.C.). Others remain unconfirmed by the drill (Smith Point, Va.; Crisfield, Md.; Lumberton, N.C.).

A major program to determine heat flow on the Atlantic Coastal Plain was undertaken during 1978 and 1979. About 50 exploratory holes, each approximately 300 m in depth, were drilled in New Jersey, Delaware, Maryland, Virginia and North Carolina under D.O.E. contract to Gruy Federal, Inc. The heat flow sites were selected by VPI&SU to represent areas with a high potential for geothermal development and to test the radiogenic model. Analysis of geothermal gradients determined by VPI&SU in these exploratory holes in the Atlantic Coastal Plain indicate that in many areas temperatures exceed 40°C at the base of the sediments. Water of such temperature, provided it is available in sufficient quantity, represents a viable geothermal resource. In the eastern United States, geothermal gradients were found to be in the range of 10 to 50°/km.

Results of the gradients on the northern Atlantic Coastal Plain have been reported by Lambiase et al., (1980) and are not repeated herein. The geothermal gradients in both the Piedmont and in the northern Atlantic Coastal Plain are tabulatedd by Costain and others in this report (Heat Flow in the Southeastern United States). Higher gradients are found throughout most of the Delmarva Peninsula where values are between 30°C/km and 46°C/km. Near Chesapeake Bay, temperature gradients are the highest of any in the five-state area with values that vary between 39°C/km and 50°C/km. The gradient of 50°C/km is at Smith Point on the western flank of a large negative gravity anomaly beneath Chesapeake Bay. The basement rock has not been sampled by the drill, but it seem likely that the anomaly is associated with unmetamorphosed granite. A gradient of 43°C/km was determined in northeastern North Carolina; all other known exploratory holes in this area have thermal gradients between 31°C/km and 35°C/km. In southeastern North Carolina, gradients are lower, and vary from 22°C/km to 32°C/km.

One promising area for geothermal development discovered by us in the northern Atlantic Coastal Plain is on the Eastern Shore between Crisfield in southern Maryland and Oak Hall in northern Virginia just east of the Chesapeake Bay gravity anomaly. Results from the exploratory heat flow determinations were used to site a 1690 m test hole that was drilled into basement and completed in June 1979 at Crisfield, Md. The Crisfield site was selected by VPI&SU because of the known high geothermal gradients there and the moderate depth to basement. Temperature at the top of crystalline basement was found to be approximately 58°C. The temperature predicted by us at the base of the Coastal Plain sediments at Crisfield was about 16 percent less than the actual temperature. The difference was due entirely to the uncertainty in estimating the higher thermal conductivity of Coastal Plain sediments in the lower 78 percent of the sedimentary sequence.

The basement rock encountered at Crisfield was not granite, but a low grade metavolcanic rock possibly equivalent to volcanics exposed in the Carolina Slate Belt. The rocks have not yet been dated. The high heat flow at Crisfield and the negative gravity anomaly just to the west beneath Chesapeake Bay suggest that granite could be the source of the heat flow anomaly. The first clue that this is a reasonable interpretation came from a VIBROSEIS survey sponsored by the U.S. Geological Survey (Contract No.14-08-0001-G-685) near Lumberton, N.C. (Pratt, 1983). Glover (1979) observed that exposed Alleghanian granitoids do not intrude very far into the low-grade metavolcanics of the Carolina Slate Belt. Core drilling, gravity, and VIBROSEIS data (Pratt et al., 1983) obtained near Lumberton for the U.S.G.S. suggest the presence of an unmetamorphosed(?) granitoid concealed beneath a belt of lower amphibolite facies volcanic rocks which comprise part of the crystalline basement under sediments of the Atlantic Coastal Plain. A 239-m core consisting of interlayered felsic and mafic metavolcanic rocks was retrieved from beneath 150 m of sediments at a site near the center of a -35 mgal Bouguer anomaly near Lumberton. Excellent seismic reflections along a nearby VIBROSEIS line shot by VPI&SU extending 16 km southeast from the Lumberton drill site are interpreted to be from these volcanics, which are 3 to 5 km in thickness. A zone transparent to seismic waves from 1.2 to 5 sec on the seismic line is believed to be associated with a large granite pluton. The heat flow is high, about 63 mW/m^2. The high heat flow and negative gravity anomaly are both characteristic of Alleghanian plutons in the southeastern U.S. At Crisfield, east of the Chesapeake Bay gravity anomaly, a pronounced negative Bouguer anomaly is not observed; however, this is not a severe constraint because density contrasts in the basement and the percentage of mafic material in the volcanic assemblage at depth are unknown. The source of the high heat flow at Crisfield is interpreted to be from unmetamorphosed granite of late Paleozoic age beneath low grade volcanic rocks.

Limited hydrologic and heat flow data made available by this program made it possible to estimate the thermal lifetime of a geothermal resource within the sediments beneath the Atlantic Coastal Plain. Using the integrated finite difference method (program CCC developed at Berkeley), numerical modeling was done by Laczniak (1980) at VPI&SU to predict the optimum spacing between a pumping well and an injection well (a single dipole). Laczniak modeled the response of a leaky doublet system to a single dipole using aquifer parameters derived from the Crisfield tests. The model was run for a simulated period of 15 years or until steady-state thermal and fluid flow was reached. A doublet with direct injection back into the reservoir was shown to be a feasible method to extract heat from the low-temperature, liquid-dominated geothermal systems of the Atlantic Coastal Plain.

Important conclusions of Laczniak's study were:

a) direct injection back into the reservoir may be necessary to maintain sufficient fluid pressure at the production well for systems with a low permeability,

b) temperature distribution within the system is only slightly affected by changes in permeability in the range 10-100 md (millidarcies),

c) resting the system for periods of six months does not result in a significant recharge of heat at the production well,

d) a doublet system with thermal and hydrologic conditions similar to those encountered at Crisfield, Maryland, a well spacing of 1000 m, a permeability of 100 md, and a pumping-injection stress of 500 gpm (injection temperature 44°C) could produce 5.5 million Btu's per hour over a period greater than 15 years.

Preliminary results of the test at Crisfield, Maryland were encouraging in that an adequate supply of water at a temperature of 56°C was produced from aquifers at a depth of 1.2 km, but further testing of the temperatures and transmissivity of the deep aquifers beneath the Coastal Plain will be required before the geothermal resource potential can be fully evaluated.

Acknowledgements

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OROGENIC STUDIES LABORATORY

PETROLOGY OF SYN- AND POSTMETAMORPHIC GRANITES IN SOUTHEAST U.S.: BASIS FOR UNDERSTANDING HEAT FLOW AND HEAT PRODUCTION

J. A. Speer, S. W. Becker, S. S. Farrar

Of 82 granitoid plutons examined by the Orogenic Studies Laboratory between 1976 and 1981, 56 are known in sufficient detail to classify them on the basis of petrography and field relations. These differences between the granite types can be extended to their heat productions, modes, major and trace element rock chemistries, mineral chemistries and isotopic compositions. The petrographic distinctions can also be applied to variations in heat flow if other factors such as pluton geometry, size, and lower crustal contribution to the heat flow are similar. The plutons and their locations are presented in Figure 1.

Our work has centered on the plutons' potential as low-temperature, geothermal heat sources. Because metamorphosed rocks tend to lose radioactive elements, we examined mostly syn- and post-metamorphic granitoid plutons which range in age from 420 to 254 Ma. The ages of metamorphism and deformation tend to decrease to the southeast (Glover and Russell, 1981; Glover *et al.*, submitted). Therefore the ages of syn-and post-deformational plutons vary according to the local age of metamorphism and deformation. Figure 2 is a map from Glover *et al.* (submitted) illustrating the regional variations in ages of ductile deformation and metamorphism.

Petrography

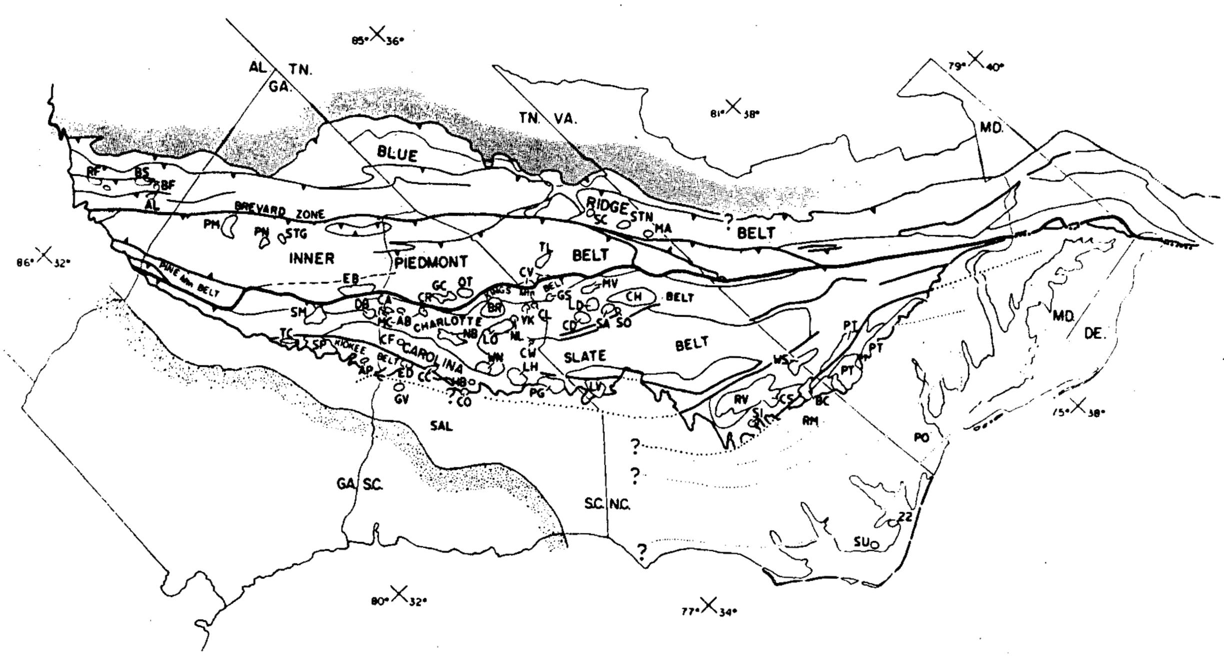
The several petrographic types of granites are:

1) coarse-grained, amphibole ± pyroxene + biotite granites with accessory allanite, titanite, ilmenite and magnetite, comprising one facies of composite plutons.

PT	Petersburg, Va.	BC	Butterwood Creek, N.C.
CS	Castalia, N.C.	RM	Rocky Mt., N.C.
LD	Landis, N.C.	BR	Bald Rock, S.C.
LH	Liberty Hill, S.C.	HB	Harbison, S.C.
PG	Pageland, S.C.	WN	Winnsboro, S.C.
GV	Vaucluse, S.C.		

Figure 1: Locations and names of the plutons discussed in this report.

(VA.)	PO	Portsmouth		
	PT	Petersburg		
(N.C.)	WS	Wise	CD	Concord
	BC	Butterwood Creek	GS	Gastonia
	RM	Rocky Mt.	CV	Cherryville
	CS	Castalia	TL	Toluca
	SI	Sims	MA	Mt. Airy
	RV	Rolesville	STN	Stone Mtn.
	CH	Churchland	SC	Spruce Pine
	SO	Southmont	LV	Lilesville
	SA	Salisbury	SU	Stumpy Point
	LD	Landis	22	NC 22
	MV	Mooresville		
(S.C.)	AB	Abbeville	HB	Harbison
	BR	Bald Rock	LH	Liberty Hill
	CA	Calhoun Falls	LO	Lowrys
	CW	Catawba	MC	Mt. Carmel
	CC	Clouds Creek	NB	Newberry
	CL	Clover	NL	North of Lowrys
	CO	Columbia	OT	Ott
	CR	Coronaca	PG	Pageland
	CF	Cuffytown Creek	SAL	Springfield
	ED	Edgefield	WN	Winnsboro
	GV	Graniteville-Vaucluse	YK	York
	GC	Grey Court		
(GA.)	AP	Appling	PN	Panola
	DB	Danberg	SM	Siloam
	EB	Elberton	STG	Stone Mtn.
	PM	Palmetto	TC	Town Creek
(AL.)	AL	Almond	BS	Bluff Springs
	BF	Blakes Ferry	RF	Rockford


AL
TN.
GA.
TN
VA.
MD.
MD.
DE.
GA/SC.
SC NC.
BLUE
RIDGE
BELT
BREVARD ZONE
INNER
PIEDMONT
BELT
CHARLOTTE
BELT
CAROLINA
SLATE
BELT
SAL
?
?
?
?

2) coarse-grained, biotite granite with accessory allanite, titanite, ilmenite and magnetite. There are two associations:

A) comprising the entire pluton

AP	Appling, Ga.	PM	Palmetto, Ga.
TC	Town Creek, Ga.	SM	Siloam, Ga.
SP	Sparta, Ga.	DB	Danberg, Ga.
CO	Columbia, S.C.	GC	Grey Court, S.C.
CL	Clover, S.C.	OT	Ott, S.C.
SAL	Springfield, S.C.	YK	York, S.C.
GS	Gastonia, N.C.	LV	Lilesville, N.C.
MV	Mooresville, N.C.		

B) comprising one facies of a composite pluton

PT	Petersburg, Va.	BC	Butterwood Creek, N.C.
CS	Castalia, N.C.	CH	Churchland, N.C.
LD	Landis, N.C.	BR	Bald Rock, S.C.
CC	Clouds Creek, S.C.	LH	Liberty Hill, S.C.
PG	Pageland, S.C.	WN	Winnsboro, S.C.

3) fine- to medium-grained, biotite granite with accessory ilmenite, allanite, molybdenite, ± REE fluorcarbonates, niobates, etc. Muscovite and epidote are arguably primary igneous. There are two associations:

A) comprising the entire pluton

EB	Elberton, Ga.	NB	Newberry, S.C.

B) comprising one facies of a composite, coarse-grained pluton

BC	Butterwood Creek, N.C.		
CH	Churchland, N.C.	BR	Bald Rock, S.C.
GV	Graniteville, S.C.	LH	Liberty Hill, S.C.
LO	Lowrys, S.C.	WN	Winnsboro, S.C.
SM	Siloam, Ga.		

The above three types of granites are comparable to the I-type granites of Chappell and White (1974). Their emplacement occurred largely during and after the Alleghanian orogeny. Fewer were emplaced during or late in the Acadian orogeny.

4) coarse-grained, cordierite + biotite granites and granodiorites with accessory ilmenite, monazite and tourmaline. There are two associations:

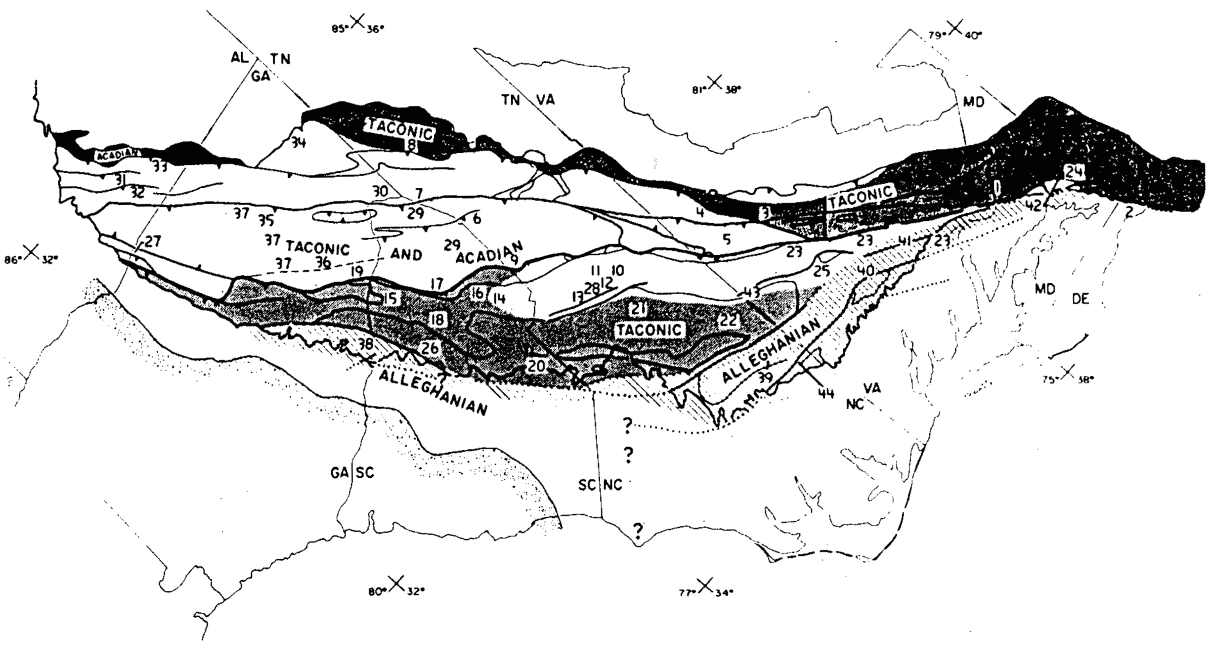


Figure 2: Map illustrating the regional variations in ages of ductile deformation and metamorphism from Glover et al. (submitted). Numbers refer to specific localities discussed in the text of that paper.

A) containing early crystallizing magnetite, which has since reacted away

CC Clouds Creek, S.C.

B) containing early crystallizing almandine, which has since become unstable

SU Stumpy Point, N.C.

5) medium- to coarse-grained, muscovite + biotite granitoid rocks. On the basis of texture, most people would interpret the muscovite as primary igneous. Accessory minerals include garnet, ilmenite, rutile, monazite and xenotime. There are four associations:

A) two-mica, plagioclase-rich granitoids (granodiorites and trondhjemites) containing few accessory minerals and having low heat productions. Garnet is present in those that are syndeformational. Some are associated with Sn and Li mineralization and contain tourmaline.

RF	Rockford, Al.	BF	Blakes Ferry, Al.
BS	Bluff Springs, Al.	AL	Almond, Al.
STG	Stone Mtn, Ga.	PN	Panola, Ga.
STN	Stone Mtn, N.C.	MA	Mt. Airy, N.C.
CV	Cherryville, N.C.	TL	Toluca, N.C.

B) two-mica granites containing modally more accessory minerals and having higher heat productions. Molydenite occurs in some.

WS	Wise, N.C.	SI	Sims, N.C.
22	NC22, N.C.	CW	Catawba, S.C.
ED	Edgefield, S.C.		

C) leucocratic, two-mica granites with accessory garnet, niobates, Li muscovite, fluorite, sulfates and carbonates. They have high Rb/Sr ratios. These are evidently the highly-evolved, fluid-rich roofs of large plutons of type 1 or 2 granites.

CF Cuffytown Creek, S.C. PO Portsmouth, Va.

D) coarse-grained, two-mica granites resulting from extensive interaction of type 1 or 2 granite with the country rock.

LH Liberty Hill, S.C. CS Castalia, S.C.

Two mica granites range in age from Taconic through Acadian to Alleghanian. Types 4B, 5A, and 5B are comparable to the S-granites of Chappell and White (1974) whereas 4A, 5C and 5D are petrologic variants of I-granites. The S-type granites are more concentrated in the Blue Ridge and Inner Piedmont but occur in every geologic belt. Many are clearly associated with major ductile thrust zones that moved at about the time of emplacement of the associated S-granites and therefore may be related to elevated thermal gradients in the crust associated with stacking of hot nappes.

6) gabbro-diorite-syenite complexes

CD	Concord, N.C.	CR	Coronaca, S.C.
MC	Mt. Carmel, S.C.	AB	Abbeville,S.C.
CA	Calhoun Falls, S.C.		

7) alkali feldspar granites

SA	Salisbury, N.C.	SO	Southmont, N.C.
NL	north of Lowrys, S.C.		

The Concord and possibly the Mt. Carmel, Abbeville, and Calhoun Falls appear to be nonorogenic Silurian granites and gabbros. They show alkaline affinities and the mafic members experienced extreme differentiation in a tectonically quiet environment.

Heat Production

Heat production of several hundred granite samples from the southeast, calculated from the U, Th, and K contents, have been obtained by the VPI&SU Regional Geophysics Laboratory. The data are tabulated in the bottom histogram of Figure 3. Heat production values in excess of 15 X 10^{-13} cal/cm^3-sec are from late-stage aplite and pegmatite dikes. These can have heat productions as high as 27 cal/cm^3-sec. The volumetrically important granitic facies have heat productions between 1 and 15 cal/cm^3-sec. U and Th are the most important contributors to the heat production.

Individual histograms for a number of different granite types and plutons are also included in Figure 3. The group with the lowest heat productions are the two-mica granites of the Inner Piedmont and Blue Ridge of Alabama, Georgia, and western North Carolina. These granites lack significant amounts of accessory minerals, minerals most likely to contain U and Th. The Wise pluton, N.C., is also a two-mica granite, but contains a modest amount of accessory minerals. It has a slightly higher average heat production than the other two-mica granites, but is still on the lower end of the range for granites of the southern Appalachians.

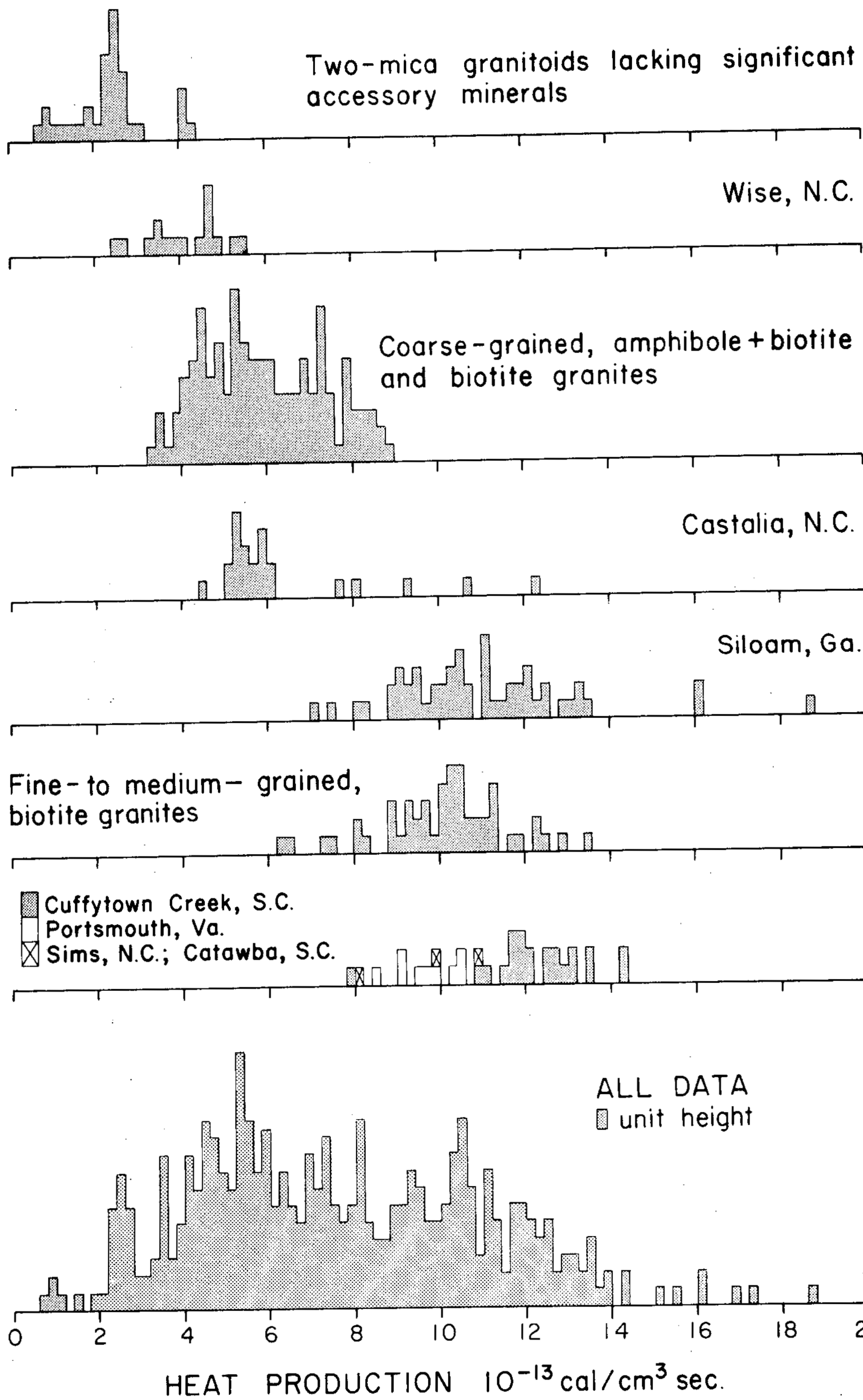


Figure 3: Histograms showing ranges of heat production values for various granite types and plutons in the southern and central Appalachians.

The third histogram is for the coarse-grained, amphibole + biotite and biotite granites, the most abundant granite types in the southern Appalachians. The fourth histogram is for the Castalia pluton, N.C., which is also a coarse-grained, biotite +amphibole granite. The majority of the Castalia heat production measurements are from a deep drillcore in the center of the pluton and have values comparable to other granites of that type. Last year, numerous shallow holes were drilled which afforded a better areal distribution of samples and, except at the center of the pluton, the Castalia generally has a higher heat production. This is believed to be a result of magma-wall rock interaction, a process also reflected in the petrography of these rocks. In a similar manner, the Siloam pluton, Ga., is a coarse-grained, biotite granite with an uniformly higher heat production (fifth histogram, Fig. 3) and is tentatively concluded to have experienced magma-wall rock reactions.

The last two histograms are for rocks that are believed to have crystallized from magmas containing the highest amounts of U and Th. One group comprises the fine- to medium-grained, biotite granites. The other group includes values of four plutons: the evolved two-mica granites with abundant accessory minerals (Cuffytown Creek and Portsmouth) and the two-mica granites containing molybdenite (Sims and Catawba).

Heat Flow

A plot of heat production versus heat flow for drill cores in the granites (Fig. 4) shows a generally linear trend. The implications of the linear relation between heat production and heat flow for the southeastern U.S.A. have been discussed by Costain and Glover (1980). Most of these plutons consist of coarse-grained, amphibole + biotite or biotite granite. They show a reasonably strong, linear trend over the limited range of heat productions exhibited by these granites. The geology and petrology of the plutons that fall away from the linear trend suggest the deviations. The Palmetto granite is thinner than the others, cut off by the Brevard zone fault (Costain and Glover, 1980). Recent work on the Castalia granite shows a range of heat production of 2.1 to 5.1 u W/m^3. The Springfield drillcore encountered some U-mineralization. In both these cases, the heat production of the freshest granite in the drillhole is an underrepresentation of the heat production of that section of the crust whose heat flow is sampled by the drillcore. Based on the new work, an estimate of the heat production of the entire Castalia pluton is 3.0 u W/m^3. This brings the Castalia nearly on the trend in Figure 4. The

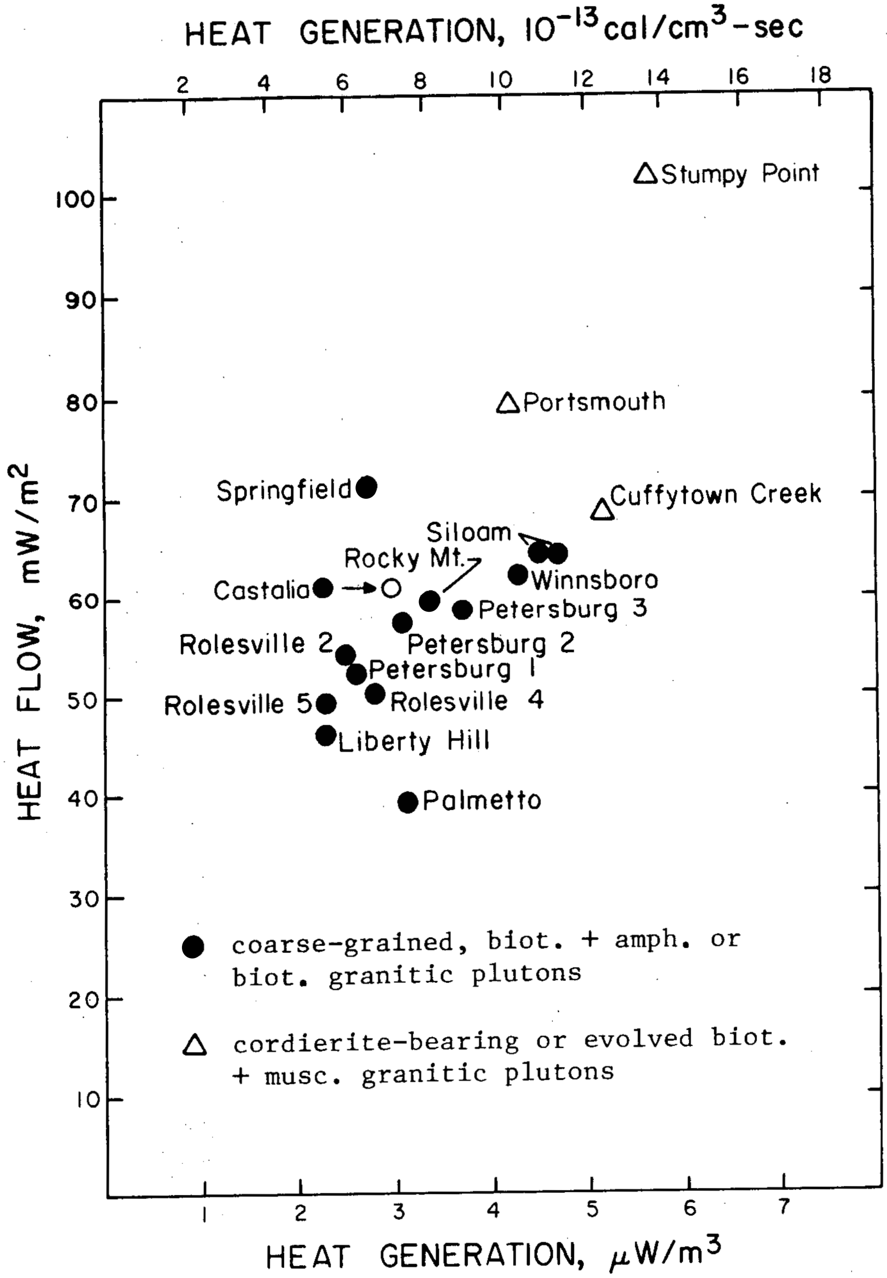


Figure 4: Relationship between heat production and heat flow for the post-metamorphic granite plutons in the southern and central Appalachians. Implications of the generally linear trend are discussed by Costain and Glover (1980). Deviations from this trend are discussed in the text.

heat production may still be underrepresented because of the greater number of samples near the center of this poorly exposed pluton.

The three remaining drillcores, those with the highest heat flows, Cuffytown Creek, Portsmouth, and Stumpy Point, differ from the other granites in being either highly evolved biotite + muscovite or cordierite-bearing granites. Deviation from the linear relation may result from other factors such as varying thickness of granite or differing heat flow provinces, but underestimation of heat production of the crustal section sampled by the heat flow measurement may also be the cause. These particular granites are associated with more abundant magmatic fluids which can be responsible for redistribution of the heat-producing elements.

Modes

The modes of these granite types vary systematically (Fig. 5a). The two-mica granites, which can also contain garnet and cordierite, are the most quartz- and plagioclase-rich with the extreme being the trondhjemitic plutons of Alabama. By contrast, the coarse-grained amphibole + biotite and biotite granitoids are the most K-feldspar rich. As mentioned above, heat production of the granites depends on the amount of accessory minerals, the minerals most likely to contain U and Th.

Chemistry

Major element chemistry of the granitoid rocks has been previously reported by Butler and Ragland (1969) and Fullagar and Butler (1979). Using these authors' average compositions, where the mineralogy of the rocks can be determined, and our unpublished chemistry, a clear distinction is evident between the peraluminous (muscovite- or cordierite-bearing) and metaluminous (amphibole- or biotite-bearing) rocks (Fig. 5b).

Figure 5a: Modal analyses of granitoid rocks of the southern Appalachians.

Figure 5b: (Ca)(Al-(Na+K)) (Fe + Mg) compositional triangle for granitoid rocks of the southern Appalachians.

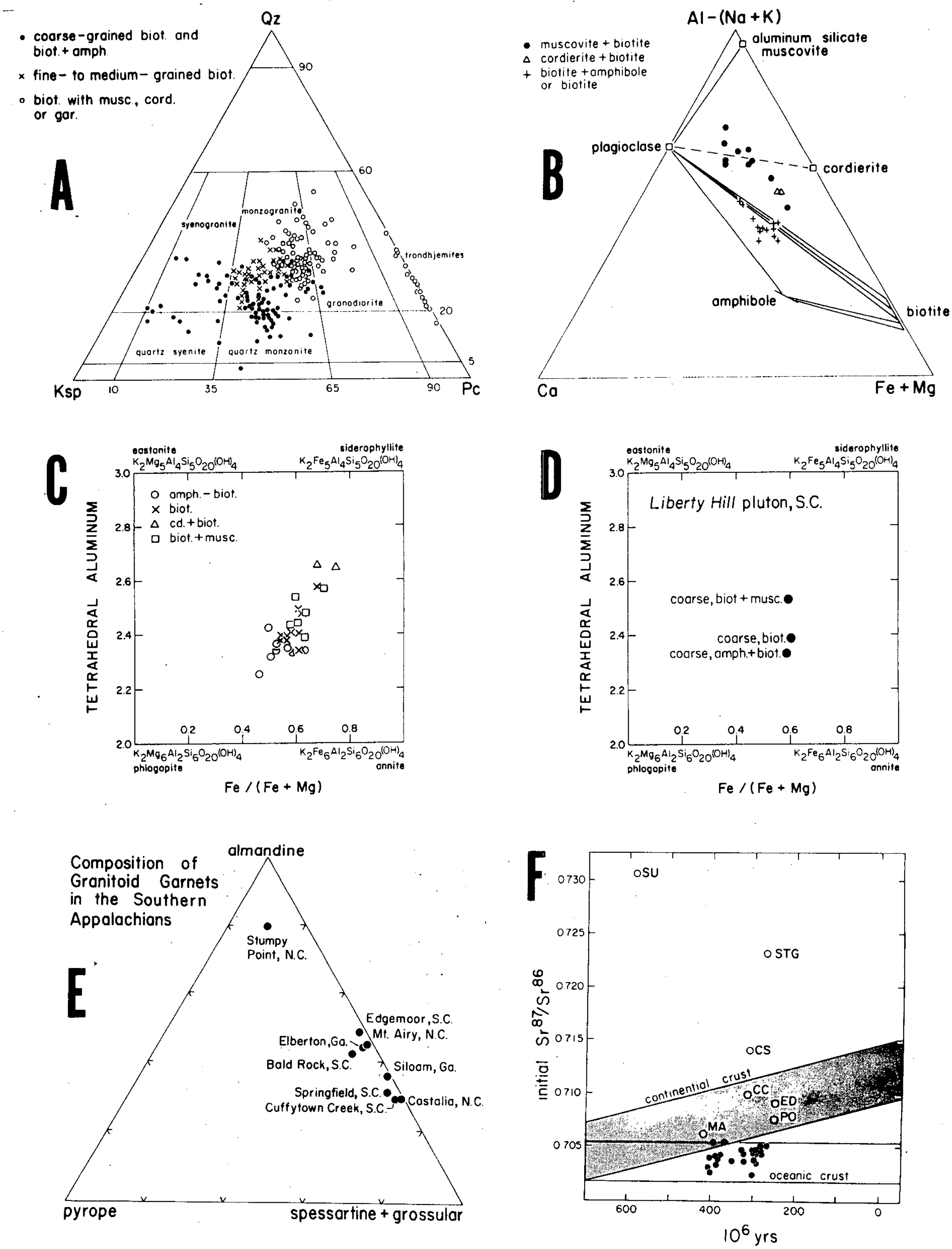
Figure 5c: Average biotite compositions for individual plutons projected onto the phlogopite-annite-eastonite-siderophyllite quadrilateral.

Figure 5d: Average biotite compositions for the various granite types in the Liberty Hill pluton, S. C.

Figure 5e: Average compositions of garnets in granitoid rocks of the southern Appalachians.

Figure 5f: Plot of initial Sr^{87}/Sr^{86} versus age for granitoid plutons of the southern Appalachians. Data is largely from Fullagar and Butler (1979) with additional unpublished OSL data.

STG	Stone Mtn., Ga.	CS	Castalia, N.C.
CC	Clouds Creek, S.C.	ED	Edgefield, S.C.
PO	Portsmouth, Va.	MA	Mt. Airy, N.C.
SU	Stumpy Point, N.C.		


A
Qz
• coarse-grained biot. and biot.+amph.
× fine- to medium-grained biot.
○ biot. with musc., cord. or gar.
90
60
20
5
syenogranite
monzogranite
trondhjemites
granodiorite
quartz syenite
quartz monzonite
Ksp
10
35
65
90
Pc
B
Al-(Na+K)
● muscovite + biotite
△ cordierite + biotite
+ biotite + amphibole or biotite
aluminum silicate muscovite
plagioclase
cordierite
amphibole
biotite
Ca
Fe + Mg
C
eastonite
$K_2Mg_5Al_4Si_5O_{20}(OH)_4$
siderophyllite
$K_2Fe_5Al_4Si_5O_{20}(OH)_4$
○ amph.- biot.
× biot.
△ cd.+biot.
□ biot.+musc.
TETRAHEDRAL ALUMINUM
3.0
2.8
2.6
2.4
2.2
2.0
0.2
0.4
0.6
0.8
$K_2Mg_6Al_2Si_6O_{20}(OH)_4$
phlogopite
$K_2Fe_6Al_2Si_6O_{20}(OH)_4$
annite
Fe / (Fe + Mg)
D
eastonite
$K_2Mg_5Al_4Si_5O_{20}(OH)_4$
siderophyllite
$K_2Fe_5Al_4Si_5O_{20}(OH)_4$
Liberty Hill pluton, S.C.
coarse, biot + musc.
coarse, biot.
coarse, amph.+biot.
TETRAHEDRAL ALUMINUM
3.0
2.8
2.6
2.4
2.2
2.0
0.2
0.4
0.6
0.8
$K_2Mg_6Al_2Si_6O_{20}(OH)_4$
phlogopite
$K_2Fe_6Al_2Si_6O_{20}(OH)_4$
annite
Fe / (Fe + Mg)
Composition of Granitoid Garnets in the Southern Appalachians
E
almandine
Stumpy Point, N.C.
Edgemoor, S.C.
Mt. Airy, N.C.
Elberton, Ga.
Bald Rock, S.C.
Siloam, Ga.
Springfield, S.C.
Cuffytown Creek, S.C.
Castalia, N.C.
pyrope
spessartine + grossular
F
initial Sr^{87}/Sr^{86}
0.730
0.725
0.720
0.715
0.710
0.705
SU
STG
CS
continental crust
CC
ED
PO
MA
oceanic crust
600
400
200
0
10^6 yrs

Mineral Chemistry

The mineral chemistry of the granites changes with petrographic type, reflecting rock chemistry and physical conditions of crystallization. The ubiquitous recorder of these phenomena is biotite, which shows a continuous enrichment in iron and tetrahedral aluminum in the series amphibole + biotite -> biotite -> biotite + muscovite -> biotite + cordierite bearing granite plutons (Fig. 5c). The aluminum enrichment trend can also be developed within a single pluton as in the Liberty Hill (Fig. 5d). In this case, the biotite and biotite + muscovite granites are border facies that have more extensively interacted with the wall rocks, producing differences in temperature, water pressure, and oxygen fugacity.

Garnet is an uncommon ferromagnesian phase in the granitoid rocks of the southern Appalachians. It most commonly occurs in later facies of the plutons; only in the Stumpy Point is the garnet an early crystallizing phase (Speer, 1981). These two types of garnet occurrence are also distinguishable by their chemistry; the early garnets are almandines whereas the late crystallizing garnets are spessartine-almandines (Fig. 5e). Not included in Figure 5e are the grossular-spessartine hydro garnets which form in the subsolidus.

Isotopic Compostions

Sr isotopic compositions of the granites, which have been summarized by Fullagar and Butler (1979), and oxygen isotope compositions by Wenner (1981) have been used to determine the source regions for these granites. The granites which would be considered "S-type" (4, 5A,5B) as well as the evolved, two-mica granites (5c) have initial $^{87}Sr/^{86}Sr$ ratios greater than 0.7055. The remaining granites, those considered to be "I-type" (1,2,3) or contaminated "I-type" magmas (5d) lie in a tight cluster just below 0.7055 (Fig. 5f). The oxygen isotope data are better resolved by the petrographic characterizations alone:

	delta O^{18}
muscovite + biotite granites	9.0 - 11.4
cordierite + biotite granite	8.9
biotite granites	7.0 - 8.4
amphibole + biotite granites	5.5 - 7.1

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AGES OF REGIONAL METAMORPHISM AND DUCTILE DEFORMATION IN THE CENTRAL AND SOUTHERN APPALACHIANS

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Abstract

A summary of available geochronology of deformational fabrics formed at greenschist and higher grades in the central and southern Appalachians is presented as a map showing the ages of regional metamorphism and ductile deformation. The ca. 480-435 Ma Taconic event affected most of the Piedmont and Blue Ridge provinces. The ca. 380-340 Ma Acadian event was milder and was centered in the west central Piedmont and eastern Blue Ridge. The 330-230 Ma Alleghanian event exhibits an abrupt ductile deformation front in the eastern Piedmont and extends an unknown distance eastward below the Atlantic Coastal Plain.

Introduction

Important constraints for orogenic models in the central and southern Appalachians are provided by determining ages and intensities of regional metamorphism and ductile deformation. We believe that the data reviewed herein imply a widespread Middle and Late Ordovician deformation that was probably generated by closure of a back arc or other ocean basin along the central Piedmont root zone. This event consolidated the entire Piedmont and Blue Ridge Provinces south of New York. If this conclusion is true, post Ordovician sutures can only exist east of the exposed Piedmont under the Atlantic Coastal Plain or on the African plate.

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In a milestone paper, Butler (1972) interpreted the age of Paleozoic regional metamorphism from isotopic, structural and petrologic data over much of the southern Appalachians. An important conclusion of that paper was that the peak of regional metamorphism was attained in the Blue Ridge at about 470 Ma. In 1972, data in the Piedmont were inadequate to closely bracket ages of regional metamorphism, although Butler correctly perceived a tendency toward younger ages of metamorphism eastward toward the Atlantic Coastal Plain. In the decade that has passed, the data base has increased and it is now possible to assemble a preliminary map of the ages of regional metamorphism and ductile deformation for most of the central and southern Appalachians.

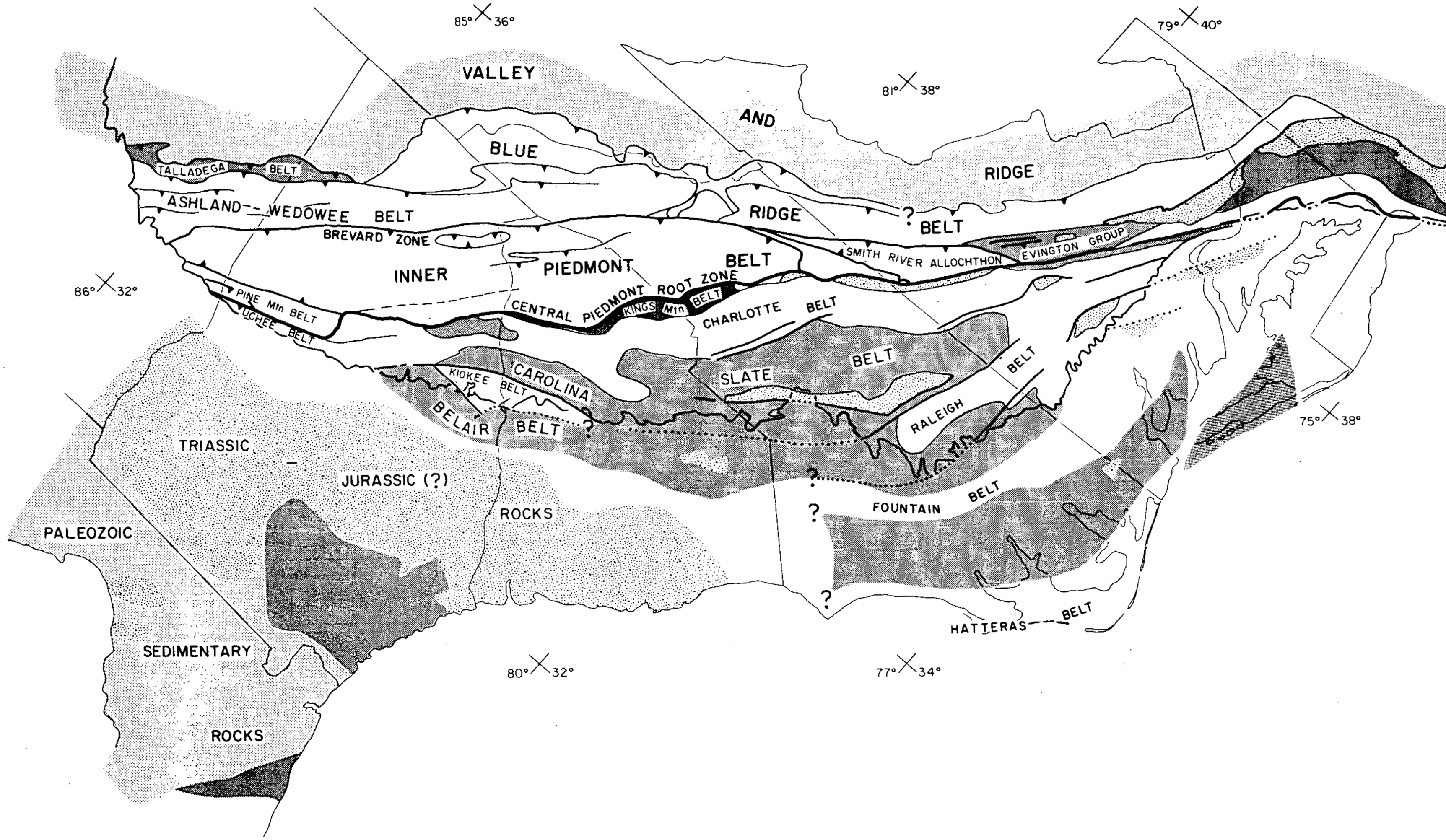
In the approach adopted here, we have concentrated on isotopic ages of regional metamorphic thermal maxima and times of compressional ductile deformation. Thus, this is not a treatment of all the available isotopic data, but only that pertaining to ages of deformation. Mineral ages are probably younger than the thermal maxima for regional metamorphism and represent cooling ages. Schistose, gneissic, and mylonitic fabrics of ductilely deformed rocks are interpreted to be the result of compressional deformation, and the age of this deformation can be bracketed by the isotopic age of these rocks and their structural relation to plutons of known age. These are some of the methods that have been tried and found useful. As outlined below, the onset of regional metamorphism is not a brief event, and most of the dating techniques only approximate the time of the thermal maximum. Deformational events, on the other hand, may be abrupt, brief, and more readily datable, as Naylor (1971) has shown for the Acadian of the northern Appalachians. In the general case, regional metamorphism probably progrades over tens or even hundreds of millions of years, whereas the isotopically datable thermal maxima generally end with the uplift and cooling, hastened by erosion, that attends compressional regional deformation. While the upper parts are abruptly cooled in response to uplift of the higher nappes and folds during deformation, deeper parts of the orogen may be depressed into regions of even higher pressure and temperature, leaving little record of their earlier P-T history. Thus much of the thermal history of an orogen is beyond reading at the present state of our science.

In the following text, the Taconic orogeny is correlated with events within the time span of 480 Ma to 435 Ma or Middle and Late Ordovician. This is consistent with the usage of Rodgers (1971) that the Taconic was a complex series of orogenic events, varying in intensity through time and along strike, that occurred in eastern North America during the Middle and Late Ordovician. Clastic sedimentary sequences indicative of the Taconic orogeny occur in the Valley and Ridge Province of the central and southern Appalachians and include the entire post-Knox and pre-Silurian section from Maryland to Alabama. Nonorogenic quartz arenite and carbonate platform strata characterize the Silurian and Early Devonian Systems.

The effects of a subsequent orogeny to the east are recorded in the Valley and Ridge of the central and southern Appalachians (Fig. 1), by downwarp of the Early Devonian carbonate platform and deposition of early Middle Devonian black mudstone of the Milboro Formation. This is succeeded by the pro-delta Brallier Formation and the rest of the classical Catskill delta facies up through the Early Mississippian Mccrady Formation. The time span for this orogeny is Middle Devonian through Early Mississippian (Meramecian) or from approximately 380 Ma to 340 Ma. Traditionally this orogeny has been correlated with the Acadian orogeny of the Canadian Maritime provinces. More recently isotopic dating (Naylor, 1971; Dallmeyer and Van Breeman, 1981) appears to have restricted the type Acadian to a brief event in the Early Devonian about 390-400 million years ago. This span of time is distinctly older than the "Acadian" orogeny as indicated from the sedimentary history of the Catskill delta and suggests at least three possible relations: (1) the orogeny started earlier in the interior regions of the Appalachian system; (2) the orogeny is diachronous along strike; and (3) the central and southern Appalachian orogeny is a distinct and younger event that we might call the Catskill orogeny. The answers to these questions are beyond the scope of this review, and until additional work is done, the name Acadian is herein continued for central and southern Appalachian orogenic events falling within the age range of about 380 Ma to 340 Ma. Following the wanning Acadian orogeny, the continental platform was briefly capped by the Early Mississippian (Meramecian) Greenbrier Limestone. Presumably this is a non-orogenic(?) deposit representing a brief lull in deformation between about 340 and 330 million years ago.

The Alleghanian orogeny (Woodward, 1958; and comments by Rodgers, 1970, p. 30fn.) has its type region in the central Appalachians where Pennsylvanian and Permian rocks were folded by it. Sedimentologically, the clastic rocks attributed to it range in age from Middle Mississippisan through Permian or about 330 Ma to 230 Ma. As indicated below, ductile deformation in the eastern Piedmont continued to produce schists, gneisses and mylonites from about 300 Ma to 250 Ma. We accept the age range of about 330-230 Ma for the Alleghanian orogeny. This is slightly larger than the 300-250 Ma range of Snoke et al. (1980) for the Kiokee belt of South Carolina. We also prefer the term Alleghanian to Hercynian (Snoke et al., 1980) because the latter has to be correlated from another continent. As shown in the following text, between about 340 Ma and 300 Ma, development of

Figure 1. Generalized geologic map of the central and southern Appalachians encompassing both the exposed rocks of the folded Appalachians and Piedmont and rocks covered by the Atlantic coastal plain. The light shading indicates Palezoic sedimentary rocks in the Valley and Ridge and in northern Florida. The intermediate shading indicates rocks with greenschist facies metamorphic mineral assemblages. The unornamorphic mineral assemblages. The darkest shading of the Kings Mtn. belt includes rocks of either greenschist or amphibolite facies. The gravel pattern indicates Triassic to Jurassic sedimentary rocks.


85° 36°
79° 40°
81° 38°
VALLEY
AND
RIDGE
BLUE
RIDGE
BELT
TALLADEGA BELT
ASHLAND--WEDOWEE BELT
BREVARD ZONE
INNER PIEDMONT BELT
SMITH RIVER ALLOCHTHON
EVINGTON GROUP
86° 32°
PINE Mtn BELT
UCHEE BELT
CENTRAL PIEDMONT ROOT ZONE
KINGS Mtn BELT
CHARLOTTE BELT
CAROLINA SLATE BELT
KIOKEE BELT
BELAIR BELT
RALEIGH BELT
75° 38°
TRIASSIC - JURASSIC (?) ROCKS
FOUNTAIN BELT
PALEOZOIC SEDIMENTARY ROCKS
HATTERAS BELT
80° 32°
77° 34°

ductile deformation in the form of mylonite and regional schistosity shifted from the southeastern Blue Ridge and western Piedmont to the easternmost Piedmont. Thus most of the ductile deformational effects of the Acadian are separated from the Alleghanian in time and in area of occurrence in the crystalline terrane of the central and southern Appalachians. Although there are clear reasons for separating the Acadian and Alleghanian in the central and southern Appalachians, it also seems quite likely that they could be two phases of one long protracted event.

High and low grade metamorphic belts, belt names, and regional faults are shown on Figure 1. contains unmetamorphosed platform and foreland basin sequences, ranging in age from Cambrian through Pennsylvanian. Deformation occurred chiefly in the Pennsylvanian and Permian according to Van der Voo (1979), but earlier deformation may have affected parts of the easternmost regions of the belt in northern Virginia. The Blue Ridge belt contains granulite- and upper amphibolite-grade, granitoid rocks belonging to the 1 Ga Grenville Province. In the western Blue Ridge of northern Virginia, Cambrian sandstone of the Valley and Ridge rests directly upon Grenville basement or is separated from it by less than a kilometer of Eocambrian(?) basalt. In contrast, eastern outcrops of the Blue Ridge belt expose as much as 9 km of dominantly deep-water, Eocambrian submarine fans of arkosic sandstone derived from the west with intercalated basalts, all resting with profound unconformably upon Grenville basement. The Ashland-Wedowee belt appears to be largely a southern extension of the Eocambrian Blue Ridge cover rocks. During the Ordovician and Devonian, parts of the Blue Ridge basement experienced retrogressive metamorphism while the cover rocks underwent prograde metamorphism of greenschist to amphibolite facies. The Talladega belt consists of epiclastic rocks with minor chert, carbonate and basalt. The possible age range is Ordovician through Devonian or younger. Only the Devonian age is certain and this assignment is from fossil evidence. Only Devonian metamorphism seems to have affected the Talladega belt. In all of these belts, thrust faults dip southeast and folds verge westward.

The Chauga belt (included here in the Inner Piedmont) is a sequence of low-to medium-grade, epiclastic rocks with intercalated basalt, lying along the west margin of the Inner Piedmont. These rocks are similar to the Evington Group, discussed below, and are probably mostly Cambrian in age. The Inner Piedmont belt is at present the least known belt of the central and southern Appalachians. It contains mica schist, paragneiss and quartzite of probable epiclastic origin, layered metavolcanic rocks and amphibolite, and rarely, marble. The central and southern parts of the Inner Piedmont are in the sillimanite zone of metamorphism, while the northwestern and southeastern flanks are kyanite-bearing. Dated intrusive rocks suggest that the layered rocks are pre-Ordovician. Subhorizontal recumbent folds and thrusts are common near the center of the belt whereas steeper dipping structures prevail on the margins. The Brevard fault zone bounds the Inner Piedmont-Chauga belt on the northwest and may reappear to the southeast at the boundary between the Inner Piedmont and the Pine Mountain belt in Georgia and Alabama. The Smith River allochthon is probably an outlier of the Inner Piedmont now disconnected by erosion across the Sauratown Mountains antiform. The Evington Group and its extension northward into the Maryland and Pennsylvania Piedmont consists of pre-Middle Ordovician, Cambrian and Eocambrian(?) mica schist, quartzite, marble, and metabasalt. These rocks are repeated by layer-parallel thrusting and along their eastern margin are in contact with melange. The Pine Mountain belt is cored by Grenville granulite basement rocks overlain by quartzite, marble and schist of undetermined age. Thrust sheets or nappes of the Inner Piedmont, Smith River allochthon and Evington group/melange seem to be rooted along the Central Piedmont Root Zone approximately as shown by the heavy line on Figure 1. Hatcher and Zietz (1980) referred to the Carolinas portion of this zone as the Central Piedmont Suture Zone, which Glover and Sinha (1973, p. 249) earlier speculated might be a suture zone.

To the east of the Central Piedmont Root Zone several named and unnamed belts of amphibolite and greenschist facies rocks (Kings Mtn., Charlotte, Carolina slate, Kiokee, Raleigh, Eastern slate, Belair, Fountain, and other unamed belts) comprise the exposed Piedmont and its extension as the basement under the Atlantic Coastal Plain. These belts are of dominately calcalkaline volcanic origin with rhyodacite, which is more abundant than basalt and andesite, probably comprising more than half of the sequence. The abundance of rhyodacite and the presence of conglomerates bearing quartz-arenite boulders suggest eruption above a subduction zone on a continental margin (Glover and Sinha, 1973). Grenville basement is exposed in the core of the Raleigh belt and its extension into northern Virginia (Glover et al., 1978, 1982; Farrar, 1982). In the Kings Mountain belt, volcanic rocks believed to be extensions of the Charlotte belt appear to grade upward into mica schist with quartzite and marble (Horton and Butler, 1982). Age range of the volcanic sequences in the eastern Piedmont from greater than 740 Ma to as young as 530 Ma, or Late Precambrian and Cambrian. Uncomformably overlying these rocks in several narrow infolds in northern Virginia are Late Ordovician to Silurian(?) black slates and schists of the Arvonia sequence. As discussed below, the belts east of the central Piedmont Root Zone were probably widely affected by Taconic ductile deformation and metamorphism, less widely by the Acadian, and by the Alleghanian predominantly in the Keokee and Raleigh belts.

The superposition of three distinct orogenic maxima, Taconic, Acadian and Alleghanian, on the area of Figure 2 is established by the data summarized in the following sections. Decay constants, errors and type of isotopic analysis for all data cited are given in Appendix 1.

Onset of Phanerozoic Metamorphism

A four to 12 kilometer thickness of Late-Precambrian to Cambrian and possible Early Ordovician sediments (the Ocoee-Murphy sequences in the Carolinas, the Lynchburg - Catoctin - Evington sequences in Virginia, and the Araby-Frederick/Grove Limestone sequences in the western Maryland Piedmont) accumulated along the eastern margin of the Blue Ridge. The top of the section is not known to be exposed on the east side of the Blue Ridge. Intermittant sediment accumulation without major interruptions may have occurred from Late Precambrian until Early Middle Ordovician time. The thickness at the beginning of the Taconic Orogeny could have been as much as 15 km. This is consistent with (1) the emplacement of the Mount Airy granite into Cambrian(?) metasediments of approximately 6 kbar pressure (de Rosset, 1978) before the Lower Devonian; (2) the ca. 3 kbar pressure of Ordovician metamorphism recorded in Cambrian/Lower Ordovician(?) Evington group northeast of Lynchburg, (R. D. Tucker, pers. comm., 1980); and (3) the pressure required for Taconic regional metamorphism in the upper greenschist to lower amphibolite facies which affected rocks as young as Early Ordovician on the west flank of the Blue Ridge. The time during which the section accumulated was at least 250 million years.

Models of temperature distribution during sedimentation (Birch _et_ _al_., 1968) suggest that in 30-50 million years after deposition, temperatures of 100°C to 450°C can be present at the base of a five to 15 km sediment pile _without_ extraordinary heat flow such as that which might be related to rifting or magmatism. But it is generally thought that the coarse turbidite and basalt sequences of the Late Precambrian-Cambrian in the western Piedmont and eastern Blue Ridge did accumulate in rift generated basins (Rankin, 1975), an environment that can be characterized by as much as three times normal heat flow (Lee, 1980). Four to twelve kilometers of this rift-fill sediment accumulated just east of the present Blue Ridge crest prior to the Cambrian (Brown, 1970; Hadley, 1970). Undoubtedly, low thermal conductivity of these rapidly accumulated and water-rich sediments, augmented by high heat flow and basalt injection associated with rift tectonics, contributed to temperature gradients even higher than those suggested by Birch's model.

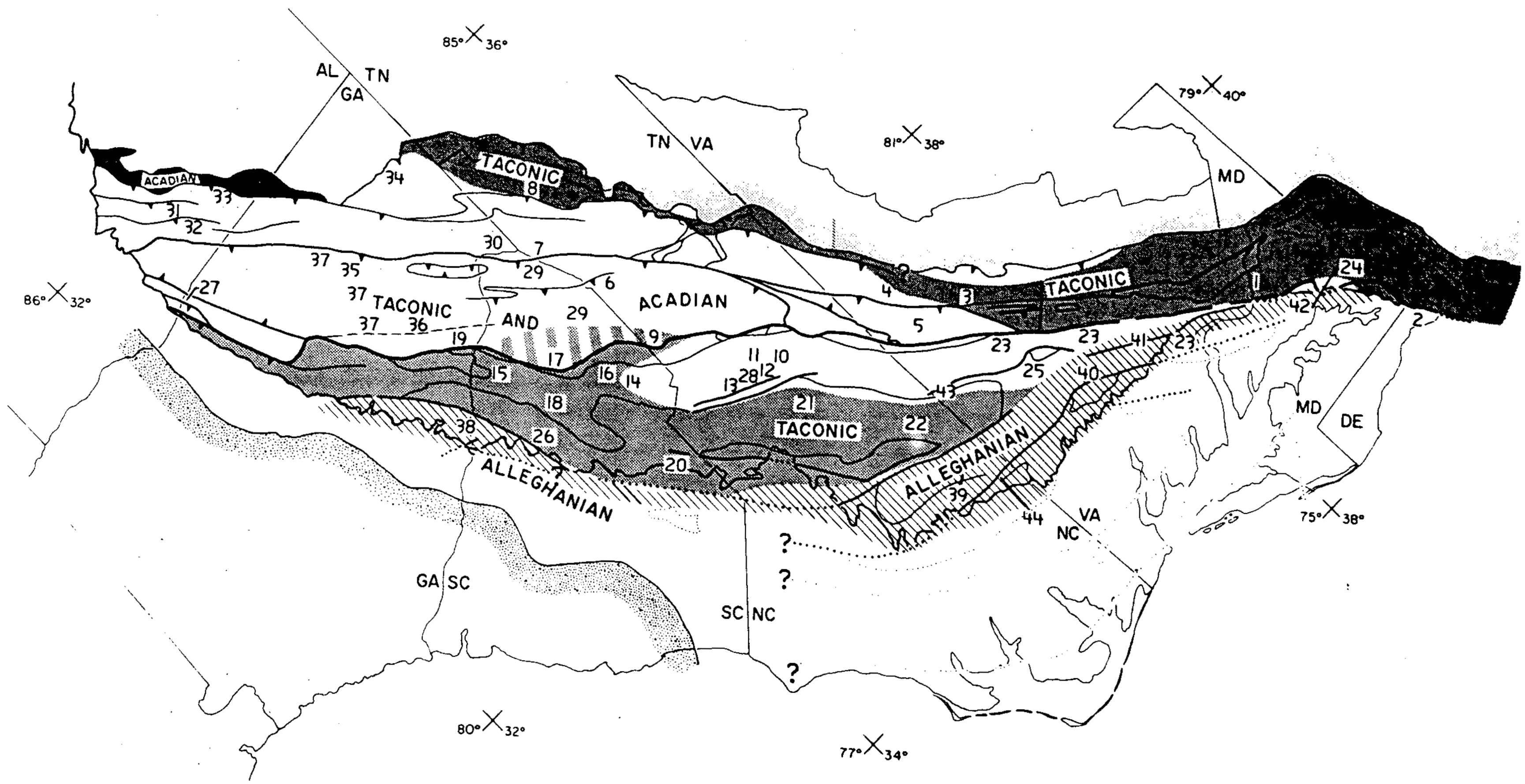


Figure 2. Timing of metamorphism of the crystalline rocks in the central and southern Appalachians. Intermediate shading indicates rocks metamorphosed during the Taconic (480-440). Unshaded areas are rocks believed to be primarily metamorphosed in the Acadian. The ruled pattern indicates rocks subjected to regional metamorphism during the Alleghanian (300-250 Ma). The numbers refer to localities discussed in the text.

Certainly by the end of the Cambrian much of the sequence was experiencing greenschist metamorphism and the lower parts may have reached the amphibolite facies.

Middle Ordovician turbidites (the Knobs Formation of southwest Virginia) and conglomerates (Fincastle Conglomerate of the Roanoke area of Virginia) contain metamorphic pebbles and sand grains resembling Chilhowie and Candler or Lynchburg (Kelberg and Grant, 1956; Cooper, 1960; Glover, unpub. data) which were derived from eastern sources, in accordance with the suggested thermal history above. Thus, Late Precambrian and Cambrian(?) rocks east of the Blue Ridge were already metamorphosed and were being eroded by Middle Ordovician time.

In the Carolina slate belt of Virginia and North Carolina (Fig. 1), approximately seven to 13 kilometers (Glover and Sinha, 1973) of pyroclastic debris, sills, dikes, and plutons accumulated between about 740 and 500 Ma. The base of the sequence is not exposed or is obscured by metamorphism and deformation in the adjacent high-grade belts. The top of the sequence is absent as a result of erosion. Although accurate thicknesses cannot be determined because of the ductile strain involved, a sediment thickness of about 13 km, that accumulated during Late Precambrian and Early Cambrian time, seems to be a reasonable estimate. The position of the accumulating Carolina slate belt in a region of high thermal gradient on a volcanicly active continental margin (Glover and Sinha, 1973), the effects of thermal blanketing by water-saturated pyroclastic debris, and 200 million years of nearly continuous injection of the volcanic pile by magmas, suggest that low to middle greenschist grade metamorphic conditions existed a few kilometers below the surface by Cambrian/Ordovician time.

The implication of this is that metamorphism accompanying Taconic (ca. 480-435 Ma) deformation was not a brief, transient event lasting a few million years, but had its inception in the late Precambrian as much as 200 million years earlier. The time of the thermal maximum is more difficult to obtain. Isotopic dating techniques generally yield ages on the cooling slope of the curve of temperature vs. time, e.g. Dallmeyer (1975). Thus it becomes a difficult task to identify the exact time at which rocks began to cool from a thermal maximum. Because of the low thermal conductivity of rock, uplift and erosion are generally considered to be the principal means by which rocks cool. During the Taconic orogeny compression-driven thrusting and folding probably accomplished the necessary uplift. In general the thermal history curve should show a long slow buildup followed by a more abrupt cooling phase driven by erosion caused by deformation and uplift.

Taconic Metamorphism and Ductile Deformation

The Taconic orogeny produced widespread regional metamorphism and ductile deformation throughout the Piedmont of the central and southern Appalachians. In the Blue Ridge only the westernmost basement rocks with a Grenville age metamorphism locally escaped intense overprinting.

According to Muth et al. (1979), the Piedmont Wissahickon Formation near Great Falls on the Potomac River (Fig. 2, No. 1) was intruded by mafic magma, multiply deformed, metamorphosed and later intruded by lamprophyric dikes and quartz veins. At the peak of metamorphism small granitic and pegmatitic bodies intruded the Wissahickon after formation of early folds and cleavages and prior to the latest folds and crenulation cleavage. Muth et al. dated muscovites from the pegmatites and granites by the Rb-Sr method and obtained ages from two samples of 469 ± 20 Ma and 469 ± 12 Ma, assuming an initial ratio of 0.707. The calculated age of the muscovites would not change more than three million years using the extreme range of initial ratios for Phaneozoic rocks (0.70 to 0.74). Muth et al. state that these dates probably record an age between intrusion and cooling of the rocks below about 500°C and provide a minimum age for the granites and high grade metamorphic rocks that they intrude. The age is interpreted as a cooling age from Taconic metamorphism and the thermal peak of metamorphism in this area may have been reached a few million years earlier than 469 Ma.

Although Fisher et al. (1979) have interpreted the geologic framework of the Wissahickon Formation north of the Potomac in terms of sedimentary facies and fold nappes, Drake and Morgan (1981) report that thrusting and ophiolitic melange comprise the dominant style of emplacement of the Wissahickon terrane just south of the Potomac. These apparently conflicting interpretations have not been resolved. A further internal conflict in the sequence of thrusting and intrusion implied by the Drake and Morgan model is that the high grade rocks that contain the 469 Ma cooling age of Muth et al. were thrust over cooler, low grade rocks prior to intrusion into the package of the Occoquan Granite which has a 494 ± 14 Ma Rb-Sr whole rock age (Mose, 1979, Mose and Nagel, 1982). If this model is correct, the muscovites that Muth dated should (and might) be older than 500 Ma in order to be intruded by a granite of this age. Because the regional structure and geochronology have not been brought into agreement it seems reasonable to consider more than one possible model: (1) that the muscovite ages indicate that the Wissahickon metamorphic terrane along the Potomac River was deformed and cooled during the Taconic orogeny a little before 469 Ma; or (2) that some of the metamorphism and deformation in the Wissahickon terrane really is older than

494 Ma and represents early deformation in a subduction wedge (melange in the sequence) as the pre-Taconic basin was closing. We tend to favor the second possibility.

Grauert and Wagner (1975) dated metamorphism of the Wilmington Complex in Delaware and Pennsylvania (Fig. 2, No. 2), using U-Pb analysis of zircons from the granulite-facies banded gneiss. The zircons in the Complex are from high grade metavolcanics(?) which are believed to be correlative with the James Run volcanics (ca. 550 Ma. or older Higgins et al., 1977) along strike to the south. The Pb-U ratios all plot close to the lower concordia intercept of the discordia at 441 Ma. The smallest size fraction is concordant within analytical error, and the upper intercept (ca. 1500 Ma) is interpreted as an inherited Precambrian zircon component. The 441 Ma age plus their estimated error of 10 percent due to "possible later disturbances" suggests 440-480 Ma for the age of Taconic metamorphism in the Wilmington Complex. Foland and Muessig (1978) have confirmed this metamorphic age by the Rb-Sr mineral isochron method. The age is broadly consistent with the data of Muth et al. above, and is the basis for their statement that, "...the geochronologic data currently available suggest that an area of the Piedmont extending at least from the Potomac to the Delaware River was affected by events associated with the Taconic orogeny."

At the Piedmont-Blue Ridge boundary near Lynchburg, Va. (Fig. 2, No. 3), Robison (1978) determined conventional K/Ar dates on biotite, hornblende, plagioclase, and muscovite in rocks of the retrogressively metamorphosed Grenville basement rocks, the Late Precambrian Lynchburg Formation and the Late Precambrian(?)-Cambrian-Early Ordovician(?) Candler phyllite of the Evington Group. Hornblende ages ranged from 484 to 356 Ma and biotite ages from 494 to 375 Ma. There was little difference between the ages for biotite and hornblende in one mineral pair, and this was interpreted as recording rapid uplift and cooling through the 500°C to 300°C temperature range. Muscovite ages as young as 320, 314 and 208 Ma suggested to Robison that there was localized retrogression following the main phase of Taconic metamorphism. Geologic reconnaissance of the Lynchburg area (L. Glover, unpub.) and detailed studies of structure along strike north of Lynchburg (Robert Tucker, pers. comm., 1980) show that post-metamorphic faulting and mylonitization at greenschist grade or lower is common in this area. It seems probable that this has disturbed some of the K/Ar ages and would be one factor to account for the scatter recorded by Robison. We concur with Robison that the older mineral ages possibly are the least disturbed and record evidence of Taconic metamorphic effects in the Blue Ridge and Piedmont near Lynchburg.

North of latitude 36°52'N (Fig. 2, No. 4) in the southern Blue Ridge of Virginia, Fullagar and Dietrick (1976) found Rb-Sr whole rock isochrons of 520, 531 and 583 Ma in the Late Precambrian Lynchburg/Ashe metasedimentary sequence. South of this latitude whole rock isochrons in the same sequence are 350 to 402 Ma. Biotite-whole rock ages on the south are 317, 329, 337 Ma and as far north as 37°N are 346 Ma (mu) and 332 Ma (bi). No mineral-whole rock pairs were analyzed north of 37°N. Some of the whole rock isochrons are from a single locality and others include samples 100 km apart. The mineral-whole rock pairs suggest that the rocks cooled from an earlier thermal maximum older than 346 Ma. (Acadian?). Additionally, mineral isotopic equilibrium was affected at least 8° north of latitude 36°52'N the boundary where whole rock isochrons responded to an Acadian event on a regional basis. No whole-rock mineral pairs were dated as far north as the Lynchburg area where Robison (1978) worked. The results of Fullagar and Dietrick suggest that Late Precambrian metasedimentary rocks from the Lynchburg area south, approximately 65 km, to 36°52'N appear to have preserved an isotopic event of dewatering or, more probably, of metamorphism that was imprinted on these rocks by Cambrian time during the interval 520 to 583 Ma. This is consistent with the model of the thermal history of this region presented earlier, and may record dewatering and metamorphism in a back arc region of high heat flow. Metamorphic facies of these rocks range from upper greenschist to lower amphibolite.

Robison (1978) speculated that in the Lynchburg area Rb/Sr whole rock systems closed on a regional basis at about 583 Ma (from evidence of Fullagar and Dietrich, 1976) and remained too hot for argon retention until relatively rapid uplift attending the Taconic orogeny at about 480 Ma.

In the western Piedmont (Fig. 2, No. 5), just east of the region studied by Fullagar and Dietrich (1976), Conley and Henika (1973) described a major thrust sequence of sillimanite-grade gneisses and schists called the Smith River Allochthon. They interpreted the metamorphosed sequence to be cut by a pre-thrust igneous complex (Leatherwood Granite-Rich Acres Gabbronorite). The granite was subsequently dated by Odom and Russell (1975) who obtained a Rb/Sr whole rock age of 464 ± 20 Ma. Thus the pre-intrusion metamorphism of the allochthon is pre-464 Ma. It was probably Taconic because the allochthonous seqence is similar to nearby para-autochthonous Late Precambrian-Cambrian rocks. The rocks of the Smith River allochthon are probably rooted on the east below the Danville Triassic basin (Fig. 1), because they are unlike any rocks in the Piedmont east of the basin. The plutonic and host metamorphic rocks are cut by the basal Ridgeway thrust (Conley and Henika, 1973), making the thrust a post-464 Ma structure. Both the Leatherwood Granite-Rich

Acres Gabbronorite and the host rocks of the allochthon were overprinted by a later, probable Acadian(?), possibly Alleghanian metamorphism.

The root zone, which marks the eastern limit of the Smith River Allochthon and similar units, is herein named the Central Piedmont Root Zone (Fig. 1) and is extended along strike to coincide with a boundary between dominantly volcanic rocks on the east and terranes containing abundant epiclastic rocks on the west. The line corresponds to the root zone adjacent to the Kings Mtn. belt shown by Hatcher et al. (1979, p. 3) and to the western thrust boundary of the Baltimore Gabbro (Fisher et al., 1979) in Maryland. It is the probable locus of Taconic suturing.

In the Inner Piedmont allochton (Fig. 2, No. 6) of western North Carolina, Lemmon (1980) has mapped the Fruitland Mountain area. Rubidium-strontium whole rock isochrons and zircon Pb/U dating of the granite and granitic gneisses was done by Odom and Fullagar (1973) Odom and Russell (1975), and Sinha and Glover (1978). One of the gneisses in the Fruitland area, the pre-S_1 foliated Henderson Augen Gneiss (535-595 Ma), is more deformed than a cross-cutting biotite granitic gneiss dated at 438 ± 22 Ma that intruded it during the Late Ordovician or Silurian. The structural history is similar to that along the nearby Brevard Zone at Rosman, N.C., where S_1 and S_2 are nearly coplanar, as they are near Fruitland. In the Brevard Zone S_1 and S_2 have been shown by Clark et al. (1978), Sinha and Glover (1978), Butler (1973), and Roper and Dunn (1973), to be Taconic and Acadian respectively. In the Fruitland area, Taconic metamorphism and deformation appear to have overprinted the Henderson gneiss prior to emplacement of the 438 Ma biotite granite gneiss. Subsequently, the biotite granite gneiss was overprinted by the younger Acadian metamorphism and deformation.

Near Rosman, N.C. (Fig. 2, No. 7), the studies cited above lead to the conclusion that Brevard mylonites formed during D_1 are Taconic whereas those formed during D_2 are Acadian. S_1 and S_2 are coplanar in the Brevard Zone and begin to diverge outside of it. Zircon Pb/U studies by Sinha and Glover (1978) date the Taconic event at about 460 Ma and Rb/Sr whole rock studies of mylonites by Odom and Fullagar (1973) dated the Acadian event at about 360 Ma. The D_1 Taconic event occurred at amphibolite facies metamorphic conditions and the D_2 Acadian event at greenschist to amphibolite facies conditions. Sinha and Glover (1978) found that the D_2 event apparently was sufficiently intense to homogenize Sr isotopes but had little effect on the Taconic mylonite zircons. The mylonite zircons studied by Sinha and Glover were recrystallized with reduction in grain size, loss of zoning, increase in U content, and their ages reset from an earlier Cambrian crystallization age during the first (Taconic) episode of mylonitization. Taconic and Acadian metamorphism, deformation and mylonitization probably occurred in the Brevard Zone rocks when they formed part of the Piedmont-Blue Ridge sole thrust. Late brittle deformation (Alleghanian) brought them to the surface along with exotic slices of carbonate autochthon at the time the Brevard formed as a splay off of the decollement (Cook et al., 1980).

Kish et al. (1976) dated the Cox No.1 pegmatite at Bryson City, North Carolina (Fig. 2, No. 8) by the whole rock Rb-Sr method at 440 ± 13 Ma. According to Kish et al. the emplacement of the pegmatite followed the peak of deformation and metamorphism, suggesting a Taconic age for the metamorphism. Dallmeyer (1975) obtained $^{40}Ar/^{39}Ar$ gas release patterns of biotite and hornblende from retrograded, kyanite-grade basement gneisses in the Bryson City area of the North Carolina Blue Ridge. Total gas ages of biotite range from 343-350 ± 15 Ma and for hornblende 415-421 ± 15 Ma, and the youngest age of peak metamorphism must be older than 421 Ma. Dallmeyer's data is consistent with that of Kish et al. (1976) and supports an Ordovician (Taconic) age for the peak of thermal metamorphism and deformation in this part of the Blue Ridge. The Bryson City area appears to be at or west of the edge of Acadian ductile deformation and metamorphism discussed below.

Along the east side of the Inner Piedmont allochthon (Fig. 2, No. 9) at the boundary with the Kings Mountain belt near the N.C.-S.C. state line, Horton (1981) reports spodumene pegmatite dikes along the newly mapped belt-boundary fault zone. The dikes are confined to the Inner Piedmont block. Some spodumene pegmatites in the zone are highly deformed with augen gneiss textures and others, presumably slightly younger, show no evidence of deformation. Kish (1977) gave a whole rock Rb-Sr age of 352 ± 10 Ma for all of these pegmatites. This Acadian age is therefore taken by Horton as the approximate time of transition from ductile to brittle deformation in the Kings Mountain shear at the belt boundary. Horton (1977) believes that regional amphibolite-facies metamorphism and the two deformation events in the adjacent Kings Mountain belt preceded the pegmatites and "were probably produced by an older Taconic(?) orogenic episode." Horton's data only requires that the earlier and more intense deformation of the country rocks be pre-352 ± 10 Ma. However a Taconic deformation is possible and consistent with the other data from the Inner Piedmont and Smith River Allochthon but is not required.

In the central and western Charlotte belt of North and South Carolina (Fig. 2, Nos. 10,11,12,13,14), Fullagar (1971) reported that five granitoid and syenitic plutons range in Rb-Sr whole-rock isochron age from 415 to 385 Ma. From north to south they are: (1) Yadkin (No. 10) 386 ± 39 Ma, (2) Salisbury (No. 11), 413 ± 9 Ma, (3) Southmont (No. 12), 390 ± 50 Ma, (4) Concord (No. 13), 413 ± 21 Ma, and (5) Lowrys (No. 14), discussed below. A sixth pluton, the Mount Carmel (No. 15), should be older than 385 Ma based on three biotite K-Ar ages discussed by Fullagar (1971). The unusual gabbro-syenite composition and "post-metamorphic" nature of the Concord and Mount Carmel suggested to Fullagar that they are of similar age, "...very likely about 410 Ma." Fullagar considered the Yadkin (386 Ma), Southmont (390 Ma) and Salisbury (411 Ma) to be foliated and metamorphosed, but thought the Lowrys, Concord and Mount Carmel to be post-metamorphic. Reconnaissance field and petrographic work by the authors suggests that these six plutons are all deformed, but less deformed and metamorphosed than their host rocks.

The rock from the quarry used in dating the Lowrys pluton (Fig. 3, No. 14) is a medium-grained, leucocratic biotite alkali-feldspar granite to syenogranite with accessory pyrite, titanite, magnetite, apatite and zircon. Muscovite and chlorite are secondary. The rock has a recrystallized cataclastic texture and in many respects is similar to the Salisbury pluton. It is unlike any facies encountered in other coarse-grained granites of the southeast and elsewhere in the Lowrys itself. Therefore it is probable that a separate, slightly metamorphosed (during Acadian?) pluton of 407 Ma age was dated. The Lowrys itself appears undeformed and may actually be a post-metamorphic pluton.

The Concord gabbro-syenite (No. 13) intrudes foliated amphibolite facies gneisses and schist (Legrand and Bell, 1966) and is discordant to the country rock structure. It is relatively undeformed. However, the Concord syenite locally displays mortar texture around feldspar grains. Thus the Concord appears to have been through some compressive deformation and metamorphism, but less than that of the host rock. It is clearly not a syn-kinematic intrusive because that would have prevented formation of the discordant, nearly circular, surface geometry of both the gabboric and syenitic facies. The available evidence suggests that the host rock had already been deformed and raised to amphibolite facies prior to emplacement of the Concord during a non-compressive, (pre-413 ± 21 Ma) stage in the development of the Charlotte belt terrane. The pre-Concord metamorphism and deformation probably was part of the Taconic orogeny. Subsequent lower grade metamorphism and deformation of the Concord could have been Acadian because it shows deformation of a 413 Ma or younger age and is adjacent to post-deformational plutons of about 300 Ma.

The Mount Carmel gabbro-syenite has been dated at 385 Ma by conventional K-Ar techniques on three biotites (Medlin, 1968). They place only a tentative younger limit on its age. It is emplaced in low grade, Carolina slate belt rocks and in higher grade, generally mafic, gneisses of the Charlotte belt (Griffen, 1978). The pluton is considered to be late syn-kinematic or post-kinematic. Some mortar texture and retrogressive mineral assemblages are present locally within the body but may possibly be late magmatic effects. A similar body to the west, the Calhoun Falls gabbro, cuts fold structures and foliations of the Carolina slate belt, but the country rock structures are deflected (by deformation during emplacement?) around the Mount Carmel body. There is a contact aureole around Mount Carmel. If the biotite ages date cooling through about 300°C, then the Mount Carmel is Early Devonian/Silurian or older and the country rock metamorphic fabric could be as late as Taconic. This is speculative but is consistent with the data from the similar body at Concord, N.C.

Subsequent to the above studies, Fullagar (1981) reported whole rock Rb/Sr ages of two more plutons in the same age group. One of these, the Bald Rock (Fig. 2, No. 16), lies in the Charlotte belt along the east boundary of the Kings Mountain belt in north-central S. C.. It is reported to be undeformed and has an age of 388 ± 6 Ma. A second body, the Gray Court Granite, (378 ± 24 Ma) (Fig. 2, No. 17) occurs just west of the Kings Mountain belt in the Inner Piedmont of South Carolina. It is reported to be a metagranite; however, thin section examination by us reveals only a slight deformation in the rocks at the quarry from which the dated samples came. In central South Carolina, the Newberry Granite (Fig. 2, No. 18) is essentially undeformed and intrudes foliated amphibolite facies rocks of the Charlotte belt. Fullagar (1981) reported a 415 ± 9 Ma whole rock Rb-Sr age from nine whole rock samples. In contrast to the very low degree of metamorphism and deformation in the Gray Court, Newberry and Bald Rock granites, the host rocks are highly deformed and contain amphibolite facies assemblages.

From the information provided by these nine plutons, it seems reasonable to conclude that the amphibolite facies metamorphism in the eastern Inner Piedmont and western Charlotte belt of the Carolinas is pre-413 Ma and probably Taconic in age. Kish et al. (1979) also favored pre-400 Ma amphibolite facies metamorphism in the Charlotte belt of the Carolinas.

The Elberton granite (Fig. 2, No. 19) was emplaced in the eastern Inner Piedmont of northwest Georgia during the interval 320 ± 20 to 350 ± 12 Ma. The first age is a zircon Pb-U date by Ross and Bickford (1980) and the second is a whole rock Rb-Sr age by Whitney et al. (1980). The 350 Ma

Rb-Sr data is inconclusive because of varying initial ratios and is also unusual in recording an older apparent age than zircon from the same pluton. The Elberton Granite cuts the amphibolite facies ductile deformation fabric of the Inner Piedmont but the granite itself is undeformed. The probable 320 Ma age of the Elberton permits interpretation that both Taconic and Acadian deformation may have preceded its emplacement. However, the lack of, or mild ductile deformation, in the Gray Court (378 Ma), Bald Rock (388 Ma) and Newberry (415 Ma) granites only 100 km to the northeast of the Elberton, suggest that the intensity of Acadian deformation may have been minimal in the Elberton area as well.

In the lower grade Carolina slate belt of South Carolina, Bell et al. (1972) reported muscovite K-Ar ages of 430 ± 15 and 401 ± 14 Ma from the Brewer Mine in Chesterfield County (Figure 2, No. 20). The mine is about 2.5 miles southeast of the 296 Ma Pageland granite. Both dated samples contain muscovite, quartz, and pyrophyllite. Among other minerals previously reported from elsewhere in the mine are topaz, kyanite and andalusite. Thus it seems likely that the mineralogy indicates a complex hydrothermal and metamorphic history, possibly beginning with pre-Taconic hydrothermal alteration related to the volcanic history of the slate belt, and continuing through mid- to late-Paleozoic time. The associated kyanite, andalusite, and pyrophyllite suggest that the pyrophyllite (associated with the dated muscovites) may have formed late by replacement of kyanite or andalusite, and that the pyrophyllite-producing event probably was below the approximately 350°C blocking temperature for loss of argon from muscovite. If this is true, the 430 Ma age may be near, but still younger than, the upper greenschist or lower amphibolite grade metamorphism, probably Taconic, indicated by the mineral assemblages.

In the central Carolina slate belt near Albemarle, N.C. (Fig. 2, No. 21), Kish et al. (1979) reported a 483 ± 15 Ma average K-Ar age for eight whole rock samples from Cambrian strata. The samples span the stratigraphic sequence and are metamorphosed to greenschist facies biotite zone or chlorite zone assemblages. "The ages are the same within the analytical uncertainty and are independent of stratigraphic position and potassium content (1.3 to 6.6% K_2O)." Kish et al. reported that metamorphic phengitic mica is the only potassium phase present in the samples dated. Because the metamorphic temperatures did not greatly exceed the approximately 300°C argon retention temperature for muscovite, they believe that 483 Ma represents the approximate time of metamorphism, rather than a cooling age. This data supports a Taconic age of metamorphism for the low-grade Carolina slate belt in central North Carolina.

North of Albemarle, the Carolina slate belt rocks are older, being predominately Late Precambrian in age. In the Durham-Chapel Hill, N.C. area metamorphic grade is chlorite zone greenschist facies and the rocks have a weak foliation. Near Chapel Hill (Fig. 2, No. 22), Rb-Sr whole rock and whole rock-mineral data (Black, 1977) suggest Taconic as the age of regional metamorphism. A whole rock Rb-Sr isochron age of 459 ± 5 Ma for Precambrian dacitic metavolcanic rocks southeast of Hillsboro suggests a Taconic metamorphic age. Additionally, a whole rock-hornblende age of 481 ± 94 Ma was found for the 705 ± 15 Ma Chapel Hill pluton.

Finally, a Taconic unconformity (Fig. 2, No. 23) occurs along the western edge of the Charlotte belt (as defined in Fig. 1) in central and northern Virginia where the Late Ordovician-Silurian(?) Arvonia Slate unconformably overlies 464-550 Ma granites and older volcanics (Brown, 1969).

From the preceding summary and the map distribution of Taconic metamorphism and deformation (Fig. 2) several conclusions can be drawn: Taconic regional metamorphism and ductile deformation of greenschist to middle amphibolite facies occurs in the Blue Ridge, Inner Piedmont, Charlotte belt and Carolina slate belt at least as far east as the present western boundary of the Alleghanian ductile deformation front. In this area, many of the exposed high and low grade rocks of the various belts had probably reached the maximum metamorphic grades that they presently show by the end of the Taconic. However their present map distribution was modified by folding and thrusting during later orogenic events. Preservation of Taconic metamorphic mineral assemblages and regional fabrics in parts of the Blue Ridge and eastern Piedmont suggests that these regions remained at relatively shallow crustal levels throughout the later orogenies. The widespread distribution of Taconic ductile deformation and metamorphism suggests that the entire exposed Piedmont and Blue Ridge was involved in the initial orogenic event. Collision probably occurred along the Central Piedmont root zone, and may have involved closing of either (a) a back arc basin, or (b) Iapetus, a major ocean basin (Glover <u>et</u> <u>al</u>., 1982).

Acadian Metamorphism and Ductile Deformation

Acadian metamorphism and ductile deformation is mostly confined to a lower grade overprint on Taconic fabrics in the west-central part of the crystalline terrane. The Talledega belt of Alabama and Georgia is exceptional in that it appears to have experienced only Acadian metamorphism. As noted previously, future studies may show that the Acadian of the central and southern Appalachians should be renamed if it is unrelated to the apparently earlier, type-Acadian event.

If Acadian ductile deformation and metamorphism affected the Maryland Piedmont its effects must be localized and minimal. Fisher (1970) concluded that greenschist retrograde assemblages developed along the Potomac River (Fig. 2, No. 1) in cleavages S_3 and S_4 might be Carboniferous based on regional mica dates. Hanan and Sinha (1976) reported an approximately 300 Ma whole rock Rb-Sr isochron from the Wissahickon schist, (Fig. 2, No. 24). This has been changed by Sinha (1981, pers. comm.) to 340 Ma. Additional work will be required to resolve the geologic significance of this age.

In the Charlotte belt of the central Piedmont of Virginia (Fig. 2, No. 25), an amphibolite facies gneiss just east of Farmville yields cooling ages by the $^{40}Ar/^{39}Ar$ technique of 300 ± 3 Ma for biotite and 324 ± 3 Ma for hornblende (J. Sutter and L. Glover, unpublished data). The amphibolite facies metamorphism of the host rock suggests a thermal maximum at or above the approximately 500°C temperature threshold of argon retention in hornblende. Therefore, the age of the thermal maximum must be at or slightly older than 324 Ma, suggesting an Acadian age. Because of the evidence for a Taconic regional metamorphism in the Piedmont, the question remains of whether the metamorphic deformation fabric east of Farmville is Acadian or is inherited from Taconic deformation in which case the age represents only cooling during Acadian uplift. Sixteen kilometers west of this location (Fig. 2, No. 23), the Late-Ordovician - Silurian (Stose and Stose, 1948, Tillman, 1970) Arvonia Slate rests unconformably upon rocks correlative with those at Farmville. This post-Taconic graphitic schist and quartzizite formation is intensely foliated and is at least at upper greenschist metamorphic facies. Therefore it seems that a post-Late Ordovician and pre-324 Ma Acadian metamorphic and deformational event occurred in the Charlotte belt of the central Virginia Piedmont. Also, the unconformity indicates that this Acadian event was superimposed on a terrane that had cooled, uplifted and eroded after an earlier Taconic metamorphism.

Just west of the Charlotte belt in southern Virginia, the Leatherwood Granite within the Smith River allochthon (Fig. 2, No. 5) was deformed subsequent to post-464 Ma thrust emplacement (previously discussed) and apparently was metamorphosed to upper greenschist or higher grade, based on the petrographic descriptions of Conley and Henika (1973, p. 24). Muscovite from a pegmatite cutting the Ridgeway fault was dated by Deuser and Herzog (1962) at 321 ± 17 Ma. This is a Rb-Sr mineral age on large (4 X 2.5 inch) muscovite books from the Knight Mine pegmatite. Deuzer and Hezog also reported a muscovite date of 276 ± 15 Ma from the nearby Spencer Mine. These place a minimum age on last movement of the Smith River Allochthon along that fault. The dates also suggest that emplacement, which could have begun during the Taconic, was probably completed between 464 and 321 or 276 Ma, although it is difficult to evaluate these old mineral ages. Work near the north end of the Danville Triassic basin (A. Gates and L. Glover, in progress) suggests initial Taconic movement of the Allochthon.

Most of the Smith River Allochthon lies just east and south of the Blue Ridge Lynchburg/Ashe terrane shown by Fullagar and Dietrick (1976) to have paragneiss whole rock Rb-Sr ages of 350 to 402 Ma (Fig. 2, No. 4). Erosion has reduced the size and areal extent of the allochthon by an undetermined amount. However it seems probable that the once hot and more extensive allochthon extended onto the Blue Ridge terrane and may be largely responsible for the Acadian whole rock cooling ages in the Lynchburg/Ashe formations south of latitude 37°N. The cooling ages therefore were probably established during Acadian deformation(?), uplift and erosion following Taconic hot nappe emplacement. Considering that a cooling interval terminated Taconic thermal events in the region, as evidenced by unconformity below the Arvonia slate, it seems likely that post-Taconic deformation of the Leatherwood granite indicates that some movement of the allochthon occurred during early Acadian. North of approximately 37°N latitude, the western Piedmont and Blue Ridge, unlike the southern Appalachians, may never have been buried by a deep pile of hot thrust nappes. This northern limit of hot thrust nappes is at the structural junction between the central and southern Appalachians defined in the Valley and Ridge (Rodgers, 1970) and suggests that the structural differences between the two segments may have already been established by the Taconic, and certainly by the Acadian. Deformation and metamorphism of the Arvonia appears to be related to reactivation of the Central Piedmont root zone and Charlotte belt during the Acadian.

Kish et al. (1979) recognized that 400-300 Ma granites in the Charlotte belt of the Carolinas were deformed less than the older rocks that they intruded. In the previous discussion of Taconic deformation and metamorphism in the

Charlotte belt it was concluded that the 415-385 Ma granites (Fig. 2, Nos. 10-18), gabbros, and syenites post-date an earlier, amphibolite facies Taconic metamorphism. Those plutons at locations Nos. 10-14, Figure 2 exhibit effects of a lower grade of metamorphism and deformation imposed on them, but as Fullagar (1971) pointed out, those to the south are essentially undeformed. The undeformed Churchland (288 Ma), Landis (298 Ma) and York (322 Ma) granites, which occur in the area of the 415-385 Ma group north of location 14, Figure 2, are clearly post-kinematic with respect to ductile deformation. These relations are interpreted to mean that a mild Acadian ductile deformation, confined in the Charlotte belt to the area north of South Carolina (Fig. 2, No. 14), occurred between 322 and 385 Ma. The Charlotte belt in South Carolina and probably in Georgia received only localized ductile and, perhaps some brittle deformation during this orogeny because the Bald Rock (Fig. 2, No. 16, 388 Ma) and Newberry (Fig. 2, No. 18, 415 Ma) granites are undeformed.

An example of Acadian ductile deformation in South Carolina may occur along the Modoc fault zone at the west edge of the Kiokee belt (Fig. 2, No. 26) where Fullagar (1981) reported a Rb-Sr whole rock age of 370 ± 8 Ma for three samples of mylonite. However, this is only a three point isochron and it is not clear how it escaped resetting during Alleghanian metamorphism. Along the Bartlett's Ferry fault zone (Fig. 2, No. 27) of Georgia in the Pine Mountain belt, Russell (1978) also found a 375 Ma Rb-Sr whole rock age of mylonites. At the Charlotte-slate belt contact of central and southern North Carolina (Fig. 2, No. 28) Butler and Fullagar (1977) found that the semiductile or ductile Gold Hill fault zone cuts and deforms the Gold Hill granite which they consider, on the basis of petrographic similarity, to be the age of the adjacent 400 Ma "Salisbury group of plutons." The southern end of the Gold Hill fault zone in South Carolina is intruded by the 329 ± 14 Ma undeformed Catawba granitoid (Butler, 1977; Fullagar and Butler, 1979). The major movement of the Gold Hill fault zone is therefore loosely constrained between 400 and 329 Ma, approximately Acadian.

From the forgoing, the evidence for Acadian deformation and metamorphism in the Charlotte belt of the central and southern Appalachians as far south as South Carolina suggests that it was dominantly of low (greenschist) metamorphic grade and tends to be localized near major fault and mylonite zones that bound the belt. Except near marginal fault and mylonite zones, the Carolina slate belt has yielded no evidence of Acadian dynamothermal metamorphism. Both the Charlotte and Carolina slate belts could contain semiductile or brittle faults and folds of Acadian age but isotopic and field studies to date have not discovered them.

At the South Carolina-North Carolina border, Horton (1981) described the Kings Mountain shear zone (Fig. 2., No. 9) that separates Inner Piedmont and Kings Mountain belts. As mentioned previously, Horton showed the Kings Mountain shear to be a zone of mylonitic deformation in which late-stage semi-brittle cleavage formed. Spodumene pegmatites, that Kish (1979) dated by Rb-Sr whole rock techniques at 352 ± 10 Ma, are only locally mylonitized where they occur in the Kings Mountain zone. Horton interprets the field relations to suggest that 352 Ma is approximately the time of last pegmatite deformation and that this time (Acadian) marked the transition from ductile to semibrittle deformation along the eastern margin of the Inner Piedmont. Three hundred Ma granites, the York, Clover, and Gastonia located nearby, are undeformed.

Just west of the Kings Mountain belt in the Inner Piedmont of South Carolina the Gray Court granite (Fig. 2, No. 17, 378 ± 24 Ma), (Fullagar, 1981) cited in the discussion of Taconic deformation, is nearly undeformed at the quarry from which the samples were collected. Thus Acadian ductile deformation was also mild at this latitude. Farther south in Georgia the 320 Ma Elberton granite (Fig. 2, No. 19) (Ross et al., 1980) also lies at the eastern edge of the Inner Piedmont and is unfoliated according to Dallmeyer et al. (1981), although it may be too young to record "Acadian" deformation.

In the central and western Inner Piedmont of South Carolina, Fullagar (1981) dated several foliated granitic gneisses (Fig. 2, No. 29). Eighteen samples yielded a Rb-Sr whole rock isochron of 421 ± 3 Ma, believed to be the time of igneous crystallization. The degree of deformation and the grade of metamorphism of these gneisses is not indicated, but apparently post-Taconic, Acadian(?) metamorphism and deformation affected a large region of the Inner Piedmont in South Carolina. Lemmon (1980), as previously discussed under Taconic deformation, found similar granitic gneisses of 438 Ma age in the western Inner Piedmont of North Carolina (Fig. 2, No. 6). The gneissic fabric of these rocks was probably formed during the Acadian. Just southwest of Lemmon's area, within the Brevard zone (Fig. 2, No. 7) at Rosman, N.C., Odom and Fullagar (1973) concluded from whole rock Rb-Sr studies of the mylonite that deformation and metamorphism homogenized Rb-Sr isotopes at about 360 Ma also during the Acadian.

Hatcher and Butler (1979) concluded that amphibolite facies Taconic metamorphism in the eastern Blue Ridge and Piedmont (Fig. 2, No. 30) is overprinted by greenschist facies Acadian metamorphism. This is based on correlation of structural elements S_2, F_2, and metamorphic grade with isotopic data discussed above for the Brevard zone.

In the Ashland-Wedowee belt of Alabama, Russell (1978) obtained a Rb-Sr whole rock isochron of 366 ± 9 my. from the Bluff Springs Granite (Fig. 2, No. 31). The Bluff Springs has a poorly to moderately developed foliation (Russell, 1978), and the deformation in the Bluff Springs must therefore be Acadian or younger.

Russell (1978) also attempted to date Ashland (Fig. 2, No. 32) and Wedowee (Fig. 2, No. 33) metasedimentary rocks by whole-rock Rb-Sr techniques. They gave an age of 366 ± 25 Ma. Russell interpreted this as the age of homogenization of strontium isotopes during a regional metamorphism. It is also consistent with the age of deformation deduced from the Bluff Springs Granite. W. Russell (pers. comm., 1981) dated mylonites from the Fort Mountain Gneiss (Fig. 2, No. 34) by the Rb-Sr whole rock method at 359 Ma. This is consistent with the dates of G. Russell and seems to confirm both regional(?) metamorphism of the Alabama Piedmont (Blue Ridge extension) and hot ductile thrusting (mylonite) during the Acadian. Kish (1982), however, got an anomalously old date of 397 Ma from seven K/Ar whole rock analyses of greenschist grade Talledega slate in Chilton County, Alabama. No evidence has been found for Taconic regional metamorphism, but it is difficult to believe from the regional relations farther north that Taconic metamorphism and deformation did not extend into Alabama.

Dallmeyer (1978) reported Rb-Sr whole rock isotopic studies of two Inner Piedmont metasedimentary localities in Georgia, one near Stone Mountain (Fig. 2, No. 35) and the other near Athens (Fig. 2, No. 36). These samples suggest homogenization of Sr isotopes at about 365 ± 10 Ma. Dallmeyer noted the similarity with Russell's data and suggested metamorphic continuity during the Acadian across the Brevard Zone. Dallmeyer (1982, pers. comm. to Glover) states that large Rb-Sr whole rock samples from the Lithonia and Athens, Ga. quarries in the Inner Piedmont give 360-380 Ma dates. Small thin-slab analyses of these samples yield much younger "cooling-type" dates. He concludes that regional metamorphism is Acadian and that slow-cooling (as reflected in K-Ar studies) may be observed in Rb-Sr systematics on a small scale. The relations of these ages to the age of the deformational metamorphic fabric is not established by the data presented, but these data are consistent with data farther north that indicate regional metamorphism and deformation of the Inner Piedmont and Blue Ridge terrane during the Acadian. There is no evidence for Taconic dynamothermal metamorphism in the Georgia Inner Piedmont; if it occurred it appears (1) to be obscured by an intense Acadian event or, (2) we haven't dated granites old enough to reveal it.

Dallmeyer (1978) studied $^{40}Ar/^{39}Ar$ incremental release ages of hornblende and biotite across the Georgia Inner Piedmont (Fig. 2, No. 37). From the Brevard Zone on the west to near the eastern edge of the Inner Piedmont biotite ages range from 317 to 236 Ma, and hornblende ages from 355 to 300 Ma. Variations are consistent and progressive, the spectra are undisturbed, and no retrogression was noted in the samples examined in the traverse. The westernmost part of the traverse is in kyanite grade rocks and the eastern 90% of the traverse is in sillimanite grade. According to Dallmeyer, the Stone Mountain granite, with a Rb-Sr whole rock age of 291 Ma (or ca. 325 Ma based on zircon Pb/U dating, Atkins et al., 1980) and near the west central part of the traverse, was emplaced in rocks that had already cooled below the retention temperature of argon in biotite. $^{40}Ar/^{39}Ar$ ages of biotite and muscovite in the Stone Mountain granite are nearly identical at 281 ± 5 Ma and 283 ± 5 Ma respectively, indicating rapid uplift and cooling. Mineral assemblages in the Inner Piedmont country rocks near Stone Mountain indicate temperatures of 700-750°C and pressures of 7-9 kb at the peak of regional metamorphism (Dallmeyer, 1978). This pressure corresponds to depths of 23-29 km at about 365 Ma, the Rb-Sr whole rock ages cited above, and slightly older than the 355 Ma $^{40}Ar/^{39}Ar$ age of oldest the hornblende dated in the traverse. Although the biotite-hornblende cooling ages suggest earlier or more rapid uplift in the western Piedmont compared to the eastern Piedmont,this may be a result of reaching argon retention temperatures earlier in the lower grade kyanite zone. Dallmeyer did not believe that this effect of metamorphic grade was large.

Dallmeyer's study specifically addresses the thermal and uplift history of the Georgia Inner Piedmont; it does not examine the nature or existance of compressional tectonics during the Acadian. It is important though to note that at a time (Acadian) when much of the Blue Ridge, Inner Piedmont, Charlotte and Carolina slate belts farther north were undergoing only greenschist or lower grade metamorphism and deformation, the Georgia Inner Piedmont was at pressures of 7-9 kb and temperatures between 700-750°C which is equivalent to sillimanite zone metamorphism. Development of compressional mylonite zones and folds along the Blue Ridge Fault, Bartletts Ferry Fault, Kings Mountain Fault and Modoc Fault at this time in Alabama, Georgia and South Carolina indicate that the region was probably undergoing compression. This should and may have affected the Inner Piedmont block also, but there is no evidence yet as to whether the regional schistosity is Taconic and cooled during Acadian or whether it is mostly Acadian. The indicated temperature and depth of the present Georgia Inner Piedmont surface at 365 Ma would provide conditions that could mask Taconic effects that might have been present. Metamorphic remobilization of an earlier Taconic terrane or even maintenance of high grade conditions from Taconic into Acadian time seem to be reasonable possibilities on the basis of Dallmeyer's study. Perhaps future studies will show that a similar explanation applies to the lack of evidence for Taconic orogeny in the higher grade belts in Alabama west of the Brevard zone. This would not include the low grade (greenschist) belt of Talledega rocks of Alabama and Georgia (Fig. 2) which is believed to contain Early Devonian fossils (Thomas et al., 1980) and therefore could have been metamorphosed no earlier than Acadian.

In summary, the effects of Acadian ductile deformation and metamorphism in the northern Virginia and Maryland Blue Ridge and western Piedmont (Fig. 2) are cryptic or absent. Parts of the westernmost crystalline Appalachians in North Carolina, Tennessee and Georgia (Fig. 2) also seem to have escaped Acadian metamorphism. Additionally, much of the low grade Carolina slate belt and Charlotte belt of South Carolina and Georgia, and Carolina slate belt of North Carolina and Virginia was probably not overprinted by Acadian metamorphism. Acadian metamorphism probably extended east of the Alleghanian ductile deformation front in Virginia (Fig. 2), but its effects are presently unknown because of Alleghanian overprinting. In the northern Charlotte belt, eastern Inner Piedmont, and Blue Ridge north of Georgia, Acadian effects at the outcropping surface appear to reflect low metamorphic grade and relatively lower strain overprinted on an earlier Taconic fabric of higher strain and metamorphic grade. In the western Inner Piedmont of Georgia, South Carolina, and perhaps North Carolina, and in the higher grade Ashland-Wedowee terrane of eastern Alabama, Acadian metamorphism probably reached the kyanite and sillimanite zone or was held at these conditions from the Taconic orogeny . In southern Virginia the northern margin of the Smith River Allochthon occurs near latitude 37°N at the southward transition from pre-Acadian to Acadian metasedimentary whole rock Rb-Sr ages. Higher grades of Acadian regional metamorphism in the Blue Ridge and Inner Piedmont also occur south of 37°N. This suggests that imbricated hot Taconic and Acadian nappes deeply buried the presently outcropping Piedmont/Blue Ridge surface. Burial occurred to depths and temperatures that could sustain metamorphic recrystallization and strontium isotope equilibration until deformation(?), isostatic rebound, erosion, and cooling during the Acadian dropped temperatures below the threshold of daughter product retention in the respective isotopic systems. The northernmost appearance in Virginia of western Piedmont and Blue Ridge hot ductile nappes at 37°N coincides with the junction between the central and southern Appalachians and suggests that this structural transition was probably established during Taconic and Acadian orogenies.

Alleghanian Metamorphism and Ductile Deformation

Prior to 1978 (Snoke et al., 1980 and references therein) it was generally believed that the youngest regional metamorphic fabric in the crystalline Appalachians south of New England was Acadian. Since that time, Alleghanian amphibolite facies regional metamorphism has been documented along most of the eastern Piedmont (Fig. 2). Along the eastern margin of the Piedmont in South Carolina (Fig. 2, No. 38), Snoke et al. (1980) found Alleghanian ductile deformation and amphibolite facies regional metamorphism in the Kiokee belt. According to Snoke et al. (1980) and Fullagar and Butler (1979), four granite plutons and orthogneisses (Lake Murray, Lexington, augen gneiss, lineated gneiss) along the west edge of the Kiokee belt have Rb-Sr whole rock ages between 313 ± 24 Ma and 284 ± 17 Ma and all contain D_2 lineations or S-surfaces. A fifth granite, the 254 ± 11 Ma Edgefield granite, contains weakly to moderately developed late Alleghanian structures consisting of a folded foliation the axis of which is said to be parallel to D_3 structures (crenulation cleavage and folds with nearly vertical axes) in the Kiokee belt. The evidence indicates that Alleghanian deformation and metamorphism continued from about 300 to 250 Ma. D_2 and D_3 deformation locally can be seen overprinting an earlier D_1 penetratiave foliation of Acadian or Taconic age in the adjacent Carolina slate belt along the western boundary with the Kiokee belt. The 319 ± 27 Ma Clouds Creek granite intruded regionally foliated Carolina slate belt and the granite was mostly deformed along its boundary with the Modoc fault zone. Apparently Alleghanian ductile deformation was mostly confined to the Kiokee belt infrastructure. The gently-southeast-dipping Augusta ductile fault thrust low grade rocks of the Belair belt over high grade rocks of the Kiokee belt. Both the Augusta and Modoc faults were subsequently reactiviated by brittle fracture (Snoke et al., 1980). Dallmeyer (1981) reported $^{40}Ar/^{39}Ar$ ages of hornblende and biotite in the Kiokee belt of 315-285 Ma that are similar to the ages of the deformed Alleghanian granites, and this is broadly consistent with the age of deformation given by Snoke et al. (1980). Dallmeyer (1981) also reported $^{40}Ar/^{39}Ar$ whole rock (mostly muscovite as Ar-bearing minerals) ages of 315-215 Ma in the Carolina slate belt just west of the Kiokee belt. These ages would confirm that the previously (Taconic) metamorphosed Slate belt was at temperatures a little above 300°C during the Alleghanian. However, nearby dated plutons including the Clouds Creek, Cuffytown Creek, and Newberry granites of the Slate belt and adjacent Charlotte belt suggest by their lack of regional foliation that the foliation is probably Taconic in the Slate belt and Charlotte belt of South Carolina.

Alleghanian tectonism and metamorphism has been documented in the Raleigh belt of North Carolina by Farrar et al. (1981). Granitic plutons in the amphibolite facies Raleigh belt (Fig. 2, No. 39) give Rb-Sr whole rock ages of 315-285 Ma an all record an S_3 cleavage. Some of these granites also intrude the adjacent greenschist facies Eastern slate belt, and in this host they are not foliated. As in the case of the Clouds Creek granite to the south (Fig. 2, No. 38) Alleghanian ductile deformation was largely confined to the amphibolite facies Raleigh belt infrastructure, and the Eastern slate belt which had cooled from a Taconic metamorphic event was sufficiently rigid to shield the young granites from deformation. The 292 ± 30 Ma Butterwood Creek granite (Fig. 3, No. 44) straddles the Raleigh belt-Eastern slate belt contact. Rb-Sr biotite-whole rock cooling ages indicate that the Raleigh belt country rock and that portion of the Butterwood Creek which intruded the Raleigh belt remained above about 300°C until 250 Ma; whereas, the adjacent Eastern Slate belt had already cooled below 300°C when it was intruded by the Butterwood Creek. Thus the D_3 event, which formed the S_3 tectonic foliation in the 315-285 Ma plutons, as well as regional F_3 folds and adjacent mylonite zones, was an Alleghanian event. Ductile deformation probably ceased with regional cooling by about 250 Ma as it did in the Kiokee belt.

Along the northern extension of the Raleigh belt near Richmond, Virginia (Fig. 2, No. 40), Wright et al. (1975) dated the Petersburg granite by the zircon Pb-U method. At that time the Petersburg was described as having a single weak foliation of uncertain origin possibly resulting from igneous flow. Therefore the granite was considered to be largely post-kinematic. Subsequently, one and possibly two tectonic foliations (Bobyarchick and Glover, 1979) have been found in the Petersburg complex of granites indicating that deformation did continue later than about 330 Ma. Durrant et al. (1980) found that $^{40}Ar/^{39}Ar$ plateau ages of hornblende and biotite in the amphibolite facies gneiss complex just west of the Petersburg granite (Fig. 2, No. 40) average 262 and 241 Ma respectively. They extrapolated the data to suggest a maximum metamorphic thermal peak at about 300 Ma, based on estimated P-T conditions for sillimanite zone metamorphism. Subsequently we have learned that the sillimanite may be relict from an earlier granulite facies metamorphism, and that the highest grade reached in the last metamorphism is more probably kyanite zone. Thus the age of the last thermal peak is about 262 Ma or older, but younger than 300 Ma. It seems reasonable to conclude that Alleghanian amphibolite facies metamorphism and ductile deformation affected the Petersburg Granite and adjacent gneisses as it did in the Raleigh and Kiokee belts farther south.

Near Fredericksburg, Virginia (Fig. 2, No. 41) Pavlides et al. (1979) and Pavlides (1980) reported strongly to imperceptibly foliated rocks in the granitic Falmouth Intrusive Suite. These rocks have been dated by zircon Pb-U and whole-rock Rb-Sr methods at 300-340 Ma. The Falmouth intrudes the Late Ordovician-Silurian(?) Arvonia slate. Near Baltimore, Md. (Fig. 2, No. 42) the work of Sinha et al. (1980) and Grauert (1973) indicates that the Gunpowder Granite is a foliated pluton of 300-330 Ma age. These relations provide evidence for extending the Alleghanian metamorphic and ductile deformation zone through the northern Virginia Piedmont into the easternmost Piedmont of Maryland.

Constraints on the Alleghanian metamorphic conditions elsewhere in the Piedmont are provided by the 320 to 269 Ma-old granites. A string of these occur immediately to the west of the Alleghanian metamorphic front and intrude into the low-grade rocks of the Carolina slate belt. Most have contact metamorphic aureoles (Speer et al., 1980; Speer, 1981). The metamorphic mineral assemblages and mineral chemistries indicate that the present erosional surface was at depths corresponding to 4.5 to 5.5 kbar at the time of intrusion. Ambient country rock temperatures were those of the greenschist facies or lower. These temperatures and pressures correspond to normal continental geothermal gradients. These granite plutons, as well as others in this age range farther west, contain zones of ductile deformation. The most strongly deformed are the Palmetto and 269 Ma-old Siloam plutons, Georgia (Speer et al., 1980) which are adjacent to regional ductile deformation zones. This would suggest that in the Inner Piedmont during the Alleghanian, ductile deformation was only possible in local shear zones.

Dallmeyer et al. (1981) presented a revised model for the Late Paleozoic - Early Mesozoic cooling history of the central Inner Piedmont of Georgia. They used $^{40}Ar/^{39}Ar$ data on country rock hornblende and biotite which they combined with petrologic constraints on P-T history from host gneisses and from the Stone Mountain and Elberton granites. They suggested that the peak metamorphic mineral assemblages of the eastern Inner Piedmont correspond to temperatures of 725 ± 25°C and 7-9 kbars. After some uplift these gneisses were intruded by the Stone Mountain Granite at a depth of 10-12 km while the country rock was maintained at a temperature above 550°C (hornblende $^{40}Ar/^{39}Ar$ data). Their calculations suggested that the geothermal gradient increased from about 30°C/km at the peak of regional metamorphism to 40°C/km at the time of Stone Mountain intrusion (325 Ma). They related this increase in geothermal gradient to Hercynian (Alleghanian) tectonothermal activity in the eastern Piedmont. Possible problems with this interpretation include: (1) Some of the P-T constraints are broad limits that do not yield convincing gradients; (2) uplift and erosion could easily outstrip deep cooling of the Inner Piedmont, in which case geothermal gradients would increase without addition of new heat sources; and (3) geothermal gradients can vary sharply from area to area because of local variations in thermal conductivity and heat generation of rocks.

From the foregoing, an Alleghanian ductile deformation front exists along most of the eastern Piedmont of the central and southern Appalachians. West of the front (Fig. 2) localized zones of mild Alleghanian deformation are imprinted on at least two ca. 270-350(?) Ma granites (i.e., Siloam and Palmetto) principally where they are cut by ductile shears. Additionally, the $^{40}Ar/^{39}Ar$ biotite and hornblende ages of 236 and 300 Ma, respectively, reported by Dallmeyer (1978) along the eastern margin of the Inner Piedmont in Georgia suggest that the Inner Piedmont was passing through a temperature range of 300-500°C at that time. However, the mild strain effects displayed in the Siloam (269 Ma) and the Palmetto granites suggest that the amphibolite facies Inner Piedmont and Charlotte belt terranes, although still quite hot, had already cooled to a sufficiently rigid state so that the Alleghanian stress field was inadequate to develop much strain in them. Apparently even high grade rocks lose much of their ductility in cooling below their prograde thermal maximum. This is probably due to their inability to recrystallize freely in the drier and lower energy environment that results from cooling.

Farther north along the Va.-N.C. State line (Fig. 2, No. 43) in Charlotte belt rocks G. Russell (unpub. data) has found ca. 260 Ma biotite-whole rock ages of biotite gneiss. These are interpreted as cooling ages when the biotite passed through about the 300°C isotherm. The data of Dallmeyer and Russell suggest that west of the Alleghanian ductile deformation front, relatively hot rocks in the crystalline terrane were deforming by differential uplift and perhaps by broad folding during the Alleghanian. As discussed previously, the Brevard splayed off of the sole fault at this time and the Blue Ridge and the Valley and Ridge stratigraphy farther west was deformed by folding and thrusting under subgreenschist conditions. Thus the Piedmont and Blue Ridge west of the Alleghanian ductile deformation front transmitted stress to the Valley and Ridge largely as a semibrittle block.

Summary

Summaries of each orogeny follow the sections on Taconic, Acadian and Alleghanian. In this summary we mention some generalities for the system as a whole.

The Taconic was the most widespread metamorphic event affecting rocks at the present level of exposure. It seems probable that it affected rocks now overprinted by the Alleghanian event and it is possible that rocks even farther east, under the Atlantic Coastal Plain, were metamorphosed at this time. Given the present data base, this seems to be the most intense thermal and deformational event experienced by the central and southern Appalachians during the Paleozoic.

The Acadian is a more localized and generally lower grade event at the present level of exposure. It seems possible that Acadian thrusting, uplift and cooling may have localized its effects, and that no large, additional increment of heat from deep subcrustal sources was available. This is strongly suggested by the data of Dallmeyer (1978) which indicate that the high temperatures of the Georgia Inner Piedmont may have been carried over from the Taconic Orogeny.

Alleghanian amphibolite facies regional ductile fabric development and metamorphism is localized along the eastern margin of the Piedmont and extends in the basement an undetermined distance under the Atlantic Coastal Plain. It presents a rather abrupt ductile deformation front against low to medium grade rocks that cooled during Taconic and Acadian(?) orogenies. Upwelling of a deep hot infrastructure or renewed crustal heating may be required to explain its development.

The general southeastward migration of thermal metamorphic events with time is consistant with the trend of ages of emplacement of Appalachians plutons. This apparent migration is mitigated to some extent by the extensive westward transport of Piedmont and Blue Ridge rocks during successive orogenies. This transport allowed the western margins of thrust plates to move out of the zone of succeeding ductile deformations.

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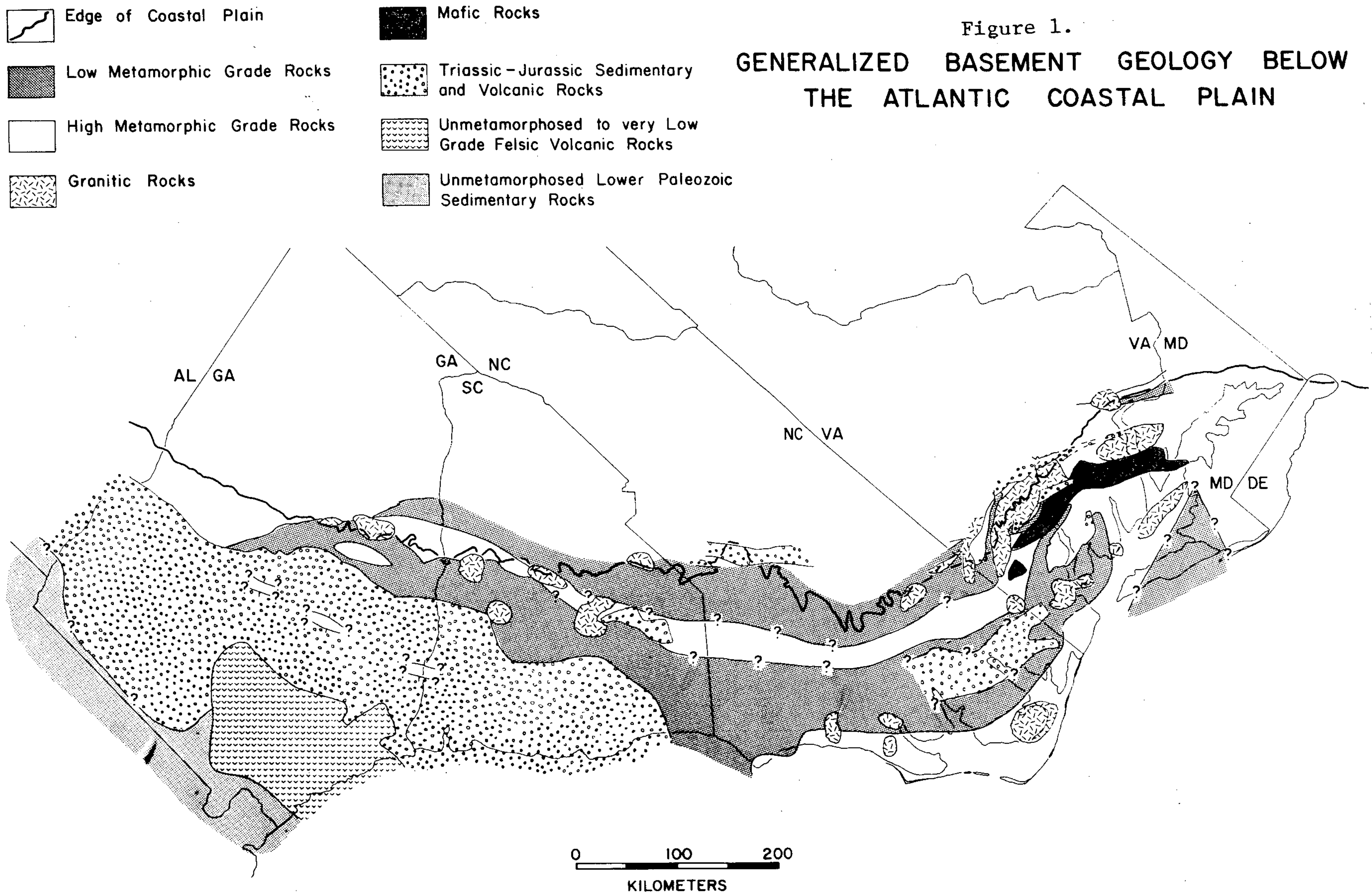
GENERALIZED GEOLOGIC MAP OF BASEMENT BELOW ATLANTIC COASTAL PLAIN

Richard J. Gleason

A generalized geologic map of the basement rocks below the Atlantic Coastal Plain sediments has been drawn (Figure 1), based on an interpretation of drill core sample descriptions (Gleason, 1979a,b), and aeromagnetic (Daniels and Zietz, 1978; Zietz _et al_., 1977) and Bouguer gravity data (VPI&SU, unpub.). This map is a preliminary version; additional drill hole data and refinement of the interpretation of potential field data will result in continuing modifications of the map.

Because of the limited number of basement samples available for direct analysis and the unreliabililty of many of the published sample descriptions, which include drillers' logs, the map is schematic. Areas indicated as "low metamorphic grade" include rocks characterized by greenschist facies or lower grade metamorphic assemblages. "High metamorphic grade" areas include rocks with metamorphic assemblages above the "biotite-in" isograd. The areas of metamorphic rocks are shown as long, continuous belts, but parts of these belts may exhibit structural complexities which juxtapose high and low grade rocks. Areas indicated as "granitic rocks" include plutonic rocks of granitic to tonalitic composition. These areas have been mapped on the basis of drill samples and Bouguer gravity data. The granitic rocks are probably more extensive in the Coastal Plain basement than indicated on the map, but their presence is undetectable due to lack of density contrast with surrounding country rocks. A major positive aeromagnetic anomaly produced by basement rocks beneath the coastal plain in Maryland and Virginia (Zietz _et al_., 1977) has been interpreted as a belt of mafic rocks. Drill samples from this area include "soapstone," metagabbro, and amphibolite. Areas of "Triassic-Jurassic rocks" are believed to be underlain by volcanic and sedimentary rocks in rift basins similar to those exposed along eastern North America, west of the Coastal Plain onlap. The large region of Triassic-Jurassic basement in Georgia and South Carolina, as well as the "felsic volcanic rocks" of Georgia and the "Paleozoic sedimentary rocks" of southern Georgia and northern Florida are discussed by Gleason (in press).

Figure 1.
GENERALIZED BASEMENT GEOLOGY BELOW THE ATLANTIC COASTAL PLAIN



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COMPARISON OF HEAT FLOW IN HIGH AND LOW GRADE METAMORPHIC BELTS BELOW THE ATLANTIC COASTAL PLAIN

Richard J. Gleason

The Piedmont province and its eastward extension buried below the Atlantic Coastal Plain sediments is composed of alternating belts of high-grade and low-grade metamorphic rocks. Glover (pers. comm.) has postulated that the high-grade belts evolved as synmetamorphic antiformal upwarps and that adjacent low-grade belts formed as synformal downwarps. In this model, the high-grade upwarps would have acted as loci for the lateral and vertical migration of large granitic masses. If such a theory is correct, it implies that there may be higher background heat flow in high-grade belts than in low-grade belts, due to the presence of large granitic masses within the cores of the high-grade belts.

To test this hypothesis, heat flow values measured from drill holes in the Atlantic Coastal Plain were compared to crystalline basement lithologies which were predicted based on a generalized Atlantic Coastal Plain basement geology map. Locations of heat flow sites are superimposed on this basement map in Figure 1. Heat flow sites overlying syn- to post-metamorphic granites have been omitted from the correlation. The average heat flow value from sites assumed to overlie low-grade rocks is 55.8mW/m^2 ($n = 15$, $SD = 9.6$). Average heat flow from sites assumed to overlie high-grade rocks is 53.8 mW/m^2 ($n = 18$, $SD = 13.8$). These values imply no significant difference in heat flow between high-grade and low-grade metamorphic belts below the Atlantic Coastal Plain.

The validity of this conclusion is limited by the accuracy of the generalized Atlantic Coastal Plain basement map. As is discussed elsewhere in this report, the basement map is based on interpretation of available basement samples, magnetic, and gravity data. Heat flow values, on the other hand, were obtained from measurements of thermal conductivities and geothermal gradients in sedimentary sections overlying the basement. Basement samples were available for only ten of the heat flow sites. Therefore, inaccuracies in the lithologic predictions obtained from the basement map for the other heat flow sites would bias the results of the high-grade/low-grade comparison.

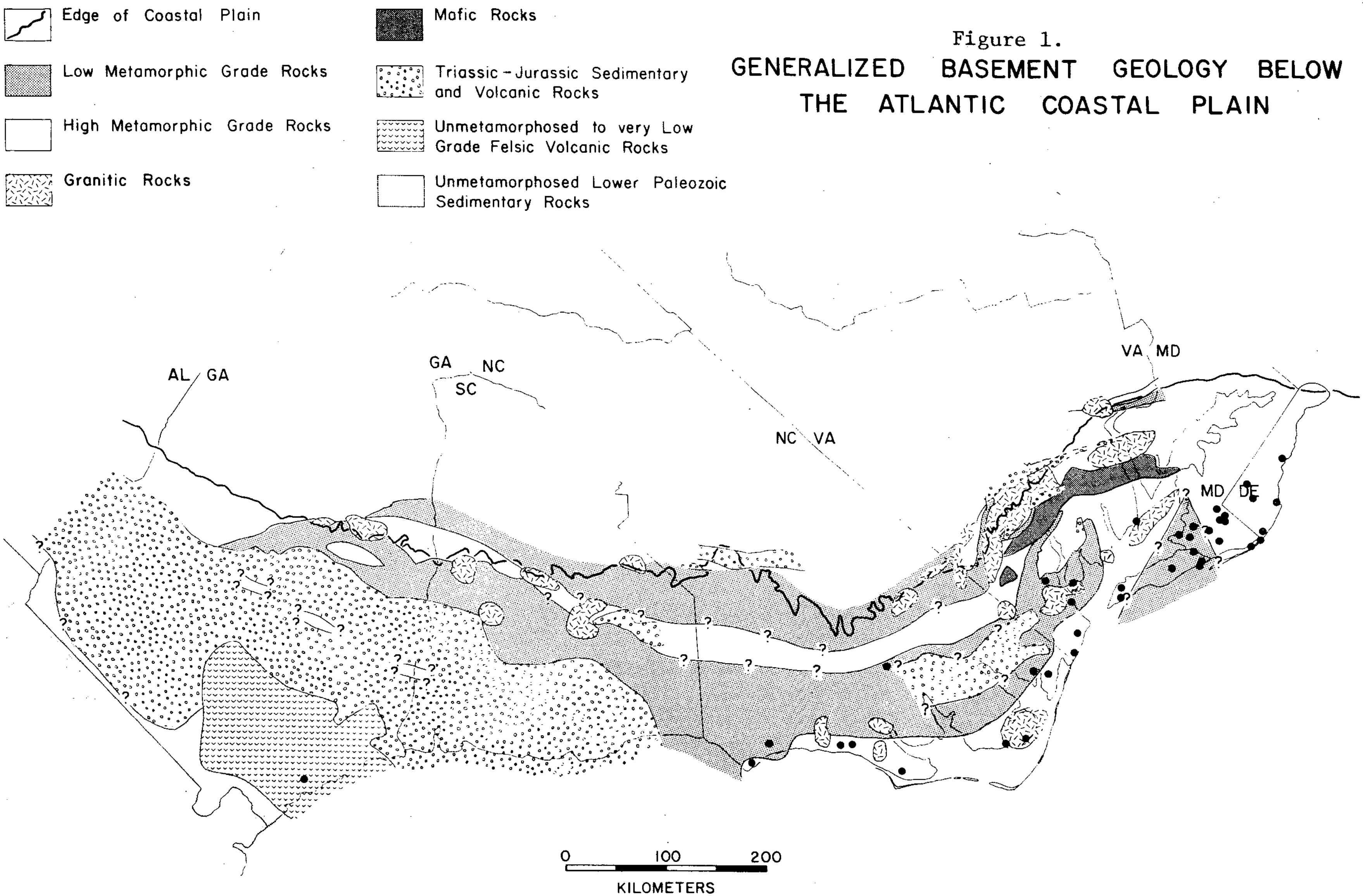

Figure 1.
GENERALIZED BASEMENT GEOLOGY BELOW THE ATLANTIC COASTAL PLAIN

The comparison of average heat flow in contrasting metamorphic belts has not been confirmed as yet in the Piedmont province. Unfortunately, nearly all of the VPI&SU heat flow measurements in the Piedmont were obtained from granites and not from metamorphic country rocks. Until a concerted effort is made to collect heat flow data directly from high- and low-grade metamorphic belts, it will remain impossible to confirm or refute a correlation between these contrasting metamorphic belts and heat flow.

STRUCTURE CONTOUR MAP OF BASEMENT BELOW NORTHERN ATLANTIC AND SOUTH CAROLINA COASTAL PLAINS

R. J. Gleason

A factor in the evaluation of geothermal resource potential in the Atlantic Coastal Plain is the thickness of Coastal Plain sediments at any given location. Sediment thickness is one criterion in the determination of the insulating capability of the sedimentary blanket overlying a basement heat source. The thickness of Coastal Plain sediments is also an important consideration in the calculation of the cost of drilling to basement. Sediment thickness may be obtained from a structure contour map of the crystalline basement surface by calculating the difference between the elevation of the ground surface and that of the basement surface. Structure contour maps of the basement surface below the North Carolina and Georgia Coastal Plains were presented in a previous report (Gleason, 1979). Similar maps of the basement in the northern Coastal Plain (New Jersey to Virginia) and in South Carolina are presented in this report as Figures 1 and 2, respectively.

In the northern Atlantic Coastal Plain (Fig. 1), basement drill-hole data is most abundant along the shallow western edge; only 69 of 391 drill-holes are located where basement elevations are less than -500'. Seismic refraction data (Ewing et al., 1937, 1939, 1940; Hansen, 1978) and reflection data (VPI&SU, unpub.; Jacobeen, 1972; Hansen, 1978) augment the drill-hole data in the deeper parts of the Coastal Plain. In South Carolina (Fig. 2), only 37 basement drill-holes were obtained from the literature. An additional 42 basement data points were obtained from seismic refraction studies of Bonini and Woollard (1960).

The structure contour maps presented as Figures 1 and 2 represent updated versions of earlier maps such as those of Flawn (1967), Maher (1971), and Brown et al. (1972). The sparseness of data points on these maps requires that the basement structure contours be generalized and commonly inferred. The addition of drilling and seismic data will result in continuous updating and modification of these maps in the future.

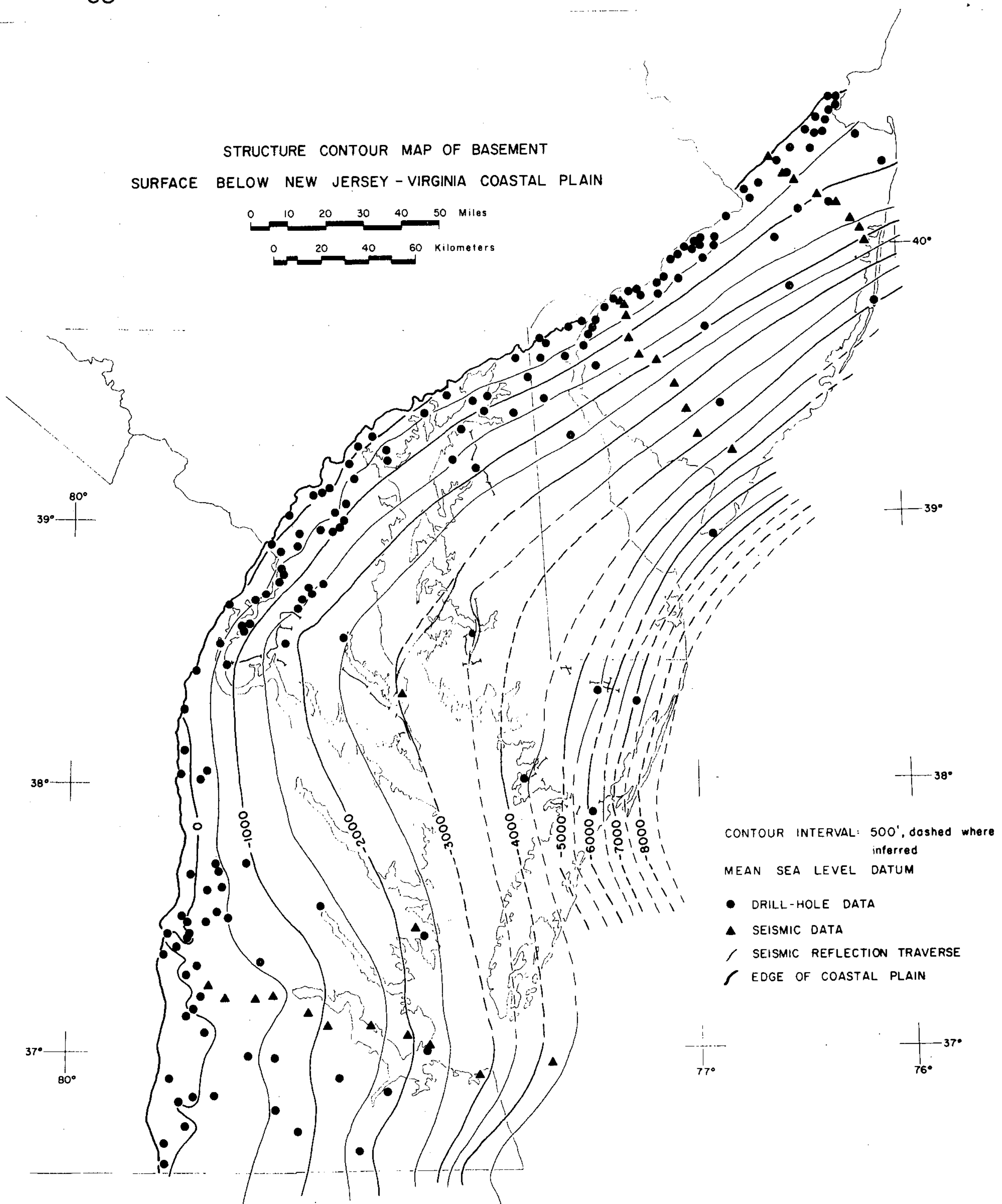

Figure 1. Structure contour map of basement surface below New Jersey - Virginia Coastal Plain.

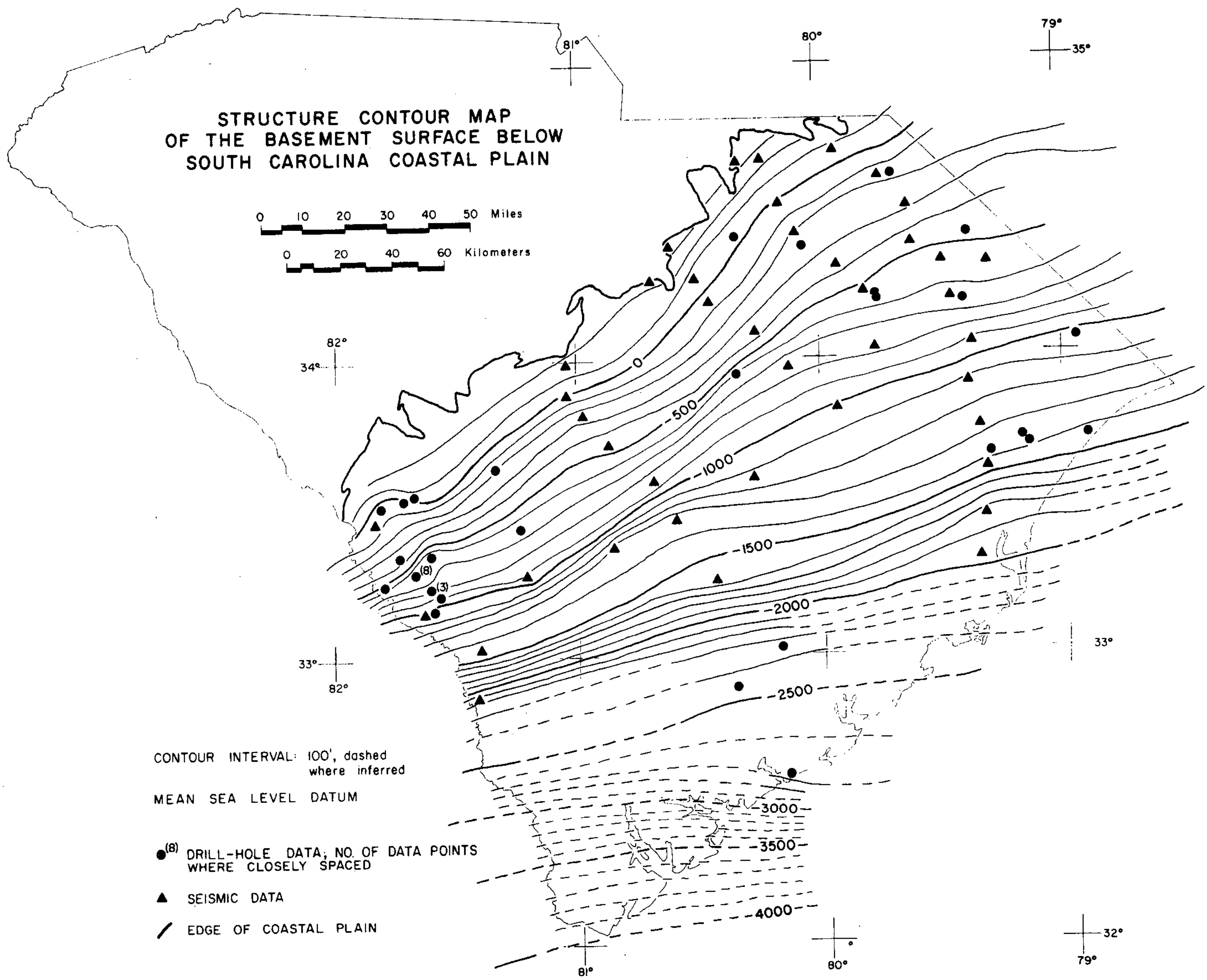

Figure 2. Structure contour map of the basement surface below South Carolina Coastal Plain.

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BASEMENT ROCKS BENEATH ATLANTIC COASTAL PLAIN: ISLE OF WIGHT, VIRGINIA

Richard J. Gleason

Introduction

The rationale for locating geothermal test drilling sites in the Atlantic Coastal Plain is based on a model (Costain et al., 1980) which predicts that granitic plutons with high contents of radioactive trace elements relative to surrounding country rock underlie the Coastal Plain sediments. As a test of this model, a drill-site (C25A) was located over the center of a circular negative Bouguer gravity anomaly near Portsmouth, Virginia (Fig. 1), thought to indicate the presence of a buried, low-density granitic pluton. The basement core obtained from site C25A is a post-metamorphic granite giving a whole-rock Rb-Sr isochron age of 263 ± 24 m.y. (Russell and Russell, 1980).

A second drill site (C26) at Isle of Wight, Va. (latitude 36°54.5', longitude 76°42.2') was located 20 kilometers from the center of the Portsmouth gravity anomaly (Fig. 1) in order to sample and measure the heat flow of the country rocks surrounding the buried pluton at Portsmouth. Drill site C26 was drilled to 325 m (1065 ft.) by Gruy Federal, Inc. between September 20 and 26, 1978. From April 1, 1980 to May 6, 1980, the hole was deepened to 512 m (1680 ft.) by VPI&SU. Core samples were obtained from 406-413 m and from 424-512 m.

The core samples of the interval from 406 to 413 m resemble saprolitized granitoid or heterogeneous arkosic sandstone. Drilling resistance increased noticeably at 424 m (1390 ft.), corresponding to a sharp spike in the gamma log (Fig. 1), and crystalline basement was cored beginning at this depth. The base of the Coastal Plain sediments has therefore been picked tentatively at 424 m.

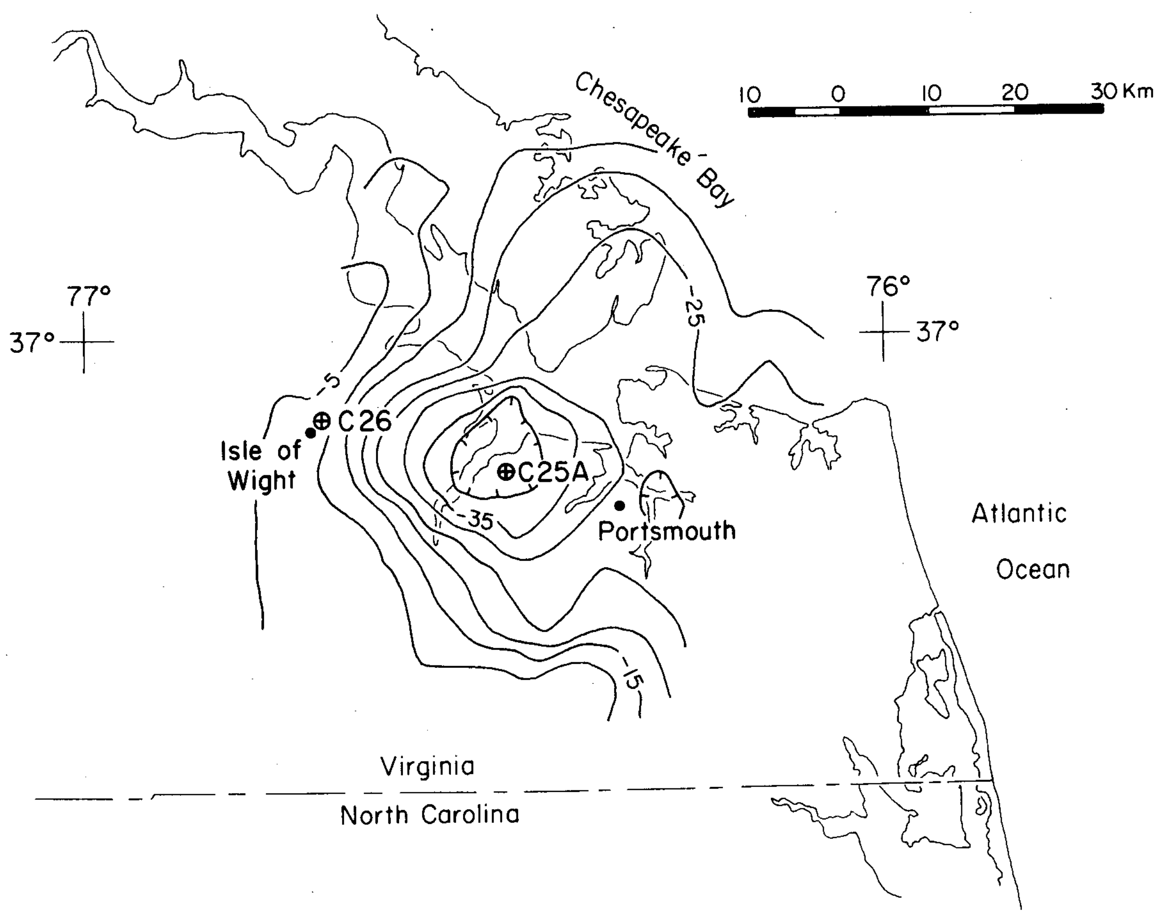

Figure 1: Location of drill sites C25A ad C26, in SE Virginia, super imposed on Bouguer gravity contours from Johnson (1977).

Macroscopic Description

Lithologic and gamma logs are presented in Figure 2. The basement consists primarily of a fine- to medium-grained (1-4 mm) foliated biotite granitoid which is cut locally by a more leucocratic facies. There are several fine-grained (<1 mm), weakly foliated melanocratic sections of the core ranging from 25 cm to 7 m in length, but contacts with the granitoid generally are fractured, and any intrusive relation to the granitoid is not readily apparent.

Figure 2 also shows the zones of alteration in the core. The core is weathered from 424-427 m. Below this zone, alteration is controlled by intricate fine fractures along which the feldspar in the core has been colored a pinkish-red. The core is brecciated along many of these fractures, and quartz or carbonate veinlets are associated with the fractured, altered zones.

Granitoid Rocks

The granitoid rocks of the drill core range modally from tonalite to monzogranite (Fig. 3). The tonalite is composed of quartz, turbid plagioclase, chloritized biotite, and accessory titanite, zircon, and opaques, with traces of microcline perthite. Color index ranges from 13 to 20. The granodiorite is mineralogically similar but contains 8-21% microcline and less biotite, with color index ranging from 3 to 10%. The monzogranite is composed of quartz, turbid plagioclase, microcline (20-32%), and accessory biotite, rutile, and opaques. Color index is 4-5.

Microscopic textures and mineral assemblages of the granitoid rocks were modified during a post-emplacement, lower greenschist facies metamorphism. Quartz grains display undulatory extinction under polarized illumination, and subgrain development in quartz and feldspar reduced microscopic grain size to 0.25-1.0 mm. Randomly oriented biotite was mostly altered to chlorite, which crystallized in a preferred orientation, imparting a weak to moderate foliation. Feldspars were sericitized and saussuritized and are turbid because of inclusions of dusty hematite(?). Porphyroblastic epidote and white mica up to 1 mm in size were produced during metamorphism. Quartz and carbonate veinlets up to 1 mm in width postdate the greenschist metamorphic assemblage.

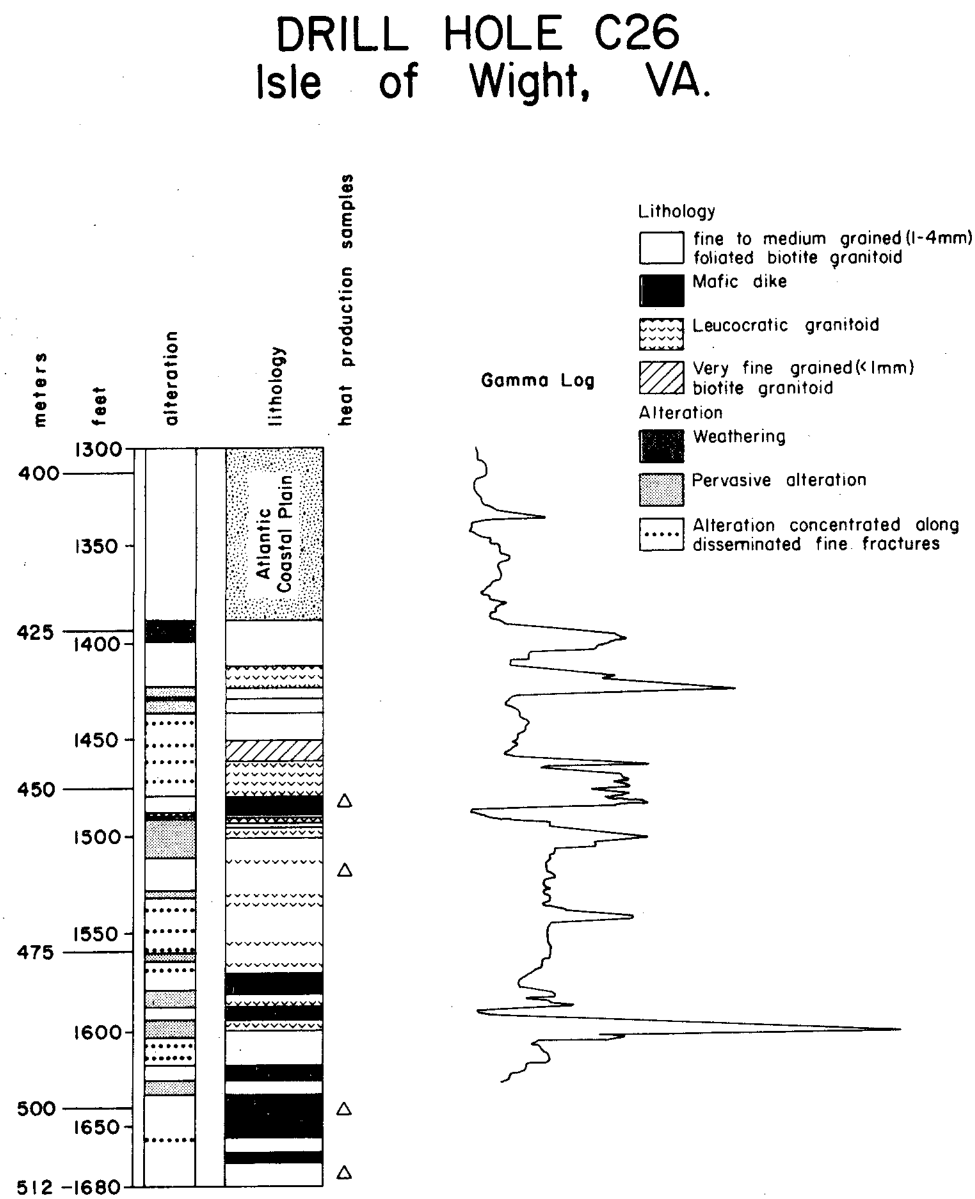

Figure 2: Lithologic and gamma logs from basement portion of geothermal test drill-site C26, Isle of Wight, Virginia.

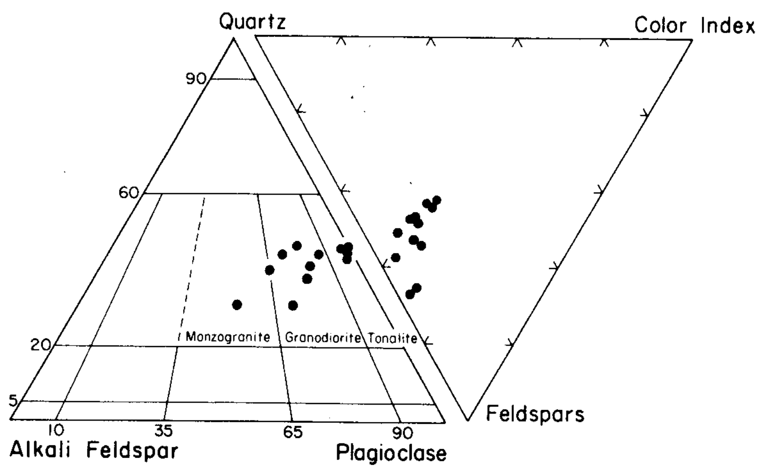

Figure 3: Modal classification (after Streckeisen, 1976) and quartz-feldspar-color index ternary diagram of basement samples from drill site C26, Isle of Wight, Virginia.

Mafic Dikes

The intrusive relation of the mafic sections to the host granitoid rocks is visible microscopically. Locally, these sections show chill textures near contacts with the granitoid, with feldspar microlites set in micro- to cryptocrystalline groundmass, whereas rocks from the centers of the wider sections coarsen to 0.5-1.0 mm in grain size. Rare small (1-2 mm) pods of the granitoid are present as xenoliths in the mafic dikes.

The dikes are pre-metamorphic and were subjected to the same greenchist facies metamorphism that affected the granitoid. The original mineralogy was partially obscured by this metamorphism, but relict mineralogical and textural evidence suggests that the dikes were basaltic, composed of calcic plagioclase, pyroxene, and hornblende, with pyrite, apatite, ilmenite-magnetite, quartz, calcite, and spinel as accessories. Subsequent metamorphism altered the plagioclase microlites to tremolite and the coarser plagioclase to turbid mats of white mica and epidote. Hornblende and pyroxene were altered to chlorite, actinolite, and epidote porphyroblasts; in scarce occurrences, the pyroxene was altered to serpentine and talc. Minor zeolite-adularia veinlets less than 1 mm in width postdate the greenschist overprint and reflect a later hydrothermal event.

Discussion

The granitoid rocks which form the basement beneath the Atlantic Coastal Plain in Isle of Wight, Virginia, originated as a deep-seated intrusion of biotite tonalite-granodiorite. With progressive crystallization of this magma, the last phase was a leucocratic monzogranite-granodiorite which intruded the earlier phases. The lack of sharp contacts between the earlier and later phases suggests that the granitoid was not fully crystallized until after the intrusion of the late monzogranite.

Following crystallization of the granitoid, the basaltic dikes were intruded. The presence of spherical, calcite-filled features resembling amygdules in one of the dikes (sample from 452 m) suggests that the emplacement of the dikes was shallow, requiring an uncertain time interval between the granitoid plutonism and the dike emplacement, in order for uplift or erosion to expose the formerly deep-seated pluton to near-surface conditions.

The greenschist metamorphic overprint evident in both the granitoid and mafic dikes shows that the basement rocks were subjected once again to burial sufficient to produce an epidote-chlorite-white mica assemblage in the granitoid rocks and an actinolite-epidote-chlorite assemblage in the mafic dikes. Strained quartz and feldspar and a weak foliation of metamorphic chlorite and white mica in the granitoid rocks indicate that the greenschist metamorphic overprint was a result of a progressive dynamothermal event and not a retrograde deuteric or hydrothermal effect following the injection of the basaltic dikes. Following the greenschist facies metamorphism, minor zeolite-facies hydrothermal alteration formed veinlets of an unidentified zeolite and a low-temperature potassium feldspar in the dikes and quartz-carbonate veinlets in the granitoid rocks. The macroscopically visible alteration indicated in Figure 2 was a result of this hydrothermal activity.

The basement at Isle of Wight is lithologically distinct from the Portsmouth granite and has undergone deformation and greenschist metamorphism assumed to have predated the 263 ± 24 Ma age of the Portsmouth pluton. Heat production from the Portsmouth granite was measured as 4.09 ± 0.3 uW/m^3, and heat flow through the sedimentary section overlying the granite was measured as 83.5 ± 10 mW/m^2 (Perry, pers. comm., 1981). The heat production of the Isle of Wight basement rocks averages 1.1 ± 0.5 uW/m^3, corresponding to a heat flow measured in the overlying sedimentary section as 54.6 ± 6.6 mW/m^2 (Perry, pers. comm., 1981). These lower heat production and heat flow values of the metamorphosed country rocks contrast with the higher values of the post-metamorphic Portsmouth granite, and provide verification of the radiogenic heat model of Costain et al. (1980).

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DESCRIPTION OF THE BASEMENT GRANITOID ROCKS OF DRILLHOLE C25A, NEAR PORTSMOUTH, VIRGINIA

J. Alexander Speer

A -40 mgal Bouguer gravity anomaly near Portsmouth, Virginia (Fig. 1) was believed to indicate the presence of a granitic body beneath the Atlantic Coastal Plain. A drill-site (CP25A) was located near the center of the circular anomaly at latitude 36°51.01' and longitude 76°29.83' to obtain heat flow and granite samples. The hole was drilled through the coastal plain sediments to 557 m (1828 ft.) by Grüy Federal, Inc. during December 1978. From January to April 1979, the hole was deepened to 611 m (2005 ft.) and a continuous, 1-1/2 inch diameter core obtained from 557 to 611 m (1828-2005 ft.). The basement rock obtained from drillcore CP25A is a post metamorphic granitoid which is described here as the Portsmouth granite. A whole-rock, Rb-Sr isochron age of 263 ± 24 Ma and initial Sr^{86}/Sr^{87} ratio of 0.7076 ± 0.0012 for this core have previously been reported by Russell and Russell (1980). Two biotite-whole rock isochrons give ages of 267 ± 7 and 267 ± 5 Ma.

The CP25A basement core is composed of a massive, orange-pink, coarse-grained, inequigranular granitoid rock. A point count of a slabbed and stained core shows the rock to be a monzogranite, with 28.6% quartz, 30.5% K feldspar, 37.8% plagioclase, and 3.1% biotite + accessory minerals. The color index of the monzogranite, based on visual estimates, varies from less than 1 to 5 along the length of the core. The alkali feldspars are an- to subhedral orange crystals up to 15 mm long. The plagioclases are euhedral to subhedral crystals up to 7 mm long and are either orange or white. The quartz is grey and anhedral, with grains less than 6 mm in diameter. Biotite flakes, 5 mm or less in length, are the most abundant mafic mineral. Primary accessory minerals of the granite include apatite, zircon, allanite, ilmenite, magnetite, pyrite, molybdenite, thorite, fluorite, and monazite. Secondary minerals are chlorite, muscovite, rutile, titanite and epidote. Average density of the granite is 2.63 gm/cm^3. The U, Th, and K contents and the calculated heat productions are given in Table 1 (Fig. 2).

The granite is cut by abundant quartz, carbonate, and muscovite veins occupying lengths of the core up to 7 cm. Disseminated replacements of the granite termed "veined gra-

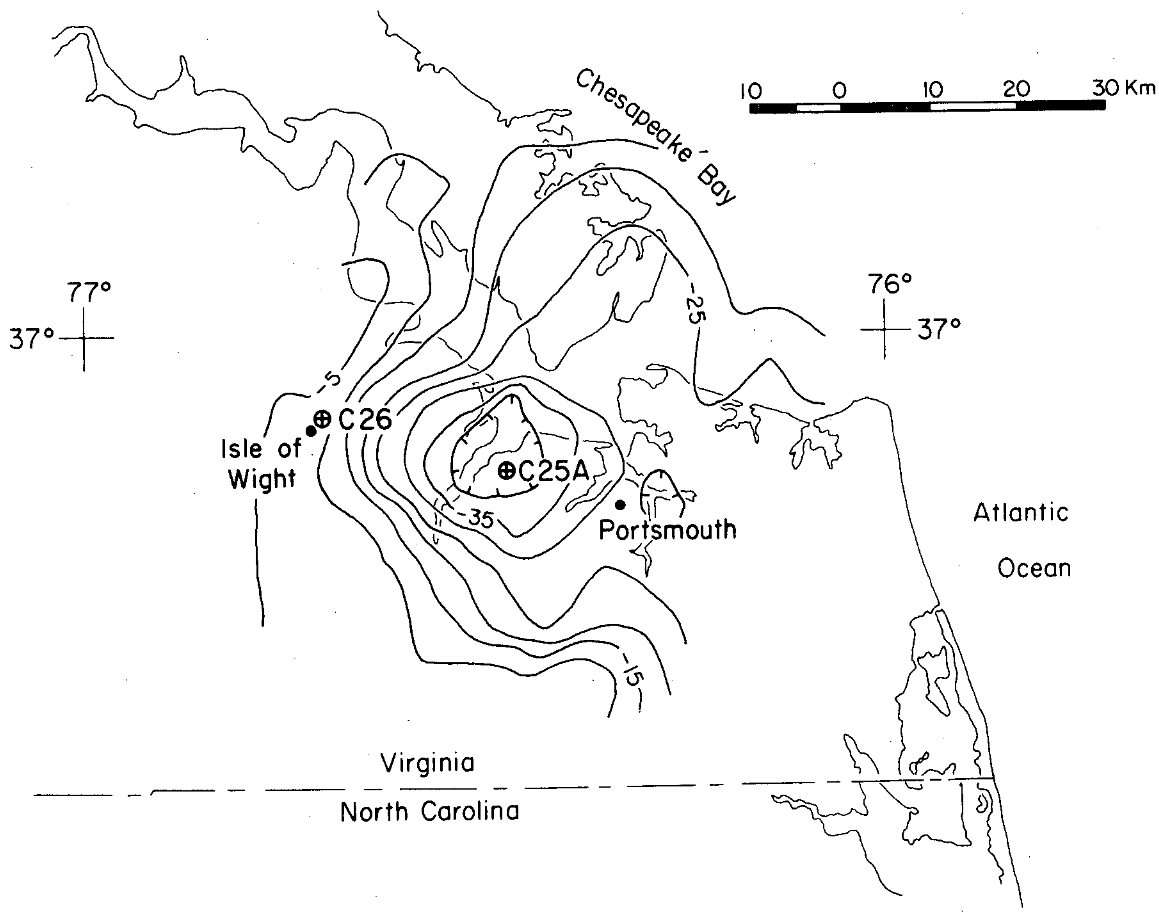

Figure 1. Location of Drill site CP25A, located in Nansemond Co., Virginia, superimposed on the Bouger gravity contours from Johnson (1977).

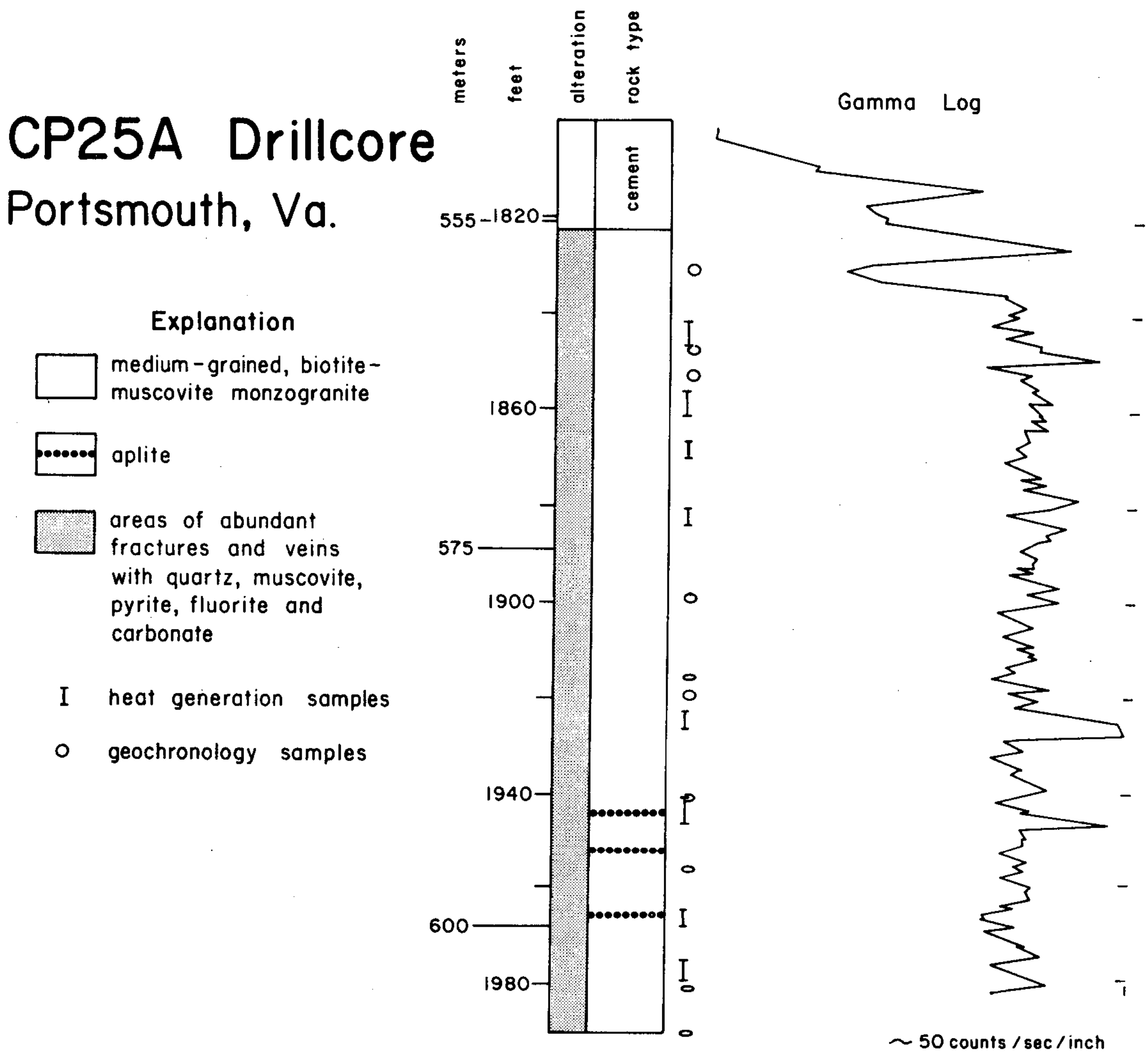

Figure 2. Lithologic and gamma logs for the basement portion of the drillsite CP25A, Nansemond Co., Virginia.

Table 1. U, Th, and K contents and heat generation for Portsmouth granite (CP25A), Portsmouth, Va. (Perry *et al*., 1980).

sample number	interval	U,ppm	Th,ppm	K,wt.%	heat generation 10^{-13}cal/cm/sec	heat generation 10^{-6}/W/m^3
562	1843-1847	9.0	24.1	4.0	10.3	4.3
566	1857-1861	9.1	20.6	3.4	9.6	4.0
569	1866-1870	8.0	20.6	3.7	9.1	3.8
573	1880-1884	7.8	17.9	4.0	8.5	3.6
586	1922-1926	9.8	19.1	3.5	9.8	4.1
591	1940-1945	9.6	23.8	3.7	10.5	4.4
599	1964-1968	9.2	24.8	3.4	10.4	4.4
602	1974-1979	8.5	20.9	3.5	9.4	3.9
	average	8.9	21.5	3.6	9.7	4.1

nites" range in length up to 10 cm. The muscovite veins or veined granites consist largely of muscovite up to 2 cm across with lesser amounts of albite and quartz and variable amounts of fluorite, pyrite, pyrophanite, chalcopyrite, and thorite. The carbonate-veined granites contain calcite, Mn-Fe calcite, siderite, barite, Ni-pyrite, and uraninite.

Mineralogy

Biotite. The biotite, pleochroic tan to brown, is extensively replaced by chlorite, rutile, and muscovite. Microprobe analyses (Table 2, Fig. 3) show the biotites to be aluminous, intermediate phlogopite-annite solid solutions with Fe(total)/(Fe(total) + Mg) between 0.48 and 0.58. The biotites are fluorian with F/(F + Cl + OH) between 0.17 and 0.27.

Feldspar. The alkali feldspar in the Portsmouth granite is microcline with locally developed microperthite. The average composition is Or(94) with a range of Or(96) to Or(92) (Table 3). Ca contents are less than An(0.2) and Ba contents range up to Cn(1.18). The plagioclase has weak oscillatory zoning with local albite rims. Compositions vary between An(21) and An(7), potassium content is less than Or(1.3). Plagioclase in the veined granite is albite, An(2.7) to An(0.5). Most plagioclase in the Portsmouth granite is replaced by variable amounts of muscovite ± epidote ± carbonates.

Muscovite. Muscovite has a variety of occurrences in the granite: in the matrix, as replacement of plagioclase, and in the muscovite veins and veined granite. They are phengites (Table 4) with Fe/(Fe + Al(VI) + Mg + Ti) up to 0.17 and Mg/(Fe + Al(VI) + Mg + Ti) of 0.13. They contain only a minor paragonite component with Na/(Na + K) less than 0.05. The muscovites' low totals and low alkali sums indicate that they are hydromicas in part or, more likely, Li-muscovites, which is suggested by their opitcal zoning. The muscovites contain variable amounts of F, ranging between 0.2 and 1.48 weight percent.

Accessory minerals. The apatites are fluorapatites. The chlorites (Table 5) are pycnochlorites and ripidolites (Hey, 1954) with Fe/(Fe + Mg) of 0.44 to 0.50. An analysis of an epidote from CP25A 1959.3 yields $Pistacite_{32}$ (Table 6). Ilmenites in the granites are nearly ideal compositions. An ilmenite from a veined granite is pyrophanite with 62% $MnTiO_3$.

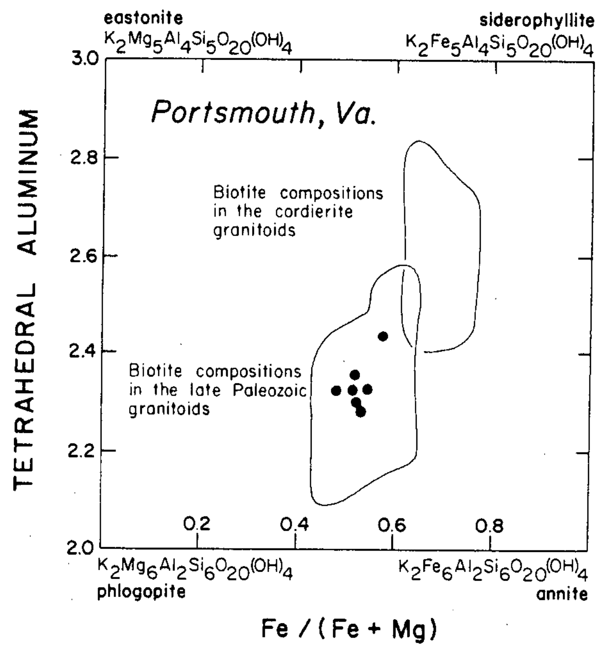

Figure 3. Average biotite compositions for the Portsmouth granite projected onto the phlogopite-annite-eastonite- siderophyllite quadrilateral.

Discussion

The well-defined, circular gravity anomaly of the Portsmouth granite, its undeformed texture, and concordant biotite-whole rock ages suggest that it is a postmetamorphic pluton. The evolved nature of the granite, indicated by its leucocratic nature, F-rich silicate minerals, extensive late magmatic or subsolidus reactions and mineral assemblages, abundant quartz veins and high U contents, suggests that the drillhole encountered the late crystallizing and fluid-rich portion, or cupola, of the magma chamber.

References

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Perry, L. D., S. P. Higgins, and M. L. McKinney, 1980: Heat Flow and Heat Generation in the Atlantic Coastal Plain, pp. B114-B153 in J. K. Costain and L. Glover, III, Evaluation and Targeting of Geothermal Resources in the Southeastern United States, National Technical Information Service VPI&SU-78ET-27001-8.

Russell, G. S. and C. W. Russell, 1980: Rb-Sr Isotopic of granitic core from Portsmouth, Virginia, pp. A64-A68 in J. K. Costain and L. Glover, III, Evaluation and Targeting of Geothermal Energy Resources in the Southeastern United States, National Technical Information Service VPI&SU-78ET-27001-8.

SUPER RECAL

TABLE 2. AVERAGE BIOTITE ANALYSES(H2O INPUTTED), PORTSMOUTH, VA.

	1	2	3	4	5	6	7
SIO2	37.06	36.30	37.18	37.69	36.63	37.40	37.36
TIO2	2.40	2.60	2.09	2.16	2.70	2.57	2.41
A2O3	15.44	18.16	15.28	15.05	15.09	15.54	15.28
FEO	20.43	19.13	18.18	20.29	19.31	19.15	19.76
MNO	0.55	0.52	0.76	0.40	0.50	0.57	0.47
MGO	9.78	9.54	11.57	10.27	10.30	10.46	10.33
CAO	0.04	0.21	0.0	0.05	0.0	0.08	0.07
BAO	0.27	0.31	0.07	0.22	0.06	0.25	0.17
NA2O	0.07	0.12	0.05	0.26	0.07	0.08	0.07
K2O	9.78	9.34	9.79	10.27	10.58	9.91	9.72
F	1.58	1.73	2.27	1.69	1.39	1.69	1.60
CL	0.05	0.07	0.08	0.05	0.04	0.04	0.08
H2O	3.16	3.09	2.85	3.17	3.23	3.15	3.16
SUM	100.61	101.12	100.17	101.57	99.90	100.89	100.48
-O= F+CL	0.68	0.74	0.97	0.72	0.59	0.72	0.69
SUM	99.93	100.38	99.20	100.85	99.31	100.17	99.79

SI	5.676	*	5.486	*	5.680	*	5.722	*	5.644	*	5.680	*	5.703	*
AL	2.324	8.000	2.514	8.000	2.320	8.000	2.278	8.000	2.356	8.000	2.320	8.000	2.297	8.000
AL	0.463	*	0.720	*	0.431	*	0.414	*	0.383	*	0.461	*	0.451	*
TI	0.276	*	0.295	*	0.240	*	0.247	*	0.313	*	0.294	*	0.277	*
FE	2.617	*	2.418	*	2.323	*	2.576	*	2.488	*	2.432	*	2.522	*
MN	0.071	*	0.067	*	0.098	*	0.051	*	0.065	*	0.073	*	0.061	*
MG	2.233	5.660	2.149	5.649	2.635	5.726	2.324	5.612	2.365	5.615	2.368	5.628	2.350	5.661
CA	0.007	*	0.034	*	0.0	*	0.008	*	0.0	*	0.013	*	0.011	*
NA	0.021	*	0.035	*	0.015	*	0.077	*	0.021	*	0.024	*	0.021	*
K	1.911	*	1.800	*	1.908	*	1.989	*	2.079	*	1.920	*	1.892	*
BA	0.0	1.938	0.0	1.870	0.0	1.922	0.0	2.073	0.0	2.100	0.0	1.956	0.0	1.925
CL	0.013	*	0.018	*	0.021	*	0.013	*	0.010	*	0.010	*	0.021	*
F	0.765	*	0.827	*	1.097	*	0.811	*	0.677	*	0.812	*	0.772	*
H	3.229	4.007	3.115	3.960	2.904	4.022	3.210	4.035	3.320	4.007	3.191	4.013	3.218	4.011
O	24.000	*	24.000	*	24.000	*	24.000	*	24.000	*	24.000	*	24.000	*
F/M		1.204		1.156		0.919		1.131		1.079		1.058		1.099
F/FM		0.546		0.536		0.479		0.531		0.519		0.514		0.524

1 CP25A 1831, PORTSMOUTH, VA.
2 CP25A 1956, PORTSMOUTH, VA.
3 CP25A 1856.5, PORTSMOUTH, VA.
4 CP25A 1959.3, PORTSMOUTH, VA.
5 CP25A 1959.3, PORTSMOUTH, VA.
6 CP25A 1981, PORTSMOUTH, VA.
7 CP25A 1991, PORTSMOUTH, VA.

SUPER RECAL

TABLE 3. FELDSPAR ANALYSES

	1		2		3		4		5		6		7		8	
SIO2	67.87		65.34		64.73		64.62		67.01		65.79		66.89		66.31	
TIO2	0.05		0.0		0.04		0.01		0.06		0.0		0.01		0.0	
A2O3	20.83		18.54		21.22		21.22		22.09		18.39		19.05		18.75	
FEO	0.07		0.08		0.02		0.02		0.02		0.03		0.0		0.0	
MNO	0.0		0.04		0.0		0.0		0.0		0.01		0.04		0.0	
MGO	0.0		0.0		0.0		0.0		0.0		0.0		0.0		0.0	
CAO	1.46		0.01		2.63		2.63		2.59		0.0		0.0		0.20	
BAO	0.0		0.26		0.0		0.0		0.0		0.0		0.0		0.0	
NA2O	9.84		0.63		9.02		9.03		8.23		0.64		0.66		10.44	
K2O	0.04		15.81		0.15		0.12		0.15		16.77		16.31		0.07	
F	0.0		0.0		0.02		0.04		0.05		0.15		0.46		0.07	
CL	0.0		0.0		0.0		0.0		0.0		0.0		0.0		0.0	
SUM	100.16		100.71		97.81		97.65		100.15		101.63		102.96		95.77	
-O= F+CL	0.0		0.0		0.0		0.0		0.0		0.0		0.0		0.0	
SUM	100.16		100.71		97.81		97.65		100.15		101.63		102.96		95.77	
SI	2.954	*	2.998	*	2.897	*	2.896	*	2.912	*	2.996	*	2.985	*	3.010	*
AL	1.068	*	1.002	*	1.119	*	1.121	*	1.131	*	0.987	*	1.002	*	1.003	*
TI	0.002	4.023	0.0	4.000	0.001	4.018	0.000	4.017	0.002	4.045	0.0	3.982	0.000	3.987	0.0	4.012
FE	0.003	*	0.003	*	0.001	*	0.001	*	0.001	*	C.001	*	0.0	*	0.0	*
MN	0.0	*	0.002	*	0.0	*	0.0	*	0.0	*	0.000	*	0.002	*	0.0	*
MG	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*
CA	0.0	0.903	0.0	0.991	0.0	0.918	0.0	0.919	0.0	0.823	0.0	1.032	0.0	0.987	0.0	0.932
BA	0.0	*	0.005	*	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*
NA	0.830	*	0.056	*	0.783	*	0.785	*	0.693	*	0.057	*	0.057	*	0.919	*
K	0.002	0.903	0.925	0.991	0.009	0.918	0.007	0.919	0.008	0.823	0.974	1.032	0.928	0.987	0.004	0.932
O	8.000	*	8.000	*	8.000	*	8.000	*	8.000	*	8.000	*	8.000	*	8.000	*
AN		7.56		0.05		13.75		13.76		14.66		0.0		0.0		1.04
AB		92.19		5.68		85.32		85.49		84.32		5.48		5.79		98.52
OR		0.25		93.79		0.93		0.75		1.01		94.52		94.21		0.43
CN		0.0		0.47		0.0		0.0		0.0		0.0		0.0		0.0
F/M		0.0		0.0		0.0		0.0		0.0		0.0		0.0		0.0
F/FM		0.0		0.0		0.0		0.0		0.0		0.0		0.0		0.0

1 CP25A 1831, PORTSMOUTH, VA.
2 CP25A 1831, PORTSMOUTH, VA.
3 CP25A 1831.8, PORTSMOUTH, VA.
4 CP25A 1831.8, PORTSMOUTH, VA.
5 CP25A 1831.8, PORTSMOUTH, VA.
6 CP25A 1831.8, PORTSMOUTH, VA.
7 CP25A 1831.8, PORTSMOUTH, VA.
8 CP25A 1837.0, CARBONATE VEIN, PORTSMOUTH

SUPER RECAL

TABLE 3. FELDSPAR ANALYSES

	9	10	11	12	13	14	15	16
SIO2	67.66	69.06	64.07	67.99	65.73	63.93	62.55	59.76
TIO2	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
A2O3	19.16	19.42	18.56	20.48	18.59	22.43	20.18	27.90
FEO	0.0	0.0	0.0	0.0	0.05	0.08	0.07	0.03
MNO	0.0	0.01	0.0	0.0	0.0	0.0	0.01	0.0
MGO	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
CAO	0.11	0.09	0.0	0.71	0.01	4.13	0.04	1.91
BAO	0.0	0.0	0.0	0.0	0.08	0.0	0.57	0.08
NA2O	11.53	11.09	0.59	9.45	0.68	8.39	0.72	8.25
K2O	0.04	0.05	16.64	0.04	16.46	0.15	15.47	0.10
F	0.05	0.04	0.0	0.22	0.0	0.0	0.0	0.0
CL	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
SUM	98.50	99.72	99.86	98.67	101.60	99.11	99.61	98.03
-O= F+CL	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
SUM	98.50	99.72	99.86	98.67	101.60	99.11	99.61	98.03

	9		10		11		12		13		14		15		16	
SI	2.997	*	3.012	*	2.977	*	2.981	*	2.995	*	2.838	*	2.915	*	2.671	*
AL	1.000	*	0.998	*	1.016	*	1.058	*	0.998	*	1.173	*	1.108	*	1.469	*
TI	0.0	3.997	0.0	4.010	0.0	3.994	0.0	4.039	0.0	3.993	0.0	4.011	0.0	4.023	0.0	4.140
FE	0.0	*	0.0	*	0.0	*	0.0	*	0.002	*	0.003	*	0.003	*	0.001	*
MN	0.0	*	0.000	*	0.0	*	0.0	*	0.0	*	0.0	*	0.000	*	0.0	*
MG	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*
CA	0.0	0.998	0.0	0.945	0.0	1.039	0.0	0.839	0.0	1.020	0.0	0.930	0.0	1.000	0.0	0.814
BA	0.0	*	0.0	*	0.0	*	0.0	*	0.001	*	0.0	*	0.010	*	0.001	*
NA	0.990	*	0.938	*	0.053	*	0.803	*	0.060	*	0.722	*	0.065	*	0.715	*
K	0.002	0.998	0.003	0.945	0.986	1.039	0.002	0.839	0.957	1.020	0.008	0.930	0.919	1.000	0.006	0.814
O	8.000	*	8.000	*	8.000	*	8.000	*	8.000	*	8.000	*	8.000	*	8.000	*

	9	10	11	12	13	14	15	16
AN	0.52	0.45	0.0	3.98	0.05	21.19	0.20	11.24
AB	99.25	99.26	5.11	95.76	5.90	77.89	6.53	87.88
OR	0.23	0.29	94.89	0.27	93.91	0.92	92.23	0.70
CN	0.0	0.0	0.0	0.0	0.14	0.0	1.04	0.17
F/M	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
F/FM	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0

9 CP25A 1837.8, MUSCOVITE VEIN, PORTSMOUTH
10 CP25A 1837.8, MUSCOVITE VEIN, PORTSMOUTH
11 CP25A 1856.5, PORTSMOUTH, VA.
12 CP25A 1856.5, PORTSMOUTH, VA.
13 CP25A 1898, PORTSMOUTH, VA.
14 CP25A 1898, PORTSMOUTH, VA.
15 CP25A 1956, PORTSMOUTH, VA.
16 CP25A 1956, PORTSMOUTH, VA.

SUPER RECAL

TABLE 3. FELDSPAR ANALYSES

	17	18	19	20	21	22	23	24
SI02	68.06	69.65	63.62	64.14	65.59	62.69	63.96	63.29
TI02	0.0	0.02	0.0	0.0	0.04	0.05	0.04	0.0
A203	19.51	19.92	18.01	21.64	21.61	17.97	18.00	18.27
FEO	0.0	0.0	0.03	0.0	0.0	0.0	0.05	0.01
MNO	0.0	0.0	0.0	0.0	0.0	0.0	0.02	0.0
MGO	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
CAO	0.09	0.53	0.0	3.25	2.63	0.0	0.0	0.0
BAO	0.0	0.0	0.0	0.0	0.0	0.0	0.64	0.0
NA20	9.80	10.29	0.52	9.45	9.54	0.66	0.58	0.64
K20	0.04	0.03	17.28	0.22	0.08	16.45	15.58	16.54
F	0.07	0.10	0.04	0.01	0.14	0.18	0.0	0.02
CL	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
SUM	97.50	100.44	99.46	98.70	99.49	97.82	98.87	98.75
-0= F+CL	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
SUM	97.50	100.44	99.46	98.70	99.49	97.82	98.87	98.75

	17		18		19		20		21		22		23		24	
SI	3.019	*	3.007	*	2.982	*	2.861	*	2.886	*	2.974	*	2.999	*	2.977	*
AL	1.020	*	1.014	*	0.995	*	1.137	*	1.121	*	1.004	*	0.995	*	1.013	*
TI	0.0	4.039	0.001	4.022	0.0	3.976	0.0	3.999	0.001	4.008	0.002	3.980	0.001	3.995	0.0	3.989
FE	0.0	*	0.0	*	0.001	*	0.0	*	0.0	*	0.0	*	0.002	*	0.000	*
MN	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*	0.001	*	0.0	*
MG	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*
CA	0.0	0.849	0.0	0.888	0.0	1.081	0.0	0.985	0.0	0.942	0.0	1.056	0.0	0.999	0.0	1.051
BA	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*	0.012	*	0.0	*
NA	0.843	*	0.861	*	0.047	*	0.817	*	0.814	*	0.061	*	0.053	*	0.058	*
K	0.002	0.849	0.002	0.888	1.033	1.081	0.013	0.985	0.004	0.942	0.995	1.056	0.932	0.999	0.992	1.051
O	8.000	*	8.000	*	8.000	*	8.000	*	8.000	*	8.000	*	8.000	*	8.000	*

	17	18	19	20	21	22	23	24
AN	0.50	2.76	0.0	15.77	13.16	0.0	0.0	0.0
AB	99.23	97.05	4.37	82.96	86.37	5.75	5.29	5.56
OR	0.27	0.19	95.63	1.27	0.48	94.25	93.53	94.44
CN	0.0	0.0	0.0	0.0	0.0	0.0	1.18	0.0
F/M	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
F/FM	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0

17 CP25A 1901.5, MUSCOVITE VEINED, PORTSMOU
18 CP25A 1901.5, MUSCOVITE VEINED, PORTSMOU
19 CP25A 1920, PORTSMOUTH, VA.
20 CP25A 1920, PORTSMOUTH, VA.
21 CP25A 1920, PORTSMOUTH, VA.
22 CP25A 1959.3, PORTSMOUTH, VA.
23 CP25A 1959.3, PORTSMOUTH, VA.
24 CP25A 1959.3, PORTSMOUTH, VA.

SUPER RECAL

TABLE 3. FELDSPAR ANALYSES

	25	26	27	28
SIO2	65.02	65.30	65.89	64.24
TIO2	0.0	0.04	0.07	0.01
A2O3	18.89	22.56	19.12	22.23
FEO	0.05	0.06	0.05	0.04
MNO	0.0	0.01	0.02	0.0
MGO	0.0	0.0	0.0	0.0
CAO	0.02	3.25	0.02	3.20
BAO	0.21	0.02	0.14	0.11
NA2O	0.46	8.49	0.78	8.21
K2O	16.27	0.08	15.99	0.11
F	0.0	0.0	0.0	0.0
CL	0.0	0.0	0.0	0.0
SUM	100.92	99.81	102.08	98.15
-O= F+CL	0.0	0.0	0.0	0.0
SUM	100.92	99.81	102.08	98.15

	25		26		27		28	
SI	2.982	*	2.864	*	2.982	*	2.866	*
AL	1.021	*	1.166	*	1.020	*	1.169	*
TI	0.0	4.003	0.001	4.032	0.002	4.004	0.000	4.035
FE	0.002	*	0.002	*	0.002	*	0.001	*
MN	0.0	*	0.000	*	0.001	*	0.0	*
MG	0.0	*	0.0	*	0.0	*	0.0	*
CA	0.0	0.999	0.0	0.882	0.0	0.997	0.0	0.873
BA	0.004	*	0.000	*	0.002	*	0.002	*
NA	0.041	*	0.722	*	0.068	*	0.710	*
K	0.952	0.999	0.004	0.882	0.923	0.997	0.006	0.873
O	8.000	*	8.000	*	8.000	*	8.000	*
AN		0.10		17.36		0.10		17.56
AB		4.10		82.09		6.88		81.51
OR		95.42		0.51		92.77		0.72
CN		0.38		0.04		0.25		0.22
F/M	0.0		0.0		0.0		0.0	
F/FM	0.0		0.0		0.0		0.0	

25 CP25A 1981, PORTSMOUTH, VA.
26 CP25A 1981, PORTSMOUTH, VA.
27 CP25A 1991, PORTSMOUTH, VA.
28 CP25A 1991, PORTSMOUTH, VA.

SUPER RECAL

TABLE 4. MUSCOVITE ANALYSES

	1	2	3	4	5	6	7	8
SIO2	48.70	48.19	46.80	47.08	46.78	47.05	47.59	47.91
TIO2	0.60	1.35	0.77	0.63	0.68	0.65	0.81	0.78
A2O3	32.00	28.68	28.04	27.56	29.60	28.26	28.61	28.23
FEO	4.51	5.32	5.57	5.49	5.52	5.70	5.83	5.78
MNO	0.02	0.05	0.15	0.08	0.21	0.22	0.15	0.05
MGO	1.45	2.14	1.73	1.85	1.91	1.80	1.86	1.94
CAO	0.0	0.0	0.0	0.0	0.0	0.0	0.04	0.0
BAO	0.0	0.0	0.0	0.0	0.01	0.0	0.0	0.0
NA2O	0.10	0.15	0.17	0.21	0.27	0.29	0.25	0.13
K2O	7.57	7.98	8.95	9.31	8.38	11.19	9.69	9.10
F	0.50	0.61	0.81	0.69	0.99	1.06	0.87	0.65
CL	0.0	0.0	0.0	0.0	0.01	0.0	0.0	0.0
H2O	4.28	4.13	3.91	3.96	3.90	3.86	3.98	4.07
SUM	99.73	98.60	96.90	96.86	98.26	100.08	99.68	98.64
-O= F+CL	0.21	0.26	0.34	0.29	0.42	0.45	0.37	0.27
SUM	99.52	98.34	96.56	96.57	97.84	99.64	99.32	98.36

	1		2		3		4		5		6		7		8	
SI	6.433	*	6.504	*	6.481	*	6.535	*	6.363	*	6.398	*	6.439	*	6.521	*
AL	1.567	8.000	1.496	8.000	1.519	8.000	1.465	8.000	1.637	8.000	1.602	8.000	1.561	8.000	1.479	8.000
AL	3.414	*	3.065	*	3.057	*	3.043	*	3.107	*	2.927	*	3.000	*	3.049	*
TI	0.060	*	0.137	*	0.080	*	0.066	*	0.070	*	0.066	*	0.082	*	0.080	*
FE	0.498	*	0.600	*	0.645	*	0.637	*	0.628	*	0.648	*	0.660	*	0.658	*
MN	0.002	*	0.006	*	0.018	*	0.009	*	0.024	*	0.025	*	0.017	*	0.006	*
MG	0.285	4.259	0.430	4.239	0.357	4.157	0.383	4.138	0.387	4.216	0.365	4.031	0.375	4.134	0.394	4.186
CA	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*	0.006	*	0.0	*
NA	0.026	*	0.039	*	0.046	*	0.057	*	0.071	*	0.076	*	0.066	*	0.034	*
K	1.275	*	1.374	*	1.581	*	1.648	*	1.454	*	1.941	*	1.672	*	1.580	*
BA	0.0	1.301	0.0	1.413	0.0	1.627	0.0	1.705	0.001	1.526	0.0	2.017	0.0	1.744	0.0	1.614
CL	0.0	*	0.0	*	0.0	*	0.0	*	0.002	*	0.0	*	0.0	*	0.0	*
F	0.209	*	0.260	*	0.355	*	0.303	*	0.426	*	0.456	*	0.372	*	0.280	*
H	3.791	4.000	3.740	4.000	3.645	4.000	3.697	4.000	3.572	4.000	3.544	4.000	3.628	4.000	3.720	4.000
O	24.000	*	24.000	*	24.000	*	24.000	*	24.000	*	24.000	*	24.000	*	24.000	*
F/M		1.753		1.408		1.856		1.690		1.684		1.846		1.804		1.686
F/FM		0.637		0.585		0.650		0.628		0.627		0.649		0.643		0.628

1 CP25A 1831.8, PORTSMOUTH, VA.
2 CP25A 1831.8, PORTSMOUTH, VA.
3 CP25A 1837.0, PORTSMOUTH, VA.
4 CP25A 1837.0, PORTSMOUTH, VA.
5 CP25A 1837.0, PORTSMOUTH, VA.
6 CP25A 1837.8, PORTSMOUTH, VA.
7 CP25A 1837.8, PORTSMOUTH, VA.
8 CP25A 1856.5, PORTSMOUTH, VA.

SUPER RECAL

TABLE 4. MUSCOVITE ANALYSES

	9	10	11	12	13	14	15	16
SIO2	49.93	47.32	49.27	46.15	46.45	48.25	48.50	45.55
TIO2	0.60	0.50	0.73	0.75	0.47	0.88	0.90	1.47
A2O3	28.23	26.98	28.91	27.65	27.92	28.87	28.80	27.04
FEO	5.77	4.92	5.70	5.49	4.64	6.27	5.62	5.34
MNO	0.15	0.04	0.15	0.18	0.14	0.22	0.10	0.09
MGO	2.24	1.69	2.82	1.78	2.33	2.43	2.22	1.82
CAO	0.0	0.04	0.0	0.0	0.0	0.0	0.0	0.0
BAO	0.0	0.21	0.0	0.0	0.0	0.0	0.0	0.0
NA2O	0.21	0.16	0.20	0.25	0.34	0.25	0.21	0.10
K2O	8.38	10.43	6.70	11.50	11.34	8.40	7.42	8.24
F	0.86	0.40	1.32	0.65	0.92	1.48	1.00	0.41
CL	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.25
H2O	4.08	4.08	3.86	3.98	3.87	3.76	3.95	3.93
SUM	100.45	96.77	99.66	98.38	98.42	100.81	98.72	94.24
-O= F+CL	0.36	0.17	0.56	0.27	0.39	0.62	0.42	0.23
SUM	100.08	96.60	99.10	98.10	98.03	100.19	98.30	94.01

SI	6.622	*	6.617	*	6.514	*	6.413	*	6.408	*	6.400	*	6.510	*	6.482	*
AL	1.378	8.000	1.383	8.000	1.486	8.000	1.587	8.000	1.592	8.000	1.600	8.000	1.490	8.000	1.518	8.000
AL	3.034	*	3.062	*	3.017	*	2.940	*	2.947	*	2.912	*	3.066	*	3.017	*
TI	0.060	*	0.053	*	0.073	*	0.078	*	0.049	*	0.088	*	0.091	*	0.157	*
FE	0.640	*	0.575	*	0.630	*	0.638	*	0.535	*	0.695	*	0.631	*	0.636	*
MN	0.017	*	0.005	*	0.017	*	0.021	*	0.016	*	0.025	*	0.011	*	0.011	*
MG	0.443	4.193	0.352	4.047	0.556	4.293	0.369	4.046	0.479	4.027	0.480	4.200	0.444	4.243	0.386	4.206
CA	0.0	*	0.006	*	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*
NA	0.054	*	0.043	*	0.051	*	0.067	*	0.091	*	0.064	*	0.055	*	0.028	*
K	1.418	*	1.860	*	1.130	*	2.038	*	1.995	*	1.421	*	1.270	*	1.496	*
BA	0.0	1.472	0.012	1.921	0.0	1.181	0.0	2.106	0.0	2.086	0.0	1.485	0.0	1.325	0.0	1.523
CL	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*	0.060	*
F	0.361	*	0.177	*	0.552	*	0.286	*	0.401	*	0.621	*	0.425	*	0.185	*
H	3.639	4.000	3.823	4.000	3.448	4.000	3.714	4.000	3.599	4.000	3.379	4.000	3.575	4.000	3.755	4.000
O	24.000	*	24.000	*	24.000	*	24.000	*	24.000	*	24.000	*	24.000	*	24.000	*
F/M		1.483		1.647		1.164		1.788		1.151		1.499		1.446		1.674
F/FM		0.597		0.622		0.538		0.641		0.535		0.600		0.591		0.626

9 CP25A 1856.5, PORTSMOUTH, VA.
10 CP25A 1898, PORTSMOUTH, VA.
11 CP25A 1901.5, PORTSMOUTH, VA.
12 CP25A 1901.5, PORTSMOUTH, VA.
13 CP25A 1901.5, PORTSMOUTH, VA.
14 CP25A 1901.5, PORTSMOUTH, VA.
15 CP25A 1901.5, PORTSMOUTH, VA.
16 CP25A 1920, PORTSMOUTH, VA.

SUPER RECAL

TABLE 4. MUSCOVITE ANALYSES

	17	18	19	20
SIO2	47.94	42.31	48.05	47.42
TIO2	1.14	0.12	0.16	0.46
A2O3	28.92	33.36	27.97	28.65
FEO	5.25	3.90	5.70	6.04
MNO	0.04	0.07	0.13	0.06
MGO	1.86	1.47	2.06	1.95
CAO	0.0	0.13	0.0	0.0
BAO	0.0	0.39	0.0	0.0
NA2O	0.11	0.18	0.11	0.15
K2O	8.24	9.96	8.95	8.37
F	0.48	0.26	0.82	0.48
CL	0.0	0.02	0.01	0.0
H2O	4.17	4.12	3.96	4.12
SUM	98.15	96.29	97.92	97.70
-O= F+CL	0.20	0.11	0.35	0.20
SUM	97.94	96.18	97.57	97.50

SI	6.508	*	5.946	*	6.571	*	6.502	*
AL	1.492	8.000	2.054	8.000	1.429	8.000	1.498	8.000
AL	3.135	*	3.470	*	3.079	*	3.131	*
TI	0.116	*	0.013	*	0.016	*	0.047	*
FE	0.596	*	0.458	*	0.652	*	0.693	*
MN	0.005	*	0.008	*	0.015	*	0.007	*
MG	0.376	4.228	0.308	4.257	0.420	4.182	0.399	4.277
CA	0.0	*	0.020	*	0.0	*	0.0	*
NA	0.029	*	0.049	*	0.029	*	0.040	*
K	1.427	*	1.785	*	1.561	*	1.464	*
BA	0.0	1.456	0.021	1.875	0.0	1.590	0.0	1.504
CL	0.0	*	0.005	*	0.002	*	0.0	*
F	0.206	*	0.116	*	0.355	*	0.208	*
H	3.794	4.000	3.880	4.000	3.643	4.000	3.792	4.000
O	24.000	*	24.000	*	24.000	*	24.000	*
F/M		1.596		1.516		1.588		1.755
F/FM		0.615		0.602		0.614		0.637

17 CP25A 1920, PORTSMOUTH, VA.
18 CP25A 1956, PORTSMOUTH, VA.
19 CP25A 1959.3, PORTSMOUTH, VA.
20 CP25A 1959.3, PORTSMOUTH, VA.

SUPER RECAL

TABLE 5. CHLORITE ANALYSES

	1	2	3	4	5	6
SIO2	25.75	27.05	27.48	26.76	27.50	26.02
TIO2	0.10	0.25	0.14	0.16	0.23	0.10
A2O3	19.60	18.47	18.86	18.98	19.45	18.51
FEO	25.42	25.53	25.03	25.33	25.54	25.11
MNO	0.63	0.62	0.71	0.65	0.69	0.61
MGO	15.29	14.44	14.41	14.71	15.12	14.31
CAO	0.0	0.03	0.0	0.01	0.02	-0.00
BAO	0.0	0.0	0.0	0.0	0.0	0.0
NA2O	0.02	0.02	0.02	0.02	0.02	0.02
K2O	0.01	0.13	0.09	0.08	0.13	0.03
F	0.17	0.34	0.14	0.22	0.30	0.13
CL	0.20	0.0	0.0	0.07	0.16	-0.03
H2O	11.07	11.02	11.20	11.10	11.31	10.88
SUM	98.26	97.90	98.08	98.08	100.47	95.69
-O= F+CL	0.12	0.14	0.06	0.11	0.16	0.05
SUM	98.15	97.76	98.02	97.98	100.31	95.64

	1		2		3		4		5		6	
SI	2.745	*	2.886	*	2.918	*	2.850	*	2.853	*	2.847	*
AL	1.255	4.000	1.114	4.000	1.082	4.000	1.150	4.000	1.147	4.000	1.153	4.000
AL	1.206	*	1.209	*	1.278	*	1.231	*	1.230	*	1.233	*
TI	0.008	*	0.020	*	0.011	*	0.013	*	0.018	*	0.008	*
FE	2.266	*	2.278	*	2.223	*	2.256	*	2.216	*	2.297	*
MN	0.057	*	0.056	*	0.064	*	0.059	*	0.061	*	0.057	*
MG	2.429	*	2.297	*	2.281	*	2.335	*	2.338	*	2.333	*
CA	0.0	5.971	0.0	5.885	0.0	5.874	0.0	5.910	0.0	5.886	0.0	5.935
NA	0.004	*	0.004	*	0.004	*	0.004	*	0.004	*	0.004	*
K	0.001	*	0.018	*	0.012	*	0.010	*	0.017	*	0.004	*
BA	0.0	5.971	0.0	5.885	0.0	5.874	0.0	5.910	0.0	5.886	0.0	5.935
CL	0.036	*	0.0	*	0.0	*	0.012	*	0.028	*	-0.005	*
F	0.057	*	0.115	*	0.047	*	0.073	*	0.100	*	0.044	*
H	7.907	8.000	7.885	8.000	7.953	8.000	7.915	8.000	7.872	8.000	7.961	8.000
O	18.000	*	18.000	*	18.000	*	18.000	*	18.000	*	18.000	*
F/M		0.956		1.016		1.003		0.991		0.974		1.009
F/FM		0.489		0.504		0.501		0.498		0.493		0.502

1 CP25A 1920, PORTSMOUTH, VA.
2 CP25A 1920, PORTSMOUTH, VA.
3 CP25A 1920, PORTSMOUTH, VA.

4 AVERAGE
5 AVERAGE PLUS SIGMA
6 AVERAGE MINUS SIGMA

SUPER RECAL

TABLE 5. CHLORITE ANALYSES

	1	2	3	4	5	6
SI02	27.52	27.21	26.42	27.05	27.51	26.59
TI02	0.12	0.10	0.09	0.10	0.12	0.09
A203	19.25	18.60	18.73	18.86	19.14	18.58
FEO	24.59	24.17	24.60	24.45	24.65	24.25
MNO	0.68	0.65	0.66	0.66	0.68	0.65
MGO	16.03	16.31	15.46	15.93	16.29	15.58
CAO	0.0	0.0	0.0	0.0	0.0	0.0
BAO	0.0	0.0	0.0	0.0	0.0	0.0
NA20	0.01	0.0	0.02	0.01	0.02	0.00
K20	0.01	0.0	0.0	0.00	0.01	-0.00
F	0.20	0.16	0.09	0.15	0.20	0.10
CL	0.0	0.0	0.0	0.0	0.0	0.0
H20	11.40	11.27	11.11	11.26	11.42	11.10
SUM	99.81	98.47	97.18	98.49	100.03	96.94
-O= F+CL	0.08	0.07	0.04	0.06	0.08	0.04
SUM	99.73	98.40	97.14	98.42	99.95	96.90

	1		2		3		4		5		6	
SI	2.862	*	2.869	*	2.836	*	2.856	*	2.857	*	2.855	*
AL	1.138	4.000	1.131	4.000	1.164	4.000	1.144	4.000	1.143	4.000	1.145	4.000
AL	1.221	*	1.180	*	1.205	*	1.202	*	1.198	*	1.206	*
TI	0.009	*	0.008	*	0.007	*	0.008	*	0.009	*	0.007	*
FE	2.139	*	2.131	*	2.208	*	2.159	*	2.141	*	2.178	*
MN	0.060	*	0.058	*	0.060	*	0.059	*	0.059	*	0.059	*
MG	2.485	*	2.563	*	2.473	*	2.507	*	2.520	*	2.493	*
CA	0.0	5.917	0.0	5.940	0.0	5.958	0.0	5.938	0.0	5.933	0.0	5.944
NA	0.002	*	0.0	*	0.004	*	0.002	*	0.004	*	0.000	*
K	0.001	*	0.0	*	0.0	*	0.000	*	0.001	*	-0.000	*
BA	0.0	5.917	0.0	5.940	0.0	5.958	0.0	5.938	0.0	5.933	0.0	5.944
CL	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*
F	0.066	*	0.053	*	0.031	*	0.050	*	0.064	*	0.036	*
H	7.934	8.000	7.947	8.000	7.969	8.000	7.950	8.000	7.936	8.000	7.964	8.000
O	18.000	*	18.000	*	18.000	*	18.000	*	18.000	*	18.000	*
F/M		0.885		0.854		0.917		0.885		0.873		0.897
F/FM		0.469		0.461		0.478		0.469		0.466		0.473

1 CP25A 1831.8, PORTSMOUTH, VA.
2 CP25A 1831.8, PORTSMOUTH, VA.
3 CP25A 1831.8, PORTSMOUTH, VA.
4 AVERAGE
5 AVERAGE PLUS SIGMA
6 AVERAGE MINUS SIGMA

SUPER RECAL

TABLE 5. CHLORITE ANALYSES

	1	2	3	4
SIO2	28.81	28.81	28.81	28.81
TIO2	0.14	0.14	0.14	0.14
A2O3	17.32	17.32	17.32	17.32
FEO	24.98	24.98	24.98	24.98
MNO	0.53	0.53	0.53	0.53
MGO	14.99	14.99	14.99	14.99
CAO	0.08	0.08	0.08	0.08
BAO	0.06	0.06	0.06	0.06
NA2O	0.05	0.05	0.05	0.05
K2O	0.06	0.06	0.06	0.06
F	0.09	0.09	0.09	0.09
CL	0.02	0.02	0.02	0.02
H2O	11.28	11.28	11.28	11.28
SUM	98.41	98.41	98.41	98.41
-O= F+CL	0.04	0.04	0.04	0.04
SUM	98.37	98.37	98.37	98.37

	1		2		3		4	
SI	3.044	*	3.044	*	3.044	*	3.044	*
AL	0.956	4.000	0.956	4.000	0.956	4.000	0.956	4.000
AL	1.201	*	1.201	*	1.201	*	1.201	*
TI	0.011	*	0.011	*	0.011	*	0.011	*
FE	2.207	*	2.207	*	2.207	*	2.207	*
MN	0.047	*	0.047	*	0.047	*	0.047	*
MG	2.361	*	2.361	*	2.361	*	2.361	*
CA	0.0	5.858	0.0	5.858	0.0	5.858	0.0	5.858
NA	0.010	*	0.010	*	0.010	*	0.010	*
K	0.008	*	0.008	*	0.008	*	0.008	*
BA	0.002	5.858	0.002	5.858	0.002	5.858	0.002	5.858
CL	0.004	*	0.004	*	0.004	*	0.004	*
F	0.030	*	0.030	*	0.030	*	0.030	*
H	7.966	8.000	7.966	8.000	7.966	8.000	7.966	8.000
O	18.000	*	18.000	*	18.000	*	18.000	*
F/M		0.955		0.955		0.955		0.955
F/FM		0.489		0.489		0.489		0.489

1 CP25A 1898, PORTSMOUTH, VA.
2 AVERAGE
3 AVERAGE PLUS SIGMA
4 AVERAGE MINUS SIGMA

TABLE 5. CHLORITE ANALYSES

SUPER RECAL

	1
SIO2	26.13
TIO2	0.03
A2O3	20.64
FEO	22.49
MNO	0.87
MGO	16.88
CAO	0.0
BAO	0.0
NA2O	0.01
K2O	0.0
F	0.34
CL	0.01
H2O	11.26
SUM	98.66
-O= F+CL	0.15
SUM	98.52

SI	2.729	*
AL	1.271	4.000
AL	1.269	*
TI	0.002	*
FE	1.964	*
MN	0.077	*
MG	2.627	*
CA	0.0	5.942
NA	0.002	*
K	0.0	*
BA	0.0	5.942
CL	0.002	*
F	0.112	*
H	7.886	8.000
O	18.000	*
F/M		0.777
F/FM		0.437

1 CP25A 1959.3, PORTSMOUTH, VA.

SUPER RECAL

TABLE 6. EPIDOTE ANALYSES

	1	2	3	4	5
SIO2	37.04	36.86	36.95	37.04	36.86
TIO2	0.16	0.02	0.09	0.16	0.02
A2O3	22.07	21.88	21.98	22.07	21.88
F2O3	15.77	16.30	16.03	16.30	15.77
MNO	0.31	0.24	0.27	0.31	0.24
MGO	0.0	0.0	0.0	0.0	0.0
CAO	22.92	22.75	22.83	22.92	22.75
BAO	0.0	0.0	0.0	0.0	0.0
NA2O	0.02	0.01	0.02	0.02	0.01
K2O	0.0	0.0	0.0	0.0	0.0
F	0.17	0.12	0.14	0.17	0.12
CL	0.0	0.0	0.0	0.0	0.0
H2O	1.79	1.81	1.80	1.80	1.80
SUM	100.25	99.99	100.12	100.79	99.45
-O= F+CL	0.07	0.05	0.06	0.07	0.05
SUM	100.18	99.94	100.06	100.71	99.40

SI	2.963	*	2.961	*	2.962	*	2.952	*	2.972	*
AL	0.037	3.000	0.039	3.000	0.038	3.000	0.048	3.000	0.028	3.000
AL	2.044	*	2.032	*	2.038	*	2.024	*	2.051	*
TI	0.010	*	0.001	*	0.005	*	0.010	*	0.001	*
MG	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*
FE3+	0.949	*	0.985	*	0.967	*	0.977	*	0.957	*
MN	0.021	3.024	0.016	3.035	0.019	3.029	0.021	3.032	0.016	3.026
CA	1.965	*	1.958	*	1.961	*	1.957	*	1.966	*
K	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*
NA	0.003	*	0.002	*	0.002	*	0.003	*	0.002	*
BA	0.0	1.968	0.0	1.960	0.0	1.964	0.0	1.960	0.0	1.967
CL	0.0	*	0.0	*	0.0	*	0.0	*	0.0	*
F	0.043	0.0	0.030	0.0	0.037	0.0	0.043	0.0	0.031	0.0
H	0.957	1.000	0.970	1.000	0.963	1.000	0.957	1.000	0.969	1.000
O	13.000	*	13.000	*	13.000	*	13.000	*	13.000	*
PS		31.12		32.07		31.59		31.83		31.35
CZ		68.19		67.40		67.80		67.49		68.11
PD		0.69		0.53		0.61		0.68		0.54

1 CP25A 1959.3, PORTSMOUTH, VA
2 CP25A 1959.3, PORTSMOUTH, VA.
3 AVERAGE
4 AVERAGE PLUS SIGMA
5 AVERAGE MINUS SIGMA

LILESVILLE CONTACT AUREOLE, NORTH CAROLINA

Nicholas H. Evans

Contact metamorphism associated with the Lilesville granite near Rockingham, N.C. is under investigation. Reconnaissance mapping has delineated an aureole of metamorphic rocks surrounding the pluton which range from greenschist regional grade in the Carolina slate belt country rocks to (andalusite + K feldspar)-bearing hornfels adjacent to the granite contact. In addition, mapping has revealed a region of high grade pelitic gneisses within the pluton. These interior gneissic rocks bear granulite facies mineral assemblages and in some places show extensive partial melting. Mineral assemblages and chemistry are being studied in the aureole rocks and in the interior gneiss in order to:

1) evaluate intensive parameters P, T, and fH_2O for emplacement of the Lilesville pluton, and

2) evaluate the degree and nature of interaction of the granite with the country rocks during emplacement, a process which often increases heat production of granites.

The generation of the partial melt phase as a product of continuous reactions among solid phases may better approximate the mechanism of granite formation than does the generation of an eutectic or cotectic minimum melt as modeled by experimental studies. The reactions and mechanisms involved in generating a partial melt have important implications for the partitioning of radionuclides between the melt and solid phase and thus for the concentration of heat producing elements in granitic bodies. The migmatitic interior gneiss affords a good opportunty to evaluate the role of partial melting in granite production.

Systematic sampling of the aureole and interior gneiss was carried out in conjunction with reconnaissance mapping. Thin sections made from the sample set have been studied petrographically in order to characterize the sequence of prograde metamorphic reactions. The petrology of the aureole and interior gneiss has been further investigated by electron microprobe analysis of selected mineral phases. Whole rock chemical analyses are being carried out on melt and restite Carolina slate belt in order to study partitioning of major elements during partial melting.

REGIONAL GEOPHYSICS LABORATORY

HEAT FLOW IN THE SOUTHEASTERN UNITED STATES

J. K. Costain, J. A. Speer, L. Glover, III, L. Perry,
S. Dashevsky, and M. McKinney

Introduction

Heat flow determinations reported here are primarily the result of a geothermal program at VPI&SU supported during 1976-82 by the Department of Energy to evaluate the low-temperature geothermal potential of the eastern U.S. from New Jersey to Georgia.

Previous Work

Diment *et al*. (1972) summarized published heat flow values for the eastern U.S. At that time, only eight heat flow values were available in Virginia, South Carolina, and Georgia. Sass *et al*. (1976) compiled a revised heat flow map which included new values in Virginia and one in South Carolina. No determinations had yet been made in North Carolina. Sass *et al*. (1981) summarized crustal heat flow in the United States. Smith *et al*. (1981) reported values in North Carolina, South Carolina, and Georgia.

Birch *et al*. (1968) showed that variations in heat flow in New England and New York are caused primarily by differences in crustal concentrations of the radioactive, heat- producing isotopes of uranium, thorium, and potassium. As shown here, a similar conclusion holds for the southeastern U. S. Many of the heat flow values reported here were determined in granites. Initial differences in concentrations of U and Th at the time of granite emplacement depend upon the composition of the granite source rock, but heat flow in the eastern U.S. and elsewhere is also affected by the metamorphic redistribution of U and Th. In both the northeastern and southeastern U.S., the highest heat flow values are associated with unmetamorphosed granites. In the northeastern U.S., the genesis, age, and mode of emplacement of many of the granites have been studied in detail. In the southeast, Speer *et al*. (1980) discussed the field relations, petrology, and chemistry of the post- and late synmetamorphic 254-330 Ma granite plutons exposed in the Piedmont of the southern Appalachians. These are the youngest exposed granitic plutons known in the southeastern U.S. They are also sites of high heat flow, and are therefore important for an evaluation of regional heat flow.

A linear relationship, $q = q^* + DA$, between surface heat flow, q, and surface heat generation, A, was discovered by Birch et al. (1968) from heat flow determinations in New England and the Adirondacks. This discovery led to the definition of continental heat flow provinces, characterized by q^* (Roy et al., 1968). The quantity q^*, defined by Roy et al. (1968) as the "reduced heat flow", is assumed to be the flux from the lower crust and upper mantle. Lachenbruch (1968) simultaneously described a similar relation in the Sierra Nevada. The empirical relationship has been observed on several continents (Sass et al., 1981). Determinations of heat flow and heat generation in the southeastern U.S. reported by Smith et al. (1981) do not define a linear relation; however, core samples were not always available for the determination of representative values of heat generation, and the drill holes used may not have been deep enough so that geothermal gradients were undisturbed by groundwater flow. These are often difficult conditions to satisfy.

The simplest models that result in a linear relation between q and A are:

I) The distribution of heat-producing elements is constant from the surface to a depth D; lateral variations in A will not significantly affect linearity. The numerical value of D is equal to the thickness of crust containing the heat-producing elements.

II) The distribution of heat-producing elements decreases exponentially downward from the surface. D is not a thickness parameter. Lateral variations in A will not significantly affect linearity.

A distinction made by Lachenbruch (1968) between the two models is that Model II will survive differential erosion between heat flow sites, but Model I will not.

Model I implies an uninterrupted, uniform distribution of U and Th down to a depth D. Model II would have to imply an uninterrupted, exponentially decreasing distribution of U and Th to a depth of approximately 3D. These models place severe constraints on crustal structure wherever a linear relationship is observed (Costain and Glover, 1980a,b).

Geothermal Gradients and Thermal Conductivity

All geothermal gradients reported here are based on temperature measurements made at discrete depths in boreholes while the sensor was moving down the hole. Temperatures in holes in the Piedmont of Georgia and beneath the Atlantic Coastal Plain of New Jersey, Delaware, Maryland, Virginia, North Carolina, South Carolina, and Georgia were measured by a thermistor assembly moving at a speed of 30 m/min. Temperatures were sampled at intervals of 0.5 m by microcomputer and digitally recorded on magnetic tape. All other temperatures measured in the Piedmont, Blue Ridge, and Valley and Ridge Provinces were determined manually at intervals of 2.5 m or greater using a Mueller resistance bridge with the thermistor assembly stationary in the borehole. In both methods a rapid-response thermistor with a time constant of approximately 1 sec was used. The thermistors are precise to better than 0.01°C, and are believed to be accurate to aproximately 0.1°C. Thermal gradients determined by each system in a 300-m hole at thermal equilibrium agree to within 0.33°C/km over the entire hole and typically to within 0.87°C/km over a 25 meter interval.

Geothermal gradients were computed from a least-squares fit of a straight line to linear intervals of the temperature versus depth curves. No corrections to the geothermal gradient for the effects of topography, topographic or geologic evolution, refraction of isotherms, or climatic change have been applied. Except for one site (SP3 at Spruce Pine, N.C.) in the Blue Ridge, topographic relief did not exceed 100 m. The correction is not important for the region of study.

Heat flow determinations reported here are based on equilibrium gradients measured in undisturbed fluid-filled drill holes. Due to the confined and semi-confined nature of aquifers beneath the Atlantic Coastal Plain, holes drilled for this program in Coastal Plain sediments were cased, and the annulus around the casing cemented to prevent vertical circulation between aquifers via the bore hole. In spite of this, deep circulation of water between fractures or aquifers was detected at sites C15 and C11.

In every case, care was taken to determine heat flow only for intervals over which the gradient was linear and apparently unaffected by groundwater flow, and which were adequately sampled for thermal conductivity. The heat flow values are, except where noted, believed to belong to Category I as defined by Sass *et al*. (1971). Category I values are accurate to within 5%.

Thermal Conductivity

Samples for determinations of thermal conductivity were selected from core from the holes. Core diameters ranged from AQ (2.7 cm) to NQ (4.76 cm). Each sample was machined to a right circular cylinder of thickness 0.197 cm. All samples were saturated with water while in an evacuated container.

Measurements were made in a divided bar apparatus using fused quartz as a reference standard. The entire fused quartz-sample stack was insulated with styrofoam. Contact resistance at the quartz-sample interface was minimized by using Vaseline petroleum jelly on the interface and axial pressures of about 1500 psi. A stack correction of 1-2% was determined by calibrating with fused quartz.

A 10°C temperature gradient was imposed across the quartz-sample stack by two constant-temperature baths. The temperature of the sample was kept within 1 C° of its *in situ* temperature. Temperatures within the stack were measured with copper-constantan thermocouples and a Leeds and Northrup Type K-3 potentiometer.

Heat Generation

Samples of core over selected depth intervals were analyzed for U, Th, and K content by comparing in a least squares sense the gamma ray spectrum of the sample to the spectra of standards with known concentrations.

Core samples were crushed and the material placed in a 15 cm diameter by 2.5 cm high plastic container, which was sealed and set aside for at least 30 days to allow radon gas (Rn^{222}) to accumulate. In general, the container was completely filled with the crushed material (approximately 700 gm) to insure that the counting geometry of the sample duplicated that of the reference standards.

Gamma-ray spectra over the energy band 100 Kev to 2.8 Mev were acquired with a 10 x 12.7 cm NaI (Th) crystal and a Nuclear Data 2200 system. The computer program ALPHA M, (Schonfield, 1966) was used to resolve the spectra. Concentrations of U, Th, and K determined assume equilibrium in the decay schemes of both the standards and the samples.

Geologic Framework

The divisions and belts of the Appalachian Mountain System have been summarized elsewhere (King, 1955; Fullagar, 1971; Hatcher, 1972; Rankin et al., 1973; Glover and Sinha, 1973; Glover, 1979). Aspects of the geologic framework important for the interpretation of heat flow have been discussed by Glover (1979), Speer et al. (1980), Glover et al. (in press), and elsewhere in this report. New heat flow values are reported here for all of the divisions of the Appalachian orogen.

An important focus of the VPI&SU geothermal program was the distribution and petrography of the post- and late synmetamorphic (254-330 Ma) granite plutons exposed in the Piedmont. Variations in heat generation and heat flow can be correlated with mineralogy and chemistry of the Piedmont granites, features which reflect the source of the magma and the physical conditions during and subsequent to crystallization. The several petrographic types of granites from which new heat flow values were obtained in the Piedmont and basement beneath the Atlantic Coastal Plain include:

1) coarse-grained, amphibole ± pyroxene + biotite granites with allanite, comprising one facies of composite plutons.

Hole	Location
PT-1,2,3	Petersburg, Va.
CS1	Castalia, N.C.
RM1	Rocky Mount, N.C.
DO1	Dort, N.C.
KR3	Liberty Hill, S.C.
PG1	Pageland, S.C.

2) coarse-grained, biotite granite with allanite, titanite,

PM1	Palmetto, Ga.
SM1	Siloam, Ga.
SL1	Springfield, S.C.

3) fine- to medium-grained, biotite granite comprising one facies of a composite pluton. Muscovite is possibly primary igneous.

SM1	Siloam, Ga.
WN1	Winnsboro, S.C.

4) Syndeformational plutons containing granites comparable to facies 1-3 above.

RL2,4,5	Rolesville, N.C.

The above four types of granites are comparable to the I-type granites of Chappell and White (1974). Their emplacement occurred largely during and after the Alleghanian orogeny.

5) coarse-grained, cordierite + biotite granites and granodiorites with ilmenite and monazite:

SU	Stumpy Point, N.C.

6) medium- to coarse-grained, (primary?) muscovite + biotite granitoid rocks that are evidently the highly-evolved, fluid-rich roofs of large plutons of type 1 or type 2 granites:

ED1	Cuffytown Creek, S.C.
C25A	Portsmouth, Va.

The cordierite-bearing granite SU is comparable to the S-granites of Chappell and White (1974). ED1 and C25A are petrologic variants of I-granites. Heat flow determinations were obtained from holes drilled into granites at the above sites except at Stumpy Point, N.C. (SU). The Stumpy Point core was recovered from basement by others a few tens of kilometers from our heat flow site C19 at Stumpy Point.

Results

Valley and Ridge and Blue Ridge Provinces

Only one new value is reported from the Valley and Ridge province, at Pearisburg, Va., in the folded Appalachians (Table 1). The heat flow is approximately 123 mW/m^2. Although no detailed structural mapping was done in the area, the high heat flow is probably the result of shallow circulation of warm water in a structural setting similar to that described by Perry *et al*. (1979). The value is not regionally representative, but does appear to support their model. The single new heat flow value in the Blue Ridge (SP3 at Spruce Pine, N.C.) was from a hole approximately 1300 m in depth and is believed to be regionally representative of the background heat flow.

Piedmont

Our new data in the Piedmont support the assumption of a single heat flow province. Heat flow and heat generation values are summarized in Table 1. Although our heat flow sites (Figs. 1 and 2) are biased toward the younger synmetamorphic and unmetamorphosed granites, enough data are available to identify a background heat flow of approximately 40 mW/m^2 in the Piedmont. The Piedmont heat flow and heat generation values, determined in granites (ED1, KR3, PT1, PT2, PT3, RL2, RL4, RL5, RM1, SM1, SM2, WN1), metagranites (RX1, RX2, RX3), and one Slate Belt site (SB1), lie in a belt approximately parallel to major structural trends in the Appalachians. These values fall on or near the regression line (Fig. 3):

$$q = 29.8 \pm 1.5 \quad + \quad 7.8 \pm 0.475\ A \qquad (R = 0.97477)$$

The postmetamorphic granites fall on the high heat flow-heat production end of the line whereas the metamorphosed granites and country rocks fall on the low heat flow-heat production end of the line. Three exceptions to the linear relation defined by Piedmont granites are the Pageland granite (PG1; N.C.), Castalia pluton (CS1; N.C.), and the Palmetto granite (PM1; S.C.) Data from these sites were not used to determine the constants (q*, D) in the linear relationship.

The temperature gradient determined in the Pageland hole was not a conduction gradient, and a reliable heat flow determination could not be made. In addition, much of the recovered drillcore was intensely altered at low temperatures and its average heat production of 3.12 uW/m^3 is higher than the average heat production of 2.48 uW/m^3 found throughout the rest of the pluton. The higher heat production of granites altered at low temperatures occurs in several granite plutons of the southern Appalachians and suggests that the heat production of the Pageland drillcore is not representative of the pluton as a whole. Using the lower, average heat production of the Pageland pluton would move the point closer to the linear relation.

The heat production of the granite from the Castalia (CS1) drillcore is 2.34 uW/m^3 whereas an estimate of the heat production of the entire Castalia pluton is 3 uW/m^3, which brings the Castalia into closer agreement with the linear relationship established from other granite sites. The CS1 drillhole is located in the center of the pluton. A number of additional holes, too shallow for heat flow determinations, were drilled in the Castalia to obtain a better distribution of samples from this poorly exposed pluton. Except for the center of the pluton, the Castalia generally has a higher heat production. The heat production may still

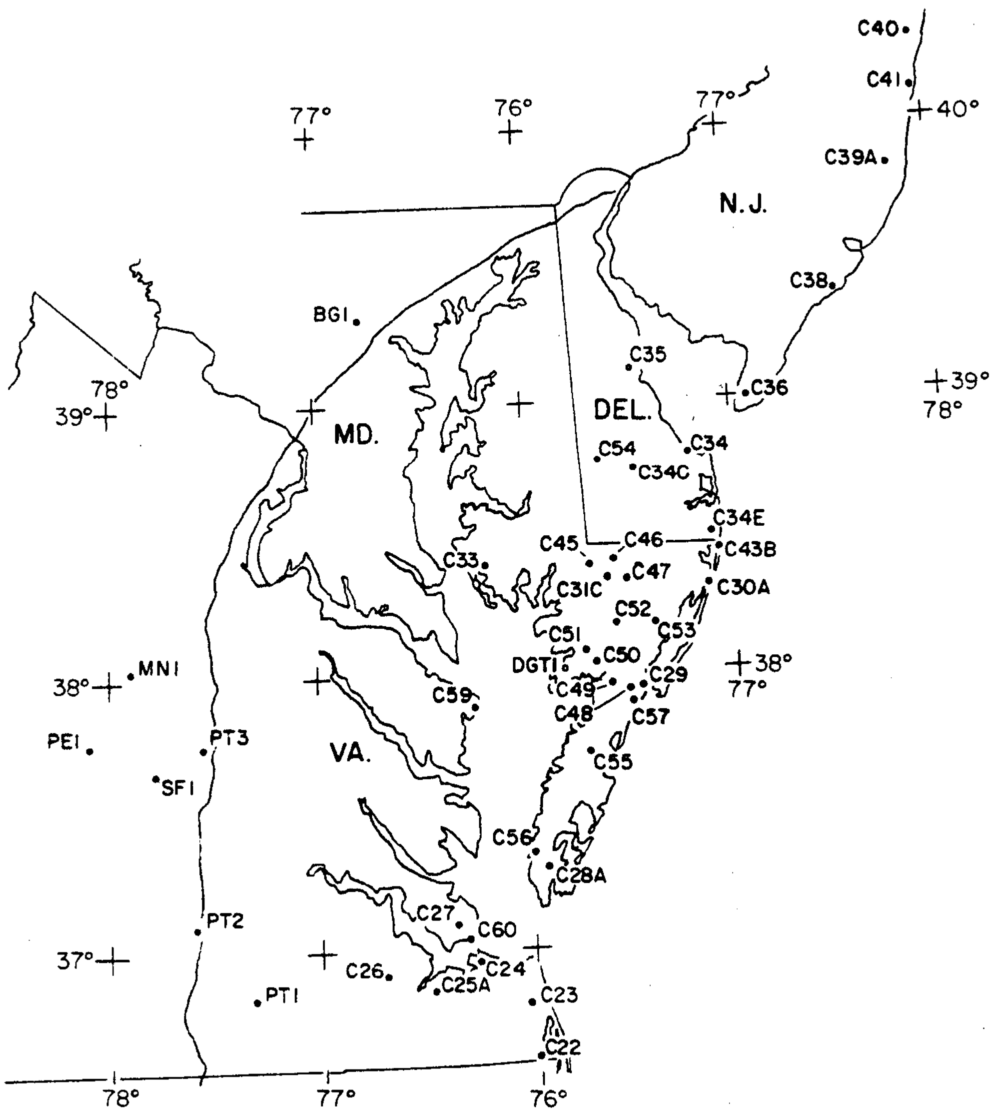

Figure 1. Heat flow sites in New Jersey, Delaware, Maryland, and Virginia.

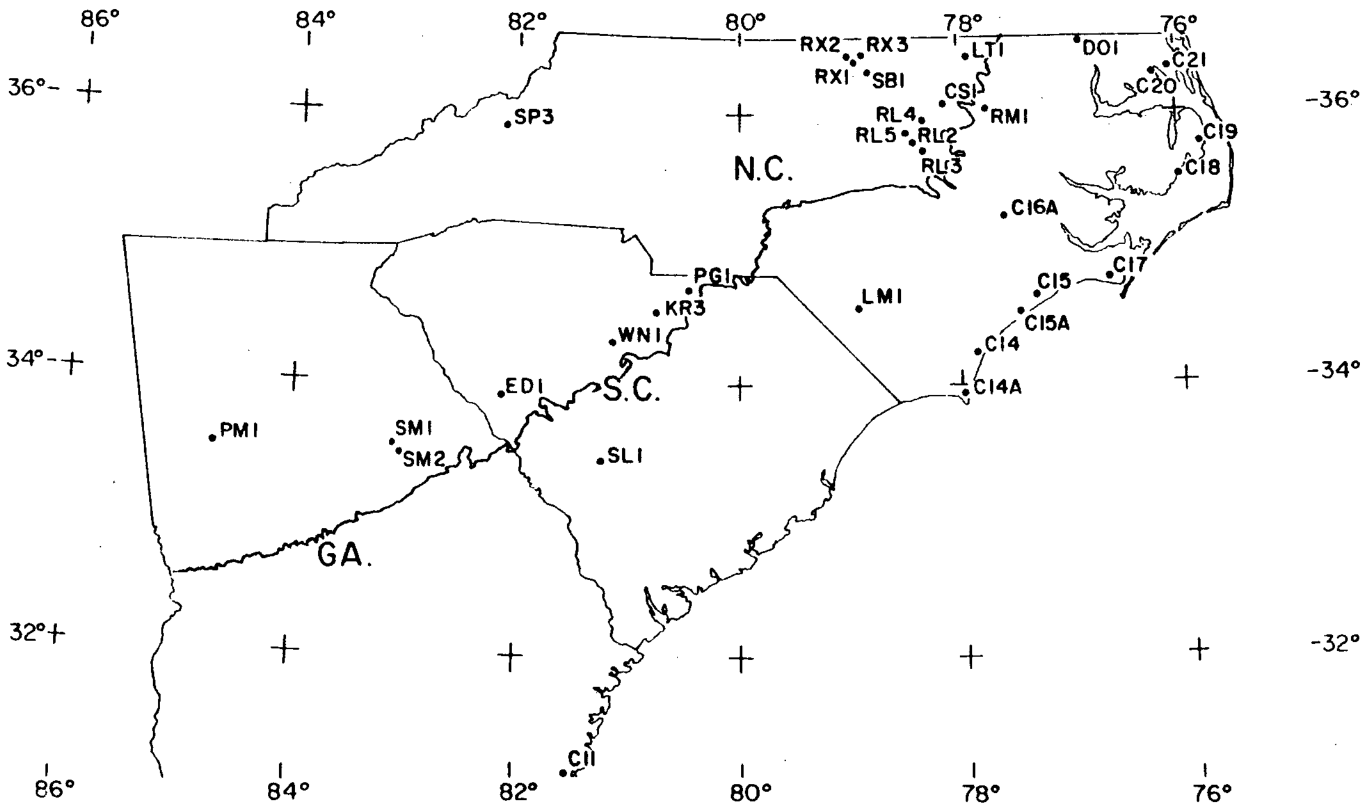

Figure 2. Heat flow sites in North Carolina, South Carolina, and Georgia.

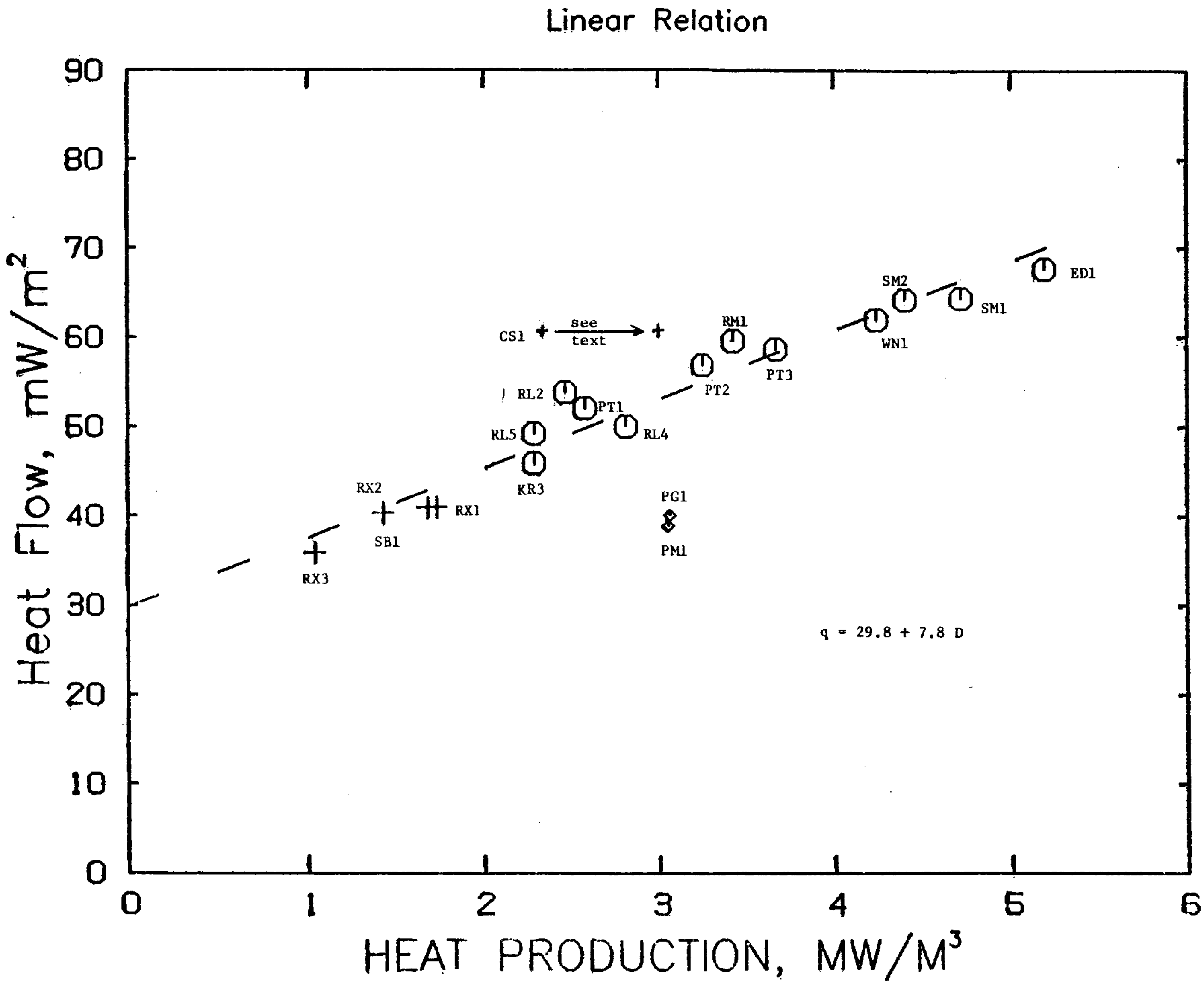

Figure 3. Linear relation between heat flow and heat generation in the Piedmont (Province 1).

be under-represented because of the greater number of samples near the center of the pluton. Redistribution of U and Th to the periphery of the Castalia pluton would not change the heat flow value (Green's theorem) but would raise the heat generation value.

The redistribution of U and Th in the Castalia pluton is indicated by a gamma-log of the CS1 drillhole. It shows a remarkably uniform distribution of heat-producing elements. With the exception of the Castalia, gamma logs of all the granite plutons register a high-level of radioactivity in aplite dikes. The Castalia is the only pluton for which a log of relatively constant activity was obtained, even across aplite dikes, implying a degree of uniformity of distribution of U and Th not found in other plutons of this age group. The uniformity of the gamma log, the absence of radioactivity from the aplite dikes, the enrichment of U + Th at the rim of the pluton, and the relatively high heat flow from the Castalia suggest that late magmatic fluids redistributed U + Th from the center of the pluton toward its rim.

The Palmetto granite is apparently sheared off at depth by the Brevard fault zone (Speer et al., 1980). The lower heat flow value from the Palmetto is consistent with an interpretation of D as a thickness parameter (Costain and Glover, 1980a,b).

Atlantic Coastal Plain

Fifty heat flow values were determined at sites located on the Atlantic Coastal Plain (Table 1). At nine of these sites, a minimum of 100 m of core was obtained from basement, and the heat flow value was determined over the cored basement interval. At all other sites on the Atlantic Coastal Plain, heat flow values were determined over an interval in the sedimentary section. All thermal conductivity determinations were made on core samples. Attempts were made to recover two 16-m cores, one sand, one clay, from each 300-m hole in Coastal Plain sediments. Core recovery was approximately 50%. Results are summarized in Table 1.

Drill holes from which basement core was recovered (BG1, DO1, LM1, SL1, C14A, C14, C15, C25A, C26, C32A) are of particular interest because of the relationship between heat flow and heat generation. Results from six of these holes (BG1, DO1, SL1, C14A, C25A, C26) define a second linear relation (Fig. 4) with a slope ($D = 8.0 \pm 0.380$ km; $R = 0.99550$) approximately parallel to the Piedmont relation, but with an intercept of 48.2 ± 0.8 mW/m^2.

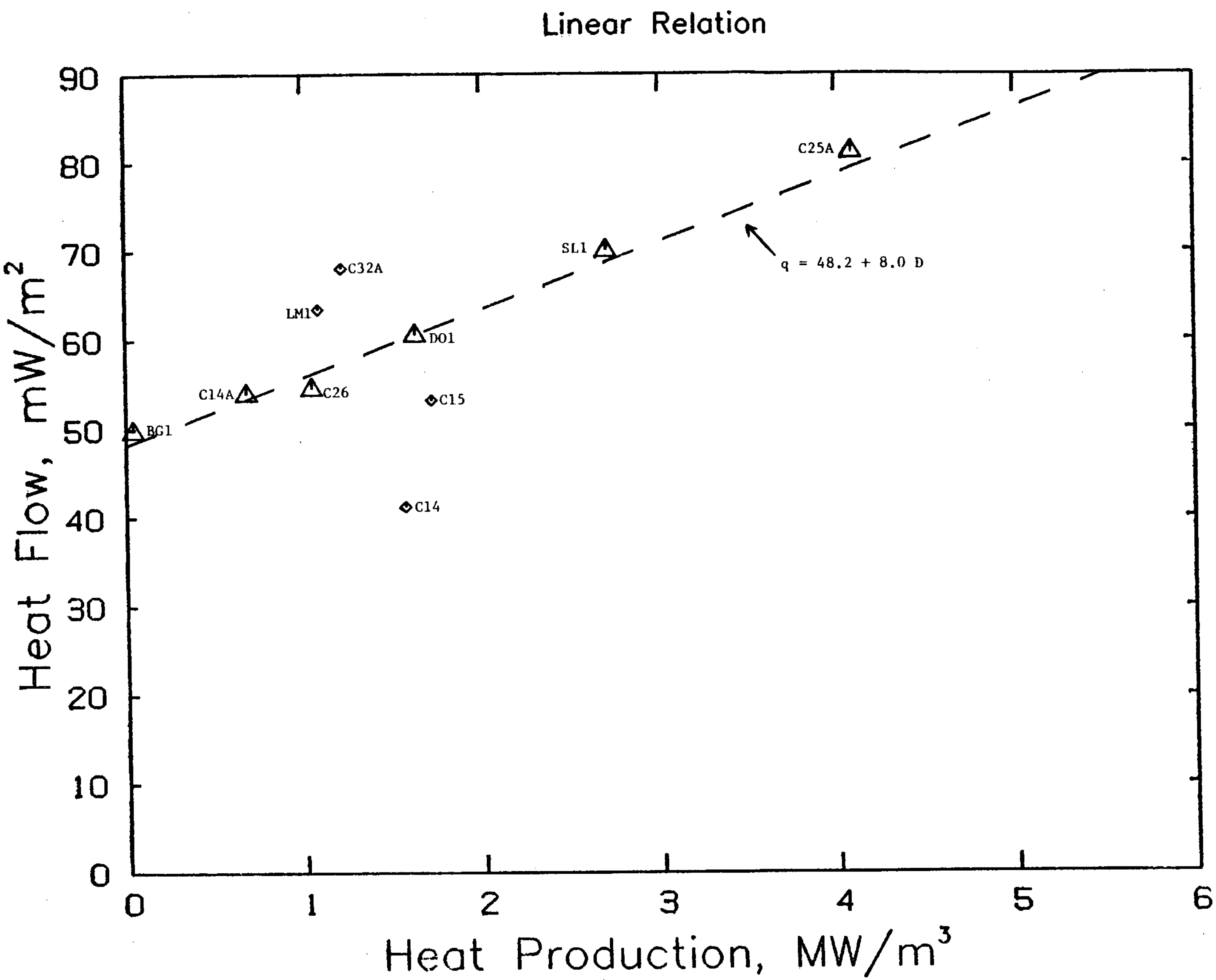

Figure 4. Heat flow versus heat production for Province II.

Drillholes in postmetamorphic granites, DO1 (Dort, N.C.), SL1 (Springfield, S.C.), and C25A (Portsmouth, Va.) lie at the high heat production-heat flow end of the line whereas the drillholes in metamorphosed rocks (C14, C14A, C15, C26, C32A, LM1) lie in a large scatter at the low end of the line. The drillcore sites for the granites were located in the centers of gravity minima and, presumably, are located in the centers of the granite plutons. The geological settings of these holes are fairly well understood, and the heat production and heat flow are believed to be representative of that section of the crust. By contrast, the drillcores in the metamorphosed rocks have generally unknown geological settings and it is not clear whether they are representative of that section of the crust. For example, the Lumberton, N.C. (LM1) drillhole penetrates metavolcanic rock but may be underlain by a granite (Pratt et al., 1982) and this would account for its lying above the line.

Discussion

Our preferred interpretation of the linear relation observed for the post- and synmetamorphic granite plutons in the Piedmont is that an essentially constant distribution of heat-producing elements extends uninterrupted to a depth, D. The granites have a homogeneous distribution of heat-producing elements across their surfaces, and evidently this same homogeneous distribution continues at depth.

Other evidence for D as a thickness parameter can be inferred from the results of recent seismic data. Clark et al. (1978) recorded large-amplitude reflections from beneath the Brevard fault zone near Rosman, N.C., approximately 150 km northwest of the Carolina slate belt. These originated from a depth of approximately 6 km and were interpreted to be from an unmetamorphosed Cambrian sedimentary sequence similar to sediments that crop out in the Valley and Ridge Province of northeastern Tennessee. According to this interpretation, below the Brevard fault zone are roots of Blue Ridge thrust nappes above a master sole fault that overrides a gently southeast-dipping autochthonous sequence of lower Cambrian carbonates and clastics. This interpretation was subsequently supported in Georgia by Cook et al. (1980). The root zone, if projected eastward, could be a major sole thrust at a depth of 6-8 km that truncates the post- to late-synmetamorphic granite plutons.

The COCORP seismic line of Cook et al. (1980) crosses the Elberton Granite in Georgia. Strong seismic events at about 2.5 sec are apparent on their record section. No events were recorded at earlier times (shallower depths) from within the Elberton Granite. For an assumed velocity of 6 km/sec for the Elberton Granite, the reflected events would originate at a depth of approximately 7.5 km in this area. A recent reinterpretation after reprocessing of the COCORP data in the southern Appalachians (Iverson and Smithson, 1982) shows a thrust fault at a depth of approximately 8 km beneath the Elberton. We have no heat flow data from the Elberton, but heat flow-heat generation values from the Siloam in Georgia and from the Winnsboro-Rion in South Carolina are from sites northeast and southwest of the Elberton and do fall on the regression line that defines the linear relationship. The age of the Elberton is 320 ± 20 m.y. (Ross and Bickford, 1980). The youngest regionally penetrative deformation that affected the Piedmont occurred near the end of the Paleozoic (Secor and Snoke, 1978; Kish and Fullagar, 1978; Snoke et al., 1980). Post-penetrative brittle deformation occurred, but the magnitude of displacement is unknown. The nature and timing of structural deformation as known to date in the Piedmont are consistent with possible truncation of post- and late-synmetamorphic granites by a master sole fault at a depth of about 8 km. Talwani (1982) notes a distinct velocity discontinuity at a depth of about 10 km in the meizoseismal area of the 1886 Charleston, S.C. earthquake, and suggests that the discontinuity may be associated with a decollement.

A sole fault at 8-9 km is indicated by recent VIBROSEIS data in central Virginia (Glover _et al_., 1982). This fault may truncate the Petersburg granite beneath site PT3 in Virginia, although the seismic line does not extend to PT3, nor far enough to the east to examine the crust beneath PT1 and PT2.

It is proposed that the occurrence of a master sole thrust that truncates granites in the allochthonous plate is directly responsible for the linear relation between heat flow and heat generation from the eastern Piedmont. The use of heat flow data to support an interpretation of D would be strengthened by seismic data at other heat flow sites in young granites. It is reasonable to expect some scatter in the linear relationship if the value of D is controlled by thickness of granite. Less scatter might be expected parallel to major structural trends of the Appalachians than across strike if large-scale allochthony of crystalline sheets is a valid explanation of the linear relationship in the southeast.

In support of D as a logarithmic decrement, the postmetamorphic granites in the southeastern U.S. are generally similar in appearance and form a subset of plutons whose field relations, mineralogy, petrography, chemistry, and crystallization histories are similar (Speer _et al_., 1980). Therefore, geochemical differences which might be expected to affect the value and interpretation of D as a logarithmic decrement should also be minimized. However, the interpretation of D as a logarithmic decrement would require an uninterrupted exponential decrease of U and Th to a depth of approximately 3D or about 20 km (Costain and Glover, 1980a,b). An uninterrupted exponential distribution of U and Th to a depth of 20 km is inconsistent with interpretations of COCORP seismic data beneath the Elberton granite in Georgia (Cook _et al_., 1981; Cook _et al_., 1980; Iverson and Smithson, 1982).

The heat flow-heat production of the basement drillholes of the Atlantic Coastal Plain could imply a different heat flow province (province II?) with a higher heat flow from the mantle and lower crust. It will be important to confirm this second linear relationship because the higher value of q* could imply a chemically different lower crust and upper mantle. Pollack and Chapman (1977) proposed a correlation between mean heat flow, q_0 and q*,

$$q^* = 0.6q_0,$$

suggesting that 60% of the mean flux comes from the lower crust and upper mantle. If applicable to the southeast, this would suggest that the mean heat flow for province II(?) is approximately 80 mW/m^2. The high heat flow (102 mW/m^2) at Stumpy Point, N.C. is consistent with this interpretation.

Piedmont gravity gradient, crust beneath Carolina slate belt and Charlotte belt, and relevence to heat flow. The Appalachian gravity gradient, approximately 100 km in width (Griscom, 1963) probably marks an abrupt fundamental change in the crust for the entire length of the Appalachians between the Gaspe Peninsula and Alabama (Gravity Anomaly Map of the United States, 1982). Perhaps the least understood aspect of the gradient is its steep slope, approximately 10 km in width. Analysis of the steep part of the gradient by Griscom indicates that the top of the source of the anomaly is in the upper crust. For much of the distance between the Gaspe Peninsula and Alabama, the gradient is approximately coincident with the western exposed margin of billion-year-old basement rocks (Rankin, 1975). East of the gravity gradient, the largest Bouguer anomalies are over the exposed Carolina slate belt in North Carolina.

A number of plate tectonic models have been proposed for the southern Appalachians. Glover _et al_. (1978) interpret the Carolina slate belt volcanics in the eastern Piedmont of Virginia as continental margin volcanics deposited on Grenville basement much like that in the Blue Ridge of Virginia and North Carolina. Butler and Fullagar (1978) and Whitney _et al_. (1978) argue for a mafic crust and an island arc origin for the volcanics. Hatcher and Zietz (1980, Fig. 6) propose that mafic crust underlies the Charlotte belt and Carolina slate belt east of the Kings Mountain belt, that the Inner Piedmont is allochthonous, but that the slate belt and Charlotte belt are autochthonous. The large (+25 mgals) positive gravity values over the Charlotte belt and Carolina slate belt are taken to indicate a thick, mafic, presumably non-Grenville, crust. Long (1979) used gravity data to interpret the Carolina slate belt in Georgia and South Carolina as the axis of a continental rift or rift system, but his model did not incorporate the later reflection seismic data of Cook et al. (1980) that suggests allochthony west of the Charlotte belt and possibly beneath the Charlotte belt. Black (1980) showed that the chemical characteristics of metavolcanics in the Carolina slate belt suggest a crustal thickness "similar to that present beneath the Carolina slate belt today"; however, if the slate belt is allochthonous, then the nature of the crust beneath the developing slate belt may be no indication of the composition of the crust beneath the slate belt today. In any event, Black presents several lines of geochemical evidence that the Carolina slate belt volcanic pile developed on a continental margin 20 to 25 km in thickness, rather than as an arc on oceanic crust.

The discovery of Grenville basement in the eastern Piedmont of Virginia by Glover et al. (1978) and Black's (1980) geochemical data suggest that orogenic models involving oceanic crust beneath the developing slate belt in the Piedmont of Virginia and North Carolina are less probable. The problem is whether the crust beneath the slate belt today is continental or mafic. None of the tectonic models proposed to date has been constrained by heat flow data which might help to choose between models with either mafic or granitic crust beneath the present Carolina slate belt.

The prominent Appalachian gravity gradient in the eastern U.S. extending from Alabama to New England has been interpreted by Hatcher and Zietz (1980) as the eastern edge of the North American craton. The steepest part of the gradient is located at about the -25 or -20 mgal contour, but varies locally. The steep part of the gradient is approximately over the southeastern boundary of the Inner Piedmont (Kings Mountain belt) in Georgia, extends to the northeast over the northwestern edge of the Charlotte belt in North and South Carolina, and is approximately coincident with the northwestern margin of the Carolina slate belt in North Carolina and Virginia. Significantly, the steep part of the gradient obliquely cuts across structural trends exposed in the Piedmont of North Carolina. Thus, the source of the gradient cannot be determined from surface exposures.

Additional geophysical data are now available to support the interpretation of the coincidence of the steep part of the Piedmont gravity gradient with the edge of a volcanic pile which, in some locations if not all, has been thickened by overthrusting. In central Virginia in Goochland County, VIBROSEIS data indicate that Chopawomsic (continental margin?) volcanics are thrust over pre-Cambrian Catoctin (rift) metamorphosed basalts and Lynchburg graywackes; the sole fault is at a depth of approximately 9 km (Glover et al., 1982). According to Nelson (1962) the thickness (duplicated by faulting) of Catoctin reaches 7 km where the steep part of the Piedmont gravity gradient coincides with the Catoctin exposed on the Southwestern Mountains in Albermarle County, Va. The James River seismic data confirm that the thickness of the volcanic pile is approximately 8 km, and that the volcanics are thrust over Grenville basement (Glover et al., 1982). East of the steep part of the gravity gradient, Reilly (1980) interpreted the more regional Piedmont gravity high (gravity field greater than +10 mgals) as resulting from crustal thinning following the crustal model of James et al. (1968).

Best et al. (1973) concluded that the steep gradient in North Carolina near the surface contact of rocks of the Charlotte belt on the northwest and Carolina slate belt on the southeast could not be caused by an anomalous mass distribution located near the base of the crust. Gradients calculated for such a body were clearly not steep enough to account for the observed slope (Best et al., 1973, Fig. 3). They concluded that the source of the steep gravity gradient in North Carolina near Lexington must be located within the upper crust, that their best model consisted of a steep-sided (90^0) semi-infinite slab 7.6 km thick located at the surface of the ground with a density contrast of + 0.213 gm/cm^3, and with the higher density rocks causing the anomaly in the Carolina slate belt; however, they also note that the surface geology of the slate belt-Charlotte belt in this area seems to contradict their inferred density contrast of +0.213 gm/cm^3. Best et al. (1973) note that the anomalous high density mass could be buried several kilometers without disturbing the fit of the model data to the actual data as long as a constant thickness-density contrast was maintained.

We suggest that the source of the steep part of the gravity gradient northeast of Lexington and elsewhere in the eastern United States is an edge of rift-related metabasalts and metasandstones (Catoctin and Catoctin equivalents), similar to those along the James River traverse in Virginia.

Eastward-dipping reflections (and same data reprocessed by Iverson and Smithson, 1982) from the Georgia COCORP line from beneath the Elberton granite (Cook et al., 1982, Fig.

7) could also originate from a thick (12 km?) sequence of thrust-duplicated rift-related(?) volcanics. The seismic signature of the COCORP reflections from beneath the Elberton on Georgia Line 1 is similar to that recorded 360 km to the northeast at Lumberton, N.C. (Pratt et al., in preparation) where drill core has tied excellent reflections to a thick (minimum 4.5 km) sequence of low-grade metavolcanics similar to those exposed in the slate belt to the northwest. The seismic data of Pratt et al. suggest that the slate belt volcanics have been thrust over another sequence of volcanics(?) (rift-related?). The sole fault postulated by Pratt et al. is at a depth of approximately 17 km. No estimate is available of the thickness of volcanics(?) below the sole fault.

Although heat flow is relatively insensitive to changes of a few km in crustal thickness, the reduced heat flow (30 mW/m^2) for the southeastern U.S. is essentially the same as that computed by Slack (1974) for the mantle heat flow beneath the North American craton and shield (28.5 mW/m^2). The linear relation between heat flow and heat generation discovered by Birch et al. (1968) defines q^* and D as 35 mW/m^2 and 6.3 km, respectively. The locations where this relation was developed encompass northeastern New York (Adirondacks) as well as south central Vermont and central New Hampshire. Current interpretations of the meaning of q^* would thus imply that no significant differences in chemical composition of the lower crust and upper mantle that would affect the distribution of heat-producing elements occur over this region, and that Grenville crust therefore underlies these heat flow sites in New York and New England. The sites are located both east and west of the gravity gradient. Grenville basement, possibly thinned, can therefore be expected to underlie that part of the southeastern U.S. where the reduced heat flow is about 30 mW/m^2.

Conclusions

Seismic and gravity data suggest that the source of the steep part of the Piedmont gravity gradient, at least in central Virginia, is an edge of Catoctin rift-related metamorphosed basalts and metasandstones, thickened by overthrusting. Crustal thinning at the Moho cannot explain the steepness of the gravity gradient, but is consistent with the higher regional gravity values east of the gradient. A linear relationship between heat flow and heat production is observed at sites east and west of the steep part of the gravity gradient in the northeast and southeast U. S. Because of the occurrence of the linear relationship between heat flow and heat production, the heat flow data suggest that the Carolina slate belt, Charlotte belt, and Raleigh belt are allochthonous (because of the value of D), and that Grenville crust underlies these areas (because of the value of q*).

A single value of q* may not be appropriate for the southeastern U. S. Deviations from q = 30 + 7.8 A at sites on the Atlantic Coastal Plain may not be attributed everywhere simply to changes in crustal thickness, D, but may instead indicate one or more heat flow provinces. Our new heat flow data support the interpretation of a second heat flow province for the southeast United States.

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Table 1. SUMMARY OF HEAT FLOW and HEAT PRODUCTION DATA
(Scriptfile HFTABLE)
February 23, 1983

Piedmont, Blue Ridge, and Valley and Ridge Provinces

HOLE	LATITUDE (North)	LONGITUDE (West)	DEPTH INTERVAL, (m)	GRADIENT, °C/Km	THERMAL CONDUCTIVITY, W/m-°C	HEAT FLOW, q,mW/m²	QUAL	HEAT PRODUCTION, A,uW/m³
CUFFYTOWN CREEK PLUTON, SC								
ED1	33°55'11"	82°07'10"	182.5-285.0	17.65 ±0.03(1.000 42)	3.84 ± 0.10(29)	67.7 ± 1.7	1	5.2 ±0.4(17)
LIBERTY HILL PLUTON, SC								
KR3	34°32'20"	80°44'51"	351.8-384.3	14.90 ±0.01(1.000 14)	3.08 ± 0.30(12)	45.9 ± 4.5	1	2.3 ±0.8(15)
PETERSBURG GRANITE, VA								
PT1	36°49'45"	77°19'15"	190.0-251.0	18.92 ±0.08(0.998 123)	2.75 ± 0.31(23)	52.0 ± 5.9	1	2.6 ±0.2(10)
PT2	37°05'44"	77°35'37"	70.1-210.1	17.30 ±0.01(1.000 276)	3.28 ± 0.18(61)	56.8 ± 3.2	1	3.3 ±0.7(18)
PT3	37°45'11"	77°33'02"	107.3-201.0	18.37 ±0.02(1.000 185)	3.19 ± 0.10(42)	58.6 ± 1.8	1	3.7 ±0.7(16)
CASTALIA PLUTON, NC								
CS1	36°04'15"	78°07'15"	142.2-209.7	19.26 ±0.03(1.000 28)	3.15 ± 0.16(26)	60.7 ± 3.1	1	2.3 ±0.1(15)
ROLESVILLE BATHOLITH, NC								
RL2	35°47'28"	78°25'04"	192.2-204.7	18.77 ±0.04(1.000 6)	2.97 ± 0.14(5)	55.7 ± 2.7		
			104.7-114.7	16.84 ±0.08(1.000 5)	3.08 ± 0.18(5)	51.9 ± 3.1		
					Best value for RL2	53.8 ± 2.7	1	2.5 ±0.3(11)
RL3	35°43'36"	78°19'45"	42.4- 94.9	13.57 ±0.15(22)	3.44 ± 0.29(12)	46.9 ± 4.6		
			97.4-129.9	13.79 ±0.10(14)	3.29 ± 0.45(15)	45.6 ± 6.7		
					Best value for RL3	46.3 ± 0.9	1	
RL4	35°57'05"	78°19'45"	155.2-192.2	16.91 ±0.05(1.000 16)	2.96 ± 0.27(16)	50.0 ± 4.5	1	2.8 ±0.5(19)
RL5	35°51'17"	78°28'54"	174.8-209.8	16.97 ±0.04(1.000 15)	2.90 ± 0.17(12)	49.2 ± 2.9	1	2.3 ±0.5(10)
ROXBORO METAGRANITE, NC								
RX1	36°23'12"	78°58'00"	149.3-169.3	10.81 ±0.04(1.000 9)	3.80 ± 0.06(10)	41.1 ± 0.6		
			229.3-246.8	10.71 ±0.07(1.000 8)	3.82 ± 0.07(10)	40.9 ± 0.8		
					Best value for RX1	41.0 ± 0.7	1	1.7 ±0.1(7)
RX2	36°25'31"	79°01'53"	131.8-209.3	11.14 ±0.03(1.000 32)	3.67 ± 0.17(28)	40.9 ± 1.9	1	1.7 ±0.1(7)
RX3	36°25'39"	78°53'42"	139.9-194.9	10.31 ±0.15(0.998 10)	3.48 ± 0.24(14)	35.9 ± 2.6	1	1.0 ±0.2(8)
SLATE BELT, NC								
SB1	36°18'42"	78°50'00"	124.2-209.2	11.99 ±0.06(0.999 26)	3.36 ± 0.28(28)	40.3 ± 3.3	1	1.4 ±0.1(5)
SPRUCE PINE, NC (Blue Ridge province)								
SP3	35°54'50"	82°07'18"		15.00		42.7		
SILOAM PLUTON, GA								
SM1	33°32'31"	83°06'40"	172.6-199.1	19.07 ±0.05(1.000 54)	3.38 ± 0.23(11)	64.4 ± 4.4	1	4.7 ±0.6(16)
SM2	33°28'41"	83°02'03"	104.5-205.5	19.30 ±0.03(1.000 203)	3.33 ± 0.29(28)	64.2 ± 5.6	1	4.4 ±0.4(14)
WINNSBORO PLUTON, SC								
WN1	34°18'48"	81°08'42"	166.7-316.7	18.34 ±0.03(1.000 61)	3.38 ± 0.10(5)	62.0 ± 1.9	1	4.2 ±0.4(14)

BALTIMORE GABBRO, MD									
BG1	39°19'15"	76°46'16"	254.6-282.6	15.56 ±0.04(1.000 56)	3.20 ± 0.32(14)	49.8 ± 5.0		1	0.0
ROCKY MOUNT, NC									
RM1	36°02'10"	77°45'14"	91.5-126.7	19.16 ±0.02(1.000 71)	3.11 ± 0.08(18)	59.6 ± 1.6		1	3.4 ±0.3(5)
PAGELAND, SC									
PG1	34°42'02"	80°27'52"	170.0-187.5	12.91 ±0.07(1.000 44)	3.10 ± 0.15(9)	40.1 ± 1.9		1	3.1 ±0.5(14)
MINERAL, VA									
MN1	38°02'14"	77°53'45"	284.0-319.3	16.26 ±0.04(0.999 111)	2.38 ± 0.19(11)	38.7 ± 3.2			
			326.5-350.1	17.34 ±0.06(0.999 79)	2.13 ± 0.31(7)	37.0 ± 5.4			
					Best value for MN1	37.8 ± 0.6		1	
PEARISBURG, VA (Valley and Ridge province. Anomalous high heat flow. See text.)									
CL1	37°20'31"	80°46'11"	113.4-171.4	18.90 ±0.04(0.999 115)	6.07 ± 0.24(28)	114.7 ± 4.5			
			188.2-195.4	21.11 ±0.16(0.999 15)	6.23 ± 0.06(3)	131.6 ± 1.7			
					Best value for CL1	123.2 ±11.9		1	
PALMETTO, GA									
PM1	33°29'55"	84°41'58"	89.5-148.5	14.74 ±0.04(118)	2.55 ± 0.42(20)	37.5 ± 3.			
			149.5-160.0	11.92 ±0.14(21)	3.13 ± 0.46(4)	37.3 ± 5.7			
			161.4-205.0	17.08 ±0.30(90)	2.45 ± 0.20(20)	41.9 ± 3.6			
					Best value for PM1	38.9 ± 2.5		1	3.1 ±0.3(16)
PEGMATITE BELT, VA									
PE1	37°45'56"	78°05'37"	121.8-201.8	15.58 ± 0.09(0.999 33)	2.63 ± 0.46(55)	41.0 ± 7.2		1	
LITTLETON, NC									
LT1	36°24'59"	77°54'59"	111.1-207.8	18.39 ±0.02(1.000 191)	2.74 ± 0.3(34)	50.5 ± 5.1		1	
STATE FARM, VA									
SF1	37°40'01"	77°46'48"	50.0- 92.5	15.29 ±0.10(0.999 18)	2.42 ± 0.3(10)	37.0 ± 4.0			
			185.0-192.5	17.60 ±0.57(0.998 4)	2.52 ±0.3(4)	44.4 ± 6.0			
			197.5-207.5	16.40 ±0.77(0.993 5)	2.37 ±0.1(3)	38.8 ± 2.6			
					Best value for SF1	40.1 ± 3.9		1	

<u>Atlantic</u> <u>Coastal</u> <u>Plain</u>

COLONEL'S ISLAND, GA									
CI1	31°08'12"	81°32'34"	478.4-627.5	19.53 ± 0.03(0.999 294)	2.04 ±0.32(5)	39.8 ± 6.2			
SPRINGFIELD, SC (C9 = SL1)									
SL1 (SL1)	33°26'12"	81°14'12"	292.5-346.9	21.32 ± 0.03(1.000 108	3.29 ±0.17 (34)	70.1 ±3.6	B	1	2.70 ± 0.3(5
DORT, NC									
DO1	36°31'43"	76°52'32"	328.3-412.8	23.19 ±0.02(1.000 167)	2.61 ± 0.22(42)	60.6 ± 5.2	B	1	1.63 ± 0.12(9)
LUMBERTON, NC									
LM1	34°34'12"	78°56'02"	249.8-274.7	21.16 ±0.023(1.000 50)	3.27 ± 0.18(5)	69.3 ± 3.8			
			291.5-301.2	22.62 ±0.08(1.000 20)	2.77 ± 0.13(3)	62.6 ± 2.9			
			307.3-327.1	19.93 ±0.03(1.000 40)	2.94 ± 0.33(4)	58.6 ± 6.6			
					Best value for LM1	63.5 ± 5.4		1	1.09 ±0.08(11)

SOUTHPORT, NC								
C14A	33°56'52"	77°59'01"	200.8-206.4	28.06 ± 0.55(.996 12)	1.91 ±0.44 (19)	53.6 ±12.5 S		
			361.6-371.3	27.36 ±0.47(.995 20)	2.01 ±0.42 (59)	54.9 ±11.4 S		
			478.2-545.4	18.62 ±0.02(1.000 133)	2.89 ±0.37 (18)	53.7 ± 6.8 B		
					Best value for C14A	54.0 ± 0.7	1	0.68 ± 0.2(13)
WILMINGTON, NC								
C14	34°13'46"	77°51'39"	257.6-262.7	26.24 ± 0.48(.997 11)	1.57 ±0.13 (7)	41.2 ±3.4 S	1	2.1±0.0
			385.0-475.0	16.63 ± 0.48				
SNEADS FERRY, NC								
C15A	34°31'36"	77°27'18"	444.0-453.6	33.83 ±0.64(.994 20)	0.98 ±0.25 (10)	33.1 ±8.43 S	3	
JACKSONVILLE, NC								
C15	34°39'07"	77°19'14"	363.5-371.1	37.39 ±1.61(.975 16)	1.48 ±0.08 (28)	55.1 ±3.9 S		
			558.5-594.6	18.80 ±0.03(1.000 72)	2.73 ±0.12 (10)	51.4 ±2.3 B		
					Best value for C15	53.2 ±2.6	1	1.72 ± 0.4(8)
KINSTON, NC								
C16A	35°14'11"	77°35'41"	81.0- 87.1	25.19 ±1.09(.980 13)	1.05 ±0.27 (11)	26.3 ±6.9 S		
			174.7-180.8	27.67 ±0.75(.992 13)	2.57 ±0.52 (11)	71.2 ±14.65 S		
					Best value for C16A	48.7 ±31.7	3	
BEAUFORT, NC								
C17	34°46'18"	76°39'42"	192.7-201.9	31.53 ±0.76(.990 19)	1.39 ±0.31 (14)	43.9 ±10.0 S	3	
ENGELHARD, NC								
C18	35°31'12"	75°59'16"	207.0-219.2	49.16 ±0.37(.999 25)	1.13 ±0.27 (31)	55.6 ±13.2 S		
			296.9-303.5	23.29 ±0.69(.990 14)	2.07 ±0.39 (22)	48.3 ±9.1 S		
					Best value for C18	51.9 ±5.2	1	
STUMPY POINT, NC								
C19	35°45'07"	75°47'39"	196.4-205.5	63.05 ±0.70(.998 19)	1.62 ±0.10 (14)	102.2 ±6.2 S	2	
SP19 (near C19) - heat production from core recovered by others from basement								2.31 ± 0.3 (7)
ELIZABETH CITY, NC								
C20	36°16'49"	76°12'35"	134.7-138.8	28.39 ±1.10(.990 9)	1.67 ±0.25 (13)	47.4 ±7.3 S		
			300.7 306.3	23.72 ±0.70(.991 12)	1.48 ±0.16 (8)	35.0 ±3.9 S		
					Best value for C20	41.2 ±8.8	2	
BELCROSS, NC								
C21	36°19'35"	76°03'32"	172.1-180.2	44.71 ±0.91(.994 17)	1.18 ±0.14 (19)	52.7 ±6.2 S		
			299.3-307.5	27.74 ±1.02(.980 17)	1.77 ±0.26 (15)	49.0 ±7.3 S		
					Best value for C21	50.8 ±2.6	1	
CREEDS, VA								
C22	36°36'23"	76°00'26"	196.7-207.4	44.04 ±0.50(.997 22)	1.30 ±0.06 (20)	57.0 ±2.7 S	1	
OCEANA, VA								
C23	36°48'05"	76°02'30"	181.5-188.1	46.15 ±1.04(.994 14)	1.30 ±0.27 (12)	60.0 ±12.5 S		
			293.2-299.8	33.66 ±1.18(.986 14)	1.33 ±0.19 (25)	44.7 ±6.7 S		
					Best value for C23	52.3 ±10.8	2	
NORFOLK, VA								

C24	36°57'24"	76°16'09"	164.4-171.5	42.10 ±0.67(.997 15)	1.44 ±0.21 (46)	60.5 ±8.7 S		
			301.0-314.2	24.93 ±0.45(.992 27)	1.89 ±0.55 (40)	47.2 ±13.7 S		
					Best value for C24	53.8 ±9.4	2	
PORTSMOUTH, VA								
C25A	36°51'01"	76°28'50"	295.8_306.0	34.33 ±0.24(.999 21)	2.43 ±0.30 (48)	83.5 ±10.4 S		
			557.0-610.4	22.70 ±0.02(1.000 106)	3.48 ±0.15 (27)	78.9 ±3.4 B		
					Best value for C25A	81.2 ±3.2	1	4.09 ± 0.3(8)
ISLE OF WIGHT, VA								
C26	36°54'31"	76°42'08"	297.5-304.5	26.83 ±0.70(.991 15)	2.21 ±0.23 (20)	59.3 ±6.5 S		
			287.1-332.4	21.77 ±0.20(.999 10)	2.29 ±0.18 (11)	49.9 ±4.0 S		
					Best value for C26	54.6 ±6.6	2	1.1 ± 0.5(7)
CHERITON, VA								
C28A	37°17'47"	75°55'52"	211.0-223.7	44.24 ±0.77(.993 26)	1.35 ±0.11 (30)	59.9 ±4.9 S		
			295.9-307.0	66.83 ±4.95(.897 23)	0.89 ±0.08 (26)	59.4 ±6.7 S		
					Best value for C28A	59.6 ±0.3	1	
WALLOPS ISLAND, VA								
C29	37°56'36"	75°27'16"	180.0-185.6	45.74 ±1.03(.995 12)	1.52 ±0.04 (6)	69.4 ±2.4 S		
			296.4-299.9	19.21 ±8.97(.696 4)	1.94 ±0.18 (8)	37.2 ±17.7 S		
					Best value for C29	53.3 ±22.8	3	
OCEAN CITY AIRPORT, MD								
C30A	38°18'37"	75° 07'04"	162.0-167.6	38.58 ±1.24(.990 12)	1.53 ±0.15 (06)	59.2 ±6.1 S	2	
SALISBURY, MD								
C31C	38°20'33"	75°36'26"	226.3-230.4	37.40 ±0.92(.996 9)	1.70 ±0.06 (6)	63.6 ±2.8 S		
			297.9-302.5	47.71 ±5.56(.899 10)	1.25 ±0.17 (20)	59.6 ±10.9 S		
					Best value for C31C	61.6 ±2.8	1	
CRISFIELD, MD (heat production from basement core of nearby deep geothermal test, DGT-1)								
C32A	38°00'58"	75°49'34"	163.4-173.0	61.23 ±0.50(.999 20)	1.11 ±0.12 (59)	68.1 ±7.3 S	2	1.2 ± 0.2(6)
GOLDEN HILL, MD								
C33	38°24'08"	76°11'11"	168.9-179.6	55.10 ±0.71(.997 22)	1.07 ±0.14 (13)	58.9 ±7.6 S	2	
ASSAWOMAN BAY, DE								
C34E	38°29'54"	75°05'49"	187.9-197.0	19.77 ±0.30(.996 19)	1.63 ±0.47 (11)	32.2 ±9.2 S	3	
ELLENDALE ST. FOREST, DE								
C34C	38°44'35"	75°27'43"	176.6-183.8	31.41 ±0.44(.997 15)	1.41 ±0.18 (10)	44.3 ±5.7 S	2	
LEWES, DE								
C34	38°48'11"	75°11'40"	159.3-172.1	26.82 ±0.35(.996 26)	1.63 ±0.33 (26)	43.8 ±8.8 S	2	
DOVER, DE								
C35	39°06'42"	75°27'41"	231.0-241.7	49.09 ±0.41(.999 22)	1.26 ±0.19 (61)	61.7 ±9.1 S	2	
CAPE MAY, NJ								
C36	39°00'02"	74°54'07"	137.1-144.2	28.59 ±0.61(.994 15)	1.58+/0.28 (5)	45.2 ±8.0 S	2	
ATLANTIC CITY, NJ								

C38	39°22'46"	74°27'14"				42.4	
FORKED RIVER, NJ							
C39A	39°50'26"	74°10'52"	222.3-229.0	33.52+/1.28(.983 14)	1.38 ±0.13 (13)	46.4 ±4.8 S	1
EATONTOWN, NJ							
C40	40°18'49"	74°03'01"	230.2-237.8	20.85 ±0.81(.979 16)	1.59 ±0.47 (10)	33.1 ±10.0 S	
			292.2-303.4	22.45 ±0.29(.997 23)	2.29 ±0.46 (14)	51.5 ±10.3 S	
					Best value for C40	42.3 ±13.0	3
SEA GIRT, NJ							
C41	40°07'15"	74°02'15"	227.9-				
OCEAN CITY, MD							
C43B	38°26'02"	75°03'34"	289.1-300.9	36.95 ±0.35(.998 24)	2.02 ±0.29 (6)	74.5 ±10.6 S	
			363.5-377.2	26.69 ±0.42(.994 28)	1.29 ±0.18 (24)	34.5 ±4.8 S	
					Best value for C43B	54.5 ±28.3	3
SALISBURY AIRPORT, MD							
C45	38°23'26"	75°41'24"	213.9-223.0	52.88 ±0.54(.998 19)	1.01 ±0.05 (17)	53.5 ±2.5 S	
			295.8-307.0	45.91 ±0.40(.998 23)	1.12 ±0.11 (24)	51.3 ±5.0 S	
					Best value for C45	52.4 ±1.5	1
SALISBURY(ZION RD), MD							
C46	38°23'58"	75°34'29"	224.1-232.2	31.80 ±0.61(.995 17)	1.59 ±0.17 (17)	50.7 ±5.5 S	2
SALISBURY-HEBRON, MD							
C47	38°19'55"	75°30'40"	233.0-242.7	31.22 ±0.64(.992 20)	1.49 ±0.24 (24)	46.6 ±7.5 S	1
WATTSVILLE, VA							
C48	37°56'04"	75° 30'36"	50-302	35			
WITHAMS, VA							
C49	37°57'31"	75°35'38"	209.2-221.4	49.93 ±0.46(.998 25)	1.31 ±0.20 (22)	65.6 ±9.9 S	2
REHOBETH, MD							
C50	38°02'07"	75°40'17"	226.1-239.8	38.43 ±0.32(.998 28)	1.13 ±0.27 (25)	43.3 ±10.3 S	
			293.8-301.4	46.57 ±1.33(.989 16)	1.36 ±0.22 (8)	63.4 ±10.4 S	
					Best value for C50	53.3 ±14.2	3
KINGSTON, MD							
C51	38°04'52"	75°43'32"	213.0-222.7	48.76 ±1.36(.986 20)	1.33 ±0.24 (26)	64.9 ±11.8 S	
			296.5-302.6	48.36 ±1.04(.995 13)	1.28 ±0.09 (10)	61.9 ±4.6 S	
					Best value for C51	63.4 ±2.1	1
PRINCESS ANNE, MD							
C52	38°10'30"	75°34'29"	212.2-223.4	52.16 ±0.78(.995 23)	0.98 ±0.23 (20)	51.1 ±12.2 S	
			300.1-310.2	52.70 ±0.86(.995 21)	1.13 ±0.11 (23)	59.6 ±6.1 S	
					Best value for C52	55.3 ±6.0	1
SNOW HILL, MD							
C53	38°10'18"	75°22'50"	234.4-242.6	33.01 ±0.51(.996 17)	1.36 ±0.31 (16)	45.0 ±10.2 S	
			297.0-306.7	40.60 ±0.84(.992 20)	1.32 ±0.10 (21)	53.7 ±4.2 S	
					Best value for C53	49.3 ±6.1	2

BRIDGEVILLE, DE							
C54	38°46'43"	75°37'49"	289.1-300.8	51.18 ±0.31(.999 24)	1.00 ±0.12 (5)	51.2 ±6.3 S	1
TASLEY, VA							
C55	37°42'32"	75°42'51"	164.4-173.6	32.26 ±0.43(.997 19)	2.05 ±0.23 (81)	66.1 ±7.5 S	1
EASTVILLE, VA							
C56	37°21'16"	75°59'34"	118.1-124.7	24.79 ±0.96(.982 14)	2.24 ±0.28 (30)	55.5 ±7.3 S	
			121.1-125.7	25.77 ±1.84(.961 10)	2.41 ±0.16 (17)	62.0 ±6.0 S	
					Best value for C56	58.7 ±4.6	1
ATLANTIC, VA							
C57	37°53'14"	75°30'02"	175.0-186.7	41.96 ±0.45(.997 24)	1.63 ±0.34(110)	68.3 ±14.5 S	2
SMITH POINT, VA							
C59	37°53'01"	76°15'05"	144.1-155.3	51.18 ±0.73(.996 23)	1.24 ±0.09 (77)	63.3 ±4.9 S	1
HAMPTON, VA							
C60	37°02'13"	76°18'54"	278.5-285.2	30.07 ±1.11(.984 14)	1.95 ±0.38 (31)	58.5 ±11.7	
			251.6-291.8	27.53 ±0.06(1.000 80)	1.95 ±0.38 (31)	53.5 ±10.6	
					Best value for C60	56.0 ±3.5	1

1) Values in parentheses are the coefficient of linear regression and the number of data pairs in the interval.

The regression coefficient of a least squares fit can be loosely interpreted as the percentage of variation in the dependent variable due to variation in the independent variable.

2) Value in parenthesis is the number of thermal conductivity values used to compute the mean.

B) Indicates a heat flow value from the basement (core) of the Atlantic Coastal Plain.

S) Indicates a heat flow value from the sediments (core) of the Atlantic Coastal Plain.

3) The standard deviation of a heat flow value calculated from two or more intervals is the standard deviation calculated from the mean of the heat flow values rather than the mean of the standard deviation of the heat flow values.

4) Category 1. Heat flow determinations of the highest quality. Temperature profiles are linear over intervals of heat flow calculations with no sign of hydrologic disturbances. Sufficient samples of rock or sediment core are available to characterize the effective conductivity of the measured section, and no variations that cannot be explained and corrected are apparent. If conductivity stratification is present, component heat flows for the various individual strata are in good agreement. Typical uncertainties for category 1 are less than ±10%.

Category 2. Heat flow values in which the uncertainty is greater than for category 1, but it probably is no greater than ± 20%. Included here are temperature profiles in which there are suggestions of local hydrologic disturbances. Also included are cases where the conductivity sample is unsatisfactory, owing to either too few samples or an unusually larger scatter in conductivity values.

Category 3. Heat flow values which are little more than rough estimates, and taken individually, do not not yield much information. Uncertainty in heat flow is greater than ±20%, but when combined with higher-quality data on a regional basis, these heat flow values can be quite useful. (Excerpted from Sass and others, 1971, J. Geophys. Res., v. 76, p. 6383.)

1 HFU (heat flow unit) = 10-6 cal/cm^2-sec = 41.84 X 10-^{3} W/m^2

1 HGU (heat generation unit) = 10-1^3 cal/cm^3-sec = 0.4184 X 10-6W/m^3

RESULTS FROM RADIOACTIVE DISEQUILIBRIUM STUDIES

L. D. Perry

A 1500 mm^2 intrinisic germanium detector (IGT) was received in the summer of 1980. In November 1980, an ND568 Gated Analog Router was purchased which allowed the IGT to be operated simultaneously with the two 4 X 5 inch NaI crystals currently used for heat production studies. A smaller 3 x 3 inch NaI crystal was also added to detect higher energy gamma rays from the same sample as the IGT. Previously, data from the two 4 x 5 NaI crystals were stored in 1024 channels of memory each. Now the data from the four detectors are stored in 512 channels of memory each. Changing from 1024 channels to 512 channels per detector required the preparation of a new library of standard spectra and some modification to the analysis software. The accuracy (± 1 ppm or 10%) of the U, Th and K measurements did not change.

The installation of the IGT was delayed when a leak in the liquid nitrogen Dewar was discovered. The Dewar was returned to Princeton Gamma Tech for repairs. When the IGT was returned in January 1981, work continued and various instrument calibrations were examined. Figure 1 shows the gain and threshold calibration currently in use for the approximate energy range 10 Kev to 300 Kev. Figure 2 shows the IGT resoultion as a function of energy.

Several FORTRAN subroutines which fit a non-linear equation to the spectrum data were obtained from the VPI&SU Research Reactor Lab and were used as the foundation for the development of an interactive graphics program for the display and analysis of the spectra from both the NaI crystals and the IGT. Figures 3 and 4 show standard U and Th spectra acquired with the IGT. Table 1 is from Szoghy and Kish (1978) and identifies the peaks in Figures 3 and 4. Figures 5 and 6 show the analysis of the 46.5 Kev peak from 210Pb in Figure 3. Figure 5 is the spectrum data and Figure 6 is the best-fitting equation determined by the analysis program.

Figure 7 is a spectrum from a sample which has been depleted in uranium and its daughters down to 226Ra. Disequilibrium to this degree is extreme and the differences between Figures 3 and 7 are obvious, but the difference would not be detected in the conventional analysis of spectra from the NaI crystals. Quantifying the degree of disequilibrium is difficult because it involves knowing the efficiency of the detector which is a function of both energy and sample geometry. Also, the absorption characteristics of the sample must be known. The techniques described by Szoghy and Kish (1978) and Ostrihansky (1976) are being employed at present. Determining an efficiency function is further complicated by a lack of agreement (see Table 1) on the number of gamma-rays emitted by any particular istope.

In conclusion, conventional determination of U and Ph with NaI spectroscopy will not yield correct values of heat production if disequilibirum is present. Disequilibrium can be detected using a germanium detector. The precision to be expected in determining the degree of disequilibrium has not yet been established.

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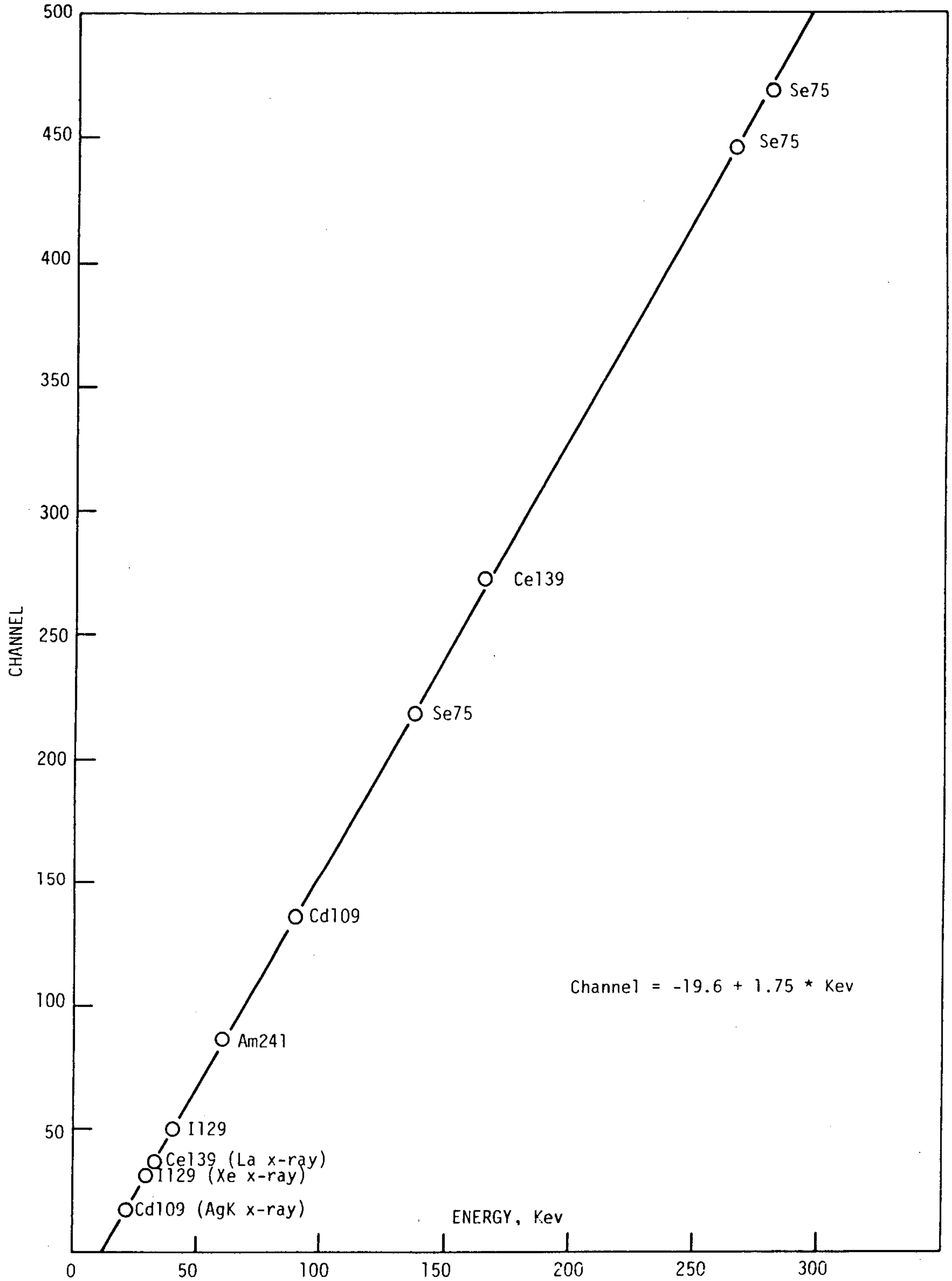

Figure 1. Points used to determine the energy calibration.

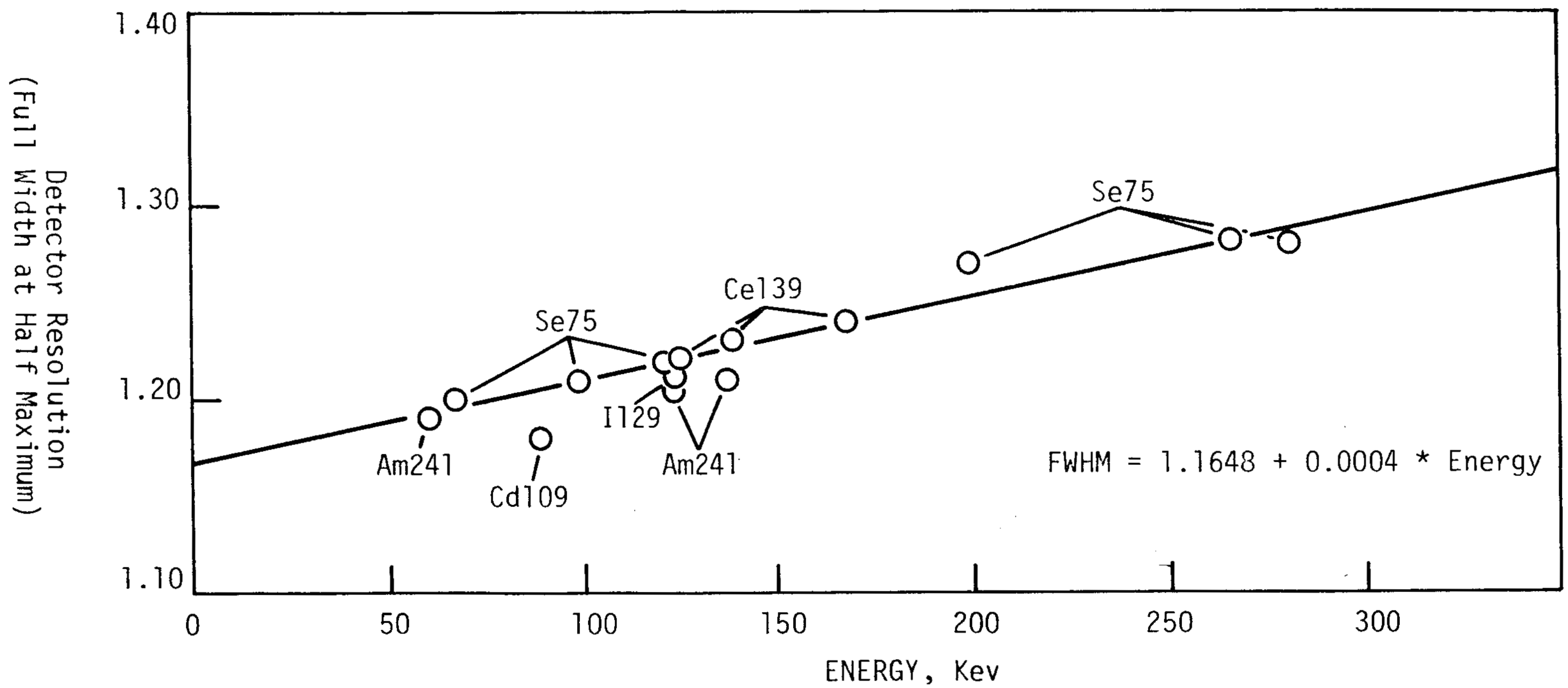

Figure 2. Points used to determine the detector resolution as a function of energy.

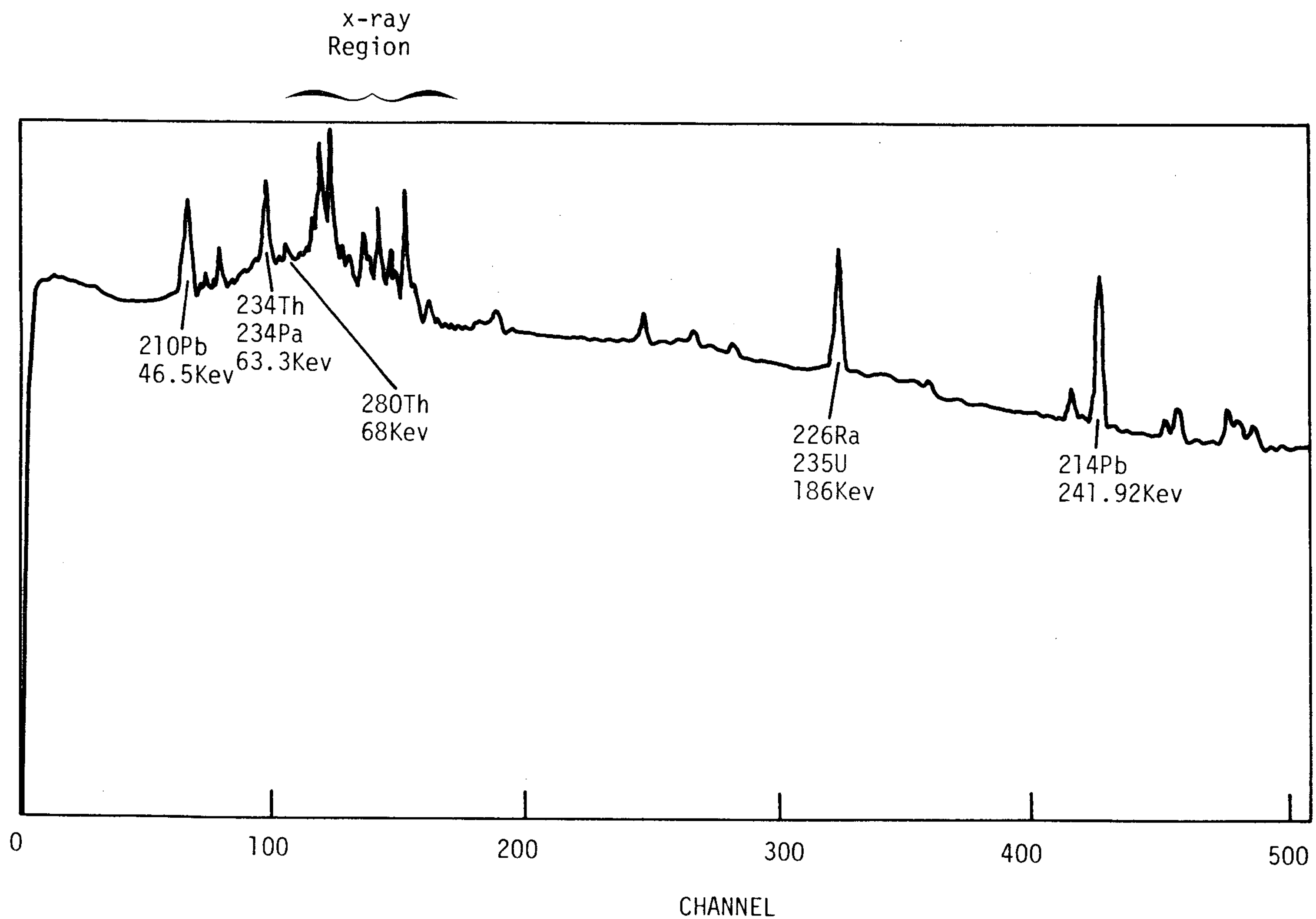

Figure 3. Spectrum from natural uranium (in equilibrium).

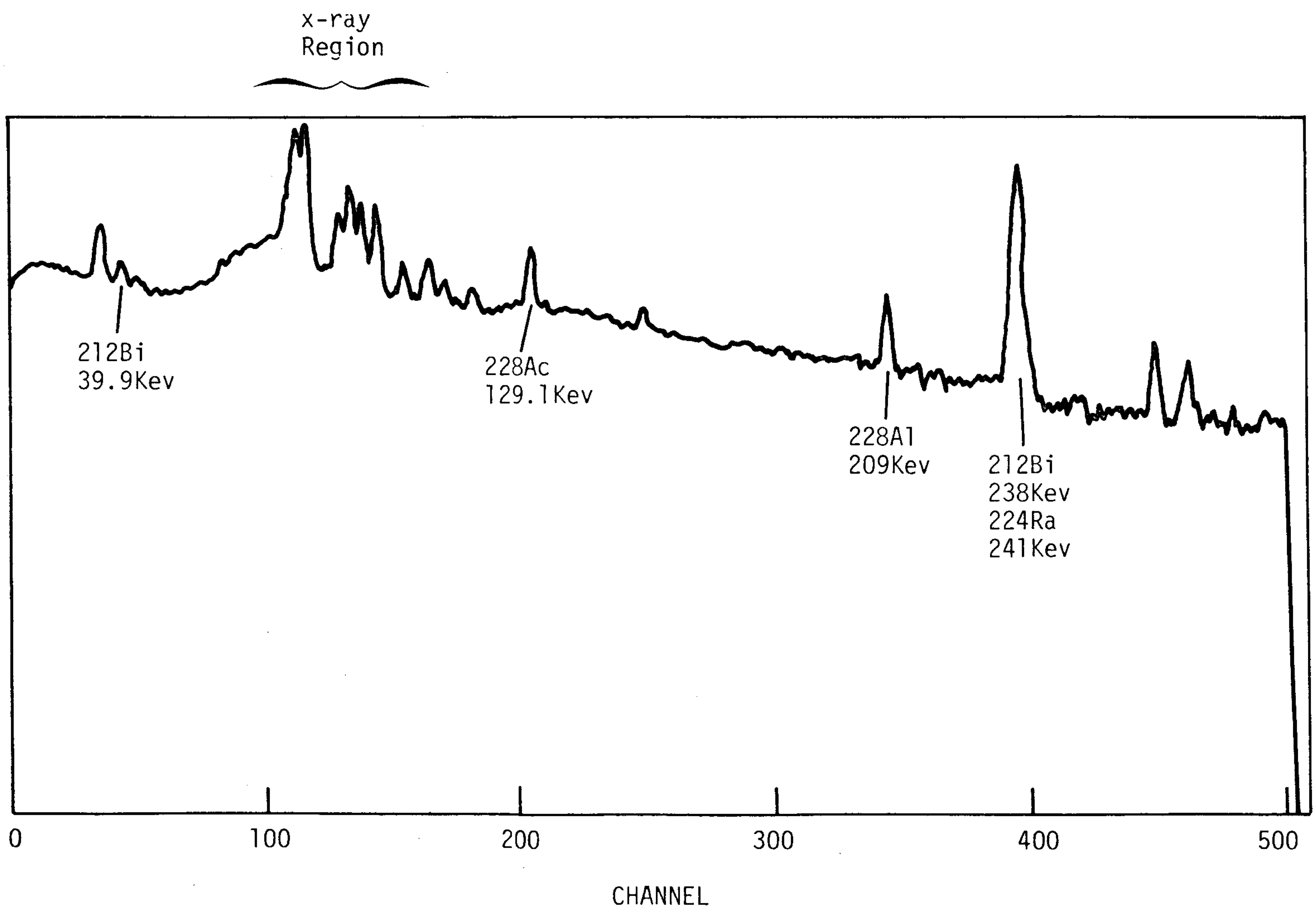

Figure 4. Spectrum from natural thorium (in equilibrium).

Peak number	Series	Group no.	X-ray emitter or mother isotope	E_{photon} (keV)	Relative detection efficiency[b]	Photons per 100 γ decay: Published	Photons per 100 γ decay: Present[c]
	Ac	1	^{235}U	26.64 (2)		12[d]	
	Ac	1	^{231}Th	25.6		9.4[d]	
	Th		^{212}Bi	39.859 (4)		1[e]	
1	U	6	^{210}Pb	46.50 (2)	80	4[d]	6 (1)
2	U	1	^{238}U	49.55 (6)	84	0.5[e]	0.3 (1)
	Ac	3	^{227}Th	50.2 (2)		8.7[d]	14 (2)
	Ac	3	^{223}Fr	50		40[d]	
3	U	2	^{234}U	53.1 (3)	88	0.7[d], 0.9[e], 0.1[e]	0.5 (1)
	U	5	^{214}Pb	53.23 (2)		2.2 (4)[f]	1.5 (5)
4	U	1	^{234}Th	63.282 (2)	96	0.6[d] 5.5[e,g]	5.3 (3)
	U	1	^{234}Pa	63.4		2.8[d]	
	Th		?	63.81 (7)		0.4[e]	
5	U	3	^{230}Th	68	95	0.59[d]	0.4 (1)
6	U	5	PbK$_{\alpha 2}$	72.8054 (13)	94.5	—	*h*
	Th		PbK$_{\alpha 2}$			—	0.9 (4)
7	U	5	BiK$_{\alpha 2}$	74.81	93.5	—	9.3 (7)
	U	5	PbK$_{\alpha 1}$	74.9706 (13)		—	*h*
	Th		BiK$_{\alpha 2}$, PbK$_{\alpha 1}$	74.9		—	7.6 (6)
8	U	5	BiK$_{\alpha 1}$, PoK$_{\alpha 2}$	77	92	15.6	15.3 (9)
	Th		BiK$_{\alpha 1}$	77.10		—	12.0 (8)
9	U	5	PoK$_{\alpha 1}$	79.30	90	—	1.3 (4)
10	Ac	3	RnK$_{\alpha 2}$	81.08	89	—	25 (1)
11	U	5	PbK$_{\beta 1}$	84.92	86	—	*h*
	Ac	1	^{231}Th	84.24 (1)		7.9[d]	5.1 (5)
	Ac	3	RnK$_{\alpha 1}$	83.8			39 (4)
	Th		^{228}Th	84.37 (5)		0.45[d], 0.9[e]	1.2 (3)
	Th		PbK$_{\beta 1}$	84.92			0.6 (3)
12	U	5	BiK$_{\beta 1}$	87.33	85	—	4.8 (4)
	U	5	PbK$_{\beta 2}$	87.34			*h*
	Th		BiK$_{\beta 1}$, PbK$_{\beta 2}$	87.3		0.7[e]	4.3 (5)
13	U	5	BiK$_{\beta 2}$, PoK$_{\beta 1}$	89.8	83	—	2.1 (3)
	Th		ThK$_{\alpha 2}$	89.94		—	*i*
	Th		BiK$_{\beta 2}$	89.83		—	3.6 (4)
14	U	1	^{234}Th	92.367 (5)	81	3.38[j] 6.7[e,j,g]	6.7 (6)
			^{234}Th	92.791 (5)		3.34[j]	
	U	5	PoK$_{\beta 2}$	92.39		—	0.2 (1)
	Th		ThK$_{\alpha 1}$	93.33		—	*i*
15	U	1	UK$_{\alpha 2}$	94.666 (4)	77	—	*i*
	Ac	3	RnK$_{\beta 1}$	94.88		—	10 (1)
16	U	1	UK$_{\alpha 1}$	98.441 (3)	74.5	—	*i*
	Ac	3	RnK$_{\beta 2}$	97.62		—	4.4 (5)
	Th		^{228}Ac	99.45 (8)		1.2[d], 0.6[e]	0.9 (4)
17	Th		ThK$_{\beta 1}$	105.59	72	—	*i*
18	U	1	UK$_{\beta 1}$	111.29		—	*i*
	Ac	1	^{235}U	109.12 (1)		1.51[d]	1.5 (1)
	Th		ThK$_{\beta 2}$	108.67	69	—	*i*
19	U	1	UK$_{\beta 2}$	114.55		—	*i*
	Th		^{212}Pb	115.19 (3)	65	0.57 (8)[f]	0.57 (8)
20	Th		^{228}Ac	129.1 (3)	57	2.5[d]	1.6 (2)
21	Ac	1	^{235}U	143.776 (10)	48	9.72 (49)[k,l]	11.7 (15)
	Ac	3	^{223}Ra	144.2		4.1[m]	6.4 (5)
22	Ac	3	^{223}Ra	154.1	42	5.5[m]	9.3 (9)
	Th		^{228}Ac	154.2		0.8 (1)[f]	0.8 (1)
23	Ac	1	^{235}U	163.363 (10)	36	4.59 (23)[k,l]	5.0 (5)
24	U	4	^{226}Ra	186.10 (10)	29	4.0[d]	[illegible]
	Ac	1	^{235}U	185.718 (11)		54 (3)[k,l]	4[illegible] (3)
	Th		^{228}Ac	181 (2)		1.6 (2)[f]	1.6 (2)
25	Ac	1	^{235}U	205.311 (12)	22	5.00 (25)[k,l]	5.0 (5)
26	Th		^{228}Ac	209.4 (3)	21	3.9[d]	2.7 (3)
27	Ac	3	^{227}Th	236.0	15	12.1[m]	12.5 (9)

Table 1. A table from Szoghy and Kish (1978) listing pertinent data for spectra in Figures 3 and 4.

Table 1 cont.

Peak number	Series	Group no.	X-ray emitter or mother isotope	E_{photon} (keV)	Relative detection efficiency[b]	Photons per 100 α decay: Published	Photons per 100 α decay: Present[c]
28	Th		^{212}Pb	238.62 (1)	14.5	43.1[d]	**38 (1)**
29	U	5	^{214}Pb	241.92 (3)	14	7.6[d]	**7.8 (6)**
	Th		^{224}Ra	240.98 (3)		3.9 (11)[n]	**3.6 (2)**
	Th		^{228}Ac	270.3 (3)	10	3.2[d]	
	Ac	3	^{223}Ra	269.4		14[d]	
	Ac	3	^{219}Rn	271.0		11[d]	

[a]See also Table 1 and its footnote.
[b]5 mm thick sample close to the beryllium window of a 5 mm thick, 130 mm² surface Ge (Li) detector.
[c]Based on the previous column and also on the present photon intensities obtained from numerous old rock samples believed to be in radioequilibrium.
[d]Bowman and MacMurdo (1974).
[e]Taylor (1973). Arbitrary normalization of photon intensities.
[f]Martin and Blichert-Toft (1970).
[g]Ellis (1970*b*).
[h]*Originating from lead content.*
[i]Nonlinear dependence on mother element concentration.
[j]Sampson (1973). Arbitrary normalization of photon intensities.
[k]Cline (1971).
[l]Cohen (1971*b*).
[m]Wapstra and Gove (1966).
[n]Ellis (1976).

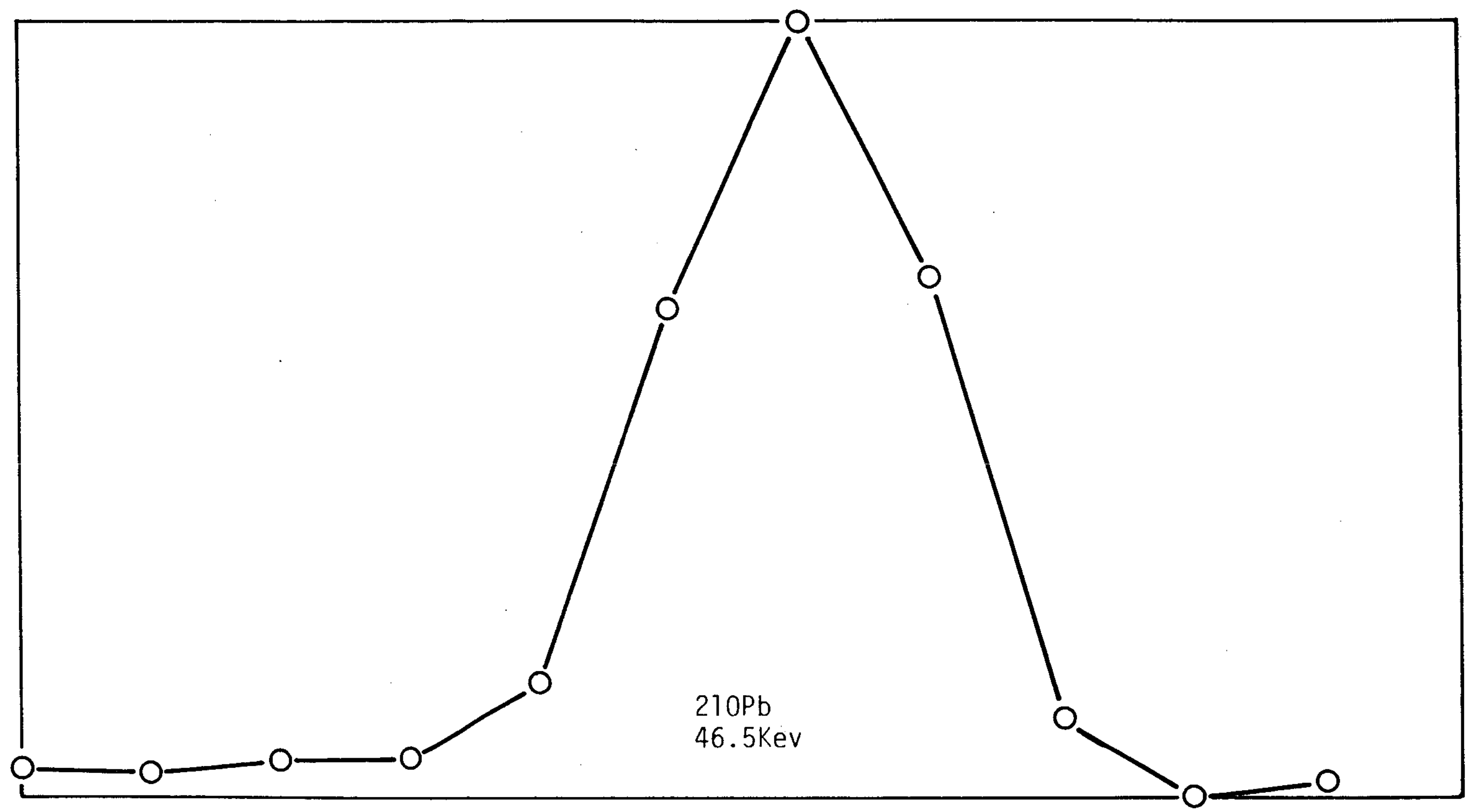

Figure 5. Blowup of the 210Pb peak at 46.5Kev in Figure 3.

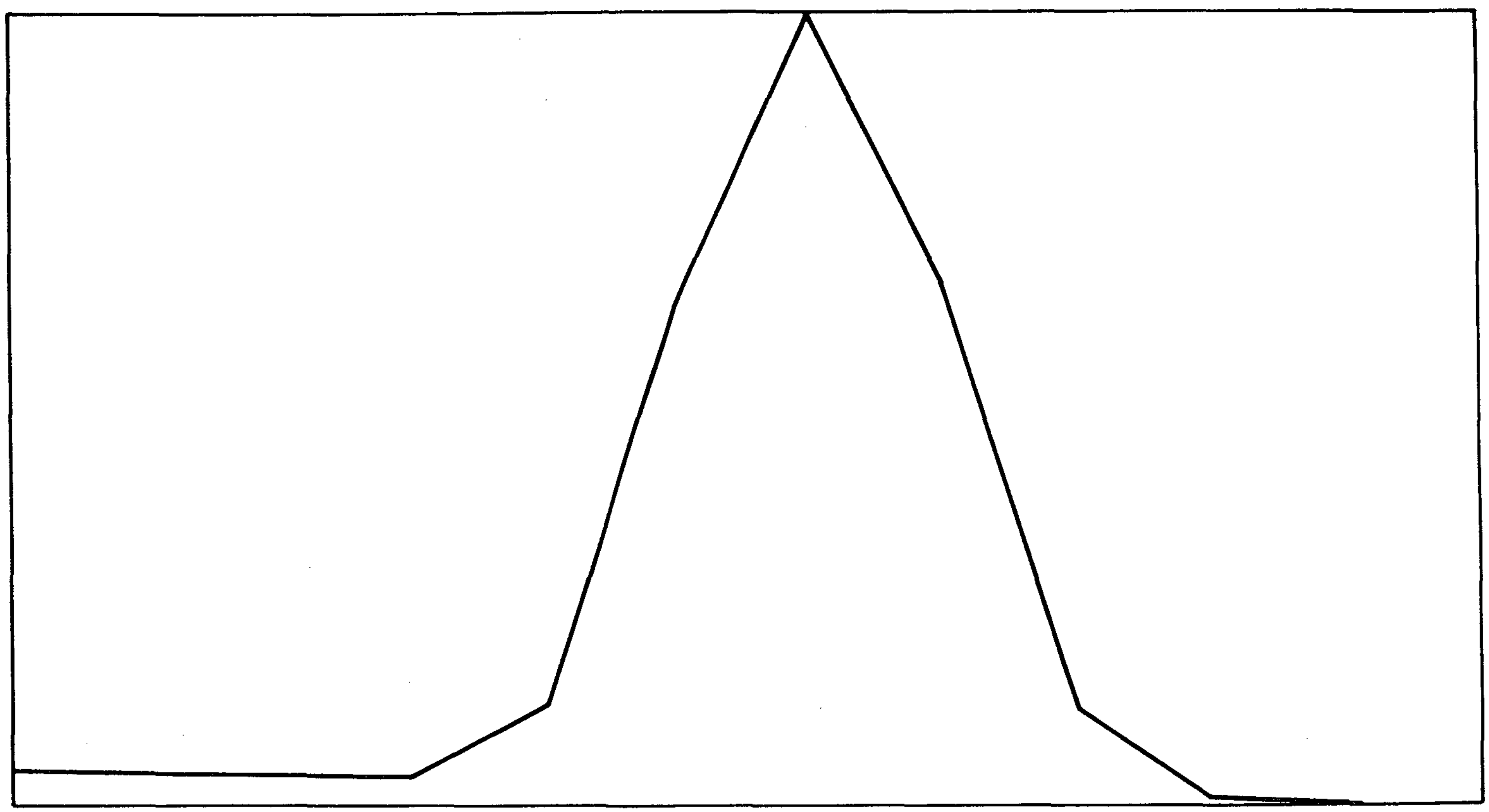

Figure 6. The best fitting function from the analysis program plotted using the same channel locations as in Figure 5. (The function is actually smoothly continuous.

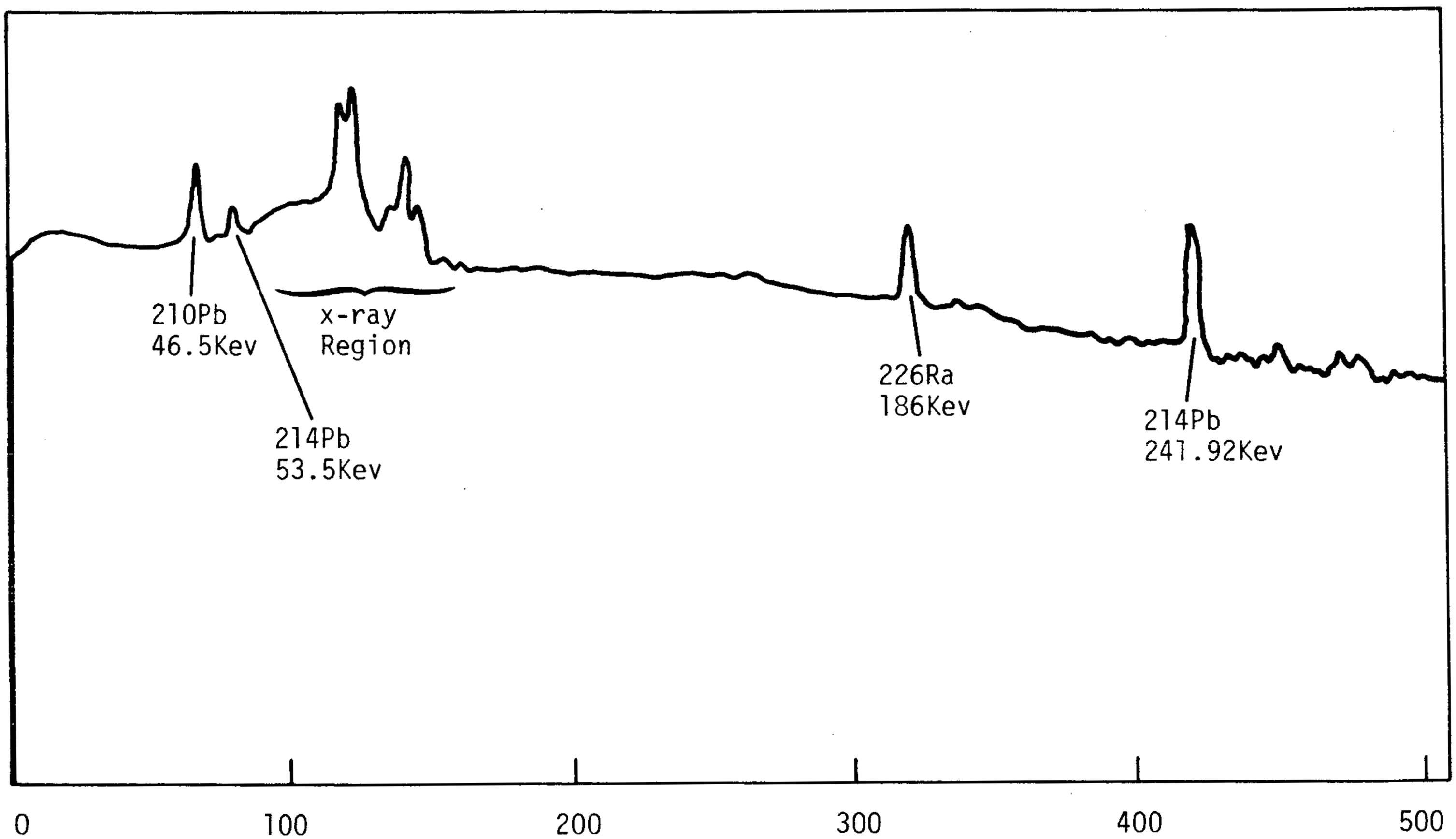

Figure 7. Spectrum from a sample in disequilibrium containing Radium in equilibrium with its daughters but from which the Uranium has been removed. Note the absence of 234Th at 63Kev and 230Th at 68Kev (and others).

GRAVITY MAPS: EASTERN UNITED STATES

J. K. Costain and L. D. Perry

Heat flow sites chosen by VPI&SU for the purpose of targeting heat-producing granites in the basement rocks beneath the sediments of the Atlantic Coastal Plain were located primarily on the basis of gravity data. The relatively young (254-420 Ma), unmetamorphosed, heat-producing granitoids generally cause negative gravity anomalies because of their relatively low density (2.65-2.67 gm/cm^3). Excellent examples of anomalies are the Portsmouth, Dort, and Chesapeake Bay (Smiths Point, Va.) gravity anomalies (Fig. 1). Heat flow values have been obtained from basement granite at Portsmouth and Dort (see chapter on heatflow, this report). The basement rocks beneath the Chesapeake Bay gravity anomaly have not yet been sampled by the drill; however, the highest geothermal gradient (50^0C/km) discovered to date in the eastern United States is at Smith Point, Va. (Fig. 1). The basement rock is most probably granite, perhaps similar to the Petersburg Granite exposed near Richmond, Va.

Negative gravity anomalies over easternmost North Carolina are probably also associated with heat-producing granite, but the geochronology and petrography of the basement rocks are poorly known. The heat flow is high (see Stumpy Point heat flow site, chapter on heatflow, this report).

Gravity data constitute the most important type of geophysical data for exploration of areas of high heat flow beneath the Atlantic Coastal Plain. A negative gravity anomaly will not be well defined, however, over a heat-producing granite emplaced in host rock that is also granitic in chemical composition. Careful three-dimensional gravity modeling and attention to the geologic framework will be required before drill sites for heat flow determinations can be selected with confidence.

Dr. Allan Cogbill developed software for the interpretation of gravity data, and interpreted gravity data from the southeastern United States (Cogbill, 1978, p. C-65 to C-123).

Dunbar (1979) developed a general modeling technique based on the gravity and magnetic effects of polygonal prisms, and applied the theory to modeling of the exposed

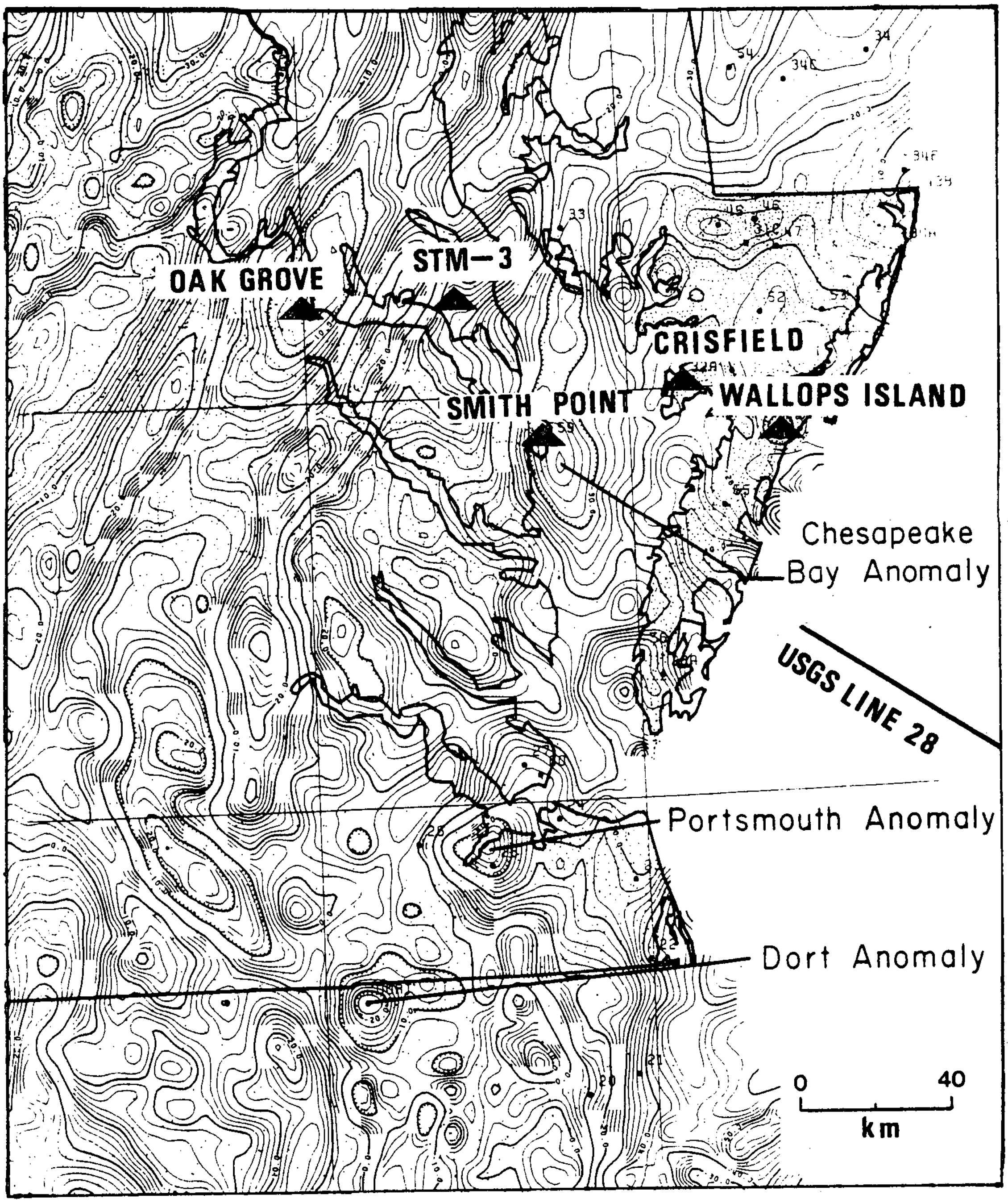

Figure 1. Bouguer gravity map showing Portsmouth, Dort, and Chesapeake Bay in Virginia and North Carolina.

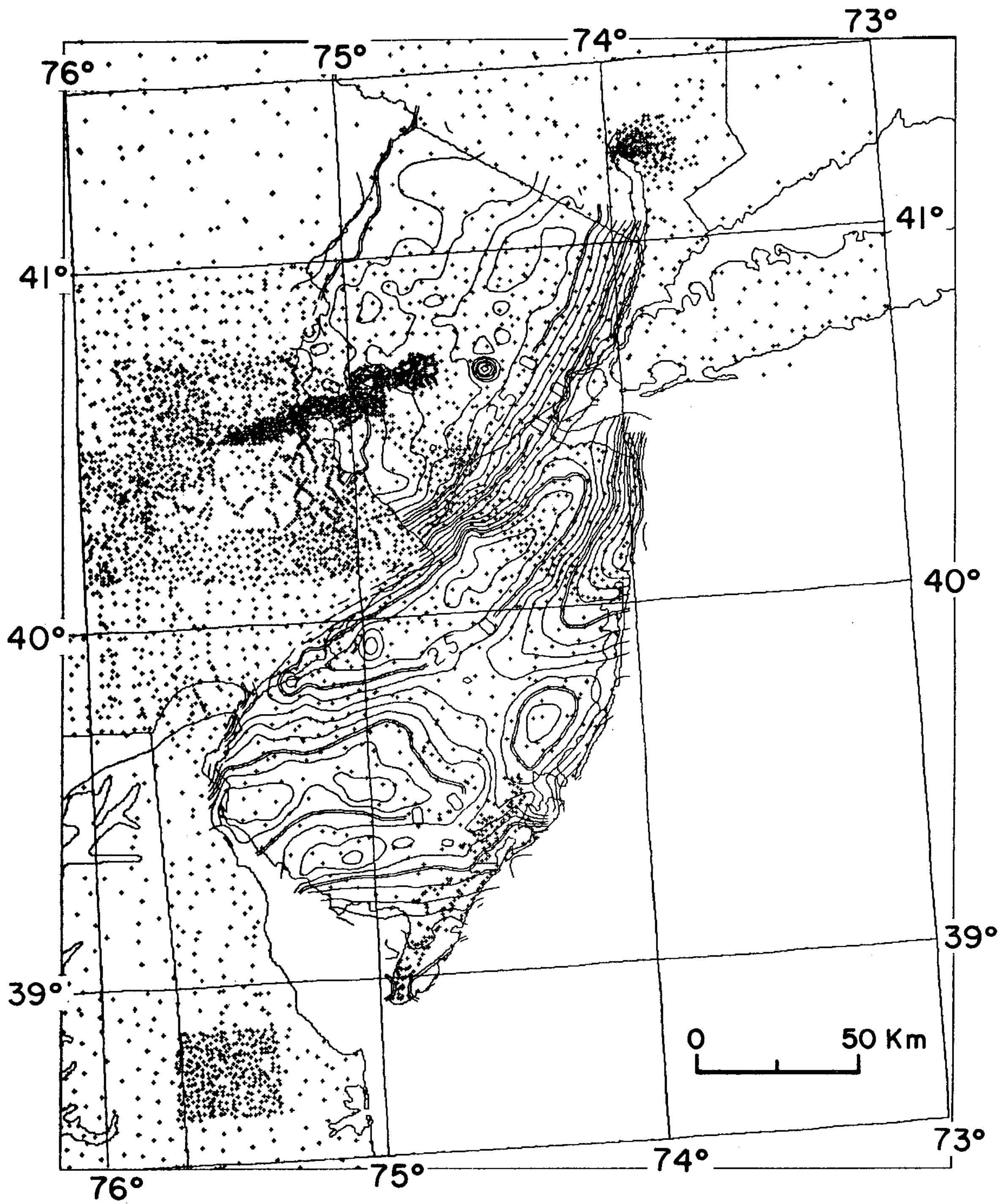

Figure 2. Distribution of gravity stations and Bouguer anomaly map of New Jersey. Contour interval 5 milligals.

Rollesville batholith and Castalia pluton in North Carolina. His results are presented in detail elsewhere in this report. Dunbar's approach and computer programs (Dunbar, 1979) are appropriate for detailed gravity and heat flow modeling of the basement beneath the Atlantic Coastal Plain.

Figures 2-6 are contour maps of the gravity field from New Jersey to Georgia. Contouring was done at VPI&SU using the IBM computer program GPCP. The map projection is Lambert conformal. The contour interval is 5 milligals.

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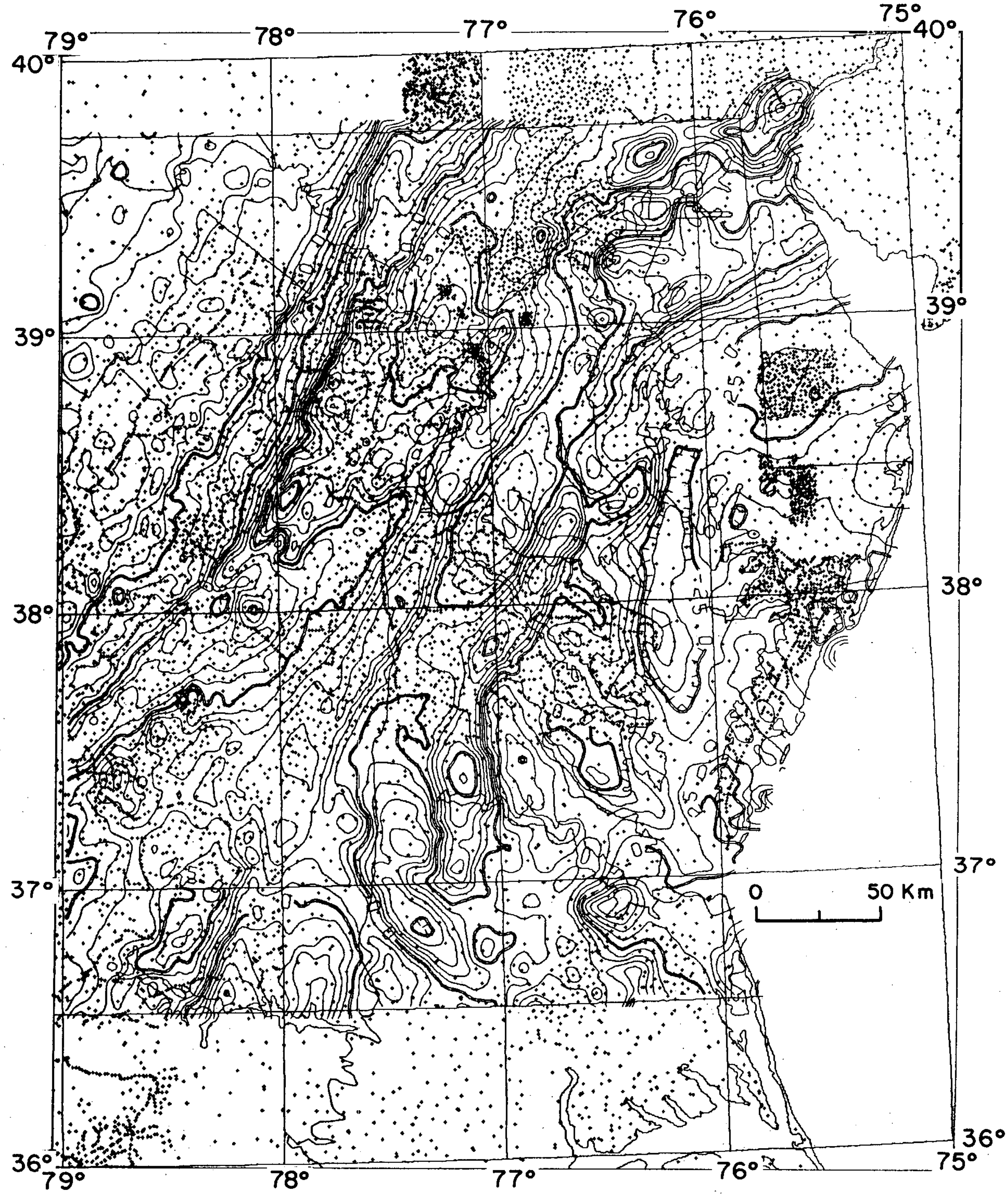

Figure 3. Distribution of gravity stations and Bouguer anomaly map of Delaware, Maryland, and Virginia. Contour interval 5 milligals.

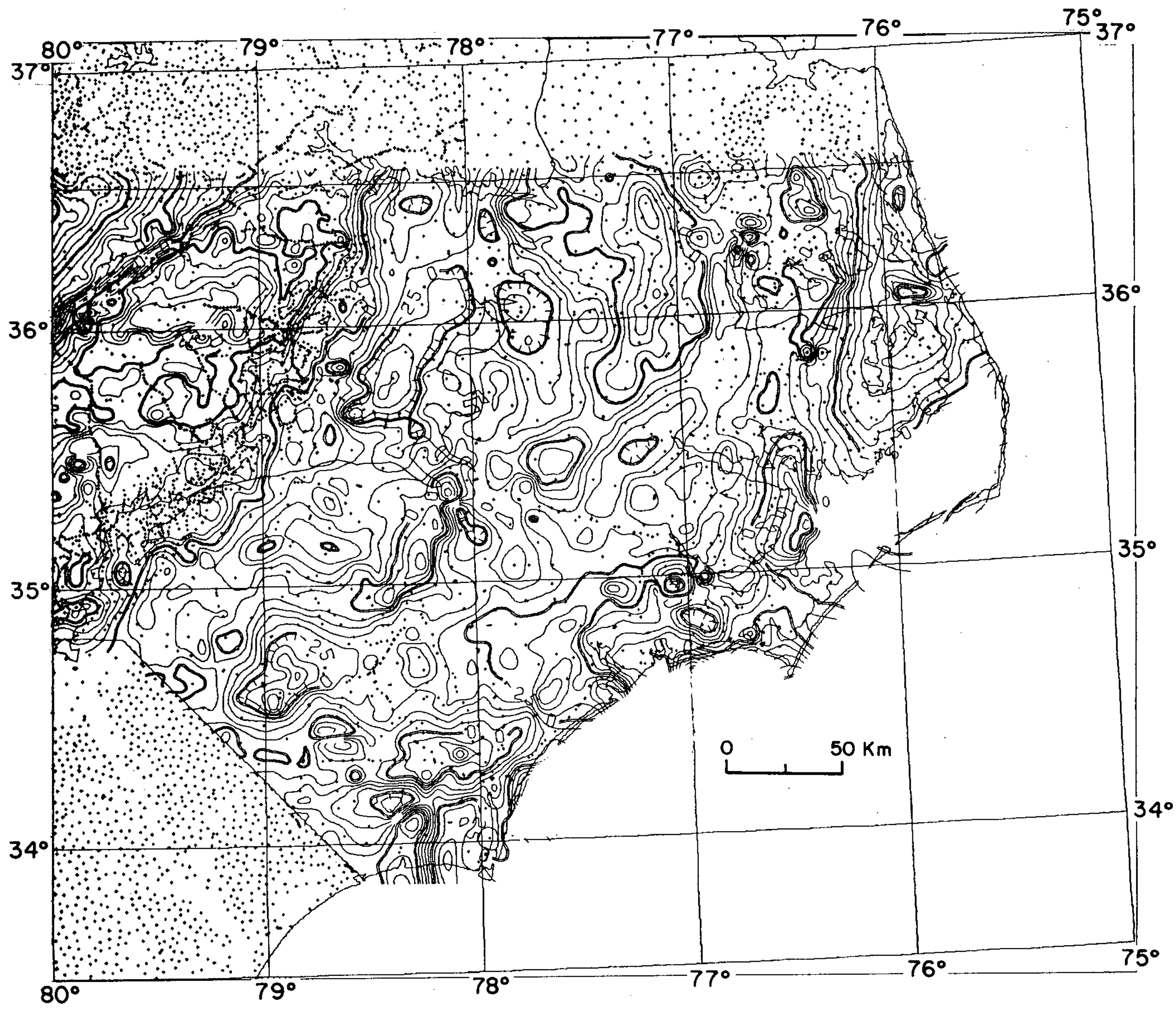

Figure 4. Distribution of gravity stations and Bouguer anomaly map of North Carolina. Contour interval 5 milligals.

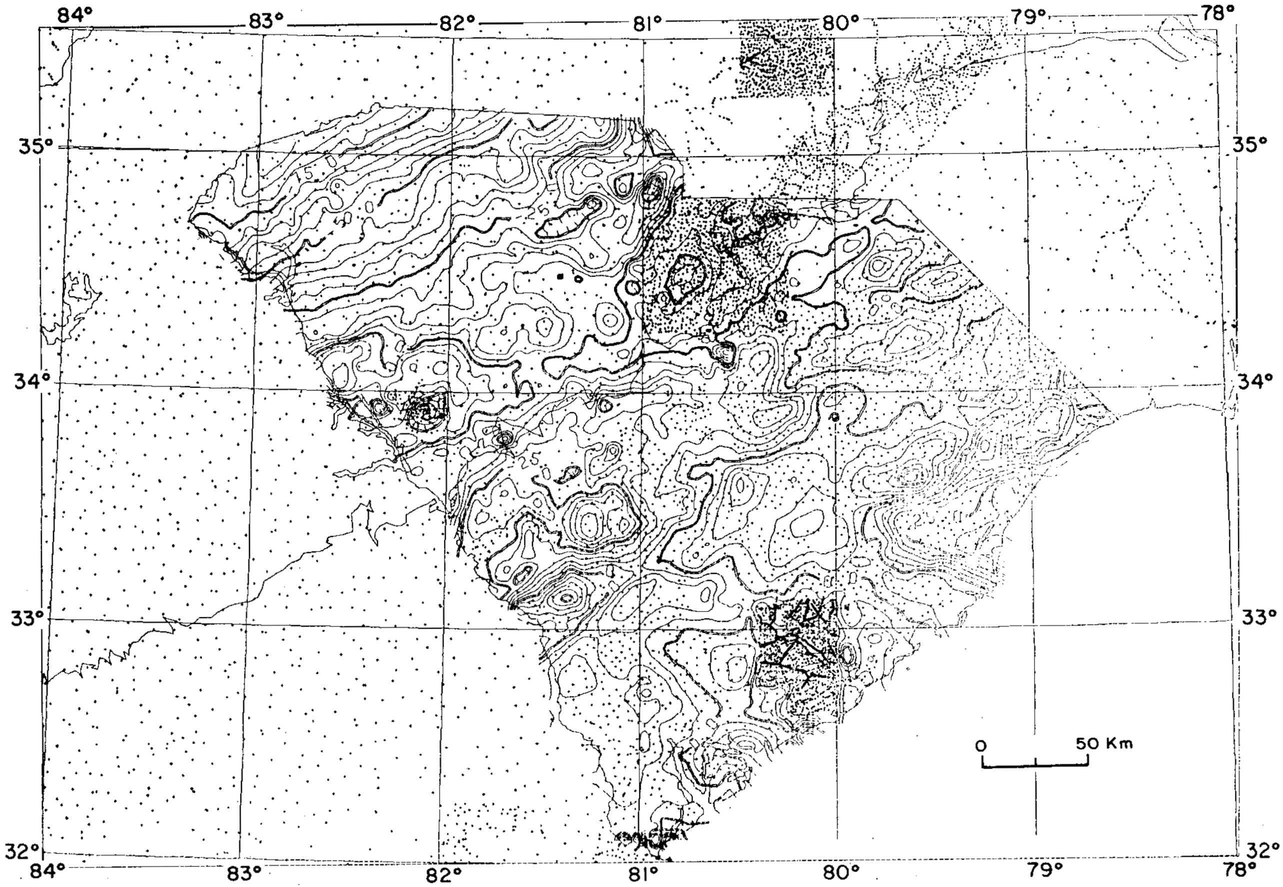

Figure 5. Distribution of gravity stations and Bouguer anomaly map of South Carolina. Contour interval 5 milligals.

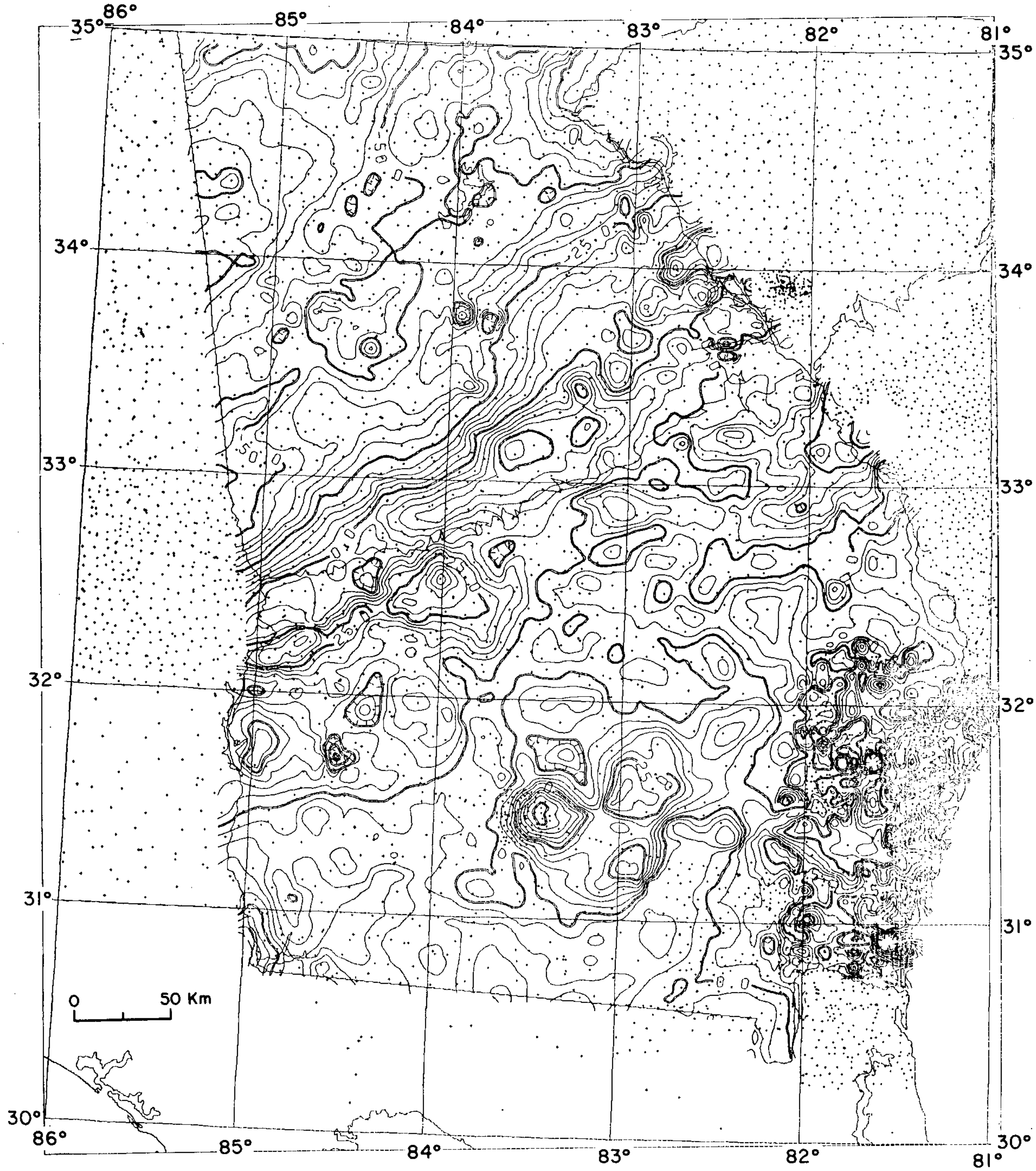

Figure 6. Distribution of gravity stations and Bouguer anomaly map of Georgia. Contour interval 5 milligals.

TEMPERATURE AND HEAT FLOW MODELING OF THREE-DIMENSIONAL BODIES IN A TWO-LAYERED HALF-SPACE

John A. Dunbar, Jr.

Introduction

Steady-state heat flow and temperature anomalies in the earth's crust can be caused by lateral variations in heat production and thermal conductivity. In many respects the problem of interpreting thermal anomalies is identical to the problem of interpreting gravity and magnetic anomalies. Heat production and thermal conductivity vary with rock type, as do density and magnetic susceptibility. As a result, geologic features such as folds, faults, and igneous intrusions can produce thermal anomalies in the same way that they produce gravity and magnetic anomalies. There are also strong mathematical similarities. Temperature fields satisfy the same governing equations as do gravity and magnetic problems. Solutions to gravity and magnetic problems then, differ from solutions to analogous temperature field problems only in the nature of the boundry conditions imposed and by constant coefficients. In particular, the heat flow anomaly caused by a body of constrasting heat production is analogous to the gravitional attraction of the same body (Simmons, 1967). Likewise, the heat flow anomaly caused by a body of contrasting thermal conductivity in a uniform heat flow field is mathematically equivalent to the magnetic anomaly caused by a body of contrasting susceptibility in a uniform inducing field (Carslaw and Jaegar, 1959, p. 425). Temperature anomalies caused by bodies of contrasting heat production and conductivity are respectively analogous to gravimetric potential and gravitational attraction anomalies.

Because of these similarities it is possible to model heat flow and temperature anomalies using the same techniques used to model gravity and magnetic anomalies. In many cases the same computer programs can be used with only minor modifications. Simmon (1965), for example, suggested a method based on the gravitational attraction of a polygonal lamina (Talwani and Ewing, 1960) for modeling heat flow anomalies due to heat production contrasts in a half space. Thermal modeling techniques of this type are faster computationally and less cumbersome to implement than the numerical techniques such as the method of finite differences and the method of finite elements. These advantages become particularly important in situations which require repeated modeling such as in solving inverse problems by trial-and-error methods.

The model proposed by Simmons does not account for the effects of contrasts in thermal conductivity between the anomalous body and the half space. It also does not account for the effects of a layer of contrasting conductivity overlying the half space. This more general problem is of current interest in the exploration for low-temperature geothermal resources. The objectives of this exploration are temperature anomalies in low-conductivity sediments overlying radioactive granites in the crystalline basement (Costain et al., 1980). In principle the sedimentary layer would act as an insulator, causing higher temperatures to occur closer to the surface. The problem is also important in the interpretation of heat flow determinations made in sea floor sediments. The latter case was considered by Lee and Henyey (1974) who used the method of finite elements to correct marine heat flow values. An analytical treatment of temperature and heat flow anomalies in a two-layer half space does not exist in the literature.

A more general modeling technique than that proposed by Simmons (1965), based on the gravity and magnetic effects of polygonal prisms (Plouff, 1976), is developed in Dunbar (1979). The technique is suitable for modeling temperature and heat flow anomalies associated with three-dimensional bodies of contrasting heat production and conductivity in a two-layered half space, and is summarized below.

Assumptions and Approximations

This study is concerned with the steady-state heat conduction problem for anomalous bodies in a half space overlain by a layer of contrasting conductivity (Fig. 1). In general, both the heat production and thermal conductivity of the body differ from that of the surrounding medium. Heat enters the system either as uniform vertical heat flow (q^*) from the base of the model region or is generated within the anomalous body and surrounding medium. Heat production in the surrounding medium extends to a finite depth. The surface of the two-layered half space is maintained at a constant temperature. All other factors which influence the terrestrial temperature field are ignored.

In lieu of an exact solution to this problem the temperature and heat flow effects of the body's heat production

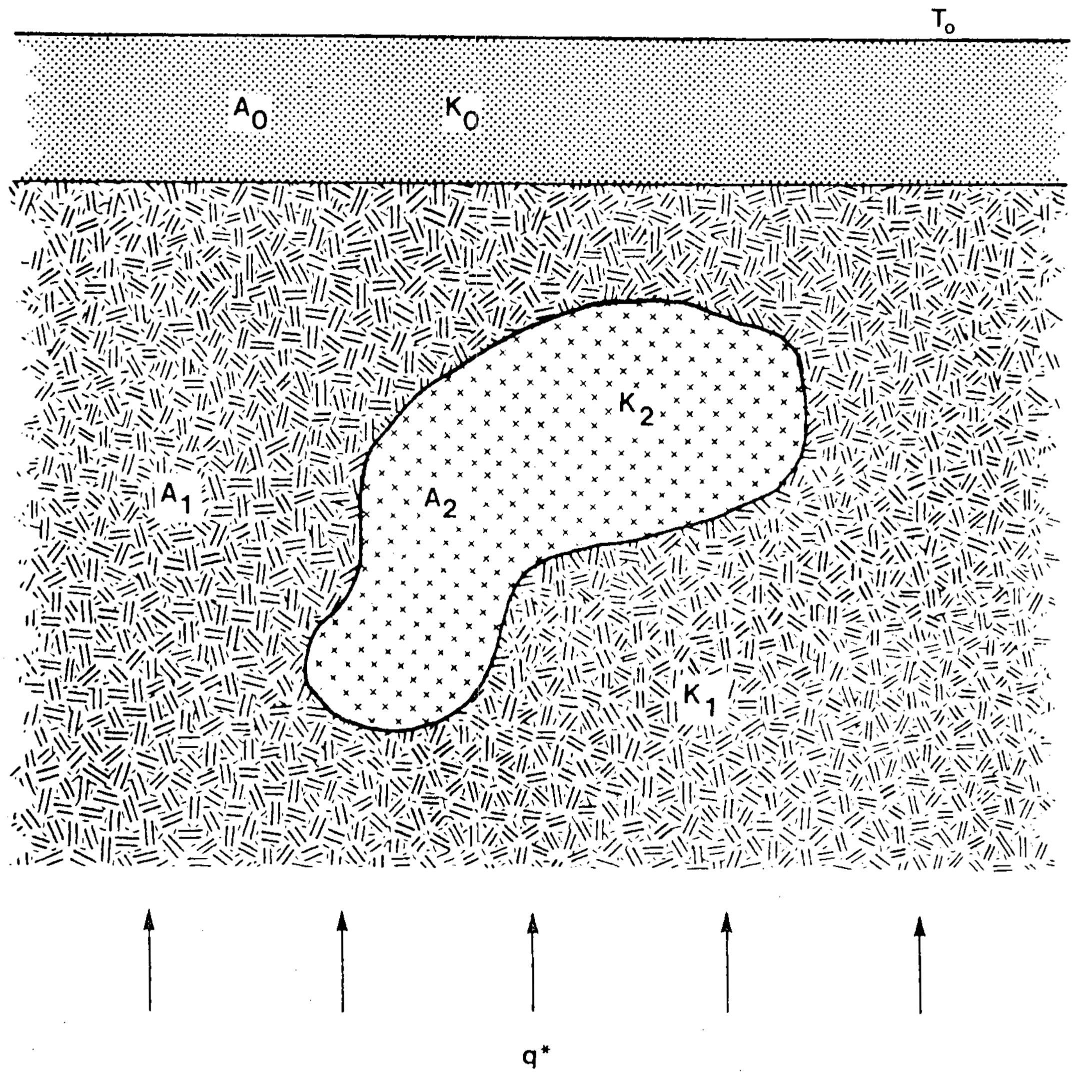

Figure 1. Schematic diagram of an anomalous body of arbitrary shape in a two-layered half space. The vertical heat flow q* enters the system at the base of the model region. The surface is held constant at temperature T_0.

and conductivity contrasts are calculated seperately and then added. The temperature field is thus approximated by the superposition of three independently calculated temperature fields: (1) the anomalous field due to a body of constrasting heat production in a two-layer half space (TA); (2) the anomalous field due to the disturbance in q* by a body of contrasting conductivity in a two-layered half space (TK); and (3) a regional field which varies only in the vertical direction (TR).

The effects of arbitrarily shaped three-dimensional bodies are approximated by summing the effects of horizontal polygonal prisms as in Plouff's (1976) gravity modeling method. Expressions for the temperature and heat-flow anomalies due to anomalous polygonal prisms in a uniform space are developed and summarized in Dunbar (1979). Expressions for the temperature and heat-flow anomalies in a uniform half space and a two-layered half space are then found by the method of images, a standard method in heat conduction (Carslaw and Jaeger, 1957, p. 273) and in electrostatics (Kellogg, 1957, p. 207). The mathematical derivations are not repeated for this report.

Thermal Conductivity Contrasts

The distortion of a uniform heat flow field by a body of contrasting conductivity is mathematically the same as the distortion of a magnetic field by a body of contrasting magnetic susceptibility. The latter problem as been dealt with by Talwani (1965) and Plouff (1976). In these studies the contribution to the magnetic field by an infinitesimal volume element in the anomalous body is assumed to be the same as a similar volume element alone in free space. The total effect of the anomalous body is then found by integrating the effects of all such volume elements which make up the body. This is not an exact solution because the magnetic field induced in a given volume element will act as an additional inducing field in neighboring volume elements. This interaction between neighboring volume elements is ignored.

The same method can be applied to the analogous heat conduction problem. The temperature effect of a single volume element can be found by considering the temperature anomaly due to a sphere of contrasting conductivity, centered in a uniform heat flow field q_0 (Carslaw and Jaeger, 1959, p. 426).

The temperature anomaly due to a spherical volume element V at an arbitrary position in a region V_0 is given by Dunbar (1979). The approximate temperature anomaly associated with a region V_0 of contrasting conductivity is then given by letting V become infinitesimal and intergrating the effects of all such volume elements in V_0.

This will be an approximation because, as in the analogous magnetic problem, the interaction between volume elements is ignored. Numerical results using this method are compared with those from exact solutions by Dunbar (1979).

For a realistic representation of the terrestrial temperature and heat flow fields, the thermal effects of the earth's surface must be accounted for. This can be done by the method of images if the earth is represented by a half-space with a constant surface temperature. Simply stated, the method of images involves constructing a system of sources and sinks in a medium with uniform material properties in such a way as to duplicate the temperature field in a region with discontinuous material properties. The combined effect of all the sources and sinks is required to satisfy the governing equation throughout the region of interest and to behave in a specified manner at the boundaries of the region.

The geometry of the source-sink system for a point source in a half-space with a uniform conductivity (one-layered half-space) is shown in Figure 2. The uniform half space model of the earth is not applicable where a layer of contrasting conductivity overlies the source region. Such a situation occurs in the Atlantic Coastal Plain, where the conductivity of the basement complex can be twice that of the overlying sediments (Costain et al., 1979). The effects of the sedimentary layer can be accounted for by placing a layer of thickness h and conductivity K_0 over the uniform half space with conductivity K_1.

Up to this point only the problem of heat production in a two-layered half space has been considered. A similar development for the problem of contrasting conductivities is given by Dunbar (1979).

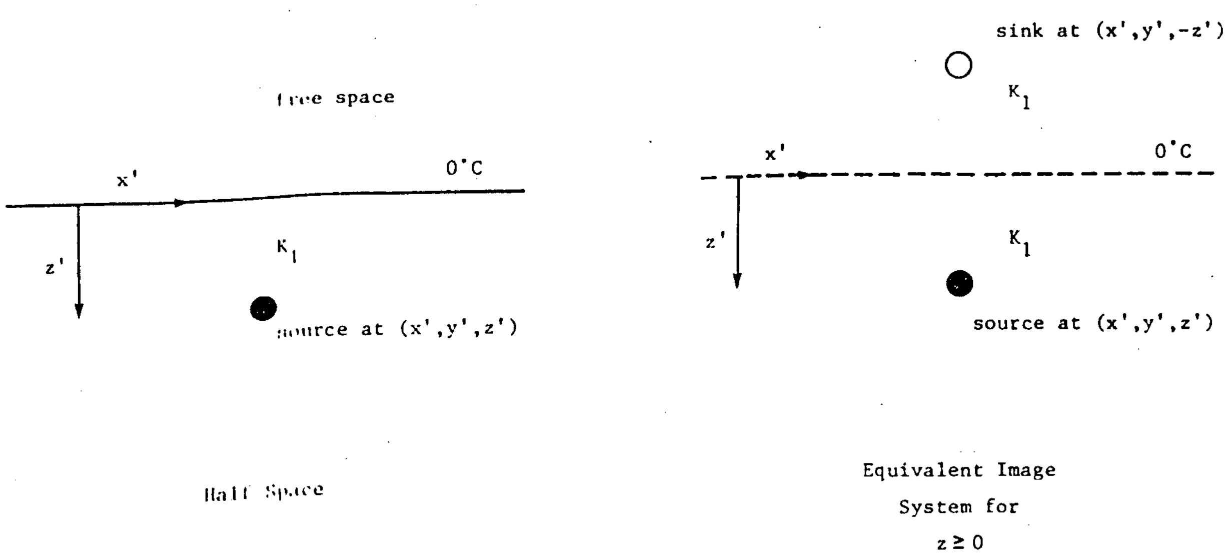

Figure 2. Point-source image system for temperature in a uniform half space.

The Regional Temperature Field

The portion of the temperature field which varies only in the vertical direction will be termed the regional field, TR(z). Because it does not vary laterally it is governed by Poisson's equation for one-dimensional heat conduction.

For the general case of a body of contrasting heat production and thermal conductivity the approximate total temperature field is given by the algebraic sum of T(x,y,z) = TA (x,y,z) + TK (x,y,z) + TR(z).

Dunbar (1979, Figs. 17 and 18) shows that the applicability of one-dimensional theory to three-dimensional sources is governed by the aspect ratio of the source and by the heat production and conductivity of both the source and the surrounding medium. For irregularly shaped bodies polygonal prism models can be used to generate sets of curves analogous to those in Figures 17 and 18 of Dunbar (1979).

Interpretation of Heat-flow Anomalies

Heat flow anomalies can be modeled by using a trial-and-error method similar to that used in gravity and magnetic modeling. The heat production and conductivity of the body and surrounding medium and the regional heat flow field are normally estimated first. The geometry of the body is then varied until a satisfactory agreement between the observed field and theoretical field is reached. As in gravity and magnetic modeling the resulting solution is not unique, but is a member of a family of possible solutions.

To illustrate the modeling process a family of heat flow models was prepared for the Rolesville batholith and adjacent Castalia pluton, in Franklin and Nash Counties, North Carolina. These granitoids were the focus of a major effort by the petrology group in the Orogenic Studies Laboratory during the DOE geothermal program at VPI&SU (Farrar, 1980). Both bodies are coarse-grained granitic intrusions emplaced in metamorphic rocks of the Raleigh belt (Fig. 3). The Raleigh belt has been interpreted as a south-plunging antiform and has a trend of increasing metamorphic grade from south to north (Farrar, 1980). Both characteristics suggest differential uplift and erosion between the southern and northern portions of the belt. The locations of five heat flow stations in the area are shown in Figure 3. The heat production, thermal conductivity, and heat flow at these sites are given in Table 1 (Costain et al., 1979). Because there is no sedimentary layer in the region the problem was treated as a one-layered half space problem. At the time of this writing data for a two-layered case are not available.

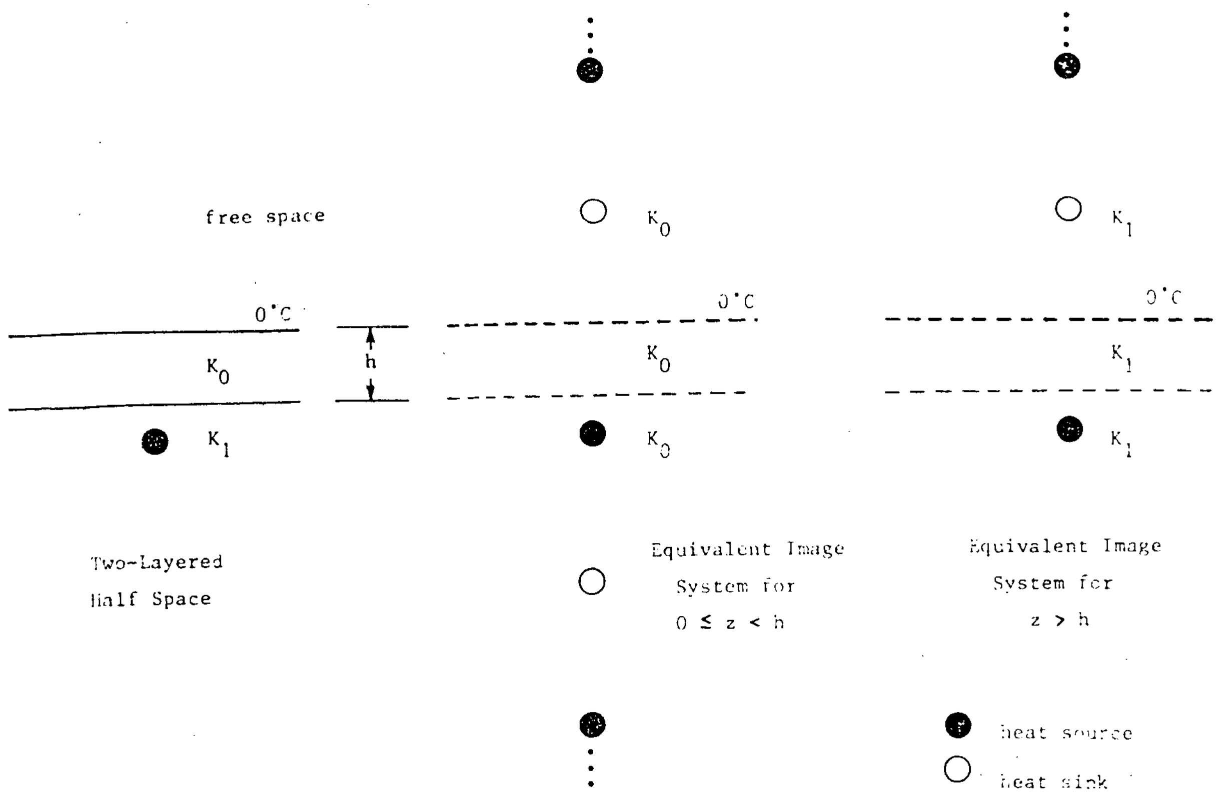

Figure 3. Geologic map of the Rolesville batholith and vicinity (simplified from Farrar, 1980). Dots indicate locations of heat-flow stations.

Table 1. Preliminary data from heat flow sites in the vicinity of the Rolesville batholith and Castalia pluton (from Costain _et al_., 1979).

Hole	Heat Production	Conductivity	Heat Flow
CS 1	5.6 HPU	7.64 TCU	1.44 HFU
RL 2	6.0	7.22	1.30
RL 3	---	8.03	1.13
RL 4	6.7	6.84	1.05
SB 1	3.3	8.03	0.94

The data in Table 1 were collected as part of a study of heat flow and heat production for plutonic rocks. As a result, heat production and thermal conductivity are relatively well determined for the granites but still almost unknown for the country rock. In the current study a range of hypothetical country rock models was considered in order to determine if one-dimensional heat conduction theory applies to the region or if the three-dimensional aspects of the granitic bodies (edge effects) must be considered.

If the Rolesville batholith was emplaced before or during deformation as Farrar (1980) suggests, erosion following the differential uplift would cause the body to be thickest down plunge. The lack of variation of heat production and conductivity at the surface of the batholith, despite the proposed differential erosion, suggest that these two parameters are nearly uniform throughout the body. In the heat-flow models both granites have a uniform heat production of 6.0 HPU and conductivity of 7.5 TCU.

The heat flow (0.94 HFU) at station SB-1 in the volcanics of the adjacent Carolina Slate Belt is typical of background (country rock) heat flow in the southeastern United States (Costain _et al_., 1979). The heat flow from the lower crust and upper mantle in the southeastern United States is approximately 0.65 HFU (Costain and Perry, 1979). The regional heat flow field was then assumed to be made up of a 0.65 HFU component derived from heat production in the upper crust. The heat production of non-granitic surface samples in the southeastern United States range from near 1.0 HPU to near 6.0 HPU (L. D. Perry, pers. comm., 1978). The average country rock heat production was assumed to be in this range. The thermal conductivity of the country rock was assumed to be 6.5 TCU; this is a representative value for schist, gneiss, and volcanics (Clark, 1966) which are the dominant rock types of the Raleigh belt country rocks.

Four heat-flow models were developed which are consistent with the surface geology (Fig. 3) and Farrar's structural interpretation and based on the four country rock heat production models shown in Figure 4. Each of the country rock models produces the required 0.29 HFU contribution to the regional heat flow. The horizontal outlines of the prisms which make up the four heat-flow models are the same for each model (Fig. 5). Only the vertical dimensions of the prisms were changed from model to model. Vertical dimensions for the prisms in models 1, 2, 3, and 4 are given in Tables 2, 3, 4, and 5, respectively. The theoretical heat flow fields produced by the four models agree with the observed heat flow to within 2.5% at the five heat flow stations. The theoretical heat-flow fields for models 1 and 2 are shown in Figures 6 and 7, respectively. Theoretical heat flow profiles and temperature cross-sections along lines A-A' and B-B' in the second model are shown in Figure 8.

Table 2. Vertical dimensions of the polygonal prisms which make up heat flow model 1, of the Rolesville batholith and Castalia pluton

Prism	Depth to Top	Depth to Base
1	0 km	1 km
2	1	10
3	10	20
4	0	2
5	2	8
6	8	22

Table 3. Vertical dimensions of the polygonal prisms which make up heat flow model 2, of the Rolesville batholith and Castalia pluton.

Prism	Depth to Top	Depth to Base
1	0 km	1 km
2	1	10
3	10	14
4	0	2
5	2	6
6	6	18

Table 4. Vertical dimensions of the polygonal prisms which make up heat flow Model 3, of the Rolesville batholith and Castalia pluton.

Prism	Depth to Top	Depth to Base
1	0 km	1 km
2	1	8
3	8	12
4	0	2
5	2	6
6	6	15

Table 5. Vertical dimensions of the polygonal prisms which make up heat flow Model 4, of the Rolesville batholith and Castalia pluton.

Prism	Depth to Top	Depth to Base
1	--	--
2	4.8 km	20 km
3	20	30
4	--	--
5	4.8	15
6	15	30

A comparison between the thicknesses of the polygonal prism models and granite thickness estimates based on one-dimensional heat conduction (Costain and Perry, 1979) is given in Table 6. The model thicknesses are highly dependent on the lateral heat production contrast between the granite and the country rock. Heat conduction is then three-dimensional in the region and edge effects must be considered.

Table 6. Thicknesses of one- and three-dimensional models of the Rolesville batholith and Castalia pluton at three heat flow stations. The thicknesses given by one-dimensional analysis are from Costain and Perry (1979).

	Model Thickness at Stations		
Model	CS1	RL2	RL4
Model 1	22 km	20 km	1 km
Model 2	18	14	1
Model 3	15	12	1
Model 4	30	30	1
1-D Model	14	11	6

Model thicknesses are also expected to depend on the conductivity contrast between the granite and the country rock. The maximum heat-flow anomaly due to the conductivity contrast of 6.5 TCU, for the four models considered is 0.11 HFU. If the average conductivity of the country rock is greater than 6.5 the model thicknesses (Table 6) would have to be increased to explain the observed heat flow. If the average conductivity is less than 6.5, thinner models could be used to explain the observed heat flow.

In all four polygonal prism models the Rolesville batholith is thickest in the south and thins northward. This northward thinning is consistent with the structural interpretation given by Farrar (1980). The large difference is the thicknesses of the Castalia pluton and the northern part of the Rolesville batholith is not supported by the gravity expression of the two bodies, however. The Bouguer gravity map of the model region (Cogbill, 1978) (Fig. 9) shows a -30 mgal anomaly at the site of RL4, where the heat flow is 1.05 HFU and the models are all 1 km thick. There is also a -30 mgal anomaly at the site of CS1, where the heat flow is 1.44 HFU and the models are 15 km to 30 km thick. Simmons (1967) gives the relationship between the vertical heat flow and gravity anomalies at the surface, caused by volume sources with both contrasting density and heat production. Heat flow varies significantly between stations RL3 and CS1 but the gravity field does not. It can be concluded that variations in gravity and heat flow over the model region cannot be explained by sources with the same shapes. A possible explanation can be found in the low density country rock in which the batholith was emplaced. In hand specimen, the densities of the dominant metamorphic rock types in Raleigh belt are not significantly different from the density of the main phase of the Rolesville granite. The gravity field is then indicative of the thickness of the entire Raleigh belt rather than the thickness of the granite alone. The heat flow field would reflect only the thickness of the granite.

Summary

New analytical solutions have been developed for temperature and heat-flow anamolies caused by a polygonal prism source in a two-layered half space. An approximate solution to the problem of a polygonal prism with contrasting conductivity in a two-layered half space was also given.

A comparison was made between the exact and approximate heat flow over an infinite semi-circular cylinder of contrasting conductivity. The two heat flow fields agree to within 5% for conductivity ratios between 0.25 and 1.5.

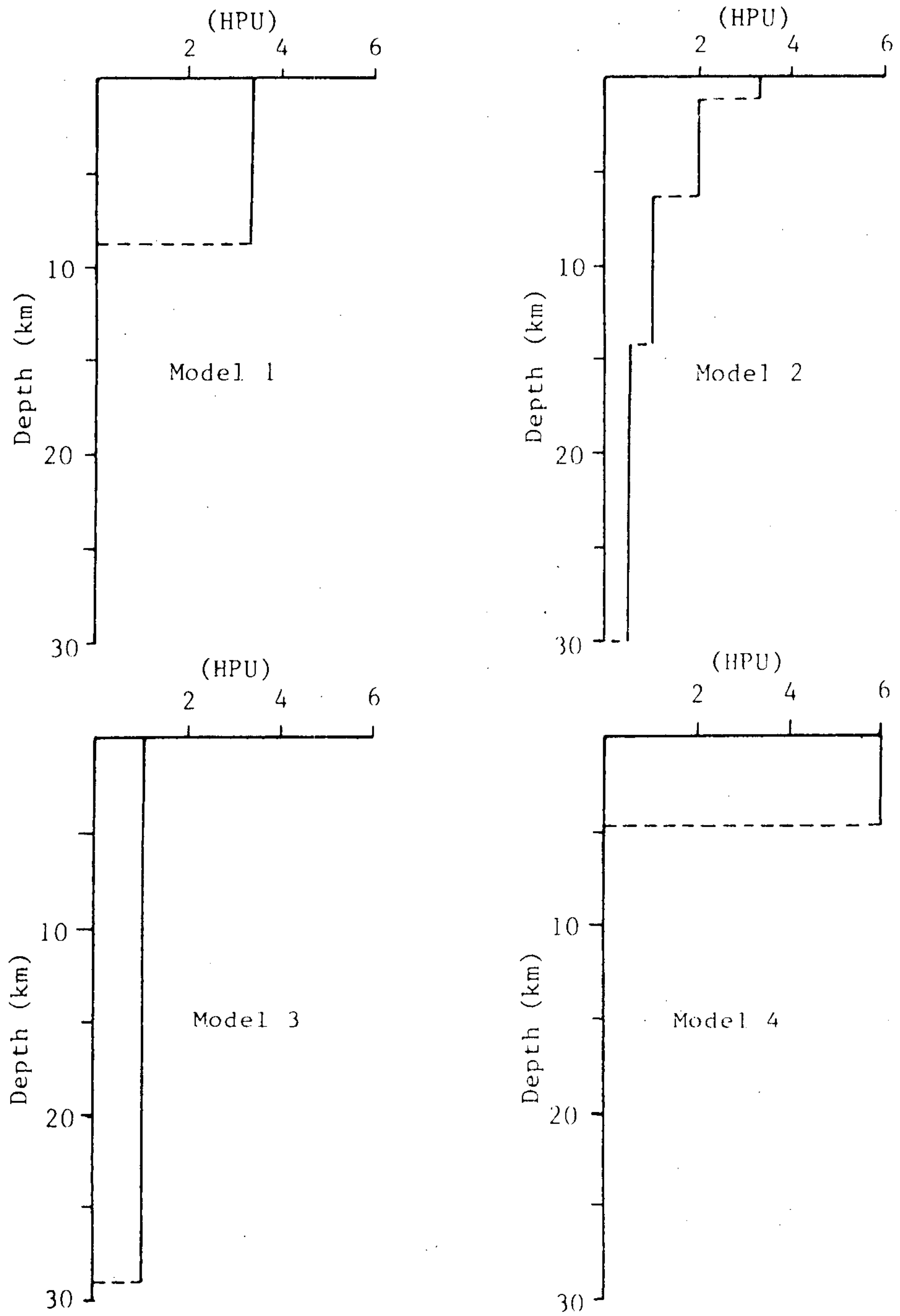

Figure 4. Vertical country rock heat production distributions for the four heat-flow models of the Rolesville batholith and Castalia.

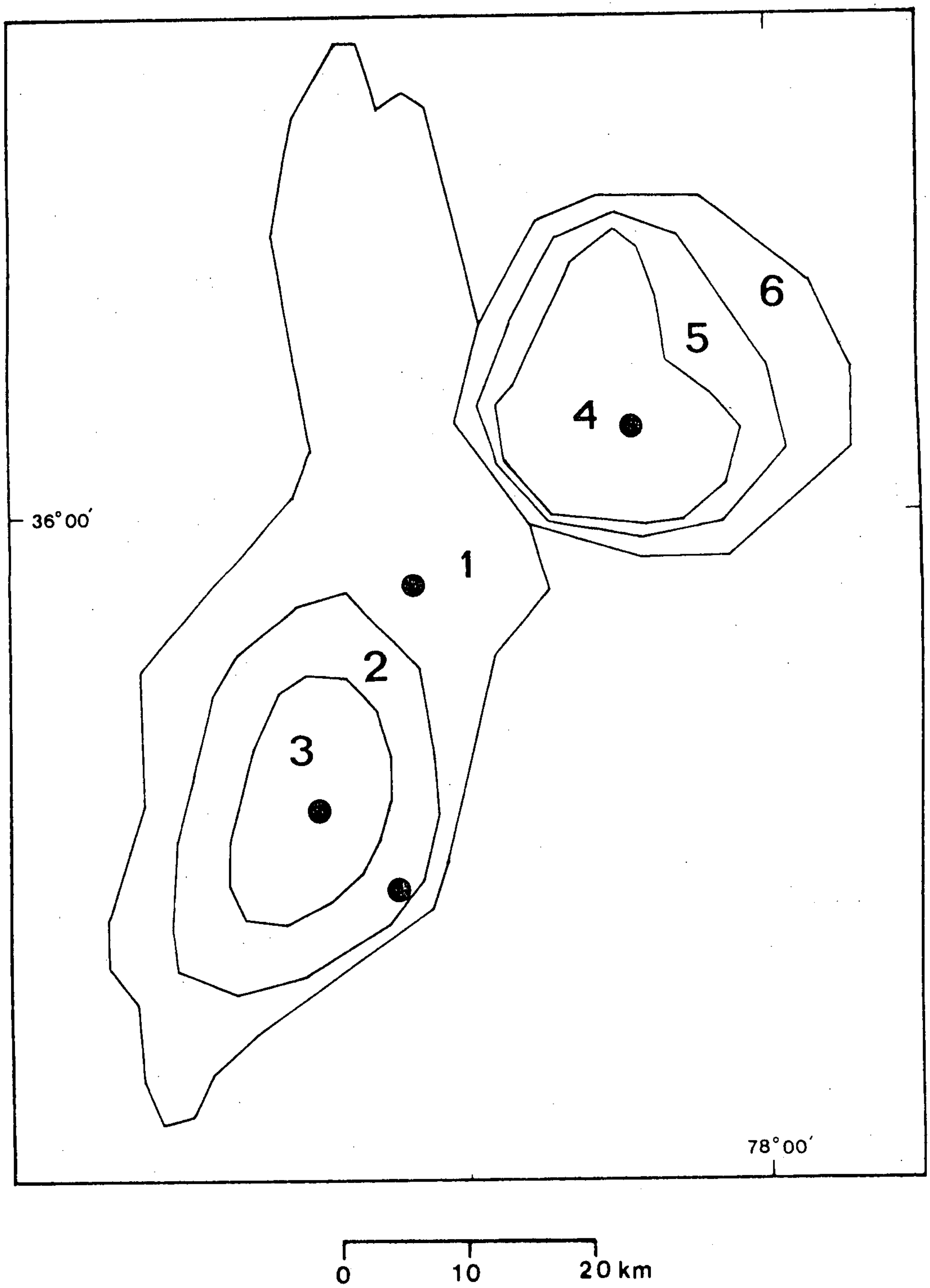

Figure 5. Outlines of the six prisms used to represent the Rolesville batholith and Castalia pluton in all four heat-flow models.

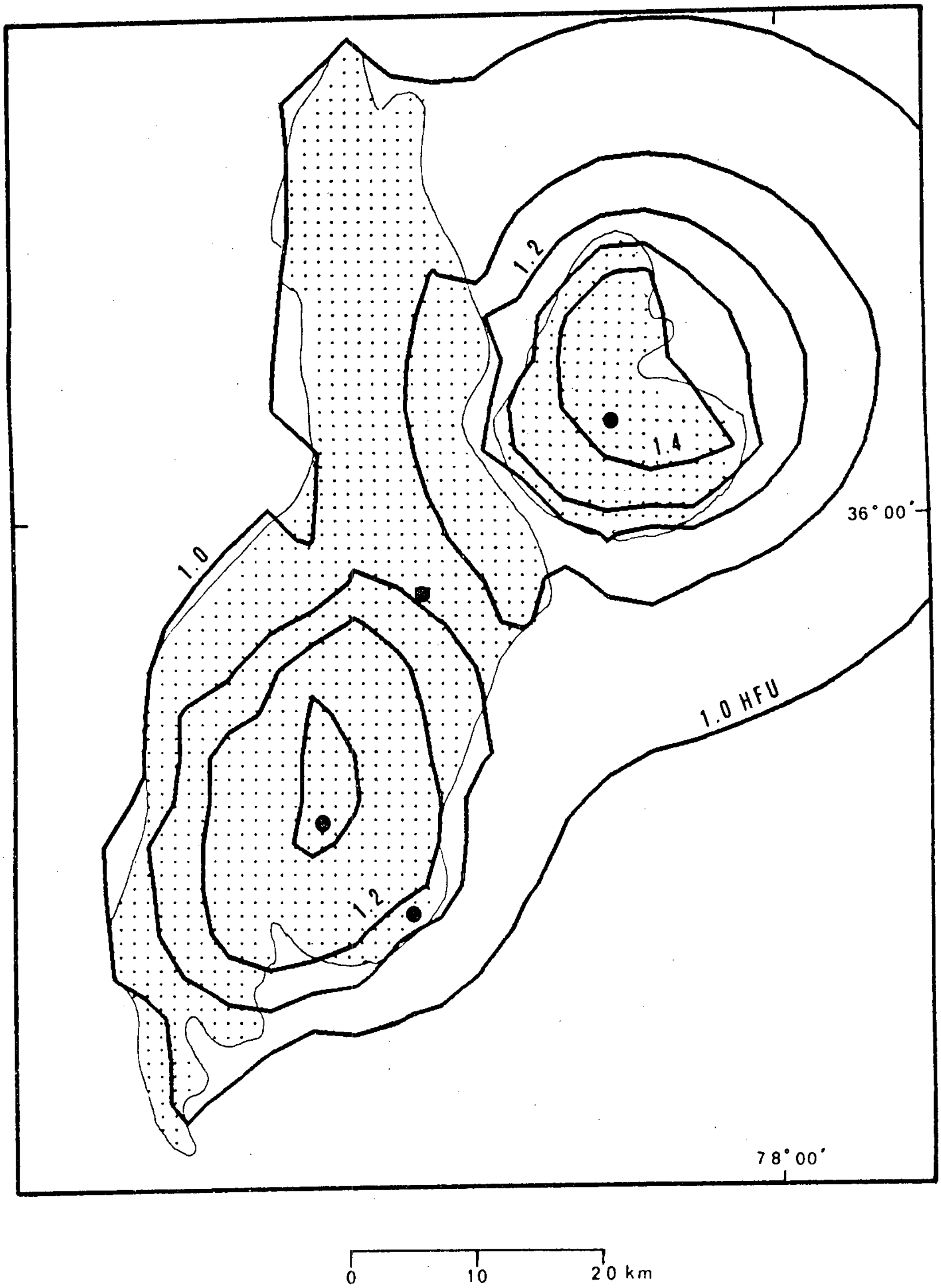

Figure 6. Theoretical heat flow map for Model 1 of the Rolesville batholith and Castalia pluton. The vertical country rock heat production distribution for Model 1 is given in (Fig.4). The polygonal prism model is shown in (Fig. 5).

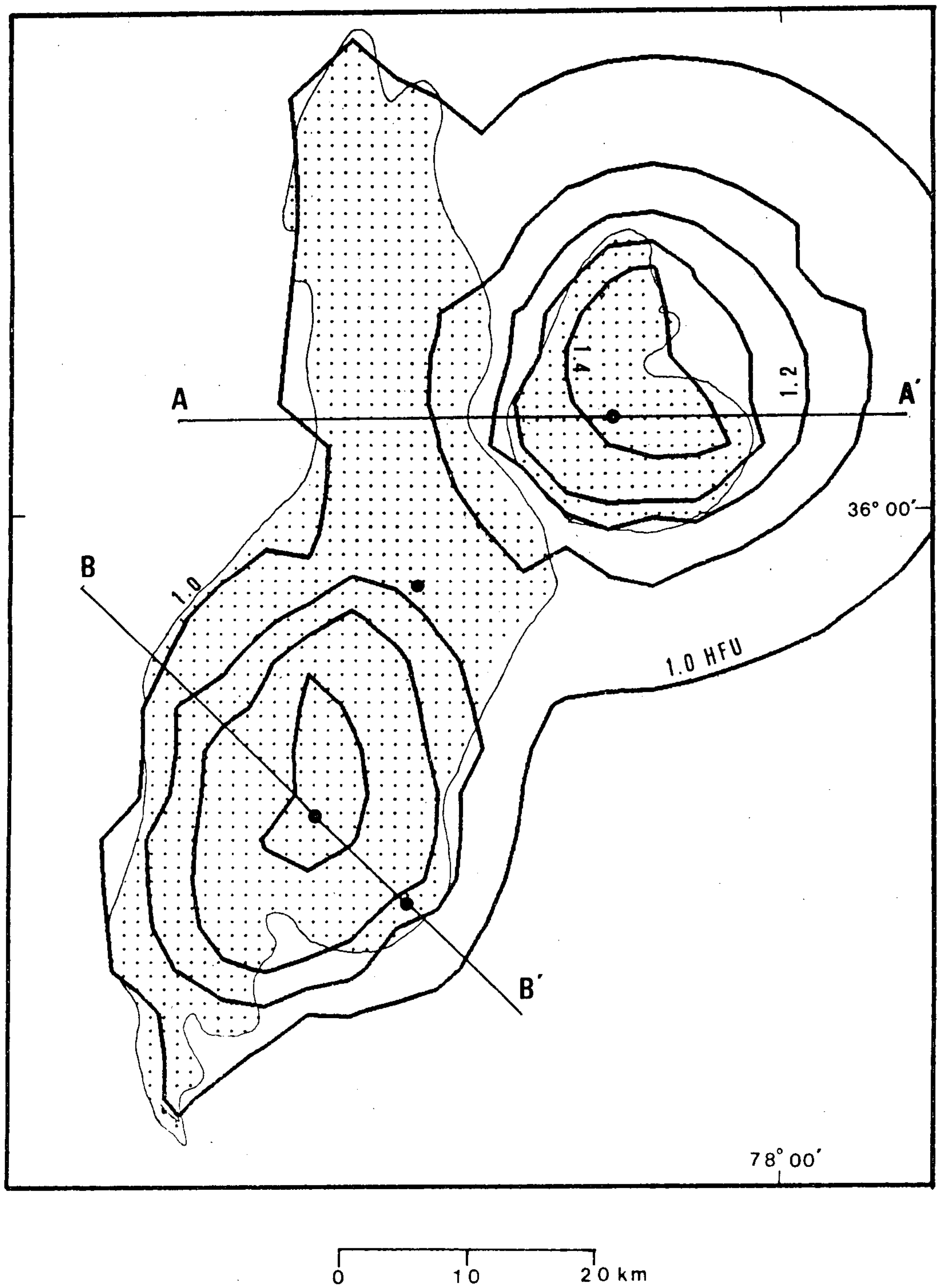

Figure 7. Theoretical heat flow map for Model 2 of the Rolesville distribution for Model 2 is shown in (Fig. 4). The polygonal prism model is shown in (Fig. 5).

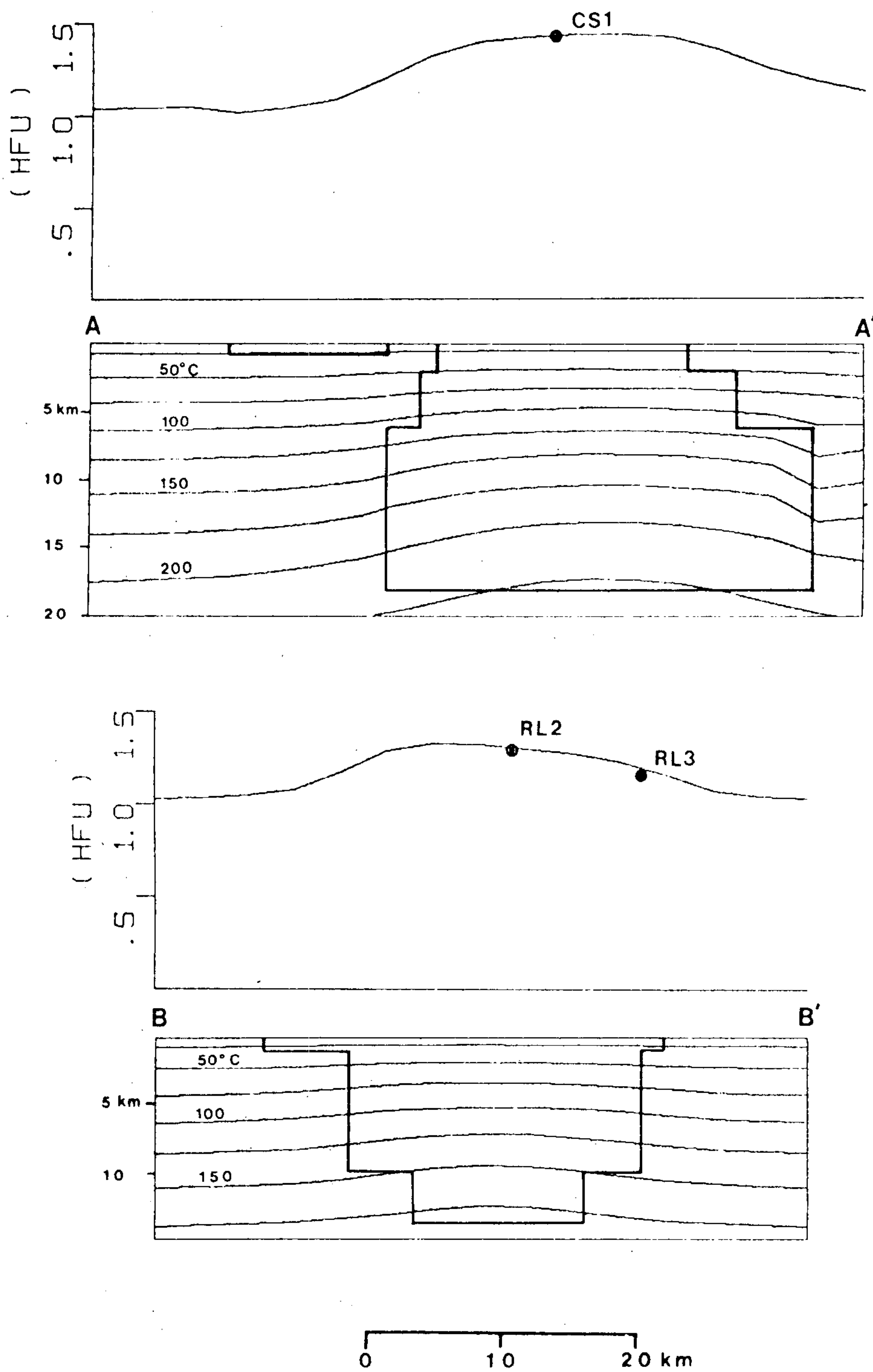

Figure 8. Theoretical heat flow profiles and temperature cross-sections for Model 2, along lines A-A and B-B (Fig. 7).

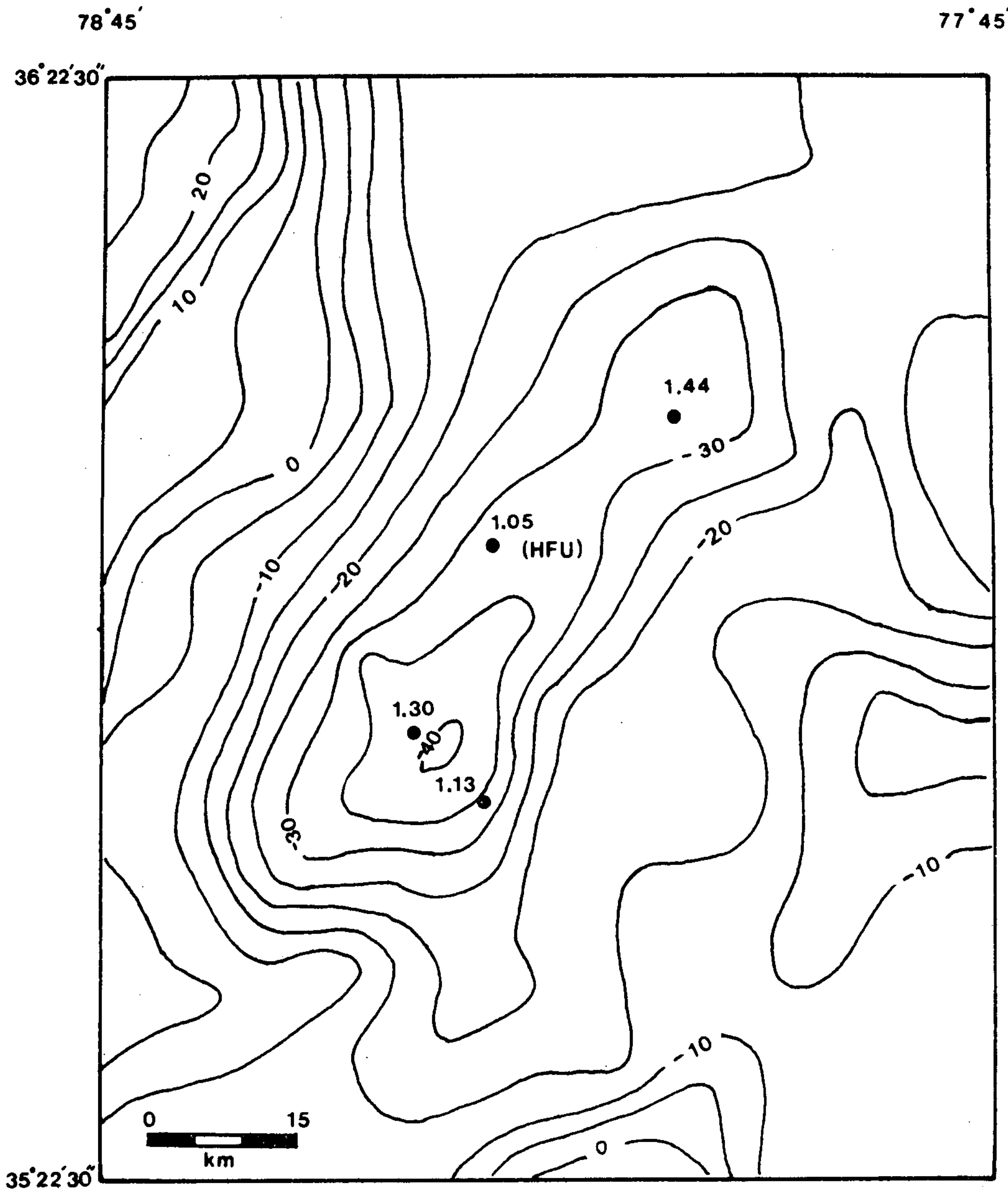

Figure 9. Bouguer gravity field (contoured in mgal) (from Cogbill, 1979) and heat flow (HFU) (from Costain and Perry, 1979) in the vicinity of the Rolesville batholith; dots indicate locations of heat-flow stations.

Comparing this error range with that for a semi-spheroid of contrasting conductivity indicates that the error level does not change rapidly with changes in the shape of the anomalous body. A survey of thermal conductivities of different rocks types (Clark, 1966) suggests that this range of conductivity ratios is sufficient for modeling many geologic problems. The uncertainty in the five heat flow values (Table 1) used is about 0.05 HPU or approximately 10% of the total heat flow anomaly in this region. This error level is typical of other heat flow determinations made in crystalline rocks (Costain and Perry, 1979). The 5% error level in the component of heat flow due to contrast in thermal conductivity is then not excessive.

The modeling technique based on these solutions was used to interpret a localized heat flow anomaly. The interpretation was carried out in the same way a gravity anomaly would be analyzed using Plouff's method (1976). Temperature on cross-sections through the model region was estimated by computing the temperature field associated with the heat-flow model. This method of estimating the amplitude and spacial extent of temperature anomalies could be useful in the exploration for low-temperature geothermal resources.

The modeling technique is also applicable to regional heat flow studies. In many regional studies (see for example Roy et al., 1968; Lachenbruch, 1970; Costain and Perry; 1979), one-dimension heat conduction is assumed. Lateral variations in heat production and conductivity cause deviations from one-dimensional conduction. The modeling technique presented in this study can be used to estimate the heat flow effects of variations in heat production and conductivity in the vicinity of a heat flow station. The heat flow value can then be adjusted accordingly.

These problems can also be solved with numerical techniques such as the method of finite differences. However, computer storage requirements of detailed three-dimensional models may exceed the space available at most computer installations. Computation time requirements can also become prohibitive, particularly when trial-and-error fitting methods are used. In the modeling technique presented in this study unknowns are calculated at points of interest only. Storage requirements are therefore not a problem. The modeling technique is also computationally efficient; the heat flow models of the Rolesville batholith and Castalia pluton required 0.26 msec computation time per field point per polygonal edge on an IBM 370/158 computer.

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ON THE USE OF COMPLEX REFLECTION COEFFICIENTS TO DETERMINE BASEMENT ROCK TYPE

John K. Costain and Cahit Coruh

Introduction

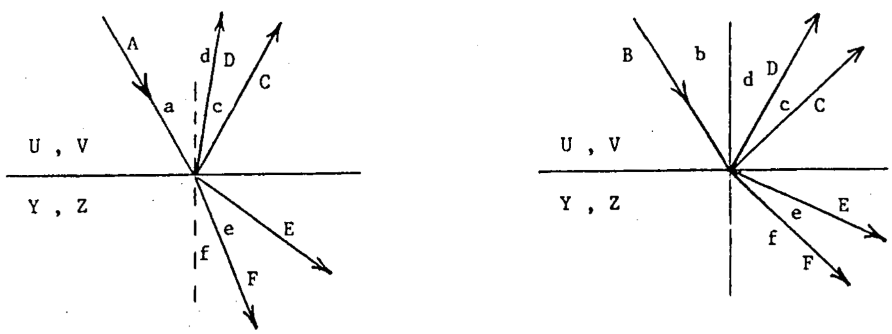
When a longitudinal or transverse elastic wave strikes at non-vertical incidence the boundary between two media within which the wave velocities and/or densities are different, four derived waves come into existence: a reflected longitudinal wave, a reflected transverse wave, a transmitted longitudinal wave, and a transmitted transverse wave. The energy of the incident wave is divided among the four derived waves in proportions which depend upon the elastic constants and densities in the two media, and on the angle of incidence. Zoeppritz (1919) derived equations for the relative amplitudes of the derived waves with respect to the amplitude of an incident plane wave. The equations comprise a system of four linear equations which can be solved simultaneously. The physical conditions assumed for the derivation of the equations are:

1) The vector sum of the displacements due to all waves along the boundary in one medium must equal the corresponding sum in the other medium. Otherwise there would be relative slip between the two media.

2) The corresponding stress components must be equal on the two sides of the interface.

The solution of these equations (Fig. 1) gives the reflection and transmission coefficients. The coefficients are sensitive to the angle of incidence, and a relatively small change in the angle of incidence can result in a large change in a coefficient. It is apparent from Figure 1 that increasing the angle of incidence of an incident P wave could eventually result in two critical angles. The first occurs when the angle of refraction of the transmitted P wave reaches 90^0; the second occurs when the angle of refraction of the transmitted S wave reaches 90^0. Similarly, inspection of the figure for an incident S wave shows that three critical angles are possible. When a critical angle is exceeded, the wave type associated with that critical an-

Incident P   Incident SV



	Velocity	Amplitude	Angle with Normal
Incident P	U	A	a
Incident S	V	B	b
Reflected P	U	C	c
Reflected S	V	D	d
Refracted P	Y	E	e
Refracted S	Z	F	f

Incident P wave

$$(A - C) \sin a + D \cos b - E \sin e + F \cos f = 0$$

$$(A + C) \cos a + D \sin b - E \cos e - F \sin f = 0$$

$$-(A + C) \sin 2a + D(U/V) \cos 2b + EK(Z/V)^2 (U/Y) \sin 2e - FK(Z/V)^2 (U/Z) \cos 2f = 0$$

$$-(A - C) \cos 2b + D(V/U) \sin 2b + EK(Y/U) \cos 2f + FK(Z/U) \sin 2f = 0$$

Incident S wave

$$(B + D) \sin b + C \cos a - E \cos e - F \sin f = 0$$

$$(B - D) \cos b + C \sin a + E \sin e - F \cos f = 0$$

$$(B + D) \cos 2b - C(V/U) \sin 2a + EK(Z^2/VY) \sin 2e - FK(Z/V) \cos 2f = 0$$

$$-(B - D) \sin 2b + C(U/V) \cos 2b + EK(Y/V) \cos 2f + FK(Z/V) \sin 2f = 0$$

Figure 1. Notation associated with Zoeppritz equations for incident P and incident S wave.

gle disappears, and the incident energy is distributed among the remaining waves. After the first critical angle is exceeded, all reflection and transmission coefficients for the remaining wave types become complex. The physical amplitude corresponding to a remaining wave type is the square root of the sum of the squares of the real and imaginary parts of the complex coefficient.

Physical Significance of a Complex Reflection Coefficient

Plane wave reflection and transmission coefficients for a single interface are independent of frequency. The coefficients apply to a single sinusoid of a single frequency incident at some angle on an interface. Before a critical angle is reached the only phase change possible for the incident sinusoid is a phase shift of 180^0, which is simply a polarity reversal. However, after a critical angle is reached one of the reflected or transmitted P to S waves exists no more (which one depends on the P velocity ratio, the S velocity ratio, the density contrast and the angle of incidence). All reflection coefficients for remaining reflected and transmitted P or S waves become complex after any critical angle is reached. This simply means that the reflected or transmitted P or S sinusoid is shifted in phase (time) with respect to the incident sinusoid. This is accompanied by a change in the amplitude of the reflected or transmitted sinusoid. Changing the phase of a sinusoid means shifting its peaks one way or the other (toward positive or negative time). The direction is given by the "time-shifting theorem" and is unambiguous.

Because any source wavelet can be decomposed into its component sinusoids, we can use the complex reflection coefficient and thus the phase change, which is the same for each component sinusoid, to compute the change in shape after reflection for any wavelet. In the frequency domain the Fourier spectrum of the original wavelet can be multiplied by the complex reflection coefficient (Fig. 2). This adds or subtracts the phase shift to the phase angle spectrum. The modulus of the complex reflection coefficient scales the original modulus for that same frequency up or down. An inverse Fourier transformation then shows the effect on the original wavelet.

A positive phase shift (counterclockwise into quadrant I from the positive real axis, or counterclockwise in quadrant III from the negative real axis) means a shift into negative time, i.e., the wavelet arrives earlier than would be expected. P.N.S. O'Brien (1963) notes that:

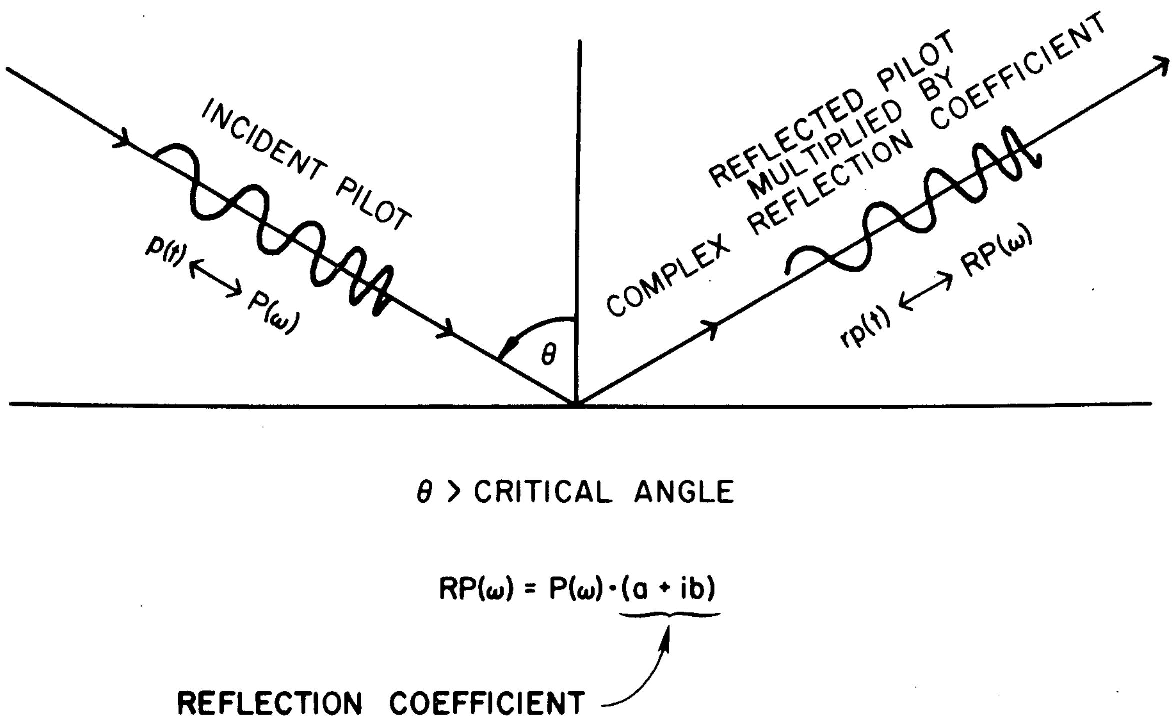


Figure 2. Reflected sweep is multiplied by complex reflection coefficient resulting in distortion of incident sweep.

"It has been objected that a phase lead, which necessitates a time lead, is physically impossible. There seems to have been no discussion of this point in the literature, but a possible explanation is as follows. When a plane pulse is incident on a boundary it acts as a line source and generates a disturbance which travels along the interface with the velocity of the lower medium. For angles of incidence greater than critical this velocity is greater than the trace velocity of the incident wave and, therefore, the interface disturbance will generate a wave in the upper medium which travels ahead of the reflected pulse predicted by ray theory. Because a plane wave means that the source lies at infinity, the interface wave will provide an infinite forerunner to the pulse traveling along the reflected ray path.

In real systems, of course, the source emits curved wavefronts from a local source and in this case a head wave exists. As this always arrives earlier than the reflected pulse the phase lead associated with wide angle reflections merely means that the disturbance between the two events never completely dies away."

There is considerable distortion of the wavelet depending on the phase shift. If the signal to noise ratio is adequate, then the observed distortion of the wavelet can be used to advantage.

Application

"Basement" beneath the Atlantic Coastal Plain, from the standpoint of possible applications for geothermal energy, is either crystalline rock (granite, gneiss, etc.) or sedimentary strata of Triassic-Jurassic age. It is well known that the velocity of Triassic rocks is considerably less than that of crystalline rocks (4-5 km/sec versus 6-7 km/sec). In addition, the density of Triassic sediments is less than that of crystalline rocks. Therefore, the onset of the critical angle for reflected P-waves will be considerably different for Triassic sediments than for crystalline basement. This is important information because although crystalline rocks offer no geothermal resource potential (other than hot-dry-rock) in the eastern United States, the possibility exists that Triassic sediments might have a permeability high enough in some areas to be considered as a possible hydrothermal target. Triassic-Jurassic sediments do not look favorable for this purpose at the present time, but Triassic basins are poorly exposed and studied, and are completely concealed beneath the sediments of the Atlantic Coastal Plain. Only drilling associated with the exploration for hydrocarbons offers hope in the near future of revealing the intrinsic permeability and degree of consolidation of Triassic-Jurassic sediments. Until such time as actual core samples become available, it might be possible to predict lithologies and the degree of consolidation by investigating pulse shape distortion beyond the critical angle, as described in this summary.

Comparison of Theoretical Waveforms with Seismic Data from Portsmouth, Virginia

Plane wave reflection coefficients computed from the Zoeppritz equations are shown in Figure 3 for the geologic model of Coastal Plain sediments over Triassic and granite basement. It is clear that the onset of the critical angle is quite different for an assumed Triassic basement versus a granite basement.

Figures 4a and 4b show typical VIBROSEIS 100% seismograms as recorded on the Atlantic Coastal Plain near Portsmouth Virginia by VPI&SU. This site is one of the best known to date in the eastern United States for confirmation of the radiogenic model (Costain et al., 1980). Data quality is excellent and allows direct examination of the 100% correlated field data for pulse distortion. Preliminary results of pulse distortion after the critical angle were given by Costain and Coruh (1980).

The theoretical basis for comparison of the real data with synthetic data is as follows. The Fourier transform of the pilot sweep is computed. This is multiplied by the constant complex reflection coefficient computed from the Zoeppritz equations. The reflection coefficient depends upon the angle of incidence of the Klauder wavelet on the basement surface. The angle of incidence is, of course, a function of the depth to basement, the spread geometry, and the velocity distribution of the overlying sediments. Synthetic data were generated using a ray tracing program that rigorously obeys Snell's law at all interfaces. The output of the ray tracing model is shown in Figure 5. The real and complex reflection coefficients corresponding to the angles of incidence of each raypath at the sediment-basement interface were determined by the ray tracing program, which incorporates the Zoeppritz equations.

The synthetic seismogram output from the ray tracing program is shown in Figure 6. Two characteristics of the synthetic data are (1) significant pulse distortion caused by the com-

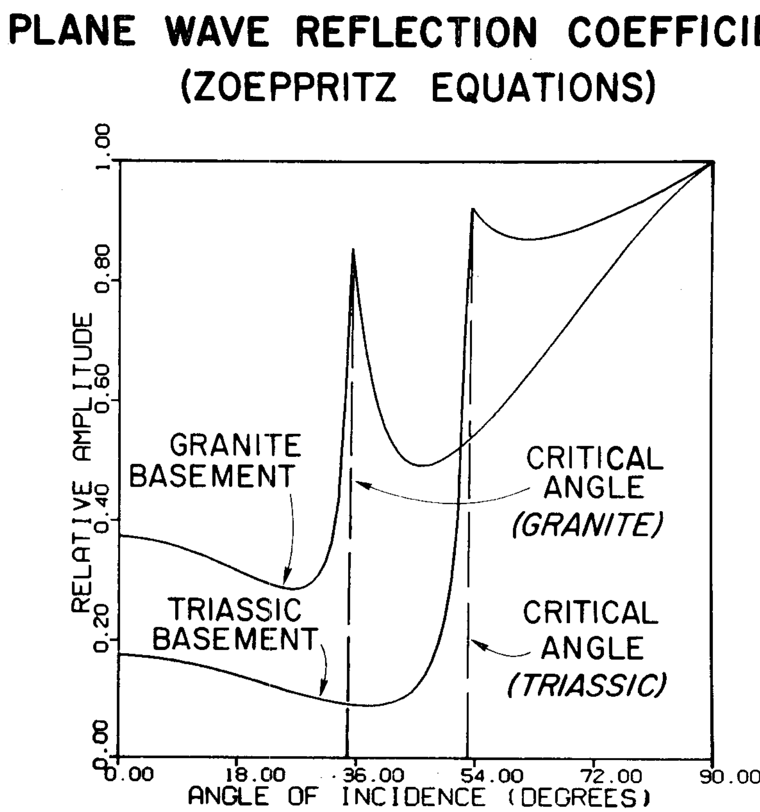
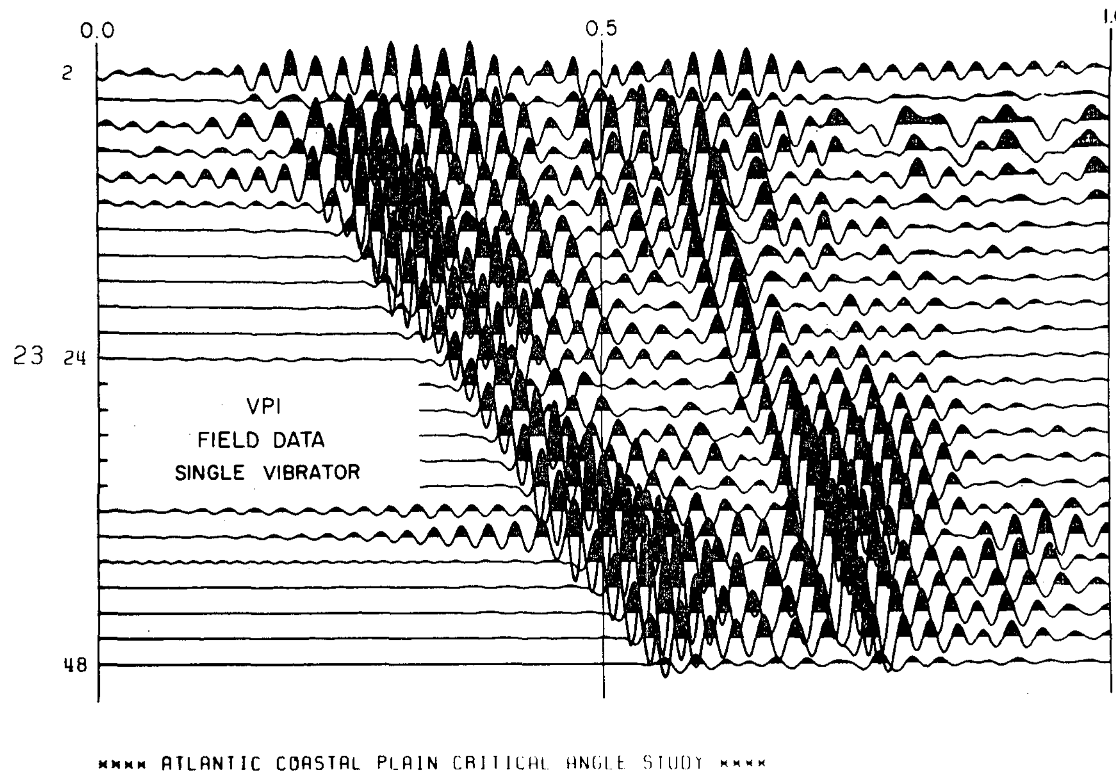


Figure 3. Plane wave reflection coefficient for model of Costal Plain sediments overlying granite basement and Triassic basement.



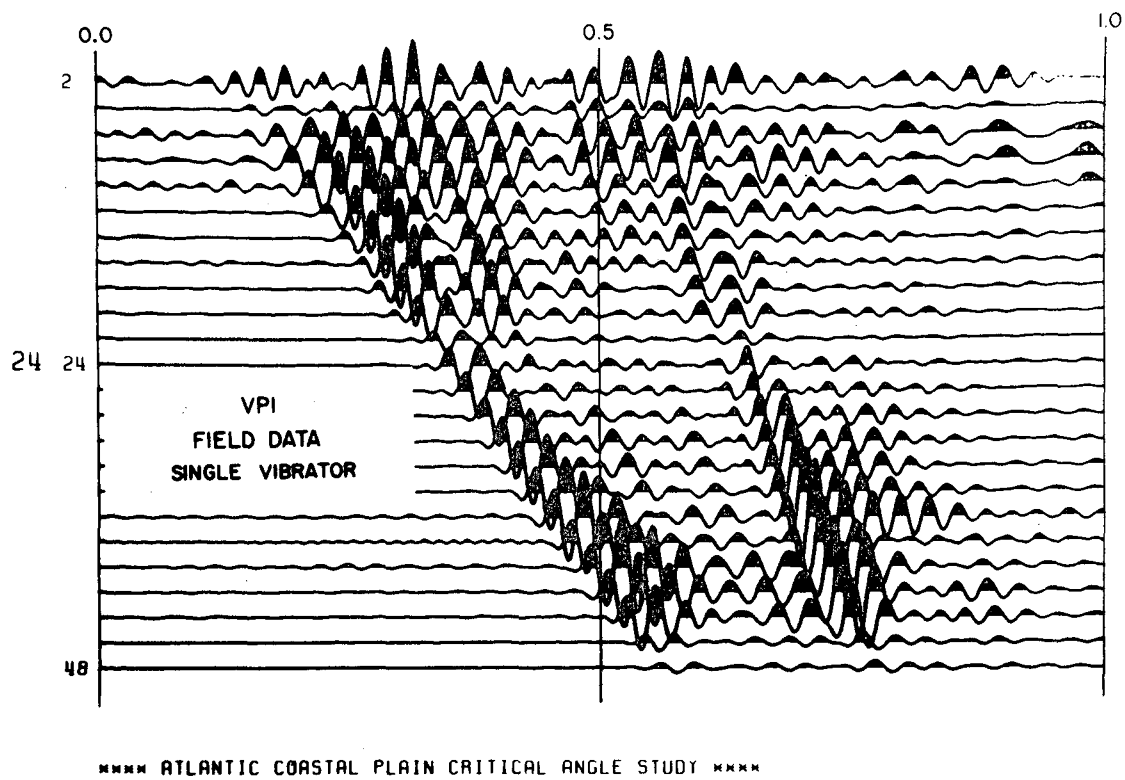


Figure 4a-b. Correlated 100% field data from the Portsmouth area.

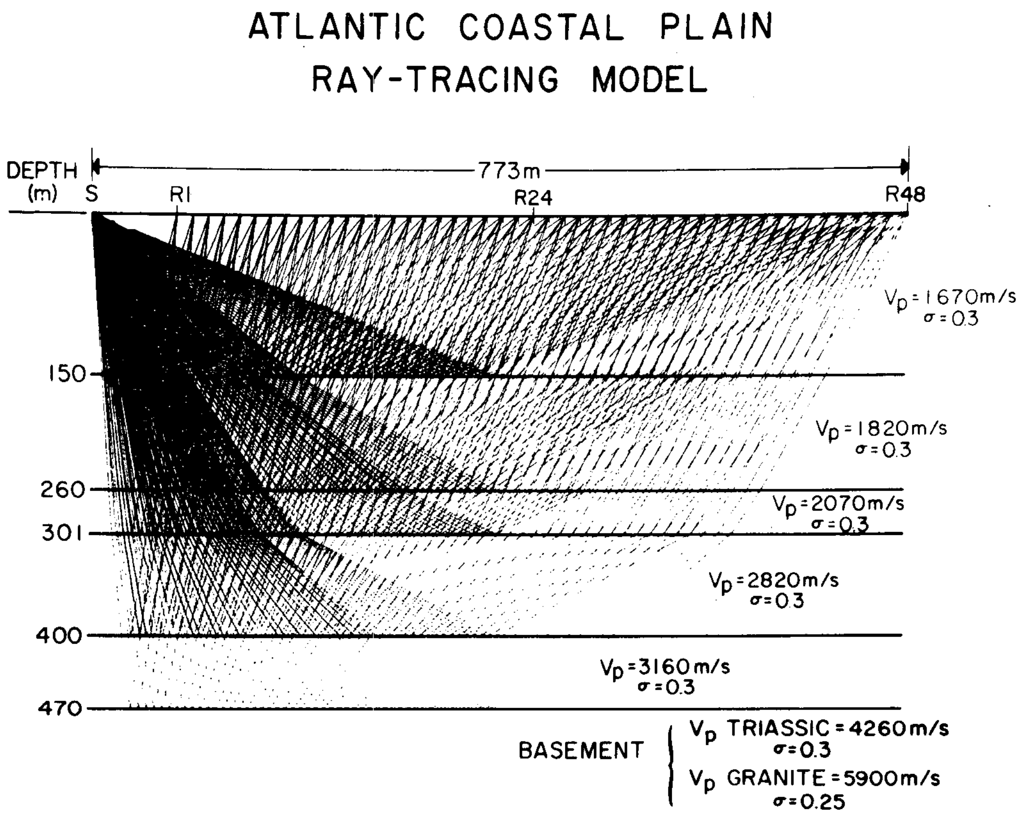


Figure 5. Output of ray-tracing program for Portsmouth, Virginia.

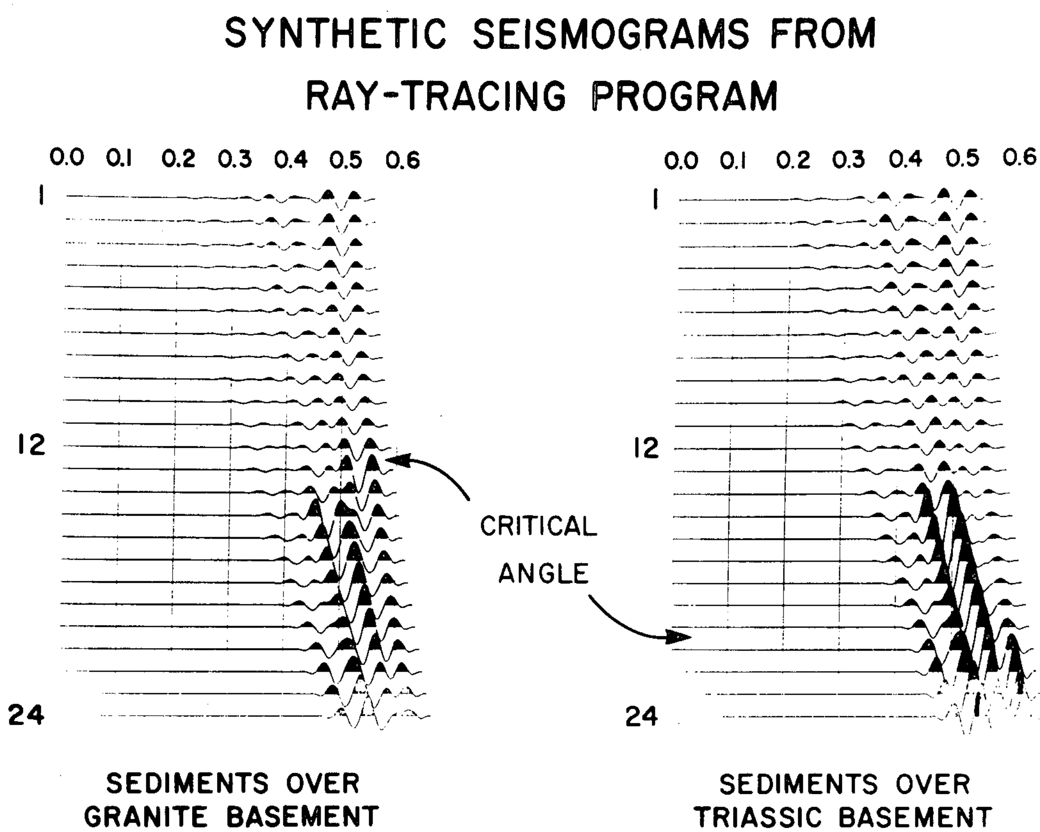


Figure 6a. Output of ray-tracing program showing onset of critical angle and increase in amplitude past the critical angle.

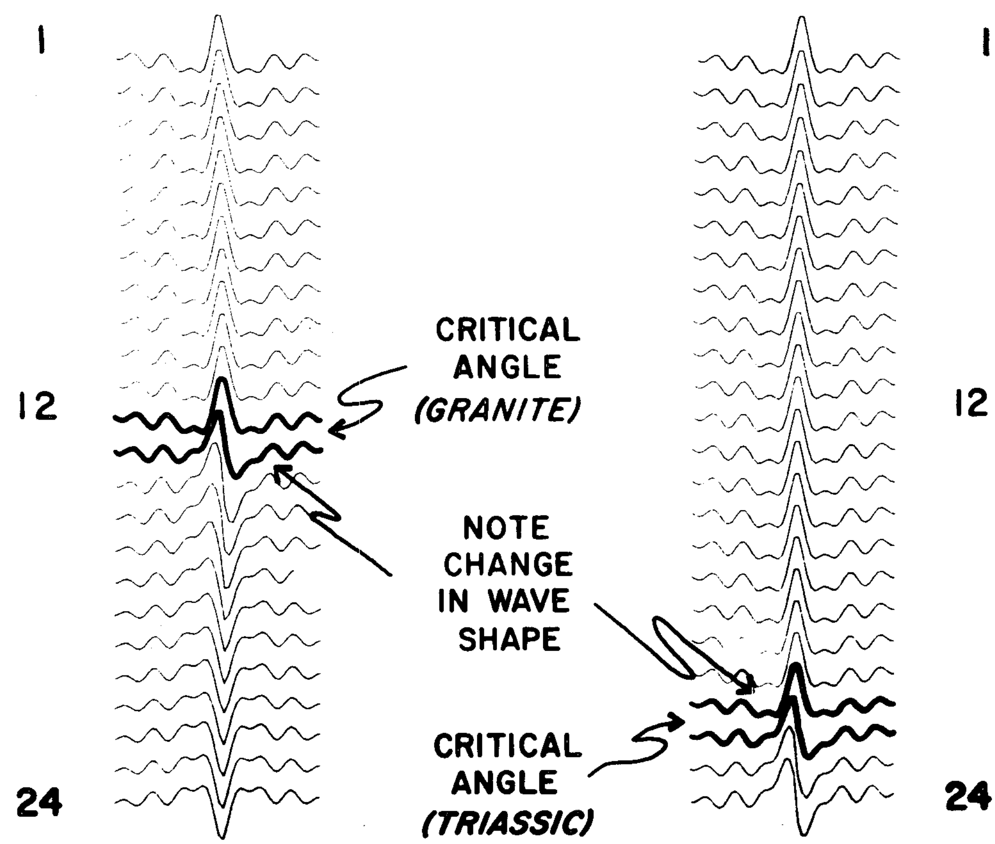


Figure 6b. Result of multiplication of Fourier transform of pilot sweep by complex reflection coefficient and subsequent inversion to the time domain.

plex reflection coefficient, and (2) significant difference in the onset of the critical angle for a granite basement versus Triassic basement. Comparison with the real data of Figure 4 shows a similar distortion in the real data. It is important to note that, if a source wavelet is minimum delay before reflection beyond the critical angle, it cannot be minimum delay after reflection because the (plane wave) reflection coefficient is independent of frequency.

Conclusions

Rather than avoiding a recording geometry that precludes recording reflections after the critical angle, such large-amplitude reflections could be used to advantage to determine basement rock type where drill control is absent or sparse. The method does require a good signal-to-noise ratio, but it apparent from the data shown in Figure 4 that this is not a problem on the Atlantic Coastal Plain. The method shows great promise for tracking major changes in basement rock type from the surface.

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RELATIONSHIP BETWEEN HYDROTHERMAL ENERGY OUTPUT AND WELL SPACING IN A TYPICAL ATLANTIC COASTAL PLAIN GEOTHERMAL DOUBLET

R. J. Laczniak

Introduction

The geothermal resources of the southeastern United States are classified as liquid-dominated low-temperature systems (Costain *et al*., 1980), characterized by temperatures less than 90°C (White and Williams, 1975). The development of an efficient method of obtaining thermal energy from low temperature systems requires an in-depth study of each unique system (Johns Hopkins University, 1981).

Major problems facing the utilization of geothermal energy in the southeastern United States are land subsidence, aquifer collapse, and the disposal of large quantities of highly saline waters. One solution is to inject the water back into the aquifer. This maintains piezometric pressures, and thus prevents subsidence and insures a continuous supply of water. Of concern is the possibility of adverse chemical and physical reactions caused by the injected fluid.

Two basic methods of reinjection are: (1) the forward and return flow method, and (2) the flow-through method (Ingenjorsbyran, 1978). The first method utilizes a single pumping-injection well. Water is pumped, stored, and later reinjected by way of a single well (Lippman *et al*., 1977; Claesson *et al*., 1978). The flow-through method uses separate pumping and injection wells which comprise a sink-source, or doublet system. The energy extracted from either system by a heat exchanger is directly proportional to the temperature of the water at the production well.

The subject of this paper is an evaluation of the thermal energy output and longevity of a doublet system. As the well spacing between the sink and source increases, engineering costs increase and land acquisition may become more difficult. Because economic considerations are crucial to the success of any energy producing system, well spacing should be minimized. The topic of well spacing has been discussed by Gringarten and Sauty (1975) for the Dogger aquifer near Paris, France, where the optimum well spacing was found to be 900 m.

As additional hydrologic and geothermal data become available for the aquifers of the Atlantic Coastal Plain, it becomes possible to investigate by means of numerical modeling on the computer, the relationship between the energy extracted and the distance between the source and sink well in a doublet system. Such modeling can aid in the evaluation of a particular doublet by defining such limiting characteristics as the thermal lifetime of the system, and the optimum well spacing for the specific application.

The subject of this paper is the relationship between well spacing and the amount of energy extracted from a "typical" Atlantic Coastal Plain geothermal system. Various well spacings, aquifer permeabilities, and pumping-injection rates are considered in the numerical simulations over a period of 15 years. The numerical method is based upon the method of Lippmann et al. (1977) and uses computer code CCC (compaction-convection-conduction) developed at Lawrence Berkeley Laboratories. Program CCC is used to calculate the thermal and fluid-flow fields for a three-dimensional rectangular liquid-dominated low-temperature system.

The Numerical Model

CCC is is a computer program written in FORTRAN IV for execution on the CDC 7600 system at Lawrence Berkeley Laboratories. It was modified by Laczniak (1980) for execution on the IBM 370 computer at VPI&SU.

Program CCC uses the "integrated finite difference method" for the numerical solution of the heat and mass flow equations (McNeal, 1953; Todd, 1959; Tyson and Weber, 1964; Cooley, 1971; Narasimhan and Witherspoon, 1976). The equations of mass and flow are coupled by interlacing them in the time domain. A complete and detailed discussion of the derivations is found in papers by Narasinham and Witherspoon (1976, 1978) and Sorey (1975).

The basic concept of the IFDM is to discretize a region into a number of smaller subregions (Fig. 1). Each element of the system is then described by a finite difference approximation to the partial differential equation governing confined transient groundwater flow (Laczniak, 1980, p. 5-6).

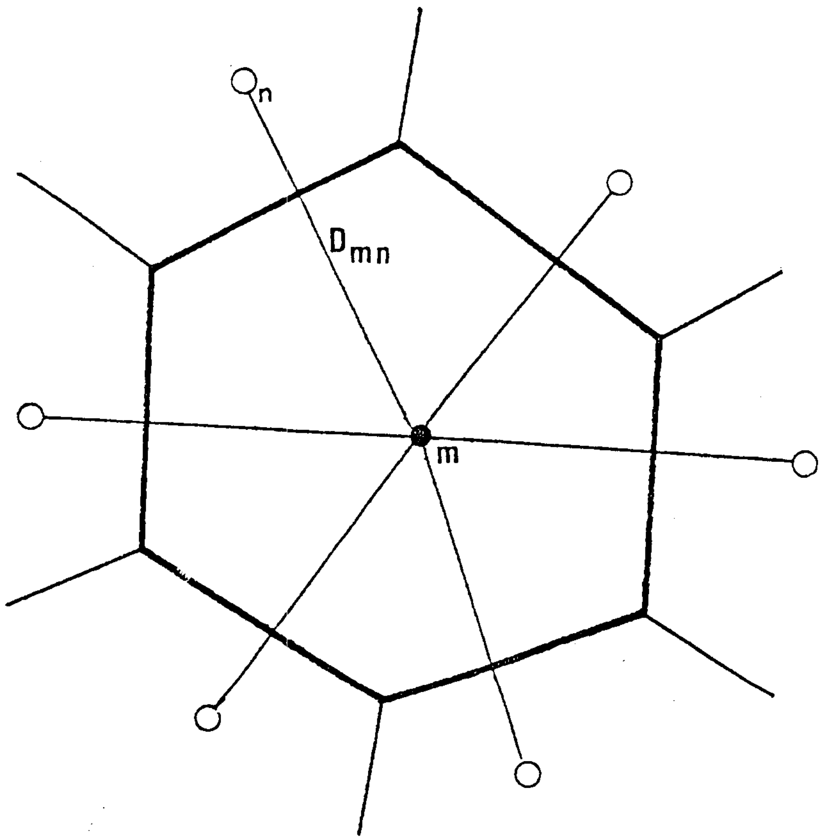


Figure 1. Illustrative representation of a descretized region showing a typical nodal element, m, and typical neighboring nodal element, n.

General Hydrogeology of the Atlantic Coastal Plain

The sediments beneath the Atlantic Coastal Plain (Brown et al., 1972) consist of a seaward-thickening wedge of interbedded sands and clays deposited by fluvial, deltaic, and marine processes. These strata consist of a sequence of moderate to highly productive aquifers overlying a pre-Cretaceous basement complex (Brown et al., 1972). The principal water-bearing units are relatively unconsolidated sands. The less permeable clays act as confining beds and subdivide the system into numerous semi-independent aquifer systems with minimally dependent values of hydraulic head.

The sediments beneath the Atlantic Coastal Plain are divided into three general subsystems according to geologic age: (1) the Quaternary system; (2) the Tertiary system; and (3) the Cretaceous system. The uppermost hydrologic unit, the Quaternary system, overlies the first major clay layer, and is a water-table, or unconfined aquifer. Production is low to moderate, and the aquifers are seldom used for anything other than domestic purposes. Aquifer thickness is generally less than 30 m.

The Tertiary and Cretaceous systems underlie the Quaternary system, and are a series of confined and semi-confined aquifers. Both are moderate to highly productive and are major sources of water for domestic and industrial use. Aquifer thickness varies from 30 to 120 m.

Regional recharge is by direct precipitation and runoff. Recharge to the underlying aquifers occurs east of the Appalachian Mountains, the major groundwater divide. Local recharge is by direct vertical leakage from precipitation. The relative contribution of regional (horizontal) recharge to vertical recharge is not well known. Vertical recharge by leakage to any particular aquifer is controlled by the vertical permeability of the sediments above the aquifer. The low permeability of the clays of Tertiary and Cretaceous age retard vertical recharge; therefore, without reinjection the potential yield of the aquifers is limited.

Water quality depends upon the mineral constituents and the chemical properties of the host formation. These vary with location. One general trend is the presence of high concentrations of total dissolved solids along the eastern shore; for example, concentrations of total dissolved solids were measured at 72,039 ppm in a deep test well near Crisfield, Md. (Hartsock, 1979).

Description and Discussion of a "Typical" Atlantic Coastal Plain Geothermal System

The following characteristics of a "typical" Atlantic Coastal Plain geothermal system are taken from data and information obtained from the Crisfield, Md., geothermal test well drilled during June, 1979 (Dashevsky and McClung, 1979) and supplemented with information from the literature (Brown and Silvey, 1972; Cedarstrom, 1945; U.S. Geological Survey, 1967). A schematic diagram of the system is shown in Figure 2.

The homogeneous aquifer is bounded on top by an aquitard consisting of semi-permeable sands, clays, and shales. The aquitard is overlain by another aquifer. The bottom of the principle aquifer is assumed to be bounded by impermeable material represented by either well consolidated sediments or crystalline rocks. In the horizontal direction, the system is unbounded (i.e., horizontally infinite). The vertical thickness of the aquifer is 50 m.

The hydraulic operation of the system is kept simple. Water is pumped out of the aquifer at the production well where it is passed through a heat exchanger. Reinjection of the cooler water takes place through an injection well located at a distance, d, from the production well.

Water entering the aquifer at the injection well flows at a rate, v(d), toward the production well. Heat is transferred by forced and natural convective and conductive processes. Cooler water flowing in the direction of the production well gains heat due to the higher ambient temperatures in conjunction with the heat flow through the system. The surrounding temperatures are lowered due to the loss of heat to the cooler injected fluid.

The point in time when the injected water lowers the temperature at the production well by 1°C is called "breakthrough." Breakthrough depends upon the well spacing and the pumping-injection rate. At some distance, d, and some pumping-injection rate, Q(p), the temperature of the injected water will recover totally, and thus inhibit breakthrough. In this special case, the geothermal resource is classified as a <u>totally renewable</u> resource.

The system is located at a depth centered around 1375 m. The system is artesian with an initial total hydraulic head producing a slight flow at the surface in a well tapping the reservoir. The overlying sediments are assumed to have an average geothermal gradient of 40°C/km, which results in a temperature of the system of 55°C. The heat flow entering the system depends upon the thermal conductivity and thermal gradient and is about 75 mW/m^2.

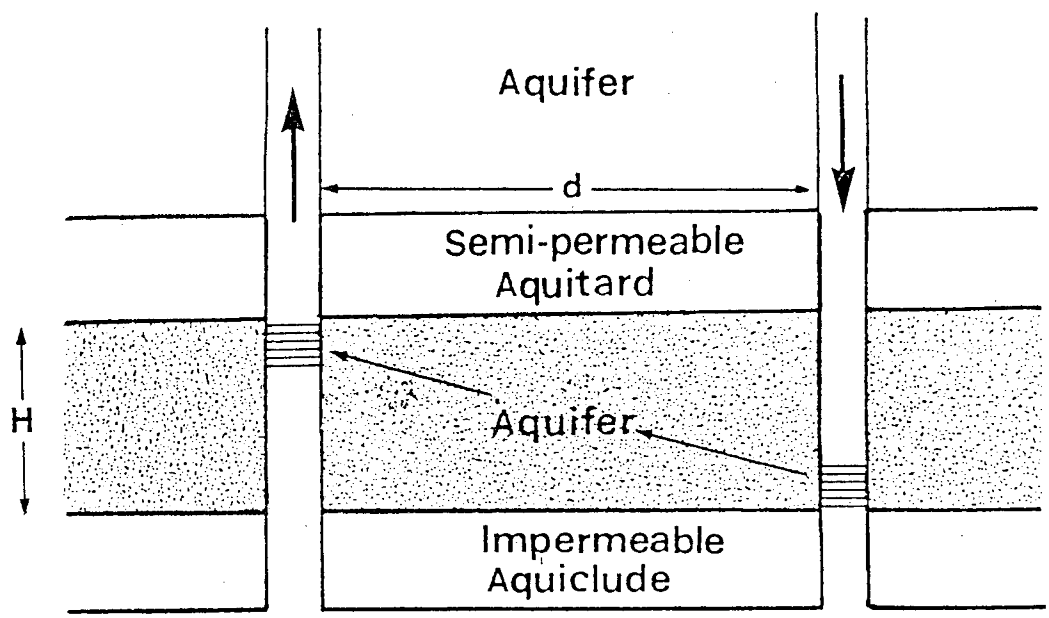


Figure 2. Schematic diagram of the doublet geothermal system used in the simulations. Hatched region shows perforated (pumping-injection) zones.

The amount of energy extracted from a geothermal system using a heat pump depends upon the thermal output and the pumping rate. Energy output in Btu's per hour is calculated by

$$\text{Btu's/hour} = Q_p \times 907.2 \times \Delta T \tag{1}$$

from Paddison et al. (1978). Q(p) is in gallons/minute.

A geothermal system which is pumped at 500 gpm, has an output temperature of 55°C, and an input temperature of 43°C (ΔT = 12°C) produces 5,443,200 Btu's per hour. The amount of energy extracted from the geothermal water could be raised considerably by inputting enough electrical energy to power a special type of heat pump called a temperature amplifier (Neiss, 1979).

Description and Development of Simulation Model

A general simulation model was constructed to simulate the pressure and temperature fields in a "typical" Atlantic Coastal Plain geothermal system. The integrated finite difference method was used to solve the mass and heat flow equations. The model makes use of the symmetry present in a doublet system by modeling one-half of the whole system.

Mesh Design

The model consists of a three-dimensional layered mesh (Fig. 3) containing 1342 three-dimensional elements each defined by a nodal point. The nodal elements are interconnected via 4172 element interfaces. Temperatures and pressures are obtained at the nodal points during the simulations.

The mesh design subdivides the geothermal system into three hydrologic units based upon the geology. The upper and lower hydrologic units each consist of one grid layer and have a vertical thickness of 25 m. The upper grid layer represents a semi-permeable unit and the lower grid layer a confining unit. The main reservoir is represented by the central five grid layers, each grid layer having a thickness of 10 m.

The design utilizes the natural convection present within the geothermal system due to water density and viscosity variations with temperature (Lippman et al., 1977). Thus, the production well is located in the upper portion of

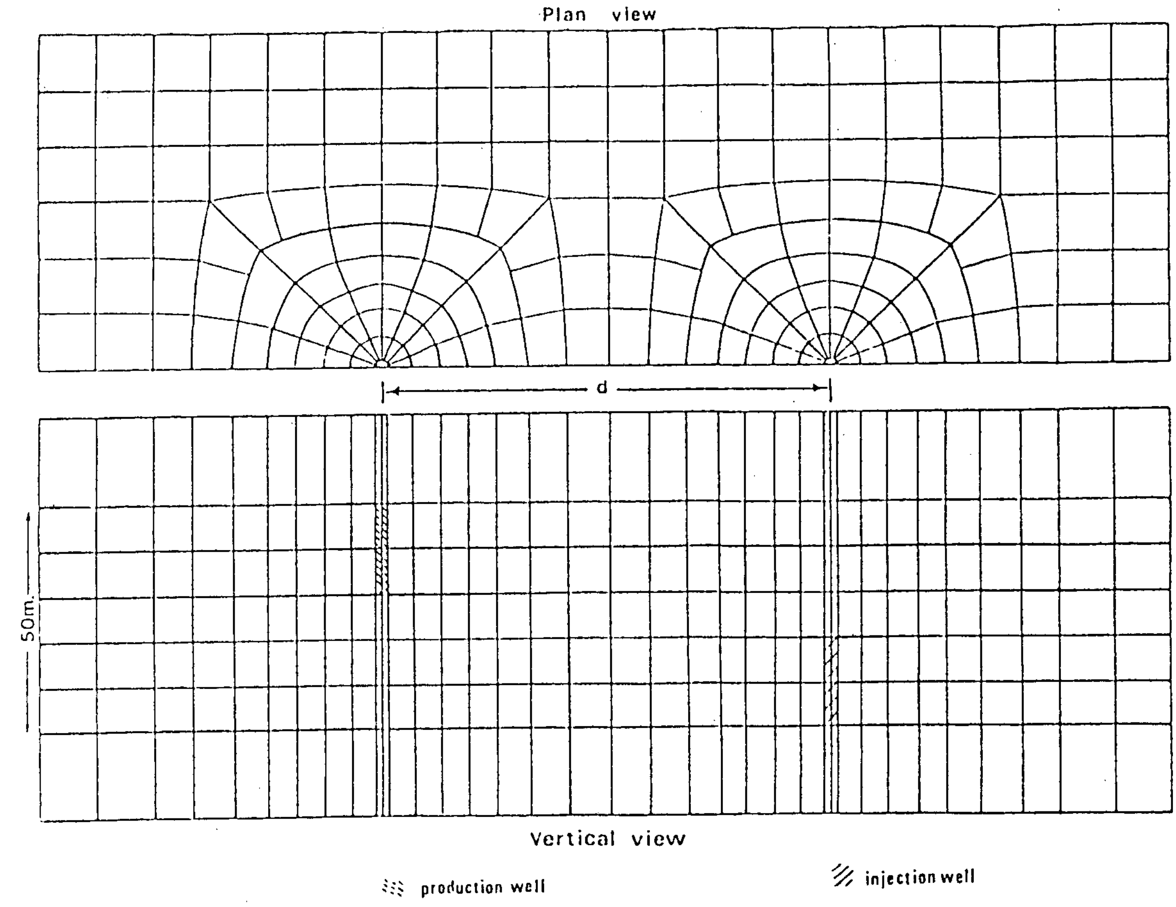


Figure 3. Mesh design used to simulate a typical Atlantic Coastal Plain geothermal system. Distance, d, represents the desired well spacing.

the reservoir with the injection well located in the lower portion of the reservoir.

The horizontal dimensions are scaled according to the horizontal distance, d, between the production and injection wells. Well spacings of 100, 250, 500, and 1000 m are tested in the simulations.

Fluid and Material Properties

Fluid and material properties used in the simulation model are chosen to describe a "typical" Atlantic Coastal Plain geothermal system. Values for these properties are obtained from the Crisfield geothermal test well (Dashevsky and McClung, 1979) and from literature pertaining to deeper wells in the Atlantic Coastal Plain (Brown and Silvey, 1972; Cedarstrom, 1945; U.S. Geological Survey, 1967).

Material properties used in the simulation are given in Table 1. The model consists of a caprock, a bedrock, and an aquifer defining an aquitard, an aquiclude, and a reservoir, respectively. Permeabilities are varied in the test simulations from 9.862×10^{-15} to 9.862×10^{-13} m^2 (10 to 1000 millidarcys).

The fluid properties (Table 2) are based upon a total dissolved solid content of 15,000 ppm. Fluid properties of the saline water are estimated from those of pure water using the equations developed by Wahl (1977) and are listed in Table 3.

The values for density, specific heat, and viscosity are functions of temperature and are calculated by the program. Density varies with temperature according to:

$$\rho = \rho_o\{1 - \gamma(T-T_o) - (T-T_o)^2\} \tag{2}$$

Variations with temperature in specific heat and viscosity are determined linearly from upper and lower limits supplied as inputs to program CCC.

Initial and Boundary Conditions

The initial and boundary conditions are determined using geothermal and hydrologic data obtained from the Crisfield well site (Dashevsky and McClung, 1979; Hartsock, 1979). The conditions are based upon the system being located at an average depth of 1.375 km. Initial values for fluid pressures and temperature are shown in Figure 4.

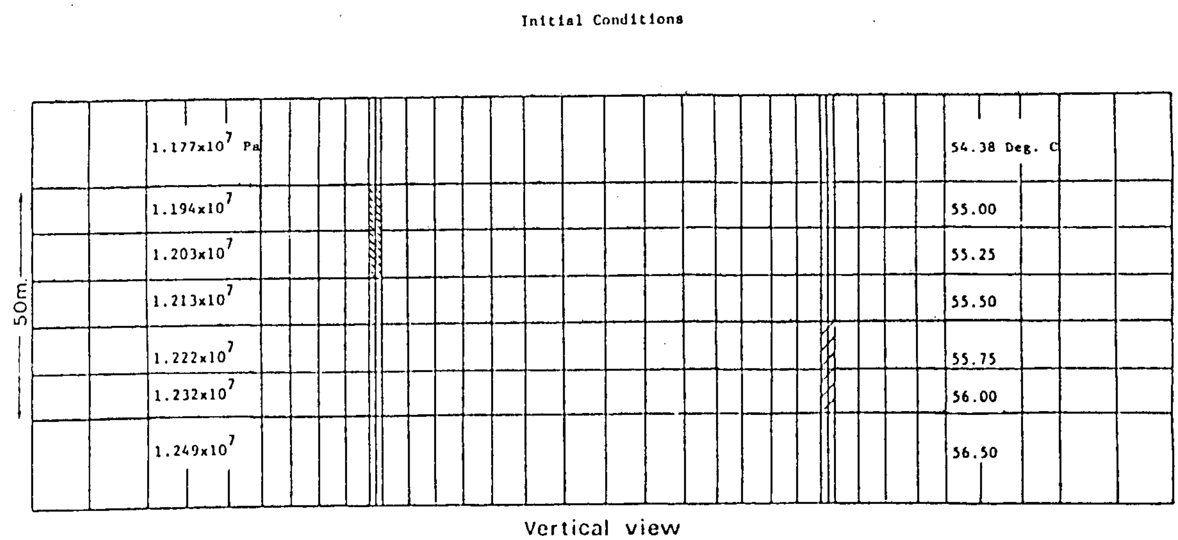


Figure 4. Inital pressure and temperature conditions used in the simulations. Values shown are input at the nodal points throughout the entire length of the grid layer. Nodal points are located in the center of the nodal elements.

Fluid pressures are chosen to simulate an artesian system. Initially, water is assumed to be under sufficient pressure to cause the piezometric surface to be located at ground level in a well tapping the aquifer. Initial pressure gradients describe a system where water velocities are zero because the value of hydraulic head is constant throughout the system. The initial temperature at the upper boundary is determined using a geothermal gradient of 40°C/km for the overlying sediments. The initial temperatures within the system are determined using the thermal gradients listed in Table 1. The initial temperature gradients are chosen in conjunction with the thermal conductivities of each hydrologic unit to produce a heat flow of 75.38 mW/m^2 (1.8×10^6 cal/cm^2-sec).

The criterion used to determine the location of the upper and lower boundary nodes relative to the rest of the system is that no more than a 10% change in heat flow should take place through the lower boundary-system interface. This criterion was satisfied in all simulations with the greatest change in heat flow being +7.2%. The boundaries enclosing all sides of the system are placed at a distance two times the distance between the production and injection wells. The boundary conditions used are:

a) the upper and lower boundaries are constant pressure (11648852 and 12608247 Pa, respectively) and constant temperature (53.875 and 56.875°C, respectively) boundaries;

b) the vertical boundaries along the caprock and bedrock are adiabatic and impermeable;

c) the reservoir boundaries along the vertical sides perpendicular to distance, d (well spacing), are constant pressure and temperature boundaries equal to the initial conditions defined for each respective layer;

d) the reservoir boundary along the vertical sides parallel to distance, d, are impermeable and isothermal with values defined by the initial conditions.

The upper constant pressure boundary simulates an overlying system of aquifers and serves as the source for recharge by leakage through the semi-confining unit. Leakage is induced only when pumping induces a hydraulic gradient across this interface. The constant pressure boundaries along the vertical sides of the reservoir perpendicular to distance, d, act as a line source or sink depending upon the hydraulic gradients across these interfaces. There is no flow through the lower boundary-system interface because of the zero permeability of the material.

Discussion of the Simulation Procedure

Well spacings of 100, 250, 500, and 1000 m are compared. Each individual well spacing is modeled using intrinsic permeabilities of 9.862 X 10^{-15}, 9.862 X 10^{-14}, and 9.862 X 10^{-13} m^2 (10,100 and 1000 millidarcys, respectively). Pumping and injection rates of 6.31 X 10^{-3} and 3.15 X 10^{-2} m^3/sec (100 and 500 gallons/minute) are considered. The water is injected into the system at a temperature of 43.5°C. Individual doublet systems are pumped for 15 years or until thermal and fluid flow steady-state conditions are reached.

This procedure is used to investigate the relationship between the temperature at the production well (energy output) and well spacing. The results produced by this procedure are also used to evaluate, in detail, a "typical" Atlantic Coastal Plain geothermal system by:

a) analyzing the temperature field to evaluate the thermal lifetime of each individual doublet system;

b) analyzing the pressure field to evaluate the ability of each system to supply adequate amounts of water;

c) analyzing the response of the system to different permeabilities; and

d) analyzing the pressure and temperature fields under relatively low and high pumping-injection stresses.

A final group of simulations considers the effect of resting on the system. A period of rest is defined as an interval of time in which the system is allowed to recover from previously-applied pumping-injection stresses. To mimic heating demands during the winter and summer seasons, six-month periods of production-injection are alternated with six-month periods of rest over a five-year interval. One well spacing of 100 m and permeabilities of 100 and 1000 md are considered. A pumping-injection stress of 100 gpm with an injection temperature of 43.5°C is applied.

Results and Discussion

Results of the simulations are displayed graphically by contour and time versus temperature in Laczniak (1980). Selected results from Laczniak are reproduced here.

Analysis of the Pressure Field

The pressure field was assumed to have reached steady-state when the change was "small," i.e., less than 2.5% of the maximum allowed pressure change over a specified number of consecutive time steps (Laczniak, 1980). Figure 5 shows a map view of the final pressure distribution through the center of the reservoir, between the grid layers containing the source and sink wells for a permeability of 100 md, a pumping stress of 100 gpm, and dipole separations of 100, 250, 500, and 1000 m. Figure 6 shows the same conditions except that the pumping stress is raised to 500 gpm. Results for permeabilities of 1000 md can be found in Laczniak (1980), but such high permeabilities are not representative of deep Coastal Plain aquifers. As expected, an increase in the pumping-injection rate or a decrease in the permeability produces a greater pressure differential within the system.

A cross-sectional view of the hydraulic head for all four well spacings is shown in Figure 7 for a permeability of 100 md and a pumping stress of 500 gpm. The influence of the injection well on the production well is displayed by the pattern of the contour lines of equal hydraulic head. The effect is less pronounced in the large separations where there is a geater independence between the production and injection wells.

Analysis of the Temperature Field

Map views of the steady state thermal distribution through the center section of the reservoir are shown in Figures 8 and 9 for a permeability of 100 md and pumping-injection rates of 100 and 500 gpm and a well spacing of 1000 m.

Figure 10 shows temperature versus time plots comparing output temperatures (temperature at the production well) in four well spacings for a permeability of 100 md and pumping-injection rates of 100 and 500 gpm. The results clearly illustrate the effect of well spacing on the temperature distribution. The general trend is an increase in the output temperature with an increase in the distance bewteen the production and injection wells. The amount of energy ex-

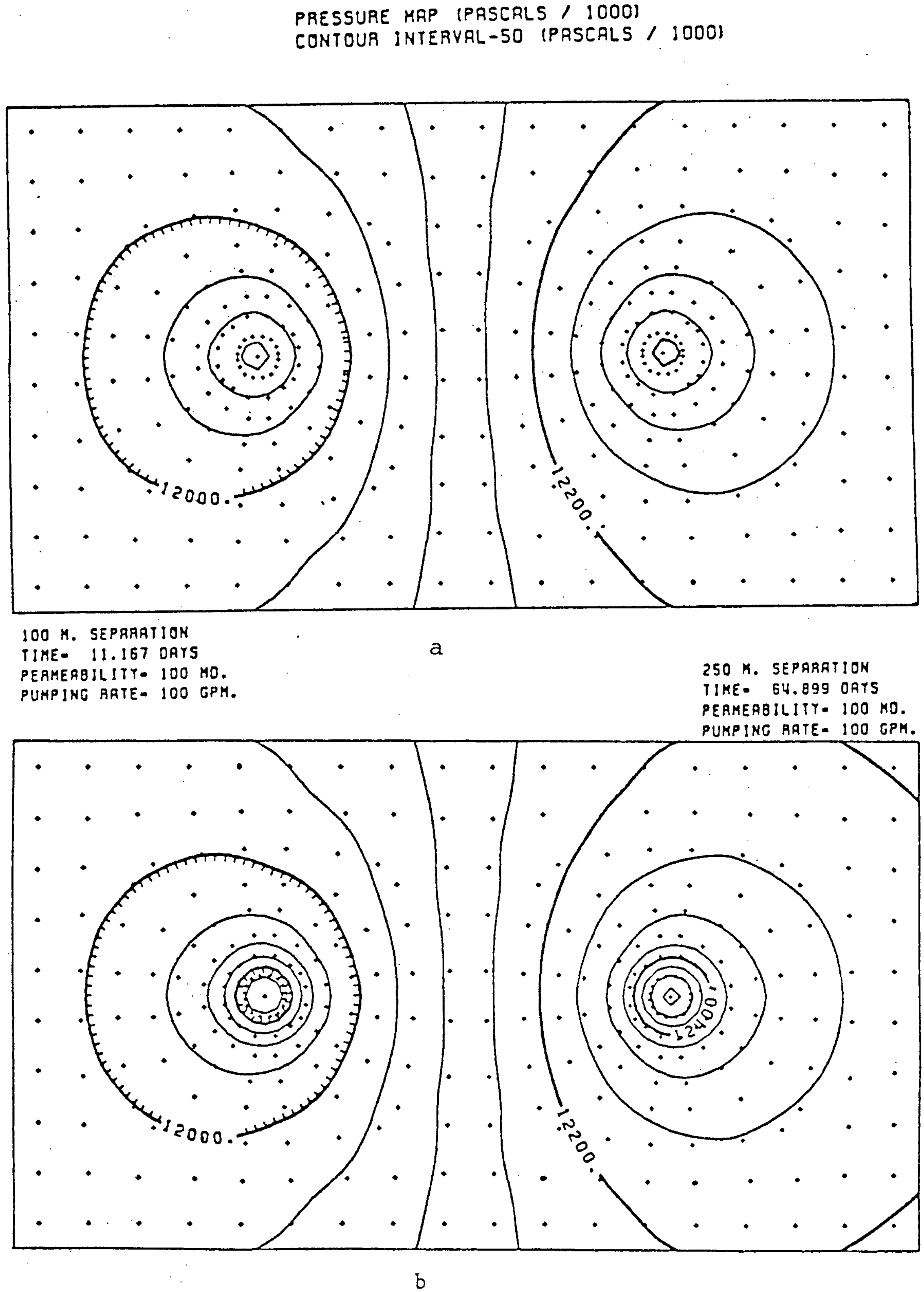


Figure 5. Pressure distribution map.

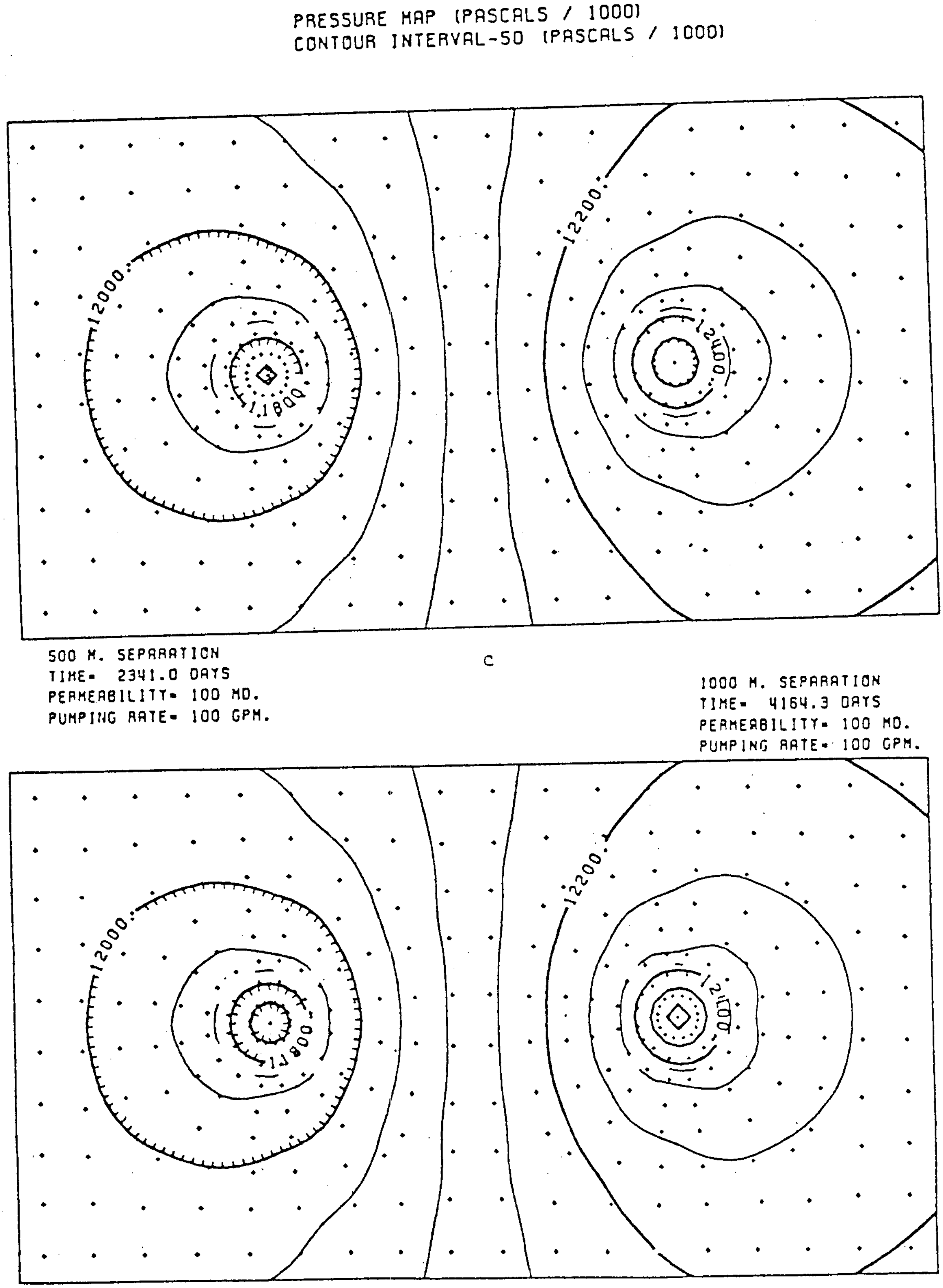
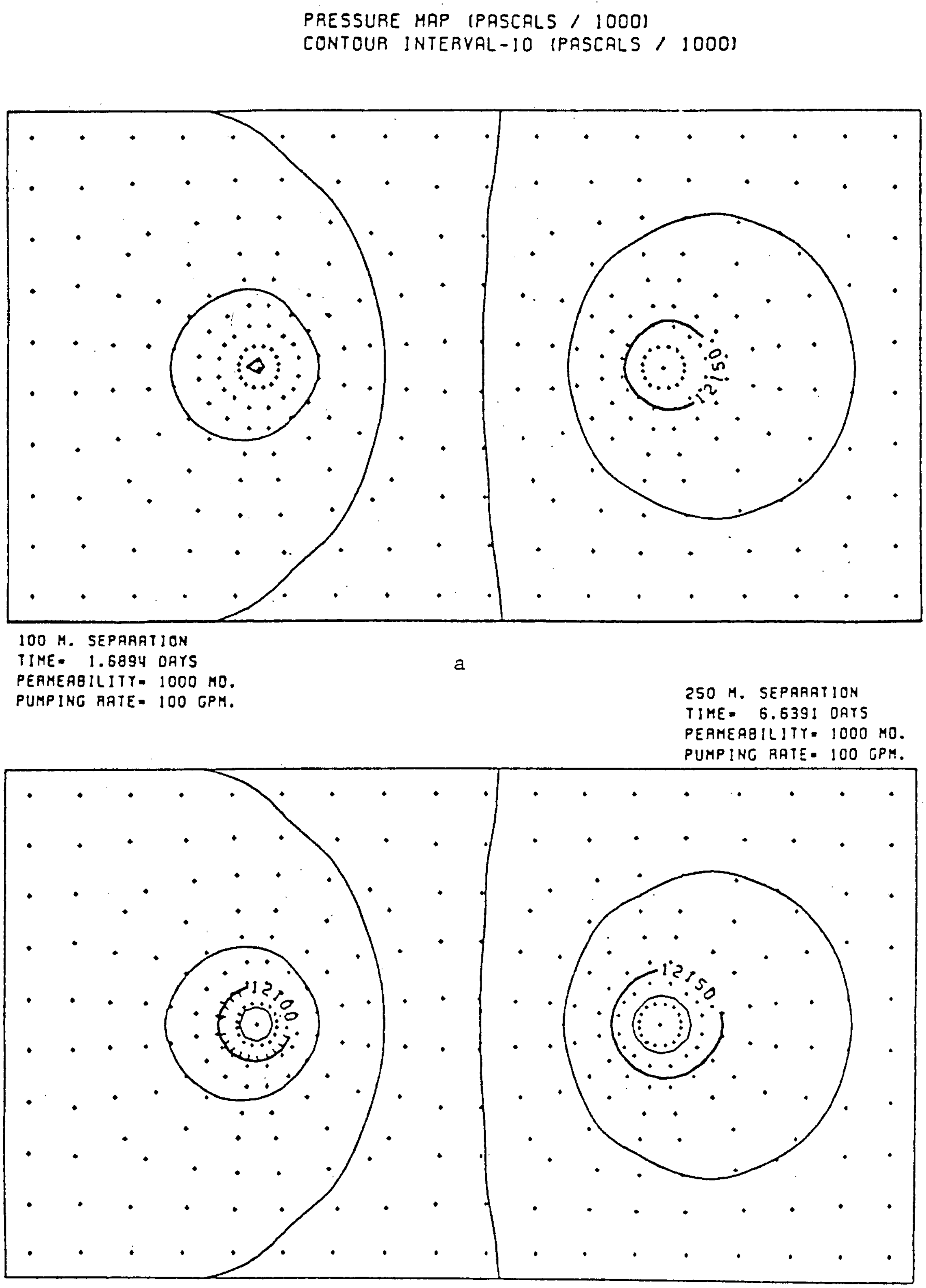
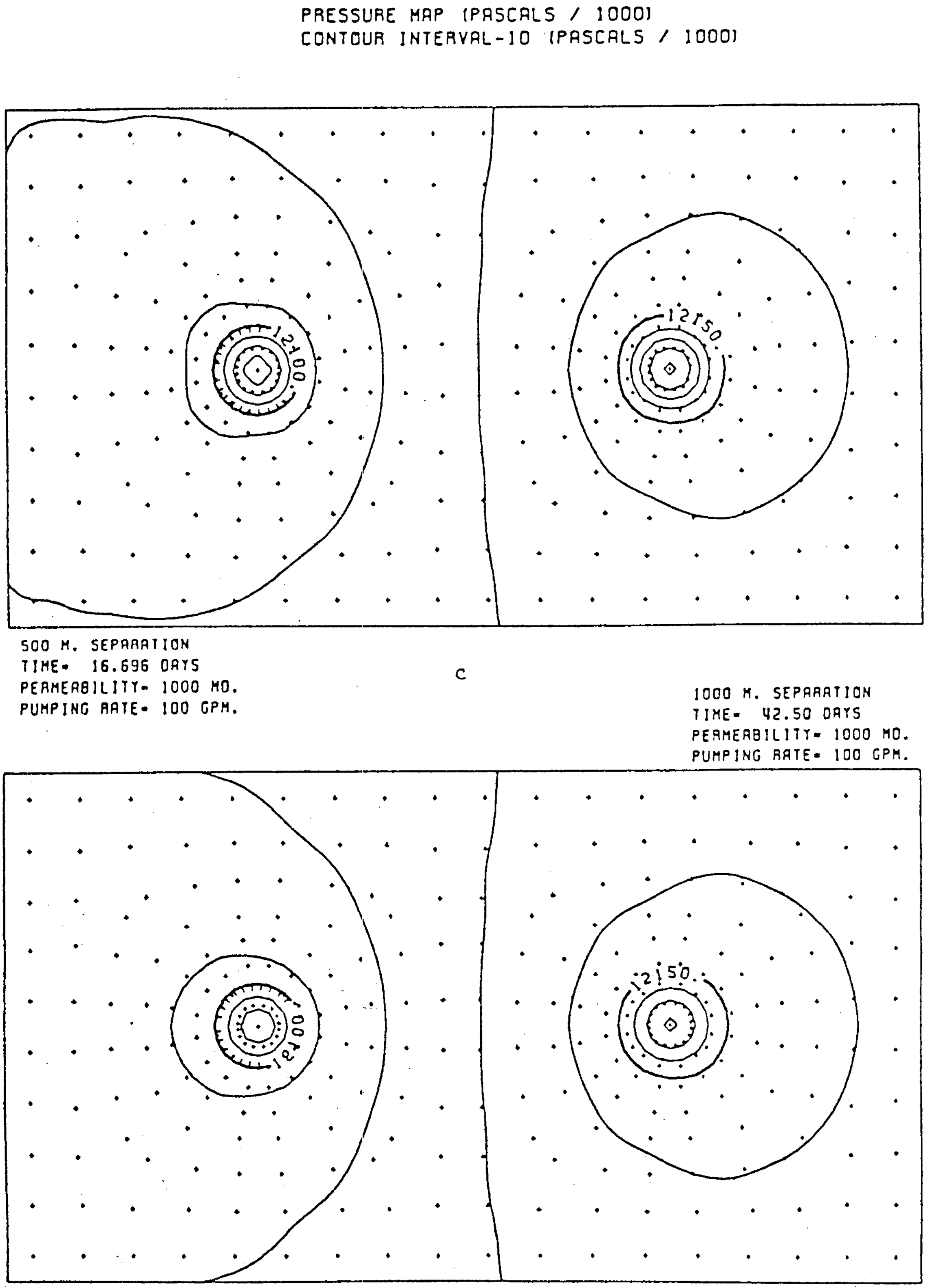


Figure 5 (continued).



b

Figure 6. Pressure distribution map as in Figure 5 except the permeability is raised to 1000 md.



d

Figure 6 (continued).

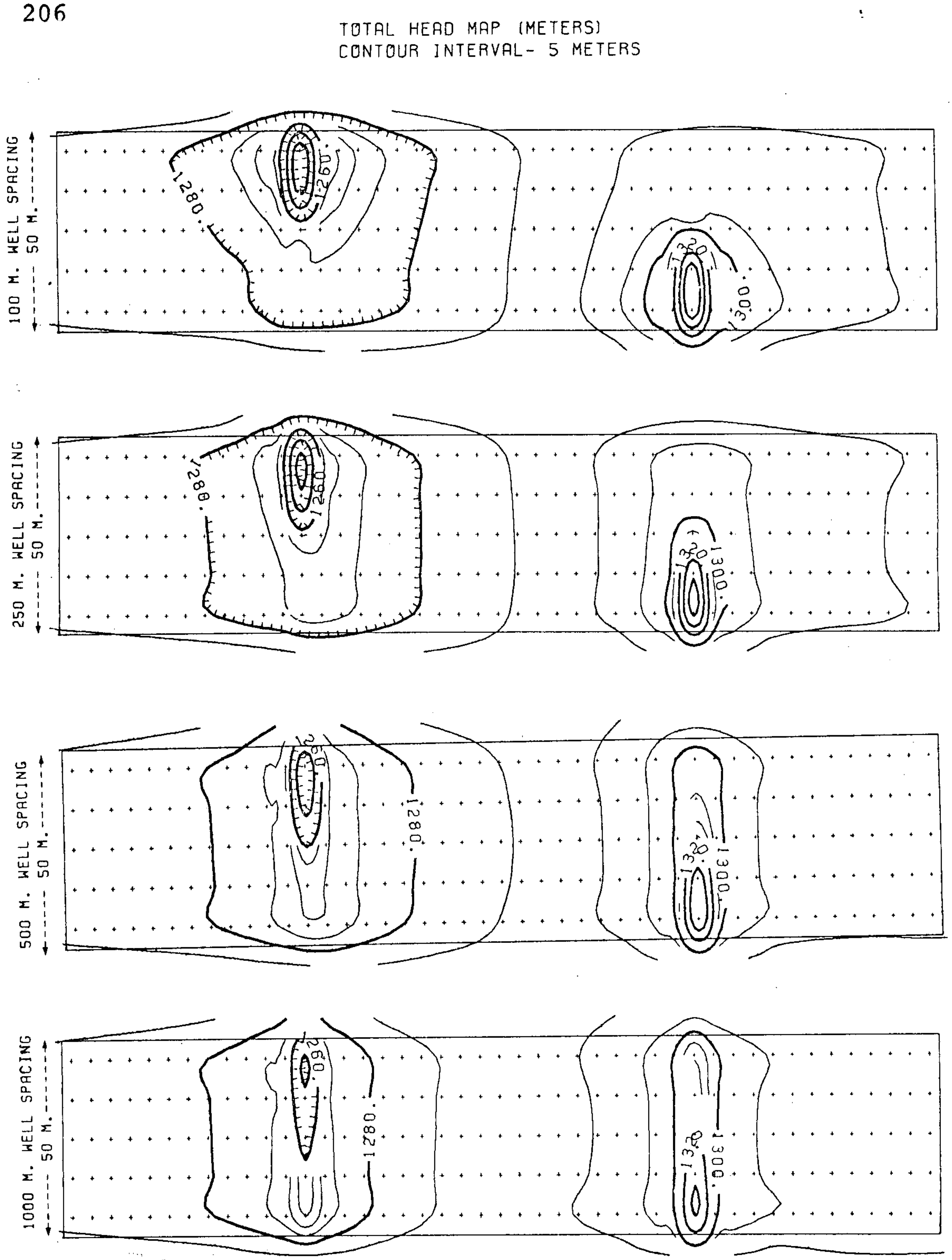


Figure 7. Cross-sectional view of the total head distribution for various well spacings in a system with a permeability of 1000 md and a pumping-injection stress of 500 gpm. Crosses indicate location of nodal points and the values interpolated between the nodal points. Typical flow paths are shown by stream lines (solid black lines with arrows) in the upper and lower sketches.

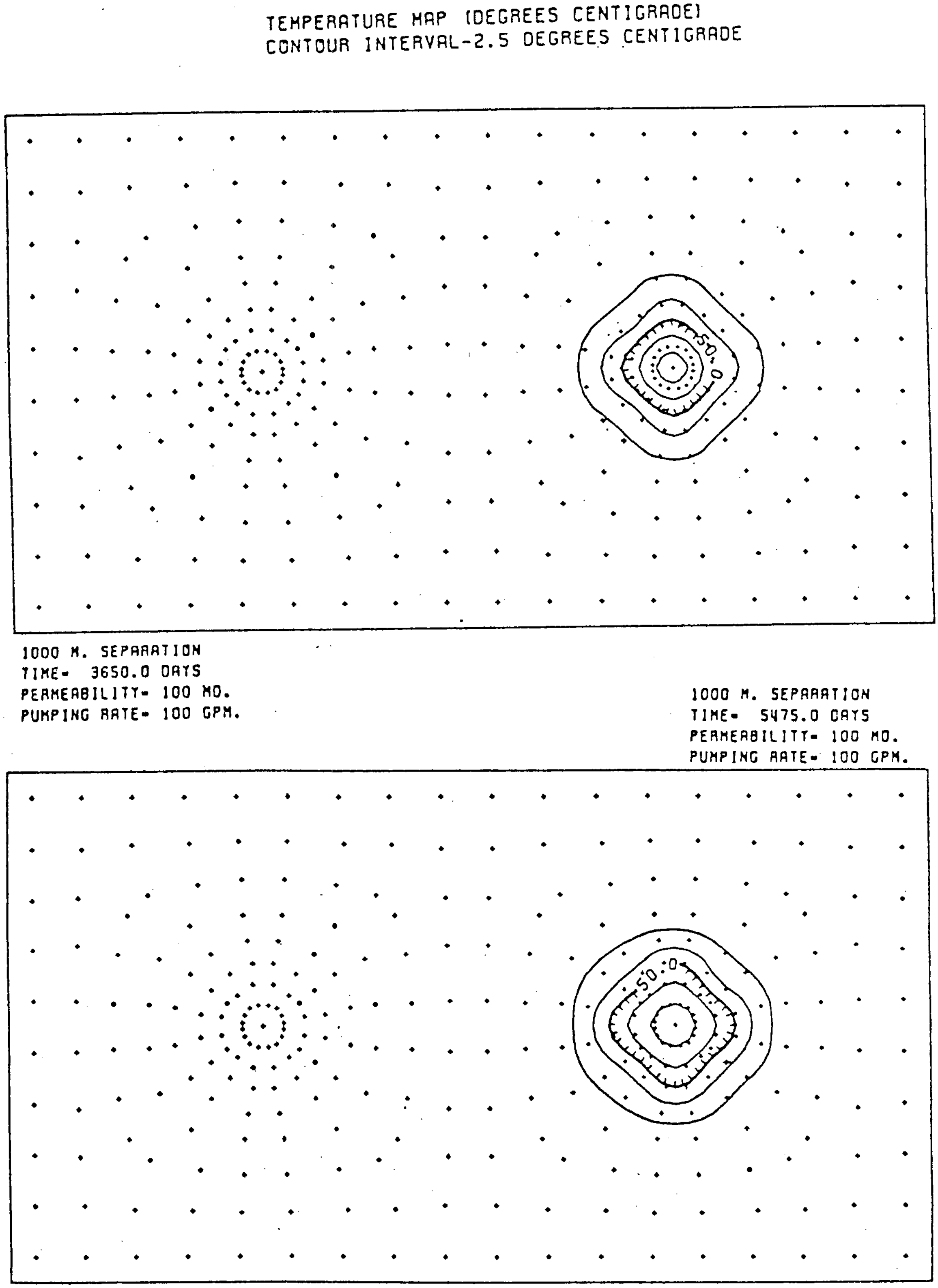


Figure 8. Temperature distribution map for a 1000 m well spacing at 10 and 15 yrs, upper and lower plots respectively.

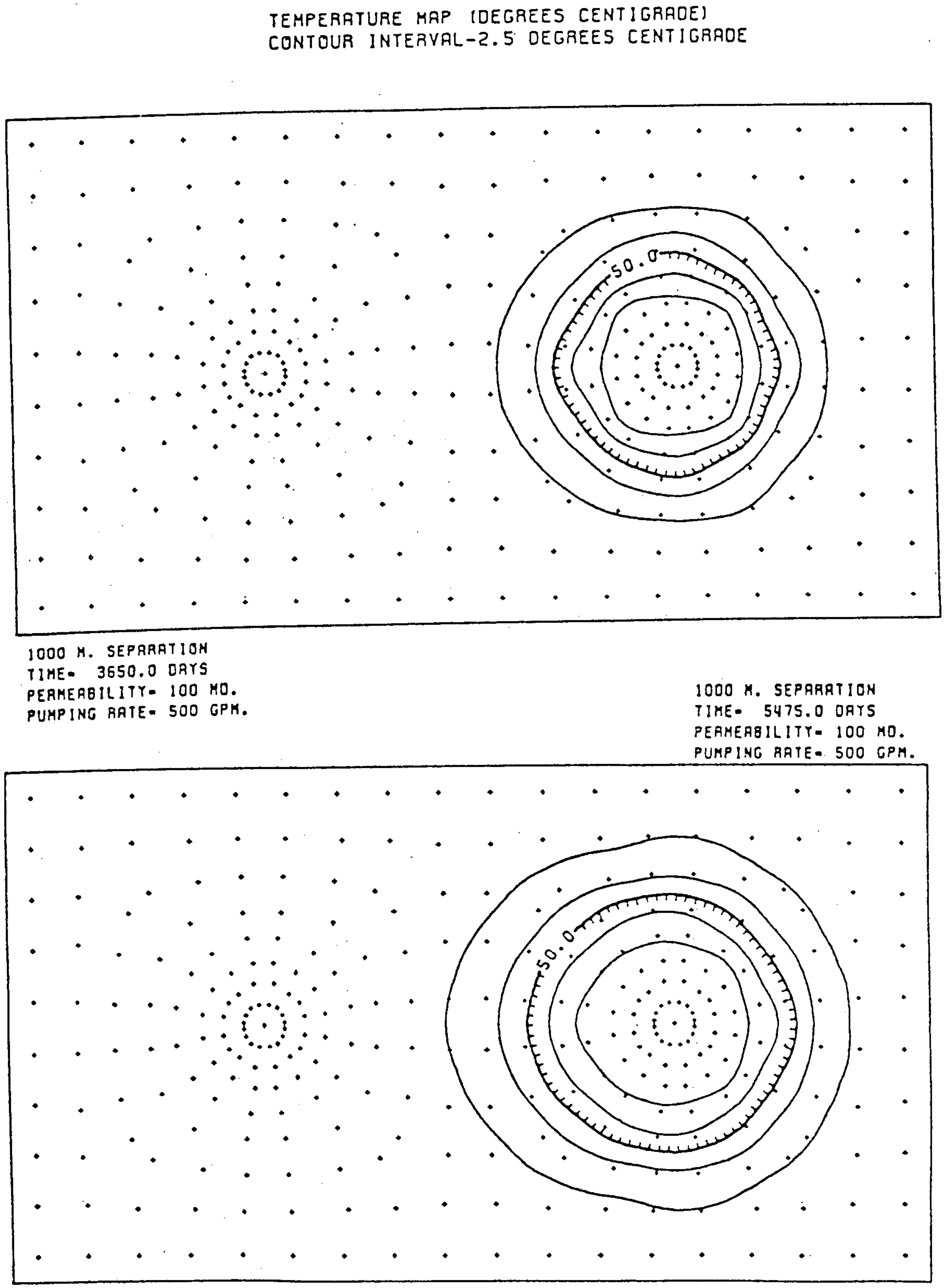


Figure 9. Temperature distribution map for a 1000 m well spacing at 10 and 15 yrs, upper and lower plots respectively.

tracted from the system, assuming no pipe loss, is calculated using equation (1). The results are listed in Table 4. For all pumping-injection rates and permeabilities, breakthrough in the 100-m spacing occurs rapidly (less than two years), while in the 1000-m separation breakthrough is altogether absent.

Differences in the pumping-injection rate also affect the temperature distribution and the time of breakthrough within the system. As expected, a greater pumping-injection rate lessens the time it takes for breakthrough to occur. The result is a decrease in the temperature differential between the pumping and injection wells.

The response of the temperature field to different permeabilities is shown through a comparison of the various temperature plots (Laczniak, 1980). A lowering of the permeability by one order of magnitude results in only a small gain in the final output energy (Table 4). This independence between the temperature field and the permeability is understood by analyzing heat transfer in conjunction with Darcy's law. Heat is transferred by conduction and convection. Conduction depends upon the physical conduction of the materials which are held constant throughout the simulations. Convection is controlled by the fluid velocity field. Darcy's law states that the velocity is proportional to both the permeability and the hydraulic gradient. For a given pumping stress, increasing the permeability decreases the hydraulic gradient; therefore the change in the magnitude of the velocity is small. The combined effect of conduction and convection results in a similar net heat transfer among systems in which all parameters except permeability are kept constant.

The placement of the pumping and injection well is reinforced by the results of the modeling. Although the pumping well is located in the coolest portion of the reservoir, warmer temperatures migrate quickly to the perforated zone of the production well. This phenomenon is displayed by the initial increase in the temperature at the production well (Laczniak, 1980, Figs. 15-18) and is due to recharge of warmer waters from the lower portions of the reservoir.

Analysis of the Effect of Resting on the System

Aside from a simple doubling of the life span, subjecting the system to alternating six-month periods of continuous pumping and resting produces only neglible changes in the overall temperatures at the injection and production wells. Laczniak (1980) compared the temperatures at the injection well in a rested system to the temperature at the

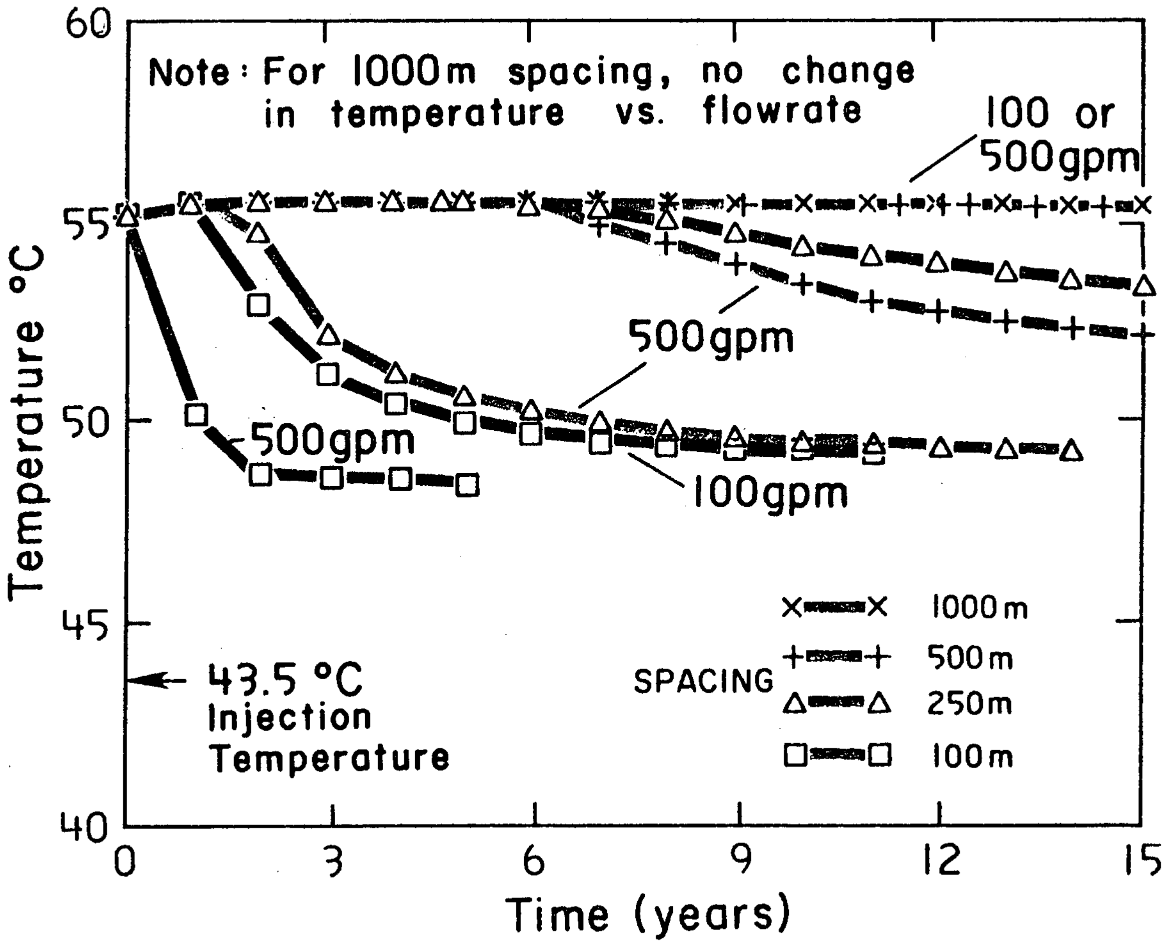


Figure 10. Plot of production temperature versus time for various well spacings in a system with a permeability of 100 md and pumping-injection stresses of 100 and 500 gpm. Production temperature is the average of the temperatures recorded at both pumping elements. Note the occurrence of breakthrough, the point in time when the temperature begins decrease.

injection well in a continuously pumped system over a period of five years. The recovery of heat from terrestrial heat flow at the injection well over the six-month rest period was minimal. Varying the permeability has little effect on the system.

Conclusions

Engineering data about fluid transmission, drilling, pumping equipment, and land acquisition costs are needed for a complete evaluation of the economics of the system. Excellent studies of this type have been carried out by the Applied Physics Laboratory of the Johns Hopkins University (1981). The results of the numerical modeling quantify, without doubt, the existence of potentially useful geothermal resources beneath the Atlantic Coastal Plain. Heat from a groundwater source (the aquifers beneath the Atlantic Coastal Plain) is adequate to run heat pumps at a high coefficient of performance (Neiss, 1979). A doublet system with direct injection back into the reservoir is shown by the numerical simulations to be a feasible method of extracting heat in the low-temperature liquid-dominated geothermal systems of the Atlantic Coastal Plain. The doublet system with a spacing, d, of 1000 m pumped at 500 gpm produces approximately 5.5 million Btu's per hour. Assuming a value of 70,000 Btu's per hour for energy expenditure to heat an average insulated home, the system would support over 75 households. If pumped for only six months a year, the thermal lifetime of the system is shown to be at least 30 years.

The optimum geologic environment for the implementation of a groundwater geothermal system is one in which the highest temperatures are encountered at the shallowest depths. Throughout much of the Atlantic Coastal Plain there exist areas where isotherms are warped upward over radioactive heat-producing granites within the crystalline basement beneath the Coastal Plain sediments (Costain et al., 1980). In the future, aquifers over these "radiogenic" sources may be located. At such locations, drilling costs would be reduced and the use of lower energy output systems might be economically justified. Systems with close well spacings and/or low pumping-injection rates could then be implemented depending upon the application.

Modifications of the injection scheme intended to increase the thermal lifetime of the system; for example, injection into an independent reservoir or injection into a zone separated by a low permeability lens should be carefully analyzed before being implemented. In the low permeability deeper reservoirs of the Atlantic Coastal Plain, the influence of the injection well is shown by this study to be necessary in order to maintain adequate fluid pressures within the system to prevent aquifer collapse. This influence of the injection well is required, and must be taken into account in the design of a successful heat extraction system.

Reinjection of the thermally spent cooler water back into an independent or semi-independent reservoir may cause adverse chemical or physical reactions within that reservoir. In this case, extensive modeling of the total system should be performed. Such modeling must incorporate not only the equations of heat and mass flow, but must simultaneously solve for hydrodynamic dispersion and heat production due to the decomposition of reactants. The sensitivity of the reservoir water to the injected water and the resulting chemical reactions should be completely understood and incorporated into the model.

In-depth studies of each proposed Atlantic Coastal Plain geothermal system are strongly recommended before any application. Such studies should include extensive hydrologic and heat flow data collection. The data should then be used in numerical models to analyze various system designs. Program CCC is applicable and appropriate for the evaluation of the low-temperature geothermal systems of the Atlantic Coastal Plain. If financial limitations inhibit adequate data collection or if computer funds are limited, it is hoped that the results of the present study will contribute to the development and design of reliable heat extraction systems beneath the Atlantic Coastal Plain.

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Table 1. Material properties used to simulate a "typical" Atlantic Coastal Plain geothermal system.

PROPERTIES*	AQUITARD	AQUIFER	AQUICLUDE
Heat capacity			
($J\ kg^{-1}\ ^{0}C^{-1}$)	0.921×10^{3}	0.103×10^{4}	0.803×10^{3}
(cgs units)	0.220	0.246	0.192
Density			
($kg\ m^{3}$)	2.60×10^{3}	2.65×10^{3}	2.60×10^{3}
(cgs units)	2.60	2.65	2.60
Thermal conductivity			
($J/m\text{-}day\ ^{0}C$)	1.63×10^{5}	2.60×10^{5}	2.17×10^{5}
(cgs units)	4.50×10^{3}	7.20×10^{-3}	6.00×10^{-3}
Temperature gradient			
($^{0}C/Km$)	40	25	30
(cgs units)	0.004	0.00025	0.0003
Permeability			
(m^{2})	$K \times 10^{-4}$	K	$K \times 10^{-25}$
(cgs units)	K	$K \times 10^{4}$	$K \times 10^{-21}$
Porosity	0.20	0.15	0.20
Specific storage			
(m^{-1})	3.00×10^{-5}	1.50×10^{-4}	4.00×10^{-5}
(cgs units)	3.00×10^{-7}	1.50×10^{-6}	4.00×10^{-7}

*The first entry for each property is in the units used by the program and the second is in cgs units.

Table 2. Fluid properties used to simulate a "typical" Atlantic Coastal Plain geothermal system. The first entry for each property is in the units used by the program and the second is in cgs units.

FLUID PROPERTIES	
Compressibility	
(m^2N^{-1})	5.000×10^{-10}
(cgs units)	5.000×10^{-11}
Thermal expansivity	
($^0C^{-1}$)	3.170×10^{-4}
Reference density	
($kg\text{-}m^{-3}$)	1003.17
(cgs units)	1.00317
Reference temperature	
(^0C)	40
** Viscosity	
(Pa-sec)	6.750×10^{-4}
(cgs units)	6.75×10^{-3}
** Specific heat	
($J\text{-}kg^{-1}\ ^0C^{-1}$)	4.115×10^{3}
(cgs units)	0.983

**Value varies linearly with temperature.

Table 3. Equations approximating physical properties of brines using properties of pure water (Wahl, 1977).

Density: $\rho = \rho_w + .00073\ (w_t)$

Specific heat: $H_c = H_{c,w}(1 - w_t/100)$

Viscosity: $\mu = \mu_w(1 + .021w_t + .00027w_t{}^2)$

Table 4. Amount of energy extracted in Btu's per hour for the various well spacings in the simulated typical Atlantic Coastal Plain geothermal system.

	100	250	500	1000 m	spacings
k			Btu's/hr.		q
1000 md	521640.	869097.	1087732.	1088640.	100 gpm
100 md	522542.	903571.	1087732.	1088640.	100 gpm
1000 md	2204496.	2644488.	3982608.	5438664.	500 gpm
100 md	2281608.	2726136.	4150440.	5443220.	500 gpm