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GEOLOGY OF THE

DEVILS HOLE AREA

FREMONT COUNTY, COLORADO

by
Richard Wright Vian

A dissertation submitted in partial fulfillment
of the requirements for the degree of
Doctor of Philosophy in the
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INTRODUCTION

STATEMENT OF PROBLEM

Rock exposures in the Devils Hole area, Fremont County, Colorado, are predominantly of Precambrian, high-grade regional metamorphic schists and gneisses and intrusive igneous bodies of granitic composition. Much of the northern part of the area is covered by a series of Tertiary volcanic rocks. The present study was undertaken with the following objectives: 1) the determination and mapping of the rock units; 2) a detailed investigation of the petrology and origin of the various rocks; 3) an interpretation of the geologic history; and 4) a description of the mineral deposits.

PREVIOUS WORK

The first general description of the geology of the region was presented by geologists of the Hayden Survey (Hayden, 1877). The area covered by the Pikes Peak folio (Cross, 1894) lies two miles to the northeast. Campbell (1922) described the scenery and sketched the geology of the part of the area along the route of the Denver and Rio Grande Western Railroad. Detailed studies have been limited to descriptions of pegmatite deposits. The Devils Hole pegmatite was mapped and described by Hanley *et al.* (1950); the Chief pegmatite was studied by Heinrich and Vian (1965). The nodular cordierite-muscovite schist occurrence east of Texas Creek was investigated by Travis (1956). Chapin and Epis (1964) conducted reconnaisance and some detailed mapping of the Tertiary volcanics.

As a part of the Arkansas Canyon - Huerfano Park mapping project being carried out by the Department of Geology and Mineralogy of the University of Michigan, four nearby areas have been studied in recent

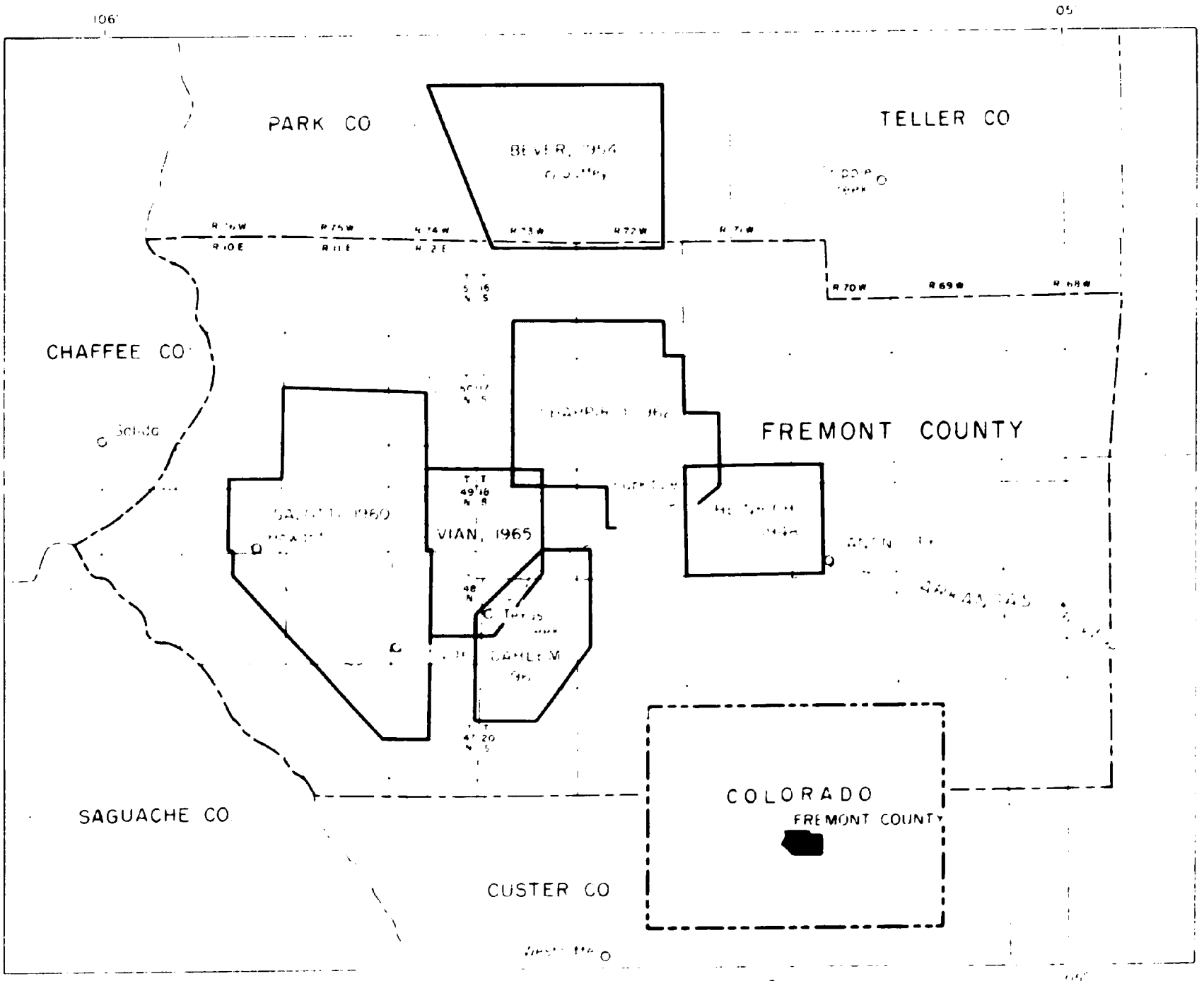


Figure 1. Index map of recent studies in the Arkansas Canyon area

years (Fig. 1). These studies and the one by Heinrich (1948) of the Eight Mile Park pegmatites (Fig. 1) involve similar rock units and structures.

PRESENT STUDY

The field work for this study was completed in the summers of 1961 and 1962. Mapping was done on U. S. Geological Survey aerial photographs (series GS-VDB, flown in 1954) and U. S. Forest Service drainage maps (27 NE and 34 SE). U. S. Geological Survey topographic maps of the Black Mountain and Cotopaxi quadrangles (15 minute) were used as base maps for the compilation of the geologic data (Plate I).

Laboratory work, employing thin-section, crushed-fragment, and x-ray techniques, was performed at the University of Michigan Mineralogical Laboratory and the Gulf Research and Development Company laboratories.

ACKNOWLEDGEMENTS

The writer is deeply indebted to Prof. E. Wm. Heinrich who suggested the problem and was a source of help, encouragement, and inspiration. The writer also wishes to express his gratitude to Profs. L. I. Briggs, W. C. Kelly, and R. M. Denning for their encouragement and thoughtful criticism of the manuscript. Prof. J. E. Bever of Miami University spent several days in the field with the writer and his knowledge of the area proved very helpful.

Financial aid, without which this study would have been impossible, was received from the Graduate Research Fund of the Horace Rackham School of Graduate Studies, from a National Science Foundation Summer Fellowship for Graduate Teaching Assistants, and from Mr. and Mrs. J. C. Vian. The Gulf Research and Development Company graciously allowed the writer free use of their facilities.

Drs. J. R. Shappirio, C. A. Salotti, and D. H. Dahlem were of much help through numerous discussions of various aspects of the problem. Messrs. Douglas Merrill and Roger Kolvoord ably assisted with the field work in 1962. Mr. Derwin Bell generously assisted in the preparation of the maps and did much of the final drafting. Local residents kindly provided information on the history of ranching and mining in the area. The writer especially wishes to acknowledge the assistance and encouragement of his wife, Janice Vian.

GEOGRAPHY

LOCATION AND ACCESSIBILITY

The Devils Hole area includes approximately 65 square miles along and north of the Arkansas River in west-central Fremont County, Colorado (Plate I). Highway U. S. 50 extends across the southern end of the area, following the south bank of the Arkansas River. The Denver and Rio Grande Western Railroad follows the north bank of the river. Colorado Highway 69 extends southward into Wet Mountain Valley from its junction with U. S. Route 50 at Texas Creek, a post office 30 miles west of Canon City, Colorado. The only bridge crossing the Arkansas River in the area is at the town of Texas Creek. The town consists of a combination post office and trading post and a railroad station. Seldom used mine and ranch roads lead back into the central part of the area from this bridge. The less rugged northern part of the area (mainly volcanic terrain) can be reached over ranch roads and trails starting well outside the area, principally from the Tallahassee Creek road (see U. S. Geological Survey, Cover Mountain Quadrangle) and the Bernard Creek - Red Gulch road (see U. S. Geological Survey, Cotopaxi Quadrangle). Much of the eastern and western portions of the area can be reached only on foot.

TOPOGRAPHY AND DRAINAGE

Elevations range from about 6000 feet on the Arkansas River at the southeast corner of the area to about 9400 feet along the north-central boundary. On the basis of topography and relief, the area can be divided into three sub-areas: 1) an area underlain by mixed igneous and metamorphic rocks, covering most of the central, eastern, and southern portions, in which the relief is highest (slopes average 1200 feet per mile) and the topography is characterized by the steep-walled valleys

and narrow divides of numerous gullies and small, intermittent streams (Fig. 2); 2) an area in which Pikes Peak granite predominates and in which the relief is less (slopes average 800 feet per mile); and 3) an area underlain by volcanics in which the relief is least (slopes average 600 feet per mile) and in which valleys and divides are relatively broad (Fig. 3). Relief is at a maximum (up to 1800 feet in less than a mile) along the contact between sub-areas 1 and 3. The erosional escarpment between these distinctly different topographic areas developed in the volcanics owing to the cap rock effect of resistant andesite flows in the upper volcanic unit (Figs. 6 and 7; Plate I).

The area lies entirely within the Arkansas River drainage basin and, indeed, the major drainage feature is the Arkansas River canyon. Six large valleys, draining an aggregate area of well over 200 square miles, enter the canyon in the area, three from the north and three from the south. Their mouths are opposite one another in pairs; Fernleaf Gulch (N) - Sand Creek (S) near the western boundary, Texas Creek Gulch (N) - Texas Creek (S) in the center, and East Gulch-Echo Canyon (N) - Spike Buck Gulch (S) near the eastern boundary.

Figure 2. General view of the topography of terrain underlain by mixed igneous and metamorphic rocks. View looking east from the divide between Reece Gulch and Texas Creek Gulch.

Figure 3. General view of the topography of terrain underlain by volcanic rocks (in distance). View looking north from Table Mountain.



Figure 2.

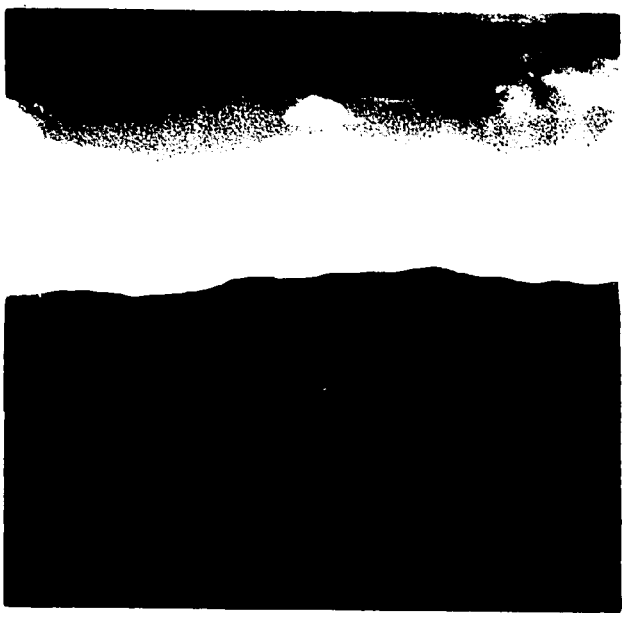


Figure 3.

GENERAL GEOLOGY

REGIONAL GEOLOGIC SETTING

The region of low mountains between Canon City and Salida lies west of the Canon City embayment, a structural re-entrant along the eastern flank of the Rocky Mountain cordillera that separates the Front Range to the north from the Wet Mountains to the south. Both ranges are elongate uplifts, arranged en echelon; the Front Range extends northward into Wyoming where it ends in the vicinity of Laramie. The Wet Mountains extend southward to the Huerfano River Valley east of Gardner, where the structural high continues to the southeast as the Apishapa uplift. On the west the Front Range and Wet Mountains are separated from other ranges of the southern Rocky Mountains by a discontinuous series of complex structural and topographic basins. The southern Front Range is separated from the Mosquito Range and the Arkansas Hills north of Salida by South Park. The Wet Mountains are separated from the Sangre de Cristo Mountains by Wet Mountain Valley and Huerfano Park.

The Devils Hole area is in the region of low mountains that forms at once the southwesternmost part of the Front Range, the northwestern part of the Wet Mountains, the southeastern and southern boundary of South Park, and the northern boundary of Wet Mountain Valley. The area cannot be assigned to any one particular physiographic or structural unit; it shares some of the geologic and topographic features of each of the bounding regions. It is, however, most closely related to the Front Range - Wet Mountain uplift structurally and the Wet Mountains topographically.

Laramide and post-Laramide uplift and erosion have caused the Precambrian core to be exposed over much of the Front Range - Wet Mountain

region. Paleozoic and Mesozoic sediments which once covered part of the region (Lovering, 1929) are preserved on the flanks of the ranges and in downfaulted blocks and downfolded inliers. Laramide and Tertiary intrusives are well exposed in the central Front Range and the ranges to the west. Formerly extensive deposits of Tertiary extrusives and sediments and Quaternary glacial material are partially preserved in the mountains and adjacent basins.

GEOMORPHOLOGY

Topographic Control

The present topographic surface in the Devils Hole area is the result of the varying degree of control that differential erosion has exerted on the development of the Arkansas River canyon. Three sub-areas have been developed generally as the result of differential erosion (see pp. 5-6). In general, the effect of structural elements in this control has been more local than that of the underlying rock-type. For example, foliation in the metamorphic rocks, faults (shear zones), and joints locally control the course of both the Arkansas River and lesser streams and gullies. Fault zones also result in topographic lineations which can be seen easily on aerial photographs. The Arkansas River flows through a series of ESE- and NE-trending valley segments which are roughly parallel to more or less well-developed sets of fault zones and joints. Fernleaf Gulch, Sand Creek, Texas Creek Gulch (Fig. 4), Echo Canyon, and Spike Buck Gulch are all localized along faults or zones of intense shearing. However, the major fault-controlled feature is Echo Canyon, a graben produced by Laramide faulting (see pp. 154-155). The relative ease with which the poorly consolidated material (Farisita (?) clastics and basaltic agglomerate) that filled this graben was removed,

Figure 4. The valley of Texas Creek Gulch, formed in the Texas Creek fault zone; view looking north from near the mouth of Texas Creek. Table Mountain in the distance.

Figure 5. Echo Canyon graben, looking south-southeast towards the Arkansas River Canyon and the mouth of Spike Buck Gulch.



Figure 4.



Figure 5.

once the cover of post-graben volcanics had been stripped away, determined its present form, a broad, low valley (Fig. 5; Plate I).

Physiographic History

The physiographic history cannot be traced back beyond the development of the surface on which the Tertiary volcanic rocks were deposited. This surface has a moderate relief of approximately 1000 feet (note contact elevations on Plate I).

If the multiple peneplain or erosional surface hypothesis (Van Tuyl and Lovering, 1935) is valid, this surface may be correlated with a surface of regional physiographic significance. Elevation and age relations indicate that the surface is correlative with either the Overland Mountain (Oligocene) or the Cheyenne Mountain (upper Eocene) surface of Van Tuyl and Lovering (1935). The surface can be no younger than Oligocene judging from the age of the volcanic rocks that overlie it (see p. 128). Other students of Front Range physiography believe that the only regionally significant erosional surface, renamed by Wahlstrom (1947) the Rocky Mountain peneplain and equivalent to the Badger Creek surface of Stark et al. (1949), was formed during the Pliocene. As pointed out above, the pre-volcanic surface in the Devils Hole area cannot be related to a Pliocene surface. Robert MacNish (pers. comm.) identified a major Eocene erosion surface in Wet Mountain Valley which he correlates with the pre-volcanic surface north of the Arkansas River. He feels that the surface ranges in age from lower Eocene in the southern part of Huerfano Park to upper Eocene north of the Arkansas River. MacNish (pers. comm.) believes that this surface should not be correlated with any of the surfaces described by Van Tuyl and Lovering (1935), Wahlstrom (1947), or Stark et al. (1949). The pre-volcanic surface is

discussed in detail by Chapin and Epis (1964).

That the Echo Canyon graben was in existence prior to the deposition of the volcanic rocks is evidenced by: 1) the lack of volcanic material in the Farisita (?) clastic rocks; 2) the occurrence of a unique basal volcanic unit (basaltic agglomerate, possibly a mud flow) in the graben, indicating a topographic low (possibly a structural low) capable of preserving a unit that was either not deposited or not preserved anywhere else in the area; and 3) the graben-bounding faults truncated by or causing relatively little displacement in the volcanic rocks. Some movement of the graben subsequent to the deposition of the rhyolite tuff unit may be evidenced by the apparent downwarping of the unit where it crosses the graben (see p. 154). The fact that the upper volcanic unit (a series of andesite flows) crosses the graben faults without apparent downwarp indicates that little movement has occurred in the graben since the tuff unit was deposited.

Quaternary landslides along the erosional escarpment at the north end of Echo Canyon (Figs. 6 and 7) have modified the topographic expression of the graben and have obscured the contact relations among the various rock units present. The landslide debris consists of large blocks and associated rubble, derived principally from the rhyolite tuff unit, which collapsed down the escarpment into Echo Canyon.

An interpretation of two stream courses sheds some light on the development of the Arkansas River canyon (including the Royal Gorge). Certain of the streams in the area are fault-controlled, but Texas Creek and East Gulch are superposed streams that have recently cut deep gorges in resistant granite (Plate I), both ignoring easier paths that were available to them. Texas Creek has been unable to alter its course in

favor of the much softer rocks in the broad Texas Creek fault zone. East Gulch has been unable to follow the Devils Hole shear zone leading south out of Devils Hole. The writer considers this to be evidence for strong uplift in geologically recent (early Pleistocene ?) time. Prior to this uplift a consequent drainage pattern had developed on a predominantly volcanic terrain. During uplift only those streams carrying sufficient volumes of water were able to maintain their courses across the exhumed Precambrian terrain. Other workers (Powers, 1935; Stark et al., 1949) have also dated the beginning of the canyon-cutting cycle as late Pliocene or early Pleistocene. MacNish (pers. comm.) feels strongly that the last major uplift (the one reflected by the cutting of the Arkansas Canyon and Royal Gorge) is even more recent -- post-Illinoian.

Physiographic features of Quaternary age, other than the landslides, are represented by two sets of terraces developed along the Arkansas River (Fig. 8). Dissected and poorly defined remnants of one set of terraces can be found in wider parts of the canyon, e.g., in the relatively open areas around the town of Texas Creek and the mouth of Echo Canyon. They are at elevations of 150 to 200 feet above the present level of the Arkansas River. The lower terraces, composed of coarse gravels, are well preserved along much of the river at elevations ranging from 20 to 40 feet above the river.

Powers (1935) found evidence for six sets of terraces along the upper Arkansas River. The lowest of these was formed during the Wisconsin glacial stage. Some of the higher terraces were formed during pre-Wisconsin glacial stages whereas others are related to unidentified episodes of terrace formation. Relative elevation and degree of preservation indicate that the lower terraces in the Devils Hole area

Figure 6. Erosional escarpment and landslides north of Echo Canyon. Tops of slides are marked by relatively flat, grassy areas.

Figure 7. Erosional escarpment and landslides north of Echo Canyon. Rhyolite tuff block at lower left is part of one slide.

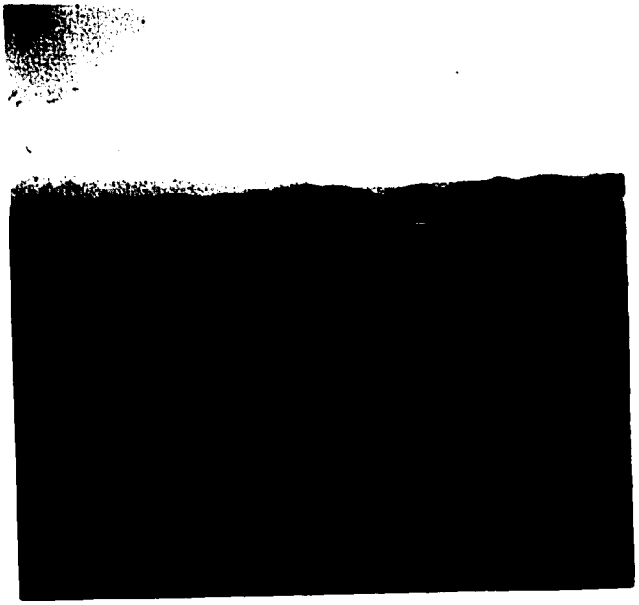


Figure 6.

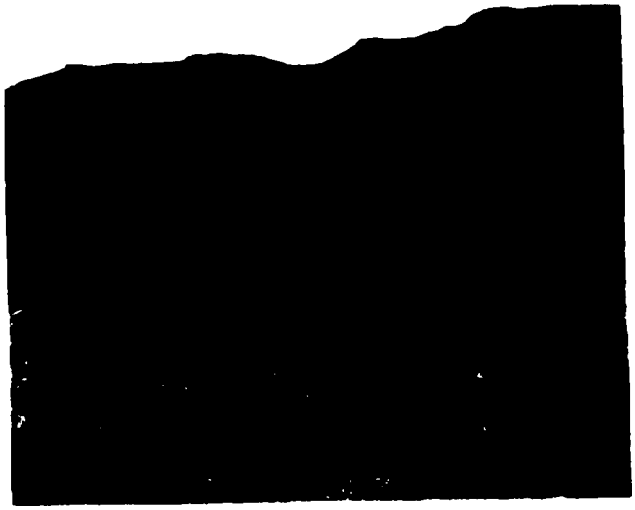


Figure 7.

correspond to the Wisconsin glacial stage terraces and that the upper terraces probably correspond to either the third or fourth lowest terrace level identified by Powers (1935), which he considers to be pre-Wisconsin but certainly Pleistocene in age.

MacNish (pers-comm.) believes that the terraces identified by Powers (1935) cannot be traced into the Arkansas Canyon and that the upper terraces were formed in early Wisconsin time and that the lower terraces were formed in late Wisconsin time.

DEFINITION OF ROCK UNITS

Precambrian Units

Precambrian igneous and metamorphic rocks are the predominant types underlying the area. Except for minor variants, the metamorphic rock types distinguished are:

biotite gneiss
- biotite schist and gneiss
- cordierite-biotite gneiss
- sillimanite-biotite gneiss
- muscovite-biotite gneiss

hornblende gneiss and hornblende amphibolite

calc-silicate gneiss

muscovite schist

These rocks are assigned to the Idaho Springs formation (Ball, 1906, 1908) on the basis of general composition, metamorphic grade, age relations to the major granitic intrusives, and continuity with rocks similarly assigned by workers in nearby areas (Fig. 1). Of the types listed, only the muscovite schist and some of the hornblende amphibolite and sillimanite-biotite gneiss occur in units large enough to be mapped

Figure 8. Upper and lower (Wisconsin) terraces at the mouth of Texas Creek Gulch.

Figure 9. Typical outcrop of biotite gneiss. East Gulch.



Figure 8.

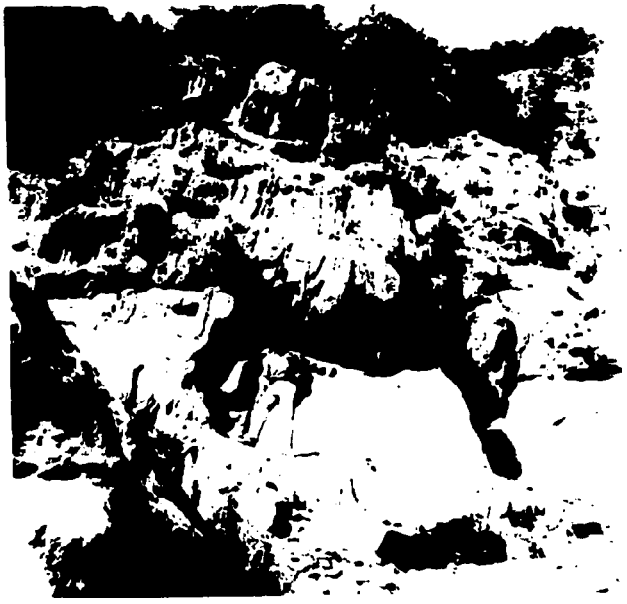


Figure 9.

separately (Plate I). Typically the biotite gneisses, hornblende-bearing rocks, and cale-silicate gneiss occur in thin, discontinuous, complexly interlayered and mutually gradational units; biotite gneiss commonly predominates. Therefore most of the Idaho Springs exposures were mapped as Idaho Springs formation, undifferentiated.

Part of the northwest corner of the area is underlain by granite gneiss (Plate I) which was intruded into or was formed by granitization of Idaho Springs rocks. This unit is tentatively correlated with the granite gneiss of Lovering and Goddard (1950) and the biotite gneiss of Bever (1954).

Three texturally distinct types of granite were recognized and mapped separately. The relation to or correlation with established granite types in the region is not equally clear for each type. Much of the west-central and southwestern part of the area is underlain by a coarse-grained, porphyritic, biotite granite that is identical with the Pikes Peak granite in its type area (Cross, 1894; Mathews, 1900) and to granites identified as Pikes Peak in adjacent areas (Fig. 1). Related granite dikes and sills and pegmatitic bodies are numerous and have locally formed migmatites with the surrounding metamorphic host rocks. A medium- to coarse-grained, nonporphyritic granite, tentatively correlated with the Cripple Creek granite, crops out east and southeast of Devils Hole. It has a strongly migmatitized aureole and is characterized by close-spaced fractures. Fine-grained, slightly porphyritic granite crops out along and west of Texas Creek and sills and dikes of similar granite intrude Pikes Peak granite and the Idaho Springs formation. The age relations and certain characteristic textural features of this rock indicate it is a Silver Plume-type granite (Lovering and Goddard, 1950).

The Idaho Springs rocks are nowhere entirely free of injected granitic material. Locally the injection and associated metasomatic effects have been so intense as to produce migmatites, here defined, for the purpose of mapping and description, as mixed rocks containing more than 50% added granitic material. On the geologic map (Plate I), areas of migmatite are differentiated with respect to the migmatitizing granite.

Diabase dikes of Precambrian age cut the metamorphic rocks and the granites. A few are large enough to be outlined on the geologic map, others are represented by an attitude symbol (Plate I).

Tertiary and Younger Rocks

The Parisita (?) clastics filling the Echo Canyon graben are best exposed along the steep walls of some of the gullies in Echo Canyon (Plate I).

The Tertiary volcanic sequence consists of the following units:

youngest 4. andesite (flows, flow breccia, and scoria)

3. volcanic conglomerate and tuffaceous sandstone

2. rhyolite tuff

oldest 1. basaltic agglomerate (restricted to Echo Canyon)

Units 2 and 4 are differentiated on Plate I. Owing to the fact that unit 3 crops out only on steep slopes held up by the cap rock effect of unit 4, it was mapped as part of unit 2. Unit 1 is largely covered by landslides and was not mapped separately.

Extensive landslides have occurred along the erosional escarpment at the north end of Echo Canyon. On the geologic map (Plate I), their upper contacts represent the change in slope along the tops of the slides and the lower contacts represent the downhill extent of material showing no evidence of water transport.

Much of Echo Canyon is filled with as much as 100 feet of unsorted debris, predominantly volcanic, derived from the landslides and talus slopes surrounding the canyon. This material, mapped as basin debris, covers most of the Farisita (?) clastics.

Travertine deposits, specifically travertine-cemented talus and alluvium, occur along some of the faults and shear zones. A few of these deposits are outlined on the geologic map (Plate I).

Quaternary terrace gravels along the Arkansas River were mapped separately. These gravels were deposited by the river during the Wisconsin glacial stage and they presently lie from 20 to 40 feet above the river.

Quaternary alluvium was mapped along the Arkansas River and some of the larger streams. Areas covered by talus or a thin blanket of poorly sorted alluvium were mapped as bedrock.

PRECAMBRIAN ROCKS

IDAHO SPRINGS FORMATION

Introduction

The following rock types occur in the Idaho Springs formation:

biotite gneiss

hornblende gneiss and hornblende amphibolite

calc-silicate gneiss

muscovite schist

The various types occur interlayered. Layer-units range in thickness from less than an inch to several hundred feet. Individual layer-units commonly display one of three types of banding; banding resulting from variation in original bulk composition, banding resulting from migmatitic injection, and miner metamorphic segregation banding. Compositional bands, typified by light-dark, biotite-poor and biotite-rich bands in biotite gneiss, are generally less than a foot thick but are relatively continuous. Segregation bands are rarely more than an inch thick and are not continuous. Layer-units are generally continuous, without marked pinching and swelling, at least on the scale of most outcrops. The general absence of marker beds hinders adequate study of continuity and structure. The contacts between both layer-units and compositional bands are gradational over short distances (commonly a fraction of an inch) and are defined by contrasting color, texture, and degree of development of the foliation.

Since the original description (Ball, 1906) most of the metamorphic rocks in central Colorado have been assigned to the Idaho Springs formation, which is generally considered to represent a thick sequence of early or early-middle Precambrian sediments that have been transformed

into moderately high grade regional metamorphic schists and gneisses. They are the eldest known rocks in the region. Migmatitization and metasomatism by granitic intrusives in many places have obscured much of the original chemical and structural characteristics of the rocks. Some of the criteria cited as evidence for sedimentary parentage have been: general parallelism (absence of cross-cutting relations) between layer-units; general parallelism of foliation and layer-unit contacts (bedding plane foliation); presence of quartzite and marble layers. The parent sediments are thought to have been largely impure sands, shales, and carbonate rocks deposited in a miogeosynclinal environment (see, for instance, Levering and Goddard, 1950; Shapirio, 1962).

In recent years United States Geological Survey geologists have adopted the policy of mapping and naming these rocks solely on the basis of lithology and have de-emphasized the use of formational names such as Idaho Springs and Swandyke, e.g., Sims and Gable (1964); Wells et al. (1964). The writer agrees that there is little proof of time-stratigraphic equivalence among units earlier assigned to the Idaho Springs formation in the central Front Range, let alone between those rocks and metamorphic units exposed in the Arkansas Canyon region. However, the writer will continue to use the term "Idaho Springs formation", as has been the custom in nearby areas (Fig. 1), to connote a thick sequence of predominantly metasedimentary rocks roughly similar (in lithology, metamorphic grade, and age relations to the major, and more positively correlatable granitic intrusives) to rocks in the vicinity of Idaho Springs in the central Front Range, the type area for the Idaho Springs formation (Ball, 1906; Bastin and Hill, 1917; Lovering and Goddard, 1950; Wells et at., 1964). Table 1 presents a suggested correlation of metamorphic rock units

Table 1. Possible correlation of Precambrian metamorphic rocks (central Front Range correlations from Wells et al., 1964).

This Report	Wells et al., 1964	Lovering and Goddard, 1950	Ball, 1906
Idaho Springs fm. biotite gneiss	biotite-quartz-plagioclase gneiss	Idaho Springs	Idaho Springs
hornblende gneiss	hornblende gneiss	Swandyke hornblende gneiss	hornblende gneiss
calc-silicate gneiss	calcium-silicate rocks	Idaho Springs	Idaho Springs
muscovite schist	mica schist and/or schist in quartzite	Idaho Springs	------

Table 2. Generalized stratigraphic sections.

Texas Creek Anticline		Antelope Gulch Anticline	
Lithology	Thickness	Lithology	Thickness
Interlayered biotite and hornblende gneisses	1300'	Interlayered biotite and hornblende gneisses	8000'
Muscovite schist	4700'	Sillimanite-biotite gneiss	3000'
Interlayered biotite and hornblende gneisses	3400'		

Table 3. Partial stratigraphic section in Hindman Gulch.

Unit	Thickness
biotite gneiss, medium gray*, grades upward into muscovite schist	10.5
hornblende gneiss	14.5
biotite gneiss, dark gray	0.6
hornblende gneiss	35.0
biotite gneiss, dark gray	5.6
hornblende gneiss	1.3
biotite gniess, dark gray	2.5
" , medium gray	8.7
" , dark gray	1.8
" , medium gray	2.5
" , dark gray	3.1
" , medium gray	5.0
granite sill	13.0
biotite gneiss, medium gray	1.0
" , dark gray	1.4
" , medium gray	1.0
" , dark gray	4.4
" , medium gray	4.6
" , dark gray	1.9
hornblende gneiss	2.6
biotite gneiss, medium gray	23.3
" , dark gray	5.0
" , medium gray	7.5
hornblende gneiss	6.7
biotite gneiss, medium gray	30.0
hornblende gneiss	2.5
biotite gneiss, light gray	11.5
hornblende gneiss	7.5
biotite gneiss, medium gray	7.5
shear zone	

*color reflects biotite content

between the central Front Range and the Arkansas Canyon.

The maximum thickness of the Idaho Springs formation in the Devils Hole area is approximately 11,000 feet. Sims and Gable (1964) report an estimated 13,500 feet maximum thickness in the Central City area and have presented a tentative stratigraphic section for the central Front Range. Very little is known of the stratigraphy in the Arkansas Canyon area. Two generalized sections, deduced from the succession of layers in the Texas Creek and Antelope Gulch anticlines, are given in Table 2. In addition, a section was measured in the most suitable area, Hindman Gulch; this is given in Table 3. It is presented to illustrate the kind of interlayering to which reference has so often been made.

Petrology

Biotite Gneiss

Occurrence. Biotite gneiss underlies nearly 30% of the area and is the most abundant and widespread unit of the Idaho Springs formation. It comprises the following rock types:

biotite schist and gneiss	50%
cordierite-biotite gneiss	1%
sillimanite-biotite gneiss	20%
muscovite-biotite gneiss	19%

The writer believes that the biotite schist was derived from an impure shale. The other rock types represent atypical bulk compositions or regional metamorphic assemblages or thermal metamorphic and metasomatic modifications of biotite schist as the result of granitic intrusions. Much of the biotite gneiss owes its banded character to the migmatitic and metasomatic introduction of felsic material along bands parallel with the schistosity.

Cordierite-biotite gneiss occurs in three areas: near the Pikes Peak contact in Sec. 11, T48N, R12E; in East Gulch, one-half mile northwest of the point where it enters Devils Hole; and south of the Arkansas River, one-half mile southwest of the mouth of Spike Buck Gulch. All three areas are "hot spots" with a granite mass or many pegmatites nearby. The cordierite may have formed in response to thermal rather than regional metamorphism (see pp. 72-80).

At least some of the sillimanite and muscovite occurring in Idaho Springs rocks has long been thought to have been formed by processes other than isochemical regional metamorphism (see, for instance, Saletti, 1960). Sillimanite-biotite and muscovite-biotite gneisses are widespread in the Devils Hole area. Both are most abundant in, but not confined to, zones near bodies of Pikes Peak granite and its associated granitic to pegmatitic sills and dikes. The writer considers these variants to have resulted mainly from Al and K metasomatism (see pp. 83-88).

Layers of biotite gneiss range from less than an inch to several hundred feet in thickness. The biotite gneisses are generally fine-grained, faintly to markedly banded, and strongly foliated (Fig. 9). The classification of these rocks as schists or gneisses is discretionary. Typically the banding reflects variation in the abundance and grain size of biotite, variation in the quartz-feldspar ratio, and, uncommonly, variation in the microcline-plagioclase ratio. The parallelism or subparallelism of biotite flakes produces the schistosity.

Petrography. The common essential minerals are quartz, biotite, and plagioclase. Microcline, cordierite, sillimanite, hornblende, and muscovite may be present in accessory to essential amounts and account for most of the major mineralogical variation. Common accessories are

apatite, magnetite, and zircon, less common ones are sphene, epidote, pyrite, tourmaline, garnet, and allanite. Modes of representative specimens are given in Table 4.

Relatively simple textures and mineral compositions were observed in most of the rocks studied microscopically. The minerals are quartz, plagioclase, and biotite. Xenoblastic quartz and plagioclase are interspersed with aligned biotite flakes (Fig. 10). Quartz grains show strain effects; plagioclase grains commonly display weak, normal zoning and a few simple twins. Grain sizes average 0.2 mm for quartz and plagioclase; biotite flakes average 0.5 mm across and 0.1 mm thick. Porphyroblasts of accessory magnetite or garnet may be present; these are commonly surrounded by a narrow, light-colored (biotite-poor) halo. Quartz-quartz grain boundaries are generally sutured and, where quartz is abundant, the sutured texture predominates. In a few of the rocks examined a high percentage of the quartz and plagioclase grains are idioblastic. Cataclastic features such as granulated or mylonitized zones and streaks and granulated grain boundaries are common only in rocks from near shear zones. Biotite gneiss appears to be recrystallized, with a marked increase in the biotite grain size, near the margins of some pegmatitic bodies and quartz veins.

These relationships are modified or nearly obliterated in much of the biotite gneiss. These modifications are discussed below, relative to the mineral produced during the modification.

Quartz replaces plagioclase, microcline, and biotite. This secondary quartz may be present as thin, intergranular films, as relatively large (2 mm in diameter) grains, or as any stage in between. The larger quartz grains are very commonly poikiloblastic and appear to have

Table 4. Modes of biotite gneiss.

	1.	2.	3.	4.	5.	6.	7.	8.
Quartz	27.4	48.0	39.3	52.6	34.3	51.4	36.2	4.0
Plagioclase	55.7 (Ab_{75})	25.8 (Ab_{60})	28.9 (Ab_{90})	1.0 (Ab_{90})	34.9 (Ab_{94})	9.4 (Ab_{55})	31.7 (Ab_{78})	25.2 (Ab_{70})
Microcline	0.4	---	20.7	1.9	12.7	3.1	---	33.8
Biotite	14.9	26.0	6.1	15.0	8.6	13.7	18.7	24.0
Muscovite	0.2	---	4.1	27.1	8.4	9.1	---	---
Sillimanite	---	---	---	1.6	---	13.3	0.2	---
Cordierite	---	---	---	---	---	---	6.9	---
Hornblende	---	---	---	---	---	---	---	12.4
Accessories	---	0.2	0.9	0.8	1.1	---	6.3	0.6

1. Biotite gneiss, mouth of Fernleaf Gulch.
2. Biotite gneiss, Texas Creek Gulch, 2½ miles northwest of Texas Creek.
3. Biotite gneiss, mouth of Hindman Gulch.
4. Muscovite-biotite gneiss, Texas Creek Gulch, 2½ miles west-northwest Texas Creek.
5. Muscovite-biotite gneiss, railroad cut ¼ mile west of Hindman Gulch.
6. Sillimanite-biotite gneiss, Hindman Gulch, 2 miles above mouth.
7. Cordierite-biotite gneiss, NE¼, Sec. 11, T48N, R12E.
8. Hornblende-biotite gneiss, Texas Creek Gulch, 2 miles northwest of Texas Creek.

Figure 10. Photomicrograph of biotite schist. Field includes biotite (B), quartz (Q), and plagioclase (P). Texas Creek Gulch. + polars. 43x.

Figure 11. Photomicrograph of biotite gneiss with secondary microcline. Field includes microcline (M), quartz (Q), biotite (B), and plagioclase (P). Fernleaf Gulch. + polars. 43x.

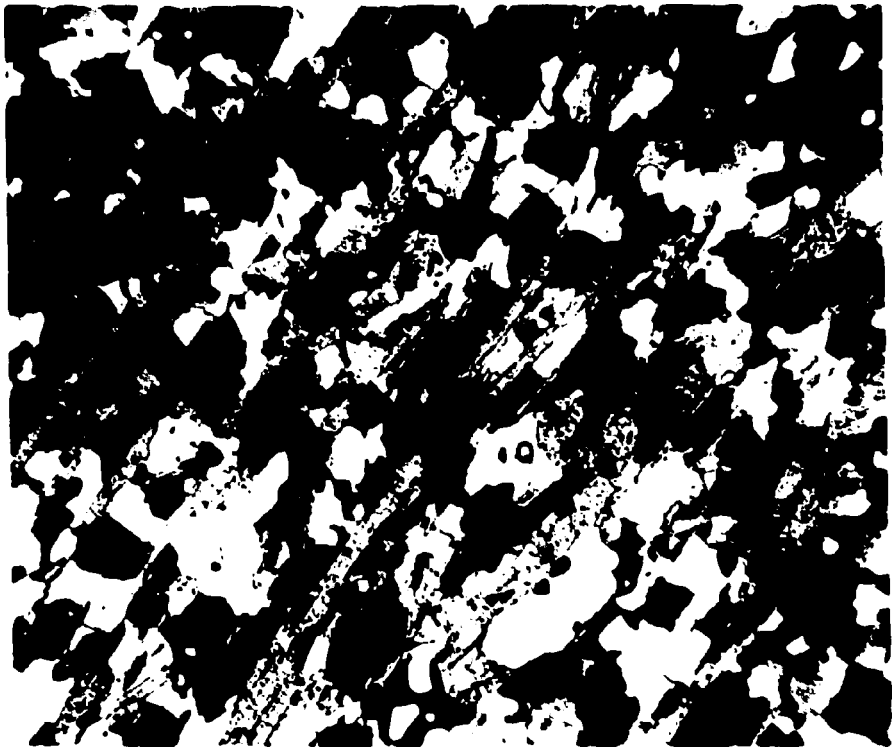


Figure 10.

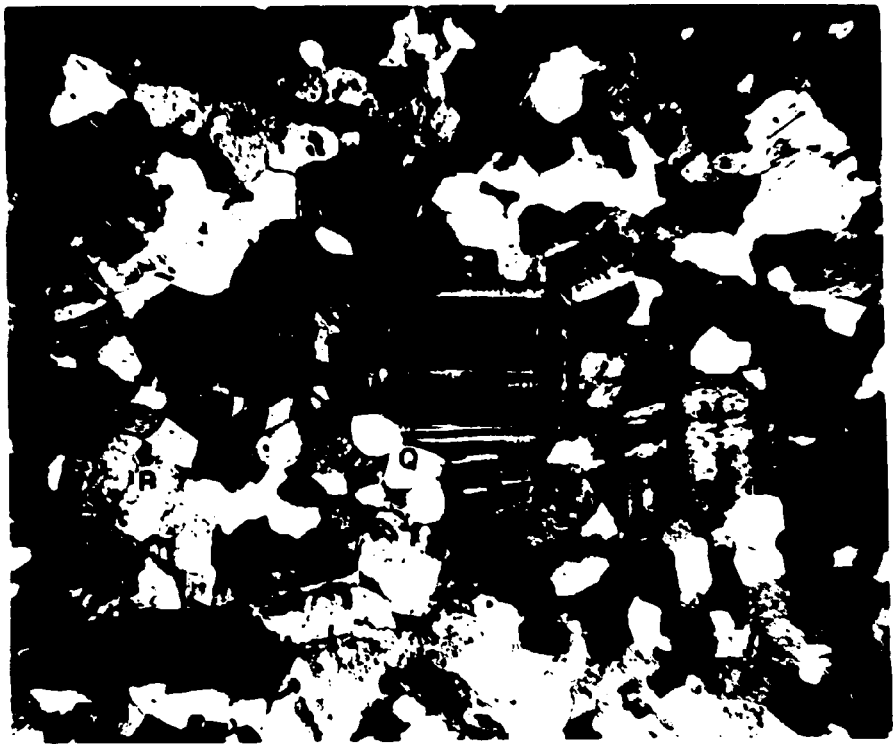


Figure 11.

engulfed preexisting mineral phases. In a few places secondary quartz has replaced nearly all of the original metamorphic rock. Secondary quartz most commonly occurs near granitic bodies and, although some of the observed replacement relations were caused by the remobilization of quartz already in the rock, most of the quartz has resulted from the metasomatic addition of silica from these bodies.

A small percentage of these rocks contains relatively large (as much as 2 mm in diameter), xenoblastic plagioclase (Ab_{90-94}) grains. These grains are probably metacrysts because they replace older, finer-grained placioclase (Ab_{60-80}), quartz, and biotite and are commonly poikiloblastic.

The variation in the abundance and textural relations of microcline accounts for much of the variation in the biotite gneiss. In some rocks microcline occurs in a granoblastic aggregate along with quartz and plagioclase, although in some rocks it is restricted to the felsic bands. With a granoblastic texture all of the microcline contact relations with other minerals are mutual and it appears to be a primary metamorphic phase. On the other hand, the presence of microcline as the result of migmatitization of the Idaho Springs rocks is very common and very strongly supported by textural evidence and field relations. Secondary microcline replaces primary metamorphic quartz, biotite, and plagioclase (Fig. 11), secondary quartz and plagioclase, and granitic quartz and plagioclase present in intrusive (migmatite) bodies in the biotite gneiss. Secondary microcline may be perthitic. Myrmekite has developed along with secondary microcline near granitic bodies. Plagioclase commonly shows strong, general alteration to sericite where secondary microcline is abundant. The presence of abundant microcline commonly results in a very uneven texture and microcline-microcline contacts may be very

complex. Microcline metacrysts are xenoblastic and commonly poikiloblastic. Secondary microcline occurs disseminated, as scattered metacrysts, in multigranular augen, or in irregular patches. It is clear that much of the microcline has resulted from metasomatism, principally the introduction of K, by granitic intrusives.

Cordierite is present as single crystals (porphyroblasts) and as multigranular aggregates; both forms are blocky or ellipsoidal with their long dimensions in the plane of the foliation (Fig. 12). Cordierite individuals average 1 mm and range up to 5 mm in diameter. Aggregates of grains range up to 15 cm in length. All cordierite grains are highly poikiloblastic and average only 47% cordierite; inclusions are quartz, plagioclase, biotite, apatite, and magnetite.

Sillimanite occurs as fibrolitic masses in granular quartz aggregates (Fig. 13) and as scattered needles in quartz. The aggregates contain as much as 70% sillimanite. They range in shape and size from phacoids a centimeter or less across and a few millimeters thick (Fig. 14) to ellipsoidal (Fig. 15) or round (Fig. 16) nodules which range up to 16 cm in length (Fig. 15). The phacoids or nodules may compose from 10 to 60% of the sillimanite-biotite gneisses. The long dimensions of the ellipsoidal aggregates lie in the plane of the foliation. In some rocks sillimanite clearly replaces biotite, and feldspar is less abundant or altogether lacking in and near the sillimanite aggregates; in others the textural relations of sillimanite are not clear, chiefly because secondary muscovite has obliterated them. Replacement of sillimanite by muscovite is common.

Muscovite characteristically occurs as large (up to 1 cm long), poikiloblastic plates which have replaced or show replacement relations

Figure 12. Photomicrograph of cordierite-biotite gneiss. Field includes poikiloblastic cordierite porphyroblast (C), quartz (Q), biotite (B), and plagioclase (P). From near Pikes Peak contact in Sec. 11, T48N, R12E. + polars. 43x.

Figure 13. Photomicrograph of sillimanite (S) -quartz (Q) intergrowth in sillimanite-biotite gneiss. Hindman Gulch. + polars. 43x.

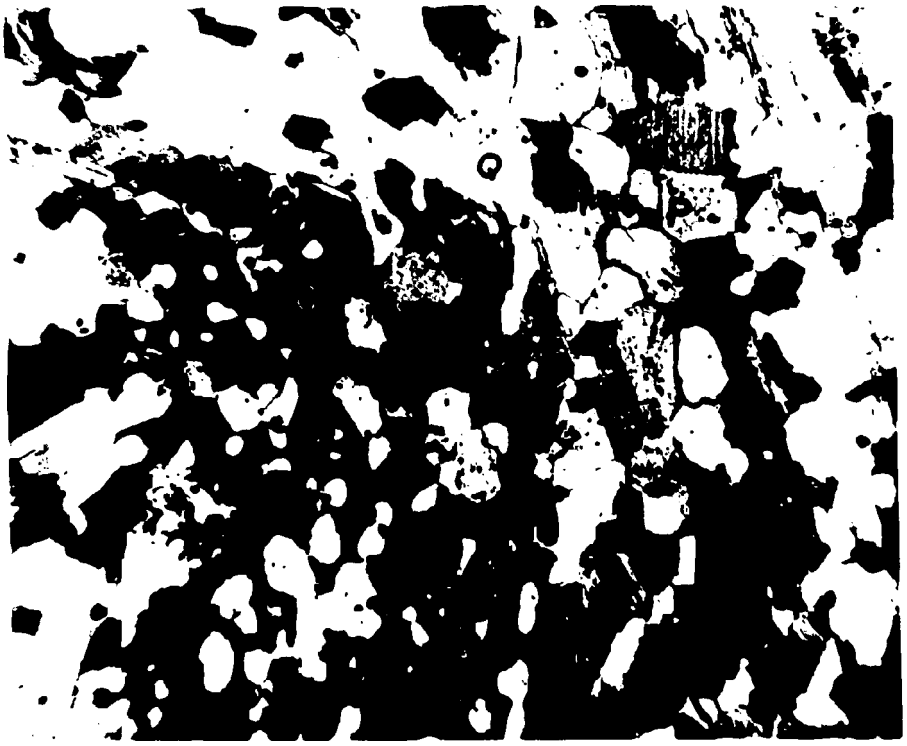


Figure 12.



Figure 13.

Figure 14. Sillimanite-biotite gneiss; phacoidal sillimanite-quartz intergrowths. East Gulch.

Figure 15. Sillimanite biotite gneiss; sillimanite-quartz intergrowths as large, ellipsoidal nodules. Upper Bull Gulch.



Figure 14.



Figure 15.

Figure 16. Sillimanite-biotite gneiss; sillimanite-quarts intergrowths as round nodules. East Gulch.

Figure 17. Hornblende gneiss layer in biotite gneiss. East Gulch.



Figure 16.



Figure 17.

with biotite, microcline, plagioclase, sillimanite, and, to a lesser extent, quartz. Muscovite flakes commonly form symplectitic intergrowths with quartz at their margins, especially against microcline grains. Muscovite flakes rarely exhibit any orientation, or only that inherited from replaced biotite and sillimanite. Muscovite is most common in rocks that contain, or once contained, sillimanite. No clearly primary metamorphic (i.e. strictly metamorphic) muscovite was observed.

Alteration products in biotite gneiss include chlorite after biotite, sericite and kaolinite after plagioclase, and hematite after magnetite and pyrite. Chlorite, kaolinite, and hematite are especially abundant in and near shear zones.

Origin. The sedimentary parentage of the Idaho Springs formation is well-established. The biotite gneisses were probably pelitic rather than quartzo-feldspathic assemblages. Other workers have presented ample evidence that the bulk composition of Idaho Springs biotite gneiss falls within the compositional range of pelitic rocks (Bever, 1954; Salotti, 1960; Shappirio, 1962). This has been done by averaging modes of biotite gneiss, converting the modes to chemical analyses, and comparing with Clarke's (1924) data for shales. The somewhat higher silica values for biotite gneiss obtained in this manner indicate that the parent shales were generally silty or sandy or possibly that some migmatitic quartz was included in the modes.

The equilibrium assemblage for biotite schist and gneiss is quartz - plagioclase (Ab_{70}) - biotite ± microcline ± almandite. This assemblage is most common for rocks of quartzo-feldspathic or pelitic parentage in the almandine-amphibolite facies (Fyfe et al., 1958). Critical minerals that would allow a more positive facies of subfacies assignment are

lacking. The absence of staurolite and kyanite and the rarity of garnet may be explained by the abundance of potassium (Fyfe et al., 1958). The common interlayering of biotite gneiss with sillimanite-biotite gneiss may indicate that these rocks were formed in the higher intensity part of the almandine-amphibolite facies. Thus Sims and Gable (1964) assigned similar rocks in the Central City district to the sillimanite-almandine subfacies. Wells et al. (1964) cite a sillimanite-muscovite assemblage and the less abundant sillimanite-microcline suite as evidence that the schists of the Coal Creek area were metamorphosed under intensities corresponding to the sillimanite-almandine-muscovite subfacies (Turner and Verhoogen, 1960). The interlayering of sillimanitic and non-sillimanitic units would reflect not a stratigraphic alternation in regional grade but a varying Al content of the parent sediments. The latter can be illustrated on an ACF diagram (Fig. 26) by varying the bulk composition between fields 1 and 2. Salotti (1960) rejected this hypothesis as an explanation of similar rocks in the Cotopaxi - Howard area. His main objection was that some sillimanitic rocks are too Al-rich to have been derived by isochemical metamorphism of any known (but not sillimanitic!) type of Idaho Springs rock. He also maintains that, "the close association of sillimanite rich rocks and intrusive granitic rocks is certainly more than fortuitous," (Salotti, 1960, p. 56).

The writer rejects a stratigraphic alternation of metamorphic grade as implausible. The interlayering of sillimanitic and non-sillimanitic rocks must be explained either as a result of varying bulk composition under constant regional grade or as a result of stratigraphically controlled, selective Al metasomatism. It is the writers experience that a large proportion of the sillimanitic rocks contains very little silliman-

ite, even counting that now replaced by muscovite, and would not yield alumina values much higher than would the biotite schists. However, certain layers are too rich in sillimanite to have been derived isochemically from anything but a nearly pure clay shale, a lithotype not in accord with the proposed sedimentary environment. Furthermore, it should be pointed out that there are, apparently, some differences between the sillimanitic rocks of the Arkansas Canyon area and those of the central Front Range. For instance, Wells et al. (1964) described the sillimanite as occurring disseminated, strongly oriented, and displaying, in places, mutual textures with muscovite and microcline. In contrast, sillimanite in the Devils Hole area occurs almost invariably concentrated as phacoidal to nodular intergrowths with quartz, shows very little preferred orientation, and does not display mutual textures with muscovite or microcline. The sillimanite-quartz aggregates appear to have grown by replacement after the relaxation of strongly directed pressure. Thus Al metasomatism has probably been an important factor in the development of sillimanite; the problem is further discussed in the section on metasomatism.

The presence of cordierite-bearing assemblages does not clarify the problem of assessing regional grade. Cordierite is stable over a wide range of temperature and pressure (Schreyer and Yoder, 1959), and although it is commonly considered as a thermal metamorphic phase it does occur in regional metamorphic suites (Eskola, 1952; Wynne-Edwards and Hay, 1963). The writer reserves the discussion of this question for the section on the origin of the cordieritic rocks.

Hornblende Gneiss, Hornblende Amphibolite, and Related Rocks

Occurrence. Rocks containing essential amphibole (predominantly horn-

blende) exhibit a considerable variation in field occurrences, textures, and composition. Typically they are interlayered with biotite gneiss and occur in layers which range from a few inches to over 100 feet in thickness. Amphibole-bearing rocks also form massive bodies that are conformable with the enclosing rock units only in a general way. In a few occurrences bodies of amphibolite transect the foliation and contacts of Idaho Springs units. These are possibly of igneous parentage.

Petrography. The interlayered amphibole-bearing rocks are of three types: 1) biotite-hornblende gneiss; 2) hornblende gneiss; and 3) hornblende amphibolite (Fig. 17). These rocks, together with those from the massive bodies, constitute 20% of the Idaho Springs formation and underlie about 8% of the area. Interlayered types commonly exhibit faint mineralogical banding and decussate arrangement of prismatic hornblende grains. Microscopically, uneven granoblastic textures are more common than porphyroblastic ones.

The equilibrium assemblages corresponding to the three interlayered rock types are: 1) hornblende-biotite-plagioclase-quartz; 2) hornblende-plagioclase-quartz; and 3) hornblende-plagioclase (columns 1, 2, and 3, Table 5). Other minerals which may be present in accessory to essential amounts are diopside, diopsidic augite, epidote, and microcline. Diopside and epidote are present in types gradational to calc-silicate gneiss. Diopsidic augite is present in an amphibolite layer near Pikes Peak granite. All of the microcline, much of the quartz, and some of the plagioclase exhibit replacement relations to minerals in the equilibrium assemblages and appear to be present as the result of migmatitic injection and related metasomatic effects. These secondary species produce the same textural variation in the hornblende-bearing rocks as they do

Table 5. Modes of hornblende-bearing rocks.

	1.	2.	3.	4.	5.
Quartz	5.9	6.0	tr	---	3.3
Plagioclase	37.0 (Ab_{67})	54.0 (Ab_{60})	35.0 (Ab_{54})	52.5 (Ab_{66})	44.4 (Ab_{51})
Hornblende	46.7	37.0	63.8	44.8	50.7
Biotite	9.4	---	---	---	1.1
Accessories	1.0	3.0	1.2	2.7	0.5

1. Biotite-hornblende gneiss, interlayered sequence, railroad cut 1 mile east of Texas Creek.

2. Hornblende gneiss, interlayered sequence, mouth of Fernleaf Gulch.

3. Hornblende amphibolite, interlayered sequence, Texas Creek Gulch, 1¼ miles north-northwest of Texas Creek.

4. Massive amphibolite, Texas Creek Gulch, 2½ miles southeast of Table Mountain.

5. Cross-cutting amphibolite, 2 miles north of Texas Creek.

in biotite gneiss (pp. 31-36). Sphene, magnetite, and apatite are common accessory minerals; scherl is present in specimens collected near pegmatite bodies. Plagioclase grains exhibit normal zoning (typically through a range of about 10 mol. % from Ab_{50-60} to Ab_{60-70}) and are commonly more complexly twinned than in biotite gneiss. Hornblende grains tend to be idioblastic and typically show a wide range in grain size, commonly from 0.1 mm to 2 mm along the C-axis, and are coarser (average 0.5 mm along the C-axis) than other constituents (plagioclase grains average 0.2 mm in diameter). The larger hornblende grains may be poikiloblastic and are commonly so in biotite-hornblende gneiss.

Along with the introduction of microcline, quartz, and minor plagioclase, hornblende and, to a lesser extent, biotite have recrystallized to form large (as much as 2 cm in diameter), poikiloblastic anhedra along contacts with granite dikes and sills. Alteration products include sericite and epidote after plagioclase, chlorite after biotite, and biotite after hornblende.

Amphibolite is the predominant rock type where the hornblende-bearing rocks form massive rather than interlayered bodies. The massive bodies tend to predominate over relatively large areas and are the only amphibole-bearing rocks that were mapped separately (Plate I). The massive character of these rocks is not ubiquitous, but is interrupted by variations in the texture of the amphibolite or by layers and lenses of hornblende gneiss, biotite gneiss, and calc-silicate gneiss. The amphibolite is commonly without megascopic or microscopic foliation, either by mineralogical banding or grain orientation. The typical texture has aggregates of predominantly idioblastic, interpenetrating hornblende prisms set in a granular matrix in which xenoblastic (rarely

idioblastic) plagioclase predominates over xenoblastic to idioblastic hornblende. The hornblende aggregates are generally blocky to ellipsoidal and only rarely do the long axes of the aggregates display a preferred orientation. Hornblende grains range from less than 0.1 mm to 2 mm along their C-axes, generally with the smaller grains in the matrix and the larger grains in the aggregates. The aggregates commonly range from 1 mm to 5 mm, rarely to 15 cm, on their long diameters (Fig. 18). Plagioclase grains average 0.2 mm in diameter. A variant of this rock type has large (up to 6 mm) plagioclase porphyroblasts scattered randomly in a hornblende-rich matrix in which the hornblende grains are not strongly aggregated. Uncommon types display weak mineralogical banding or have uneven granoblastic textures wholly without mineral segregations.

Hornblende and plagioclase constitute the equilibrium assemblage and, with rare diopsidic augite, are the only minerals present in essential amounts (column 4, Table 5). Diopsidic augite, present in an amphibolite body near Pikes Peak granite, occurs as scattered subhedral to anhedral grains that appear to have grown, in part, by replacing hornblende and have subsequently been partly replaced by hornblende. Sphene and magnetite are common accessories, apatite and titano-magnetite are less common. Much of the plagioclase is sericitized or saussuritized.

Other amphibole-bearing rocks, although similar in mineralogy and texture or occurrence to the massive amphibolites described above, have characteristics which warrant their receiving separate consideration. One of the rocks examined from the area of massive amphibolite southeast of Table Mountain contains remnant portions of augite grains within aggregates of replacing hornblende grains, which suggests a relict ig-

Figure 18. Hornblende amphibolite with large hornblende aggregates. Long Gulch, SW¼, Sec. 21, T49N, R12E.

Figure 19. Photomicrograph of hornblende amphibolite from massive amphibolite 1½ miles southeast of Table Mountain. Remnant augite (A) grains within aggregate of later hornblende (H). Plagioclase (P) completes the field. Blastoglomeroporphyritic texture? 43x.



Figure 18.

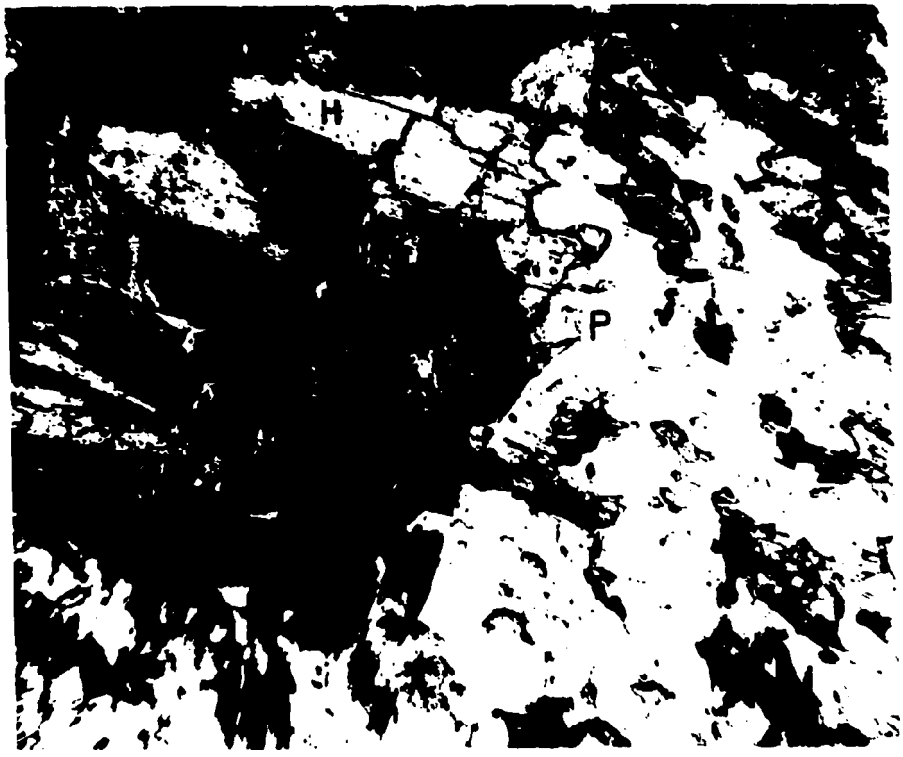


Figure 19.

neous (glomeroporphyritic) texture (Fig. 19). A nearly identical rock crops out a half mile east-northeast of the Texas Creek Post Office. Associated with the augite-bearing amphibolite in this area are rocks that grade through one in which the hornblende is largely replaced by actinolitic hornblende and the plagioclase is almost completely saussuritized to a massive actinolitic hornblende - chlorite rock. Apparently the latter was derived from the augite-bearing amphibolite through local retrogressive metamorphism. The retrogressive effects increase to the east-southeast along the direction of strike in the enclosing foliated rocks. Cataclastic effects also increase because the Texas Creek fault zone is approached in that direction. The writer believes that the retrogressive changes are related to the fault zone and are Precambrian in age. The related cataclastic effects range in age from Precambrian to Laramide.

The augite-bearing amphibolites and related greenschists provide textural and mineralogical evidence that some of the amphibole-bearing rocks may have had an igneous parentage. Evidence of a more convincing nature is provided by the previously mentioned bodies of amphibolite that transect the foliation and contacts in the enclosing Idaho Springs rock units (Fig. 20).

The cross-cutting amphibolites are very similar to the massive amphibolites. They contain essential hornblende and plagioclase and accessory sphene, titano-magnetite, biotite, and pyrite (column 5, Table 5). In one, the texture is granoblastic with only slight clustering of hornblende grains; in another, aggregates of hornblende grains occur in a plagioclase-hornblende matrix. The possibility of a meta-igneous origin for some of the amphibole-bearing rocks will be discussed

in more detail below (pp. 56-58).

It is interesting to note that the pyroxenes present in the hornblende-bearing rocks are distributed in the following manner:

1) augite ($\gamma \wedge c$ = ca. 50°) occurs as remnants in what may be a blastoglomeroporphyritic texture that is also found in cross-cutting amphibolite bodies;

2) diopsidic augite ($\gamma \wedge c$ = ca. 45°) occurs irregularly distributed in rocks close to the Pikes Peak granite and may have formed in response to locally higher temperatures;

3) diopside ($\gamma \wedge c$ = ca. 40°) is present in lenses and layers of calc-silicate gneiss.

Cummingtonite-Hornblende Amphibolite. Cummingtonite is present in foliated amphibolites in the areas of predominantly massive amphibolite southeast and northwest of Table Mountain. The foliation is produced by weak to strong mineralogical banding and by the decussate arrangement of both hornblende and cummingtonite. Essential minerals are hornblende, plagioclase, and cummingtomite; accessory species are magnetite, rutile, apatite, and quartz. The cummingtonite occurs in the following textural relations: as grains having the same distribution as hornblende grains, although smaller in size and less abundant; in crystallographic continuity with hornblende in the same grain, the domains of the two phases being separated by irregular surfaces or by planes parallel with (100) and (001) (Fig. 21); as grains showing marginal overgrowths of hornblende or replacement by hornblende along cleavages and fractures. Thus the formation of hornblende appears to have occurred during and continued after the formation of cummingtonite. Bose (1959) has described the bleaching of hornblende along planes and zones parallel to (001). The colorless

Figure 20. Hornblende amphibolite body transecting foliation and layering in biotite gneiss. Near railroad, ½ mile northeast of the mouth of East Gulch.

Figure 21. Photomicrograph showing cummingtonite (C) lamellae along (001) in hornblende (H) from cummingtonite-hornblende gneiss, 1½ miles southeast of Table Mountain. Also note the less regular intergrowths. Plagioclase (P) and magnetite (black) complete the field. 330x.



Figure. 20



Figure 21.

hornblende he describes has higher indices than the colored hornblende, contrasts with it in interference colors, and is optically negative. The colorless amphibole lying along (001) planes in hornblende in the Devils Hole rocks is optically positive, has lower indices, but does contrast somewhat in interference colors. It appears identical to the granular cummingtonite and the writer must consider it to be cummingtonite. Vernon (1962) described very similar cummingtonite-hornblende relations in an amphibolite from Duchess, Queensland, Australia. Shappirio (1962) found similar cummingtonite-hornblende relations, with the exception of the intergrowths parallel to (001), in the Tallahassee Creek area. In addition, he found exsolved hornblende lamellae parallel with (100) in cummingtonite.

Cummingtonite and hornblende are partly replaced by light brown biotite. Other secondary materials are sericite, epidote, and clay, all after plagioclase.

Origin. Both igneous and sedimentary parents have been suggested for hornblende gneisses and amphibolites (see summary by Bever, 1954). The name Swandyke hornblende gneiss (Lovering, 1935) has been applied to such ortho-gneisses within the Idaho Springs formation in Colorado. In the Montezuma quadrangle (Lovering, 1935) the Swandyke gneiss consists of a thick series of metamorphosed basaltic to dioritic flows and sills, largely overlying but in part interlayered with upper Idaho Springs units. Bodies of hornblende gneiss and amphibolite that show discordant relations to the foliation and contacts in Idaho Springs rocks have also been assigned to the Swandyke gneiss (Lovering, 1935). The Swandyke gneiss is thought to be younger than the Idaho Springs formation but still pre-metamorphic in age.

It also has been proposed that hornblende gneiss and amphibolite units derived from sedimentary parents form an important part of the Idaho Springs formation in several places (Bastin and Hill, 1917; Tweto, 1947; Scott, 1963; Sims and Gable, 1964).

In recent years the Swandyke hornblende gneiss concept has been largely discarded by United States Geological Survey geologists working in the central Front Range (Harrison and Wells, 1959; Sims and Gable, 1964; Wells et al., 1964). They propose a sedimentary parentage for most of the hornblende gneiss and amphibolite, although they do not exclude the possibility of a metavolcanic origin for some of these rocks.

In the Devils Hole area both genetic types of amphibole-bearing rocks may be present. The interlayered amphibole-bearing rocks appear to be part of the metasedimentary sequence for the following reasons:

1) conformity with layering and foliation in the biotite gneiss.
2) similarity of bedding features to those of biotite gneiss (thickness, extent, contacts).
3) presence of rock types gradational between hornblende gneiss and biotite gneiss.
4) close association with calc-silicate gneiss.
5) lack of relict igneous textures or structures.

The cross-cutting amphibolites and minor amounts of massive amphibolite which contain what appear to be relict igneous textures are possibly meta-igneous in origin.

Most of the massive amphibolites show some of the characteristics and associations of both the interlayered (metasedimentary) and the possibly meta-igneous types. The writer favors assigning most of the massive amphibolites to the Idaho Springs formation because:

1) they are generally conformable with Idaho Springs units;

2) they contain layers of biotite gneiss and calc-silicate gneiss;

3) they lack relict igneous textures and structures.

The fact that a minor amount of the massive amphibolite and some associated greenschists appear to be meta-igneous does not warrant ascribing all of the massive amphibolite to such an origin.

The equilibrium assemblage for the hornblende gneisses and hornblende amphibolites is plagioclase (Ab_{60}) - hornblende ± quartz ± biotite ± diopside ± epidote ± cummingtonite. This assemblage is characteristic of the moderate to high intensity subfacies of the almandine-amphibolite facies and could have been derived from a variety of parent materials (Fyfe et al., 1958). Because field relations favor a sedimentary parentage for the interlayered hornblende gneiss and amphibolite and much of the massive amphibolite, they were probably formed from impure sediments such as calcitic, dolomitic, or sideritic mudstones. Variations in the original carbonate-clay-quartz ratio could account for the mineralogical variation in the present rocks. This can be illustrated in an ACF diagram (Fig. 26) by varying the bulk composition among fields 3, 4, and 5.

The rare cummingtonite-bearing hornblende amphibolites were derived from locally magnesium-rich sediments, the required excess of magnesium present in the clay-chlorite fraction. No other origin for the rock is suggested by either field or microscopic relations. The same interpretation has been advanced for similar rocks in nearby areas (Shappirio, 1962; Dahlem, 1965).

Anthophyllite-Cordierite Gneiss

Anthophyllite-cordierite gneiss crops out on the ridge east of Echo

Canyon. It occurs as a thin layer in sillimanite-biotite gneiss. The rock contains essential amounts of anthophyllite, cordierite, and light brown biotite. The cordierite is present as large anhedra as much as 1.2 cm in diameter. Anthophyllite forms slender prisms that range up to 5 mm in length and occur both scattered and in clusters. The only foliate feature of the rock is the general alignment of strings of apatite grains and elongate magnetite grains. Biotite occurs as unoriented flakes up to 1 mm across and has replaced much of the anthophyllite. Other secondary species are sericite and quartz after cordierite and chlorite after biotite and anthophyllite.

Possible origins for this rock are discussed in the section on the origin of the cordieritic rocks.

Calc-silicate Gneiss

Occurrence. Calc-silicate gneiss is widespread but constitutes less than 5% of the Idaho Springs formation. Over 90% of its occurrences are within hornblende gneiss and hornblende amphibolite, the rest are within biotite gneiss. Calc-silicate gneiss forms pods, lenses, and bands that are regular in outline and orientation where foliation in the enclosing rocks is well developed and are irregular where foliation is poorly developed. These structures are interpreted to result from the stretching and plastic flowage of the incompetent carbonate beds from which these rocks were derived (Bever, 1954; Salotti, 1960). Individual pods and lenses are generally less than a foot across and a few inches or less in thickness; bands range from less than an inch to several feet in thickness and are rarely traceable for more than ten feet along strike. However, zones of hornblende-bearing rocks that contain more than 10% calc-silicate gneiss may be as much as 300 feet thick.

Petrography. Contacts with enclosing rocks range from sharp to diffuse. Internally, the calc-silicate bodies may be texturally and mineralogically homogeneous, banded, or irregular. In homogeneous and banded types grain size averages range from 0.2 mm to 1 mm; in irregular types coarse-grained pockets, which generally occur in the central portions of thicker lenses and layers, may contain garnet and calcite crystals several centimeters across. A characteristic feature of most occurrences is extreme and abrupt variations in grain size. Both granoblastic and porphyroblastic textures are common. Garnet and calcite are more commonly segregated into bands or pockets than are other minerals of the calc-silicate gneisses.

The characteristic minerals, and the only coarse-grained ones, are diopside, epidote, garnet, calcite, and quartz; these occur in various combinations, i.e., diopside alone, diopside-epidote-quartz, diopside-epidote-garnet-quartz, all together. Sphene is very common; apatite and magnetite may be present; scapolite is rare. Porphyroblastic species are garnet and epidote, both of which are commonly poikiloblastic. Hornblende and plagioclase are present where contacts are diffuse, but the hornblende is commonly bleached and the plagioclase badly sericitized and partly replaced by epidote. Replacement textures are common. Hornblende and diopside appear to be mutual in general, but diopside commonly has hornblende overgrowths. Epidote replaces garnet, diopside, hornblende, and plagioclase. Quartz and calcite appear to be later than other minerals in some occurrences. Augite, instead of diopside, is present in rocks subjected to considerable migmatitization.

Origin. Calc-silicate assemblages contain one or more of the following; diopside, epidote, garnet, quartz, calcite, and may contain hornblende

and plagioclase (oligoclase-andesine). These assemblages were derived from impure calcareous or dolomitic sediments and are characteristic of the almandine-amphibolite facies (Fyfe et al., 1958). The presence of epidote rather than calcic plagioclase may indicate that the intensity at which equilibrium was most nearly achieved was somewhat below that of the sillimanite-almandine subfacies. On the other hand, epidote rarely appears in mutual textures with other phases, especially plagioclase; rather, it appears to be a later, perhaps retrograde, phase.

The skarn development in calc-silicate rocks in nearby areas (Bever, 1954; Salotti, 1960; Shappirio, 1962) was not observed in the Devils Hole area.

Muscovite Schist

Occurrence. Muscovite schist crops out along a belt extending northwestward from a shear zone in Spike Buck Gulch two miles south of the Arkansas River to a shear zone and the Cripple Creek (?) granite contact south of Devils Hole, a distance of about four miles (Plate I). The unit, characterized by abundant muscovite and a well-developed schistosity, is the most continuous of all the Idaho Springs units. The muscovite schist can be subdivided into three parallel layers of contrasting mineralogy and texture (Plate I). From southwest to northeast the layers, and their thicknesses in Hindman Gulch, are:

sillimanite-muscovite schist	710'
nodular sillimanite-muscovite schist	2780'
nodular cordierite-muscovite schist	1280'

The muscovite schist lies conformably between zones of interlayered biotite gneiss and hornblende gneiss; it also contains a few conformable layers of biotite gneiss. Granitic intrusives form about 25% of the unit.

Contacts between the muscovite schist variants are gradational over a few tens of feet; contacts with other Idaho Springs rocks are gradational over a few feet where observable but are generally obscured by shear zones.

Petrography. Sillimanite-muscovite schist contains essential quartz, muscovite, and biotite; sillimanite and plagioclase are present in accessory to essential amounts (see column 1, Table 6). Characteristic accessories are magnetite, apatite, zircon, and schorl. Magnetite commonly occurs as strings of grains parallel to the foliation. The rock is very uneven-grained. Near the southwestern contact of the unit plagioclase and some of the biotite and quartz occur in a granoblastic texture similar to that in biotite gneiss, but this association is minor, having been nearly completely destroyed by recrystallization of biotite and quartz and replacement by quartz, sillimanite, and muscovite.

The sillimanite is fibrolitic, and alignment of bundles of needles contributes to the foliation of the rock. Sillimanite occurs intergrown with quartz. Quartz grains range up to 2 mm in diameter and may be flattened in the plane of foliation; quartz-quartz grain boundaries are sutured. Quartz may be very abundant; a few thin quartzite layers are present in Hindman Gulch. Most of the biotite occurs as large flakes (up to 5 mm across) that are subparallel and contribute to the foliation. Biotite may poikiloblastically include quartz and plagioclase. The coarse, recrystallized biotite appears to have formed earlier than the sillimanite-quartz intergrowths because both sillimanite and quartz replace biotite. Muscovite is present as flakes as much as 5 mm across that are generally aligned in the plane of foliation. Some of it is later than and has replaced biotite and sillimanite, but much of it appears to be a primary metamorphic phase, like the biotite in

Table 6. Modes of muscovite schist.

	1.	2a.	2b.	3a.	3b.
Quartz	22.5	33.7	52.0	21.0	35.0
Muscovite	62.2	0.7	31.0	0.7	45.0
Sillimanite	1.1	49.6	3.7	---	0.9
Cordierite	---	---	6.9	70.7	12.9
Biotite	10.4	13.0	3.5	---	tr
Accessories	3.8	3.0	2.9	7.6	6.2

1. Sillimanite-muscovite schist, Hindman Gulch, ¼ mile northeast of mouth.

2a. Nodule - nodular sillimanite-muscovite schist, Hindman Gulch, ½ mile northeast of mouth. Nodules form 20-40% of the rock.

2b. Matrix - nodular sillimanite-muscovite schist, Hindman Gulch, ½ mile northeast of mouth.

3a. Nodule - nodular cordierite-muscovite schist, Hindman Gulch, 1 mile northeast of mouth. Nodules form 20-40% of the rock.

3b. Matrix - nodular cordierite-muscovite schist, Hindman Gulch, 1 mile northeast of mouth.

biotite schist.

Alteration products are sericite after plagioclase, chlorite after biotite, and hematite after magnetite. They increase in abundance toward the shear zones at the southwest contact of the unit.

Nodular sillimanite-muscovite schist (Fig. 22) contains essential muscovite, sillimanite, quartz, and biotite. Sillimanite-quartz intergrowths form the nodules (Fig. 23). The matrix is schistose and consists predominantly of muscovite and quartz. Biotite is present as unoriented porphyroblasts (average 2 mm across) and small flakes (average 0.2 mm across) that are about equally abundant in the matrix and the nodules. Cordierite is present as poikiloblastic porphyroblasts (average 0.5 mm in diameter) and is restricted to the matrix and peripheral zones of the nodules. Magnetite, apatite, zircon, and schorl are the characteristic accessories. Andalusite is present in the matrix as widely scattered, strongly muscovitised grains. Modes are given in Table 6.

The nodules have very irregular surfaces, projecting into and embayed by the matrix. Foliation planes in the matrix in part bend around and in part are transected by the nodules. The major and intermediate axes of the nodules lie in the plane of schistosity of the matrix but the major axes are unoriented in that plane. The nodules range up to 15 cm in diameter and form from 20 to 40% of the rock. Sillimanite occurs as fibrolitic masses and as prisms which range up to 2 mm in length and 0.2 mm in width. Both forms are unoriented. The sillimanite is intergrown with granular quartz (Fig. 23); quartz grains may have smooth or sutured boundaries and range from less than 0.1 mm to 0.4 mm in diameter. In the matrix the strongly oriented muscovite flakes average 0.3 mm across and less than 0.1 mm thick; quartz grains are xenoblastic

Figure 22. Nodular sillimanite-muscovite schist; small, dark knots are biotite porphyroblasts. Hindman Gulch, ½ mile northeast of mouth.

Figure 23. Photomicrograph of sillimanite (S)-quartz (Q) intergrowth in nodule of nodular sillimanite-muscovite schist. Hindman Gulch, ½ mile northeast of mouth. + polars. 43x.



Figure 22.

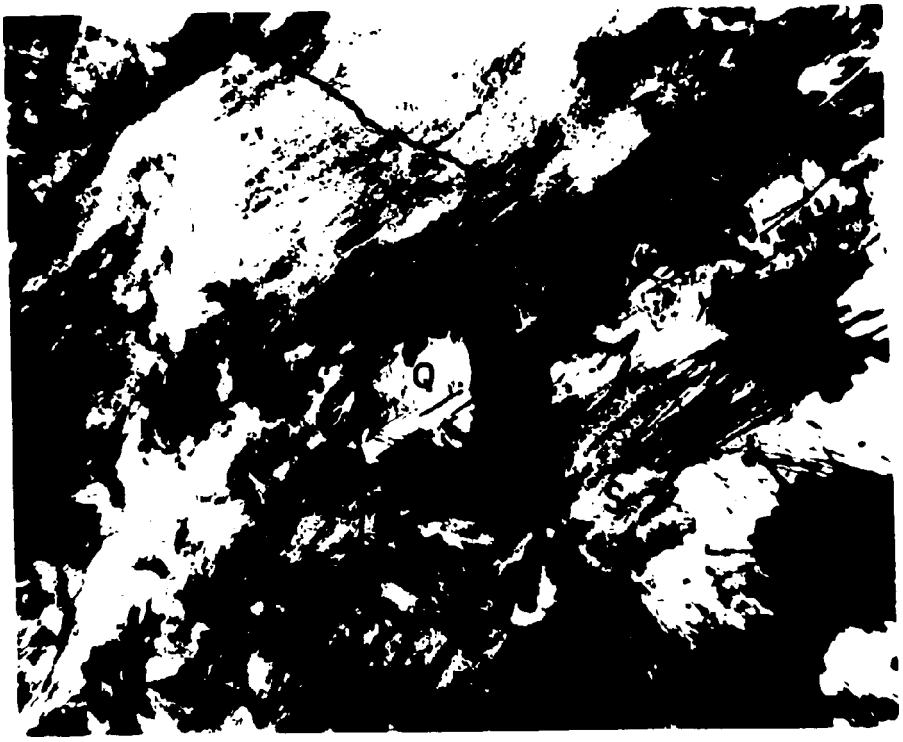


Figure 23.

and average about 0.2 mm in diameter.

Some of the paragenetic relations appear to be: the nodules are later than and have grown by replacing and pushing aside the matrix, sillimanite and quartz partly replace biotite (both porphyroblasts and smaller flakes), and minor late unoriented muscovite replaces sillimanite, biotite, and andalusite.

Nodular cordierite-muscovite schist has cordierite nodules set in a schistose, muscovite-rich matrix similar to that in nodular sillimanite-muscovite schist. The two rocks intergrade over a distance of about 300 feet. The sillimanite-quartz intergrowths become smaller and the cordierite porphyroblasts become larger. Biotite porphyroblasts are less abundant or are absent. Modes are given in Table 6.

The cordierite nodules exhibit a range in size and shape in any one layer and in different parts of the unit. They may be round and average 3 cm in diameter, ellipsoidal and average from 5 to 15 cm on the major axis, or blocky (Fig. 24) and average 5 cm in length (see also the description by Travis, 1956). No regular variation in nodule shape could be established. Ellipsoidal and round types are multigranular; blocky types are single-crystal porphyroblasts. All cordierite grains are highly poikiloblastic; inclusions are quartz and accessory magnetite, apatite, schorl, zircon, colorless (in thin section) garnet, biotite, and sillimanite. The inclusions are concentrated in bands parallel to the foliation in the matrix (Fig. 25). Ellipsoidal and blocky nodules have their long axes in the plane of foliation, but the long axes are unoriented in that plane (Fig. 24). Nodules constitute from 20 to 40% of the rock. Individual cordierite grains range from 2 mm to 3 mm or more in diameter. Quartz inclusions average 0.1 mm in diameter. The

Figure 24. Nodular cordierite-muscovite schist with blocky cordierite porphyroblasts. Small, dark knots are biotite and small cordierite porphyroblasts. Hindman Gulch.

Figure 25. Photomicrograph of nodular cordierite-muscovite schist. Field includes cordierite (C), quarts (Q), muscovite (M), and sillimanite (S). Hindman Gulch. + polars. 43x.

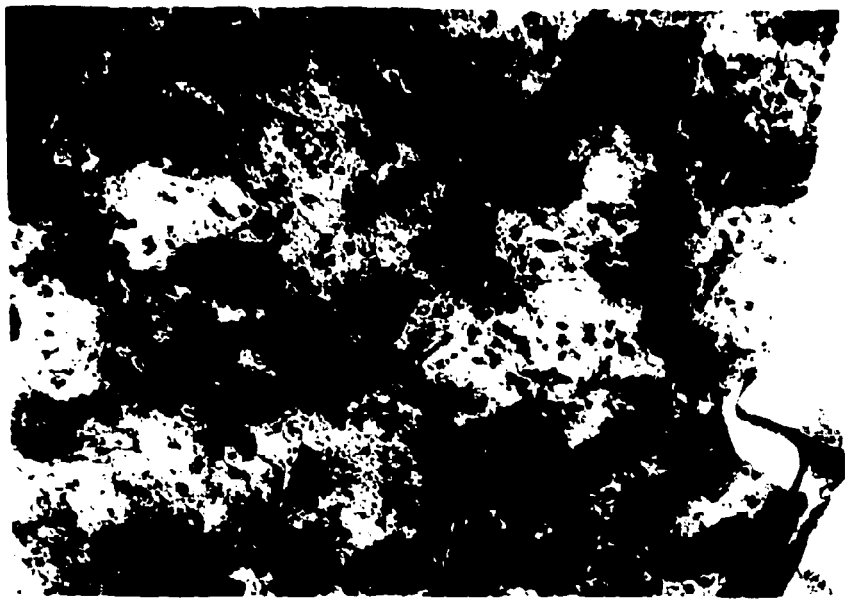


Figure 24.

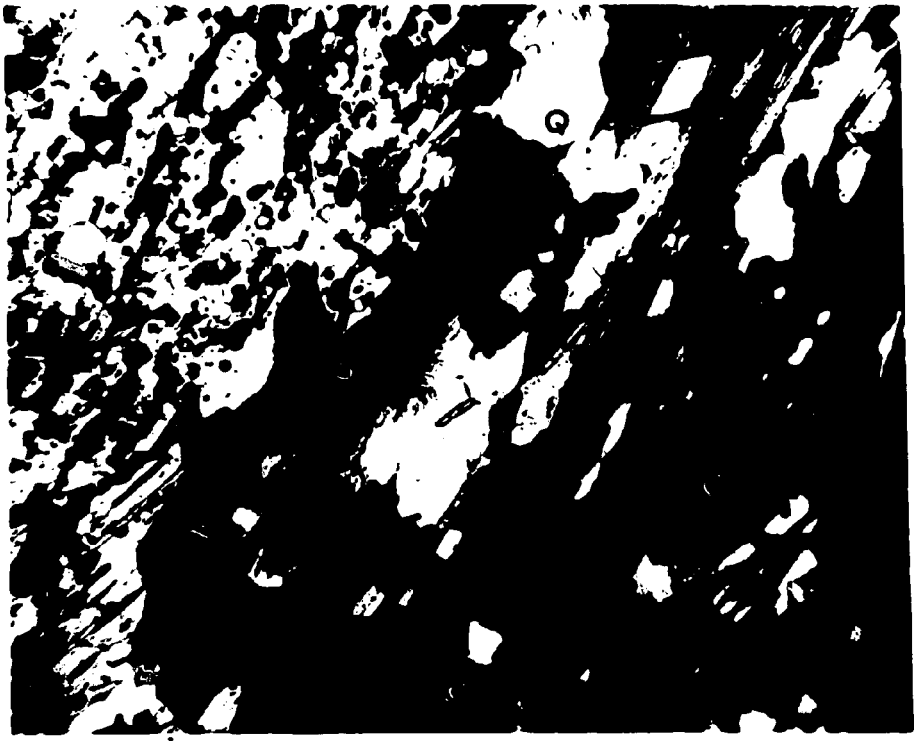


Figure 25.

cordierite commonly exhibits polysynthetic (multilamellar) twinning.

The matrix contains essential muscovite, quartz, biotite, and sillimanite (Fig. 25). Andalusite may be present in accessory amounts; it occurs as poikiloblastic porphyroblasts up to 1 cm in length. Plagioclase (Ab_{68}) may occur in amounts up to 10%. Accessories are magnetite, apatite, zircon, schorl, colorless (in thin section) garnet, rutile, and spinel. Muscovite flakes average 0.5 mm across and less than 0.1 mm thick. Quartz grains are xenoblastic and average 0.2 mm in diameter. Sillimanite-quartz intergrowths up to 5 mm in diameter are scattered throughout the matrix (Fig. 25). Biotite is very light brown in color; it occurs as unoriented poikiloblastic porphyroblasts (average 2 mm across). Foliation in the matrix in part bends around and in part is transected by the nodules.

Some of the paragenetic relations are: biotite is partly replaced by sillimanite; sillimanite, cordierite, and andalusite may be intergrown but generally occur separately, cordierite and andalusite appear to be the earliest and sillimanite the latest; and late, unoriented muscovite replaces biotite, cordierite, andalusite, and sillimanite. Alteration products include chlorite after biotite, and pinnite and hematite after cordierite.

Origin. The following points are pertinent in an interpretation of the origin of the muscovite schist:

1) the three porphyroblastic species (cordierite, biotite, and andalusite) are earlier than sillimanite. That they formed in a foliated rock is clearly shown by the oriented inclusions in cordierite and andalusite. Biotite is generally unoriented and this may indicate the subordinant role of stress at the

time of its formation;

2) the prevalence of accessory tourmaline and the abundance of pegmatites indicate strong migmatitic influence;

3) the rock is conspicuously low in calcium-bearing minerals;

4) there is an inverse relationship in the abundances of biotite and cordierite.

The unit was derived from a parent of somewhat abnormal bulk composition; the present paucity of calcium may be original or may have resulted from metasomatic removal. Also, the uniqueness of the unit in the Idaho Springs formation indicates that it was probably different to begin with. The writer suggests hat a fine-grained clay-chlorite-quartz sediment was a likely parent.

Most of the muscovite, plagioclase, and quartz appear to be strictly regional metamorphic in origin. The porphyroblastic species, rare or absent in other Idaho Springs units, further indicate an abnormal bulk composition and/or metamorphic history for the muscovite schist. The unit must have been subjected to the same regional metamorphic intensities as were the enclosing biotite and hornblende gneisses. The cordierite-andalusite association is more characteristic of thermally metamorphosed rocks, but it is difficult to postulate a set of conditions under which only the muscovite schist unit was thermally metamorphosed. The abundant granitic to pegmatitic sills in the unit are the only possible heat source presently exposed. This problem is further discussed in the section on the origin of the cordieritic rocks.

The sillimanite-quartz aggregates and minor late, unoriented muscovite appear to have been superimposed on both cordieritic and non-cordieritic assemblages in the muscovite schist. They are probably

metasomatic in origin (see section on metasomatism).

The schist in the quartzite-schist sequence in the Coal Creek syncline in the central Front Range, Colorado is similar in many respects to the muscovite schist. Wells et al. (1964) relate the metamorphism of the quartzite-schist sequence to the structural history of the area. They postulate an earlier period of plastic deformation characterized by pervasive, probably deep-seated regional metamorphism followed by a later period of deformation accompanied by granitic intrusion and somewhat higher-grade metamorphism. In the schist layers the earlier period of deformation is represented by staurolite remnants now within andalusite grains and by relict microfolds outlined by inclusions in cordierite porphyroblasts. The second period of deformation and accompanying intrusion produced the cordierite, andalusite, "and other components of the high-grade metamorphic rocks as they now exist" (Wells et al., 1964, p. 021). They consider sillimanite-muscovite with andalusite and cordierite to be the critical (thus equilibrium?) assemblages and to represent progressive regional metamorphism to sillimanite-almandine-muscovite subfacies intensities.

Metamorphism

Origin of the Cordierite Rocks

Occurrences of cordieritic rocks in the Devils Hole area can be summarized as follows:

1) Cordierite-biotite gneiss: cordierite porphyroblasts up to 3 cm in length occur irregularly scattered in biotite gneiss near the Pikes Peak granite in Sec. 11, T48N, R12E. The cordierite occurs as far as 300 feet from the granite contact.

2) Cordierite-biotite gneiss: layer-units of cordierite-biotite

gneiss, interlayered with biotite schist, sillimanite-biotite gneiss, hornblende gneiss, and calc-silicate gneiss, crop out in two small areas, one west of Devils Hole, the other southwest of the mouth of Spike Buck Gulch. The only intrusive bodies in either area are pegmatites (sills and dikes of the Pikes Peak type west of Devils Hole and largely sills of the Cripple Creek (?) type southwest of the mouth of Spike Buck Gulch). However, the cordierite rock units are not restricted to zones along, or even very near, the pegmatite contacts. The cordierite is present as single-crystal porphyroblasts and multigranular nodules.

3) Nodular muscovite schist: cordierite occurs as single-crystal porphyroblasts and multigranular nodules throughout the upper (northeast) one-half to two-thirds of the muscovite schist unit. The cordierite occurs with two other porphyroblastic species, biotite and (minor) andalusite, in a muscovite-quartz matrix. Where cordierite is abundant, biotite is rare or absent. Nodular sillimanite-quartz aggregates are minor to major components of this cordierite-bearing schist.

4) Anthophyllite-cordierite gneiss: xenoblastic cordierite and prismatic anthophyllite are the major constituents in a thin gneiss layer within sillimanite-biotite gneiss on the ridge east of Echo Canyon.

Types 1, 2, and 4 are very minor rock types in the area; type 3 represents a major rock type and is a mappable unit.

It is not likely that four such distinctly different rocks -- different in texture, mineralogy, and chemical composition -- were formed by the same process. The scattered cordierite porphyroblasts in biotite gneiss near the Pikes Peak granite apparently formed in response to locally higher temperatures and may be the only strictly thermal metamorphic cordierite in the area.

Cordierite-biotite gneiss and nodular cordierite-muscovite schist occur in areas where pegmatites and medium-grained to pegmatitic granite sills and dikes are the only exposed igneous bodies. Such bodies are not noted for their exothermic effects. In the Devils Hole area, the only thermal effect definitely associated with pegmatites is minor recrystallization of gneisses (chiefly biotite) near their contacts. This recrystallization rarely extends more than a few inches away from the contacts. Also, the large Pikes Peak body has produced only minor thermal metamorphic cordierite, so it is unlikely that the sills and dikes caused extensive thermal metamorphism. In addition the pegmatites near the nodular corderite-muscovite schist are probably of Cripple Creek (?) derivation. This type is notorious for its poor zoning, simple mineralogy, and lack of wall rock effects.

Travis (1956) concluded that the cordierite-bearing muscovite schist exposed along the south side of the Arkansas Canyon in the Devils Hole area was formed by contact metamorphism of a quartz-mica schist by the Pikes Peak granite. This conclusion cannot be substantiated. Schreyer (1959) includes the nodular cordierite-muscovite schist in his summary of natural cordierite-bearing rocks. He cites the lack of muscovite and biotite within cordierite porphroblasts as evidence that they may have been consumed in the formation of cordierite. Such a reaction, also considered by the writer in an attempt to explain the formation of both cordierite and andalusite, might have been:

$$4KAl_3Si_3O_{10}(OH)_2 + 2K(Mg,Fe)_3AlSi_3O_{10}(OH)_2 \rightarrow$$

muscovite biotite

$$3(Mg,Fe)_2Al_4Si_5O_{18} + Al_2SiO_5 + 2SiO_2 + 6KOH + 3H_2O$$

cordierite andalusite

This is not an isochemical reaction because no potassium-bearing product is present. Therefore, there is the problem of accounting for the excess potassium.

The occurrence of cordieritic biotite gneiss and muscovite schist in layer-units suggests that the original bulk composition determined whether or not cordierite formed. Thus these cordieritic rocks may have formed by isochemical metamorphism.

Schreyer and Yoder (1959) have shown that the stability field of cordierite is very large but that, at least in the presence of water, pressures may be too high and temperatures too low for cordierite to be common in regional metamorphic assemblages. They feel that the association of cordierite and andalusite indicates low temperatures and low pressure. However, Wynne-Edwards and Hay (1963) have listed occurrences of cordierite in regional metamorphic rocks and have shown them to be more common than was earlier believed. They relate the occurrence of cordierite to bulk composition, intensity environment, and stability relative to garnet; thus,

> "both cordierite and garnet are stable at the high pressures of regional metamorphism, but cordierite develops only in lime poor, magnesia-rich rocks which are less commonly represented in the geological column than those in the compositional field of garnets. With decreasing pressure, the compositional stability field of garnet narrows, so that in contact metamorphism the situation is reversed and the commonest pelitic rock compositions fall in the cordierite field" (Wynne-Edwards and Hay, 1963, p. 475).

Perhaps significantly, the muscovite schist has a very low calcium content.

It can be seen on an ACF diagram (Fig. 26) that the assemblages biotite-plagioclase-quartz-cordierite (cordierite-biotite gneiss) and muscovite-quartz-cordierite-andalusite (cordierite-muscovite schist)

are possible. In the latter assemblage, biotite and aluminum silicate should be mutually exclusive, so that the assemblage muscovite-quartz-sillimanite-cordierite-biotite-andalusite (nodular sillimanite-muscovite schist) indicates disequilibrium, even if the sillimanite is metasomatic or the occurrence of both sillimanite and andalusite is explained by fluctuations in intensity near the sillimanite-andalusite join.

The origin of the anthophyllite-cordierite assemblage cannot be established with certainty. Most of the known occurrences of this assemblage have been ascribed to the metasomatism (generally, addition of Mg or Mg and Fe and removal of Ca and alkalies) of a number of rock types (see summaries by Bever, 1954; Deer et al., 1962). Bever (1954) decided that cordierite-anthophyllite gneiss in the Guffey area resulted from the metasomatic addition of Mg and Fe and removal of Ca and CO_2 from impure carbonate rocks. Cordierite-anthophyllite rocks interlayered with hornblende and hornblende-cummingtonite rocks occur in scattered lenses, at places adjacent to thin layers of amphibolite, in microcline gneiss in the Central City quadrangle, Colorado. Sims and Gable (1963) ascribe the rock to the dynamo-thermal metamorphism of sedimentary rocks or possibly to Mg metasomatism of Fe-rich sediments. The Devils Hole occurrence is not particularly similar to either of the above occurrences. Examination of the one small outcrop does not allow a positive interpretation but suggests that the rock may have resulted from the high-grade regional metamorphism of a sedimentary layer of anomalous (Mg-rich) bulk composition rather than some special process such as Mg metasomatism.

A compilation of Colorado cordierite occurences, made for the purpose of comparison, appears in Table 7. It can be seen that a metasomatic origin has been favored for cordierite-anthophyllite assemblages

Table 7. Colorado cordierite occurences.

Location	Association	Origin	Remarks	Reference
Devils Hole area, Fremont Co.	scattered porphyroblasts in biotite gneiss near Pikes Peak granite	Thermal metamorphism		this report
Devils Hole area, Fremont Co.	with andalusite in muscovite schist and in biotite gneiss	regional metamorphism?	poikiloblastic porphyroblasts, nodules up to 6" long	this report
Devils Hole area, Fremont Co.	anthophyllite	regional metamorphism?		this report
Betty Mine, Park Co.	Lime silicates, sulfides	skarn	large masses and a few long, euhedral, prismatic crystals	Bever, 1954
Guffey area, Fremont and Park Co.'s	with anthophyllite	Mg-Fe metasomatism of carbonate rocks	large grains, extremely poikiloblastic both optically + and - types	Bever, 1954
Guffey area, Fremont and Park Co.'s	in quartz-biotite-sillimanite schist	"late origin"	poikiloblasts, optically +	Bever, 1954
Climax pegmatite, Micanite pegmatite Dist., Fremont and Park Co.'s	With feldspar, mica, and quartz in intermediate zone of pegmatite along hanging wall side of core.	igneous (pegmatite)	occurs as altered masses as much as three feet across.	Heinrich, 1950
Eight Mile Park, Fremont Co.	xenolith in pegmatite	contact metamorphism?		Heinrich, pers. comm.

Table 7. Cont'd.

Location	Association	Origin	Remarks	Reference
Cotopaxi-Howard area, Fremont Co.	quartz-mica-albite schist, Mitchell Gulch group	contact metamorphism, albite-epidote hornfels		Salotti, 1960
Central Front Range, unnamed localities	metasedimentary gneisses, with sillimanite, almandine, and microcline	progressive regional metamorphism		Gable, Sheridan, and Wrucke in Nolan, 1964
Central Front Range, unnamed localities	cordierite-bearing rocks occurring near igneous bodies	thermal metamorphism		Gable, Sheridan, Wrucke in Nolan, 1964
Central City quadrangle	biotite-cordierite-garnet-sillimanite gneiss	regional metamorphism, intensity spans boundary of amphibolite and granulite facies	porphroblastic	Sims and Gable, 1963
Central City quadrangle	in biotite gneiss, with sillimanite, garnet, spinel, and microcline in vicinity of dioritic and granodioritic bodies	"intense metamorphism"		Gable, in Nolan, 1962
Coal Creek, Front Range	in quartz-muscovite schist with andalusite	regional metamorphism, sillimanite-almandine subfacies		Wells, Sheridan, and Albee, 1964
Central City quadrangle	anthophyllite	regional metamorphism or possibly Mg metasomatism of Fe-rich sediments		Sims and Gable, 1963

Table 7. Cont'd.

Location	Association	Origin	Remarks	Reference
Pearl Dist., Jackson Co.	nodular mica schist	?	ellipsoidal nodules, 1-4" long	Eckel, 1961 Read, 1903
Monarch District Chaffee Co.	with biotite and quartz in hornfels	contact metamorphic	poikiloblastic	Eckel, 1961
Road cut along U. S. 50, 47.2 miles west of Poncha Springs	spotted cordierite-bearing schist, metasedimentary	?		Donnell, 1960
Gold Brick district, Gunnison Co.	Mica schist	?	dark gray, poikiloblastic subhedra up to 2½" long	Eckel, 1961

and that interpretations of the origin of the cordieritic mica schists are split about equally between thermal and regional metamorphism. It cannot be overemphasized that cordierite is stable over a wide range of temperatures and pressures, especially where p_{H_2O} is less than p_{total}. Also, Schreyer and Yoder (1961, p. 152) have pointed out,

> "Because of its central location in the systems $MgO-Al_2O_3-SiO_2$ and $FeO-Al_2O_3-SiO_2$, which contain the most common components present in natural rocks, cordierite can occur in stable assemblages with a great variety of mineral phases ranging from felsic to mafic members".

Regional Metamorphic Aspects

One of the main objectives of this and similar studies has been the description and interpretation of the regional metamorphic aspects of the Idaho Springs formation. Regional grade, presence or absence of metamorphic zoning, and the chronological position of the metamorphism in relation to other events, e.g., emplacement of the Pikes Peak batholith, are of particular interest.

It has been shown that many of the assemblages are characteristic of the almandine-amphibolite facies. Owing to the lack of certain critical minerals or assemblages, assignment to a specific subfacies is uncertain. The stability of sillimanite is evidence that intensities of the sillimanite-almandine subfacies prevailed. However, the co-existence of an intermediate plagioclase and an epidote mineral in calcareous assemblages is characteristic of the staurolite-quartz or the kyanite-muscovite-quartz subfacies (Fyfe et al., 1958). Calcic plagioclase and the absence of epidote would be more characteristic of the sillimanite-almandine subfacies. Using similar arguments, Salotti (1960) and Shappirio (1962) decided that the assemblages were characteristic of the staurolite-quartz subfacies.

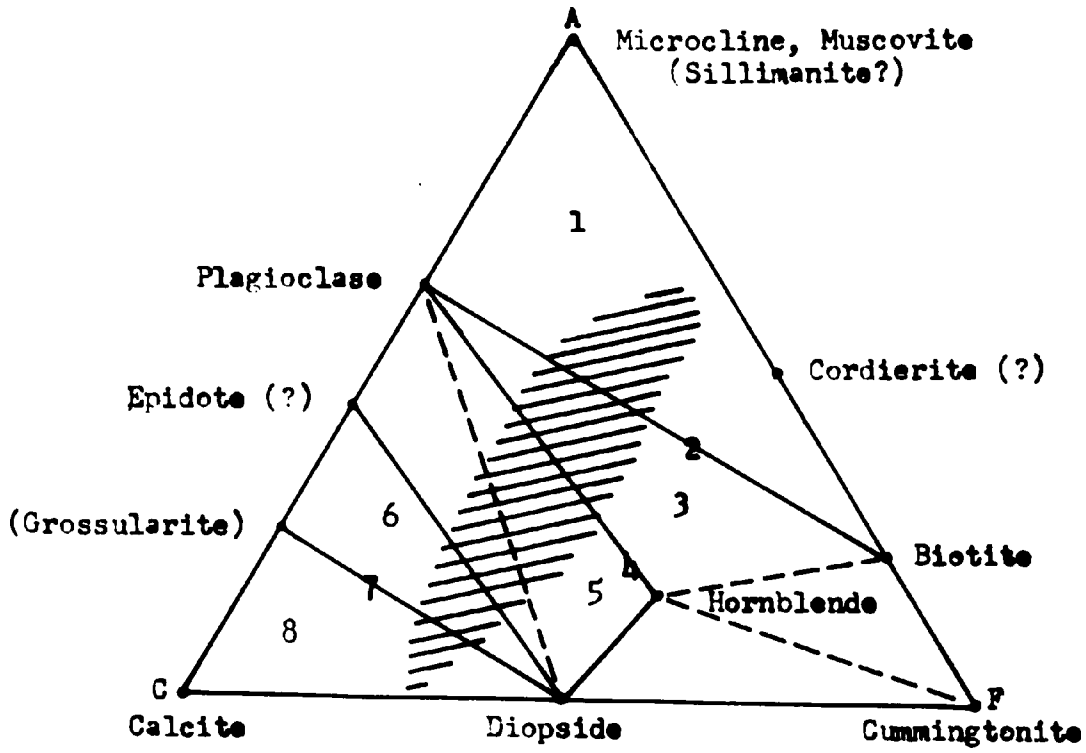


Fig. 26. Schematic ACF diagram empirically derived from assemblages occurring in the Devils Hole area. Excess K_2O and SiO_2. Shaded area covers proposed range in bulk composition of parent sediments. Quartz is a possible additional phase.

Biotite gneiss, Fields 1 and 2.

Hornblende gneiss and hornblende amphibolite (including cummingtonite-hornblende amphibolite), Fields 3, 4, and 5.

Calc-silicate gneiss, Fields 5 through 8.

Muscovite schist, Field 1.

In the Devils Hole area most of the epidote appears to be secondary, a product of retrograde processes. The sillimanite occurs in quartz-sillimanite aggregates that replace primary metamorphic species, the unoriented habit of most of the sillimanite indicating its formation subsequent to the relaxation of strongly directed pressure. Thus there seems to be a total lack of a critical equilibrium assemblage with which to assess more accurately the intensity environment. No systematic variation in regional intensity was found.

The relationship of metamorphic assemblage and parent rock type can be shown on a schematic ACF fiagram empirically derived from assemblages occurring in the Devils Hole area (Fig. 26).

Thermal Metamorphic Aspects

Thermal metamorphism has not been an important process in the development of the rocks in the area. Near the Pikes Peak granite, minor occurrences of cordierite in biotite gneiss and diopsidic augite in amphibolite are thermal metamorphic in origin. Minor recrystallization of the metamorphic rocks has occurred along the contacts of some of the pegmatites.

Migmatitization and Metasomatism

Idaho Springs rocks are nowhere entirely free of injected granitic material. A rock was mapped as migmatite where it appeared, over a large enough area, to be more than 50% granitic material. Rocks mapped as migmatite not only contain many granitic sills and dikes of a large range in size, but commonly appear as though they have soaked in a granitizing solution and have consequently been impregnated and partly replaced by migmatitic quartz, microcline, and rarely, sodic plagioclase. In areas of intense injection the entire rock mass apparently

behaved plastically, giving rise to ptygmatic folding and boudinage structures (Figs. 27 and 28). The degree of foliation was an important factor controlling susceptibility to injection. Thus muscovite schist contains many pegmatites, biotite schist is generally more highly injected than hornblende gneiss, and hornblende gneiss more than massive amphibolite. Also, granitic bodies are more regular in shape (tabular) and continuous in strongly foliated country rocks (Fig. 29). The development of migmatite will be discussed further in the sections on the individual granites.

In preceeding sections it has been suggested that sillimanite has formed as a result of Al metasomatism. Arguments supporting this conclusion are summarized in the following paragraphs.

Sillimanite is present in Idaho Springs rocks as follows: 1) in sillimanite-biotite gneiss layers; 2) in muscovite schist; and 3) in biotite gneiss, but not in particular layers, near granite contacts (minor occurrences). The three occurrences have several features in common:

1) sillimanite is present as spatially separate concentrations of various shapes and sizes; it is not present as a homogeneously dispersed phase.

2) quartz is the only essential mineral in or near the sillimanite concentrations that exhibits mutual textures with sillimanite.

3) feldspars are subordinant or absent in and near the sillimanite concentrations.

4) although discoidal or ellipsoidal concentrations of sillimanite commonly have their two long axes in the plane of foliation, individual fibres and prisms are very commonly unoriented.

Figure 27. Migmatitized Idaho Springs biotite gneiss near Pikes Peak granite, 2½ miles southeast of Table Mountain.

Figure 28. Ptygmatic folding and boudinage structure in migmatitized biotite gneiss, Texas Creek Gulch.



Figure 27.



Figure 28.

The sillimanite formed in response to one of three processes:

1) by progressive regional metamorphism in locally more intense environments.

2) under constant regional metamorphic grade (sillimanite-almandine subfacies), the presence of sillimanite simply reflecting sufficient Al in the parent sediments.

3) through Al metasomatism of favorable layers.

It is not possible, within the limits prescribed by the available data, to completely reject any of these processes. However, the similarity in the relations of the sillimanite in its three occurrences suggests that only one of the processes was dominant in all three occurrences.

Progressive regional metamorphism has some possible signigicance because of the close spatial relation of many of the sillimanitic rocks to granitic intrusives, where locally higher temperatures must have existed. However, this cannot explain the interlayering of sillimanitie and non-sillimanitic rocks.

Variation in the bulk composition of parent sediments under constant regional grade, with high alumina beds resulting in sillimanite-bearing rocks, offers an appealing solution to the unquestionable layered character of most of the sillimanitic rocks. With any other process one has to explain the selectivity which caused only certain layers to be sites of sillimanite formation. But one cannot, in this manner, explain the close spatial relation of sillimanitic rocks to granitic intrusives nor the fact that the sillimanite appears to be secondary and post-stress. However, according to Turner (Fyfe et al. 1958, pp. 230-231),

> "The sillimanite zone is a focus of granite injection or granitization, so that metamorphic temperatures here over-

> lap the magmatic range. Possibly because the metamorphic history of rocks in this zone is complicated by perimagmatic processes, or possibly for some other reason unknown, obvious departures from equilibrium are rather common".

It is the writer's opinion that metasomatism has been the dominant process in the formation of sillimanite. Salotti (1960) has presented a very detailed analysis of the problem. He concludes that the sillimanite was formed by the metasomatic addition of Al to and removal of Fe, Mg, K, OH, F, and perhaps Si from the biotite schist by solutions emanating from the Pikes Peak granite. These conclusions apply, with some modification, to the Devils Hole area. Although there is some aureolic development of sillimanite at the borders of the large mass of Pikes Peak granite, the bulk of the sillimanitic rocks occur in areas where abundant pegmatite dikes are the only granitic bodies present. Although many layers of sillimanitic gneiss occur hundreds of feet from the nearest pegmatite body it is more common to find a close relationship between the number and thickness of pegmatite sills and dikes and the number and thickness of sillimanitic layers.

Under the microscope, the quartz-sillimanite aggregates appear to have replaced the original biotite and feldspar but the rest of the rock (matrix) appears unaffected. A detailed examination of the transition from biotite schist to sillimanite-biotite gneiss shows only the appearance of the quartz-sillimanite aggregates, the enclosing matrix remaining mineralogically and texturally identical to the biotite schist. Thus the process would seem to involve the addition of Al and Si and removal of Mg, Fe, Ca, Na, K, OH, and F.

In summary, sillimanite, more exactly quartz-sillimanite aggregates, formed in some of the Idaho Springs units through metasomatic addition of Al and Si from hydrothermal solutions. The source of the solutions

was the Pikes Peak granite, and the solutions closely followed the same channelways that carried the pegmatite-forming fluids. The aureolic development of some of the sillimanite and the proximity of many sillimanitic layers to pegmatites suggest that local increases in grade were contributing factors. The predominance of the layered character of sillimanite rocks indicates a selectivity in the development of the sillimanite that was in part compositionally controlled, the original bulk composition favorably affected either the reactivity of the layer or its permeability to the hydrothermal solutions (through rock schistosity) or both. The knotted or nodular character of all sillimanite occurrences resulted from the nucleation of sillimanite in scattered biotite flakes which then acted as centers for further sillimanite development, the remainder of the host rock (present matrix) remaining essentially unaffected.

The widespread muscovitization of sillimanite is very common in other areas where sillimanite-bearing units of the Idaho Springs formation are present (Salotti, 1960; Shappirio, 1962). The process has been described in detail (Salotti, 1960) and it is generally agreed that the muscovite results from the fixation of potassium by sillimanite during postmagmatic alkali metasomatism. The source and mode of transport of K is assumed to be the Pikes Peak granite and hydrothermal solutions therefrom. Salotti (1960, p. 74) concludes,

> "...a solution containing a hydrous alkali silicate phase separated from the magma and permeated the enclosing rocks. Large unoriented muscovite metacrysts replaced earlier sillimanite and formed complex myrmekitic intergrowths with quartz. Considerable quantities of K^+, and OH^-, F^- were added to the country rock during this period."

Retrogressive Metamorphism

Overall, retrogressive metamorphism has not been an important

process in the area. Retrograde assemblages are widespread, however, because they occur in and near many of the shear zones. The major retrograde reactions involve the replacement of biotite and hornblende by chlorite, plagioclase by epidote, and plagioclase and microcline by sericite. Excessive shearing and brecciation in the major shear zones (Texas Creek fault, Echo Canyon graben bounding faults, Devils Hole fault) has produced various cataclasites as well. Low temperature hydrothermal alteration, chiefly silicification, calcification, and kaolinitization, has also occurred in some of the shear zones.

It is possible that the muscovitization that has occurred used potassium liberated during the earlier replacement of biotite by sillimanite, the potassium having remained in solutions in or recycled through the rock. In a similar manner, epidote may have formed by reactions with calcium-bearing solutions, the calcium having been liberated during the replacement of plagioclase by microcline. In the epidotization of amphibolites, Ca was available from hornblende.

GRANITE GNEISS

Introduction

In the Devils Hole area, as in many other areas of central Colorado, there occurs a granitoid unit generally younger than the Idaho Springs formation but older than the major granites. Rocks assigned to this unit, although by no means positively correlatable, have in common the composition of a biotite granite and a gneissic structure; several examples are: the microcline-quartz-plagioclase-biotite gneiss of Moench (1964), the biotite gneiss of Bever (1954), the granite gneiss and gneissic aplite of Lovering and Goddard (1950), certain of the granitic gneisses of Christmen et al. (1959), and the granite gneiss

and gneissic granite of Boyer (1962). The writer uses the name granite gneiss for this unit and intends for the name to be descriptive but without genetic implications.

Petrology

Occurrence

Granite gneiss is poorly exposed over about three square miles in the northwest corner of the area (Plate I). It is covered by volcanics north of the mapped area. The exposure area has the shape of a rough half circle. Foliations in the granite gneiss and surrounding Idaho Springs biotite gneiss are conformable and dip away from the center of the granite gneiss body, forming a steeply plunging anticlinal (or a domal) structure (Plate I; p. 149). The contact of the granite gneiss with the Idaho Springs formation is poorly exposed but is gradational where observable. A thick zone of sillimanite-biotite gneiss encloses the granite gneiss to the south and west, except where faulted out by the Texas Creek fault zone. This "mantle" of sillimanite-biotite gneiss fingers out into biotite gneiss which lies southeast of the granite gneiss.

Petrography

Granite gneiss is generally medium-grained but is noticeably more coarse-grained than most Idaho Springs rocks. Aplitic to coarse-grained and pegmatitic phases are not rare. Foliation is commonly weak but ranges from undetectable to moderate. It is caused by alignment of biotite flakes and, to a very minor extent, by alternate banding of biotite-rich and biotite-poor layers. Foliations are weak because of the generally low biotite content throughout the unit. Aplitic to pegmatitic variants of the rock conform to the foliation and have diffuse and gradational contacts with the common medium-grained type.

Granite gneiss is mineralogically simple and rather uniform. Quartz, plagioclase (Ab_{75}), microcline, and biotite are the essential minerals, although biotite may amount to as little as 5% of the rock (see mode, Table 9). The accessory suite is comprised of magnetite, zircon, and apatite. Muscovite and sericite are present in small amounts.

In addition to the range in grain size, microscopic textures are very irregular. In general, an even granoblastic texture prevails, but this appears to be a composite texture made up, in varying degrees of importance, of aspects of igneous (granitic) and metamorphic (xenoblastic, lepidoblastic, and metacrystic) textures. Paragenetically, plagioclase, biotite, accessories, and some of the quartz appear to be early whereas microcline and much of the quartz appear to be later.

The feldspars are weakly sericitized. Widely scattered sheets of muscovite replace biotite and feldspars. Some of the muscovite forms a symplectite with quartz.

Origin

In its present state the rock is a gneiss of granitic composition. Earlier workers (Ball, 1906; Lovering, 1935) considered similar rocks to be orthogneisses. More recent investigators (Tweto, 1947; Bever, 1954; Moench, 1964) have considered them to be paragneisses or have been noncommittal. The writer, too, would prefer to remain noncommittal. The origin of the rock is undoubtedly complex, but one of two very general processes was involved:

1) granitization, by migmatitic injection and associated metasomatism (primarily the addition of K and Si and removal of Fe, Mg, and Ca) of Idaho Springs units;

2) emplacement, early in the granite cycle, of a concordant quartz-monzonitic body with considerable subsequent migmatitic and metasomatic modification by granites later in the cycle.

The Devils Hole area is not suited to the resolution of this problem owing to the relatively small exposure of the granite gneiss body.

GRANITIC INTRUSIVES

General Statement

Precambrian granitic rocks, intrusive into the metasedimentary schists and gneisses, are abundant along the length of the Front Range and Wet Mountain uplifts. In the Arkansas Canyon, as in the Front Range proper, it has been the practice to correlate granitic bodies with one of three more or less distinct granite types. These are:

Boulder Creek granite (Lovering and Goddard, 1950) -- pink to gray, predominantly coarse-grained, well foliated, biotite granite; granodioritic to tonalitic phases are common; occurs in small plutons generally satellitic to larger bodies of Pikes Peak granite or as marginal phases in Pikes Peak plutons; reported in Arkansas Canyon area by Bever (1954) and Dahlem (1965);

Pikes Peak granite (Lovering and Goddard, 1950) -- pinkish gray, coarse-grained, commonly foliated, commonly porphyritic (microcline), biotite granite; occurs in one large batholith and in smaller, outlying plutons; recognized in the Arkansas Canyon area by Heinrich (1948), Salotti (1960), Dahlem (1965), and the writer;

Silver Plume granite (Lovering and Goddard, 1950) -- white to red, medium-grained, commonly porphyritic (K-spar), biotite granite; conspicuous parallelism of K-spar laths in many occurrences; biotite is rarely oriented; occurs in small plutons, generally less concordant than Bould-

er Creek and Pikes Peak plutons; identified in the Arkansas Canyon area by Bever (?)(1954), Salotti (1960), Dahlem (1965), and the writer (?);

Cripple Creek granite (Mathews, 1900) -- reddish, fine-grained, biotite-poor granite; prominant lath-shaped K-spar; considered equivalent to the Silver Plume granite; occurs as a small batholith adjacent on the southwest to the main Pikes Peak batholith; recongnized in the Arkansas Canyon area by Shappirio (1962) and, questionably, by the writer.

Field relations and isotope ratio age dates generally indicate that the Boulder Creek granite is the oldest and the Silver Plume granite and its equivalents the youngest, but that all may be phases of one long period of granite formation and emplacement. Most structural and petrological interpretations indicate that the peak of regional metamorphism and granite emplacement were syntectonic. Salotti's (1960) interpretation of the Mitchell Gulch group and earlier interpretations of the Coal Creek quartzite (Lovering and Goddard, 1950) would require the regional metamorphism of the Idaho Springs formation and the period of granite intrusion to be distinct events.

Three distinct types of granite have been recognized in the Devils Hole area;

Pikes Peak granite -- pinkish gray, medium- to coarse-grained, foliated (biotite), porphyritic (microcline), abundant biotite; characteristic weathering to rounded boulders; correlation based on continuous outcrop with granite mapped as Pikes Peak by Salotti (1960), similarity to Pikes Peak granite in Eight Mile Park (Heinrich, 1948) and near Pikes Peak (type area), and similarity to descriptions in the literature (Mathews, 1900, Lovering and Goddard, 1950);

Cripple Creek (?) granite -- buff-colored, medium-grained, non-foliated, non-porphyritic, contains less biotite than the Pikes Peak granite; coarse-grained to pegmatitic phases amount to about 5% of the exposed rock and occur as irregular masses and less common dikes; conspicuous close-spaced fractures are characteristic of the rock; possibly correlatable with the Cripple Creek granite because it is identical to Cripple Creek (?) granite mapped by Shapiro (1962); also, it is somewhat similar to the Cripple Creek granite described by Mathews (1900);

Silver Plume (?) granite -- reddish, fine- to medium-grained, unfoliated, slightly porphyritic (microcline), biotite-poor; characterized by microscopic streaks of granulation; tentatively correlated with the Silver Plume granite because it is definitely younger than the Pikes Peak granite and is somewhat similar to Silver Plume types described in the literature.

Age relations among the granites in the Devils Hole area are not completely clear. Dikes and more equant bodies of Silver Plume (?) character cut the Pikes Peak granite and the migmatites associated with the Cripple Creek (?) granite. Pikes Peak and Cripple Creek (?) granite were not observed in contact with each other. The extensive migmatite associated with the Cripple Creek (?) granite, as well as its more altered character, make it appear to be the older of the two, which is contrary to the proposed correlation. As far as field relations indicate, all three granites could actually be variants of only one granite, namely the Pikes Peak. Age dates have been obtained for some of the rock types in the Arkansas Canyon area; these are listed in Table 8.

Table 8. A^{40}/K^{40} age dates for some rocks in the Arkansas Canyon area (Prof. E. Wm. Heinrich, pers. comm.).

Rock Type --- Dated Mineral	Specimen Location	Age
Boulder Creek granite --- Biotite	Where Wilson Creek enters Wilson Park.	1350(±40)x10^6 years
Pikes Peak granite --- Biotite	8 mi. west of Canon City, along U.S. 50.	1370(±60)x10^6 years
Silver Plume granite (?) --- Biotite	Texas Creek area.	1320(±60)x10^6 years
Cripple Creek granite(?) --- Biotite	Top of Miners Gulch Road, Tallahassee Creek area.	1290(±40)x10^6 years
Red aplitic granite gneiss --- Biotite	Smiths Gulch Tallahassee Creek area.	1165(±40)x10^6 years
Cordierite-muscovite schist --- Muscovite	Texas Creek area.	1375(±30)x10^6 years
Sericite schist, Mitchell Gulch group --- Muscovite	Mitchell Gulch, Cotopaxi area.	1290(±40)x10^6 years
Sericite schist, Gooseberry Gulch fm. --- Sericite	South side of Wilson Park.	1290(±50)x10^6 years

Pikes Peak Granite

Occurrence

Pikes Peak granite underlies nearly one-fourth of the area (Plate I). Exposed masses are part of essentially one body, a small batholith that extends westward from the western part of the Devils Hole area (see Salotti, 1960, Plate I). Near the Texas Creek Post Office the body forms the core of a large domal or anticlinal structure the sharply plunging nose of which is repeated to the east by the Texas Creek fault (Plate I).

There are a great many sills and dikes in the metamorphic rocks that are associated with the Pikes Peak granite (Fig. 29). These are granitic in composition and range in character from identical to the Pikes Peak through medium-grained types to pegmatites. The first are obviously related to the Pikes Peak granite because of their similarity of appearance and the fact that they invariably occur close to Pikes Peak contacts. Correlation of the medium-grained, non-porphyritic sills and dikes with the Pikes Peak granite is less certain. They are more biotite-rich than known Cripple Creek (?) dikes and do not contain the intense granulation characteristic of Silver Plume (?) dikes, and they are more numerous near Pikes Peak contacts. Pegmatite dikes are even less easily correlated with the Pikes Peak. There are only two likely sources for the many pegmatites in the area; the Pikes Peak and Cripple Creek (?) granites contain interior pegmatites, the Silver Plume (?) granite does not. There are also many pegmatites in the metamorphics near the margins of the Pikes Peak and Cripple Creek (?) granites but not the Silver Plume (?) granite. Only the Pikes Peak granite contains zoned interior pegmatites and it is this feature that provides the best evidence that most of the large zoned exterior pegmatites in the area

are related to the Pikes Peak granite. Many of the exterior pegmatites are zoned, and many that are not are otherwise mineralogically and texturally similar to those that are. Therefore, it is the writer's opinion that most of the pegmatites in the area were deposited by residual fluids derived from the Pikes Peak magma.

The greatest concentration of Pikes Peak pegmatites lies in a zone extending from Table Mountain east-northeast to the north end of Devils Hole. Nearly all of the mineralogically interesting pegmatites and those that have been profitably mined are in this zone. These pegmatites are similar to those of the Eight Mile Park area which Heinrich (1948) related to a Pikes Peak source. In south-central Colorado the Pikes Peak granite is the only granite common to the various pegmatite districts.

Structure

Pikes Peak granite very commonly has a distinct foliation (Fig. 30) produced by the alignment of biotite flakes and clusters and, less importantly, by alignment of elongate microcline phenocrysts and microcline aggregates. The foliation is stronger and more regular near contacts and is commonly faint and irregular in the central part of the Pikes Peak mass. The foliation parallels the contacts of the granite and the foliation and layering in the enclosing metamorphics.

Xenoliths are common near the margins of the batholith and may be abundant. Most are small but some are 100 feet or more in length. They range from unaltered rock with sharp contacts to largely assimilated ghosts indicated only by a higher than normal biotite concentration. The only xenolithic rock types observable are biotite gneiss and hornblende gneiss. The foliation and elongation of the xenoliths commonly

Figure 29. Granite sill of Pikes Peak type in Idaho Springs biotite gneiss, along highway ½ mile east of Texas Creek.

Figure 30. Foliation in Pikes Peak granite, Reece Gulch.



Figure 29.



Figure 30.

parallels the foliation of the granite (Fig. 31).

Variations in mineralogy and texture are common but generally irregular. A biotite-rich facies occurs along Red Gulch at the western edge of the area. There is a fine-grained, non-porphyritic border phase along the eastern margin of the batholith in Sec. 11, T48N, R12E. The zone ranges from a few feet to about 300 feet in thickness. It grades into normal Pikes Peak granite and may well be a chilled margin.

Pikes Peak granite is cut by three kinds of dike rocks; pegmatites (Fig. 32); red, fine-grained granite; and diabase (Fig. 33). There is no regularity to the distribution of any of these dike types. The pegmatites almost certainly are late derivatives of the Pikes Peak magma. The fine-grained granite dikes are mineralogically and texturally identical to the Silver Plume (?) granite and are, therefore, probably dikes of that younger granite. The diabase dikes are much younger, being in fact the youngest Precambrian rocks in the area.

The contacts of the Pikes Peak granite are, in general, conformable with the layering and foliation in the enclosing metamorphics. They are sharp or gradational over a few feet.

Contact Effects

Biotite gneiss has a rough, knotted appearance near the Pikes Peak contact in Sec. 11, T48N, R12E. This is caused by scattered, small cordierite porphyroblasts which the writer believes have formed in response to locally higher temperatures.

Migmatitization of country rocks has accompanied the intrusion of the Pikes Peak granite in places, but it is not as intense or as widespread as that associated with Cripple Creek (?) granite. The main areas of Pikes Peak-related migmatites are along Fernleaf Gulch (Fig. 34),

Figure 31. Biotite schist xenolith in Pikes Peak granite. Pasture Gulch.

Figure 32. Small interior pegmatites in Pikes Peak granite. Reece Gulch.



Figure 31.



Figure 32.

Figure 33. Diabase dike in Pikes Peak granite. Reece Gulch.

Figure 34. Migmatite near mouth of Fernleaf Gulch.



Figure 33.



Figure 34.

west of Table Mountain, and southeast of Table Mountain (Fig. 27). The first two are elongate roof pendants, the third area is a highly injected patch of foliated rocks surrounded by massive amphibolite (Plate I).

Metasomatic microcline and quartz are common in gneisses near Pikes Peak granite and its associated granite and pegmatitic sills and dikes. The metasomatic phases may be noticeable for only a few inches from the contacts or they may persist throughout large areas, as in the migmatites. Sillimanite-quartz aggregates in biotite gneiss near the Pikes Peak contact in Sec's. 2 and 11, T4S8N, R12E, are present as a result of Al metasomatism or locally higher metamorphic intensities or both.

The Pikes Peak magma apparently was the source of the solutions responsible for the metasomatism of the Idaho Springs rocks. The close association of Pikes Peak pegmatites and sillimanitic rocks indicates that the pegmatite-forming and Al-rich fluids followed the same channelways.

Petrography

Quartz, microcline, plagioclase (Ab_{80}), and biotite occur in essential amounts; magnetite, apatite, zircon, and allanite are common accessories. A modal analysis is given in Table 9. Plagioclase is weakly sericitized and is embayed and veined by microcline. It commonly has clear, albitic rims against microcline or is separated from it by myrmekite. Microcline displays a great range in grain size and may occur as phenocrysts or aggregates a centimeter or more in length. It is perthitic and contains both exsolved, clear plagioclase blebs and larger, remnant patches of highly sericitized plagioclase.

Biotite flakes and aggregates are crudely aligned and, with elongate aggregates of microcline grains, produce the foliation. The average

Table 9. Modes of Precambrian igneous rocks.

	1.	2.	3.	4.	5.
Quarts	38.5	33.5	27.5	28.0	---
Plagioclase	40.0 (Ab_{76})	26.0 (Ab_{80})	35.0 (Ab_{74})	28.0 (Ab_{80})	49.0 (Ab_{50})
Microcline	9.0	31.0	30.0	39.0	---
Olivine	---	---	---	---	23.5
Augite	---	---	---	---	24.5
Biotite	10.5	8.0	4.5	4.0	---
Muscovite	2.0	---	3.0	1.0	---
Accessories	tr	1.5	tr	tr	3.0

1. Granite gneiss, Antelope Gulch.
2. Pikes Peak granite, Fernleaf Gulch.
3. Cripple Creek (?) granite, East Gulch.
4. Silver Plume (?) granite, quarry west of Texas Creek.
5. Diabase, Reece Gulch.

grain size of the rock is about 2 mm. There is a minor amount of granulation and suturing along grain boundaries.

Pikes Peak granite dikes are more even-grained, contain more of the felsic intergrowths (perthite, myrmekite, and graphic granite), have more highly sericitized plagioclase, and show more granulation along grain margins.

Cripple Creek (?) Granite

Occurrence

Cripple Creek (?) granite occurs in what is apparently a phacolith covering about two square miles in the center of the area. The best exposures are in East Gulch, southeast of Devils Hole (Fig. 35). It is difficult to distinguish with certainty the medium-grained to pegmatitic sills and dikes of Cripple Creek (?) affinity from those of Pikes Peak affinity, especially west of the Cripple Creek (?) body. East of the body sills and dikes are numerous and are identical to the granite in the main body. They contain accessory garnet, which seems to characterize the Cripple Creek (?) dike rocks (but not the Cripple Creek (?) granite). Both interior and exterior pegmatites are unzoned. The many medium-grained to pegmatitic sills southwest of the Cripple Creek (?) body could not be related definitely to either one of the two granites. A few of the pegmatites show some zoning (small quartz cores), indicative of Pikes Peak types, but most of them do not appear to be Pikes Peak types. Many of the medium-grained granite sills are garnet bearing, indicative of Cripple Creek (?) types. It is possible that both types are present, the unusual abundance of sills in this area resulting from the overlap of two different zones of granitic intrusion.

Figure 35. Cripple Creek (?) granite along East Gulch at the south end of Devils Hole.

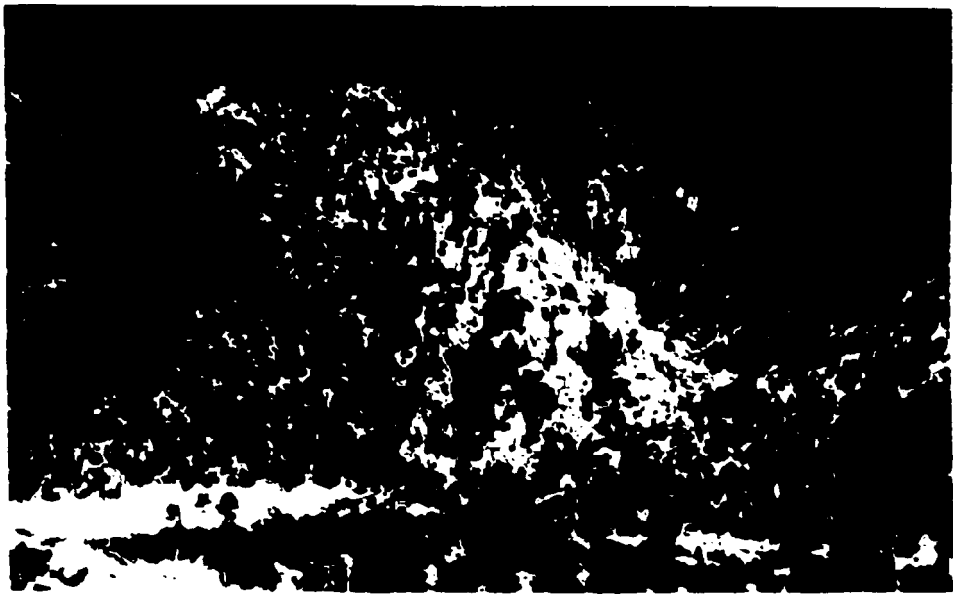


Figure 35.

Structure

The Cripple Creek (?) granite contains very little measurable regular internal structural features. The rock is non-foliated, exhibits no regular variation in mineralogy or texture, and xenoliths are rare. Coarse-grained and pegmatitic phases occur as irregular masses and as tabular dikes. The only prominant structural feature is the horizontal sheeting of the rock. This is especially prominant in East Gulch, and a specimen from that locality cannot be distinguished from a specimen of Cripple Creek (?) granite from the Tallahassee Creek area (Shappirio, 1962).

There is a great difference between the contacts on either side of the nearly vertical phacolithic body (excluding the northwest contact which is faulted out); the convex southwest contact (hanging wall ?) is sharp and the enclosing rocks have not been migmatitized, although sills are numerous. The concave eastern contact (foot wall ?) can only be arbitrarily drawn because the enclosing rocks are extensively migmatitized.

Contact Effects

The migmatitization associated with the Cripple Creek (?) granite is the most extensive and most intense in the area. Near and up to 300 yards away from the granite contact granitic material predominates greatly, and the metamorphics appear as scattered lenses and layers between thick sills of granite. Further away, granite and metamorphic rock are nearly equal in abundance. The granite occurs as aplitic, medium-grained, and pegmatitic sills and dikes that range in thickness from less than an inch to tens of feet. Both metamorphic rocks and granite bodies commonly exhibit ptygmatic folding, but straight sills and dikes are also

abundant. The straight dikes offset ptygmatically folded ones. The metamorphics consist of biotite gneiss and small amounts of hornblende gneiss and sillimanitic gneiss, the appearance of the metamorphics is somewhere between that of a biotite schist and that of a fine-grained granite. In contrast to other migmatites and less injected Idaho Springs rocks, the Cripple Creek (?) migmatites are generally reddish in color; this is caused by hematite dust in the K-feldspars.

Petrography

Cripple Creek (?) granite contains essential quartz, microcline, plagioclase (Ab_{86}), and biotite, and accessory muscovite and sircon. Apatite and magnetite are rare. A modal analysis is given in Table 9. Plagioclase is moderately to strongly sericitized and is embayed and otherwise partly replaced by quartz and microcline (Fig. 36). It commonly has clear, albitic rims against microcline. Myrmekite blebs may be present between plagioclase and microcline but are less abundant than in other granites of the area. Microcline shows a wide range in grain size; this causes the rock to have an uneven texture. Microcline partly replaces some quartz grains; it is weakly perthitic and contains few graphic quartz inclusions. Muscovite is more abundant than in other granites in the area; it occurs as shreds and large flakes, It may be intergrown with biotite and much of it appears to replace plagioclase and biotite. The biotite is unoriented and about half of it is altered to chlorite.

The rock is uneven-grained but not porphyritic; grain size averages between 1 and 2 mm. Grain boundaries are very uneven. The microscope reveals many parallel, through-going fractures, but little granulation along these fractures or along grain boundaries (Fig. 36).

Figure 36. Photomicrograph of Cripple Creek (?) granite, East Gulch. Note microcline(M) embaying quartz(Q) and plagioclase(P). Also note the close-spaced fractures. + polars. 43x.

Figure 37. Photomicrograph of Silver Plume (?) granite, Race Path Gulch. Note streaks of granulation; field includes microcline(M), quartz(Q), and plagioclase(P). + polars. 43x.

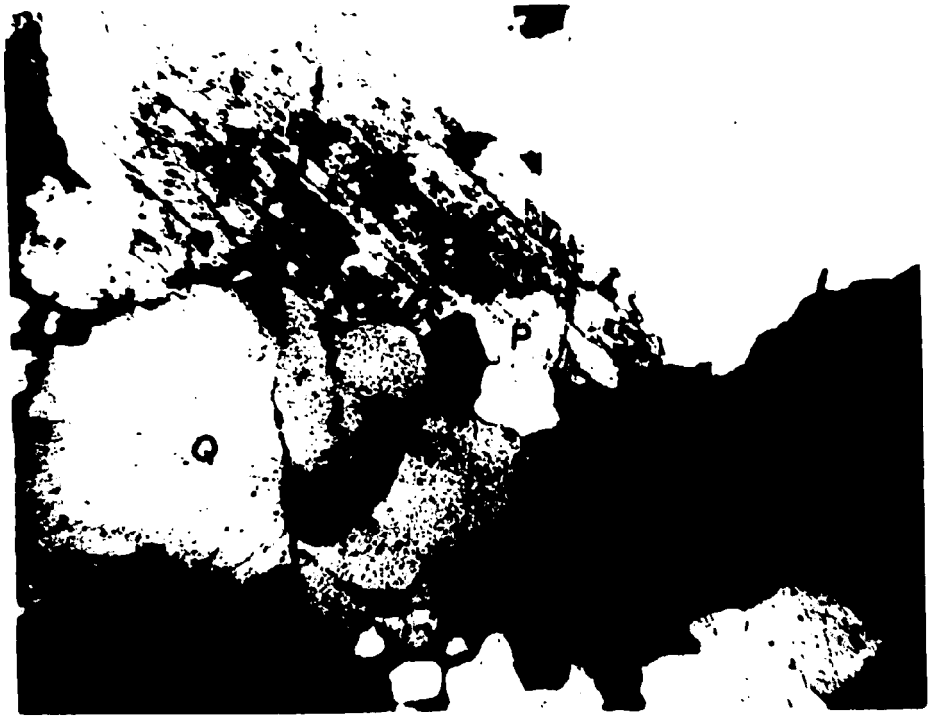


Figure 36.

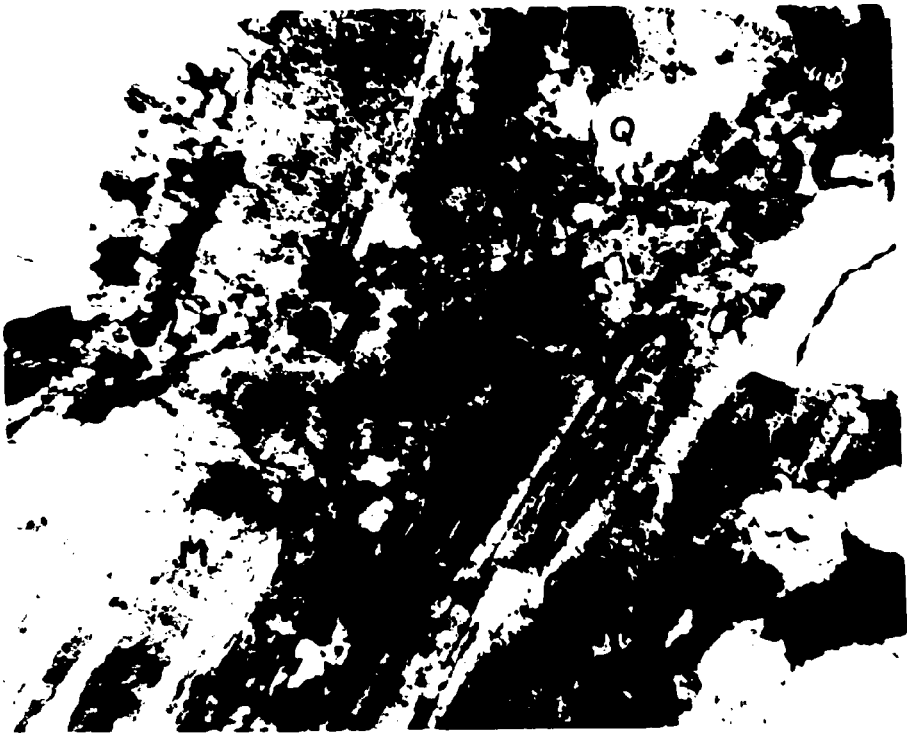


Figure 37.

Dike rocks are more even-grained, contain more muscovite and less biotite, and contain accessory garnet.

Silver Plume (?) Granite

Occurrence

The largest body of Silver Plume (?) granite is west of Texas Creek in the southern part of the area (Plate I); detailed mapping may reveal this body to be a thick sill. Many dikes, sills, and small plugs of a very similar granite are exposed in the southern half of the area. Such rocks transect both the Pikes Peak granite and the Cripple Creek (?) migmatite. There is little variation in the mineralogy, texture, or structure of this granite in any of its occurrences.

Structure

Abundant, close-spaced granulated streaks are the only internal structure observed. Alignment of streaks may constitute a foliation but to establish this would require a study of many oriented thin sections. The granulation occurred during the emplacement of the granite and was caused by the late movements of the largely crystalline magma (autoclastic texture). This is evidenced by the fact that the granulation is characteristic of the rock in different places in the same mass and in different bodies of the same (?) rock from widely separated localities. The granulation has not resulted from local or regional cataclastic deformation because many of the host rocks contain no such evidence of granulation.

Silver Plume (?) contacts are the sharpest of all of the granite types. Some bodies display narrow chilled margins. Xenoliths of biotite and hornblende gneisses are uncommon; they are fresh in appearance and also exhibit sharp contacts with the granite.

Contact Effects

No thermal or metasomatic effects were noted along Silver Plume (?) granite contacts. The granite has not formed any migmatite in the area, although Silver Plume (?) dikes and sills approach 50% of the total rock in some areas (in the northern half of Sec. 24, T48N, R12E, and along both sides of the Arkansas River for a mile west of Texas Creek).

Petrography

Silver Plume (?) granite consists of essential quartz, microcline, and plagioclase (Ab_{80})(see column 4, Table 9). Biotite, zircon, apatite, magnetite, epidote, and allanite are common accessories. Plagioclase occurs as equant to ragged grains that are embayed by and contain patches of replacing microcline and quartz. Plagioclase is moderately to strongly altered to sericite and calcite, and is clouded with dusty hematite which imparts the characteristic red color to the rock. Microcline grains range in size from tiny blebs to phenocrysts a centimeter in length. Myrmekite is abundant along microcline-plagioclase contacts, commonly as fan-shaped blebs projecting into the microcline. Microcline is markedly perthitic; it contains few quartz inclusions. Quartz may occur as separate grains but, more characteristically, it occurs as aggregates of sutured, interlocking grains that resemble quartzite.

The outstanding microscopic feature of this rock is the presence of numerous streaks of granulation and markedly granulated and sutured grain boundaries (Fig. 37). Granulation has involved quartz, plagioclase, and accessories, the hematitic plagioclase imparting a red color to the streaks. Microcline is at least in part later than the granulation and has partly engulfed some of the streaks.

ALTERED PERIDOTITE

A body of actinolite-chlorite greenschist crops out two miles north-northeast of the Texas Creek Post Office. The outcrop is on a divide and exposures of this body are so poor that its relationship with enclosing rocks could not be determined.

The actinolite occurs as slender prisms that range from unoriented to highly oriented in different parts of the rock mass. Chlorite ranges in size from minute shreds to sheets up to 2 mm in diameter. Relict portions of once millimeter-size grains of olivine are present; the olivine is largely replaced by actinolite, chlorite, limonite, and calcite. A small amount of idioblastic and only slightly altered anthophyllite is present in this rock.

The rock may be an altered peridotite. The possibility of such an origin for this unusual rock type was brought to the attention of the writer by E. Wm. Heinrich, who has mapped a similar occurrence in the gorge of Currant Creek, due east of Gribble Mountain (Heinrich, pers. comm.). The peridotite (?) is probably post-granite, although this is certainly not clear. In the Currant Creek occurrence the peridotite (?) apparently cuts migmatites formed by the Pikes Peak granite (Heinrich, pers. comm.).

DIABASE

Occurrence

Mafic dike rocks exhibiting diabasic textures are widely distributed in the area (Plate I). They cut both the metamorphic and granitic rocks, and are the youngest of the Precambrian rocks (Fig. 33). Their Precambrian age cannot be proved but is indicated by the fact that they are found only in Precambrian rocks and by their similarity to dike

rocks so interpreted in nearby areas (Fig. 1). In Eight Mile Park (Heinrich, 1948) they do not cut Paleozoic rocks and in the Lookout Mountain area (Dahlem, 1965) a diabase dike is cut by a Precambrian lamprophyre dike.

The diabase dikes are most commonly tabular bodies ranging from a few inches to ten feet in thickness and traceable along strike for a few tens of feet. The rock weathers deeply and good exposures of fresh rock are rare. Commonly the presence of a dike is indicated only by a train of altered, dark brown cobbles lying on the ground. Where fresh, the dikes are black to dark gray or dark brown, aphanitic to medium-grained, and even-grained to porphyritic. Very commonly they are fine-grained and non-porphyritic. The diabasic texture may be visible in hand specimens. Plagioclase, pyroxene, and olivine may occur as megascopic phenocrysts. Many dikes exhibit a variation in grain size, and fine-grained, chilled margins are common.

Diabase dikes generally appear to have followed strong, local fractures, but no regional set or system is evident. The dikes may branch; and small offshoots may be locally so numerous as to give the country rock the appearance of a diabase-cemented breccia. Several of the larger dikes have marginal zones rich in xenoliths and xenocrystic quartz and feldspar.

Many dikes were intruded along shear zones. These are commonly highly altered and have been brecciated or sheared by subsequent movement in the zone (Fig. 38).

Petrography

Microscopic examination reveals that the dikes have gabbroic compositions and diabasic to porphyritic and microporphyritic textures.

None are lamprophyres, although lamprophyres have been mapped in adjoining areas (Saletti, 1960; Dahlem, 1965).

Plagioclase (Ab_{46}) and titanian augite are the main constituents. Olivine is commonly present in accessory to essential amounts. Brown hornblende may be present and in some dikes it is abundant. Apatite and magnetite are common accessories. A modal analysis is given in Table 9. Plagioclase grains are predominantly lath-shaped, but phenocrysts are equant. Euhedral to anhedral augite grains fill the spaces between plagioclase laths; less commonly, large augite subhedra are molded around the plagioclase laths (ophitic texture). Olivine is commonly phenocrystic but also occurs in even-grained rocks. Hornblende occurs as stubby prisms that are generally smaller than grains of the other mafic species in the rock.

Alteration is generally strong and may be complete. The following is a list of the primary minerals and their alteration products;

plagioclase — sericite, calcite, chlorite;
augite -- chlorite, iron oxide, hornblende, biotite;
olivine — serpentine, talc, chlorite, magnetite;
hornblende — chlorite, biotite.

Figure 38. Quartz-cemented diabase breccia, Devils Hole fault some ¼ mile west of Fernleaf Gulch.

Figure 39. Echo Canyon graben. View looking east from divide between Devils Hole and Echo Canyon. Light streaks are outcrops of Farisita (?) formation.



Figure 38.



Figure 39.

TERTIARY ROCKS

FARISITA (?) FORMATION

A group of poorly consolidated clastic sediments, found only in the Echo Canyon graben (Fig. 39), are the oldest sediments in the area. They are very poorly sorted and range from mudstone to conglomerate, with coarse to conglomeratic arkoses predominating (Figs. 40 and 41). Bedding is nearly absent; in most exposures vaguely defined gradations in grain size constitute the only bedding feature observable. The apparent lack of bedding is in part real and in part reflects the limited extent of most outcrops. The net result, in any case, is that the lithostratigraphy and structural attitude of the sediments cannot be deciphered. There is a general trend from conglomerate at lower elevations near the south end of the canyon to mudstones at higher elevations near the north end.

The finer sediments are brown to greenish gray in color and contain conspicuous mica and quartz. In places they also contain carbonized and iron-stained, non-specific, plant debris. The coarser sediments are yellow to tan in color and contain abundant quartz, feldspar, and granite pebbles and boulders. Of the possible granitic source rocks in the area, the fragments most closely resemble the granite gneiss and the Cripple Creek (?) granite. No material of certain volcanic origin was found in any of these sediments.

The sediments are restricted to Echo Canyon and there are exposed only along the sides of some of the larger gullies and streams. The poor exposures result from the unresistant character of the sediments and the thick cover of erosional debris (basin debris) which extends over most of Echo Canyon (Fig. 41).

Figure 40. Typical outcrop of Farisita (?) arkose, Echo Canyon.

Figure 41. Farisita (?) arkose and conglomerate overlain by basin debris. Large, light boulder (granite) is in Farisita (?) conglomerate; large, dark boulder (andesite) is at the base of the basin debris. Echo Canyon.



Figure 40.



Figure 41.

The sediments are tentatively correlated with the Farisita formation (Johnson and Wood, 1956). This correlation is based on the great visual similarity of the two rocks (Robert MacNish, pers. comm.) and on their relation to the volcanics. MacNish (pers. comm.) believes the Farisita sediments were deposited by a southerly flowing major drainage in middle to upper Eocene time.

The sediments are older than the volcanic rocks in the area because they contain no volcanic material. A maximum age cannot be established with any certainty. The Farisita formation is probably late Early Eocene to late Eocene in age (Berner and Briggs, 1958; MacNish, pers. comm.). Similar sediments have been mapped by Shappirio (1962) a few miles to the northeast of Echo Canyon. Shappirio (1962) tentatively correlates those sediments with the Denver formation, a prominant rock of Eocene age in South Park (Stark, et al., 1949) and east of the Front Range. The basis of this correlation is the tentatively assigned Eocene or Paleocene age of some plant fossils he collected from a mudstone unit (Shappirio, 1962). The sediments in both areas are older than the volcanics which are younger than middle Paleocene but no younger than earliest Oligocene (Chapin and Epis, 1964). The problem of correlation rests with the question whether the sediments are a very local accumulation or whether they represent a locally preserved remnant of a regional deposit.

Chapin and Epis (1964) have noted arkosic sediments filling local topographic lows on the pre-volcanic surface; they cite one such deposit in the Tallahassee Creek area which is probably the same one described by Shappirio (1962). They also refer to pre-volcanic arkose at the base of the north wall of the Devils Hole (Echo Canyon of this report). Chapin and Epis (1964) do not assign a formational name or an

exact age to these deposits.

VOLCANIC ROCKS

General

Tertiary volcanics cover the northern part of the area and cap Table Mountain and the ridge between Bull Gulch and Antelope Gulch (Plate I). The best exposures are on Table Mountain and in the cliffs north of Echo Canyon (Fig. 7). The following units have been identified:

youngest	andesite (flows, flow breccia, and scoria)
	volcanic conglomerate and tuffaceous sandstone
	rhyolite tuff
oldest	basaltic agglomerate

The main units are rhyolite tuff and andesite. They are widespread in the area and to the northeast, north, and northwest. They are nearly everywhere separated by a thin, rarely exposed layer of volcanic conglomerate. A typical section was measured on Table Mountain and is shown in Table 10. The section in Echo Canyon graben is not typical (Table 11, compare with Table 10); it is much thicker below the andesite and contains units not found elsewhere. In Echo Canyon the basal unit is basaltic agglomerate, elsewhere it is rhyolite tuff. The volcanic conglomerate is much thicker in Echo Canyon and beds of tuffaceous sandstone occur within the conglomerate unit.

The volcanics are part of the Thirty-nine Mile volcanic field, and lie in the southern part of that field (Bever, 1954; Chapin and Epis, 1964). Of the north wall of Echo Canyon, Chapin and Epis (1964, p. 153) state,

"A 2000-foot section, beginning with pre-volcanic arkose and

Table 10. Volcanic section, Table Mountain

Unit	Thickness
//////////	
dark gray to reddish black, brown and maroon, scoriaceous to dense andesite flows.	135'
————	
Medium to dark gray, aphanitic andesite flow.	95'

Volcanic conglomerate, not exposed.	40'

Light pinkish brown, slabby bedded, moderately welded tuff; contains many Precambrian fragments.	30'
————	
Medium brown, moderately to strongly welded tuff; contains many Precambrian fragments.	75'
————	
Light buff, light-weight, slightly welded to unwelded rhyolite tuff; contains acid volcanic and Precambrian gneiss fragments.	140'
//////////	
Precambrian	

Table 11. Volcanic section, northwest end of Echo Canyon, and its relation to the composite stratigraphic column of Chapin and Epis (1964, Plate II).

Thickness	Unit	Chapin and Epis (1964, Plate II).
160'	Medium to dark gray, dense, aphanitic andesite.	Upper andesite.
10-20'	Red and maroon, scoriaceous andesite.	
75'	Volcanic conglomerate as below, upper 2' are reddish brown siltstone and ss.	
70'	Very fine-grained tuffaceous sandstone, lower 3' is coarse-grained.	Upper conglomerate.
175'	Volcanic conglomerate, predominantly andesite cobbles in sandy, calcite-cemented matrix. Upper 2' is coarse, brown sandstone.	(Trump - Wagontongue formation ? -- upper Miocene - Pliocene ?).
80'	Reddish brown, dense, moderately welded rhyolite tuff.	
190'	Pinkish brown, moderately welded rhyolite tuff; commonly near base of volcanics in rest of area.	Ash flows 5, 6, and 7.
25'	Glassy, strongly welded tuff.	
105'	Thin to massive bedded tuff and sandy tuff; granite pebble horizon near top.	Antero formation - lower Oligocene.
210'	Basaltic agglomerate, volcanic conglomerate, mud flow (?); angular, basic volcanic fragments in a dark matrix; a few Precambrian cobbles; a few acid volcanic fragments; more conglomeratic in upper part of unit.	
60'	Olivine basalt and other dark volcanic rocks and minor acid volcanic rocks in a brownish, reddish, purplish, or greenish, altered, fine-grained matrix; agglomerate - mud flow?	Lower andesite.
190'	Covered, same as above?	
-?-?-?-?-	Precambrian.	

> extending into the first flows of the upper andesite, is also well exposed. This locality is perhaps the most critical of the entire Thirtynine Mile volcanic field, for it is here that the thickest and most complete stratigraphic sequence is preserved and exposed for study."

Table 11 contains a volcanic section measured at the northwest end of Echo Canyon and its correlation with the composite stratigraphic column presented by Chapin and Epis (1964, Plate II).

Chapin and Epis (1964) conclude that the source of the Thirty-nine Mile volcanics was near Guffey, where they believe they have found a caldera. They state that the oldest units are no older than middle Paleocene, and that the youngest units are late Miocene or early Pliocene in age. Potassium-argon age dates on volcanic rocks in Wet Mountain Valley range from 33.2×10^6 to 41.2×10^6 years -- early Oligocene (Robert MacNish, pers. comm.). Chapin and Epis (1964) correlate these volcanic rocks (largely tuffs) with the Thirty-nine Mile volcanic field.

Basaltic Agglomerate

A much altered and highly variable unit forms the base of the volcanic section in Echo Canyon. It varies from agglomeratic to conglomeratic and tuffaceous. The matrix of the rock varies from dense or pasty to sandy, and its color is generally mottled in dark shades of red, brown, purple, or green and black. The fragments set in this matrix are aphanitic to porphyritic andesite and basalt. They may be angular or round and range from less than an inch to about 3 feet in diameter. The unit shows very little sorting.

Microscopic examination shows the matrix to be a highly altered lithic tuff. The volcanic rock fragments show a wide range in texture and mineralogy, but most appear to be basalt or andesite.

Chapin and Epis (1964) consider this unit to be a lahar originating

in volcanoes near Guffey, which is also the opinion expressed to the author by Dr. J. E. Bever (pers. comm., 1962). The very chaotic nature of the rock lends support to this view. The Guffey volcanic center is 15 miles to the north. This is perhaps too far for mud flows to travel and may indicate a nearer source for the agglomerate.

Rhyolite Tuff

The most widespread basal volcanic unit is rhyolite tuff. The unit generally consists of three layers, distinguished by differences in the degree of welding. The lower layer is white to light buff in color, highly porous and light-weight, and unwelded (Fig. 42). It may be missing but where present ranges up to about 150 feet in thickness. The lower layer is commonly overlain by a thin (up to 25 feet thick where present), strongly welded (Fig. 43), medium to dark gray layer that in some places resembles obsidian. The upper layer may be the only one present; it ranges up to about 500 feet in thickness. It is buff to pinkish brown in color and is moderately welded (Fig. 44).

All layers are vitric tuffs which have the general composition: 50% glass, 30% crystal fragments, and 20% lithic fragments. The glassy constituents are shards and pumice fragments, which are commonly devitrified and altered. Axiolitic textures are common (Figs. 46 and 47). Crystal fragments commonly present are quartz, sanidine, plagioclase, microcline, augite, hornblende, biotite, sphene, and magnetite. Most rock fragments are acid to basic volcanics but Precambrian gneiss fragments are common. The rock fragments range up to about a foot across and attest to the violence of the volcanic eruptions.

Most of the exposed tuff shows no evidence of water transport and is therefore interpreted as an ash flow deposit. Chapin and Epis (1964)

Figure 42. Photomicrograph of unwelded rhyolite tuff from upper East Gulch. Field includes shards(S), pumice fragment(P), and quartz crystal fragments(Q). 43x.

Figure 43. Photomicrograph of strongly welded rhyolite tuff from northwest wall of Echo Canyon. Note collapse of pumice fragments(P), molding of shards and pumice fragments around crystal fragments, and embayment of quartz crystal fragment(Q) by glassy matrix. 43x.

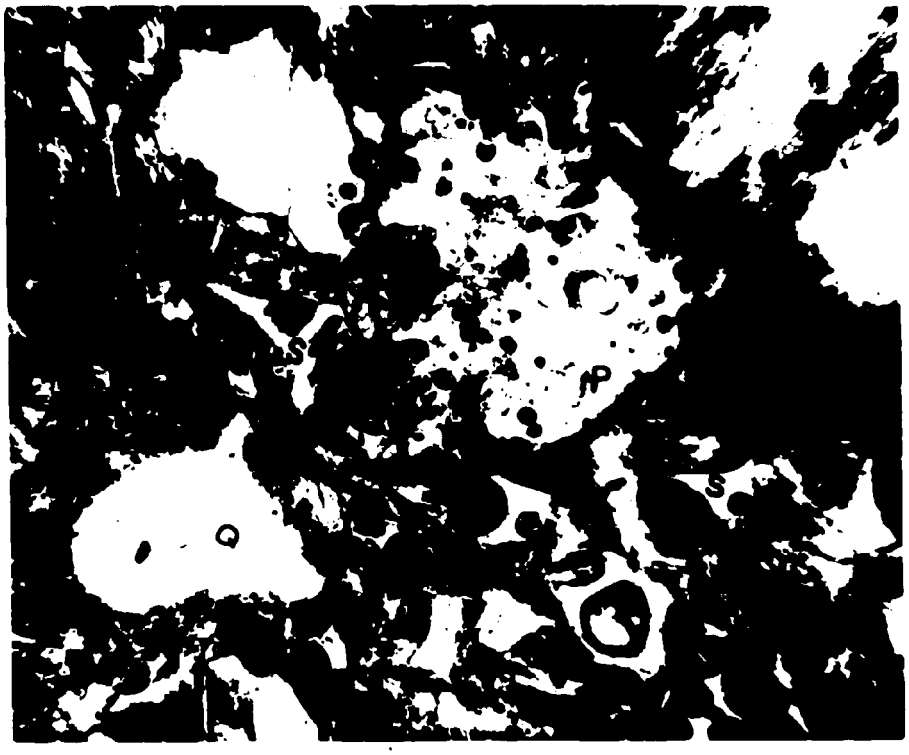


Figure 42.



Figure 43.

Figure 44. Photomicrograph of moderately welded rhyolite tuff. Shards are still distinct, although they have been flattened and molded around quartz crystal fragments(Q). 43x.

Figure 45. Photomicrograph of andesite. Field includes microphenocrysts of pyroxene(P) and partly altered olivine(O) and groundmass(G) of plagioclase, pyroxene, and magnetite. 43x.



Figure 44.

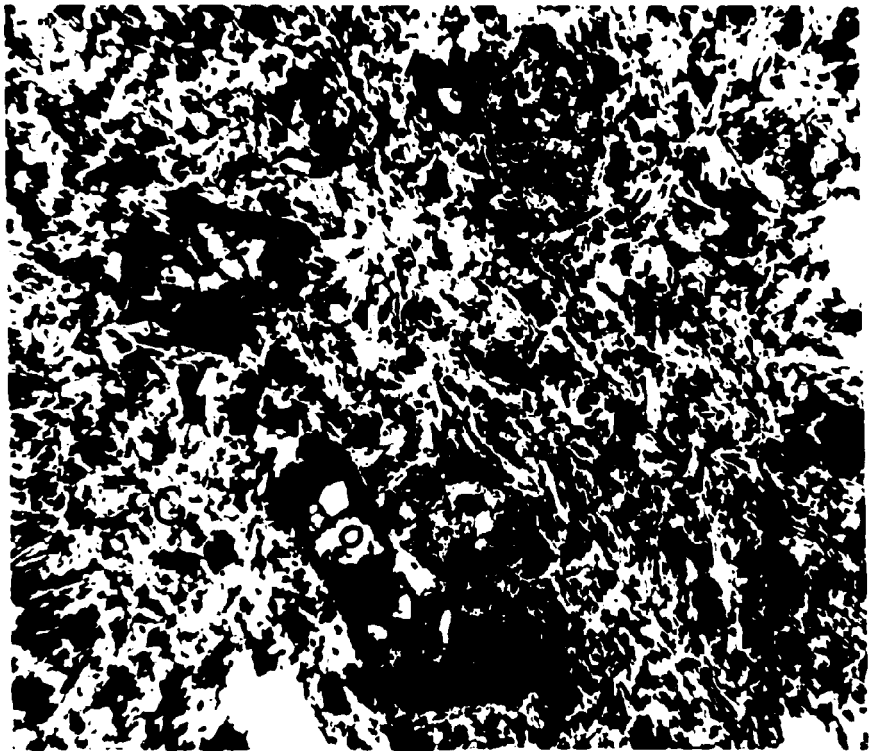


Figure 45.

Figure 46. Photomicrograph of devitrified and altered, slightly welded tuff. Field includes pumice fragments(P) and shards(S). 43x.

Figure 47. Photomicrograph of same field as in Fig. 46 but with crossed polars, showing axiolitic texture in devitrified pumice fragment and shards. 43x.

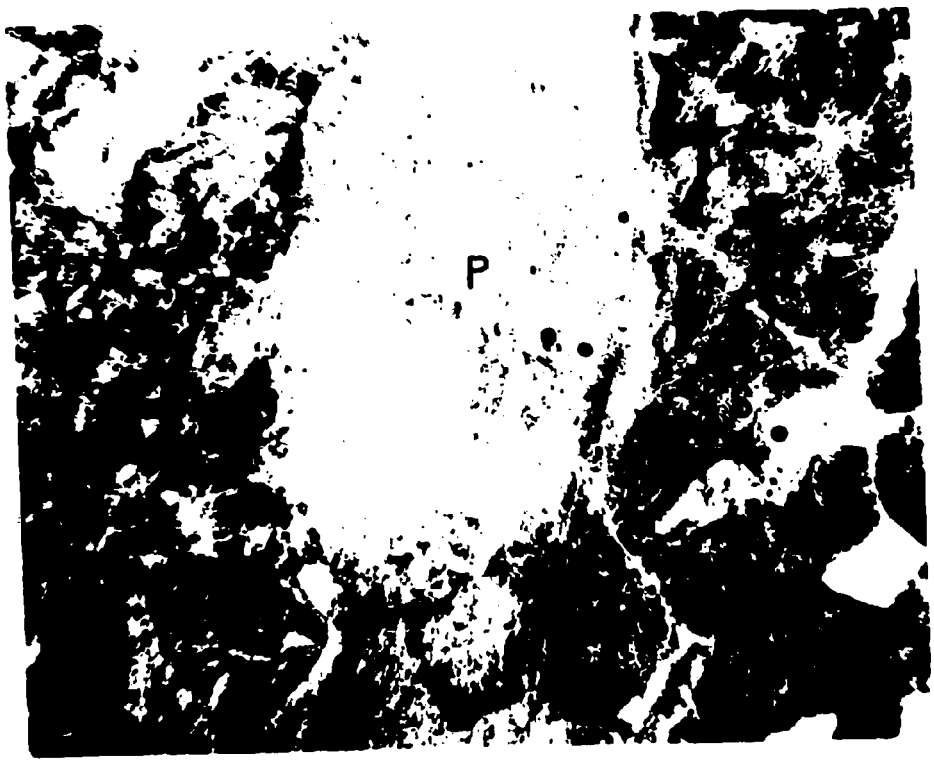


Figure 46.

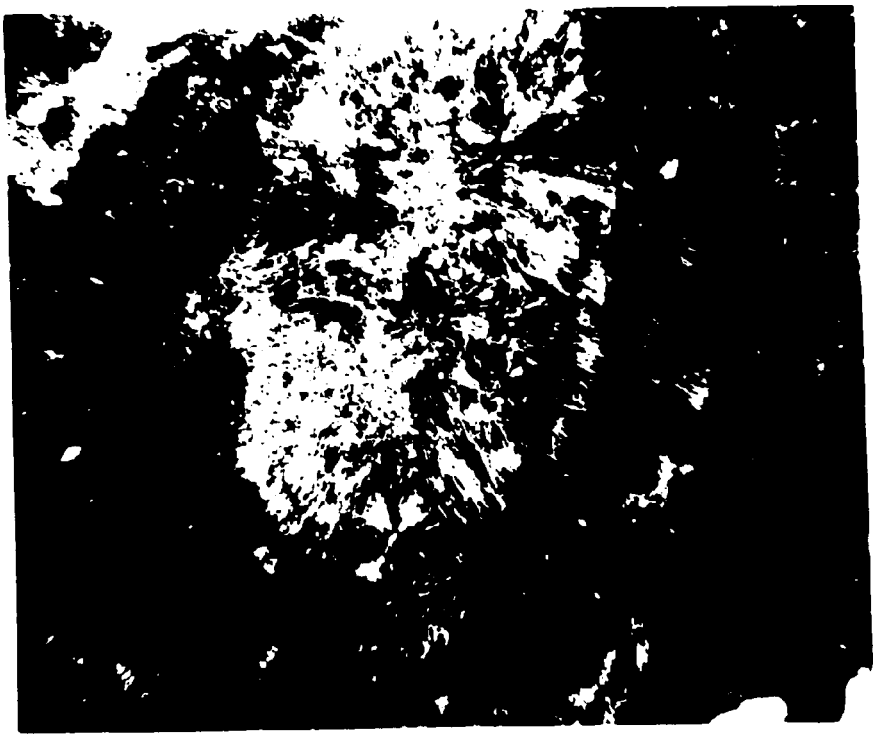


Figure 47.

feel that the ash flows of the Thirty-nine Mile volcanic field represent an estimated 30 cubic miles of material erupted from vents near Guffey. The lower part of the unit in Echo Canyon is sandy, stratified, and shows some sorting; it is probably a water-laid deposit or an ash flow that was reworked by water. Chapin and Epis (1964) correlate this layer with the Antero formation (see Table 11).

Intravolcanic Sediments

Volcanic Conglomerate

A bed of volcanic conglomerate separates the underlying rhyolite tuff from the overlying andesite flows. The unit varies from about 100 to over 300 feet in thickness where it is exposed in the cliffs north of Echo Canyon (Fig. 48). Elsewhere the presence of the unit is revealed only by abundant, well-rounded volcanic cobbles on slopes between outcrops of rhyolite tuff and andesite and thicknesses can only be estimated.

In the cliffs north of Echo Canyon the unit consists of cobbles of andesite, basalt, and, less common, Precambrian rocks. The cobbles are set in a sandy to pebbly matrix that is weakly calcite-cemented. The conglomerate is thickest in what is apparently a channel in the rhyolite tuff unit at the northwest end of Echo Canyon, where it also contains beds of tuffaceous sandstone. The lower contact of the unit is not exposed in the area but the upper contact is well exposed at the northwest end of Echo Canyon. There the conglomerate is sharply overlain by a two foot layer composed of beds of reddish brown siltstone and sandstone; this layer is overlain by andesite scoria (Fig. 49). The abundant andesite cobbles indicate a source to the north where andesite flows had already been deposited and subjected to erosion.

Figure 48. Contact between volcanic conglomerate (below) and tuffaceous sandstone (above), northwest wall of Echo Canyon.

Figure 49. Contact between volcanic conglomerate (below) and scoriaceous andesite (above), northwest wall of Echo Canyon.



Figure 48.



Figure 49.

Shappirie (1962) describes volcanic conglomerate occupying a similar stratigraphic position. Chapin and Epis (1964) noted the same conglomerate in the Tallahassee Creek area and designate it "Lower Conglomerate" in their composite stratigraphic column. However, the conglomerate in the Devils Hole area is unquestionably the "Upper Conglomerate" of Chapin and Epis (1964), which they tentatively correlate with the Trump and Wagontongue formations (upper Miocene - Pliocene ?) of South Park.

Tuffaceous Sandstone

Lenticular beds of tuffaceous sandstone are exposed in two places along the cliffs at the north end of Echo Canyon (Fig. 48). Both occurrences are in filled channels in the rhyolite tuff unit. The beds overlie and are overlain by volcanic conglomerate (Table 11). The sediment is distinctly layered and ranges from coarse- to very fine-grained tuffaceous sandstone. The beds are white to light buff and salt and pepper gray. The sandstone is poorly consolidated, and coarse-grained types are extremely friable. Bedding planes within the unit and the upper and lower contacts of the unit are essentially horizontal (Fig. 48).

The rock is composed predominantly of quartz and fresh to highly altered glass fragments. Feldspars are also abundant and biotite and hornblende grains are the dark specks in the rock. Hornblende grains are subrounded in contrast to the angular shapes of the other constituents; they were probably derived from Precambrian rocks and have undergone more working by water.

Chapin and Epis (1964) also noted "tuffaceous strata" in the conglomerate exposed in the cliffs north of the Devils Hole (Echo Canyon). They ascribe the origin of the volcanic conglomerate and tuffaceous sandstone to the filling of channels eroded in the rhyolite tuff. It

is possible that post-tuff faulting in the graben localized the channels because they occur between the graben-bounding faults.

Andesite

Andesite flows, flow breccia, and scoria constitute the upper volcanic unit. They are best exposed on Table Mountain and north of Echo Canyon (Plate I). Dense flows and scoria beds are about equal in abundance, and scoriaceous, brecciated, oxidized lava tops are exposed in several places. The unit seems to be characterized by a great variety of textural types and must represent a very complex sequence of eruption and lava flow distribution. Some flows are thick and massive, others are thin and have a well developed platy fracture.

Fresh surfaces of dense flows are typically colored dark gray to black. Highly oxidized flow tops and some scoria units are a deep brick red; this oxidation probably occurred during extrusion because it is general, whereas obvious weathering phenomena are superficial.

Microscopic examination reveals that textures range from intergranular to porphyritic, from dense to vesicular, and that flow orientation of plagioclase microlites is common. Augite and olivine are common phenocrystic species (Fig. 45). Rare plagioclase phenocrysts are labradorite, but groundmass plagioclase brings the average plagioclase composition into the andesine range. Clinopyroxene and plagioclase are the predominant groundmass constituents. Accessory magnetite is common and brown glass is present interstitially in some samples. Rare phenocrystic species are oxyhornblende, biotite, and hypersthene. Xenocrystic quartz may be present. Alteration products include clay after volcanic glass and iddingsite after olivine.

An olivine-bearing andesite dike, exposed on the Table Mountain

road (Plate I), is the only Tertiary intrusive in the area. It contains olivine microphenocrysts set in a groundmass of clinopyroxene, plagioclase, and magnetite. The dike was intruded along the contact of a granite dike with biotite gneiss. Plagioclase microlites are strongly flow oriented near the margins of the dike.

QUATERNARY DEPOSITS

TERRACE GRAVELS

Remnants of terrace gravels occur along much of the Arkansas River (Fig. 50). They lie at elevations ranging from 20 to 40 feet above the present river. The gravels were deposited during the Wisconsin glacial stage (see p. 17).

BASIN DEBRIS

Much of Echo Canyon is covered by up to 100 feet of erosional debris (Fig. 51; Plate I). The material is thicker and much more chaotic and heterogeneous than typical talus or aluvium. The debris was derived predominantly from the volcanics to the north, but some also came from the high ridge to the east and the lower ridges to the west. The unit thins in the southern part of the canyon, but everywhere obscures most of the Farisita (?) clastics (Fig. 41). The material is very poorly sorted and was probably just dumped by many small streams heading in the surrounding high areas.

LANDSLIDE DEBRIS

Numerous landslides have occurred along the erosional escarpment developed in the volcanics in the northeast part of the area (Plate I). The landslides consist of huge slumped blocks, several hundreds of feet on a side, and chaotic heaps of small blocks and fragments. The tops of the slides are flat or dip gently opposite to the direction of the slope of the escarpment (Figs. 6 and 7). The lower ends of the slides merge indistinctly with talus or, in Echo Canyon, with basin debris. The slides have involved primarily the rhyolite tuff unit. The outlying patches of rhyolite tuff south of Table Mountain (Plate I) appear to be in place but may actually be remnants of landslides from Table Mountain;

this would explain their low elevation.

The landslides are not very recent. They are grown over with vegetation and no scars are visible. Also, in the Little Hole area east of Echo Canyon a depression (noted as a "dry lake" on the United States Geological Survey topographic map of the Royal Gorge quadrangle) caused by the reversal in slope at the top of a large landslide has been partly filled by alluvium. This may indicate a climatic change, the climate having been much wetter when the landslides occurred. The climatic change probably corresponds to the end of the Pleistocene.

TRAVERTINE

Travertine-cemented talus and alluvium occur along some of the faults and shear zones (Plate I)(Fig. 52). The cementing is not tight and the travertine occurs interstitially rather than in pure masses. The travertine-cemented material occurs in thin patches that are generally elongate parallel with the associated fault zone. The deposits may be thinly covered with uncemented talus or alluvium, indicating that the travertine deposition is not very recent. However, travertine deposits in Eight Mile Park near the north rim of the Royal Gorge appear to be very recent (Heinrich, pers. comm.) The travertine was probably deposited by ground water or hot springs rising along the fault zones. Carbonatite dikes are numerous south of the Arkansas River (Dahlem, 1965) and may be an important source of the $CaCO_3$.

ALLUVIUM

The Arkansas River and some of the large streams have deposited extensive blankets of alluvium on their flood plains or valley floors (Plate I). These deposits are thick and well sorted relative to surficial deposits elsewhere in the area.

Figure 50. Gravel terrace of Wisconsin age, railroad cut 1½ miles northeast of Texas Creek.

Figure 51. Basin debris, exposed along a gulley in Echo Canyon.



Figure 50.

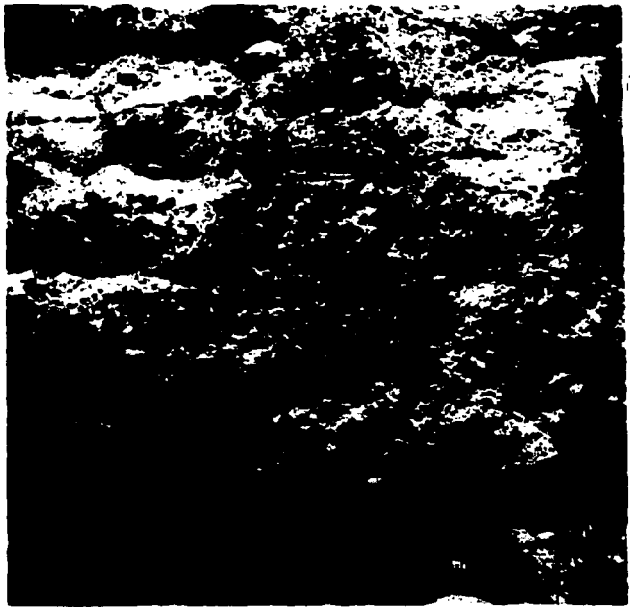


Figure 51.

Figure 52. Travertine-cemented alluvium along Colorado Highway 69. Note thin cover of uncemented alluvium.

Figure 53. Foliation at angle to compositional layering in biotite schist, 2 miles northeast of Texas Creek.



Figure 52.



Figure 53.

STRUCTURE

PRECAMBRIAN STRUCTURE

Structure of the Idaho Springs Formation

Foliation and Lineation

Foliation in the Idaho Springs formation ranges from strong to undetectable. Mineral alignment, especially parallelism of biotite flakes, is the predominant foliate feature. As in other areas where Idaho Springs rocks are exposed, the foliation is parallel with the compositional layering and is interpreted as bedding plane foliation (Lovering, 1935). This parallelism is constant, no matter what the present orientation of the layers, and indicates that the bedding planes were the planes of least stress during metamorphic recrystallization, no matter what the relative attitudes of the bedding and stress field. Rare deviations from this parallelism (Fig. 53) occur in "hot spots" where the rocks may have become so plastic that bedding planes ceased to exert dominant control.

Lineation is prominent in some rocks and, rarely, constitutes the only oriented feature of the rock. Elongate biotite flakes, hornblende prisms, sillimanitic pods, and axes of ptygmatic folding are the linear elements present in the area. Lineations invariably plunge at the maximum angle in the plane of the foliation where foliation is detectable.

Major Folds

Two major anticlinal or domal structures dominate the western two-thirds of the area (Plate I). In the southwest, the Texas Creek anticline is outlined by the contact of the central mass of Pikes Peak granite and the foliation (bedding attitudes) in the enclosing metamorphics. The steeply plunging nose of the Pikes Peak core is repeated to

the east by the Texas Creek fault. The Antelope Gulch anticline is outlined by comformable contacts and foliations in the northwest. The folds are pre-faulting in age and probably also pre-granite in age. The writer believes that the folds controlled the emplacement of the Pikes Peak granite and the granite gneiss (if it is meta-igneous) but were broadened and otherwise modified by the intrusives.

Primary Magmatic Structures

The internal structure of the individual granites has been described. It is necessary, here, only to point out that the shape and attitude of granite bodies was largely controlled by pre-existing structures in the metamorphics. The only prominent internal structural feature, the foliation of the Pikes Peak granite, is probably primary flow foliation. This interpretation is supported by the lessening degree and regularity of the foliation inward from the contacts and the micro-textural relations (igneous, not metamorphic) of the essential minerals.

Faults

The age of the faulting in the area cannot be established with certainty. All fault zones are apparently post-granite (Fig. 54). That some are pre-diabase is evident in that many diabase dikes were intruded along them. That many of these diabase dikes are strongly sheared indicates post-diabase movement in the fault zones (Fig. 38). It is probable that most of the faults were formed in Precambrian time. They were possibly reactivated several times in the Precambrian, again in the Paleozoic during the uplift of the Ancestral Rockies, and certainly many were reactivated during the Laramide (Tweto and Sims, 1963).

Figure 54. Shear zone with abundant quartz veins in Pikes Peak granite. Upper Reese Gulch.

Figure 55. Shear zone between biotite gneiss (left) and Pikes Peak granite dike (right), Texas Creek Gulch, 1½ miles northwest of Texas Creek.

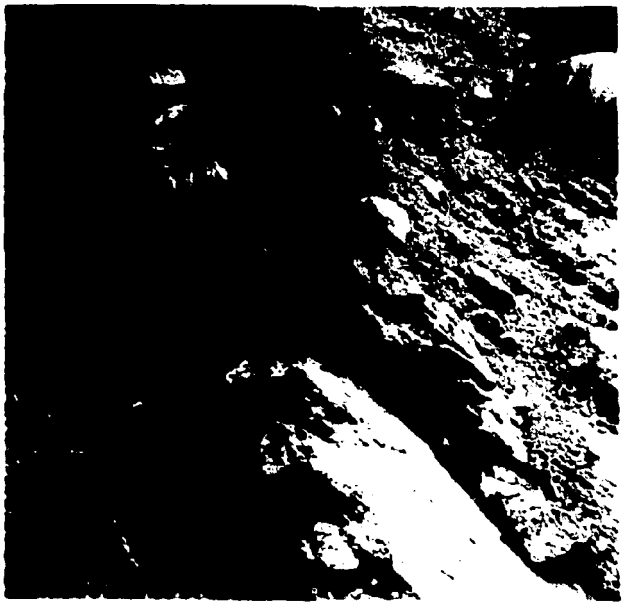


Figure 54.

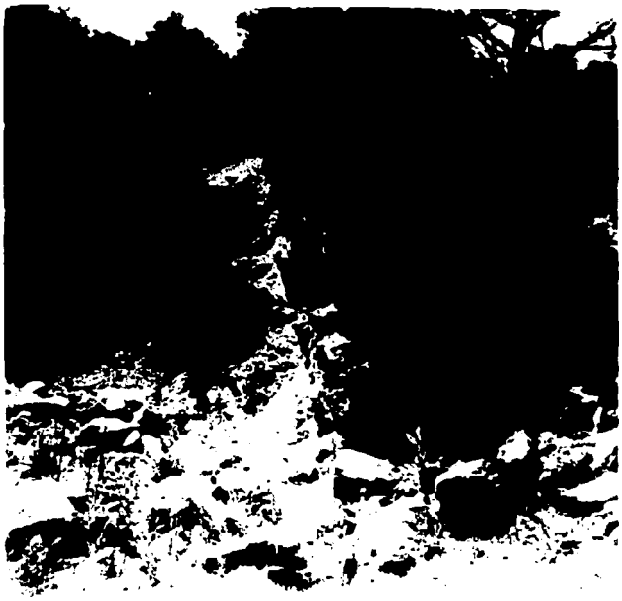


Figure 55.

The north-northeast trending Texas Creek fault zone is the major fracture in the area (Figs. 4 and 55). It cuts Pikes Peak granite and older rocks but passes under the volcanics on Table Mountain without displacing them. It extends southward into Wet Mountain Valley but its northward extent is obscured by the Tertiary volcanics. It is upthrown on the east; the repetition of the nose of the Texas Creek anticline represents a vertical displacement of as much as 10,000 feet.

The major northeast trending fracture is the Devils Hole fault zone. It crosses the Texas Creek fault zone but their relative ages are not apparent. Dahlem (pers. comm.) has found that the northeast faults offset the north-northwest faults in the Lookout Mountain area. Northeast-trending Precambrian shear zones are prominent in the Colorado mineral belt (Tweto and Sims, 1963).

Other, smaller fault zones generally parallel the above major trends.

SANDSTONE DIKE

An exposure of what is interpreted to be a sandstone dike occurs on the east side of one of the forks of Texas Creek Gulch in the NW¼, Sec. 2, T48N, R12E. The dike, which is about two feet wide and 20 feet long, occurs in the highly sheared and altered rocks of the Texas Creek fault zone. The contacts between the dike and the chloritic wall rock trend N10°E and dip 50°E. A nearby brecciated zone strikes N10°E and dips 80° to the east.

The sandstone is an ortho-quartzite; the quartz grains range from well- to sub-rounded and are from 0.07 to 1.02 mm in diameter (average 0.3 mm in diameter). Some grains display dust rings and oriented quartz overgrowths. The sediment was moderately-well sorted and, in thin section, appears very similar to the photomicrograph shown by Hutchinson

(1960, Figure 7, Plate 3, p. 173). The rock is criss-crossed by close-spaced, calcite-filled fractures and zones of granulation. Granulation along grain boundaries also is common. Wavy extinction of the quartz grains is not pronounced, and sutured grain boundaries occur only in partly recrystallized granulated streaks. These features are characteristics of a sheared sedimentary ortho-quartzite, not those of a metamorphic quartzite. The sandstone dike was formed through the filling of a fissure in the Precambrian rocks by sand sifting down from overlying sediments, then consolidated and later subjected to minor fracturing. The following general conclusions can be drawn:

1) That the sand entered the fissure by a relatively gentle sifting and settling action rather than by forceful intrusion is indicated by the lack of apophyses from the dike into the highly fractured wall rocks.

2) The unmetamorphosed character of the sandstone indicates that it is Paleozoic or younger in age.

3) The existence of a fissure connected to the surface is evidence of a shallow environment. At the present time the dike outcrop may be from 200 to 1000 feet below the pre-volcanic surface, depending on how one interprets the rhyolite tuff outlyers south of Table Mountain (see p. 142).

4) The fissure probably formed in response to small adjustments along the Texas Creek fault zone.

5) Some movement in the fault zone subsequent to the formation of the dike is evidenced by the shearing and granulation of the sandstone.

6) Possible sources of the sand include the Sawatch (Cambrian) and Dakota (Cretaceous) formations. Both are basal sediments in nearby

areas. The possibility of a Fountain (Pennsylvanian-Permian) or a Farisita (?) (Eocene) source cannot be ignored but is unlikely in view of the clean and relatively well-sorted nature of the sand. A Sawatch source has been suggested for the sandstone dikes in the Platte Canyon and Kessler quadrangles (Peterson and Scott, 1960).

LARAMIDE AND TERTIARY STRUCTURE

Faults

The Echo Canyon graben, although perhaps originating in the Precambrian, shows the strongest evidence of Laramide activity. There are no large faults observable in the volcanics overlying the graben to the north, therefore, the faulting is largely pre-volcanic. However, the rhyolite tuff unit appears to be lower in the graben and, in places, is separated from the andesite by a much greater thickness of conglomerate than separates the two units outside the graben. Whether this apparent down-warping has resulted from the deposition of the tuff on an uneven topographic surface (lower in the graben), from channeling in the tuff unit, from the slumping of large blocks of tuff, or from post-tuff faulting is not clear.

The contact relations between the Farisita (?) clastics and the Precambrian are obscure. If the Farisita (?) - Precambrian contact is a fault contact, as is probable, then most of the faulting occurred between the deposition of the Farisita (?) clastics and the deposition of the rhyolite tuff. There is at least 1000 feet of relief between the Precambrian surfaces within and outside of the graben. It is highly unlikely that such a sharp topographic feature could be preserved from Precambrian, or even Paleozoic, faulting. One cannot call for the present canyon to be an exhumed Precambrian or Paleozoic feature because

present erosion is only now exposing the Farisita (?) clastics.

GEOLOGIC HISTORY

The oldest rocks exposed in the area are the schists and gneisses correlated with the Idaho Springs formation. They represent a thick accumulation of mudstones and impure calcareous sediments and impure sandstones in a miogeosynclinal environment. They reached their present state through regional metamorphism and subsequent thermal and metasomatic modification by a series of granite intrusives.

The Idaho Springs sediments were probably deposited in middle Precambrian time and are at least 1.3 billion years old. After deposition and burial the sediments were converted to schists and gneisses by regional metamorphism under intensities corresponding to the almandine-amphibolite facies. The Idaho Springs rocks were folded into broad, northwest-trending anticlines and synclines. The limbs of the folds are steeply dipping and the fold axes have moderate to steep plunges. Metamorphic recrystallization and folding were probably synchronous, but foliations were controlled by inherited bedding plane weakness rather than by the orientation of the stress field and structural position in the folds.

A period of granite formation and emplacement represented by three, and possibly four, distinct granites began after the Idaho Springs rocks had become regionally metamorphosed and folded, but the granite cycle was probably part of the tectonic period that produced the metamorphism and folding. The granite bodies are, in the order of emplacement indicated by field relations; granite gneiss (if it is of igneous parentage), Cripple Creek (?) granite, Pikes Peak granite, and Silver Plume (?) granite. The emplacement of the main granite bodies and associated sills and dikes was controlled by but caused modification of the pre-existing

structural configuration and marked the end of strongly directed regional stress. Migmatites were produced by sill and dike injection associated with the Cripple Creek (?) granite and, to a lesser extent, the Pikes Peak granite. Minor thermal metamorphism by the Pikes Peak granite produced cordierite in biotite gneiss and pyroxene in amphibolite. Later, Al-rich residual fluids emanating from the Pikes Peak bodies caused extensive sillimanite formation in favorable layers. Still later, K-rich solutions produced widespread muscovitization, especially of silimanite.

Faulting, possibly reflecting adjustments at the close of the same Precambrian tectonic period, produced broad shear zones with north-northwest and northeast trends. Diabase dikes, representing the last Precambrian intrusive activity recorded in the area, were emplaced along strong fractures and shear zones.

The Paleozoic and Mesozoic history of the area is unrecorded. Sediments preserved in down-faulted blocks to the east (Shappirio, 1962), northeast (Bever, 1954), and west (Salotti, 1960) indicate that the area may have been thinly covered by lower Paleozoic sediments but that these were probably stripped away upon the uplift of the Ancestral Rockies. Upper Paleozoic sediments preserved to the west and north may have covered the area. The area probably was positive throughout much of the Mesozoic but may have been covered by the Cretaceous Dakota, Pierre, and younger formations. In any case, Laramide uplift and subsequent erosion have resulted in the removal of any Paleozoic and Mesozoic sediments that may have been present in the Devils Hole area.

The area contains a relatively complete record of Cenozoic events. The Farisita (?) clastics of probable Eocene age are a local accumula-

tion or are locally preserved in the Laramide Echo Canyon graben. In either case the clastics were probably produced during the period of erosion which formed the pre-volcanic surface. That surface is one of moderate relief onto which were deposited the volcanics of the Thirty-nine Mile volcanic field.

The earliest phase of volcanic activity is represented by basaltic agglomerate. This unit is possibly a mud flow or flows (lahar) that originated on the slopes of volcanoes to the north; it is preserved only in the Echo Canyon graben. Volcanic activity continued with the accumulation of rhyolitic ash flow tuff deposits. The lower layers of tuff in Echo Canyon were water laid and may be correlatable with the Antero formation of lower Oligocene age (Chapin and Epis, 1964). The tuff beds are overlain by volcanic conglomerate locally containing beds of tuffaceous sandstone. These sediments were produced by the erosion of andesite flows which had succeeded the eruptions of ash in the Guffey area. The conglomerate may be correlative with the Trump and Wagontongue formations (upper Miocene-Pliocene?) in South Park (Chapin and Epis, 1964). The final phase of volcanism is represented by the complex andesite flows which overlie the conglomerate.

Pliocene-Pleistocene uplift and subsequent erosion have produced the present topography. The volcanics have been largely stripped from the area and the underlying Precambrian rocks have been moderately to deeply disected. In many places the present volcanic front is an erosional escarpment held up by the cap-rock effect of the andesite flows. This front has receded as much as 8 miles north from the Arkansas River. Landslides have occurred along much of the escarpment, principally through the slumping of large blocks of the rhyolite tuff unit.

Events of Pleistocene to recent age include the formation of two sets of terraces along the Arkansas River, the lower, gravel terraces forming during the Wisconsin glacial stage; the partial filling of Echo Canyon by erosional debris washed down from surrounding high areas, principally from the landslides and escarpment to the north; the deposition of travertine by waters rising along some of the fault zones; and the alluviation of the floors of the larger valleys.

ECONOMIC GEOLOGY

GRANITES

A small amount of Silver Plume (?) granite has been quarried west of Texas Creek, about a mile and a half south of the Arkansas River. The operation is recent but no quarrying was being done in the summers of 1961 and 1962.

The large pegmatitic granite plug at the mouth of Echo Canyon is quarried north of the Arkansas River by the Denver and Rio Grande Western Railroad. The pegmatite is mined in two open cuts; the larger blocks blasted from the cuts are loaded by crane onto flat cars. The operation is served by a side track connecting with the Echo siding. Blocks of the pegmatite can be seen along the railroad in the area; presumably, their purpose is to protect the railroad bed from undercutting by the Arkansas River.

PEGMATITES

General

Many pegmatites are exposed in a zone extending east-northeast from Table Mountain to the north end of Devils Hole. The zone is about a mile wide and is over 3 miles long. Many of the pegmatites are zoned and at least eleven of them have been explored by open cuts. Three of these have been more extensively mined than the others and are described below.

Devils Hole Pegmatite

The Devils Hole pegmatite has been the most important economically of all the pegmatites in the area. It has been mapped and described in detail by Heinrich and Hanley (Hanley et al., 1950). As of 1944, at least 300 tons of beryl, 16,000 tons of feldspar, 1,600 tons of scrap

mica, and about 300 pounds of columbite had been mined. Since that time no extensive mining has been done. The mine is presently operated by Ralph Pierce of Canon City. In the summer of 1962 the mine was intermittently worked for small amounts of beryl and columbite and an occasional truck load of rose quartz, which is valued as an ornamental stone.

Chief Lithium Pegmatite

General

The Cheif pegmatite is in the SE ¼, Sec. 17, T.18S., R.73W., approximately one mile north-northeast of the Devils Hole pegmatite. The deposit lies on the northern side of East Gulch, about ¼ mile northwest of the point where East Gulch enters Devils Hole. In the vicinity of the pegmatite the country rocks are gneisses and schists of the Idaho Springs formation.

The pegmatite, which is on several of a group of five claims (Chief Nos. 1-5), was located by P. A. Weatherson and Sidney G. McClue on June 11, 1957. The early history of the deposit is not well known. Apparently it was worked for feldspar in the early or middle 1950's. During that time some feldspar was sold to the feldspar mill of the International Minerals and Chemicals Corporation in Canon City, Colorado. However, lithium minerals were rumored to have been found in this area as early as 1956. The Chief pegmatite was explored for lithium and beryllium in the late 1950's by William McClelland.

The deposit was first examined by E. Wm. Heinrich in 1959 and mapped by him in 1960. It was restudied by the writer and a joint paper has been published (Heinrich and Vian, 1965).

The Chief pegmatite has been explored by means of two open cuts.

The eastern cut, from which the feldspar was mined, is approximately 100 feet long with a generally east-west axis, about 25 feet across, and as much as 35 feet deep at the face. The western cut, in which the lithium pegmatite is exposed, is about 25 feet across and has a maximum depth of about 12 feet (Fig. 56). There are, in addition, several small prospect pits and trenches.

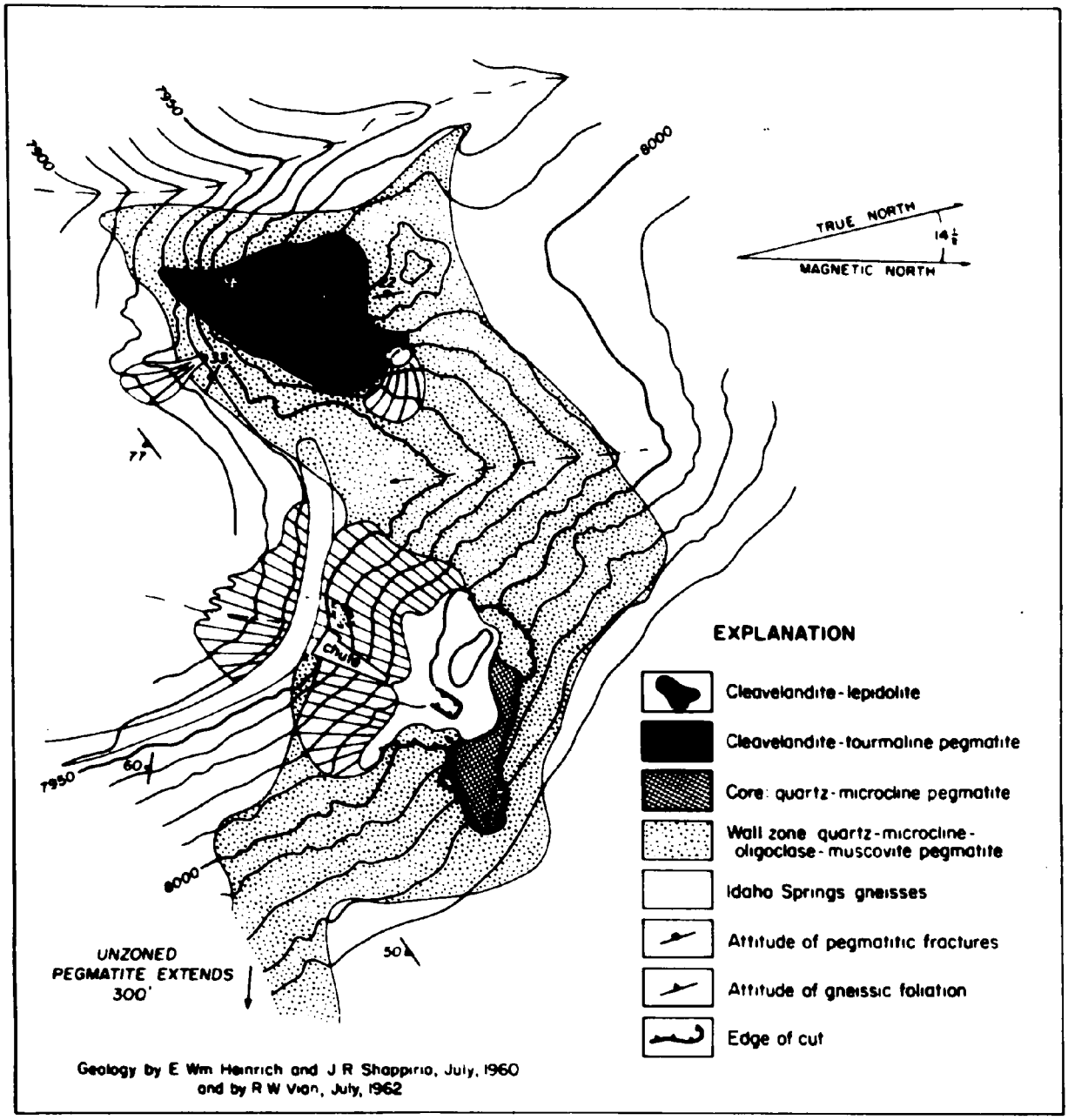
The entire pegmatite is much larger than the part that has been mapped. The mapped section is the thick, arcuate, western end of a generally northeasterly trending dike (Fig. 56). Beyond the mapped area to the northeast the pegmatite continues for at least another 300 feet, but is neither commercially nor mineralogically interesting.

The mapped part of the pegmatite, which is the thickest section of the dike and the only part that is distinctly zoned, has the form of a letter "C", open to the south (Fig. 56). The eastern half of this arcuate segment trends south of east and dips northeast (i.e. into the slope of the hill) at moderate angles. The western "tail" trends at right angles to this quarried part, striking just east of north and dipping generally westward. East of the area mapped the dike strikes northeast, dips moderately northwest and is not zoned.

Internal Structure

Within the pegmatite, five main rock types are represented (Fig. 56).

1) Wall zone rock which consists of fine- to medium-grained pegmatite containing quartz, microcline, plagioclase, muscovite and minor amounts of biotite, garnet, and tourmaline.
2) Quartz-feldspar core. This is exposed only in the eastern cut and represents the feldspar ore body previously mined.
3) Irregular small units of black tourmaline and muscovite. Some



GEOLOGIC MAP OF THE CHIEF Li PEGMATITE, DEVILS HOLE, COLORADO

0 50 100 150
FEET

CONTOUR INTERVAL 10'

Figure 56

of the tourmaline crystals are several inches across and some of the books of muscovite are as much as 1¼ feet across. This unit, which is exposed only in the eastern cut, is too small and too irregular to be shown on the scale of the map (Fig. 56).

4) Cleavelandite-tourmaline rock.

5) Cleavelandite-lepidolite rock.

Units 4 and 5 are found only in the western "tail" of the pegmatite, and Unit 5 is exposed chiefly in and near the western open cut.

The cleavelandite-rich units grade into each other and show diffuse and gradational contacts with wall-zone rock. In the marginal parts of the cleavelandite-tourmaline unit corroded microcline remnants are common. It is noteworthy, however, that the albitization is not localized along a zone contact, as is so commonly the case in albitized pegmatites of some other districts (see, for example, Heinrich, 1948; Jahns, 1946).

Within the westward dipping "tail" of the pegmatite there is developed a conspicuous planar structure consisting of a series of parallel fractures that also strike northerly and dip westward. Such fractures are absent elsewhere in the dike. The cleavelandite units are believed to have been formed as replacement units of wall-zone pegmatite, localized along this multiple fracture system.

Mineralogy

Wall zone. Most of the wall zone consists of anhedral granular microcline, quartz, and subordinate sodic plagioclase. Books of muscovite several inches across are clustered in some places. Some of these include needles of scherl. Locally the rock is peppered with rounded spessartite grains, as much as ½ inch in diameter, set in a conspicuous dark halo. This halo consists of a rim of granular blue tourmaline and

a fine-grained flaky aggregate of a bronze mica which optical and chemical tests indicate is zinnwaldite (γ = 1.564 - 1.570 ± 0.002). No lepidolite appears in the wall zone other than a very narrow outer rim overgrown on some of the larger muscovite books.

Cleavelandite-tourmaline pegmatite. The larger of the two cleavelandite-rich units contains chiefly cleavelandite, quartz, microcline, muscovite, black tourmaline and green tourmaline. Green tourmaline, which is very abundant locally, appears as:

1) Rims and outer sh lls on schorl crystals.

2) Slender prisms, 6 inches or less in length, or prismatic clusters.

3) Veinlets of very fine anhedral grains (called "epidote" locally).

The microcline and muscovite and much of the quartz are relict from wall-zone rock.

Cleavelandite-lepidolite pegmatite. Within the cleavelandite-tourmaline unit, near its northeastern corner, is a small, irregular, tabular mass which, at this spot, appears to underlie the main body of cleavelandite-tourmaline pegmatite, although to the south, cleavelandite-tourmaline rock appears beneath it. Thus originally it was probably enclosed in cleavelandite-tourmaline pegmatite.

This body measures 55 by 20 feet in plan. It is as much as 15 feet thick at its northern end but is very poorly exposed along the south side of the western cut. It appears to dip about 60° to the west, and its long axis may plunge at a somewhat more shallow angle to the southwest.

Within it are exposed two distinctly different types of pegmatite. The upper part of the exposure, which forms a very pronounced erosional knob, consists almost entirely of cleavelandite and flakes of lepidolite that range in diameter from ¼ to 1 inch, the mica being rather uniformly

distributed through the cleavelandite. Beneath this rock the unit consists chiefly of cleavelandite with scattered larger books of lepidolite, a few of which are as much as 10 inches across, although most measure 2 to 6 inches.

The two lepidolites compare as follows:

	γ	Li_2O
Flakes, $\frac{1}{4}$ - 1"	1.561 ± 0.002	2.4%
Coarse, thick books	1.560 ± 0.002	5.3%

Other minerals present as accessories in the cleavelandite-lepidolite unit are:

beryl, a few scattered prisms, very pale blue, up to 3 inches long;

topaz, white to pale buff (subhedra up to 6 inches long) rimmed by a film of fine-grained rose muscovite;

apatite, pale blue to gray, anhedral masses several inches or less across;

columbite, thin plates as large as $\frac{1}{4}$ x $\frac{1}{2}$ inch, and

bismuth carbonate (bismutite?), interstitial patches a fraction of an inch long.

Paragenetic note. Both the micas and the tourmalines appear in several species and varieties, which may be paired paragenetically as follows:

	Position	Mica	Tourmaline
Oldest	1. Wall zone (rare)	biotite	-----
	1. Wall zone	muscovite	schorl
	2. Wall zone, altered	zinnwaldite	blue tourmaline
	3a. Cleavelandite tourmaline unit	---	green tourmaline
	b. Cleavelandite lepidolite unit	lepidolite (flakes)	----
	c. Cleavelandite lepidolite unit	lepidolite (books)	----
Youngest	4. do. , altered	rose muscovite	----

Lithium concentration. In the Montgary pegmatite of Manitoba, Wright (1963) has shown that lithium has been concentrated toward the hanging wall and up the dip of the pegmatite. In the Chief pegmatite lithium mineralization is restricted to the lowest part of the pegmatite, close to its keel. The lithium is confined:

1) to the thickest section of the dike
2) to that part of the thickest section that lies at right angles to the main trend of the dike
3) to the strongly fractured part of this "tail".

Isabel Pegmatite

The Isabel pegmatite is just east of and below the volcanics on Table Mountain (SW¼ of Sec. 26, T49N, R12E). Three small open cuts have been made in the pegmatite (Plate II). The pegmatite has not been worked economically, nor is it likely ever to be. Small amounts of scrap mica and feldspar are stockpiled near the workings. It is claimed that several hundred pounds of beryl have been removed from the largest cut, but the presence of beryl in the pegmatite was not substantiated.

The pegmatite is an irregular dike that trends northeast. It

pinches or feathers out to the southwest, and rounds off abruptly to the northeast in the thickest part of the dike. The pegmatite contacts are steeply dipping and transect the foliation in the enclosing Idaho Springs gneisses.

The pegmatite is zoned and consists of the following units:

1) wall zone, quartz-microcline-biotite pegmatite;

2) intermediate zone, quartz-microcline-albite pegmatite;

3) core, quartz-microcline pegmatite;

4) replacement unit, fracture-controlled muscovite.

The replacement unit (Fig. 57) occurs in the wall zone and, less commonly, in the intermediate zone, but where it occurs it is invariably close to the contact between the wall zone and the intermediate zone (Plate II). The attitudes of the units are obscure, but the core apparently strikes northwest, across the trend of the pegmatite (Plate II).

COPPER

Many of the shear zones in the area have thin crusts of malachite along some of the fractures and contain small amounts of chalcopyrite. Most of these shear zones have been explored by small shafts or adits. None of these prospects are presently being worked.

URANIUM

4-Star Mine

The 4-Star mine was a prospectional development on a radioactive vein in Texas Creek Gulch (Plate I). The operation consists of a head frame and hoist house (Fig. 58) and a large, well-timbered shaft. Unfortunately, there is no ore to be mined. The shaft is filled with water, thus the extent of underground workings is not known to the writer.

The shaft was sunk in a small shear zone (part of the Texas Creek

Figure 57. Fracture-controlled muscovite replacement unit, Isabel pegmatite.

Figure 58. Head frame and hoist house, 4-Star mine.



Figure 57.



Figure 58.

fault zone) which contains feldspar-calcite veins. The country rocks are biotite and hornblende gneisses. In the vicinity of the shear zone the biotite and hornblende are altered to chlorite and the plagioclase is strongly sericitized and colored red from minute dusty hematite inclusions.

The veins are irregularly distributed in the shear zone. They consist of plagioclase and calcite and minor amounts of magnetite, pyrite, and sodic amphibole. The plagioclase grains range from subhedral to euhedral and average about 1 cm in length (they range up to 4 cm in length). They are moderately to strongly sericitized and are brick-red in color owing to inclusions of dusty hematite, but they have clear, unaltered rims against calcite grains. Plagioclase in the wall rocks has a composition ranging from andesine to calcic oligoclase; the plagioclase in the veins is sodic oligoclase, and the clear rims against calcite are sodic albite. The chloritic wall roks are impregnated with sodic oligoclase adjacent to some of the veins.

Calcite occurs as coarse crystals interstitial to the feldspar in the veins and as cross-cutting, finer-grained veins and patches.

Sodic amphibole is abundant in some of the wall rocks and is a minor constituent of some of the veins. It appears to have replaced the original hornblende grains rather than their chlorite pseudomorphs because much of it is controlled by the orientation of the original hornblende grains. The amphibole is pleochroic from light yellowish brown (X) to dark greenish blue (Z), has an extinction angle of about 10°, and has a strong dispersion.

Radioactivity of specimens on the mine dump ranges from undetectable over background to about 1.0 MR/Hr. The radioactivity is irregu-

larly distributed in both wall rocks and veins although the veins are, if anything, less radioactive. Autoradiographs were made from several specimens in an attempt to isolate the radioactive mineral(s). However, the greatest darkening was caused by small, widely scattered, black to dark brown patches containing chlorite as the only identifiable mineral. Analysis of specimens with an X-ray spectrograph showed the radioactivity to be from a uranium source; no thorium was detected.

The origin of this deposit is not clear. The mineral composition of the veins and the presence of sodic amphibole in the wall rocks suggests a carbonatitic origin related to similar feldspar-calcite dikes and associated fenites in the McClure Mountain alkaline complex to the south (Dahlem, 1965; Heinrich and Dahlem, 1966). If the veins are, indeed, carbonatites they are at least 3 miles north of the recognized northern boundary of the McClure Mountain complex. Also, thorium rather than uranium would be more in keeping with the known geochemistry of carbonatites.

Some of the veins are brecciated and the chlorite appears to be later than the sodic amphibole, this is further evidence of multiple activation of the Texas Creek fault zone.

South 4-Star Prospect

The South 4-Star prospect is in Texas Creek Gulch about a half-mile south of the 4-Star mine. The workings consist of a small pit, a tunnel, and a shaft (Fig. 59). The tunnel was driven into a hillside to intersect a feldspathic vein that crops out higher up on the hillside (Fig. 59). The vein strikes N20°W and dips 75°NE. The tunnel intersects the vein and a shaft was sunk along the vein at the intersection. The shaft is about four feet square and is water-filled.

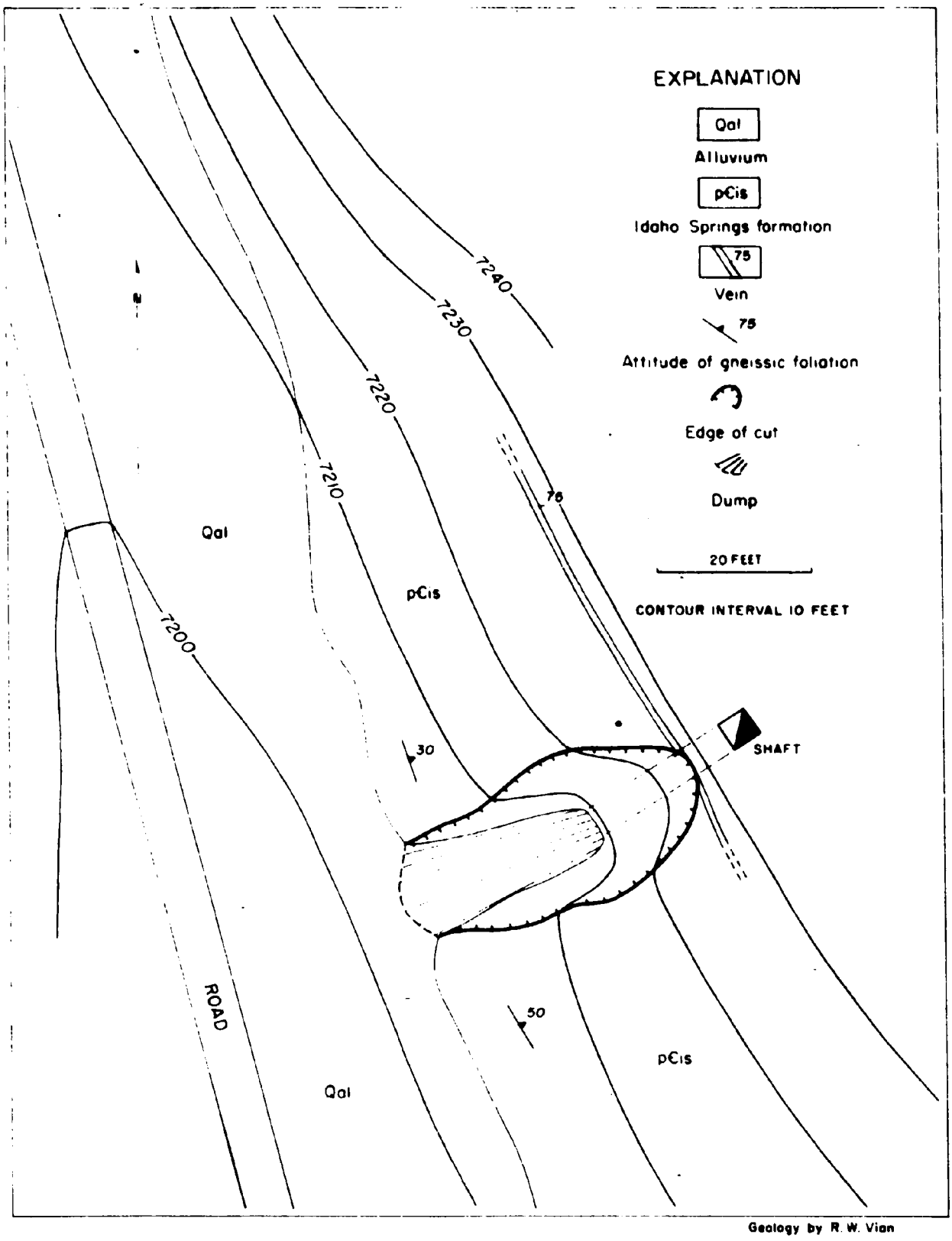


Figure 59. Geologic map of the South 4-Star prospect, Fremont County, Colorado

The feldspathic vein is in one of the many small shear zones that make up the broad Texas Creek fault zone. The vein consists predominantly of sodic oligoclase but includes minor amounts of calcite and remnant patches of highly altered (chloritized) wall rocks. Identifiable wall rocks include biotite gneiss, amphibolite, and pegmatite. The plagioclase is brick-red in color from abundant dusty hematite inclusions. It has clear rims against the few calcite grains present, indicating an origin similar to that of the feldspar-calcite veins in the 4-Star mine.

Some of the rocks in the dump are moderately radioactive and it is this feature which probably stimulated the development of the prospect.

CONCLUSIONS

The study of the geology and petrology of the Devils Hole area has led to the following conclusions:

1) The area is underlain by schists and gneisses of the Idaho Springs formation which represent at least 11,000 feet of sediments of miogeosynclinal character that were regionally metamorphosed under almandine-amphibolite facies intensities.

2) Most of the sillimanitic rocks have been formed by local aluminum metasomatism superimposed on biotite gneisses by solutions derived from the Pikes Peak magma.

3) The origin of the cordieritic rocks is open to question; some were formed by thermal metamorphism, but most are apparently the products of regional metamorphism.

4) The Idaho Springs rocks were intruded by granitic rocks in the sequence: granite gneiss, Cripple Creek (?) granite, Pikes Peak granite, Silver Plume (?) granite.

5) Migmatites were produced by both the Pikes Peak and Cripple Creek (?) granites.

6) Metasomatic effects are assignable chiefly to the Pikes Peak granite, indicating, along with similarly assignable metasomatic effects and related pegmatite formation in other areas, that the Pikes Peak magma was very rich in volatiles. Pikes Peak pegmatites are commonly large, zoned, and may have concentrations of rarer minerals. In contrast few, if any, metasomatic effects can be ascribed to the Cripple Creek (?) magma. Cripple Creek (?) pegmatites are small, poorly zoned, and mineralogically simple.

7) If the correlation of the sediments in the Echo Canyon graben

with the Farisita formation is correct, the Farisita environment extends farther north than was previously known.

8) The volcanics are part of Thirty-nine Mile volcanic field, their source was to the north. If the basaltic agglomerate is indeed a lahar, a source closer than the Guffey caldera may be indicated.

9) The Echo Canyon graben-bounding faults were active in Laramide time but may have originated in the Precambrian, as did most of the other major faults in the area. The Echo Canyon faults are, however, pre-andesite in age and thus may have been last active in Miocene time.

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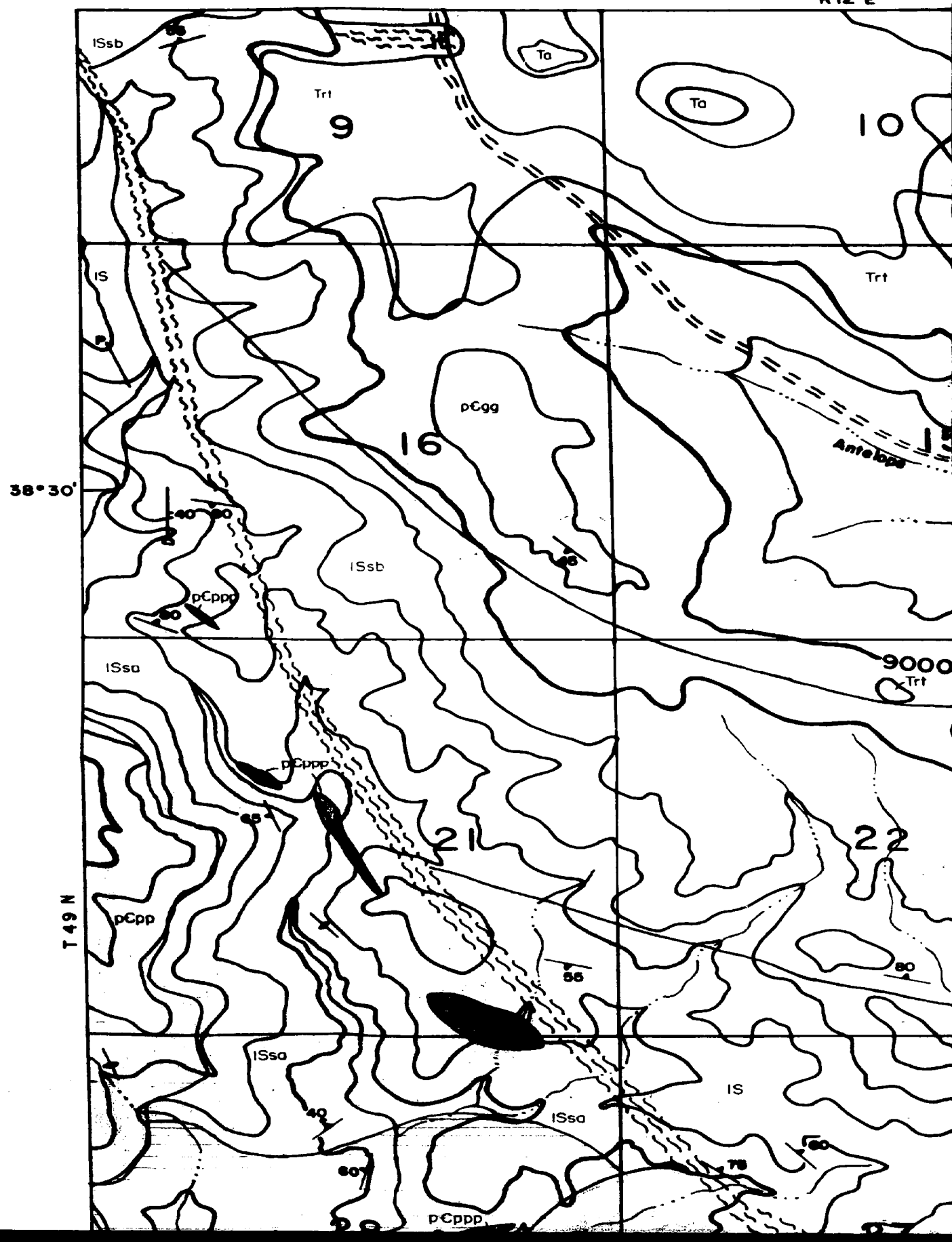
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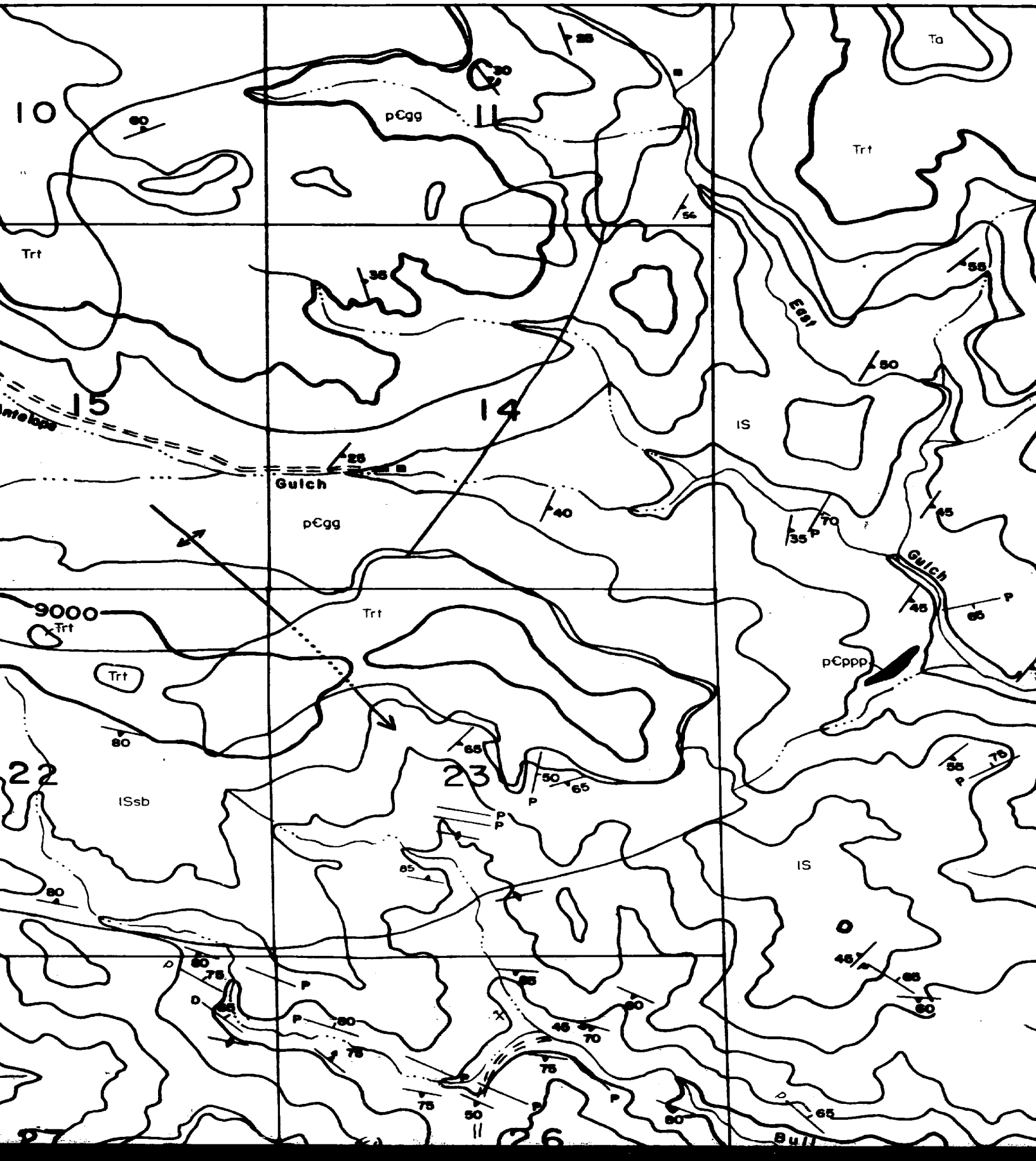
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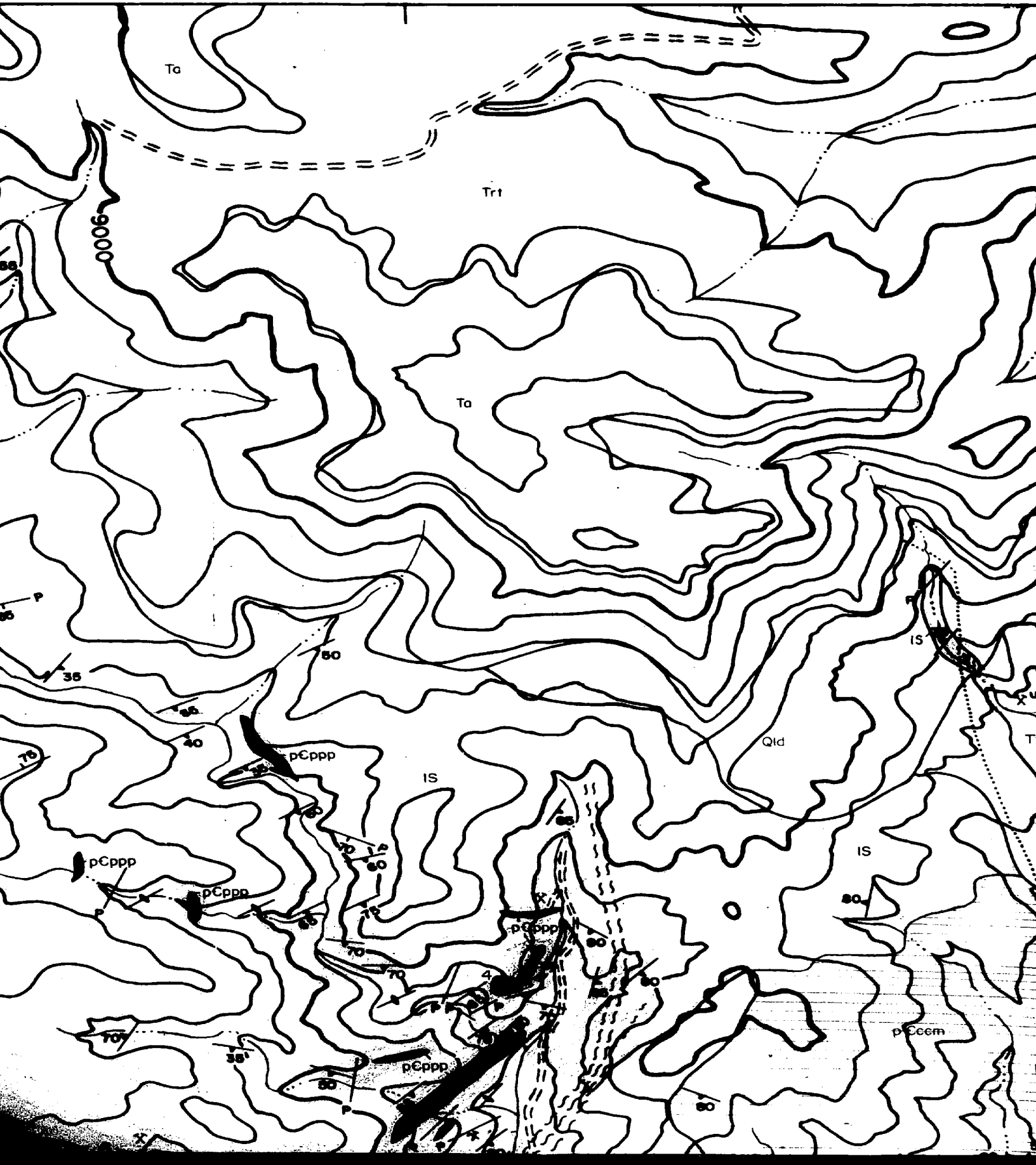
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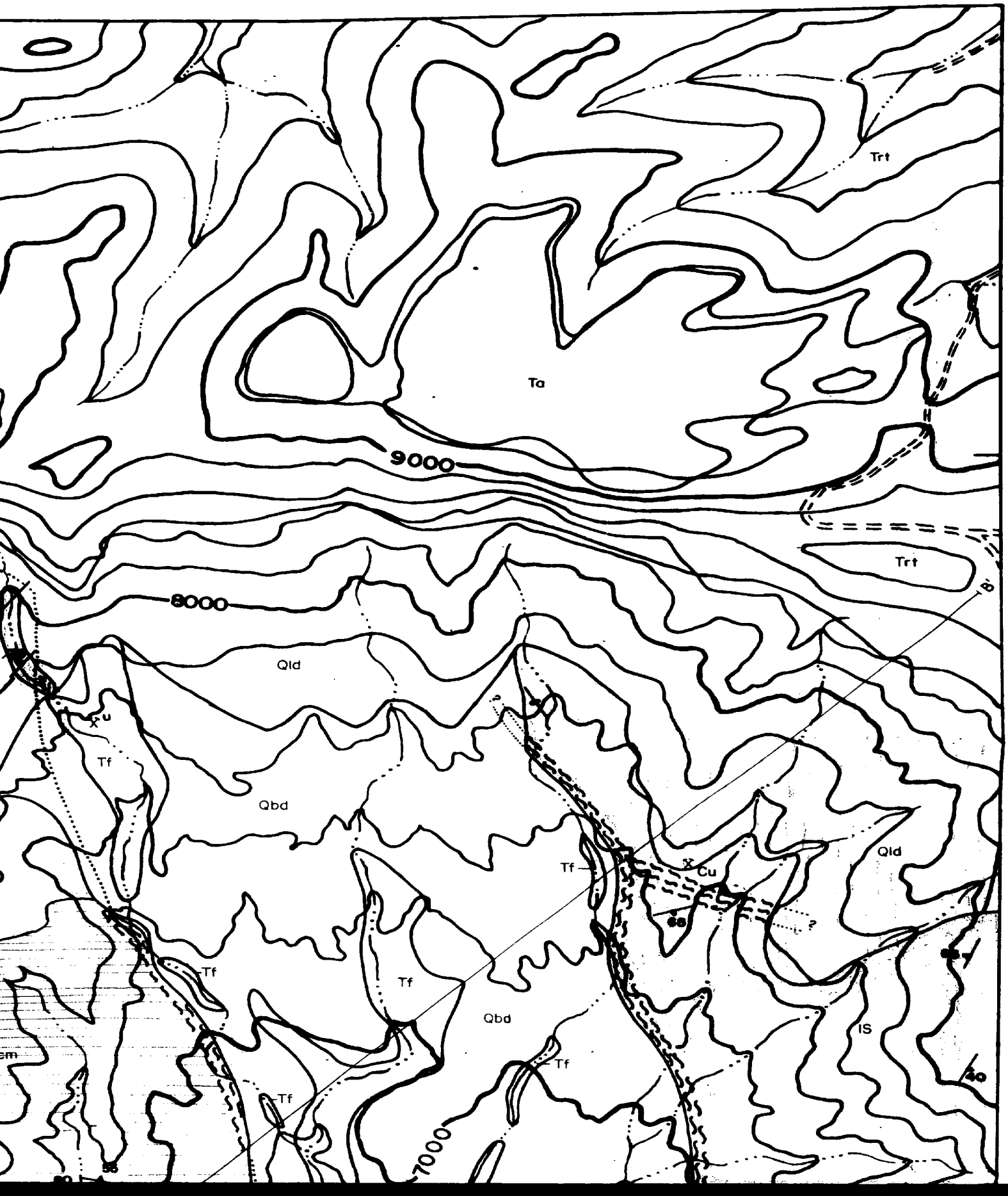
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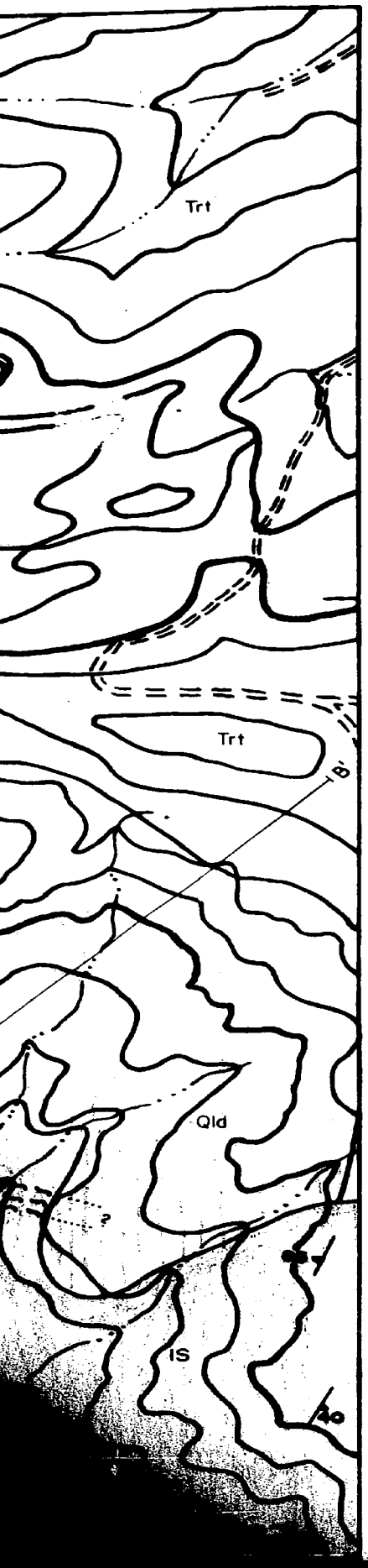
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EXPLANATION

QUATERNARY

Qal
Alluvium

Qt
Travertine

Qld
Landslide debris

Qbd
Basin debris

Qtg
Terrace gravel

TERTIARY

Ta
Andesite flows, flow breccia, and scoria

Tad
Andesite dike

Trt
Vitric rhyolite tuff, unwelded to strongly welded

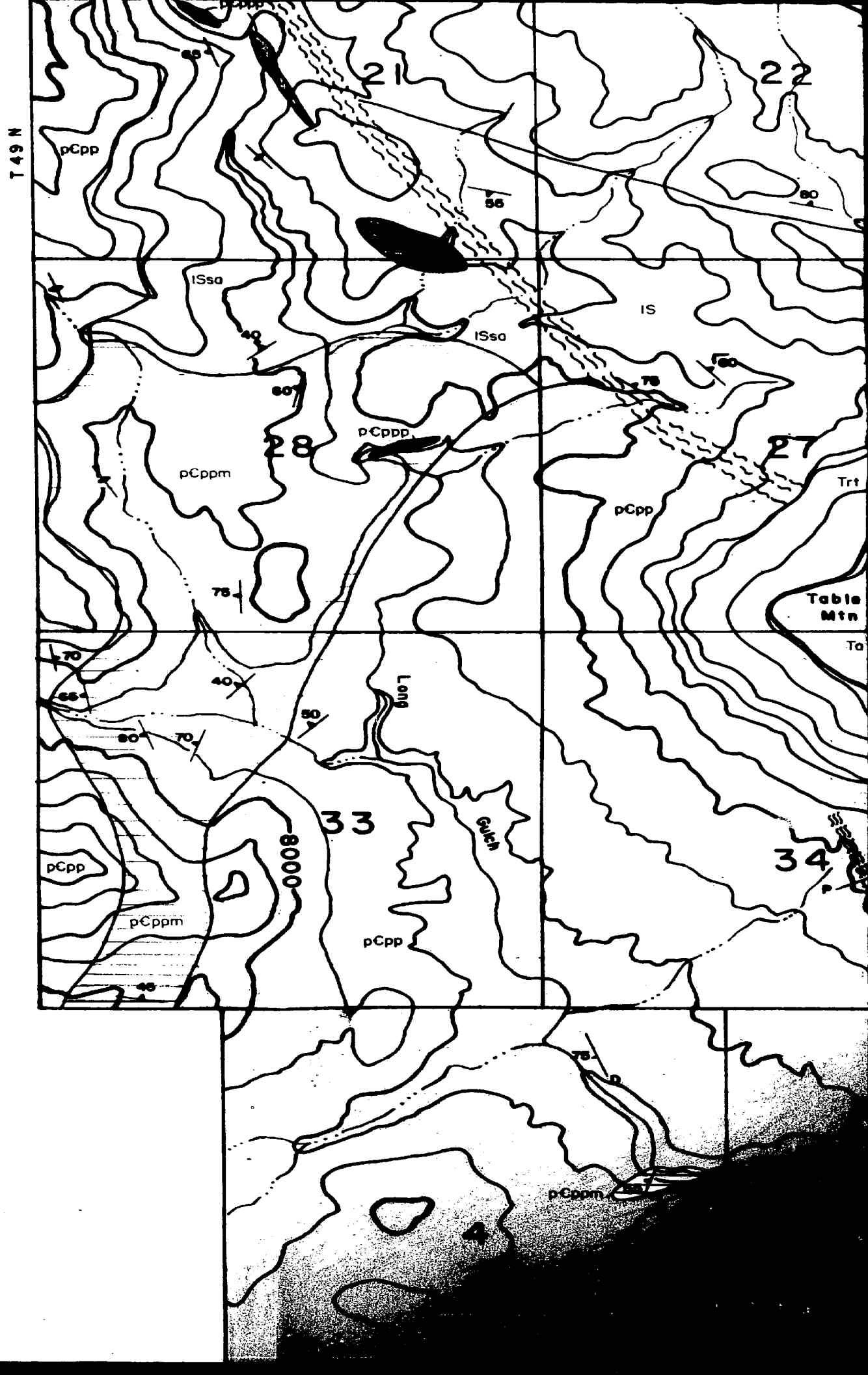
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Farisita (?) formation

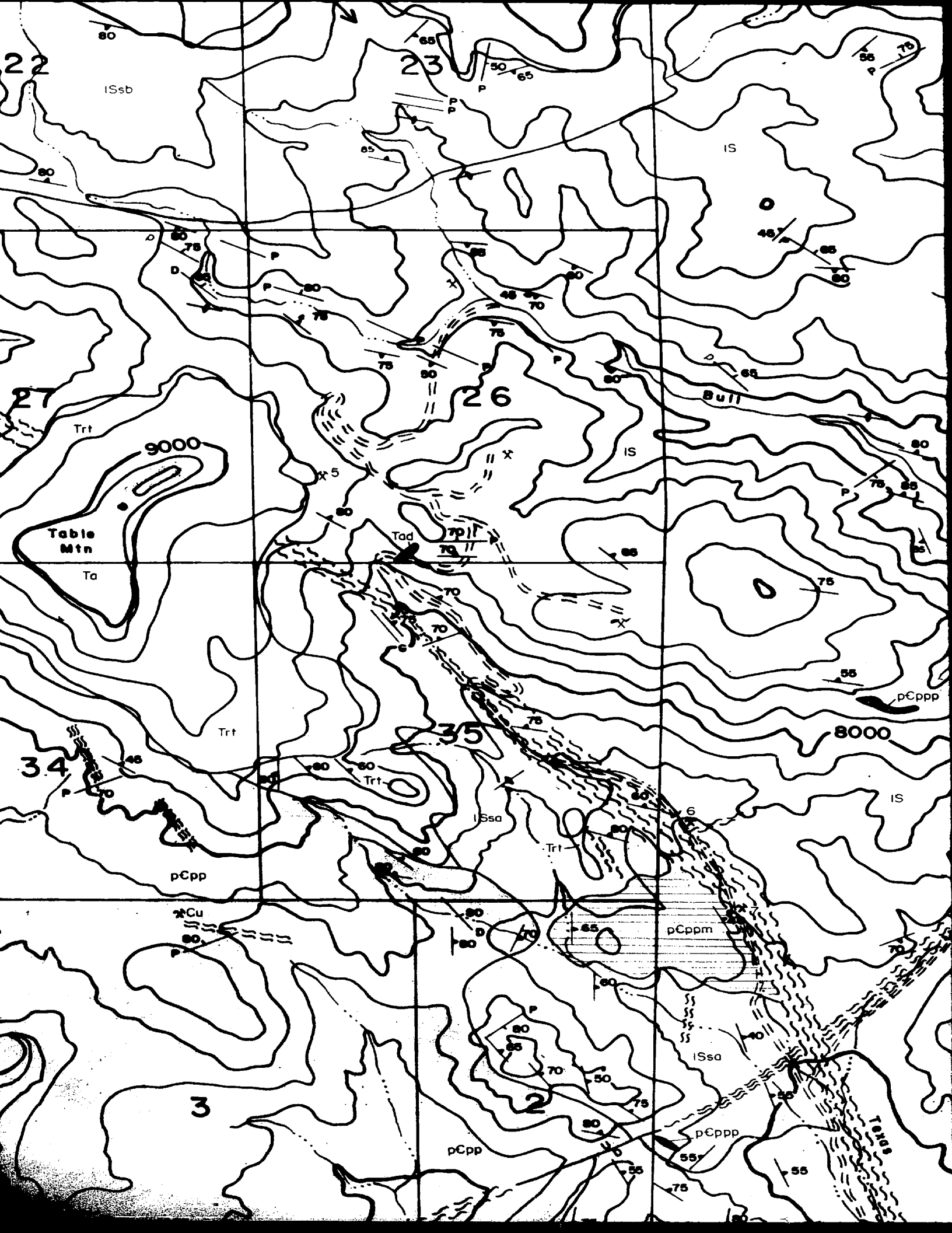
pCd
Diabase dike

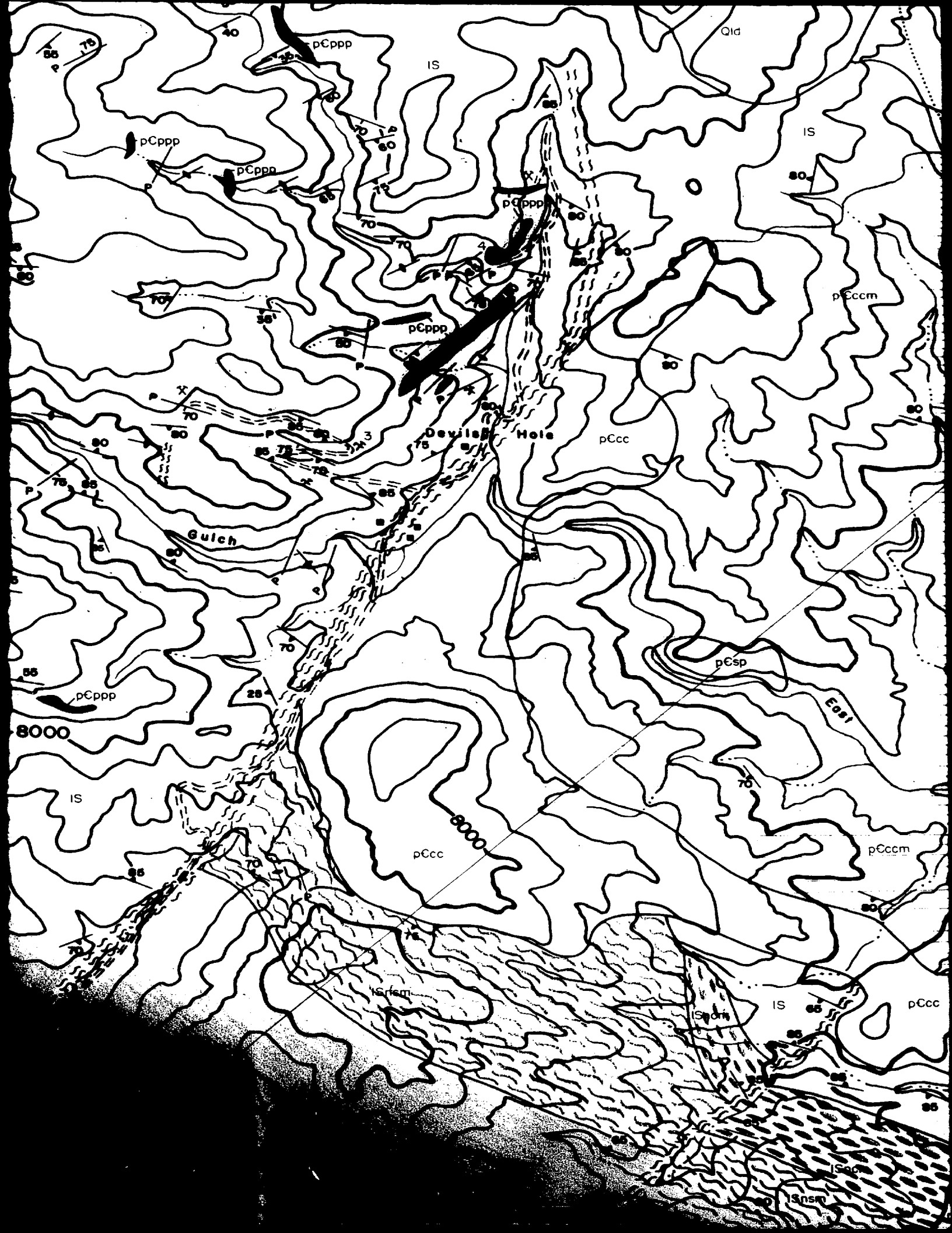
A
Silver Plume (?) granite sills and dikes

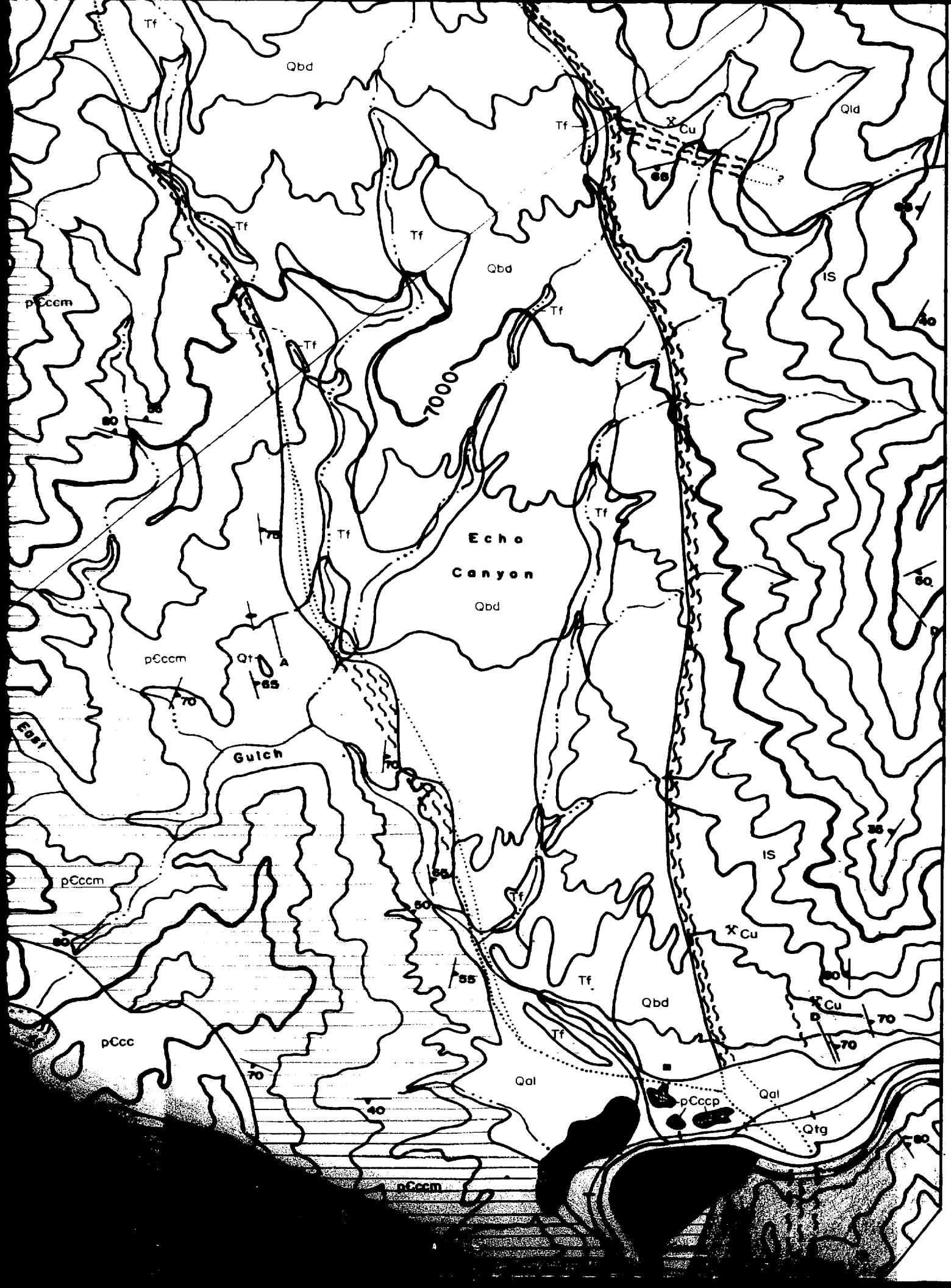
pCsp
Silver Plume (?) granite

G


T 49 N
21
22
28
27
33
34
pCpp
pCppm
pCppp
ISsa
IS
Trt
Table Mtn
Long
Gulch
8000
80
70
60
50
40
30
20
10


22
23
27
26
35
34
3
2
ISsb
IS
Trt
9000
Table Mtn
Ta
Tad
Bull
pCppp
8000
ISsa
pCpp
pCppm
Cu
Texas


Qld
IS
pCppp
pCppp
pCppp
pCppp
pCppp
pCccm
pCccm
pCcc
pCcc
pCsp
Devils Hole
Gulch
East
8000
8000
IS
IS
IS
ISnrsm
ISnsm
ISnsm


Tf
Qbd
Tf
Cu
Qld
Tf
Tf
Qbd
IS
pCccm
Tf
7000
Tf
Tf
Echo
Canyon
Qbd
pCccm
Qt
East
Gulch
pCccm
IS
Tf
Cu
Tf
Qbd
Cu
pCcc
Qal
pCccp
Qal
Qtg
pCccm
T 18 S



Vitric rhyolite tuff, unwelded to strongly welded

Tf

Farisita (?) formation

pCd

Diabase dike

Silver Plume (?) granite sills and dikes

pCsp

Silver Plume (?) granite

Granite sills and dikes undifferentiated

Cripple Creek (?) granite pegmatite

pCcc

pCccm

Cripple Creek (?) granite and associated migmatites

pCppp

Pikes Peak granite pegmatite

pCpp

pCppm

Pikes Peak granite and associated migmatites

Granite gneiss

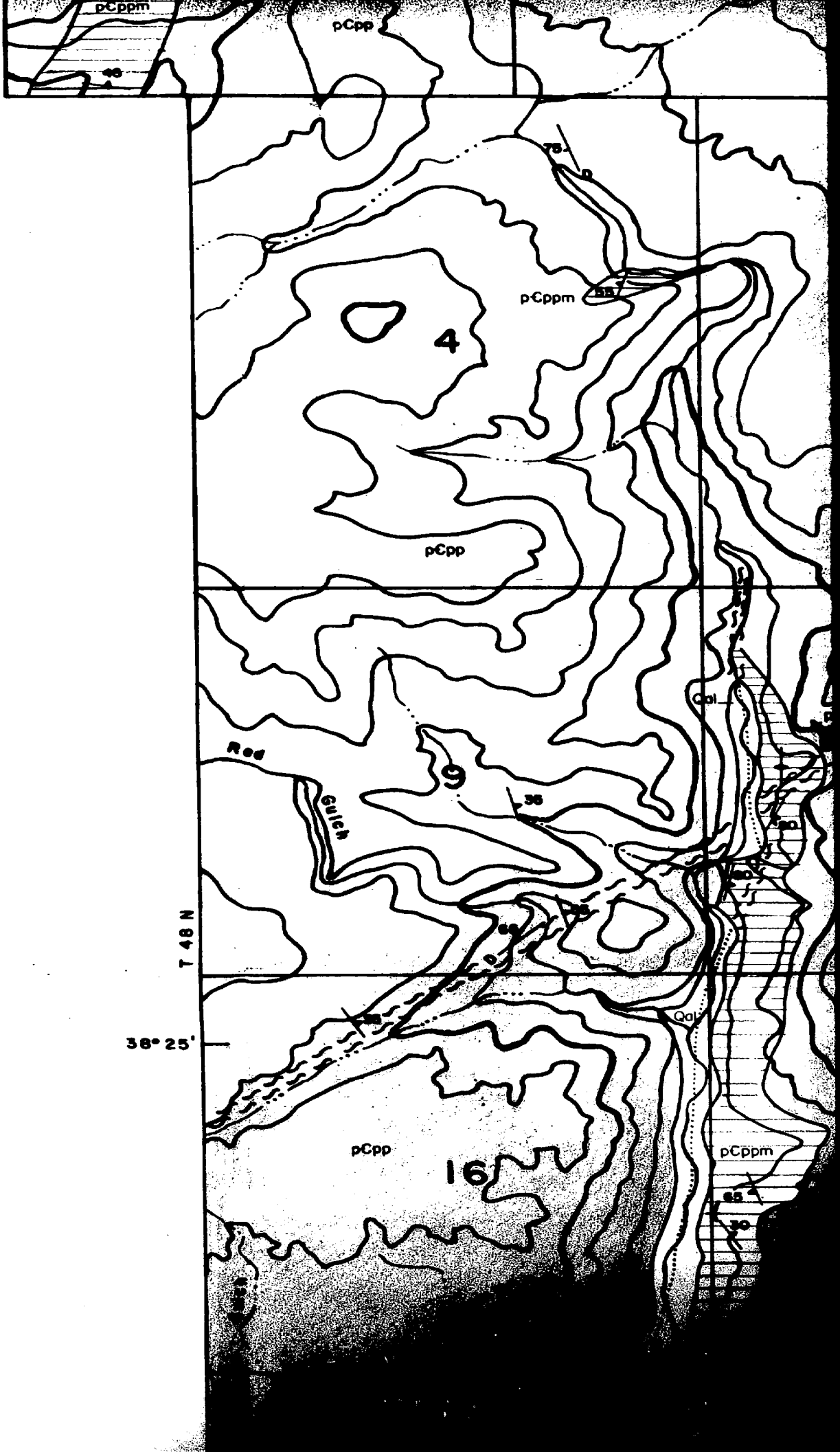
Idaho Springs nodular, cordierite-muscovite schist

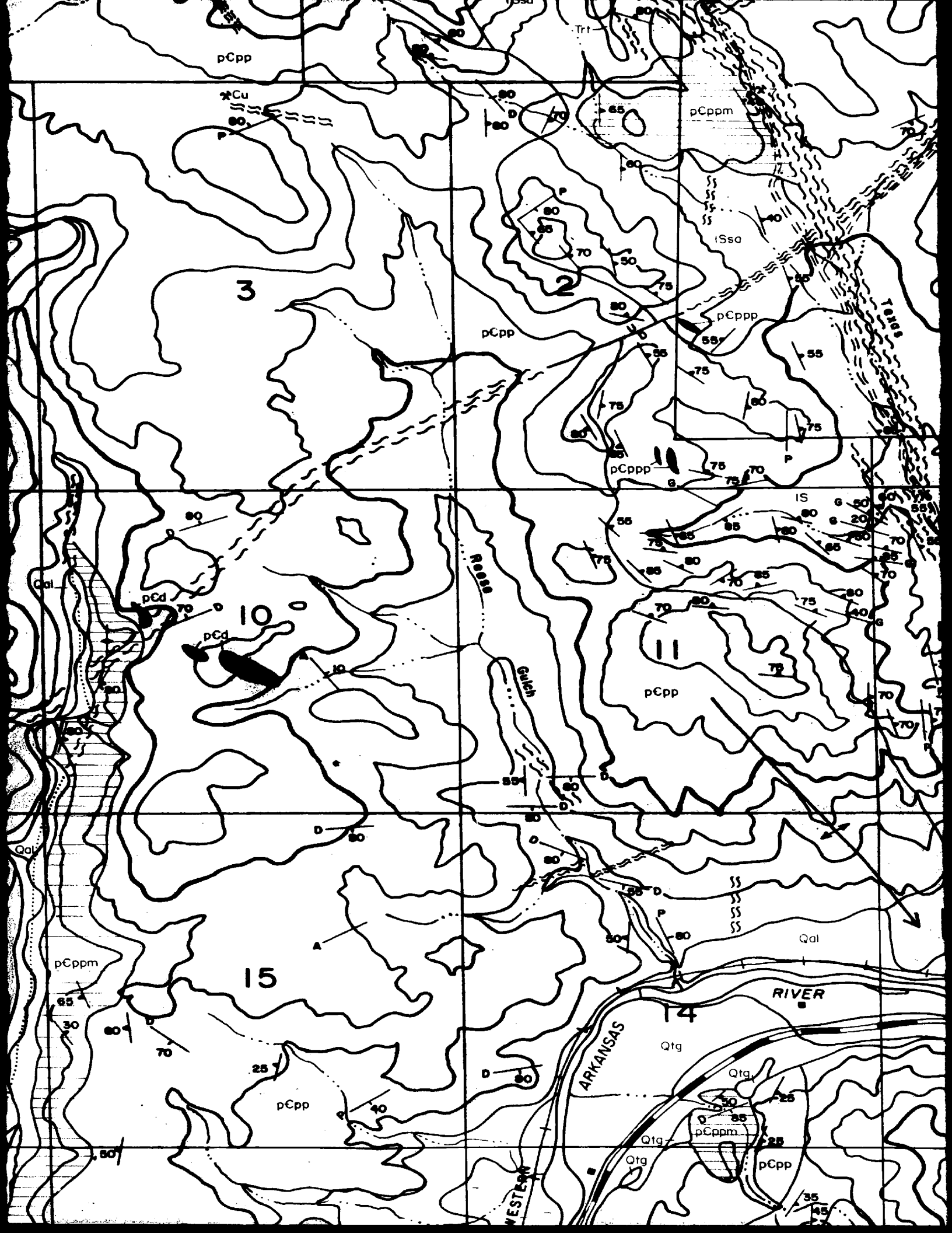
Idaho Springs nodular, sillimanite-muscovite schist

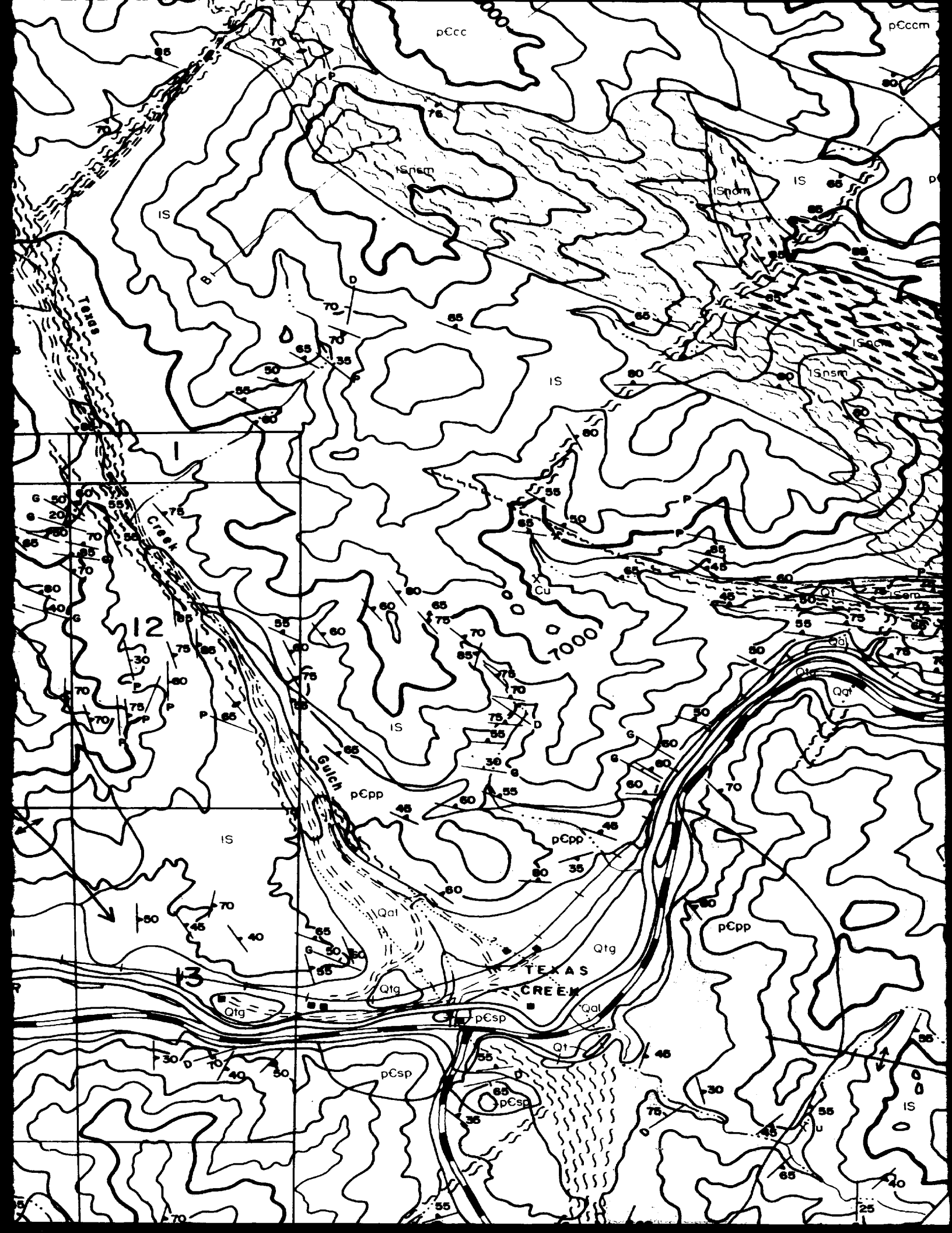
Idaho Springs undifferentiated metamorphic rocks

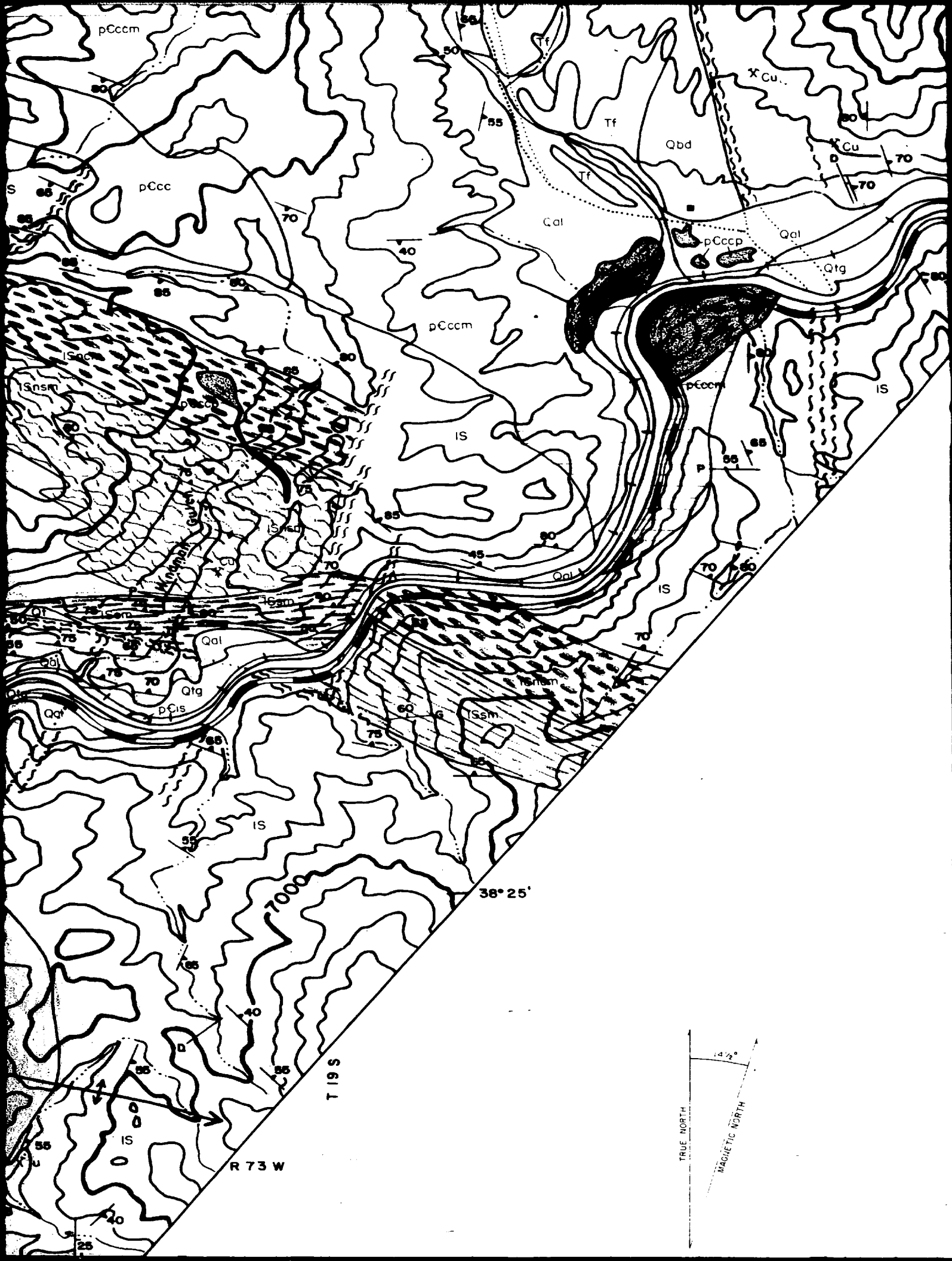
Idaho Springs metamorphic rocks

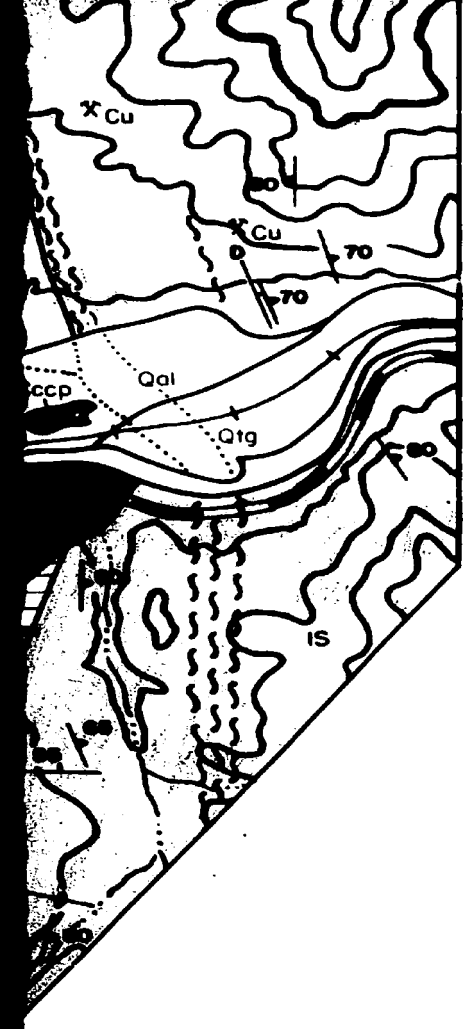
PRECAMBRIAN


pCppm
pCpp
75
pCppm
55
4
pCpp
Qal
Red
Gulch
9
35
T 48 N
38° 25'
Qal
pCpp
16
pCppm
65
30


pCpp
*Cu
Trt
pCppm
ISsa
3
2
pCpp
pCppp
Tenge
pCppp
IS
Qal
pCd
10
pCd
Reese
Gulch
11
pCpp
Qal
pCppm
15
14
RIVER
Qal
ARKANSAS
Qtg
Qtg
pCppm
Qtg
Qtg
pCpp
pCpp
WESTERN


pCcc
pCccm
ISnsm
ISnom
IS
ISsc
Texas
Creek
Gulch
pCpp
Qal
Qtg
Qt
Qaf
Cu
TEXAS
CREEK
pCsp
7000
1
12
13


pCccm
Tf
Cu
Qbd
pCcc
Cal
pCccp
Qal
Qtg
IS
Wingnook Gulch
Qal
Qtg
pCis
TSsm
38°25'
T 19 S
R 73 W
14½°
TRUE NORTH
MAGNETIC NORTH



ISccp

Idaho Springs nodular, cordierite-muscovite schist

ISsm

Idaho Springs nodular, sillimanite-muscovite schist

ISsm

Idaho Springs sillimanite-muscovite schist

ISa

Idaho Springs massive amphibole

ISsb

Idaho Springs sillimanite-biotite gneiss

IS

Idaho Springs formation undifferentiated

25

Strike and dip of foliation

Strike of vertical foliation

Fault

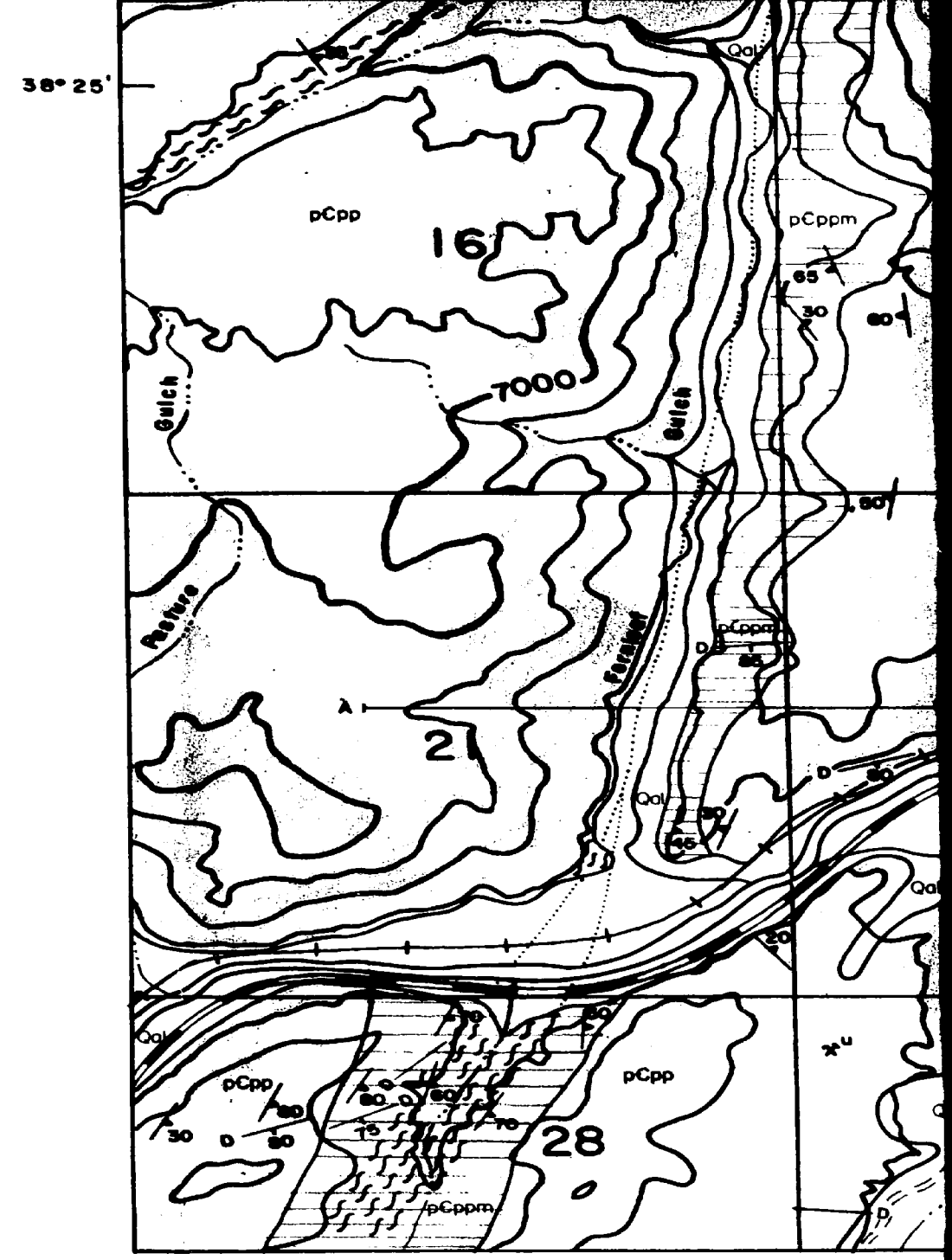
Fault zone, dotted where concealed

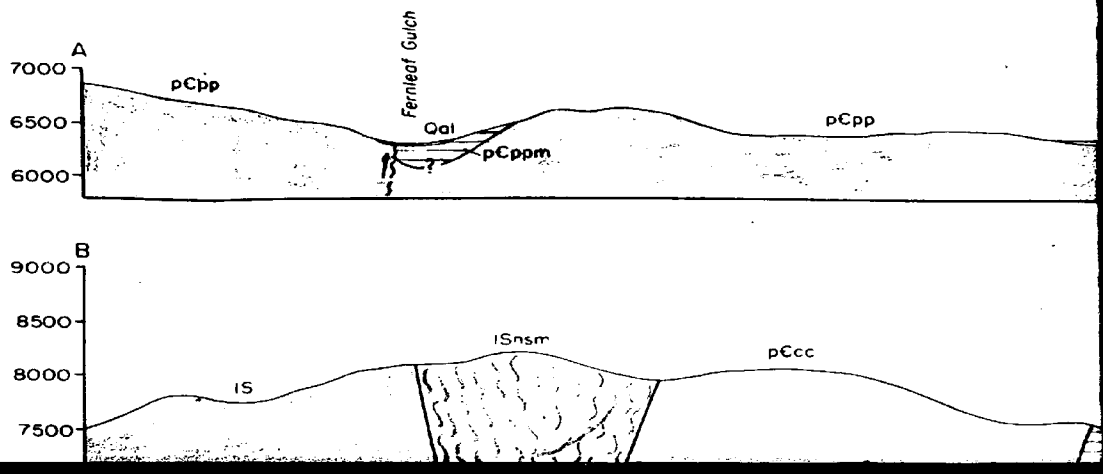
Strike and plunge of anticlinal axis

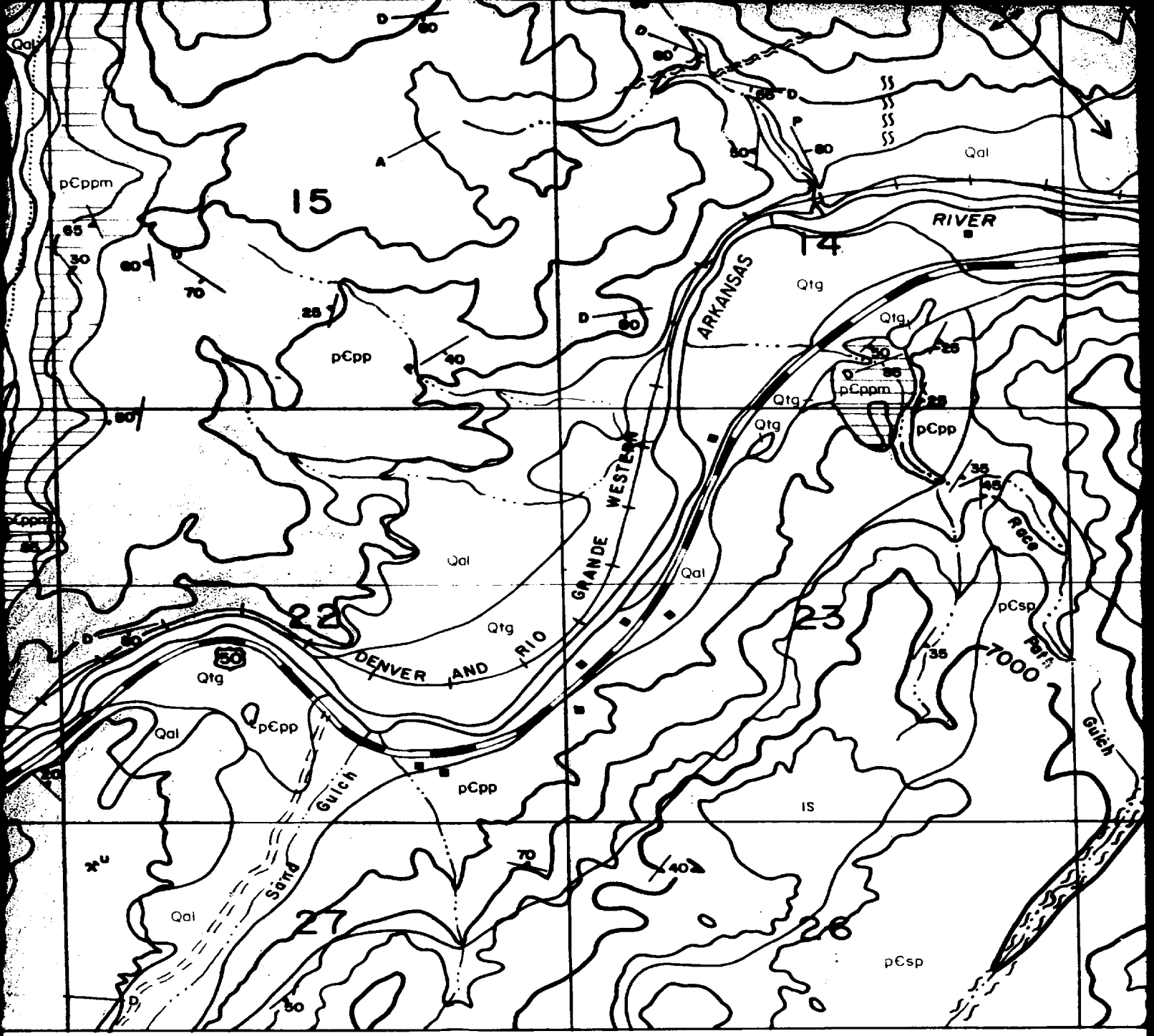
Mine, prospect or quarry
Cu - copper
u - uranium

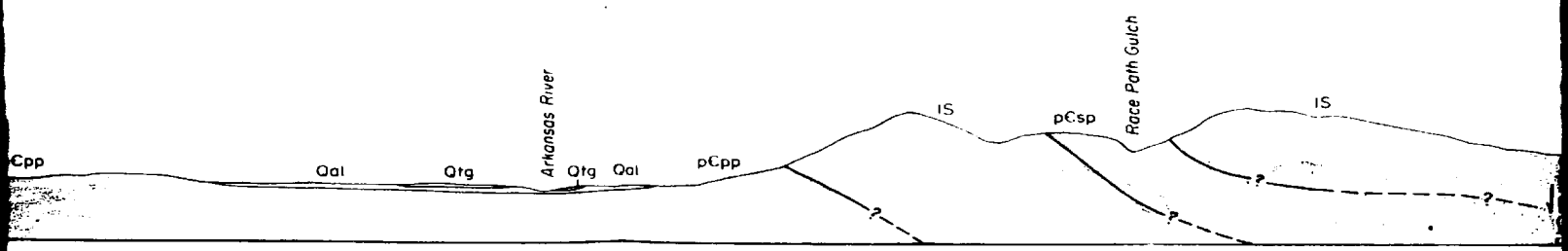
1. Silver Plume (?) granite quarry
2. Railroad quarry
3. Devils Hole pegmatite
4. Chief pegmatite
5. Isabel pegmatite

MAGNETIC NORTH

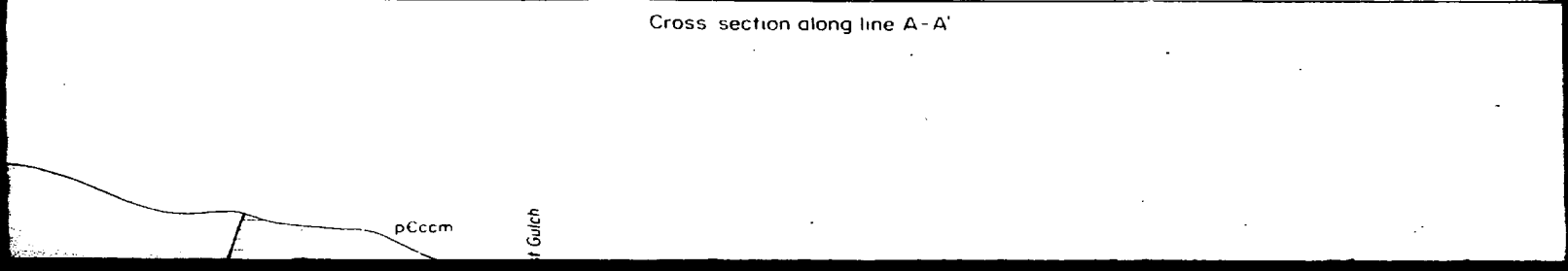

38° 25'
Qal
pCpp
16
pCppm
7000
Gulch
Gulch
Fernleaf
pCppm
A
2
Qal
pCpp
pCpp
28
pCppm

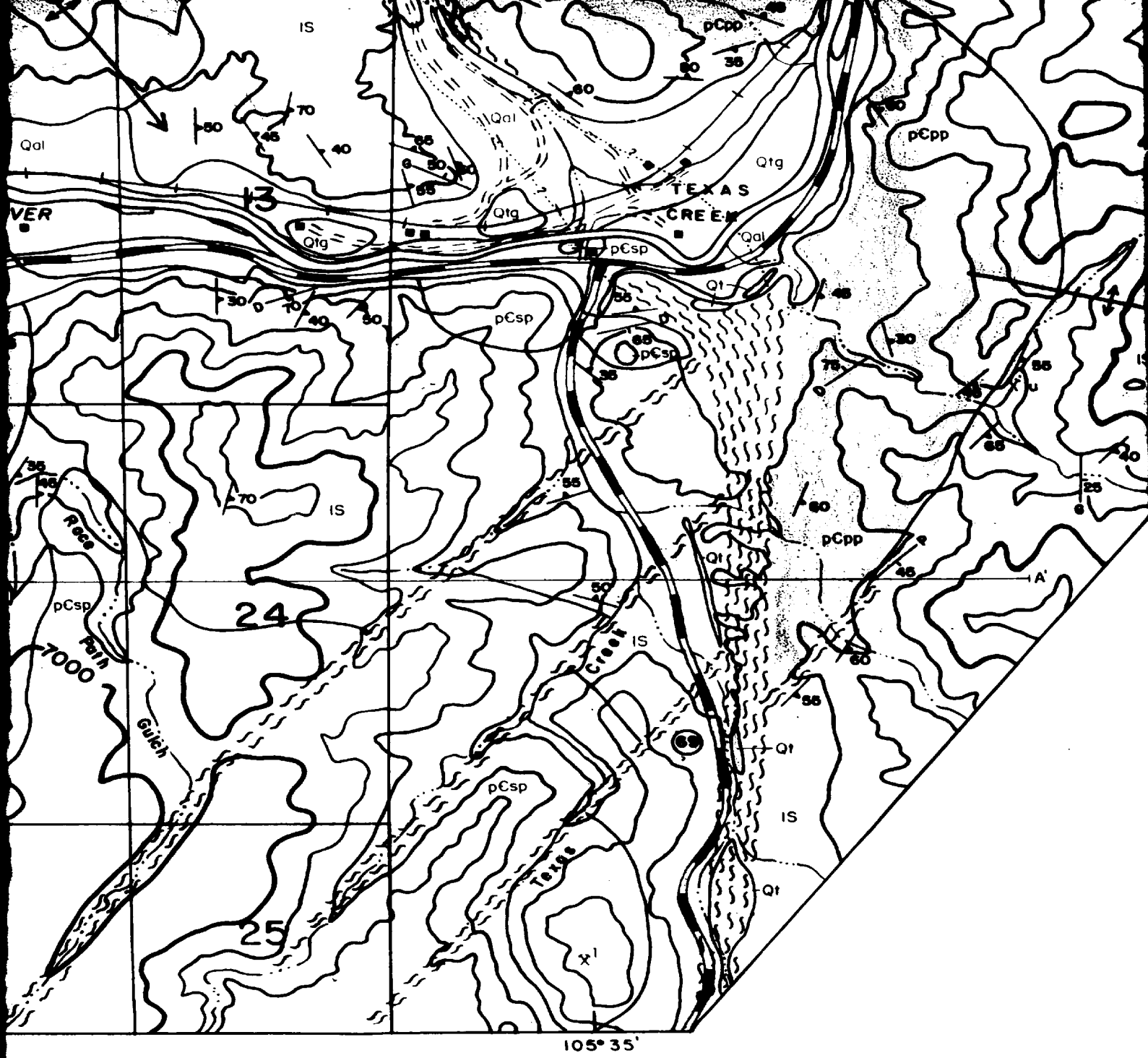

A
7000
6500
6000
pCpp
Fernleaf Gulch
Qal
pCppm
pCpp
B
9000
8500
8000
7500
IS
ISnsm
pCcc

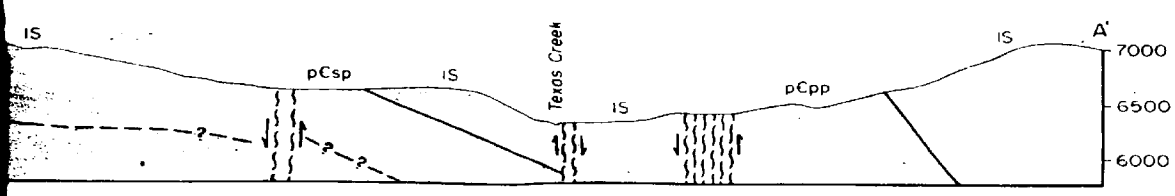




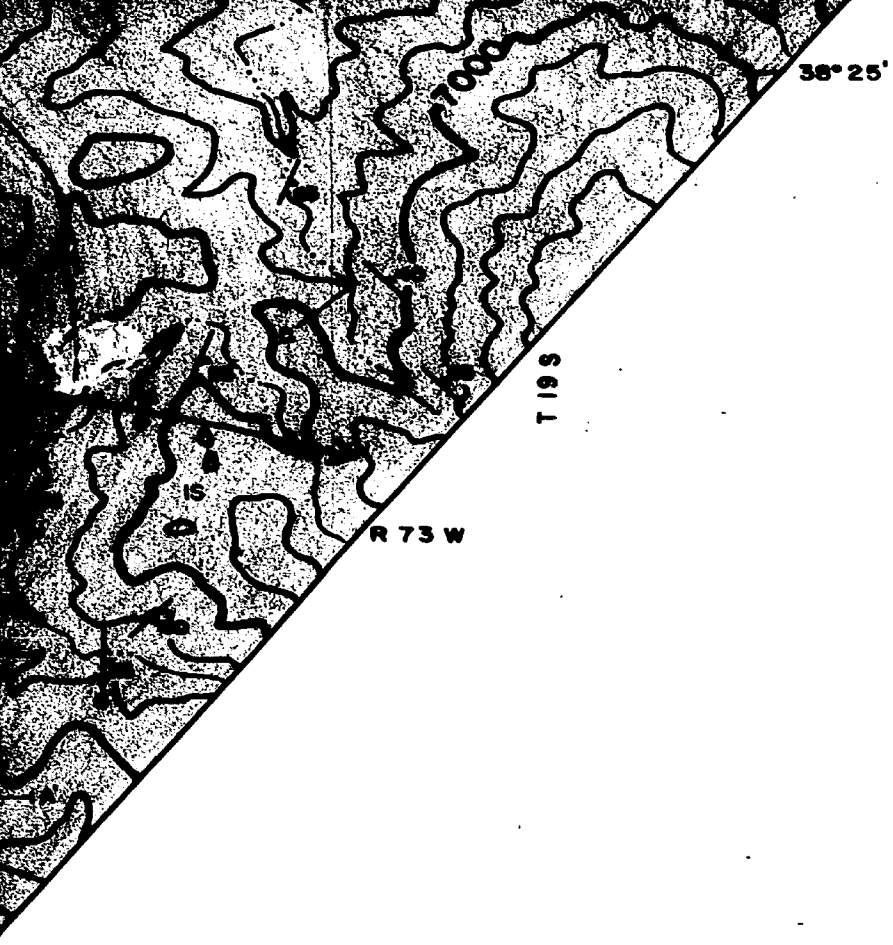
Cross section along line A-A'

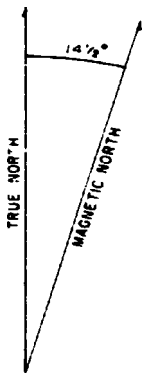



IS
Qal
Qtg
pCpp
pCsp
Qt
TEXAS
CREEK
Rock
Path
Gulch
Creek
Texas
24
25
13
7000
105°35'
A'

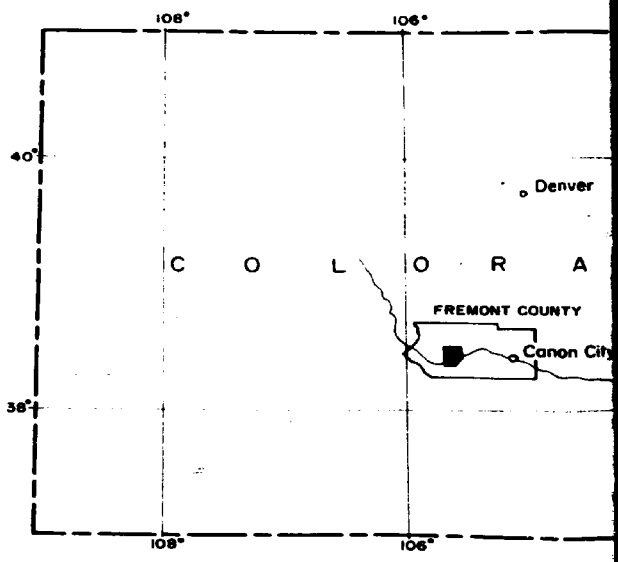

IS
pCsp
IS
Texas Creek
IS
pCpp
IS
A'
7000
6500
6000


Trt
Qld
IS
Qbd
Tf
Qba





APPROXIMATE MEAN DECLINATION



MAP OF COLORADO SHOWING LOCAT

Strike and plunge of anticlinal axis

Mine, prospect or quarry
Cu - copper
u - uranium

1. Silver Plume (?) granite quarry

2. Railroad quarry

3. Devils Hole pegmatite

4. Chief pegmatite

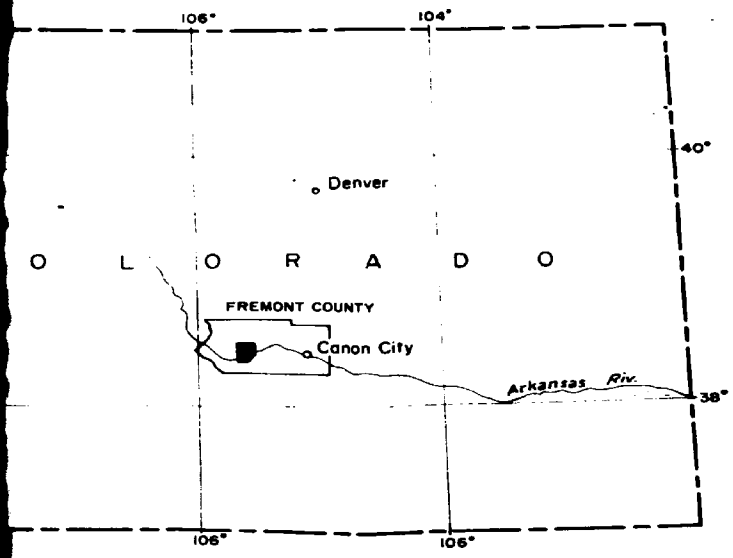
5. Isabel pegmatite

6. 4-Star mine

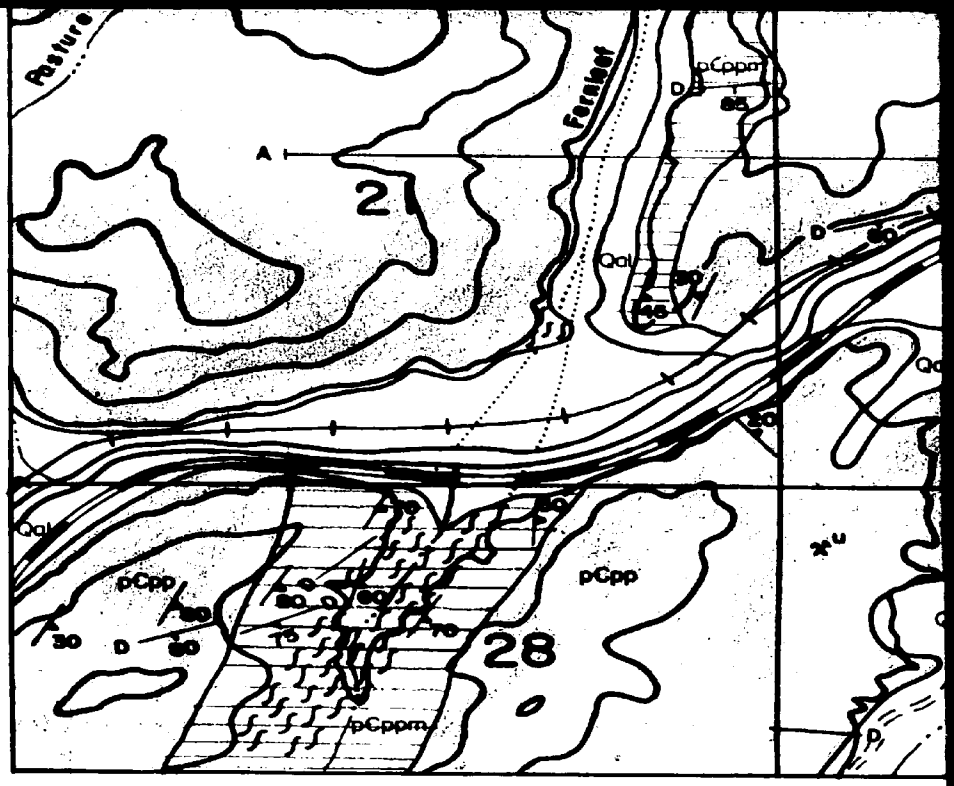
7. South 4-Star prospect

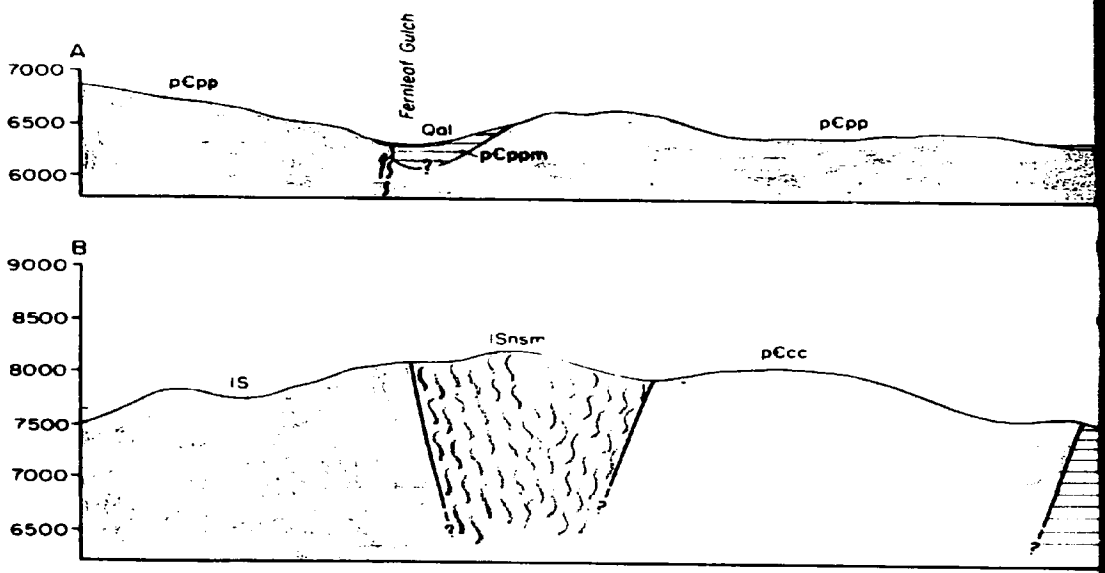
MAGNETIC NORTH

ECLINATION

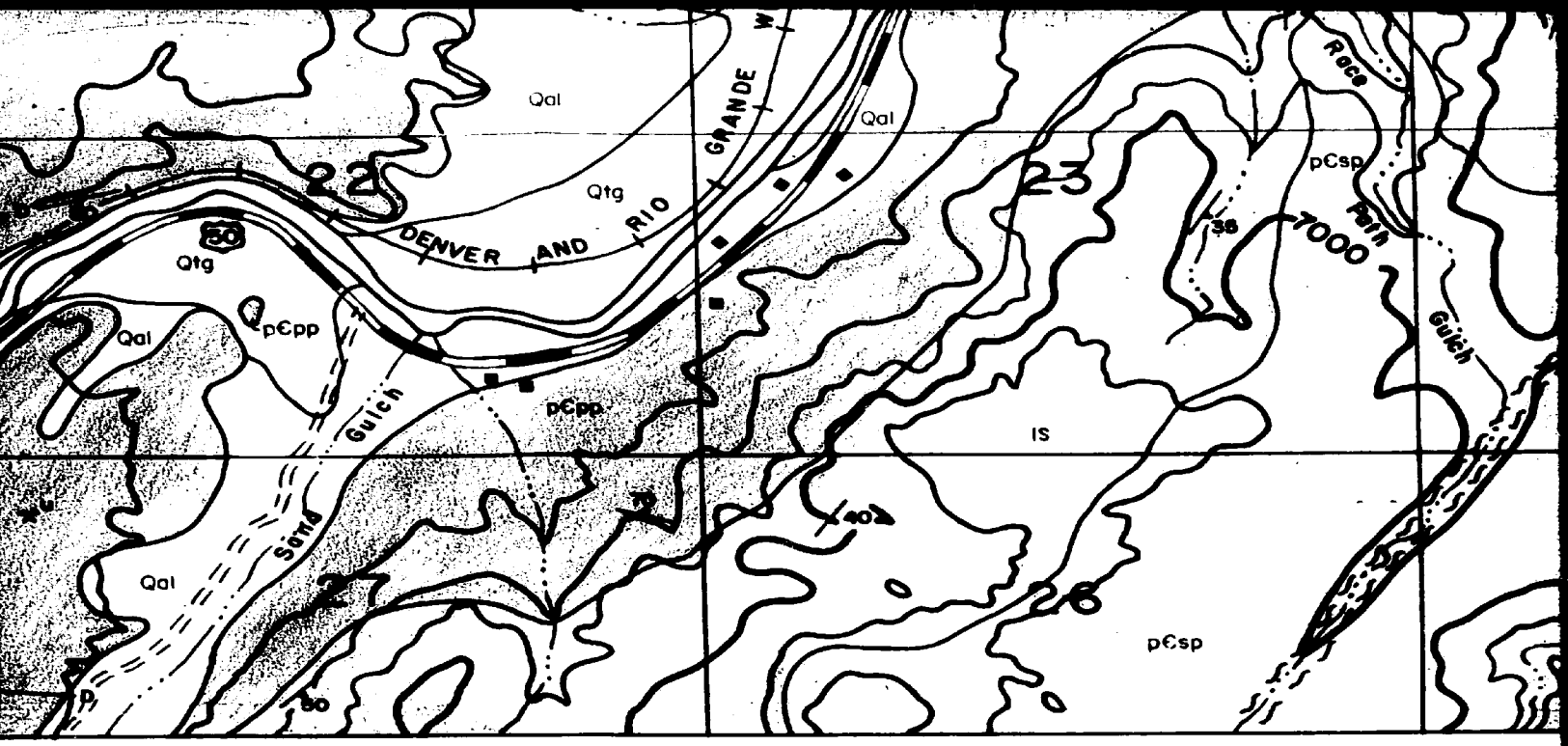


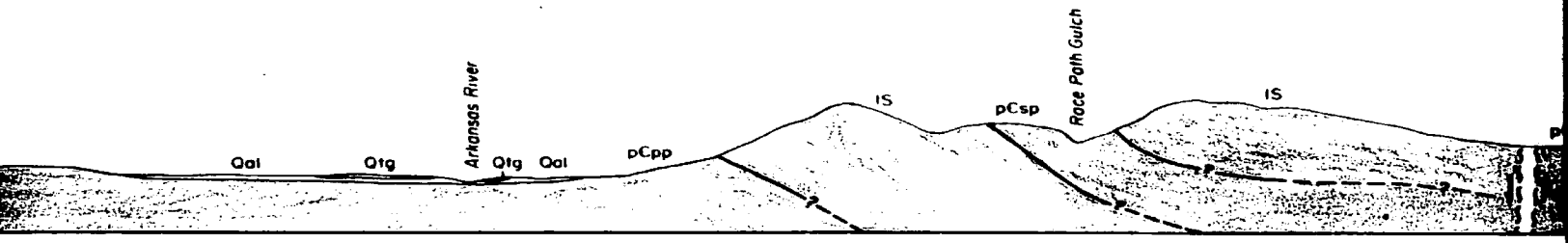
OF COLORADO SHOWING LOCATION OF AREA



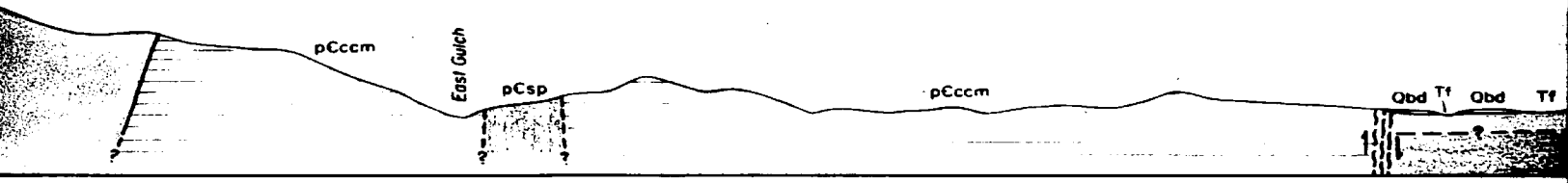


Base map from U.S.G.S. Black Mountain and Cotopaxi quadrangles



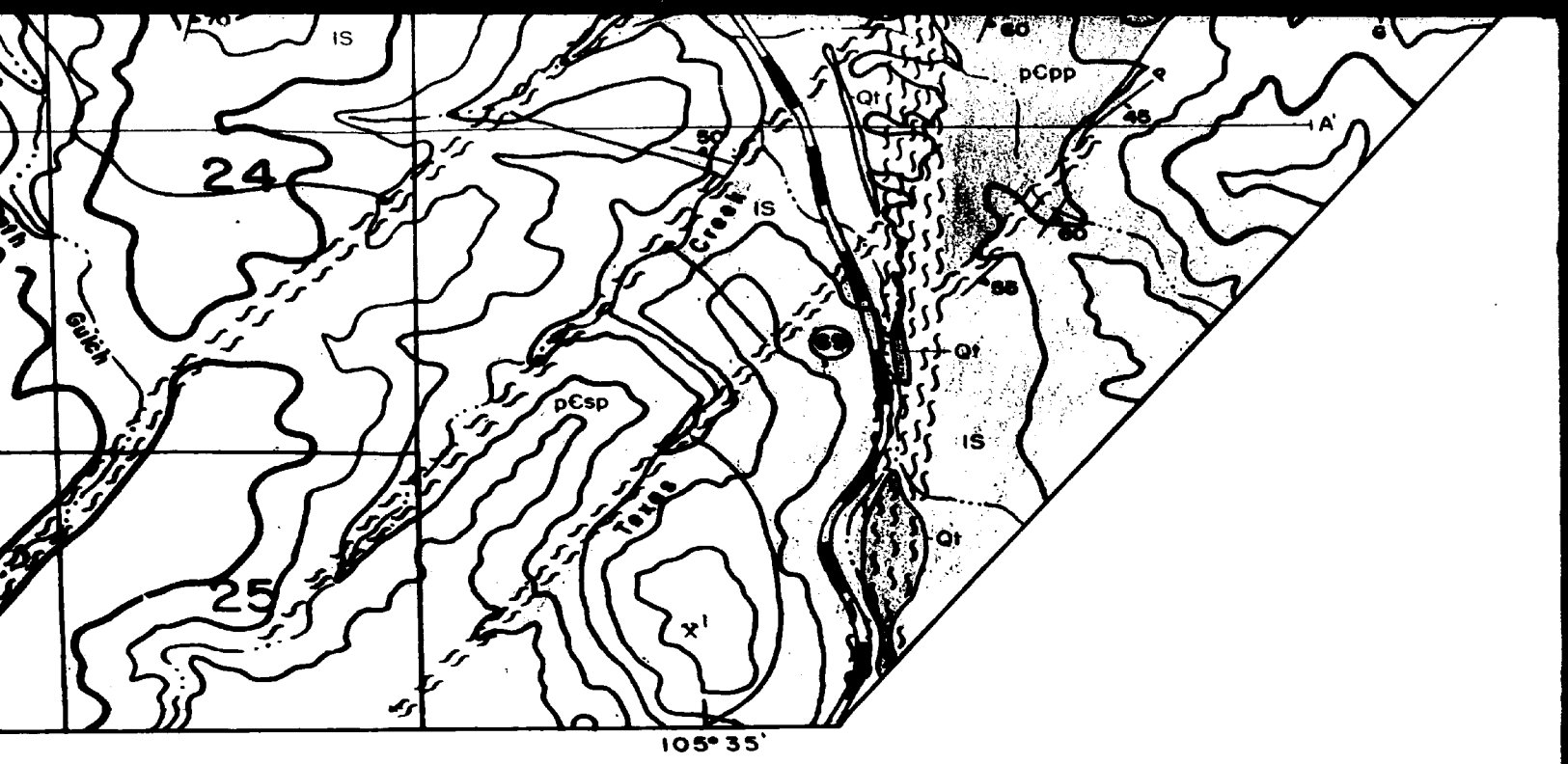


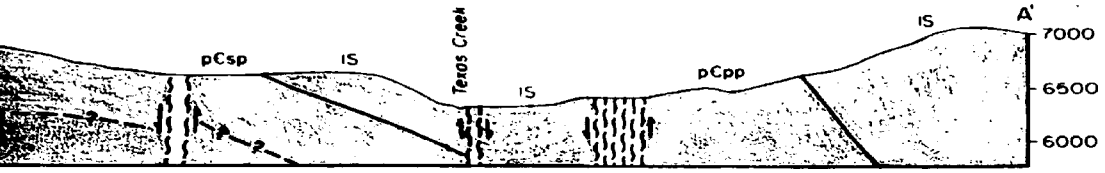
Cross section along line A - A'

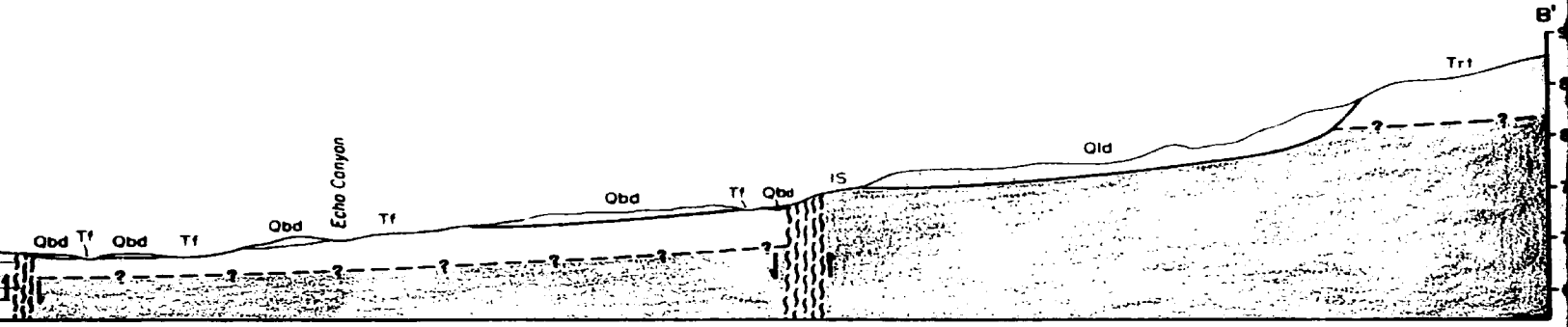


Cross section along line B - B'

GEOLOGIC MAP AND SECT







AND SECTIONS OF THE DEVILS HOLE AREA, FREMONT COUNTY, COLORADO

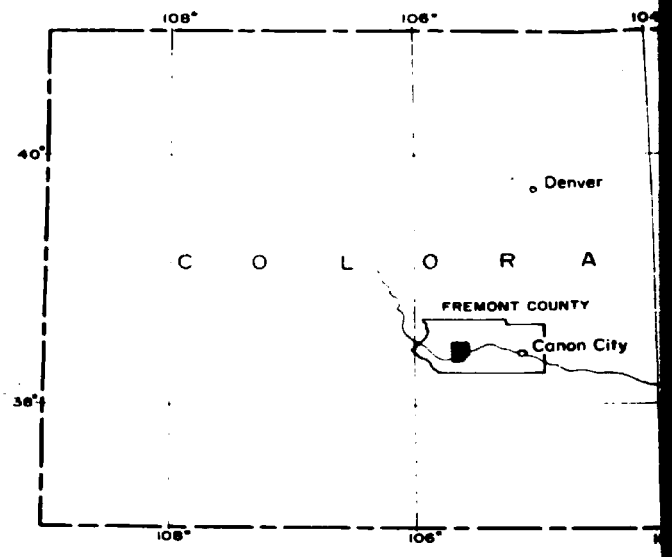
SCALE

0 1

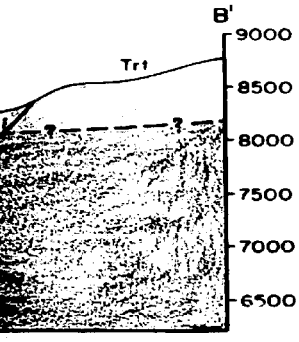
Miles

CONTOUR INTERVAL 200 FEET

APPROXIMATE MEAN DECLINATION

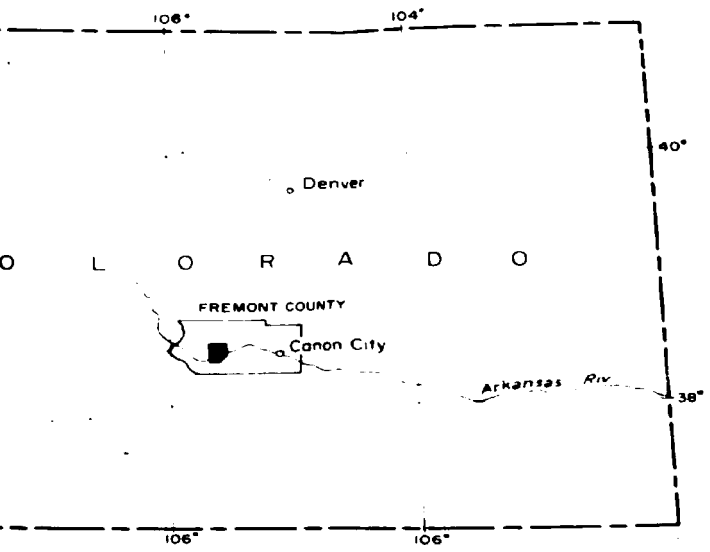


MAP OF COLORADO SHOWING LOCATION



, COLORADO

7 South 4 - Star prospect



OF COLORADO SHOWING LOCATION OF AREA

Geology by Richard W. Vian, 1961-62