TEC-7

GEOTHERMAL RESOURCES OF VIRGINIA AND WEST VIRGINIA

Geological and Geophysical Environment of Thermal Springs

By

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With a

PETROGRAPHIC REPORT ON THREE ROCKS FROM

HIGHLAND AND BATH COUNTIES, VIRGINIA

By

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PART I: Text and Figures

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CONTENTS

Part I: Text and Figures

Contents continued..

SUMMARY

Seventy-two springs and groups of springs in the Appalachians of Virginia and West Virginia discharge thermal waters whose temperatures exceed 15°C (59°F), at least part of the year. Most of the springs issue from Ordovician and Silurian carbonate rocks in anticlinal valleys of the Valley and Ridge and the Appalachian Plateau physiographic provinces. The center of the thermal activity is located in Warm Springs anticline (Allegheny, Bath and Highland Counties, Virginia), where a chain of spas has utilized the waters for almost 200 years. Spout Spring at Hot Springs, Virginia, is the warmest, discharging at 41°C (106°F). Most of the springs are reported to resemble cold springs in chemical composition; however, some contain high concentrations of sulfates and lithium.

The thermal area overlies an estimated 9km of Paleozoic sediments laid down in the Appalachian eugeosyncline. Subsequent thrusting associated with uplift of Blue Ridge Precambrian rocks to the east has contorted the beds into a series of parallel NE-SW-trending folds undercut by low angle decollements and splay faults dipping toward the southeast. Whether or not the underlying basement has experienced folding and faulting has not yet been resolved.

A number of regional geophysical anomalies centered around the thermal area suggest a causal relationship with the springs. These include an unusual thickening of the crust to

60km (greatest in the U.S.), deduced from travel-time delays of seismic waves; broad gravity and magnetic lows; an absence of seismicity; and an extreme damping of response from distant earthquakes. In addition, geomorphic evidence points to an uplift of the Eocene erosion surface in the vicinity of the springs. The evidence is corroborated by the presence of outcropping volcanics of Eocene age (47 million years) at the north end of Warm Springs anticline. These are the youngest igneous rocks found east of the Mississippi. Dennison and Johnson (1971) conclude that a heated pluton underlies the area.

Examination of the consequences of crustal thickening lead to the deduction that, even under conditions of a typical low, Eastern-type geothermal gradient (13°C/km), molten granite or partially molten basalt should be present below crustal depths of 35 to 45km. Thus a broad lens of molten material 10 to 20km thick--the consequence of heating by a low geothermal gradient in an abnormally thick crust--is likely to underlie the thermal area. Such a pool would be compatible with the geophysical evidences and would account for the attenuation of seismic waves emerging from the Mohorovic discontinuity, since the major component of the earthquake signal is the shear waves; these do not pass through fluids.

Local upward migration of magma in the past would account for the presence of the volcanic outcrops. Such ascendants today would result in <u>localized</u> heating of the subsurface and deeply penetrating groundwater. Rising magmatic waters may also contribute to the temperature and composition of the springs. The likely seat of a geothermal reservoir is the buried paleokarst of the Ordovician Beekmantown formation.

A recommended program for determining the source of heat and location of possible geothermal reservoirs include: 1) the geochemical analysis of spring waters and travertines to determine temperatures at depth; 2) geological reconnaissance for fracture controls; 3) monitoring of microearthquakes associated with fault movement at depth; 4) heat-flow measurements; 5) exploration of Warm River Cave, an enterable thermal spring; and 6) gravity, electrical, ground-noise and geochemical surveys in the vicinity of the springs. The evidences presented and the potential Eastern market for geothermal power warrant these investigations.

INTRODUCTION

The numerous hot springs and geyser fields of Western United States have monopolized the attentions of persons exploring for geothermal resources. In many cases their exploration is directed to areas having the lowest population densities in the conterminous U.S., in the expectation that the power generated will find a distant market. In view of the remoteness of the coal-fueled power-generating facilities of Farmington, New Mexico, this hope is probably justified. Nevertheless, a well-known region of thermal springs adjacent to the vast market of the Eastern seaboard has evidently been ignored. These springs are localized in the Valley and Ridge and Appalachian Plateau physiographic provinces, extending along the border of Virginia and West Virginia. The reasons for the neglect are probably the relatively low water temperatures at the discharge points of the springs (less than 41°C (106°F)*; low heat-flow readings in Eastern U.S. (none have yet been made in the thermal area); and the century-old explanation that the waters are heated by the normal geothermal gradient at depth, without abnormal heat sources being involved.

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* In answer to the first point, we note that the <u>maximum</u> water temperature of the surface emanations from the Monte Amiata geothermal reservoir in Italy, which today produces power from dry steam, is 50°C (122°F), only 9° higher than the water at Hot Springs, Virginia.

The past decade of regional geophysical measurements over the Middle Appalachians, and the discovery of Eocene volcanism in Highland County, Virginia, lead to the deduction of an alternative source of heat to explain the thermal waters. This report analyzes the evidence accumulated to date and interpretes it in terms of potential geothermal reservoirs.

GEOLOGIC SKETCH

Physiography

The thermal springs of the central Appalachians occur in the Valley and Ridge province and to a lesser extent in the Appalachian Plateau to the west (Figure 1). The Valley and Ridge system extends generally northeast-southwest along the western border of Virginia and consists of parallel synclinal and anticlinal valleys between ridges of about 1200m (4000ft) elevation (Plate 1,2A). A maximum altitude of 1480m (4850ft) is attained in the Elkins Valley anticline of Pochontas County, West Virginia. Typical relief from ridge tops to canyon bottoms is about 600m (2000ft). The Valley and Ridge province merges into the Appalachian Plateau westward, where the undulations gradually decrease in amplitude and the bedding becomes essentially horizontal across central West Virginia. The Valley and Ridge province is terminated abruptly on the east by the uplifted Blue Ridge of Precambrian crystalline rocks separating it from the metamorphics of the Virginia Piedmont--a terrain of rolling hills.

Stratigraphy

Sedimentary rocks of the Valley and Ridge province range in age from the Cambrian Chilhowee group of the Shenandoah Valley, adjacent to the Blue Ridge, to the Pennsylvanian Pottsville series, in synclines of the Plateau

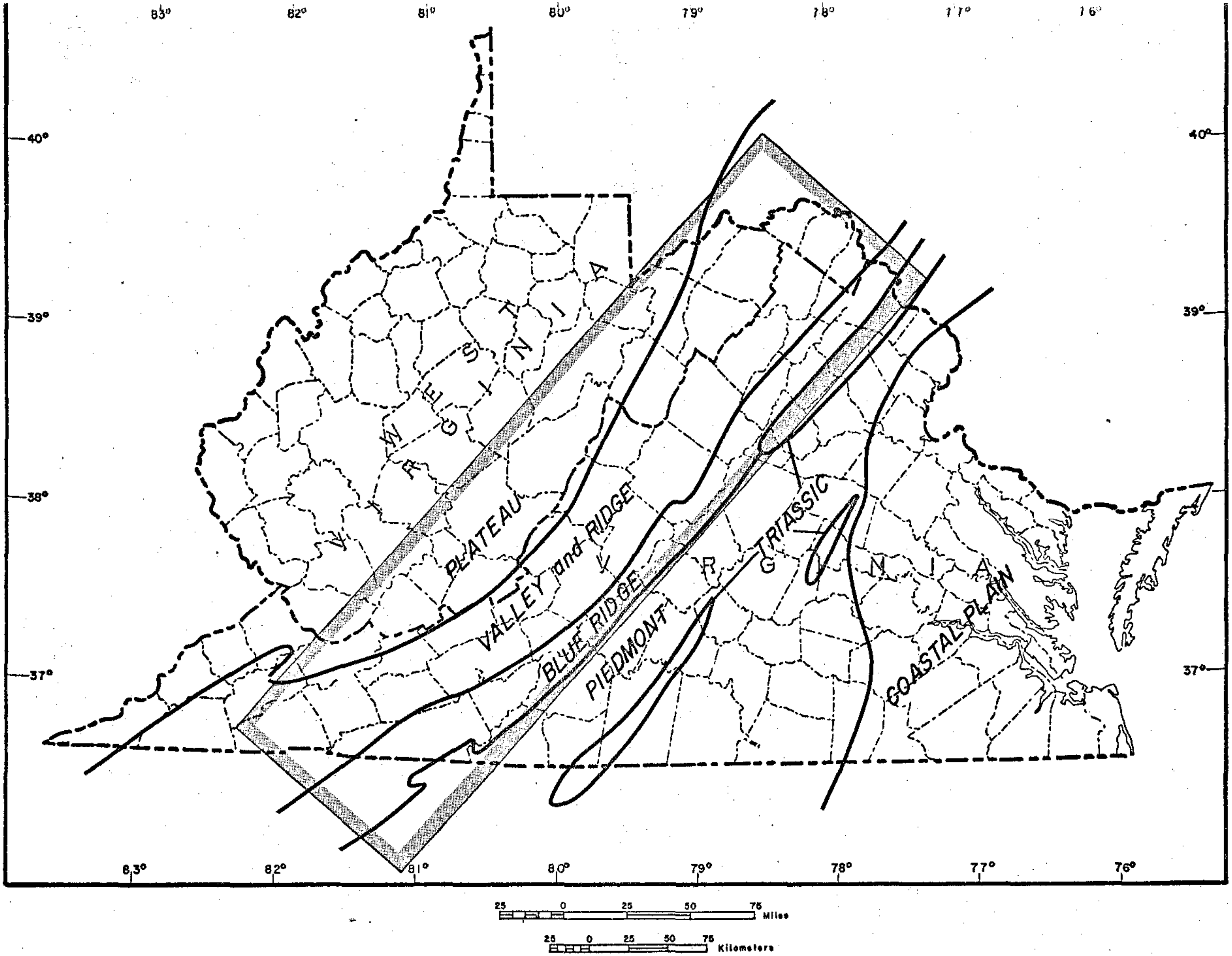


Figure 1. Physiographic provinces of Virginia and West Virginia, region of thermal springs outlined as a block.

of West Virginia (Plate 3,2A). Because the two provinces overlap, we consider the eastern portion of the Plateau as part of the Valley and Ridge. Table 1 (Gwinn, 1964) provides a generalized summary of the Central Appalachian sediments. In the vicinity of Hot Springs, Virginia, (essentially Highland, Bath and Allegheny Counties), rock strata range in age from Ordovician Beekmantown limestones and and dolomites (outcropping only along the axes of anticlinal valleys) to the Mississippian Pocono Formation appearing along synclinal axes. A detailed section through Highland County is given in Table 2.

The beds accumulated in the subsiding Appalachian eugeosyncline during the Paleozoic era. The Tertiary and Holocene regolith is made up of alluvium and colluvium derived from the local environment and adjacent Blue Ridge. Solution is a major erosional agent, and is the dominant one on the carbonates. Residuum from the Cambrian Tomstown dolomite has yielded economic deposits of iron and manganese in the Shenandoah Valley (Hack, 1965). Travertine deposits are associated with many of the springs.

Precambrian intrusive and metamorphic rocks make up the Blue Ridge and Piedmont to the east. Except for a thin layer of Tioga bentonite of Devonian age--believed derived from volcanism to the east (Dennison & Johnson, 1971)--igneous activity in the Valley and Ridge was confined to dikes and sills of Mesozoic and early Tertiary ages. The latter, represented by Eocene intrusives in Highland County, are discussed later in this report.

System	Lithic Unit	Approximate thickness (feet) West	East
Permian	Cyclic sandstone, limestone, mudstone, minor coal	0–800	0 (Eroded)
Pennsylvanian	Cyclic sandstone, limestone, mudstone, and coal	1400	2000+–3000+
Mississippian	Mauch Chunk redbeds, Greenbrier limestone, Pocono sandstone	500–700	1500–5000
Devonian	Catskill redbeds	0–200	1200–2000
	Devonian "shale," in descending order, Chemung sandstone-shale "facies," Portage shale "facies," Tully limestone, Hamilton shale	2000–3250	4500–6500
	Onondaga limestone (NW.), Huntersville chert (SE.) Oriskany sandstone Helderberg limestone and shale	100–600	750–1000
Silurian	Salina group (NW.). Tonoloway limestone, Wills Creek shale, and Bloomsburg redbed facies (SE.): salt, gypsum, shale and carbonates in Salina; evaporites wedge out east and west	1000–1700	900–1300
	Lockport dolomite (NW.), McKenzie-Rochester shale (SE.)	200–300	300–500
	Clinton group (NW.), Rose Hill formation (SE.) Tuscarora sandstone (SE.), Medina sandstone (NW.)	450–600	1200–1500
Ordovician	Upper: Queenston shale (N. and NW.), Juniata red shale and sandstone (SE.) Oswego sandstone (eastern Plateau only)	450–1050	1000–1800
	Martinsburg shale	700–1000	1000–2000
	Middle: carbonate rocks, including Trenton, Black River, and Chazy	900–1200	1200–1600
	Lower: Beekmantown dolomite	1000	3000–4500
Cambrian	Middle and Upper Cambrian carbonates (including Conococheague and Elbrook to the east)	1100–1600	4000–5500?
	Waynesboro Formation Tomstown-Shady dolomite Chilhowee group	Thickness in Plateau not determined by drilling; units outcrop only in Blue Ridge and Great Valley to the southeast; only Upper Cambrian rocks are present in western West Virginia and western Pennsylvania because of Cambrian onlap from the southeast.	
Precambrian	Catoctin greenstone		

Table 1. Generalized stratigraphy of the Central Appalachians (from GWINN, 1964).

Age		Name	Character	Generalized thickness in feet
Quaternary	Holocene	Alluvium	Sand to rounded boulders in stream valleys.	0-10
Quaternary	Pleistocene	Colluvium	Angular rubble to subrounded boulders on valley slopes and capping terraces.	0-20
Tertiary	Eocene	Felsic dikes and sills	Andesite, andesite porphyry, and trachyte.	
Triassic(?)		Mafic dikes and plugs	Basalt, olivine-basalt porphyry, and diatreme breccias.	
Mississippian		Pocono Formation	Chiefly yellowish-gray sandstone with siltstone and shale.	400
Devonian		Hampshire Formation	Nonmarine, reddish-gray siltstone, sandstone, and shale with some yellowish-gray sandstone.	1000
		Foreknobs formation (Dennison, 1970)	Marine, light-olive gray sandstone and siltstone with some reddish streaks and conglomeratic layers.	2300
		Brallier Formation	Marine shale and turbidite siltstone; weathers light-olive gray.	1500
		Millboro Shale	Black, mostly platy, shale.	230
		Tioga Bentonite	Brownish-gray shale with tuff layers as coarse as sand-size.	30
		Needmore Formation	Dark- to light-olive gray, chippy shale; tongue from west of Huntersville chert near middle.	100
		Oriskany Formation	Coarse, fossiliferous, marine sandstone: important as gas reservoir.	130
		Licking Creek Limestone	Limestone; dark and cherty in lower part.	195
		New Scotland Formation	Limy chert and cherty limestone.	25

Age	Name		Character	Generalized thickness in feet
Devonian	New Creek Limestone (Bowen, 1967; formerly Coeymans Limestone.)		Coarsely crystalline and crinoidal limestone.	50
	Keyser Fm.	Upper member	Fossiliferous limestone.	120
Silurian	Keyser Fm.	Big Mountain Shale member	Dark, calcitic shale.	28
	Keyser Fm.	Lower member	Fossiliferous limestone, wavy-bedded.	75
	Tonoloway Formation		Thin-bedded to laminated, fine-grained limestone.	450
	Wills Creek Formation		Calcitic shale and argillaceous limestone.	120
	Williamsport Sandstone		Whitish, fine-grained, marine sandstone. Important gas reservoir rock.	20
	McKenzie Formation		Limestone and shale.	170
	Rochester Shale		Gray, very calcitic shale.	25
	Keefer Sandstone		Very limy, locally hematitic, sandstone.	8
	Rose Hill Formation		Reddish- to olive-gray sandstone, and shale with marine fossils.	200
	Tuscarora Formation		White orthoquartzite.	110
Ordovician	Juniata Formation		Reddish-gray siltstone, sandstone, and shale; non-fossiliferous.	700
	Oswego Formation		Olive-gray sandstone; *Orthorhynchula* abundant at base.	50
	Martinsburg Formation		Dark shale with interbedded limestone at base and turbidite siltstones more abundant toward top.	1400
	Trenton Group		Limestone and some shale, with metabentonites.	200
	Black River Group		Limestone, micritic to argillaceous.	165
	New Market Limestone		Cherty limestone, micrite, and magnesian limestone.	350
	Beekmantown Formation		(Only top 200 feet is exposed in axial region of Hightown anticline.)	2000

Table 2. Geologic formations in western Highland County, Virginia (from JOHNSON, ET AL., 1971).

STRUCTURE

Two distinct concepts of Appalachian structure have evolved, hinging on the question of whether or not basement rocks are involved in the thrusting and folding of the range. These are referred to as the "thin-skinned" hypothesis (little or no basement distortion), championed by Rodgers (1949,1972), Gwinn (1964) and others; and the "thick-skinned" hypothesis, in which movement of the basement is said to have played an important role in the folding of the sedimentary rocks. Advocating the latter concept, Cooper (1968) argues that "...vertical movements--more specifically sharp, differential downwarps--were the overshadowing, major tectonic feature of the Appalachian miogeosyncline as it evolved into a geosynclinorium." Thus, with subsidence of the trough concomitant with sedimentation, older rocks along the edges are supposed to have ridden the margins of the structure and become subsequently downwarped and in some cases overturned (Figure 2). The "thin-skinned" people argue for a decollement zone or series of bedding plane faults along which eastern strata were transported over beds of the central and western portions of the basin (Gwinn,1964). These sole thrust zones occur largely within incompetent strata in the Middle-Cambrian, Upper Ordovician, Upper Silurian and Middle and Upper Devonian, and are said not to involve the base of the Cambrian or the Pre-cambrian basement. The anticlines were formed as a result of the piling up of strata "...where bedding plane thrust faults shear upward as they rise to a higher decollement-glide zone...", or (in the case of lower amplitude anticlines) where splay faults branch from a major sole thrust (Gwinn, 1964; p. 890)(Figure 3A).

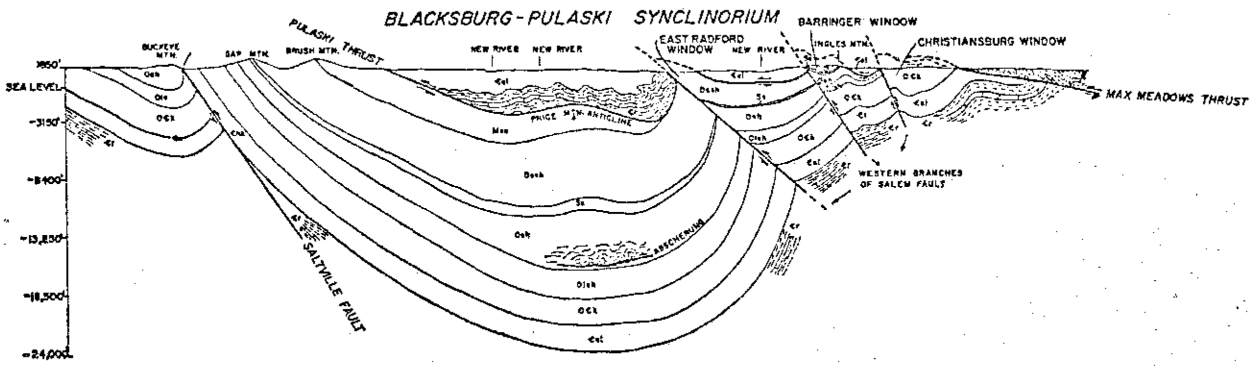


Figure 2. Cooper's view of thrust formation in a subsiding geosyncline (Cooper, 1968: Figure 7).

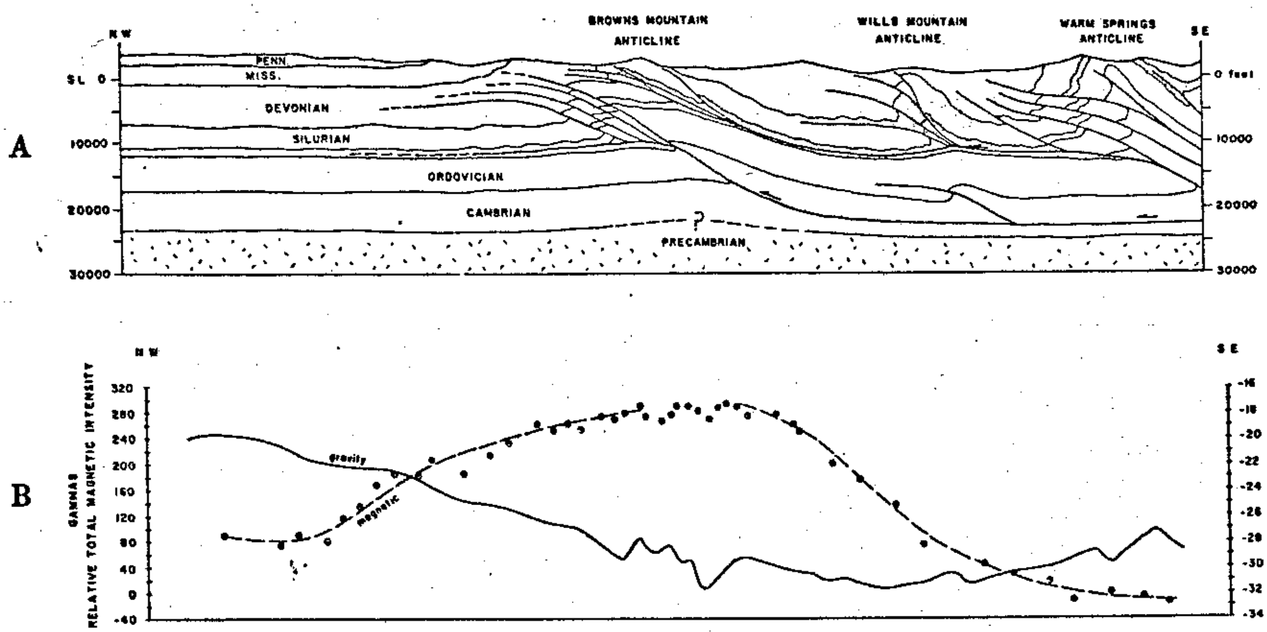
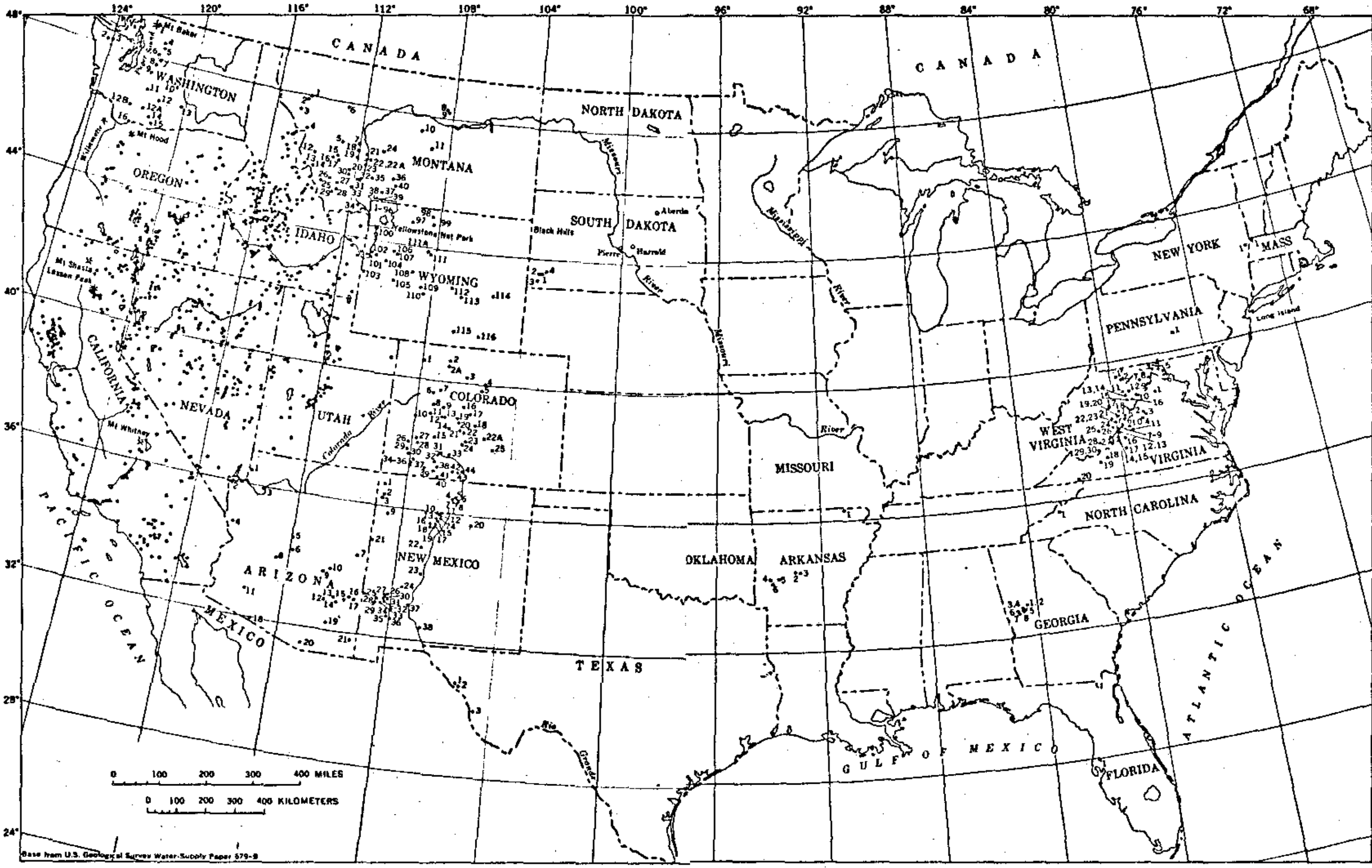


Figure 3: A. NW-SE section through Browns Mountain, Wills Mountain and Warm Springs anticlines, showing nature of decollements and splay faults. The line is through the median of the detailed gravity survey of the Kulander & Dean strip,shown in Plate 9. B. Bouguer gravity and total intensity ground magnetic survey profiles along same line (Kulander & Dean, 1972: Figure 5).

Inducement of the thrusting is attributed by Cloos (1972) to movement of the Precambrian basement in the vicinity of the Blue Ridge and Piedmont about 10km northwestward and upward during the Paleozoic, due to thus-far-anonymous forces originating farther to the east. The resulting compression drove the overlying Cambrian rocks to a depth of 20 to 30 kilometers. Spencer (1972) demonstrates, however, that part of the Blue Ridge complex, itself, was thrust over Cambrian strata along the Blue Ridge frontal fault a distance of some 16km; hence, the plot thickens!

Rodgers (1972) mediates the two epidermal views by allowing for some deformation of the basement in light of recent data on the Jura Mountains; thus:

> I have retreated, however, from insistence on a rigid and entirely undeformed basement, in the light of recent data on the Jura Mountains (see Rodgers, 1964). Basement irregularities exist there, and probably also in the Appalachians, but they formed earlier than the folds and hence could not have caused them, although they may have influenced the distribution of strata responsible for the decollement zones and then of the culminations and depressions in the folds. Likewise, the stratigraphic evidence makes me willing to extend the period of deformation in the Valley and Ridge province over much of the Carboniferous period (in addition to an Ordovician period of folding), but it does not convince me that folding and thrust faulting were produced only by continued uneven sedimentation in the subsiding basin, but rather, that after a period of extensional tectonics accompanying the main subsidence, a period of true compression ensued in late Mississippian to early Permian time, during which the final group of clastic sediments were deposited and the whole section was deformed.



<u>Figure 4</u>. Thermal springs of the U.S. (from WARING, 1965: Figures 2,3).

THE THERMAL SPRINGS

Distribution

Thermal springs are relatively uncommon in Eastern U.S. compared with the West; however, in terms of density, the Valley and Ridge province of Virginia and West Virginia contain about as many as an equivalent thermal area of the West (Figure 4). The measured temperatures, however, average considerably less than those of Western hot springs. For the purposes of this study, thermal springs are those whose maximum measured temperature exceeds 15°C (59°F)--substantially higher than the mean annual temperature of the region, which is 9-12°C (48-54°F) (Reeves, 1932). This allows for some elevation of ground temperature on southern exposures.

The distribution of thermal springs and groups of springs appears in Plate 5, and is documented in Table 3.* The circle size corresponds to the magnitude of discharge according to Meinzer's Scale (Todd, 1959: p. 34) averaged over different observations. Maximum recorded temperatures are colored according to the spectrum of the legend. In the case of a cluster of springs of differing temperatures, such as Hot Springs, Virginia, the proportion of the total discharge falling within a particular temperature range is depicted by the angle of its

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* Temperature conversion charts are provided in Appendix I.

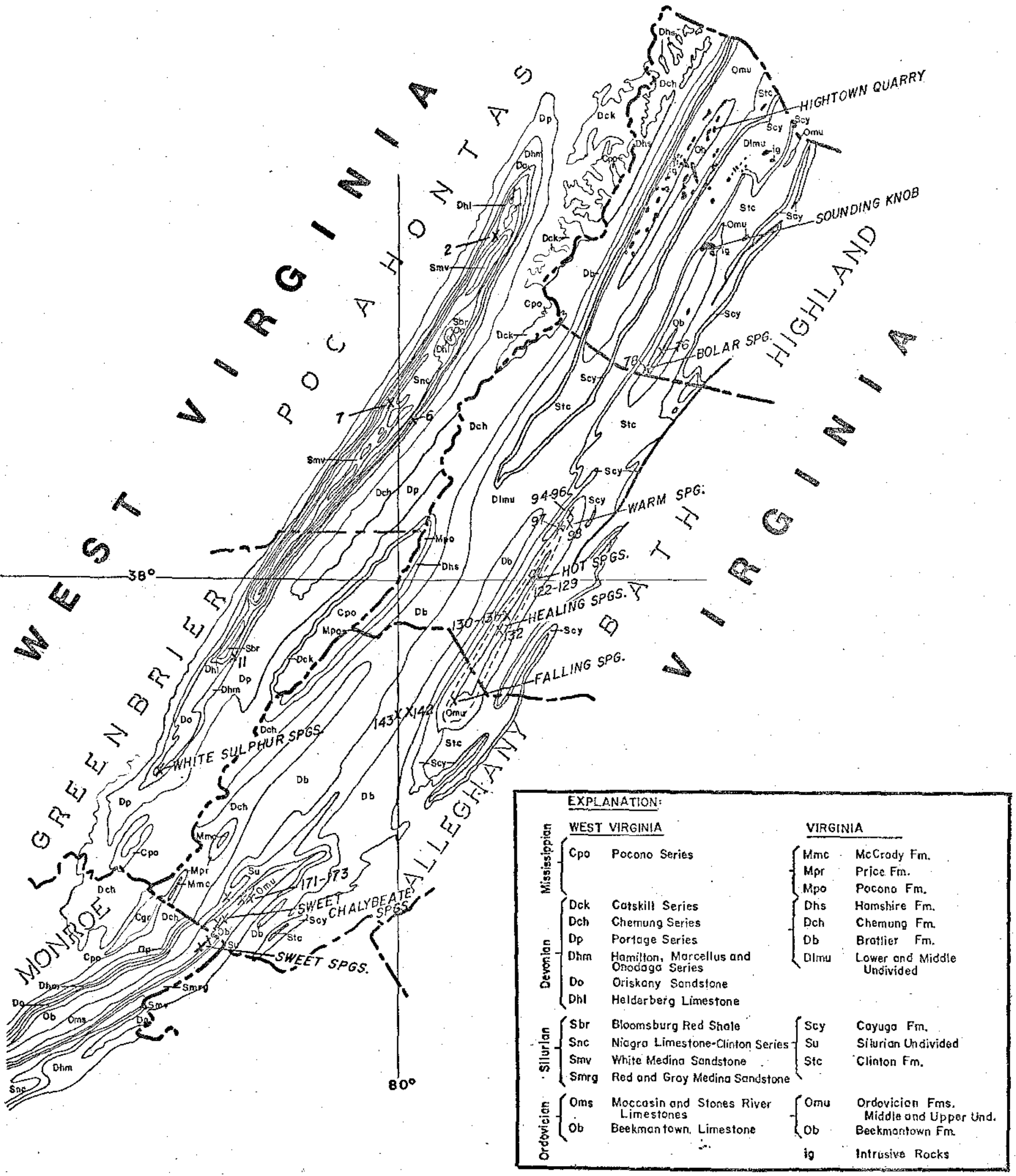


Figure 5. Relationship of major springs to anticlinal structure. Spring numbers are those assigned by Reeves (1932), and in Virginia correspond to numbers of Plate 5. In West Virginia, No. 2 is our 188-189; No. 6 is 181; No. 7 is 180; and No. 11 (Anthony Cave) falls below our temperature threshold.

Table 3. Analyses of waters from warm springs in Bath County, Virginia (sampled, July 1929). Numbers refer to springs in Table 3 (from REEVES, 1932: Table 4).

NO.	NAME AND LOCATION	DATE	GAL. PER MIN.	TEMPERATURE (°F)	REMARKS
20	Buttermilk Spring, 2 miles northeast of Sweet Springs, Monroe Co.	April 1,1928 Sept. 11,1928	100 10	70.5°) 71.6°)	Lowville limestone on faulted anticline.
21	Old Sweet Spring, drinking fountain and pool, Sweet Springs, Monroe Co.	April 1, 1928 Sept. 11, 1928	400 400	71.8° 73.4°	do.
76	Bragg Spring, 2.2 miles northeast of Bolar, Highland Co.	Feb. 23, 1928 Sept. 13,1928 Nov. 7, 1928	60 30 30	74 ° 74.8° 73.8°	Lowville limestone on northwest limb of anticline.
78	Bolar Spring, 3 miles northeast of Bolar, Highland Co.	Feb. 23, 1928 Sept. 13, 1928 Nov. 7, 1928	3,500 1,500 1,500	66 ° 68.8° 71.8°	Lowville limestone on northwest limb of anticline.
94	Ladies' pool at Warm Springs, 0.6 mile northeast of Warm Springs post office Bath Co.	Feb. 22, 1928 Sept. 12, 1928 Nov. 2, 1928		94 ° 95.3° ----	Lowville limestone on, or near crest of anticline.
95	Drinking fountain at Warm Springs, 0.6 mile northeast of Warm Springs post office, Bath Co.	Feb. 22, 1928 Sept. 12, 1928 Nov. 2, 1928	1,000 to 1,200	95 ° 95.3° 95.5°	do.
96	Gentlemen's pool at Warm Springs, 0.6 mile northeast of Warm Springs post office, Bath Co.	Feb. 22, 1928 Sept. 12, 1928 Nov. 2, 1928		95.2° 96.2° 96 °	do.
97	Meadow spring at Warm Springs, 0.4 mile northeast of Warm Springs post office Bath Co.	Mar. 21, 1928 Sept. 12, 1928	100 100	91 ° 91.2°	do.
122	Octagon Spring, Homestead Hotel, Hot Springs, Bath Co.	Feb. 22,1928 Sept. 12,1928 Nov. 2, 1928	67 67 67	95.1° 98.6° 98.6°	Lowville limestone near axis of anticline.
123	Boiler Spring, Homestead Hotel, Hot Springs, Bath Co.	Feb. 22, 1928 Sept. 12, 1928 Nov. 2, 1928	168 168 168	104 ° 104.9° 104.5°	do.
124	Hot Sulphur Springs, Homestead Hotel, Hot Springs, Bath Co.	Feb. 22, 1928 Sept. 12, 1928 Nov. 2, 1928	40 40 40	97.3° 98.5° 97 °	do.
125	Magnesia Spring, Homestead Hotel, Hot Springs, Bath Co.	Feb. 22, 1928 Sept. 12, 1928 Nov. 2, 1928	30 30 30	95.8° 96.5° 95.9°	do.
126	Spout Spring, Homestead Hotel, Hot Springs, Bath Co.	Feb. 22, 1928 Sept. 12, 1928 Nov. 2, 1928	36 36 36	103 ° 105 ° 105.8°	do.
127	Soda Spring, Homestead Hotel, Hot Springs, Bath Co.	Dec. 4, 1928	140	72.2°	do.
128	Warm Spring, Homestead Hotel, Hot Springs, Bath Co.	Feb. 22, 1928 Sept. 12, 1928	15 15	90 ° 91.8°	do.
129	Cold spring, Hot Springs, Bath Co.	Feb. 22, 1928 Sept. 12, 1928	350 350	56 ° 60.8°	do.
130	Drinking spring, Healing Springs post office, Bath Co.	Feb. 22, 1928 Sept. 12, 1928 Nov. 2, 1928	2 2 2	77.2° 84.2° 83 °	Lowville limestone on northwest limb of anticline.
131	Bathing pool, Healing Springs post office, Bath Co.	Sept. 12, 1928 Nov. 2, 1928	10 10	84.2° 78 °	do.
132	Bottling spring, Healing Springs post office, Bath Co.	Sept. 12, 1928 Nov. 2, 1928	1 1	86 ° 85 °	do.
134	Falling Spring, 8 miles southwest of Healing Springs, Alleghany Co.	Feb. 21, 1928 Sept. 12, 1928 Nov. 27, 1928	7,000	65 ° 66.7° 74 °	do.
141	Spring on west bank of Jackson River 2 miles south of Falling Spring Alleghany Co.	Mar. 29, 1928 Sept. 11, 1928	100 100	59.8° 63.1°	Oriskany sandstone on southeast limb of anticline.
142	Layton Spring, on east bank of Jackson River 2 miles south of Falling Spring Alleghany Co.	Mar. 29, 1928 Sept. 11, 1928	100 100	70.8° 71.6°	do.
171	Spring on Snake Run 3 miles north of Sweet Chalybeate, Alleghany Co.	Apr. 2, 1928 Sept. 11, 1928	200 200	56.5° 62.6°	Oriskany sandstone on northwest limb of anticline.
172	Iron Hill Spring, 200 ft east of summer hotel, 2.8 miles northeast of Sweet Chalbeate, Alleghany Co.	Apr. 2, 1928 Sept. 11, 1928	60 60	63 ° 67.5°	Lowville limestone on southwest limb of faulted anticline.
173	Iron Hill Spring, 400 ft east of summer hotel, 2.8 miles northeast of Sweet Chalybeate, Alleghany Co.	Apr. 2, 1928 Sept. 11, 1928	20 20	65.5° 68.2°	do.
174	Lee Carter Spring, 1.4 miles northeast of Sweet Chalybeate, Alleghany Co.	Apr. 2, 1928 Sept. 11, 1928	30 20	61 ° 62.9°	do.
175	C.B. Hunter Spring, 0.5 mile north of Sweet Chalybeate, Alleghany Co.	Apr. 2, 1928 Sept. 11, 1928	30 10	58 ° 59.9°	do.
185	Principal spring at Massanetta Springs, 4 miles southeast of Harrisonburg, Rockingham Co.	May 1, 1928 Sept. 15, 1928	400 300	59 ° 59.9°	Beekmantown or Elbrook dolomite near fault.
189	Warm Spring, 1 mile south of Bridgewater, Rockingham Co.	Feb. 8, 1928 Sept. 5, 1928	500 500	63.5° 62.9°	Chambersburg or Athens limestone.
195	Dices Spring, 1 mile southeast of Burketown	Feb. 8, 1928 Apr. 24, 1928 Sept. 14, 1928	2,000 1,500 1,500	63.2° 64 ° 64.4°	Copper Ridge dolomite.
196	Seawright Spring, 3.6 miles southwest of Burketown, Rockingham Co.	Feb. 8, 1928 Sept. 14, 1928	1,200 1,000	59 ° 59.7°	Elbrook limestone.
202	Fitzgerald Spring at Middle River bridge, 2.2 miles west of Fort Defiance, Rockingham Co.	Mar. 28, 1928 Sept. 14, 1928	60 60	61.2° 60 °	do.
251	Westernmost of 3 springs in Panther Gap, 1.6 miles west of Goshen Augusta Co.	Mar. 21, 1928 Sept. 13, 1928 Nov. 2, 1928	60 50 50	58 ° 59.9° 59.5°	Helderberg limestone.

Table 3, continued.

No.	NAME AND LOCATION	DATE	GAL. PER MIN.	TEMPERATURE (°F)	REMARKS
252	Easternmost of 3 springs in Panther Gap, 1.6 miles west of Goshen Augusta Co.	Mar. 21, 1928 Sept. 13, 1928 Nov. 2, 1928	50 50 50	66 ° 66.2° 66 °	Helderberg limestone.
267	Lithia Spring, Rockbridge Baths Augusta Co.	Feb. 25, 1928 Sept. 14, 1928	200 200	71.8° 72.1°	Beckmantown dolomite.
268	Magnesia Spring, Rockbridge Baths Augusta Co.	Feb. 25, 1928 Apr. 26, 1928 Sept. 14, 1928	600 500 300	71.8° 71.5° 71.6°	do.
269	Warm spring in North River, Rockbridge Baths, Augusta Co.	Mar. 29, 1928 Sept. 14, 1928	60 60	63 ° 64.4°	do.
301	Lithia Spring, 3.2 miles east of Gala Rockbridge Co.	Feb. 20, 1928	300	64.6°	Oriskany sandstone on nose of anticline.
E-28	Piercy's Cave Spr., 2 mi. N.W. Asbury, Greenbrier Co., W. Va.	July 27, 1945 Oct. 16, 1945	630 1,770	68 °	Water milky
E-35	Big Springs at White Sulphur Springs, Greenbrier Co.	Sept. 12, 1945 Oct. 15, 1945	840 670	62 ° 61 °	
E-37	Sterett Springs at White Sulphur Springs, Greenbrier Co.	Sept. 12, 1945 Oct. 15, 1945	610 550	61 ° 60 °	
E-42	Capacon Springs at Capacon Springs Hamshire Co.	July 25, 1944	200	64 °	Discharge estimated.
E-45	Cold Stream Run, 1 mile west of Cold Stream, Hamshire Co.	July 24, 1945	700	64 °	Water slightly cloudy.
E-96	Beaver Pond Spring at Bluefield near State Boundary line: probably in Virginia	July 20, 1945 Oct. 8, 1945	380 300	52 ° 52 °	Used by W. Va. Water Service Co. for emergency supply.
E-121	Springs on Right fork Trout Branch 6 miles southeast of Gap Mills. Monroe Co.	July 25, 1945 Oct. 14, 1945	310 360	64 ° 47 °	Represents combined discharge of several small springs.
E-122	Springs on left fork Trout Branch, 6 miles Southeast of Gap Mills Monroe Co.	July 25, 1945 Oct. 14, 1945	240 210	60 ° 45 °	Represents combined discharge of several small springs.
E-129	Berkeley Springs at Berkeley Springs Morgan Co.	July 23, 1945 Oct. 7, 1945	1,230 1,010	72 ° 72 °	Combined discharge of two large springs, Water temperature is the average for the two.
E-133	Roaring Springs at Circleville Pendleton Co.	Aug. 20, 1945 Oct. 16, 1945	5,500 5,050	65 ° 52 °	Combined discharge of several springs in group.
E-139	Roaring Springs, 1 mile north of Onego	Sept. 6, 1945 Oct. 2, 1945	850 2,100	61 ° 50 °	
E-141	Spring on Arbogast Farm, 3 miles north of Onego, Pendleton Co.	Sept. 6, 1945 Oct. 12, 1945	500 400	61 ° 50 °	
E-154	Thorn Ck. at mouth, 2 miles north of Onego, Pendleton Co.	Sept. 6, 1945 Oct. 12, 1945	7,700 8,000	71 ° 47 °	Represents combined discharge of springs tributary to Thorne Ck.
E-160	Trout Rock Springs at County line, 3 miles south of Hopeville, Pendleton Co.	Aug. 28, 1944 Oct. 12, 1944	510 650	61 ° 59 °	
E-176	Big Springs on Big Springs Fork, 2 miles west of Linwood, Pocahontas Co.	Aug. 29, 1945 Oct. 19, 1945	1,140 1,710	63 ° 57 °	
E-180	S.P. Curry Spring No. 1 at Huntersville, Pocahontas Co.	Sept. 12, 1945	230	64 °	
E-181	Minnehaha Spring at Camp Minnehaha, 4 miles south east of Huntersville, Pocahontas Co.	Sept. 11, 1945 Oct. 17, 1945	580 550	69 ° 67 °	
E-182	Peter McCarthy Spring on Browns Creek 5 miles northeast of Huntersville, Pocahontas Co.	Sept. 12, 1945	230	64 °	
E-183	Ruckman Run, 6 miles east of Huntersville, Pocahontas Co.	Sept. 11, 1945 Oct. 18, 1945	300 320	62 ° 50 °	Represents combined discharge of several small springs.
E-184	Guy Run, 4 miles southwest of Frost. Pocahontas Co.	Sept. 11, 1945	240	69 °	do.
E-185	Mill Run, 2½ miles southwest of Frost. Pocahontas Co.	Sept. 11, 1945 Oct. 18, 1945	400 460	65 ° 51 °	do.
E-188	Meadow Spring, 1/2 mile east of Dunmore, Pocahontas Co.	Sept. 10, 1945 Oct. 18, 1945	280 320	65 ° 64 °	
E-189	Drinking Spring and Pool Supply Spring at Dunmore, Pocahontas Co.	Sept. 10, 1945 Oct. 18, 1945	840 910	63 ° 62 °	
E-195	Cold Run Spring on left bank of Shavers Fork at Bowden, Randolph Co.	Sept. 14, 1945 Oct. 25, 1945	1,000 1,190	60 ° 52 °	
E-201	Big Spring 1/2 mile southwest of Harmon	Sept. 7, 1945 Oct. 24, 1945	2,290 2,820	61 ° 53 °	
F-18	Swan Pond Springs, 5 miles east of Martinsburg, Berkeley Co.	July 21, 1945	100	72 °	Discharge estimated; observed 1/2 mile below pond.
F-30	Spring on Everett Fruit, 5 mi. southeast of Romney, Greenbrier Co.	July 26, 1945	70	57 °	Discharge estimated; water slightly sulphurous.
F-50	Warm or Boiling Spring on Waites Run 4 mi south of Wardensville, Hardy Co.	Aug. 8, 1945	100	61 °	
F-123	Gillis Spring at Terra Alta, Preston Co.	Sept. 11, 1945	40	64 °	Formerly used for Terra Alta water supply. Discharge from pumpage record.
S-1	Limestone Springs, near Compton, Page Co.	No Data	No Data	61-66 °	3 or more springs on different properties; local use.
S-17	Blueridge Springs, Bufords Gap Bedford Co.	No Data	No Data	75 °	3 spgs; local use. Formerally Buford's Gap Spgs.
S-18	New River White Sulphur Springs Eggleston, Giles Co.		3	85 °	3 springs - resort.
S-19	Alum Springs, Sassin, 8½ mi north of Pulaski, Pulaski Co.	No Data	No Data	72 °	2 springs - resort.
S-20	McHenry's Spg., Near North Fork of Fork of Holston River,	No Data	No Data	68 °	Local use

Note: No letter - Reeves (1932) Va. G.S. Bull. 36.
E - Erskine Major Springs
F - Erskine Minor Springs } Principal Springs of W.Va., W. Va. Conservation Comm., 1948.
S - Stearns (1935). U.S.G.S. Water-Supply Paper 679.

sector. A reduced overlay of Plate 5 at the scale of the reduced maps is supplied in the pocket.

Temperatures

The warmest spring in the area is 105.8°F (41°C): Spout Spring at the Homestead, Hot Springs. Reeves provides a table of water temperature distribution of the springs, reproduced here.

Warm springs:	Number
100 - 105.8°F.	2
90 - 98.6	8
77.2- 86	3
60 - 75	27
55 - 59.5	45
Cold springs	
50 - 54.9°F.	120
35 - 49.5	116

Table 3. Thermal springs of Virginia and West Virginia exceeding 15°C (59°F) in temperature. Sources of data at bottom of table.

The warmest springs, for the most part, maintain a fairly constant temperature and volume. Those that vary appreciably with volume undergo a decrease in temperature with increasing volume in almost all cases; for example, Bolar Springs, #78; Big Springs, #E-176; and Harmon Big Spring, #E-201; Table 3. This behavior is attributed to dilution of a relatively constant flow of warm water by near-surface cold waters of variable flow, particularly since the higher temperatures occur during the dry season of September.

Discharge

Discharge of the springs ranges from slow seepages of 1 to 2gpm (.06-.12 liters/sec) to flows of several thousand gpm. The combined flow of the springs at the Homestead Hotel, Hot Springs, Virginia, amounts to about 300gpm (19 liters/sec) (Reeves, 1932), while those at Warm Springs amount to 1000-1200gpm (63-75 liters/sec). The largest flow occurs in Falling Springs, whose discharge is 6000-7000gpm (380-440 liters/sec). (See discussion below). Those at Healing Springs and White Sulphur Springs are mere seepages.

Chemistry

Several old reports on the chemical composition of the springs are available (Reeves, 1932; Price, et al., 1936; Watson, 1924; Froehling & Robertson, 1904). Reeves cites the U.S.G.S. analyses of 150 samples of water from warm and cold springs of the region. In general, calcium carbonate predominates, while the content of sodium and chloride is low and the sulphate content varies over a wide range. Mineral matter in solution ranges from less than 25 to more than 2500ppm, although most show less than 300ppm. Reeves quotes the analyst (M.D. Foster) thus: "...there is no definite relationship between mineral content and temperature, although most of the waters that contained more than 300ppm of dissolved mineral matter came from warm springs." A table of Reeves is reproduced here as an example of the measured contents of several of the thermal springs of Warm Springs Valley.

Composition (Parts per million)	Boiler Spring at Hot Springs (No. 123) b	Soda Spring at Hot Springs (No. 137)	Drinking Spring at Warm Springs (No. 95)	Drinking Spring at Healing Springs (No. 93-B)	Falling Spring (No. 134)
Silica (SiO_2)	24	16	24	25	20
Iron (Fe)	.30	.15	.11	.30	.13
Calcium (Ca)	138	103	118	140	153
Magnesium (Mg)	38	24	22	38	31
Sodium (Na)	7.4	4.4	6.5	8.4	7.4
Potassium (K)	6.6	5.1	2.4	7.3	6.5
Carbonate (CO_3)	0	0	0	0	0
Bicarbonate (HCO_3)	455	330	192	471	310
Sulphate (SO_4)	131	90	228	131	261
Chloride (Cl)	3.0	2.2	2.4	3.6	2.9
Nitrate (NO_3)	.05	.75	0	0	1.7
Total dissolved solids	586	433	525	596	672
Total hardness as $CaCO_3$ (calculated)	501	353	385	506	510

Table 4. Springs of Virginia and West Virginia grouped according to temperature (Reeves, 1932).

Reeves goes on to state, "Hot Sulphur, Spout and Magnesia Springs at Hot Springs, Virginia, are similar in composition to Boiler Springs. Cold Magnesia Spring is much lower in mineral content." He remarks also that those having temperatures above 80°F (26.7°C) seem to remain constant in composition through the year, while those below tend to vary. Noticeable gases escape from most of the warm springs and some of the cold springs. Nitrogen predominates. Much lesser amounts of oxygen and CO_2 occur, with traces of helium appearing in several of the warm springs.

RELATIONSHIP OF THE THERMAL SPRINGS TO GEOLOGY

Drainage relationships

The greatest density of thermal springs occurs in an *en echelon* series of anticlines along the highest altitudes of the Valley and Ridge Province in Virginia and West Virginia. These include the Hightown, and Warm Springs anticlines of the Wills Mountain anticline of Highland, Bath and Allegheny Counties, Virginia, and the Browns Mountain anticline of Greenbrier County, West Virginia. These features as well as an anticline to the southwest of Covington, running through Allegheny County, Virginia, and Monroe County, West Virginia, are shown in detail in Figure 5, with positions of thermal spring discharge points marked as accurately as possible from the available descriptions. It is evident that in this area of about 2 dozen thermal springs and groups of springs, all but 2 springs (#142,143) discharge in the anticlinal valleys. Within the breached anticlines, the springs issue (predominantly from the oldest exposed soluble rocks--the Ordovician and Silurian limestones and dolomites: an observation noted as early as 1840 by Rogers (1840-42)). The exceptions are several springs in Browns Mountain that discharge from the lower Devonian Oriskany Sandstone possibly on or very near the contact with the limestones. The non-thermal springs, likewise, tend to discharge from the same carbonate aquifers, as seen on Reeves' plot (1932). The thermal springs are not evidently associated with mapped faults; hence, we must conclude that they are stratigraphically controlled, though as with most limestone caves, their position within the formation is probably determined by local faults or jointing. In Warm Springs anticlines, the spas--Warm Springs, Hot Springs, Healing Springs and Bolar Springs--are all located at outlets of their respective valley segments; that is, at points where the valley wall has been breached and streams flow westward into the adjoining synclinal valley. These points are thus loci of minimum elevation along the valley axis. Wright (1925) describes these streams as being resequent; that is, they are former consequent streams that flowed down the limb of the original anticlinal surface. Under headward erosion they eventually pirated the subsequent stream that developed along the breached axis of the anticline, leaving it segmented. Almost all of the resequent streams drain to the west, because the west limbs of the anticlines are steeper; hence, the paths are shorter than to the east. It is nevertheless possible that faulting has helped direct the course of the lateral drainage. Thus, the drainage pattern may be an important clue in locating additional thermal waters that have not broken through to the surface.

If the source of the cold spring waters is the anticline limbs which rise 100 to 300 meters above the valley floor, then their collecting terrain would be the Devonian sandstones or soluble rocks younger than those of the valley floor, demonstrating that locally water does move down-section. Swallets and springs along the talwegs demonstrate also karstic circulation along the axial strata.

Deep-well data

Before attempting conjecture on the source of the ascending warm waters, we should review the available geologic information and interpretations of the substructure in the area of interest. Gravity, magnetic and seismic profiles were conducted along a WNW strip extending from Warm Springs anticline westward beyond Browns Mountain (Kulander & Dean, 1972). These data together with a well log to 7905ft (2590m) in Browns Mountain and another 13,001ft (4270m) in the Hightown anticline have provided sufficient information to construct a geologic profile along this strip (Figure 3A). Precambrian basement was detected at approximately 7200m (22,000ft) below Browns Mountain with a slight upwarp of its surface suspected beneath the anticline. Well-log data from the Sponaugle Well in Pendleton County, West Virginia, just north of Hightown, has been published and interpreted by Perry (1964). His interpretation is shown in Figure 6.

A detailed section through Browns Mountain is shown in Figure 7. Here both decollements along the incompetent Martinsburg shales have doubled the Ordovician section, and additional thickening has been effected by the splay faults above and within the glide zone. A gravity model of this section was calculated by Kulander and Dean, which matched reasonably the measured response of Figure 3B. The same

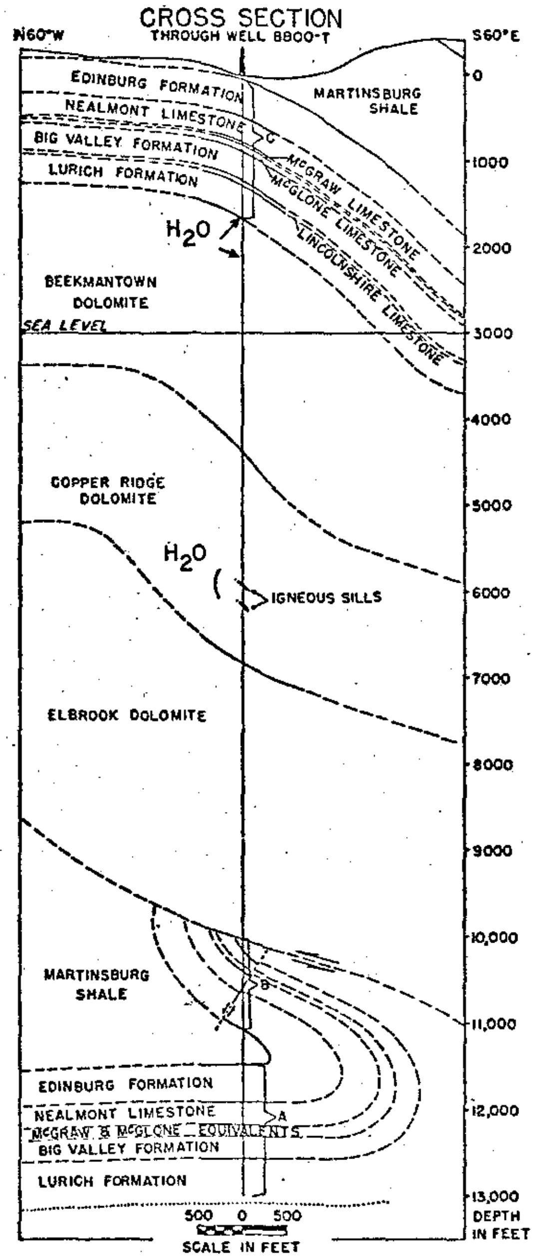


Figure 6. Cross-section through Sponaugle Well, Hightown anticline, West Virginia, showing points at which water was encountered (from PERRY, (1964: Figure 3).

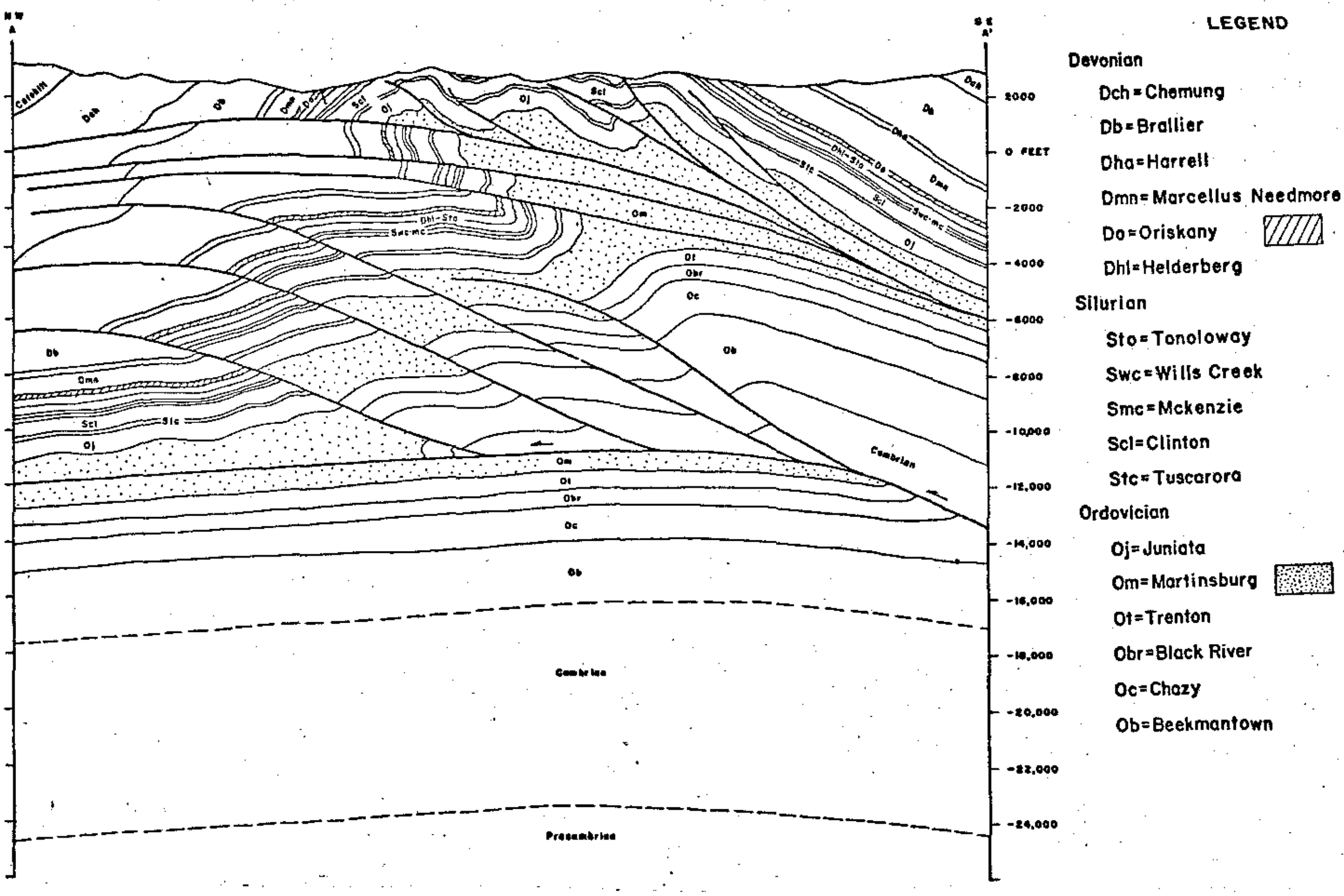


Figure 7. Imbricate thrust faulting in the core of Browns Mountain anticline, Pocahontas County, West Virginia. The faults are assumed to originate from decollements in the Martinsburg (stippled) (Kulander & Dean, 1972: Figure 4).

authors extended their geologic interpretation eastward to the Warm Springs anticline (Figure 3A).

The Sponaugle well (Figure 6) was spudded in the eastern limb of the Hightown anticline in Ordovician Martinsburg shale and penetrated 10,035ft (3160m) before encountering the upper decollement. It then passed through a nappe in the Ordovician, including the Martinsburg, and terminated at 13,001ft (3970m) in the Ordovician 1529ft (463m) stratigraphically beneath the collar level. Another well has been drilled (VDMR Well--W-1432) by Shell in a syncline of Rockingham County, Virginia. I did not have time to copy the log, but noted that it was spudded in the Devonian Brallier shales and bottomed at 13,830ft (422m) in the Upper Cambrian Conococheague limestone.

Tertiary Intrusives

A series of igneous dikes and sills outcrop in the vicinity of Monterey in Highland County (Figure 5). These occur within strata ranging from the Lower Devonian to the Ordovician Beekmantown Formation. The rocks, described as a porphyritic albite felsite have been dated by Fullagar and Bottino (1969, 1971) as being 47 million years old (K-Ar). This date corresponds to middle Eocene, making these rocks the youngest igneous rocks in Eastern U.S. The sills seem to have favored the weak Devonian Tioga bentonite for their emplacement. A baked upper surface confirms that these are indeed intrusives.

Fullagar and Bottino (1971) point out the occurrence of a basaltic dike cutting a felsite "dike" in a quarry northeast of Hightown. Robert Wright and I collected specimens of both rocks at the site, whose analyses are given in Appendix III. The younger rock has been submitted to Dr. Fullagar for dating.

The low Sr^{87}/Sr^{86} ratio of about .7 for felsites are said by them to be characteristic of mantle magma.

Darton and Keith (1898) mention a prominent outcrop of "basaltic" rock approximately 7 kilometers due south of Monterey:

> The outcrop No. 16 at Sounding Knob is on the top of a high anticlinal mountain of Tuscarora (Medina-Oneida) quartzite, the igneous rock rising as a steep-sided neck about 80 feet above the crest line. The altitude of its summit is nearly 4500 feet...
>
> Section No. 16, taken from Sounding Knob, displays the most clearly fluidal arrangement. In this area of the basalt it is evident, from the appearance of the rock mass upon the ground, that it was an old volcanic neck and was produced by an injection of considerable height.

To my knowledge, no further analysis or dating has been performed on this provocative occurrence, which outcrops on the axis of the Jack Mountain anticline containing Bragg and Bolar thermal springs, northeast of Warm Springs.

Transcurrent dikes farther east in Augusta County have been dated at about 145 to 153 million years (Jurassic). Zartman, et al (1967) suggest that these and the Monterey, Virginia, igneous rocks define the eastern end of a lineament of igneous activity stretching westward along about the 38th Parallel to eastern Kansas. The area also marks the termination of Gwinn's (1964) conjectured Burning Springs lineament extending approximately east-west across West Virginia. Gwinn's transcurrent faulting was postulated to account for the *en echelon* distribution of the anticlines. The tear faults supposedly bound independent thrust blocks within the Paleozoic sequence.

Of particular interest in the log of the Sponaugle Well is the vuggy nature of the top of the Beekmantown Formation, which Perry reads to be an erosional unconformity. A karsted surface developed in Ordovician time can be expected to be accompanied by subterranean solution-work in the form of caverns and conduits. Fresh water was encountered just above and below this unconformity and at two igneous sills in the Copper Ridge dolomite (about 6000ft, 1900m)(Figure 6). Whether or not these sills are related to Tertiary intrusives several kilometers to the south is not known. The water was characterized as being highly resistive, and likely represents meteoric water descending from the axis of the anticline. Water was not encountered below the decollement.

GEOPHYSICAL EVIDENCE

Published and other available geophysical data on the thermal area of Virginia and West Virginia have been assimilated and recompiled to a consistent map scale and presented herewith. On a regional scope, each of the maps reveal quite consistantly an anomalous zone trending along the area of highest temperature emanations and centering around Bath and Highland Counties, where occur the major thermal springs and Eocene volcanics. A number of these anomalies was pointed out by Dennison and Johnson in their 1971 article. We have here built upon their observations and correlations using the latest available data.

Seismic Refraction Measurements

In 1965 the ECOOE Experiment was conducted to measure crustal properties along the Eastern seaboard, utilizing nearly 1000 seismic records from over 100 shots on land and at sea. Stations were spaced at about 35km intervals on a grid extending across Virginia and adjacent states. Deep refractions of compressional waves (P_n) along the Mohorovicic Discontinuity consist of three travel-time intervals, or "time-terms"; namely, t_1: the time required by the shortest time ray to reach the Moho; t_2: the time of travel within the upper zone of the mantle, and t_3: the travel time from the Moho to the recording station (Figure 8). Multiple recording stations permit the determination of each of these

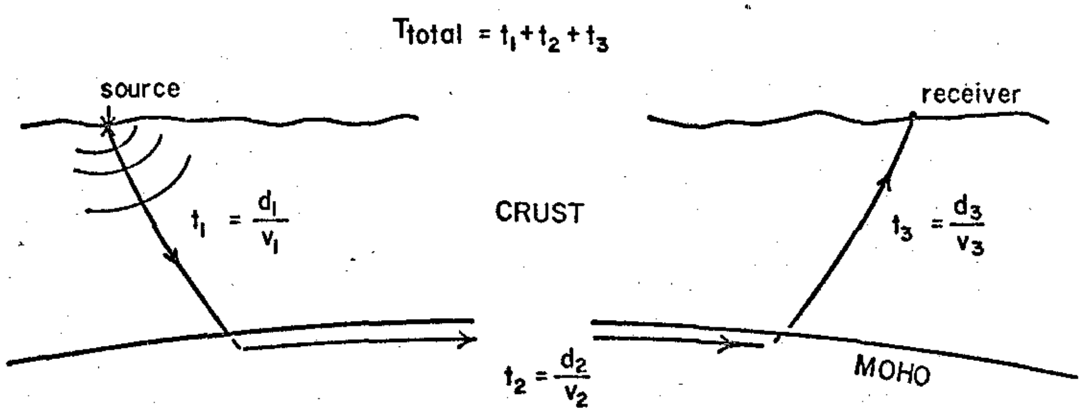


Figure 8. Travel-time terms of a seismic wave refracted along the Mohorovicic Discontinuity.

terms from the total travel time. Calculation of the first and third terms provides a measure of the crustal thickness beneath the source and receiver, respectively, when mean crustal velocities are introduced.

The time-term anomaly map resulting from these experiments was prepared by James, et al (1968) and is reproduced in Plate 6. It represents departures from the travel time through a 30km crust. On it, we see that the maximum delay times occur along the crust of the Valley and Ridge province and the thermal zone.

Crustal Thickness

With certain modifications, and the application of the best available information on mean crustal velocities (shown by the zones on Plate 6), the time-term map has been transformed into a crustal thickness map (Plate 7). The dashed contours are derived from other surveys and are regarded to be of lesser reliability. The crust beneath the crest of the Appalachians reaches almost 60km in thickness in the thermal area of Bath, Allegheny and Greenbrier Counties--a value not exceeded anywhere in the U.S. This computation modifies the previously deduced profile along the 37th Parallel by more than 15km (Willden, et al., 1968), shown at the bottom of the map.

Gravity Surveys

The regional Bouguer gravity map (Plate 8) is compiled from the U.S. map (Woolard & Joesting, 1964) and the Transcontinental Survey strip between Latitudes 35 and 39°N (U.S.G.S., 1968). The Plate shows a prominent gravity low of -80mgals extending over the principal thermal area of Bath and Allegheny Counties. This low is part of a longitudinal gravity low running along the crest of the Appalachians. More profound lows are observed to the southwest in the Valley and Ridge Province. Minus 100mgal pockets extend into the North Carolina high country. In Virginia the gravity lows correspond approximately to the crustal thickness maxima inferred from seismic measurements of Plate 7. The low gravity is characteristic of acidic basement rocks.

Two detailed Bouguer gravity maps for parts of the thermal area have been published, and appear in Plate 9. The map by Johnson & Pearson (1972), surveyed at 1 to 5km intervals, mostly along lines parallel to the topography, confirms the broad regional low and reveals several local pockets reaching a minimum of -89.61mgals in the extreme northwest corner of Bath County. A gravity depression of -80mgals also extends for about 7km just south of Hot Springs. This low centers on the ridge of Warm Springs Mountain and, as drawn, does not intercept the discharge points of the springs. The gravity strip of Kulander & Dean (1972) also confirms the regional low, but on this structure are superimposed ripples of about 1 mgal amplitude over the anticlines, including over Warm Springs (Figure 3B, profile). As a result of computer modeling the Browns Mountain anticline structure, they conclude:

> Anomalies of this nature observed over Browns Mountain and other major structures, therefore, are most likely due to mass distributions in the sedimentary cover above the lowermost decollement plane. In fact, the smallest anomalies could only be reproduced and rearranged by shuffling the 0.1g/cm^3 density contrast blocks in approximately the upper 2,000 feet of strata, and most of the 1 to 3mgal anomalies are most likely caused in large part by density variations in the upper 4,000 feet...
>
> Similar less defined ±1 mgal anomalies were detected over the Warm Springs anticline indicating a structural situation simulating that beneath the Browns Mountain flexure.

In general they attribute the small negative anomalies to near-surface thickened shales and the broader positives to tectonically thickened Cambro-Ordovician carbonates.

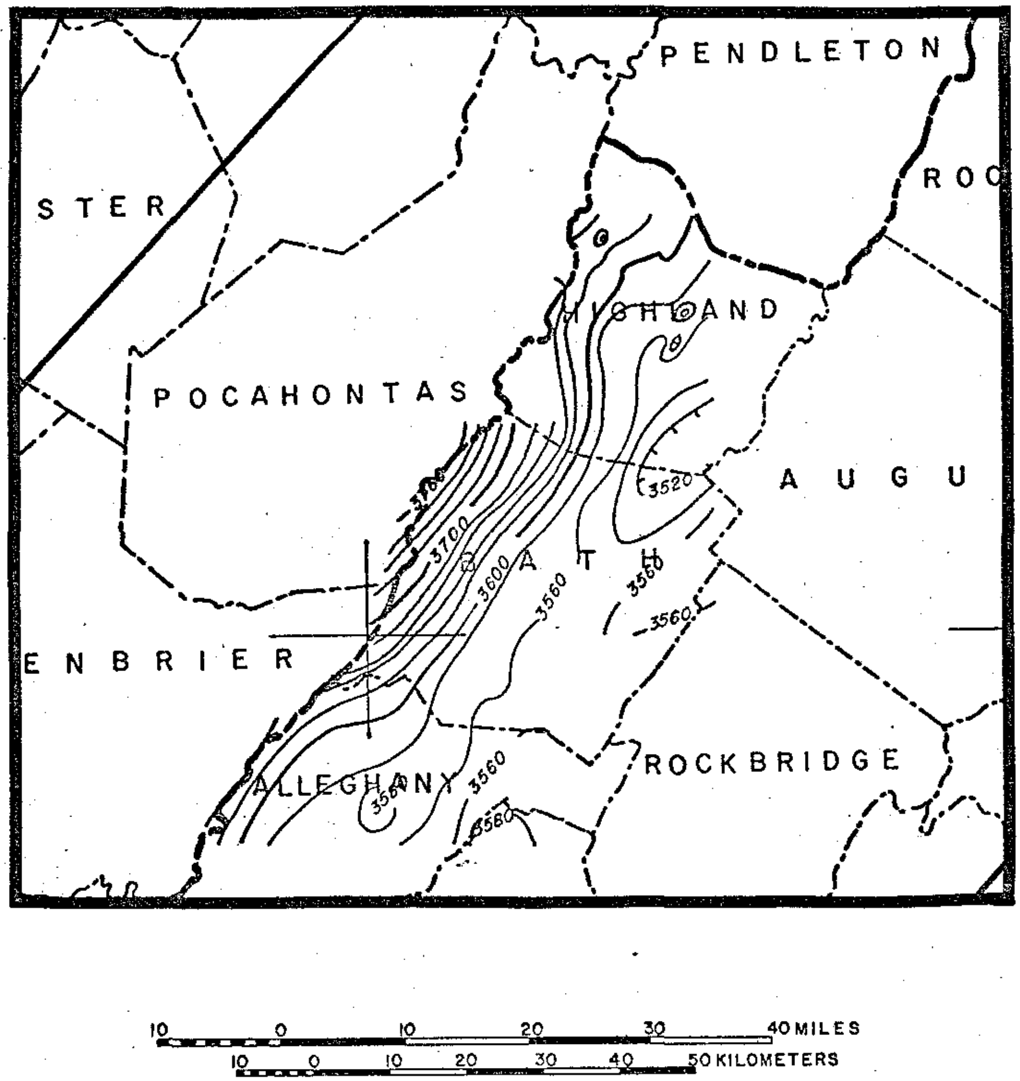
Magnetics

The regional aeromagnetic map of Plate 10 was compiled from two sources: Zietz, et al. (1966 and 1968). The contours of the individual maps do not mesh well.

The thermal area in the vicinity of Bath County is characterized by a very broad magnetic low centered on the crest of the Valley and Ridge province. To the north the low is not well defined. The Blue Ridge, composed of more mafic rocks, is expressed by a longitudinal magnetic high characteristic of a relatively high magnetite content. The low over the Valley and Ridge correspondingly can be attributed to the presence of acid volcanics in the basement.

Some detailed aeromagnetic mapping has been done by the Virginia Division of Mineral Resources (1972). Data from 5 15' quadrangles have been compiled in Figure 9. Additional quadrangles are on open file, but have not yet been procured. The regional low is here again suggested, with an axis lying just west of the thermal zone of Bath County. Two bulls-eye anomalies appear near the center of Highland County. The southern anomaly (+ 45γ) appears over Sounding Knob, elevation 4300ft ASL (1310m), about 700ft (213m) below the flight height. It may express the lookout tower on the Knob, or the basic volcanics that Darton and Keith (1898) described above as a volcanic neck. The anomaly to the north (-260gammas) on a spur of Jack Mountain (elevation 3883ft, 1166m) correlates with no mapped geological feature or cultural object on the 1969 Monterey 7½' Quadrangle. It may be a noise spike on the record.

DETAILED AEROMAGNETIC MAP



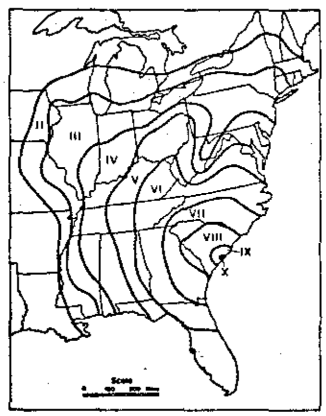
Open file report Virginia Geologic Survey, 1972
Contour Interval 20 gammas
Altitude 5000'A.S.L.

Figure 9. Detailed aeromagnetic survey over parts of Bath and Highland Counties, Virginia, flown at 5000ft (1640m) at a line-spacing of 3mi (4.8km). Contour interval: 20gammas with the regional removed.

Seismicity

One or two earthquakes are felt each year in Virginia, and others occur elsewhere along the Appalachians and in a zone across South Carolina. An insufficient density of seismic stations has prevented calculating Richter magnitudes and depths for most of these events; however, new installations this year should improve the resolution of future data. A seismicity map of the south-central Appalachians was published by Bollinger (1969), of Virginia Polytechnic Institute, who supplied us with his plots of Southeastern seismicity. Plate 11 was prepared from the latter. The epicentral locations on many of these events, particularly the older ones, are very crude. Some may be off by tens of kilometers. The symbol sizes correspond to intensities on the Modified Mercalli Scale, calibrated according to observations by the general public. Intensity IX, the most severe on the map, is defined by partial or total destruction of buildings.

It will be seen from the plot that the thermal area around Bath County is characterized by an absence of earthquakes. It is also reported to be a zone in which earthquakes are seldom felt, or at least the shocks are severely dampened. The phenomenon is illustrated by isoseismal or intensity maps of two earthquakes. The very destructive Charleston Earthquake of 1886 was felt over an area of 2 million square miles (5,200,000km^2), equivalent to that of the 1811 New Madrid, Missouri earthquake, regarded as the country's most severe. The pattern of isoseimals (Figure 10), however, shows a zone of lesser intensity extending along the Valley and Ridge province of Virginia and West Virginia (Bollinger, 1972). A similar lobe of damping appears in the map of the Elgood, West Virginia earthquake of 1969 (Bollinger & Hopper, 1970). Dr. Bollinger does not explain this phenomena.



ISOSEISMAL MAP (Rossi-Forel), from Dutton, 1889

Rossi-Forel	X	IX+	VIII+-IX-	VIII	VI+-VII	V-VI	IV-V	III	I-II
Mod, Mercalli	X	IX	VIII	VII	VI	V	IV	III	II

CHARLESTON, S.C. EARTHQUAKE - AUGUST 31, 1886

Intensity: X   Felt Area: 2 million sq. miles

[1811 New Madrid Eq.: XII, 2 million sq. mi.
1906 San Francisco Eq.: XI, 375,000 sq. mi.]

Deaths: ~ 60   Damage: ~ $5 million

Epicentral Effects

- Ground fissures, and crateriets
- Water, sand, and mud fountains
- Railroad rails bent, tracks displaced
- Loud earthquake sounds
- Earth and water waves
- Sulfur gas released

Unusual Aspects

- Region essentially free of shocks for preceding 200yrs.
- Large felt area
- Dual epicentral points
- West Virginia "low intensity"

Figure 10. Isoseismal map showing response pattern of 1886 Charleston, South Carolina earthquake. Note the prominent lobe of seismic damping extending along the Virginia/West Virginia border. (From BOLLINGER, 1972).

Heat-Flow Measurements

Although the most direct measurement that can be made in the exploration for a geothermal reservoir is that of heat flow, no such measurements have been made in the thermal areas of Virginia and West Virginia. Only one measurement has been completed in the Valley and Ridge of Virginia.

Heat flow is measured in micro-cal/cm^{2}sec, or HFU (heat flow units); that is, the number of microcalories (10^{-6}cal) passing vertically through a square centimeter each second. The measurement is obtained from the product of the rock conductivity (measured in the laboratory from core samples) and the thermal gradient, or change in temperature with depth, usually expressed in ^{o}C/km. The mean value of heat flow over the world is 1.5HFU. By comparison, the maximum near-surface value of heat flow measured in the Niland thermal area at the Salton Sea, California is 12HFU (Douze & Sorrells, 1972). Typical values of geothermal gradient are 23^{o}C/km in the Basin-Range and 13^{o}C/km in eastern U.S.

Available heat-flow values in Virginia and West Virginia are shown in Plate 5 beneath the gradient values. Values are mostly low, with one in the Blue Ridge south of Roanoke reported to be .78HFU (John Costain, personal communication), characteristic of mafic rocks having low radioactive components. A series of wells near Cripple Creek, Wythe County, southern Virginia, also yielded a low value (1.03HFU) in the

Cambrian Shady dolomite of the Valley and Ridge. Reiter and Costain (1973) attribute their result to Precambrian erosion of the radioactive materials of the crust prior to deposition of the sediments. Another value obtained near Grundy, Buchanan County, Virginia, just inside the Plateau produced a heat-flow value of 1.7, higher than normal. The logged wells were in sandstones and shales of the Pennsylvanian Pocohontas Foramtion (Reiter & Costain, 1973).

OTHER EVIDENCES

Schooley Erosion Surface

The highest ridges of the Plateau and the Valley and Ridge have long been recognized as a former erosion surface, originally described by Wright (1925, 1934) as the Upland, or Schooley Peneplane. He determined the approximate age for this surface to be Cretaceous or early Tertiary. Dennison and Johnson (1971) simplified and extended Wright's map (1925) of the erosion surface contours (Figure 11) noting:

> The position of maximum uplift of a dome on the Schooley surface with 1000ft of arching is remarkably close to the presumably younger (Eocene) intrusions in Highland County. This association is even more striking when it is realized that the general scarcity of preserved uplands in Virginia prevents detailed reconstruction of the Schooley surface.

From this relationship they infer the following:

> Thermal and mechanical energy (diapiric action) of an igneous intrusion below could upwarp the erosion surface. The near-coincidence of maximum domal uplift with the present known site of Eocene intrusion suggests a causal relationship.

Caves

Virginia and West Virginia are renowned for their limestone caverns, many of which have been popular tourist attractions since the early 19th Century. Most of the caves are found in the Valley and Ridge Province and the folds of the Plateau, where the carbonate rocks have undergone

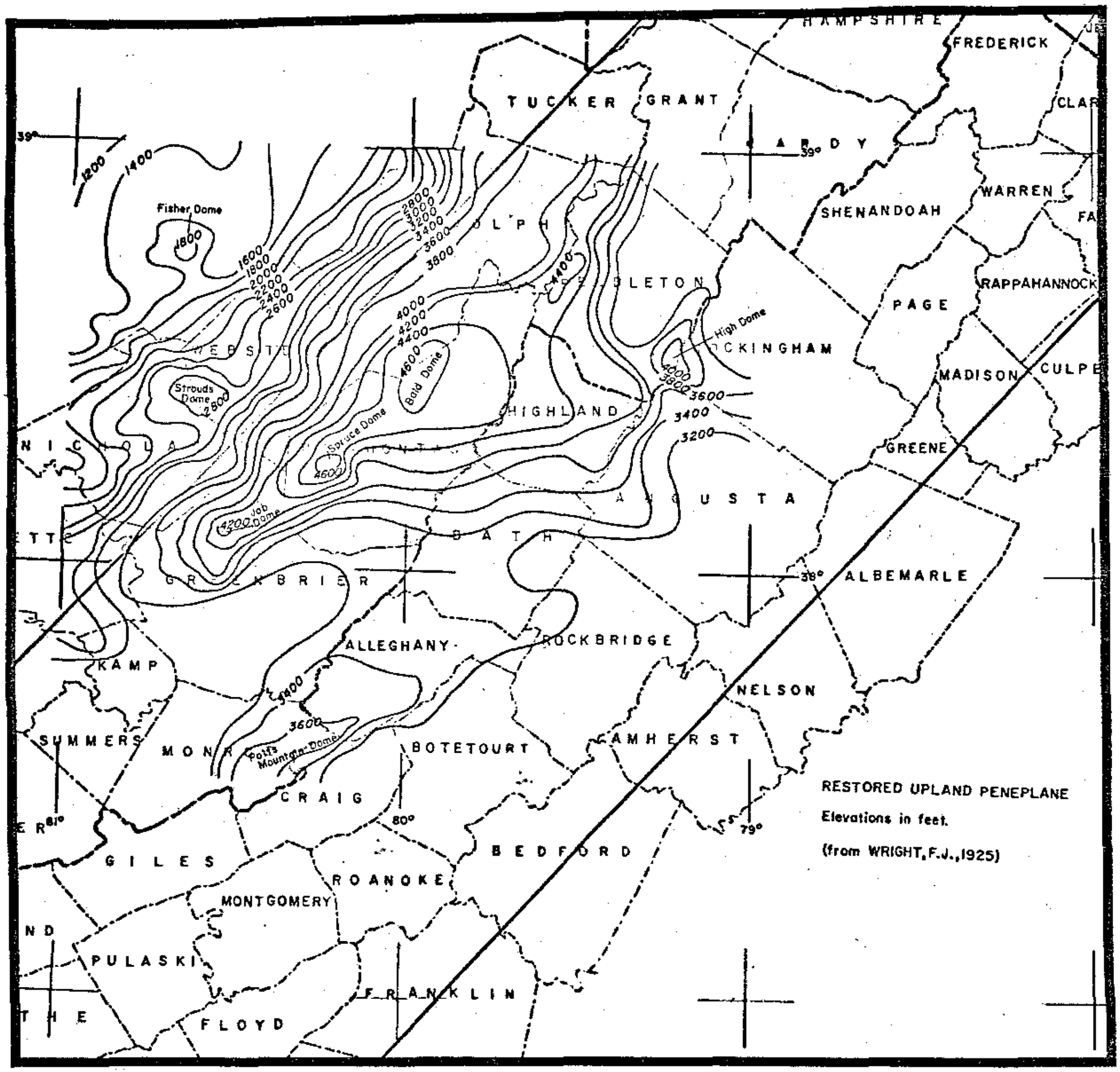


Figure 11. Elevation contours of the Eocene Schooley erosion surface, modified by Dennison & Johnson(1971) from Wright(1925).

fracturing and have been exposed to erosion and dissolving by meteoric waters. Highland, Bath and Allegheny Counties contain a relatively high number of caves--95, 74, and 84, respectively, according to Douglas (1964)--and no doubt more are known today.* Almost all of these exhibit normal air and water temperatures, and their origin has always been attributed to solution by streams and groundwater of normal temperatures. Solution by thermal waters, as evidenced in a number of western caves, can not be ruled out as playing a role in the development of at least some of these caves. Warm River Cave--the source of Falling Springs (#134, Plate 5), Allegheny County--provides the evidence as well as a clue to the underground behavior of the thermal waters.

The waters from Warm River Cave discharge from rubble at the base of a hill of Ordovician limestone, 11 miles (18km) SW of Hot Springs, at the southern end of Warm Springs anticline. Reeves (1932) reports a discharge as high as 7000gpm (440 liters/sec) in February of 1928. The springs fill a small pond (said to be too warm to sustain trout); hence, the water flows as a surface stream along the valley floor about 1 km before plunging over a waterfall and on down a steep canyon to the town of Falling Springs. The falls result from the buildup of travertine over many years.

* At least one more is known, because while touring the area, around Burnsville, Bath County, we were asked by the Highway Department if we would explore a cave that had just been opened by blasting in the road bank. We obliged, and found it to be passable for about 300ft (130m). Part of the cave extended under the road, to the dismay of the foreman.

The resulting cliff is about 30m high and extends for at least several hundred meters across the south end of the valley. This feature provides evidence that the carbonate-laden waters of Warm Spring Valley (be they warm or cold) have discharged in this manner for a very long time, and in so doing, may have elevated the local base level by the construction of the travertine dam. The water temperatures measured at the spring varied from 65°F (18.3°C) in February, 1928 to 74° (23.3°C) in November, according to Reeves. I recorded a temperature of 65°F (17°C) on March 15, 1973.

The cave was first entered by digging through the rubble of the spring and wading upstream about 100m where rockfall prevented further exploration (Dreese, 1956). On a later assault, the explorers entered the cave via a hillside sinkhole about 0.4km distant. They descended to a point about 55m below the surface, where they reached the main stream, having a temperature of 76°F (24.4°C). Upstream, it led to a pool whose temperature was normal (50-55°F) (10-13°C), but along the way it is joined by a small tributary (reported by Gurnee (1960) to be 2 liters/sec in volume). Its temperature registered 84°F (28.9°C). Their experiences in exploring this stream in July, 1956, deserve retelling in Dreese's own words:

> Past the formations and through a tight squeeze we came into a passage which permitted easier traveling. Working our way upstream by half swimming and half crawling we reached a siphon, but its two-foot length was easily managed and we continued on. As we progressed the temperature increased gradually from 85 to 95 degrees.

Eventually, the major volume of water was found to siphon out from a deep pool in the floor, while a small stream of even warmer water flowed down to meet it. At this point the ceiling rose to about 50 feet and several rooms and passages could be seen. However these upper levels were very high and reaching them would be a difficult task. A current of air was traced to a short side passage, in which we believe an opening could be made without too much work.

Up to this time most of our exploration had necessitated working through short siphons and swimming through "sewer-like" channels that provided so little head room we could not use our caving hats. As we continued beyond another siphon we found the space between the cave's roof and the water even more restricted. The heat became almost unbearable as it soared to about 100 degrees and we were beginning to find it difficult to breathe. Although the cave appeared to continue as far as we could see, it did not seem wise to continue the exploration any further. The extreme physical discomfort which finally forced us to turn back might have been caused by the very warm temperatures and the fatigue of our respiratory muscles as a result of continued water pressure rather than oxygen deficient air. But - no matter what the cause - further exploration of this particular portion of the cave would be very difficult and probably dangerous.

A later exploration of the cave (February 22, 1959) (Gurnee, 1960) found the temperature at the upstream end of the warm water passage to be 83^{o}F (28.3^{o}C). He makes a rough calculation of heat transfer:

The normal cave temperature in this latitude is approximately 56^{o}F. Assuming the walls and ceiling of the cave to have a normal mean temperature of 56^{o}F, we are subtracting from the stream and adding to the walls 35,000 BTU's in the two thousand feet from the "Float Way" to the "Dressing Room".

Additional explorations and temperature measurements are being made today by local speleological groups, who will be contacted.

Piercy's Cave (Plate 5: No. E-28) and Anthony's in Greenbrier, County, West Virginia (E-38) are associated with warm springs (Erskine, 1948), but in Davies' (1949) compendium of West Virginia caves, water temperatures are not discussed. Other caves may contain warm waters; in fact, most of the thermal springs discharging from limestone or dolomite may be regarded as issuing from caves or solution channels. Warm River Cave demonstrated the important role that dilution can play in lowering the temperatures of discharge, and the dramatic change in water temperature that can occur within about 50m of the surface.

THE SOURCE OF HEAT: DISCUSSION

Artesian Circulation

The high density of thermal springs in the vicinity of Bath County, Virginia and their paucity elsewhere in the Appalachians suggests that the local geologic conditions are unique. In the past, the thermal waters were ascribed to meteoric water circulating down synclines and emerging in the anticlines after having been heated by the normal geothermal gradient at depth. Reeves (1932) calculated that at a depth of 5000ft (1520m) the earth temperature exceeds that of the hottest springs in the region. Adopting the normal Eastern gradient of $13^{o}C/km$, mean ground surface temperature of $50^{o}F$ and the temperature of the hottest spring as $105^{o}F$ the minimum depth required for the required temperature rise of $28^{o}C$ would be 2km (6560ft), and this calculation assumes that no dilution by cold water takes place and that the water does not lose any heat during its ascent. Reeves was hampered by the lack of adequate topographic mapping. The topographic profile of Plate 2B drawn approximately normal to structure, shows that the only possible artesian source along that line for waters at Hot Springs would be the southeast limb of the anticline around Warm Springs Mountain, less than 2km distant. It seems doubtful that descending waters would attain the necessary depth over such a short path, and if they did, they would tend to lose much of acquired heat as they rose (unless they resided for a long time at the calculated depth). Furthermore, as Dennison & Johnson (1971) argue, we might expect thermal springs up and down the length of the Appalachians, since the structure at Warm Springs Valley is quite typical.

A further argument against Reeves' thesis is that continuity of bedding is interrupted by thrusts within the folds, so that simple arcuate flow paths probably do not occur. As seen on the cross-section of Figure 3A, descending water would tend to move eastward, ascending water westward. One might speculate that the thermal waters in Warm Springs anticline derived from the generally higher Browns Mountain anticline to the west after traveling down the thrusts to great depth (this is conceivable), but whence come the thermal springs of Browns Mountain itself?

Another possibility is that waters enter an anticlinal structure near its northern, usually higher end, penetrate to the necessary depth and ascend within the same structure to the south. Adequate path length is available in the case of the Warm Springs and Browns Mountain anticlines; however, in such a system we would expect to find warmer temperature discharges with increasing distance from the collecting area (Figure 12). Perhaps before dilution this is the case; however, with the available temperature readings of the springs, no apparent relationship exists.

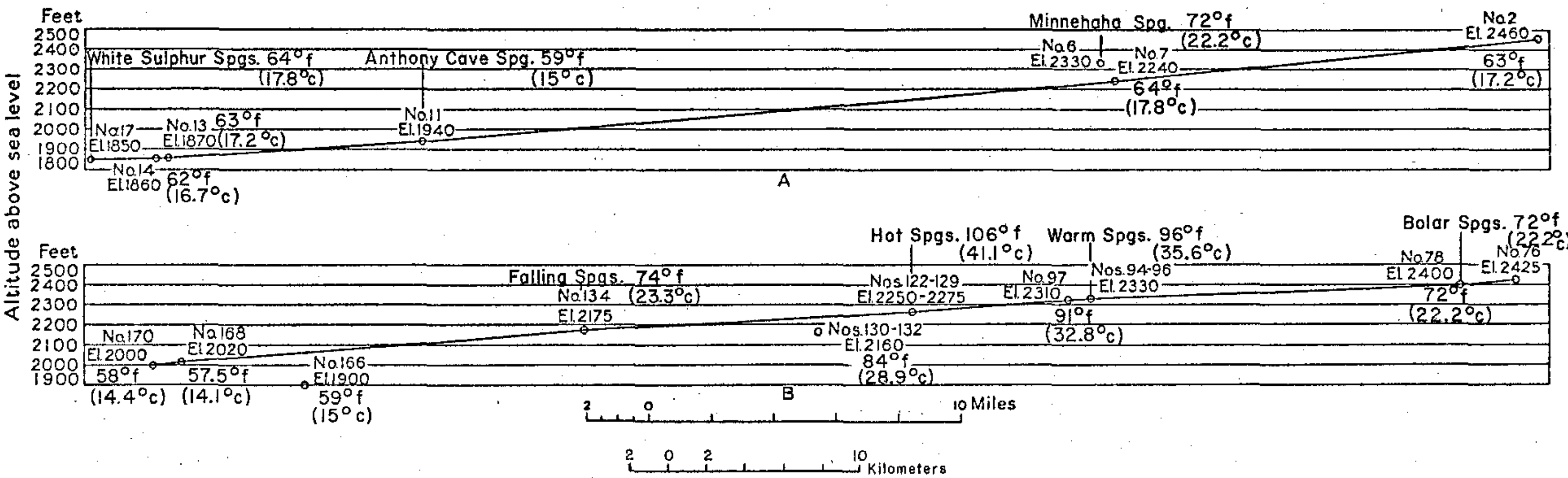


Figure 12. Southward decrease in elevation of discharge point of thermal springs in Browns Mountain and Warm Springs anticlines, respectively (from REEVES,1932). Compare with Figure 5. Temperatures of springs have been added.

Alternative Heat Sources

Radioactivity: Three minor contributors to conditions of elevated heat flow are 1) Radioactive decay of uranium, thorium and potassium; 2) oxidation of sulfides; and 3) friction of fault movement. None of these are generally considered adequate to account for thermal waters. Though increased radioactivity is associated with granitic basement rocks, evidence points to a decrease in heat generation with depth due to this phenomenon, possibly an exponential decrease (Lachenbruch & Bunker, 1971); hence, as in the case of the Sierra Nevada batholith, the inferred great crustal thickness in Virginia would not account for a local increase in heat generation or flow.

Sulfide oxidation: Reeves (1932) reports that some of the thermal springs have a high sulfate content (e.g., White Sulphur Springs), associated with a high bicarbonate content. These were all from limestones, and he attributes this to oxidation of pyrite to ferrous sulfate, with precipitation in water of ferric oxide and the formation of free sulfuric acid. Since not all of the springs exhibit an abnormal sulfate content, the process is occurring only locally, and cannot be invoked as an appreciable source of water-heating. Magmatic origin of the sulfates is also possible.

Fault-movement: Although stresses along thrusts may be building up, they are not known to produce felt earthquakes in the heart of the thermal area. Since low-angle thrusts are universal throughout the Valley and Ridge, there is no reason to expect an anomalous buildup of heat from the stress-accumulation here. In general, fault friction is regarded as a very minor heat source.

Heated pluton or magma chamber

A number of geophysical observations were presented above that display anomalous conditions in the vicinity of the thermal region. We might expect a relationship between the hot waters and these anomalies. Consider first the crustal thickness increase to nearly 60km in this area (Plate 7), deducted from the travel-time anomalies of Plate 6. As James, et al. (1968) demonstrate, such a thickening also accounts for the low gravity anomaly observed (Plate 8), because the denser mantle material would here be displaced. Such a gravity low is characteristic of a granite batholith (Bott & Smithson, 1967). We might also explain by this the broad magnetic low, since acidic crustal rocks would contain less magnetite than mafic mantle rock displaced at depth. But we have not yet explained the damping of seismic waves or the presence of thermal waters.

Dennison and Johnson (1971) deduce from consideration of the Eocene intrusions, gravity low, thermal-spring distribution and the apparent updoming of the Schooley erosion surface, that:

> ...the large negative gravity anomaly..., one of the greatest anomalies in the Appalachian region, is related to a deep felsic pluton that provides heat to the thermal springs centered in Bath County. The pluton is either the source of the andesite dikes in Highland County, or possibly represents a later thermal (volcanic) pulse related to the same deep crustal fracture. Unitl more subsurface data and detailed geophysical measurements are available, the assumption is made that the negative simple Bouguer anomaly, reaching a minimum of -83mg in western Bath County, overlies this pluton. It is suggested that the -70mg line approximates the general shape of the underlying pluton.
>
> In summary, we consider these phenomena all related to the Eocene intrusive activity in Highland County: the site of maximum uplift on the Schooley erosion surface, the culmination of both temperature and number of thermal springs, and the minimum value of the large negative gravity anomaly. Each of these phenomena is centered within 20mi of the point common to the boundary of Bath, Highland, and Pocahontas Counties. We consider that the near coincidence of these four geologic phenomena either represents surface expression of the Eocene volcanic pulse observed in the Highland County andesites, or perhaps a later manifestation of continuing spasmodic local plutonism related to a deep fracture zone.
>
> ...The data on thermal springs and gravity suggest that the center of the Eocene pluton is some 20mi southwest of the position of outcropping felsite dikes and sills in Highland County. Information presently available does not show us whether the thermal springs and gravity minimum are relict phenomena related to the Eocene volcanism or whether they possibly represent a still younger deep source of heat (volcanism).

In their analysis, they do not take up the matter of the crustal thickening or the seismic attenuation problem. From a heated mass of granite, however, we can expect to observe not only a decrease in density due to thermal expansion (except in the case of quartz, which begins to contract with temperatures above 570^{o}C (Skinner, 1966)), but also a decrease in the velocities of elastic waves (Horai & Simmons, 1968; Hughes & Maurette, 1956). For example, the latter authors found that in laboratory experiments on a biotite granite from Vermont, the compressional wave velocity at a pressure of 5000bars (equivalent to about 19km depth) decreased from 6.31km/sec at 25^{o}C to 6.08km/sec at 300^{o}C, temperatures to be expected at 19km in eastern U.S. But this normal elevation of temperature with depth is already taken into account in the velocity gradient. If, at that depth, the temperature were abnormally higher by an additional 300^{o}C, then an anomalous decrease in P-wave velocity of about 4% might be expected, corresponding to a travel-time delay of about 0.006sec over a depth range of 1km. This value is about 2 orders of magnitude less than the observed travel-time delays of about 0.5sec per km; hence, even allowing for the limitation of our calculation to 5000bars, we conclude that the effect of abnormal heating on seismic velocity (barring change of state) is negligible and will not replace thickening of the crust as an explanation of the travel-time delays measured.

Let us examine at this point the behavior of the geothermal gradient in a crust of 60km thickness. In Figure 13, two pairs of temperature-depth curves are shown, with a water-pressure scale corresponding to the depth scale. The solid curves represent models of constant heat generation in an 8km-thick upper layer of crust; the dashed curves correspond

TEMPERATURE GRADIENTS and MELTING CURVES

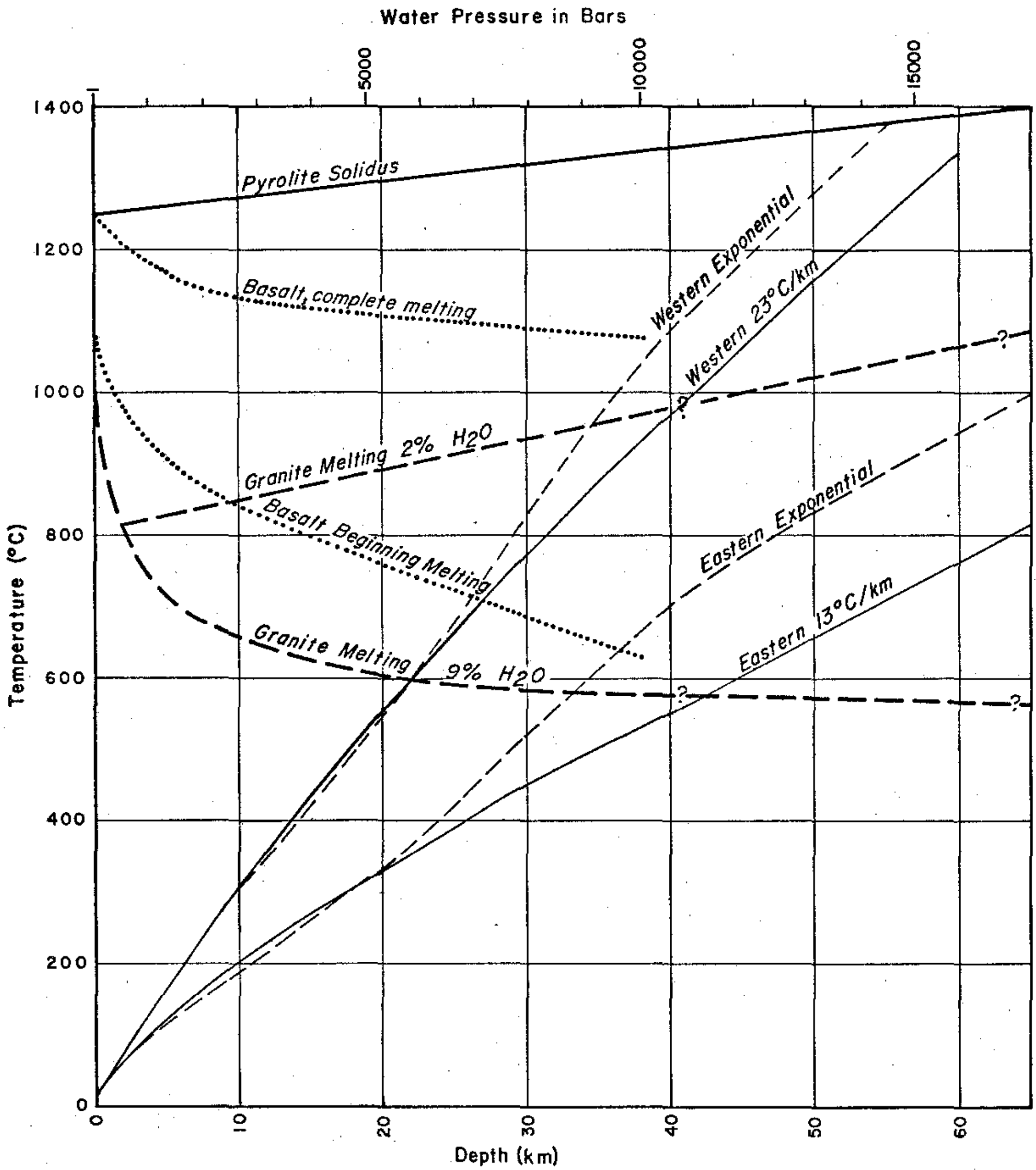


Figure 13. Typical eastern and western geothermal gradients for alternate heat-generation models. At depths where these gradients intercept the melting curves,molten or partially molten material can be expected.Sources:COMBS & SIMMONS(1973;CLARK(1966);&TUTTLE & BOWEN(1958).

to Lachenbruch's exponential heat-generation model (Lachenbruch & Bunker, 1971) for the same rate of heat flow (cf., Combs & Simmons, 1973: Figure 11). The upper gradient pair represents a heat flow of 2.0 units, typical of Western U.S., corresponding to a gradient of about 23°C/km. The lower pair is a typical Eastern U.S. condition of 1.4HFU and a gradient of about 13°C/km. The Western gradient intercepts the line of initial melting (solidus) of dry pyrolite, a possible mantle material, at a depth of about 60km (at a lesser depth if water is present). Combs and Simmons argue on this basis that the presence of a molten zone at shallow depths in the mantle of the northern Great Plains would explain anomalous travel-time delays observed in that area.

Also shown in Figure 13 are the melting-point range for basalt (Clark, 1966: Figure 15-16) and the anhydrous and hydrous melting curves for granite (Tuttle & Bowen, 1958: Figure 62). We observe that in the case of the Eastern geothermal gradient we may expect to find basalt and granite (in the presence of water) beginning to melt at a depth between 35 and 45km. Thus, in the thermal area of Virginia, where crustal thickness exceeds 55km, we may expect to find a 10- to 20km-thick layer of molten or partially molten crustal material atop the mantle. This deduction is in agreement with conclusions drawn by Tuttle and Bowen:

> If the water content of a segment of the earth's crust is on the order of 1-2 percent by weight and the composition is approximately that of the average granite a zone of melting at least 10 km thick will result.

It is not unreasonable to expect that at depths of 35-40km basalt could become partially liquid and that the liquid would have a composition not far from that of the andesitic rocks. If the unmelted portion then settled out, and the andesitic liquid, displaced upward, crystallized in an environment where an opportunity for settling of crystals still prevailed, liquids of granitic composition could result. This process of crystal fractionation by settling of crystals gives a progressively more salic liquid until granite liquids are produced--here the process stops. The liquid has now reached the bottom of petrogeny's residua system, and as crystallization of this granitic liquid proceeds the bulk composition of the crystals as well as the liquid has the composition of granite. Except for very special circumstances this is the end of the road ofr fractional crystallization, and consequently the composition of the liquid will remain essentially unchanged throughout crystallization.

It is therefore suggested that this zone of melting, where temperatures are high enough to melt granite completely and more basic compositions at least partially, may offer a mechanism for producing large batholitic masses of granite.

Some authors, e.g., Turner and Verhoogen (1951, p. 358) and Jeffries (1941, p. 834), have proposed geothermal gradients which give rise to a temperature of only 500°-600°C. at a depth of 40km. Perhaps this is reasonable for an average gradient for the earth as a whole, but it is certainly unrealistic in so far as the granitic rocks and the generation of granitic liquids are concerned. Granitic liquids have been formed in great abundance in the earth's crust in the past, and it is unlikely that such liquids were generated at depths below 40km. Granitic rocks are common in geosynclinal areas where a relatively thick blanket of sediments was present, but few, if any, geologists would suggest that the sediments ever attained a thickness of 40km. We are not here concerned with average geothermal gradients for the earth as a whole, but with gradients in geosynclinal regions where thick blankets of sediments serve as thermal insulators. It is probable that granitic liquids exist at all times beneath these thick sedimentary sections if the chemical composition is appropriate.

A thick blanket of Paleozoic sediments (estimated at 30,000ft or 9km) does exist in the thermal area (see Figure 3A), not to mention Precambrian metamorphics that likely occur in the upper part of the basement.

The presence of a lens of molten material, be it granite or basalt, would not in itself result in higher heat flow or thermal waters, since the pool is a result of the normal geothermal gradient. However, because of the length and breadth of the 55km crustal thickness anomaly (approximately 80 x 30km) we might suspect local vertical migrations along crustal fractures, resulting in localized heating in the upper crust, and occasional volcanism, such as occurred during the Jurassic period and Eocene epoch. The center of the Eocene volcanism (Sounding Knob) in Highland County falls within the maximum crustal thickness anomaly (Plate 7) and on a flank of the updoming of the Eocene Schooley surface (Figure 11). This volcanic neck lies on the structural axis of the Warm Springs thermal zone, 13km north of Bolar Springs, and 38km north of Warm Springs (Figure 5). Whether or not the present source of heat underlies the Knob or Warm Springs Valley, as Dennison and Johnson (1971) suggest, is not evident; but is conceivable that local magma chimneys somewhere under the anticline are able to heat groundwater and possibly set up convection movements within the aquifers. I would suspect similar conditions to prevail in the other anticlines containing thermal springs. As indicated by the presence of volcanic outcrops and higher temperature springs, the magma is likely to be closer to the surface under the Warm Springs anticline.

The presence of molten material beneath the thermal area would explain one anomalous condition that crustal thickness alone does not; namely, the seismic damping that appears in the isoseismal maps of earthquakes (Figure 10). Elastic waves are attentuated to some extent at lithologic boundaries, and moreso at phase boundaries; however, shear waves are absorbed when they enter a liquid; hence, only the compressional wave could pass upward from the mantle, through the molten material to the surface. Since the shear wave of an earthquake shock has normally 1 1/2 to 2 times the amplitude of the compressional wave, the major energy component would not be felt on the surface above such a reservoir, except for diffraction around the edges. The result would be a seismically "dead" area, such as Bollinger (1972) reports from Highland and Bath Counties. No doubt some attenuation and velocity decrease is also suffered by compressional waves within fluid, but I have no measurements on these properties in magma. Furthermore, since earthquakes take place more commonly around the edges of batholiths, rather than within them, we may account for the absence of shocks recorded from within the main thermal area (Plate 11).

That fractures do occur in the basement is demonstrated by the volcanic outcrops. Whether or not water circulates there or not, we do not know. We note, however, that in the

Sponaugle well, water discharged from two igneous sills at about 1900m depth within the Copper Ridge dolomite; so that some water movement is known to take place along igneous masses. Meanwhile, the principal aquifer for circulation of groundwater is the Beekmantown Formation. If a reservoir of thermal water exists, it would most likely occupy the karsted channels of this formation.

Part of the surface emanations may comprise magmatic water or condensed vapors, mixed with meteoric water. Rather high lithium and sulfate contents of some of the springs (Froehling & Robertson, 1904) suggest this. But in other localities, according to White (1957):

> Near the surface, waters of volcanic association clearly consist dominantly of meteoric water. The surface waters of New Zealand, Lassen Park, Yellowstone Park, Steamboat Springs, and Iceland differ isotopically from each other, but the hot springs of each area are very sililar to the surface water of the same area. In the waters that have been studied, the volcanic component probably does not exceed 5-10 percent of the total, and the isotope evidence does not actually demand the existence of any volcanic water.

Nevertheless, refined geochemical measurements of such quantities as silica content, K/Na and Li/Na ratios are used to determine temperatures at depth and origin of thermal fluids; and should prove to be very valuable in identifying the source of the Virginia waters.

RECOMMENDATIONS FOR EXPLORATION

The following studies should be conducted in order to ascertain the nature and source of heating of the thermal waters:

1) Collection and analysis of thermal waters and nearby cold waters in order to determine temperatures at depth and amount of dilution; temperatures and discharge of the springs must be recorded with each sample. Measurements should be repeated at another time of year when flow conditions differ. In addition to the major thermal springs of Warm Springs Valley and Warm River Cave, sampling should include springs of divers chemistry and geology. Dye tests may be required to identify the sources of dilution.

2) Examination of the geologic locale of the major springs, volcanic outcrops and magnetic features in Highland County. Particular scrutiny should be applied to surface-drainage routes of the anticlinal valleys for possible trans-current faults that might control the locations of the springs. Warm River Cave should receive special attention.

3) Microearthquake monitoring in the thermal region: The mapping of fault movements is a valuable tool for ascertaining the paths along which hydrothermal fluids move, since repeated fracturing is believed necessary for maintaining circulation.

<u>Micro</u>earthquakes are recorded in many areas where <u>macro</u>-earthquakes are not felt. All geothermal sites thus far monitored exhibit microearthquake activity (Ward, 1972; Lange & Westphal, 1969) and many show associated ground noise over the reservoir. Dr. Bollinger, seismologist of Virginia Polytechnic Institute, has proposed that he monitor the area at suitable sites over a period of at least two weeks, using one or two Institute seismographs. If micro-earthquakes are present, these should be subsequently mapped, using a network of four or more seismographs. Particulur attention should be given to the presence or absence of shear waves of distant earthquakes. While some of this work can be performed without cost to AMAX, as a part of the Institute's monitoring program; Dr. Bollinger indicates that support for a graduate student in extending the survey would be appreciated. This work could commence in May or June 1973.

4) Heat flow measurements in the thermal area. No such measurements have as yet been made; however, Dr. John Costain, also of VPI, is interested in attempting some. Difficulties in obtaining reliable measurements in the carbonate rocks are anticipated, due to groundwater circulation in the drill holes. I would hope that some pattern of heat flow could be obtained using shallow hole measurements, or by thermistor measurements in the walls of natural caves, which are plentiful in the area. Financial support would be required. It is

unlikely that Costain would perform more than a few trial measurements before Autumn 1973.

5) Petrographic analysis and age-dating. Analysis of travertines deposited by thermal and cold springs should be carried out on a number of springs as an index to present and former water temperatures and their origins. The presence of certain flora and fauna--diatoms, algae, bacteria, and gastropods--might be utilized for determining former temperatures (White, et al 1964). (See Appendix III, for an analysis of stream travertine from Warm Springs). Petrographic and geochemical analyses should be performed on igneous rock from the volcanic neck of Sounding Knob and an age-date obtained. Drs. Dennison and Fullagar at the University of North Carolina presently are attempting to date the igneous dike sample from the Hightown Quarry. (See Appendix III: Sample A-2).

6) Detailed geologic, geophysical and geochemical properties of the known thermal springs should be determined, in the hope that these will lead to the recognition of relict springs and confined waters. Among the properties that should be investigated: A) Gravity response of travertines and detectability of caverns; B) Electrical response of spring waters by resistivity surveys; C) Ground noise associated with thermal waters (Douze & Sorrels, 1972). The alimentary conduits may be mappable; however, it is likely that local

surface-and cold water streams will obscure the signals of thermal waters (Lange, 1972); D) The geochemical environment of the springs and vegetation. The vegetation itself may express an ecological association with thermal waters, as White (1964: p. 20) observes in the case of Steamboat Springs, Nevada, where pine trees favor rocks leached by sulfuric acid.

The above techniques are relatively inexpensive compared with deep exploration drilling and heat-flow measurements at depth. Hopefully, the preliminary surveys will provide sufficient information to determine the temperature of the waters at depth, nature of origin, and the existence of geothermal reservoirs.

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APPENDIX I

TEMPERATURE CONVERSION TABLES

Fahrenheit to Centigrade

	0	1	2	3	4	5	6	7	8	9
30			0	0.5556	1.111	1.667	2.222	2.778	3.333	3.889
40	4.444	5	5.556	6.111	6.667	7.222	7.778	8.333	8.889	9.444
50	10	10.56	11.11	11.67	12.22	12.78	13.33	13.89	14.44	15
60	15.56	16.11	16.67	17.22	17.78	18.33	18.89	19.44	20	20.56
70	21.11	21.67	22.22	22.78	23.33	23.89	24.44	25	25.56	26.11
80	26.67	27.22	27.78	28.33	28.89	29.44	30	30.56	31.11	31.67
90	32.22	32.78	33.33	33.89	34.44	35	35.56	36.11	36.67	37.22
100	37.78	38.33	38.89	39.44	40	40.56	41.11	41.67	42.22	42.78
110	43.33	43.89	44.44	45	45.56	46.11	46.67	47.22	47.78	48.33
120	48.89	49.44	50	50.56	51.11	51.67	52.22	52.78	53.33	53.89
130	54.44	55	55.56	56.11	56.67	57.22	57.78	58.33	58.89	59.44
140	60	60.56	61.11	61.67	62.22	62.78	63.33	63.89	64.44	65
150	65.56	66.11	66.67	67.22	67.78	68.33	68.89	69.44	70	70.56
160	71.11	71.67	72.22	72.78	73.33	73.89	74.44	75	75.56	76.11
170	76.67	77.22	77.78	78.33	78.89	79.44	80	80.56	81.11	81.67
180	82.22	82.78	83.33	83.89	84.44	85	85.56	86.11	86.67	87.22
190	87.78	88.33	88.89	89.44	90	90.56	91.11	91.67	92.22	92.78
200	93.33	93.89	94.44	95	95.56	96.11	96.67	97.22	97.78	98.33
210	98.89	99.44	100	100.6	101.1	101.7	102.2	102.8	103.3	103.9
220	104.4									

Centigrade to Fahrenheit

	0	1	2	3	4	5	6	7	8	9
0	32	33.8	35.6	37.4	39.2	41	42.8	44.6	46.4	48.2
10	50	51.8	53.6	55.4	57.2	59	60.8	62.6	64.4	66.2
20	68	69.8	71.6	73.4	75.2	77	78.8	80.6	82.4	84.2
30	86	87.8	89.6	91.4	93.2	95	96.8	98.6	100.4	102.2
40	104	105.8	107.6	109.4	111.2	113	114.8	116.6	118.4	120.2
50	122	123.8	125.6	127.4	129.2	131	132.8	134.6	136.4	138.2
60	140	141.8	143.6	145.4	147.2	149	150.8	152.6	154.4	156.2
70	158	159.8	161.6	163.4	165.2	167	168.8	170.6	172.4	174.2
80	176	177.8	179.6	181.4	183.2	185	186.8	188.6	190.4	192.2
90	194	195.8	197.6	199.4	201.2	203	204.8	206.6	208.4	210.2
100	212									

Figure I-1. Temperature conversion tables.

APPENDIX II

GOVERNMENT LANDS

APPENDIX II

GOVERNMENT LANDS

On Plate 12, appear the outlines of U.S. National Parks & Forests, Wildlife Management, Recreational and other federal lands, as well as State Parks and reserves. These are areas in which geothermal leases might be required or development might be precluded entirely. The information is derived from base maps of the individual national forests.

Homesteads, private and otherwise patented lands do exist within some of these areas, particularly in the National Forests. Their extent is shown on maps issued by the Forest Service for individual districts. A copy of the Warm Springs District map of George Washington National Forest is enclosed (Plate 13), showing the borders of these lands within the Forest boundary. This covers the important thermal areas of Allegheny, Bath and Highland Counties. Land plots showing individual ownership would have to be procured from the county offices in Monterey, Warm Springs and Covington.

APPENDIX III

PETROGRAPHIC REPORT ON THREE ROCKS FROM HIGHLAND AND BATH COUNTIES, VIRGINIA

By

D. A. Andrews-Jones

PETROGRAPHIC REPORT ON THREE ROCKS FROM HIGHLAND AND BATH COUNTIES, VIRGINIA

INTRODUCTION

Three rocks were submitted for petrographic examination. Two of these are from Highland County, Virginia and are "felsites" from the volcanic suite associated with the Tioga Bentonite Horizon. A 1 was taken from a "felsite" which appears to be concordant with the strata. This outcrop occurs about 2½ miles N.W. of Monterey. A 2 was taken from a felsitic dyke, which reportedly cuts across the strata. The third rock; designated A 3, is a travertine taken from a tufa deposit which forms a cliff associated with the thermal waters of Warm Springs in Bath County, Virginia. Thin sections of the three rocks were prepared commercially. A thin section is also being prepared from the acid soluble residue derived from the Travertine sample. Geochemical trace element analyses and total silica analyses are also in progress.

DISCUSSION

There is some controversy as to the character of the concordant "felsites" of Highland County, Virginia. They are considered to be either intrusive sills or lava flows. Potassium/argon age determinations of 47 my ± 3 my favor the former origin, since the associated strata are of Devonian age. The dykes are thought to represent the feeder channels for the concordant material.

The two felsitic rocks A 1 and A 2 are similar mineralogically, but differ considerably in texture (Plate I). A 1 is a biotite soda trachyte porphyry composed predominantly of albite phenocrysts with the interstices infilled with feldspar microlites. Biotite also forms phenocrysts. These mineral grains are in random orientation, and the phenocrysts vary considerably in size from fine-grained up to coarse (A 1 Plate II). A 2 is also a biotite soda trachyte porphyry, but has a strongly trachytic texture. That is, the mineral grains have a strong parallelism. The phenocrysts of A 2 are predominantly coarse grained, and also tend to be aligned with the matrix minerals (A 2 Plate II). They consist of both albite and biotite.

Both rocks are unaltered, except for carbonate replacement of feldspar, and along fractures. A 1 is more fractured than A 2 and shows more carbonate replacement. The carbonate was probably derived from the limestone sequence which they intrude.

The trachytic texture exhibited by A 2 is usually found associated with lava flows, and is thought to result from the flowage of the molten material. However, A 2 is reportedly derived from a dyke. In this case, the parallelism may be due to the restricted passage of the igneous material between the dyke walls. The random orientation of the mineral grains shown by A 1 is more likely to have occurred by cooling in place of intruded material, as in a sill.

A 3 is a typical travertine, and is composed almost entirely of calcium carbonate. Impurities such as silica are restricted to an estimated less than 1%. Only quartz grains and iron staining have been identified. No collodial or precipitated silica has been observed. However, determination of the acid solid residue, and total silica analysis will give a more accurate figure for these impurities.

DESCRIPTION OF THE ROCKS

A 1: This is a light grey porphyritic rock, composed of white feldspar, and black biotite phenocryst laths in a very fine-grained matrix (A 1, Plate I). It can be termed a biotite soda trachyte porphyry. The phenocrysts reach up to 8.0 mm in long axis.

In thin section, the feldspar insets can be seen to be albitic, and are frequently strained and/or multi-zoned (A 1, Plate III). It was not possibly to determine the nature of the zoning. Biotite also forms phenocryst laths, and is of the reddish brown variety. These phenocrysts vary considerably in grain-size from coarse to fine-grained, and form the bulk of the rock. Biotite is subordinate to albite. Feldspar microlites infill the interstices, and appear to be mainly potash feldspar. All of these mineral grains have a random orientation (A 1, Plate II).

Accessory minerals include apatite and zircon. Some epidote also occurs, but may have resulted from alteration of the plagioclase.

The rock is relatively fresh except for the introduction of carbonate along fractures, and minor replacement of feldspar by carbonate. Some marginal limonitic replacement of opaque material has occurred. The limonite forms halos around opaque grains. The latter are liberally disseminated throughout A 1, mainly as a fine dust. It is probably predominantly magnetite, and is estimated to amount to less than 1%.

A 2: The rock is a light medium-grey porphyritic biotite soda trachyte similar in hand specimen to A 1. It is composed of white feldspar, and black biotite phenocryst laths, up to 8.0 mm in length, in a fine-grained matrix (A 2, Plate I). However, in thin section there are major textural differences. A 2 has a strong parallel fluidal texture (i.e., it is trachytic) with virtually no interstitial material (A 2, Plate II).

The albite and biotite phenocrysts are mainly coarse-grained, the albite showing multi-zoning and strain (A 2, Plate III), and are set in a matrix of closely packed parallel to sub-parallel fine-grained feldspar laths, which do not exceed 0.1 mm in length. This matrix predominates over the insets, and contains a little primary biotite. The phenocryst tend to conform to the parallelism, but may oppose it. The matrix laths are then deflected around the inset/insets. Opaques again occur as a liberal dusting. A 2 is also relatively fresh, except for a similar development of carbonate to A 1. However, this is less intense, probably due to the less fractured condition of the rock. It contains apatite and zircon as accessory minerals, but no epidote was observed.

A 3: This is a buff colored calcareous sinter or travertine. It is composed of concentric layers of calcareous material enveloping numerous cavities in sponge-like form. Some pisoliths are also present, usually lining a cavity.

In thin sections, the rock is composed overwhelmingly of calcium carbonate. These form interlocking grains of very variably size, but which conform to the banding or layering. Thus, coarser and finer bands may alternate, or the structure may pass from very fine, through fine, to medium grained layers (Plate IV). The maximum grain size observed was around 1.0 mm. Pisolitic aggregates occur within the layering. These form ovoid grains each of which is composed of radiating carbonate grains arranged around a center, usually a hole (Plate V).

This travertine is probably + 99% pure carbonate. It is iron stained in places, and contains a few sparse and very isolated quartz grains or aggregates. These are estimated to represent less than 1.00% of the rock.

PETROGRAPHIC REPORT

Sample No. A1.

Rock Type Biotite Soda Trachyte Porphyry.

Alteration Carbonate + Limonite - Light

MEGASCOPIC DESCRIPTION

Light grey porphyritic "felsite". Composed of white feldspar, + black Biotite phenocrysts in a very fine-grained matrix. Phenocrysts are variable in size up to an observed maximum of about 8.0 mm.

MICROSCOPIC DESCRIPTION

Albite + Biotite phenocrysts ~~are~~ very variable in grain-size (from very fine-grained up to 4.0 mm observed), with interstitial microlitic material, which appears to be mainly K-spar. Carbonate replaces the feldspar + also occurs along cracks + interstitially within the matrix. Grain orientation is random throughout. The inset material predominates + all other material occurs only in the interstices.

ALTERATION

Apatite + zircon occur as accessory minerals, with some epidote (after feldspar?). Opaques occur as very finely disseminated dust.

Carbonate replaces feldspar, + limonite forms halos around opaques (probably mainly magnetite)

Essential Minerals

Quartz None seen.

K-Spar K-spar staining indicates dominant in matrix.

Plagioclase Albite phenocrysts greatly dominant.

Mica Red-brown biotite as phenocrysts + in matrix.

Other Accessory apatite, zircon, epidote, Carbonate liberal

Economic Minerals Opaques as very fine dust. < 1%.

PETROGRAPHIC REPORT

Sample No. A 2.

Rock Type Biotite Soda Trachyte Porphyry (Keratophyre)

Alteration Light Carbonate replacement.

MEGASCOPIC DESCRIPTION

Light medium-grey porphyritic "felsite", from Highland County, Virginia. Composed of white feldspar, + black biotite laths (up to 8.0mm). in a very fine-grained light grey matrix. Some feldspar laths exhibit resorbsion.

MICROSCOPIC DESCRIPTION

Coarse, strained + multi-zoned plagioclase (albite), with biotite phenocryst laths in a fine-grained groundmass (<0.1%), which composed predominantly of felted plagioclase (albite) with some biotite with a strong trachytic texture. Some carbonate occurs along fractures, interstitially, + replacing plagioclase. The matrix also contains accessory apatite + zircon, + veinlets, composed of breciated rock fragments in a sideritic matrix.

ALTERATION

Virtually none, except for a little replacement of plagioclase by carbonate, which also has developed along cracks. Staining indicates

Essential Minerals the presence of some potash feldspar.

Quartz None seen

K-Spar Some indicated in the matrix.

Plagioclase Predominant both in matrix + as insets.

Mica Primary orange-brown biotite in matrix + as inset

Other Apatite + zircon as accessories. Carbonate replacement

Economic Minerals Liberal fine dusting of opaques in matrix probably < .1%

PETROGRAPHIC REPORT

Sample No. A3.

Rock Type Travertine

Alteration None

MEGASCOPIC DESCRIPTION

Buff colored sinter or travertine. Composed of concentric layers of calcareous material enveloping numerous cavities in sponge-like form. Some pisoliths are present, and some ferruginous staining. From Bath County, Virginia.

MICROSCOPIC DESCRIPTION

Predominantly composed of interlocking grains of carbonate of very variable grain-size, from dust up to a maximum of about 1.0 mm in length. Frequently arranged into fine scale layering. Some layers enclose pisolithic aggregates composed of ovoid "grains" each formed of radiating carbonate grains arranged around a center, usually a hole. Also contains a few medium-grained quartz grains + some ferruginous staining. Impurities probably less than 1%.

ALTERATION

NONE.

Essential Minerals

Quartz Sparsely scattered quartz grains < 1%.

K-Spar None

Plagioclase None

Mica None

Other Carbonate dominant

Economic Minerals