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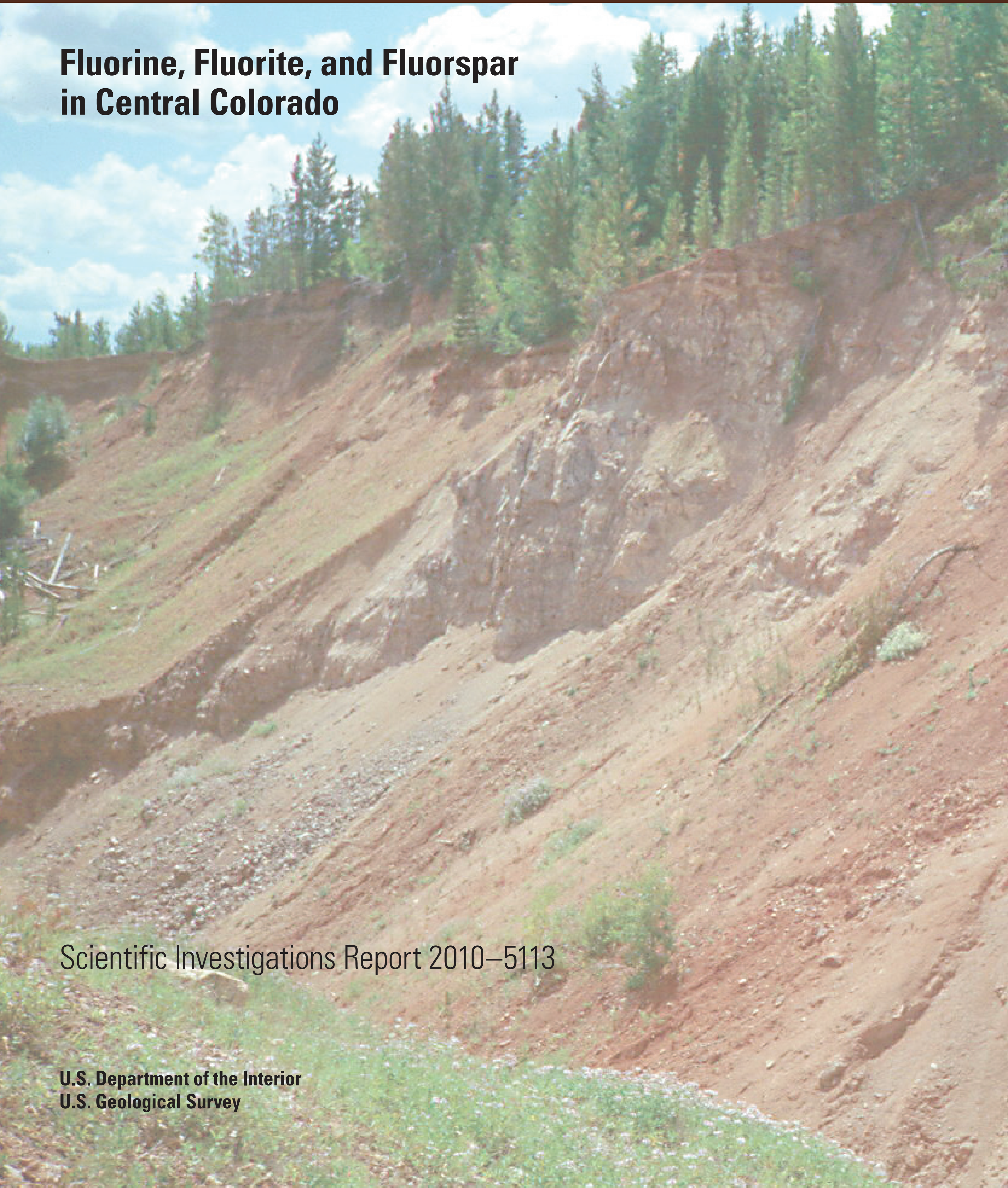

Fluorine, Fluorite, and Fluorspar in Central Colorado

Scientific Investigations Report 2010–5113

U.S. Department of the Interior
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By Alan R. Wallace

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Abstract

Fluorine (F) is a widespread element that was deposited in a variety of rocks, minerals, and geologic environments in central Colorado. It occurs as a trace element, as a major component of the mineral fluorite (CaF_s), and as a major economic source of fluorine in fluorspar deposits, which are massive concentrations of fluorite. This study has compiled available geochemical analyses of rocks, both unmineralized and mineralized, to determine the distribution of fluorine in specific age-lithologic categories, ranging from 1.8-giga-annum (Ga) metamorphic rocks to modern soils, throughout central Colorado. It also draws upon field studies of fluorine-rich mineral deposits, including fluorspar deposits, to decipher the nearly two-billion-year-long geologic history of fluorine in the study area, with implications for mineral-resource evaluations and exploration. The resulting compilation provides an important inventory of the naturally occurring levels and sources of fluorine that ultimately weather, erode, and become part of surface waters that are used for domestic water supplies in densely populated areas along the Colorado Front Range.

Most commonly, fluorine is a trace element in virtually all rocks in the region. In the 3,798 unmineralized rocks that were analyzed for fluorine in the study area, the average fluorine content was 1,550 parts per million (ppm). The median was 640 ppm, nearly identical to the average crustal abundance of 650 ppm, and some high-fluorine rocks in the Pikes Peak area skewed the average to a value much greater than the median. Most unmineralized age-lithologic rock suites, including Proterozoic metamorphic rocks, 1.7- and 1.4-Ga granitic batholiths, Cambrian igneous rocks, Phanerozoic sedimentary rocks, and Laramide and Tertiary igneous rocks, had median fluorine values of 400 to 740 ppm fluorine. In all suites, however, a small number of analyzed samples contained more than 1 percent (10,000 ppm) fluorine. The 1.1-Ga plutonic rocks related to the Pikes Peak batholith had a mean fluorine content of 1,700 ppm, and primary magmatic fluorite and fluorite-bearing pegmatites are common throughout that igneous mass.

Fluorine was deposited in many types of economic mineral deposits in central Colorado, and it currently is a significant trace element in some thermal springs. In the fluorspar deposits, fluorine contents were as high as 37 percent. Some fluorine-rich porphyry systems, such as Jamestown, had fluorine values that ranged from 200 ppm to nearly 37 percent fluorine, and veins in other deposits contained hydrothermal fluorite, although it was not ubiquitous. For the 495 samples from non-fluorspar mining districts (and excluding Jamestown), however, the median fluorine content was 990 ppm. This is above the crustal average but still relatively modest compared to the fluorspar deposits, and it indicates that the majority of the mineralizing systems in central Colorado did not deposit large amounts of fluorine. Nevertheless, the fluorine- and fluorite-rich mineral deposits could be used as guides for the evaluation and discovery of related but concealed porphyry and epithermal base- and precious-metal deposits.

The Cenozoic geologic history of central Colorado included multiple periods during which fluorine-bearing rocks and mineral deposits were exposed, weathered, and eroded. This protracted history has released fluorine into soils and regoliths, and modern rainfall and snowmelt interact with these substrates to add fluorine to the hydrosphere. This study did not evaluate the fluorine contents of water or make any predictions about what areas might be major sources for dissolved fluorine. However, the abundant data that are available on fluorine in surface water and ground water can be coupled with the results of this study to provide additional insight into natural sources of fluorine in domestic drinking water.

Introduction

Fluorine (F) is a widespread element that is found in a variety of rocks, minerals, and geologic environments. On average, the Earth's crust contains only about 650 parts per million (ppm) fluorine (Krauskopf, 1979). The geologic history of Colorado spans more than 1.8 billion years, ranging from the deposition of Paleoproterozoic sediments and volcanic units to modern erosion and sedimentation. Fluorine is present in all mineralized and unmineralized lithologies in central Colorado, and fluorine concentrations range from less than a hundred parts per million (ppm, with 10,000 ppm equaling 1 percent) to nearly 40 percent. Primary fluorite (CaF_2), as compared to fluorite introduced by later mineralizing processes, is present in many igneous and metamorphic rocks, and fluorine also is a common trace element in minerals in those rocks. Fluorite occurs in fluorspar deposits and as an accessory mineral or trace element in other hydrothermal systems. Fluorite was deposited from the Paleoproterozoic to the Holocene, and the settings varied as widely as the ages.

Central Colorado was a leading producer of fluorine from a number of fluorspar deposits during parts of the 20th century. Although the fluorspar deposits currently are not economic compared to those in other countries, the deposits and other fluorine-rich geologic environments provide important information about the geologic history of the region.

This report summarizes the geology, origin, and distribution of fluorine, fluorite, and fluorspar in central Colorado. This study is part of a larger U.S. Geological Survey (USGS) project to inventory and assess the geology and mineral resources of several national forests in the heavily mineralized central part of the State. The study area includes much of the mountainous part of Colorado, extending north to south from the southern Wyoming border to the northern New Mexico border, and east to west from the plains/mountains interface to a more arbitrary border in west-central Colorado that was defined for the USGS project as a whole (fig. 1). This inventory and discussion of fluorine environments in the study area can be used as a metallogenic guide at a regional scale, and as a baseline for other studies of natural sources of fluorine in Colorado watersheds. In addition, areas with naturally high fluorine levels are important considerations in discussions about water quality.

This report provides two major databases—a compilation of fluorine geochemical data and a database on fluorspar deposits in central Colorado—that include information on the geologic setting of the fluorine-bearing rocks and deposits. Data used for this study included past publications on the geology and mineral deposits of central Colorado, reconnaissance field work and geochemical analyses done by the author for this and related studies, fluorite and fluorspar records from the USGS Mineral Resources Data Systems (MRDS) database, and fluorine analyses in the USGS rock geochemistry database; the two USGS databases are described in the "Fluorine Data from Central Colorado" and "Fluorspar Deposit Database" sections. In addition, the report provides a table and brief summaries of the locations and geologic settings of significant topaz deposits in the region, although topaz was treated simply as another fluorine-bearing mineral.

Numerous reports summarize the geology of various geographic areas and time periods. Ogden Tweto's geologic map of Colorado (Tweto, 1979a) served as basis for the regional geology, and Wallace (1990) provided a summation of the geologic history of the central Colorado mineral belt. Each time-based section in this report gives a summary of the geologic history of that period, especially as it pertains to fluorine-rich rocks and mineral deposits of that age. Earlier reports (Aurand, 1920; Vanderwilt and others, 1947; del Rio, 1960; Brady, 1975; Shawe, 1976, in part) described the geology of fluorspar districts in Colorado, and the results of those studies were used for the current project.

Although the mineral fluorite (CaF_2) was the most common fluorine-bearing substance produced from fluorspar deposits, colloform, non- to semicrystalline calcium fluoride was a significant source of fluorine during mining. In this report, the term fluorspar is used when describing the actual fluorine-rich ore, regardless of its crystalline or noncrystalline state. The term fluorite is used for the actual mineral, which is present in the fluorspar deposits, other mineral deposits, and as a primary mineral in some igneous rocks. Fluorine simply refers to the element itself with no specific mineralogical or economic connotation.

Fluorine Data From Central Colorado

The U.S. Geological Survey maintains a database of geochemical analyses of rocks (referred to as the "USGS rock database" in the rest of the report). This database contains fluorine analyses for 4,293 rock and soil samples collected in and near the central Colorado study area. The database of fluorine analyses included 3,798 samples of unmineralized materials and 495 samples of mineralized rocks. Most of the samples were collected from surface exposures of the rocks. A minority of the samples came from drill cores, underground mines, and soils. Mineralized samples included both vein materials and altered rocks. Except for samples from the Browns Canyon and Jamestown fluorspar deposits, none of the samples were collected with fluorine in mind and thus represent unbiased sampling for that element. Samples were collected where geoscientists happened to be working, and some areas received more study than others. As a result, the sampling sites are not evenly distributed across the study area (fig. 2).

Geoscientists who submitted the samples for analysis generally provided sample information, such as rock type and age, and coordinate locations for many of the samples. Many samples contained more specific information, including formation name and other attributes. A number of the samples contain only partial information, such as only "granite" or "sandstone" or the geologic period. Specific ages and lithologies were added to most of these samples by comparing the sampling sites with the State and larger scale geologic maps (Tweto, 1979a). Additional descriptions of the sample information are provided in the **Appendix** (click on the link to access the Excel file).

The locations provided for some sampling sites were no more specific than the latitude and longitude of the southeast corner of the 7.5-minute quadrangle from which the sample was collected. In many of these cases, the scientist also provided enough lithologic and age information to enable classification of the samples when compared with geologic maps of the quadrangle. In some cases, however, little or no additional information was provided, and the samples were categorized as "unknown." Additional fluorine data collected during various geoscience studies have been published by USGS and non-USGS authors. Those data are described in the relevant sections of the report, but they were not included in the master database. The main reasons for omitting them were the general absence of specific sample-location data and incomplete details on the geologic context of the samples. Because this report focuses on fluorine in rocks, the database also does not include many published analyses of fluorine in surface water and groundwater in central Colorado by USGS and other researchers.

Base map from National Geographic Society TOPO! Maps

Figure 1. Location of the central Colorado study area (dark outline) and place names mentioned in the text. The shaded relief base map and the study area outline shown here serve as the base map for subsequent figures.

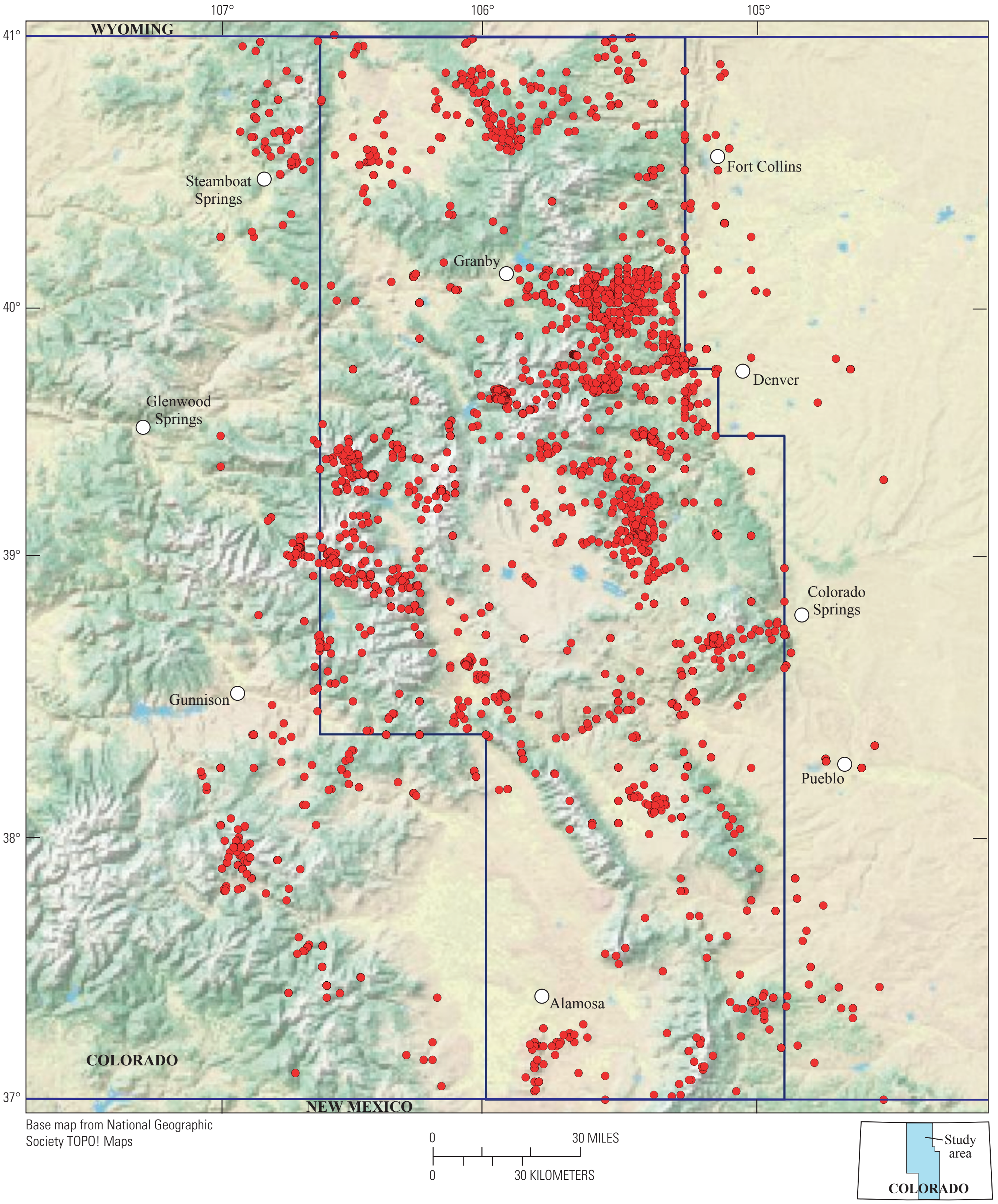


Figure 2. Locations of all samples (red dots) that were collected in and near the central Colorado study area and analyzed for fluorine. Refer to figure 1 for additonal geographic locations.

Results

All of the data are listed in the **Appendix** (click on the link to access the Excel file), which is an Excel spreadsheet. The data can be sorted by any combination of criteria, such as age and primary lithology, including igneous, metamorphic, sedimentary, soil, tephra, tectonic, or ore. These data and categories led to a number of general and age- and lithology-specific results, which are discussed briefly in this section and in later sections of this report. A summary of the data is given in table 1, and the average and median fluorine contents in the various types and ages of rocks are shown in figure 3.

The average fluorine content of all 4,293 samples in the database was 3,270 ppm, which is 6.6 times the average crustal abundance of 650 ppm (Krauskopf, 1979). Fluorine values ranged from 40 to 368,000 ppm. However, the median fluorine content was 680 ppm, close to the crustal average, and 3,903 samples (91 percent) had less than the 3,270 ppm average for the database. The high fluorine contents of a comparatively small number of Neoproterozoic igneous and Tertiary fluorspar ore samples created the high average compared to the median.

The 3,798 samples of unmineralized rocks, regardless of lithology, age, or source, contained an average of 1,550 ppm fluorine, about 2.4 times the crustal average. Similar to the database as a whole, however, the median fluorine content of unmineralized samples was much lower (640 ppm), which again is close to the crustal average; 3,076 (81 percent) of the samples had less than the 1,550 ppm average. High-fluorine samples from the Mesoproterozoic Pikes Peak intrusive complex, the average for which is 3,620 ppm fluorine, were the primary reason for the high average for the database as a whole.

The 495 samples of mineralized rock contained an average of 16,940 ppm fluorine, with a range of 100 to 368,000 ppm and a median of 990 ppm. Of these, 112 samples contained 10,000 ppm or more fluorine. The Jamestown fluorspar district in Boulder County accounted for 96 of those 112 samples (about 20 percent of all of the mineralized rocks). The Jamestown samples, many of which came from drill cores in orebodies, averaged 69,780 ppm fluorine. When that average was treated as a single, high-fluorine “sample” in order to reduce the bias induced by this single source, the revised average of mineralized rocks was 4,420 ppm fluorine, which is still much higher than the median for the mineralized rocks as a group. Many samples (248) contained less than 1,000 ppm fluorine, indicating that not all mineralizing systems produced elevated levels of fluorine.

In summary, while many samples in the central Colorado database contained significantly large amounts of fluorine, the median values indicate that fluorine levels in the regional and various age-dependent subsets are fairly normal when compared to the crustal average. Specific events, such as the intrusion of fluorine-rich magmas in the Neoproterozoic and the concentration of fluorine into some Tertiary fluorspar deposits, produced areas and sites that are very rich in fluorine.

Fluorspar Deposit Database

The U.S. Geological Survey maintains a database of information, called the Mineral Resources Data Systems (MRDS) database, on thousands of mineral deposits and prospects in the United States. This database contains 121 records from central Colorado that indicate that fluorite was a primary, major, or minor commodity during mining. Fluorite was present in trace amounts in other mineral deposits and occurrences in the area, on the basis of various published reports, although that information was not mentioned in the MRDS records of those deposits. The fluorite records include the major fluorspar mines in the region, such as those in the Northgate, Jamestown, Browns Canyon, Poncha Springs, and Wagon Wheel Gap districts. They also include numerous mines from which fluorite was a byproduct mineral or in which fluorite was listed as a component of the ore deposit. These records include limited information regarding name, location, characteristics, major and minor commodities, and mining, as well as some references to publications on the deposit or occurrence, if available. These records originally were created several decades ago, and they have been updated inconsistently since their creation.

For the present study, a modified database was created that used the basic location and commodity information, updated other fields, and added new fields that pertained to various geologic characteristics. The new fields include the district, the general category of deposit, the general deposit type, and the known or approximate age of the deposit or occurrence. These new fields also collectively provide information on the associated igneous rocks and, in some cases, the tectonic setting. These new data were derived from a wide variety of published reports on the geology, geochronology, and metallogenesis of the region. Almost all of the fluorspar districts and many of the fluorite-bearing deposits were visited by the author in the early 1990s, and those data and observations were incorporated into the database and the age-based descriptions that comprise the later sections of this report. The modified database is shown in table 2, and the deposit locations are shown in figure 4.

The goal of creating the modified database was to place the fluorspar deposits in a genetic context for use in metallogenic studies and mineral assessments of fluorspar and fluorite-bearing mineral deposits in the study area. Much of the organization of the ensuing sections of this report is based on the general age and geologic categories listed in the database.

Table 1. Summary of fluorine contents (in parts per million) of rocks in the central Colorado study area by age and lithology.

[Ga, giga-annum; --, no data; ppm, parts per million]

Lithologic suite	Total No. samples	Total average	No. samples (without extremes[1])	Average (without extremes[1])	High (without extremes[1])	Low[2]	Median (without extremes[1])
Paleoproterozoic metamorphic rocks	816	1,180	--	--	81,000	50	645
Paleoproterozoic (1.7 Ga) igneous rocks	463	2,480	459	*1,100*	*260,600*	70	*500*
Mesoproterozoic (1.4 Ga) igneous rocks	285	1,750	284	*1,116*	*170,000*	60	*500*
Mesoproterozoic (1.1 Ga) igneous rocks	274	2,500	273	*2,140*	*13,300*	60	*1,700*
Neoproterozoic-Cambrian igneous rocks	80	970	--	--	12,300	60	400
Neoproterozoic-Paleozoic diatremes	35	2,070	--	--	8,600	20	1,500
Sedimentary rocks (Phanerozoic)	587	1,990	448	*1,600*	*62,000*	40	*560*
Cenozoic intrusive rocks	593	1,220	592	*1,170*	*33,900*	90	*740*
Cenozoic volcanic rocks	334	930	331	*930*	*6,900*	40	*600*
Cenozoic tephras	76	820	--	--	3,100	20	600
Soils	209	630	--	--	7,160	40	510
Miscellaneous	15	880	--	--	2,400	100	810
Unidentified rocks	31	740	--	--	4,500	50	600
All non-ore rocks	**3,798**	**1,550**	**--**	**--**	**260,600**	**40**	**640**
All ore and altered rock[3]	**495**	**16,940**	**--**	**--**	**368,000**	**100**	**990**
All samples	**4,293**	**3,270**	**--**	**--**	**368,000**	**20**	**680**
Crustal average		650					

[1] Some age/lithology suites included a small number of samples (four maximum) that had fluorine contents that were extremely high compared with the rest of the data set. When these samples were removed, then the resulting average and median more closely represented the vast majority of the samples. Some suites did not contain samples such as these, and the two sets of data shown above are the same. Normal font is used for values that represent the total suite; italic font is used for values that represent averages, high values, and medians of data bases from which a few extremely high values were eliminated.

[2] The low is the same for both the total and subset data bases for each category.

[3] Includes 96 high-fluorine (greater than 1 percent) samples collected from the Jamestown fluorspar district.

[4] Uses one averaged sample for the Jamestown district to remove the bias produced by that large, high-fluorine data set. The average for the 96 Jamestown samples is 69,780 ppm F, with a median value of 56,000 ppm.

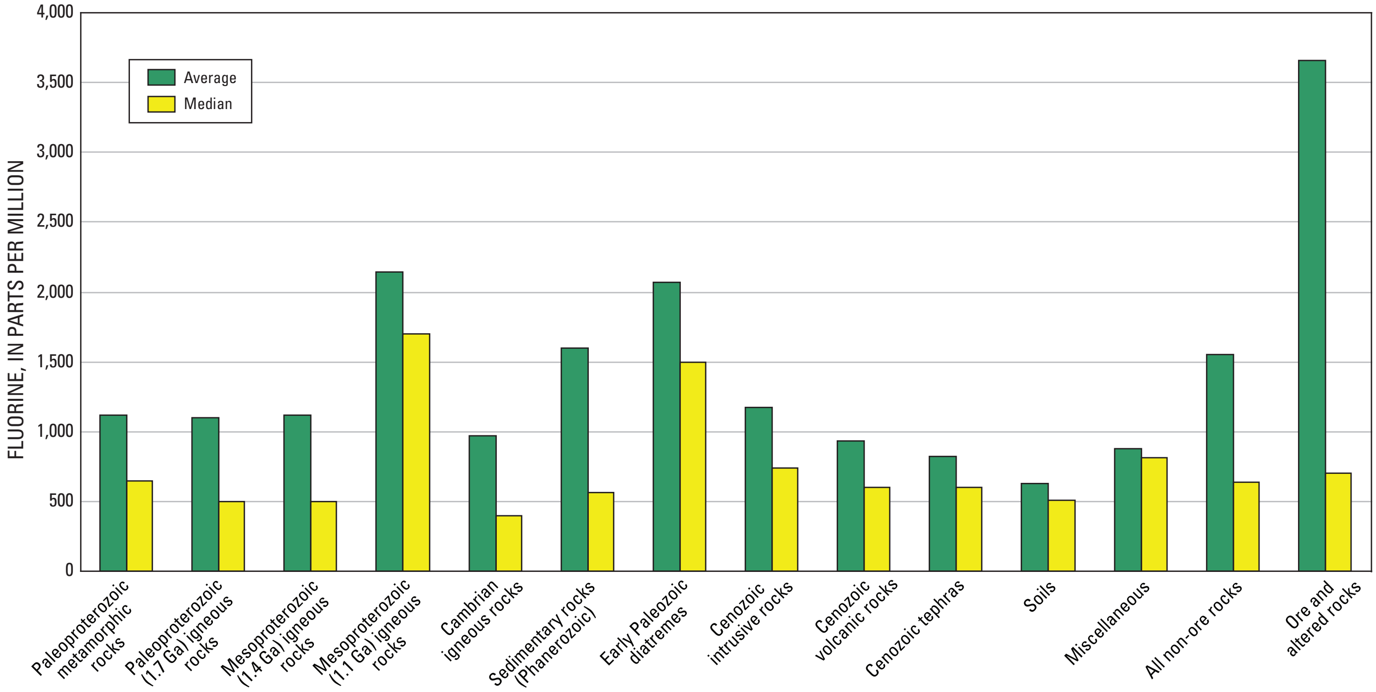


Figure 3. Average and median fluorine values for different age and lithologic groupings of rocks found in the central Colorado study area. All of the data are available in the Appendix.

Proterozoic Plutonic and Metamorphic Rocks

Geologic History

The underpinnings of central Colorado started out as Paleoproterozoic sedimentary and volcanic rocks that were intruded and metamorphosed by Paleo- and Mesoproterozoic plutons and batholiths (fig. 5). The crust began to be assembled at about 1.8 Ga as various mantle-derived volcanic arc and backarc terranes accreted against the Archean Wyoming province craton in southernmost Wyoming (DePaolo, 1981; Reed and others, 1987; Bickford, 1988). The resulting supracrustal rock assemblage was composed dominantly of volcanic and volcaniclastic rocks in the northern and southern parts of Colorado and more abundant graywackes, pelites, and lesser quartzites in the Front Range, Wet Mountains, and other areas (fig. 5; Tweto, 1980). In a very general way, almost all of the sediment-dominated supracrustal rocks are north of an east-northeast-trending line extending across the area south of Salida, and supracrustal rocks to the south are predominantly volcanic in origin (fig. 5). In early studies of Proterozoic rocks in the Front Range (cf. Lovering and Goddard [1950] and other studies into the 1960s), the metamorphic rocks in that area were called the Idaho Springs Formation, but that term no longer is used (Tweto, 1987).

The metavolcanic rocks include both tholeiitic basalt and rhyolite, typical of a bimodal assemblage (DePaolo, 1981; Bickford, 1988; Aleinikoff and others, 1993); intermediate-composition volcanic rocks are scarce. Isotopic data suggest that virtually all of the magmas were derived from a ~1.8-Ga mantle source, with little to no contribution from older crustal material, even in areas adjacent to the Archean craton (Aleinikoff and others, 1993; Premo and Loucks, 2000).

The metasedimentary rocks had both local and exotic sources. Most of the volcaniclastic sediments in the Paleoproterozoic assemblage were derived locally from the volcanic rocks. However, the felsic volcanic rocks were not voluminous enough, when partially eroded, to produce the large amounts of pelitic sediments (graywacke, shale, quartzite) in the Front Range and Wet Mountains areas. Instead, dating of zircons in the pelitic rocks points to Paleoproterozoic cratonic sources far to the northeast and east, with minimal to no contribution from the Archean Wyoming province to the north (Aleinkoff and others, 1993).

Three suites of Proterozoic plutons and batholiths intruded the sedimentary and volcanic rocks, roughly at 1.7, 1.4, and 1.1 Ga (fig. 5). Isotopic data (DePaolo, 1981; Bickford, 1988; Aleinikoff and others, 1993) show that all three suites were derived from sources no older than 1.8 Ga. The 1.7-Ga suite (Boulder Creek Granodiorite, granitic rocks of the Rahwah batholith, Kroenke and Denny Creek Granodiorites, Cross Creek Quartz Monzogranite, and others) was derived from partial melting at depth of the volcanic and sedimentary rocks of the accreted terrains, with no melt contribution from a subducted Archean slab. Isotopic and chemical differences between these plutons suggest that the relative amounts of volcanic and non-volcanic materials in the source areas affected the compositions of the resulting magmas.

Table 2. Fluorspar deposits and localities with fluorite as a byproduct commodity in the central Colorado study area.

[Updated from data in the USGS MRDS database (see text). Au, gold; Ag, silver; Ba, barite; Be, beryllium; Bi, bismuth; Cu, copper; F, fluorine; Fld, feldspar; Gem, gemstone; Gar, garnet; Mic, mica; Mn, manganese; Mo, molybdenum; Pb, lead; Qtz, quartz; REE, rare earth elements; Th, thorium; U, uranium; V, vanadium; W, tungsten; Zn, zinc; Zr, zircon. Development: DP-I, developed producer, inactive; IP, inactive producer; O, occurrence]

No.	Site name	Latitude (dec. deg.)	Longitude (dec. deg.)	District	General category
D001076	Sweet Home mine	39.3000	–106.1181	Alma	Climax-type Mo deposit
ARW03	Unknown	40.4200	–105.9632	Teller City	Climax-type Mo deposit
DC04930	Fluorspar	38.9861	–106.4278	Winfield (La Plata)	Climax-type Mo deposit
ARW04	Fortune	38.9779	–106.4662	Winfield (La Plata)	Climax-type Mo deposit
D002198	Fair Day mine	40.0667	–105.4000	Gold Hill	Laramide alkalic pluton
DC03451	Alice mine	40.1225	–105.3981	Jamestown	Laramide alkalic pluton
DC04754	Argo	40.1369	–105.3981	Jamestown	Laramide alkalic pluton
DC04740	Blue Jay	40.1100	–105.3867	Jamestown	Laramide alkalic pluton
DC03091	Brown Spar	40.1225	–105.3981	Jamestown	Laramide alkalic pluton
DC03369	Burlington	40.1225	–105.3981	Jamestown	Laramide alkalic pluton
DC03092	Chancellor	40.1225	–105.3981	Jamestown	Laramide alkalic pluton
DC04756	Consolation mine	40.1369	–105.3981	Jamestown	Laramide alkalic pluton
DC04761	Emmet	40.1225	–105.3981	Jamestown	Laramide alkalic pluton
DC04753	Gladstone, Alice, McKinley	40.1369	–105.3981	Jamestown	Laramide alkalic pluton
DC05012	Invincible	40.1247	–105.3867	Jamestown	Laramide alkalic pluton
DC04719	L.Evans Lease	40.1247	–105.3867	Jamestown	Laramide alkalic pluton
D003043	Orion and Saddle No. 6 fluorspar	40.1250	–105.3833	Jamestown	Laramide alkalic pluton
DC03580	Orion, Saddle No. 6	40.1225	–105.3981	Jamestown	Laramide alkalic pluton
DC04737	Poorman	40.1100	–105.3867	Jamestown	Laramide alkalic pluton
DC04728	Spruce Gulch	40.1247	–105.3300	Jamestown	Laramide alkalic pluton
D002196	Victory Lode Claim	40.1167	–105.3750	Jamestown	Laramide alkalic pluton
DC04755	Warren Spar	40.1369	–105.3981	Jamestown	Laramide alkalic pluton
DC03453	Yellow Girl	40.1225	–105.3981	Jamestown	Laramide alkalic pluton
DC04718	Yellow Girl	40.1247	–105.3867	Jamestown	Laramide alkalic pluton
DC05724	Hill, Bunkhous	39.7767	–105.5181	Central City	Laramide calc-alkaline pluton
DC03242	Blue Stone Mines, Inc.	39.9356	–105.6608	Eldora	Laramide calc-alkaline pluton
DC04675	Unknown	39.4264	–105.8019	Jefferson	Laramide calc-alkaline pluton
D010369	Cheyenne district	38.7475	–104.9111	St. Peters Dome	Mid-Tertiary alkalic pluton
D010307	Duffields deposit	38.7383	–104.9086	St. Peters Dome	Mid-Tertiary alkalic pluton
D010309	Mattie B. mine	38.7439	–104.9092	St. Peters Dome	Mid-Tertiary alkalic pluton
D010311	Timberline tunnel	38.7294	–104.9094	St. Peters Dome	Mid-Tertiary alkalic pluton
DC02462	Stumbling Stud mine	37.7728	–105.0185	Gardner Hill	Mid-Tertiary alkalic pluton
DC01168	Fluorspar Lode	38.6011	–106.4500	Tincup	Mid-Tertiary calc-alkaline pluton
DC01515	Graphite	38.6839	–106.4571	Tincup	Mid-Tertiary calc-alkaline pluton
D010520	Unnamed graphite deposit	38.6736	–105.9800	Turret	Mid-Tertiary calc-alkaline pluton
DC04273	Unknown	38.8031	–104.8969	Woodland Park area	Mid-Tertiary calc-alkaline pluton
DC01503	Jocomo	38.0214	–105.2050	Antelope Creek	Mid-Tertiary volcanism
D009468	Texas Creek prospect	38.4553	–105.6076	Texas Creek area	Mid-Tertiary volcanism
DC02260	Wagon Wheel Gap	37.7478	–106.8178	Wagon Wheel Gap	Mid-Tertiary volcanism
D010674	Wagon Wheel Gap fluorspar mine	37.7461	–106.8258	Wagon Wheel Gap	Mid-Tertiary volcanism
D008884	Beryl deposit	38.0936	–105.7456	Blake (Mirage)	Pre-extension calc-alkaline
DC01373	Unknown	37.8614	–105.5911	Liberty (Music)	Pre-extension calc-alkaline
D010545	Spaulding mine	40.7122	–105.3131	Stove Prairie area	Proterozoic granite
DC05111	Aspen Lode pegmatites	39.3397	–105.3944	Buffalo Creek area	Proterozoic pegmatite
D010210	Buffalo mine	39.3922	–105.2856	Buffalo Creek area	Proterozoic pegmatite
DC01603	Unknown	39.3958	–105.2875	Buffalo Creek area	Proterozoic pegmatite
DC01604	Unknown	39.3958	–105.2875	Buffalo Creek area	Proterozoic pegmatite
D004859	Cascade-Ute Pass	38.8867	–104.9653	St. Peters Dome	Proterozoic pegmatite
D010308	Eureka shaft	38.7508	–104.9008	St. Peters Dome	Proterozoic pegmatite
D010379	Stove Mountain mineral locality	38.7761	–104.9008	St. Peters Dome	Proterozoic pegmatite
DC04167	Unknown	38.7597	–104.9156	St. Peters Dome	Proterozoic pegmatite
DC03182	Fern Creek	39.1642	–105.1011	Devils Head Pegmatite	Proterozoic pegmatite
DC03177	Rusty Pick	39.2075	–105.0453	Devils Head Pegmatite	Proterozoic pegmatite
DC00316	Kyner	38.9808	–105.3347	Lake George area	Proterozoic pegmatite
DC00317	Teller pegmatite claim	38.9708	–105.3694	Lake George area	Proterozoic pegmatite
DC01611	Butte #1	39.3814	–105.2131	South Platte Pegmatite	Proterozoic pegmatite
DC01609	Catherine No. 1 claim	39.3814	–105.2131	South Platte Pegmatite	Proterozoic pegmatite
D010175	Dazie Bell mine	39.3806	–105.2011	South Platte Pegmatite	Proterozoic pegmatite
DC01598	Delbert claims #1 & 2	39.3958	–105.2131	South Platte Pegmatite	Proterozoic pegmatite
D010172	Judy Ann mine	39.3833	–105.2017	South Platte Pegmatite	Proterozoic pegmatite
D010171	Little Patsy mine	39.4106	–105.2297	South Platte Pegmatite	Proterozoic pegmatite

Table 2.—Continued

No.	Deposit type	Deposit age	Primary	Major commodity	Minor commodity	Development commodity
D001076	Vein/replacement	26 Ma	W	W Cu Pb	DP-I	
ARW03	Vein	<28 Ma	F	F	O	
DC04930	Climax Mo	39 Ma	F	F	IP	
ARW04	Climax Mo	39 Ma	F	F	O	
D002198	Vein	56 Ma	U	U Pb Ag F	IP	
DC03451	Vein	56 Ma	Pb	Pb Ag Au Cu F	IP	
DC04754	Vein	56 Ma	U	U F Au	O	
DC04740	Vein	56 Ma	U	U F	O	
DC03091	Vein	56 Ma	F	F U	IP	
DC03369	Vein	56 Ma	F	F U	IP	
DC03092	Vein	56 Ma	F	F Au Pb Ag	IP	
DC04756	Vein	56 Ma	F	F	O	
DC04761	Vein	56 Ma	Au	Au F U	IP	
DC04753	Vein	56 Ma	Au	Au F	IP	
DC05012	Vein	56 Ma	Au	Au F	O	
DC04719	Vein	56 Ma	F	F	IP	
D003043	Vein	56 Ma	F	F Pb Cu	O	
DC03580	Vein	56 Ma	F	F	O	
DC04737	Vein	56 Ma	Au	Au F U	O	
DC04728	Vein	56 Ma	F	F	O	
D002196	Vein	56 Ma	U	U F	IP	
DC04755	Vein	56 Ma	F	F	IP	
DC03453	Vein	56 Ma	F	F	IP	
DC04718	Vein	56 Ma	F	F	O	
DC05724	Vein	55–60 Ma	Au	Au Ag F	O	
DC03242	Vein	55–60 Ma	F	F	O	
DC04675	Vein	Paleogene	F	F	O	
D010369	Vein	~28 Ma?	F	F Pb Zn Au Ag	Th Gem REE	IP
D010307	Vein	~28 Ma?	F	F Pb Zn Au Ag	DP-I	
D010309	Vein	~28 Ma?	F	F Pb Zn Au Ag	DP-I	
D010311	Vein	~28 Ma?	F	F Pb Zn Au Ag	DP-I	
DC02462	Replacement	24 Ma?	U	U V	DP-I	
DC01168	Vein	39 Ma?	F	F	O	
DC01515	Replacement	~35 Ma	Gar F	Gar	F	IP
D010520	Replacement	35 Ma?	Gar F	Gar	F	DP-I
DC04273	Vein	Oligocene?	Au	Au	F	O
DC01503	Vein	~33 Ma	F	F	IP	
D009468	Volcanogenic U	<33 Ma	U	U	F	DP-I
DC02260	Vein	20-25 Ma	F	F	Ba	IP
D010674	Vein	20-25 Ma	F	F	DP-I	
D008884	Vein	26 Ma?	F	F	Mn	O
DC01373	Vein	26 Ma?	F	F	IP	
D010545	Vein	Mesoproterozoic	W	W	O	
DC05111	Pikes Peak pegmatite	Mesoproterozoic	F	F	O	
D010210	Pikes Peak pegmatite	Mesoproterozoic	Au	Au Ag Pb F	DP-I	
DC01603	Pikes Peak pegmatite	Mesoproterozoic	F	F	O	
DC01604	Pikes Peak pegmatite	Mesoproterozoic	F	F	O	
D004859	Pikes Peak pegmatite	Mesoproterozoic	F	F	DP-I	
D010308	Pikes Peak pegmatite	Mesoproterozoic	F	F	Gem	DP-I
D010379	Pikes Peak pegmatite	Mesoproterozoic	Gem	Gem REE F Be	Nb TI SN Th	IP
DC04167	Pikes Peak pegmatite	Mesoproterozoic	F	F	O	
DC03182	Pikes Peak pegmatite	Mesoproterozoic	Fld	Fld Qtz Mica F	IP	
DC03177	Pikes Peak pegmatite	Mesoproterozoic	Fld	Fld Mica F	O	
DC00316	Pikes Peak pegmatite	Mesoproterozoic	F	F	O	
DC00317	Pikes Peak pegmatite	Mesoproterozoic	REE	REE F	IP	
DC01611	Pikes Peak pegmatite	Mesoproterozoic	Fld	Fld Qtz F	IP	
DC01609	Pikes Peak pegmatite	Mesoproterozoic	Fld	Fld Qtz Mica F Gem	O	
D010175	Pikes Peak pegmatite	Mesoproterozoic	Qtz	Qtz	F REE	DP-I
DC01598	Pikes Peak pegmatite	Mesoproterozoic	Fld	Fld Qtz Mica F	IP	
D010172	Pikes Peak pegmatite	Mesoproterozoic	Fld	Fld Qtz Mica	F REE	O
D010171	Pikes Peak pegmatite	Mesoproterozoic	Qtz	Qtz Fld	F REE	DP-I

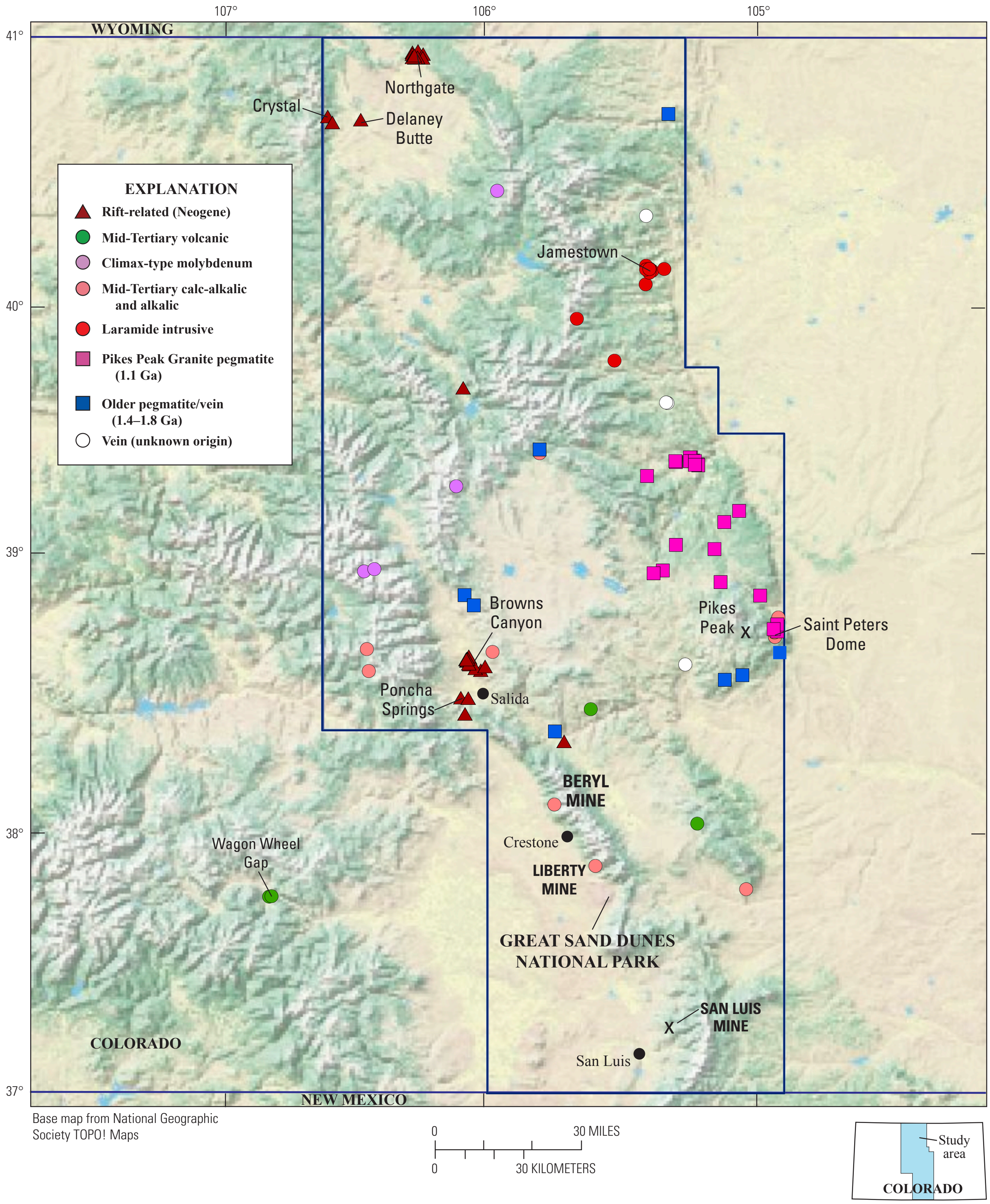
Table 2. Fluorspar deposits and localities with fluorite as a byproduct commodity in the central Colorado study area.—Continued

[Updated from data in the USGS MRDS database (see text). Au, gold; Ag, silver; Ba, barite; Be, beryllium; Bi, bismuth; Cu, copper; F, fluorine; Fld, feldspar; Gem, gemstone; Gar, garnet; Mic, mica; Mn, manganese; Mo, molybdenum; Pb, lead; Qtz, quartz; REE, rare earth elements; Th, thorium; U, uranium; V, vanadium; W, tungsten; Zn, zinc; Zr, zircon. Development: DP-I, developed producer, inactive; IP, inactive producer; O, occurrence]

No.	Site name	Latitude (dec. deg.)	Longitude (dec. deg.)	District	General category
DC01591	Little Patsy Nos. 1&2, Fawn No. 1	39.4103	–105.2317	South Platte Pegmatite	Proterozoic pegmatite
DC01595	Luster Lode claim	39.3958	–105.2317	South Platte Pegmatite	Proterozoic pegmatite
D010170	Luster No. 1 mine	39.3981	–105.2206	South Platte Pegmatite	Proterozoic pegmatite
DC01594	Oregon #1 & 2 & 3 claim	39.3958	–105.2317	South Platte Pegmatite	Proterozoic pegmatite
D010180	South Platte pegmatite district	39.3956	–105.2144	South Platte Pegmatite	Proterozoic pegmatite
DC01610	Susy Belle	39.3814	–105.2131	South Platte Pegmatite	Proterozoic pegmatite
DC03811	Quartz lode, Dands #2 quarry	39.0781	–105.2858	Unnamed	Proterozoic pegmatite
D010230	Snowflake mine	39.0631	–105.1383	Woodland Park area	Proterozoic pegmatite
D004970	Unnamed mine	38.9375	–105.1167	Woodland Park area	Proterozoic pegmatite
DC03748	Jones & Young lease	38.3706	–105.7442	Cotopaxi	Proterozoic pegmatite
ARW02	Mill Canyon fluorite area	38.5848	–105.0338	Phantom Canyon area	Proterozoic pegmatite
D009874	Phantom lode	38.5667	–105.1000	Phantom Canyon area	Proterozoic pegmatite
D010247	Luella pegmatite	38.8494	–106.0511	Trout Creek Pass area	Proterozoic pegmatite
D010231	Yard mine	38.8875	–106.0861	Trout Creek Pass area	Proterozoic pegmatite
DC04179	Unknown	38.6717	–104.8928	St. Peters Dome	Proterozoic pegmatite
DC04674	Donna Lou, Guernsey	39.4408	–105.8019	Jefferson	Proterozoic pegmatite
DC05604	Unknown	39.6747	–106.0917	Frisco	Rio Grande rift
D010130	Alderman deposit	38.6253	–106.0661	Browns Canyon	Rio Grande rift
DC03007	Arkansas claim, Bapp mine, American Fluorspar mine	38.6294	–106.0703	Browns Canyon	Rio Grande rift
D010257	Blue Stone deposit	38.6486	–106.0750	Browns Canyon	Rio Grande rift
DC01007	Brown Canyon deposit	38.6575	–106.0703	Browns Canyon	Rio Grande rift
D010255	Browns Canyon district	38.6472	–106.0750	Browns Canyon	Rio Grande rift
DC05277	Browns Canyon or Dostal mine	38.6294	–106.0703	Browns Canyon	Rio Grande rift
DC05253	Chimney Hill	38.6294	–106.0703	Browns Canyon	Rio Grande rift
D010254	Chimney Hill mine	38.6439	–106.0786	Browns Canyon	Rio Grande rift
DC05282	Chimney Hill, Manganese Hill	38.6444	–106.0793	Browns Canyon	Rio Grande rift
D010253	Colorado-American mine	38.6325	–106.0661	Browns Canyon	Rio Grande rift
D010250	Delay adit	38.6417	–106.0742	Browns Canyon	Rio Grande rift
D010249	Last Chance mine	38.6519	–106.0661	Browns Canyon	Rio Grande rift
DC05280	Lionell	38.6472	–106.0728	Browns Canyon	Rio Grande rift
DC05254	Manganese	38.6294	–106.0703	Browns Canyon	Rio Grande rift
D010246	Manganese Hill mine	38.6450	–106.0800	Browns Canyon	Rio Grande rift
D010244	Puzzle deposit	38.6478	–106.0728	Browns Canyon	Rio Grande rift
D010239	Snowflake deposit	38.6386	–106.0764	Browns Canyon	Rio Grande rift
D010260	Unnamed fluorite deposit	38.6042	–106.0239	Browns Canyon	Rio Grande rift
D010261	Unnamed fluorite deposit	38.6117	–106.0453	Browns Canyon	Rio Grande rift
D010262	Unnamed fluorite deposit	38.6186	–106.0089	Browns Canyon	Rio Grande rift
D010232	White King deposit	38.6489	–106.0772	Browns Canyon	Rio Grande rift
DC05403	Blue Jay	38.3347	–105.7094	Cotopaxi	Rio Grande rift
ARW01	Delaney Butte	40.6900	–106.4800	Crystal	Rio Grande rift
DC01705	Non magnet, Joe, Fluorine	40.6794	–106.5872	Crystal	Rio Grande rift
DC01704	Non-magnetic fluorspar mine	40.7025	–106.6053	Crystal	Rio Grande rift
DC01732	Camp Creek Claims	40.9522	–106.2622	Northgate	Rio Grande rift
DC01734	Fluorine Claims	40.9392	–106.2433	Northgate	Rio Grande rift
D002496	Fluorite Nos 1–8	40.9250	–106.2500	Northgate	Rio Grande rift
DC01738	Fluorspar	40.9272	–106.2831	Northgate	Rio Grande rift
DC01733	Fluorspar Claims	40.9397	–106.2828	Northgate	Rio Grande rift
DC01737	Gero	40.9244	–106.2744	Northgate	Rio Grande rift
DC01735	Hankins & Follett prospects	40.9253	–106.2439	Northgate	Rio Grande rift
DC01729	Northgate Ozark	40.9469	–106.2831	Northgate	Rio Grande rift
DC01736	Penbar	40.9258	–106.2636	Northgate	Rio Grande rift
DC01739	Springer Pit	40.9342	–106.2764	Northgate	Rio Grande rift
DC02710	Poncha mine	38.4974	–106.0713	Poncha Springs area	Rio Grande rift
D010245	Poncha Pass Claim	38.4387	–106.0842	Poncha Springs area	Rio Grande rift
D010237	Unnamed fluorspar deposit	38.5001	–106.1000	Poncha Springs area	Rio Grande rift
D010219	Augusta Lode mine	39.6172	–105.3166	Evergreen	Unknown
DC02404	Bull Hill claim	39.6178	–105.3231	Evergreen	Unknown
DC01920	Bashor mine	40.3250	–105.3981	None	Unknown
D004852	Blue Nugget	38.6250	–105.2500	Unnamed	Unknown

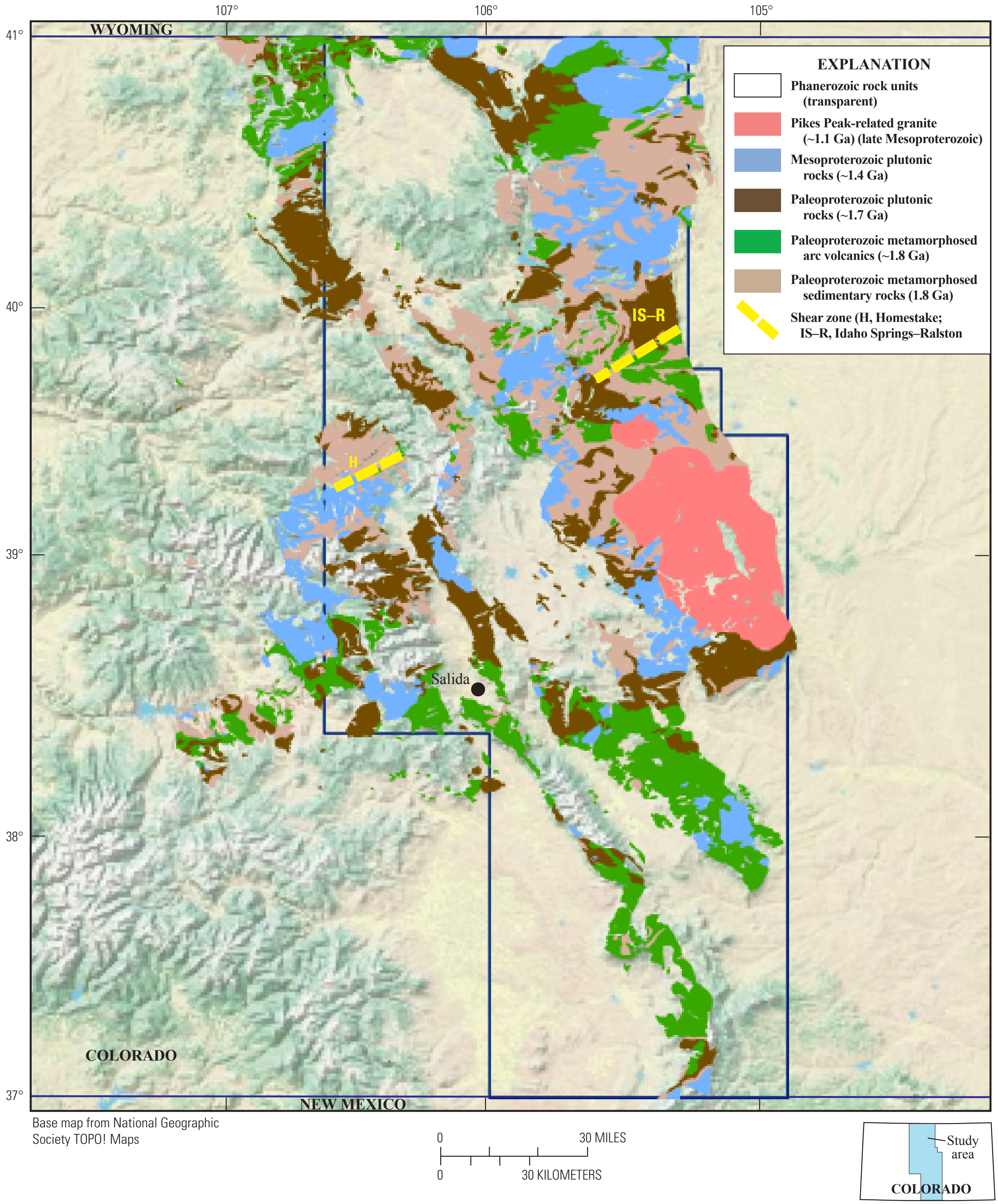
Table 2.—Continued

No.	Deposit type	Deposit age	Primary	Major commodity	Minor commodity	Development commodity
DC01591	Pikes Peak pegmatite	Mesoproterozoic	Fld	Fld Mica Qtz F Gem	IP	IP
DC01595	Pikes Peak pegmatite	Mesoproterozoic	Fld	Fld Mica Qtz F Gem	IP	IP
D010170	Pikes Peak pegmatite	Mesoproterozoic	Fld	Fld F	DP-I	DP-I
DC01594	Pikes Peak pegmatite	Mesoproterozoic	Fld	Fld Qtz Mica F Gem	IP	IP
D010180	Pikes Peak pegmatite	Mesoproterozoic	Fld	Fld Qtz	F REE Gem	IP
DC01610	Pikes Peak pegmatite	Mesoproterozoic	Fld	Fld Mica Qtz F Gem	O	O
DC03811	Pikes Peak pegmatite	Mesoproterozoic	Fld	Fld F REE	IP	IP
D010230	Pikes Peak pegmatite	Mesoproterozoic	Fld	Fld	DP-I	DP-I
D004970	Pikes Peak pegmatite	Mesoproterozoic	F	F Zr	IP	IP
DC03748	Paleoprot. pegmatite	Paleoproterozoic	F	F	Fld Qtz Be	DP-I
ARW02	Paleoprot. pegmatite	Paleoproterozoic	F			IP
D009874	Paleoprot. pegmatite	Paleoproterozoic	Fld	Fld Mica	DP-I	
D010247	Paleoprot. pegmatite	Paleoproterozoic	Fld	Fld	REE BI	DP-I
D010231	Paleoprot. pegmatite	Paleoproterozoic	Fld	Fld	REE BI	DP-I
DC04179	Pikes Peak pegmatite	Mesoproterozoic	Fld	Fld Gem F	Mica	O
DC04674	Mesoprot. pegmatite	Mesoproterozoic	F	F		O
DC05604	Hot Spring	Quaternary	F	F		IP
D010130	Vein	Neogene	F	F		DP-I
DC03007	Vein	Neogene	F	F		O
D010257	Vein	Neogene	F	F		DP-I
DC01007	Vein	Neogene	F	F		O
D010255	Vein	Neogene	F	F		DP-I
DC05277	Vein	Neogene	F	F		O
DC05253	Vein	Neogene	F	F		O
D010254	Vein	Neogene	F	F		DP-I
DC05282	Vein	Neogene	F	F		O
D010253	Vein	Neogene	F	F		DP-I
D010250	Vein	Neogene	F	F		DP-I
D010249	Vein	Neogene	F	F		DP-I
DC05280	Vein	Neogene	F	F		IP
DC05254	Vein	Neogene	F	F		O
D010246	Vein	Neogene	F	F		DP-I
D010244	Vein	Neogene	F	F		DP-I
D010239	Vein	Neogene	F	F		DP-I
D010260	Vein	Neogene	F	F		O
D010261	Vein	Neogene	F	F		O
D010262	Vein	Neogene	F	F		O
D010232	Vein	Neogene	F	F		DP-I
DC05403	Vein	Neogene	F	F		IP
ARW01	Vein	Neogene	F	F		O
DC01705	Vein	Neogene	F	F		O
DC01704	Vein	Neogene	F	F		IP
DC01732	Vein	Neogene	F	F		O
DC01734	Vein	Neogene	F	F		IP
D002496	Vein	Neogene	F	F		IP
DC01738	Vein	Neogene	F	F		O
DC01733	Vein	Neogene	F	F		O
DC01737	Vein	Neogene	F	F		IP
DC01735	Vein	Neogene	F	F		O
DC01729	Vein	Neogene	F	F		IP
DC01736	Vein	Neogene	F	F		O
DC01739	Vein	Neogene	F	F		O
DC02710	Vein	Neogene	F	F		DP-I
D010245	Vein	Neogene	F	F		O
D010237	Vein	Neogene	F	F		O
D010219	Vein (?)	Laramide?	Cu	Cu Ag F		DP-I
DC02404	Vein (?)	Laramide?	F	F		IP
DC01920	Vein (?)	Proterozoic?	F	F		O
D004852	Unknown	Unknown	F	F		IP



Base map from National Geographic Society TOPO! Maps

Figure 4. Locations of fluorspar deposits and major fluorite occurrences in and near the central Colorado study area (modified from the U.S. Geological Survey's Mineral Resource Data System [MRDS]). The deposits are categorized by age-magmatic-tectonic associations, as described in the text and listed in table 2. Refer to figure 1 for general geographic locations.



Base map from National Geographic Society TOPO! Maps

Figure 5. Distribution of Paleoproterozoic metamorphic and igneous rocks and Mesoproterozoic plutonic rocks in the central Colorado study area. Also shown are the Homestake (H) and Idaho Springs–Ralston (IS–R) shear zones, which formed in the Paleoproterozoic and were reactivated later in the Proterozoic, Late Cretaceous, and Tertiary. The geology was simplified from Tweto (1979a). Refer to figure 1 for general geographic locations.

The 1.4-Ga granitic plutons (Silver Plume, Saint Vrain, Sherman, Mount Evans, St. Kevin, San Isabel, Mount Ethel) are two-mica granites that largely were produced from melting of the older crustal units during extension, with little to no mantle contribution (Peterman and others, 1968; DePaolo, 1981; Aleinikoff and others, 1993). Anderson and Thomas (1985) concluded that these plutons were anorogenic and formed in an extensional environment, but Nyman and others (1994) showed evidence for a more compressional orogenic setting. These plutons typically are largely peraluminous, similar but not identical to S-type granites, and the source rocks thus may have had a significant pelitic component (Anderson and Thomas, 1985). Strontium isotopes, however, indicate that the source rocks had been depleted in rubidium and water prior to magma generation, possibly during Paleoproterozoic metamorphism (see summary and discussion in Anderson and Thomas, 1985). Several of the 1.4-Ga plutons and related pegmatites contain primary magmatic fluorite, as described below (Tweto and Pearson, 1964; Snyder and Hedge, 1978).

The host rocks for the 1.7-Ga plutons generally, but not always, reflect medium- to high-grade, multistage metamorphism and deformation. In contrast, the host rocks adjacent to the 1.4-Ga plutons were much less metamorphosed and deformed, and pluton emplacement did not affect the 1.7-Ga metamorphic assemblages (Cole, 2004). This difference could be a combination of coeval tectonics and the mechanisms and depths of emplacements. The juxtaposition of these different emplacement features may indicate that the 1.7-Ga plutons were raised to shallower crustal depths prior to the more shallow (8–9-km-deep; Anderson and Thomas, 1985) emplacement of the younger suite.

The 1.1-Ga plutons, including the dominant Pikes Peak Granite and subordinate plutons, are entirely within the Pikes Peak batholith west of Colorado Springs (fig. 6; Tweto, 1980). Actual intrusion ages range from 1.07 to 1.1 Ga (Schaerer and Allegre, 1982; Smith and others, 1999). The Pikes Peak batholith is composed of three main intrusive centers: (1) the Pikes Peak center in the southeastern part of the batholith, (2) the Lost Park center in the western lobe, and (3) the Buffalo Park center in the northern part of the batholith (fig. 6; Hutchinson, 1976). The intrusive rocks are mostly peraluminous syenogranite, but the Buffalo Park center also contains a large area of monzogranite. Isotopic data show that the granitic magmas were derived from anatectic melting of a tonalitic lower crust, such as the 1.7-Ga Boulder Creek Batholith, with little to no mantle contributions, and that they evolved through fractional crystallization (Smith and others, 1999). The monzogranite phase in the Buffalo Park center may have been produced by mixing of the anatectic melts with more mafic magmas that underplated the area. Differential Phanerozoic erosion of the batholith has exposed the three centers at different levels of emplacement. As such, the deeper intrusive levels are exposed in the Buffalo Park center, the intermediate levels in the Lost Park center, and the upper levels in the Pikes Peak center.

Late-stage intrusions in the Pikes Peak batholith include a potassic series and a sodic series (fig. 6; Wobus, 1976; Smith and others, 1999). The distribution of these plutons is independent of that of the main Pikes Peak phases, and they may have been controlled by deep linear fracture systems related to regional Mesoproterozoic extension. The potassic series is composed of biotite granite, and it geochemically and isotopically is similar to the Pikes Peak Granite. Potassic intrusives are present in the Redskin and Lake George stocks at the south edge and southeast of the Lost Park center, respectively, and in several other plutons. These magmas were derived from anatectic melting of tonalitic parent rocks, specifically the Boulder Creek Batholith. They evolved through fractional crystallization of feldspars (Smith and others, 1999) and thus are true A-type granites.

The sodic plutons are composed of syenite, fayalitic granite, and alkali gabbro. They include the Tarryall and Lake George plutons near the potassic plutons in the Lost Creek center; the Mount Rosa pluton in the Pikes Peak center near Saint Peters Dome; the small Spring Creek pluton at the southwest margin of the batholith near Cripple Creek; the West Creek stock in the eastern part of the batholith; and the Sugarloaf stock at the southern margin of the Buffalo Park center. The sodic plutons are products of fractional crystallization of a mantle-derived mafic magma, with some crustal contamination and late-stage release of fluorine-rich volatiles and pegmatitic fluids (Barker and others, 1975; Smith and others, 1999).

The Paleoproterozoic supracrustal rocks of central Colorado are part of the Yavapai crustal province, which extends southwest into Arizona. As noted above, this part of the province was accreted progressively to the Archean craton to the north along the Cheyenne belt, a northeast-striking collision zone (Hills and Houston, 1979). Within the province, geologic and isotopic data reveal intraprovince deformational zones, including the multiply-reactivated, northeast-striking Homestake and Idaho Springs–Ralston shear zones in the Sawatch and Front Ranges, respectively (fig. 5; Tweto and Sims, 1963; McCoy and others, 2001; Shaw and others, 2001), and a possible east-northeast-trending collision or backarc zone in the Gunnison-Salida area (Boardman and Condie, 1986; Bickford, 1988). The Homestake shear zone was active both during emplacement of the 1.7-Ga Cross Creek Quartz Monzonite and the 1.4-Ga St. Kevin Granite (McCoy and others, 2001), as well as during the Phanerozoic. As noted by Tweto and Sims (1963), the Homestake and Idaho Springs–Ralston shear zones helped to localize Cenozoic magmatism and mineralization that define the Colorado mineral belt.

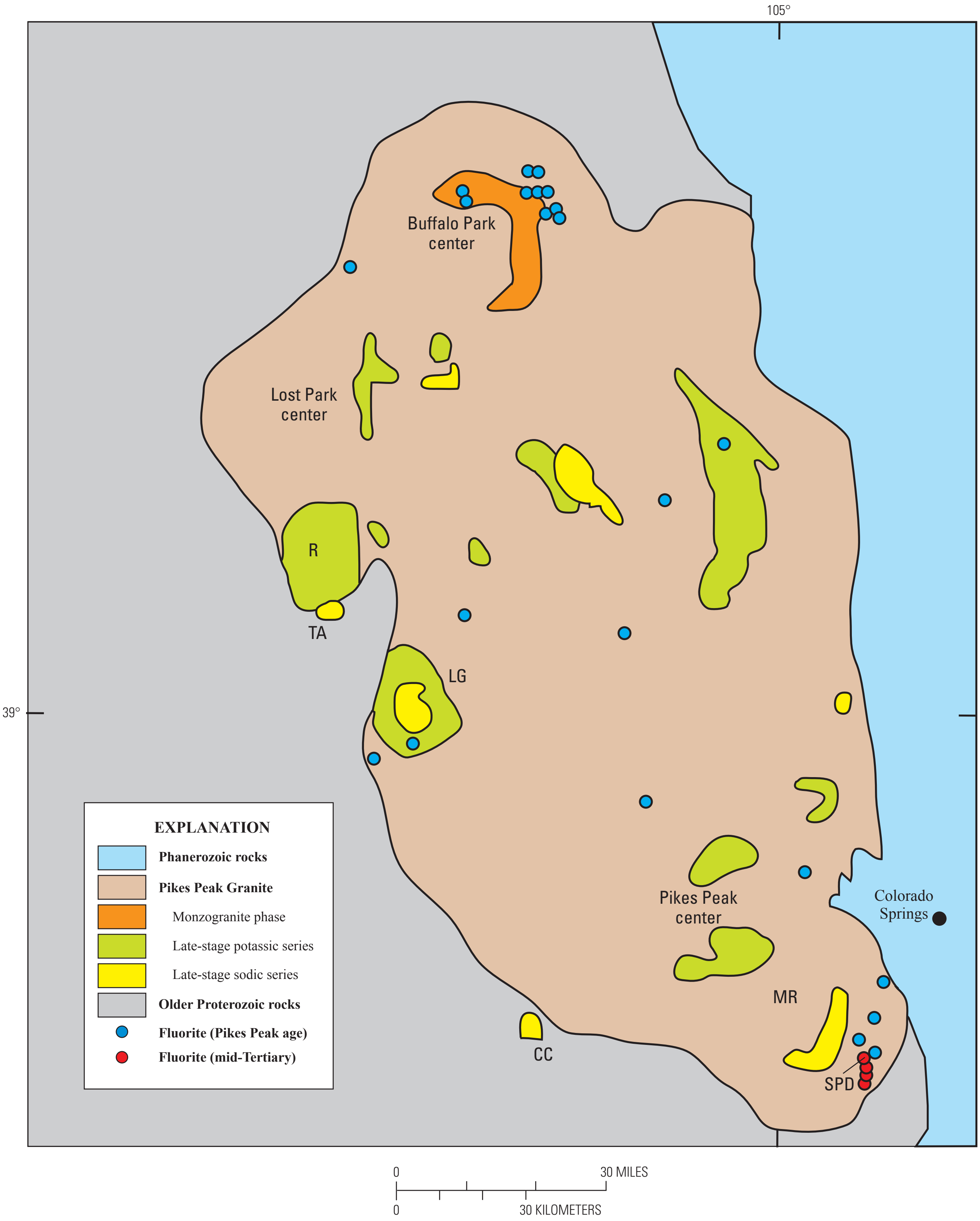


Figure 6. Generalized map of the Pikes Peak batholith, showing the three main intrusive centers (Pikes Peak, Buffalo Park, and Lost Park) and related phases and series. Intrusive phases and locations discussed in the text include: CC, Cripple Creek; LG, Lake George; MR, Mount Rosa; R, Redskin; SPD, Saint Peters Dome; and TA, Tarryall. Also shown are the locations of fluorite sites (blue circles) related to the batholith and its phases, and of fluorite deposits at Saint Peters Dome (red circles) that are related to middle Tertiary intrusive activity. The figure was simplified from Smith and others (1999). The fluorite locations are from the MRDS database (see table 2).

Fluorine in Paleoproterozoic Igneous and Metamorphic Rocks

Primary fluorite has not been documented in the Paleoproterozoic metamorphic and plutonic rocks in central Colorado. The 816 fresh metamorphic rocks in the USGS rock database average 1,180 ppm fluorine, with a range of 60 to 81,000 ppm (table 1; figs. 3, 6). The median is 645 ppm, similar to the average crustal abundance of 650 ppm. Seven samples contain more than 1 percent (10,000 ppm) fluorine; two were collected from topaz-bearing gneisses with possible fluorine-rich skarn or rhyolite protoliths, and the other five were from hornblende to granitic gneisses with no information on the possible site of the fluorine. The remaining metamorphic rock samples have an average fluorine content of 900 ppm, with a median value of 630 ppm and a range of 60 to 8,300 ppm. These rocks range in composition from leucocratic quartz-feldspar gneiss to biotite- or hornblende-rich biotite schist and amphibolite. The brief to missing lithologic descriptions given in the database preclude any correlations between the gross lithology and the fluorine content of the rock.

The 463 analyses of Paleoproterozoic plutonic rocks have an average fluorine content of 2,481 ppm, ranging from 70 to 266,000 ppm (table 1; figs. 3, 7). Four samples contain more than 100,000 ppm fluorine and must reflect the presence of secondary, but undescribed, fluorite. The remaining 459 samples contain an average of 1,100 ppm fluorine, with a median value of 500 ppm.

On the basis of isotopic data, the Paleoproterozoic magmas were derived from partial melting of the metamorphic protoliths at depth, as well as contributions from 1.7-Ga mantle materials. The prevalence of volcanic or sedimentary units in the source area affected the ultimate composition of the magma. While it is not possible to link each pluton sample in the fluorine database with a specific protolith, the ranges and averages of fluorine values in both the metamorphic and plutonic suites of samples are very similar and support derivation of the fluorine in the plutons from the combined sedimentary and volcanic protoliths. The 1.7-Ga magmas were the first to have formed by partial melting of these rocks, and much of the fluorine may have remained in the restite.

Fluorine and Fluorite in Mesoproterozoic (1.4 Ga) Plutonic Rocks

Several Mesoproterozoic plutons contain primary magmatic fluorite, either intergrown with other plutonic minerals or as late-stage veinlets, or they contain elevated fluorine contents. The best-studied of these is the Mount Ethel pluton in the Park Range and North Park area in northern Colorado (fig. 1; Snyder and Hedge, 1978). The pluton contains several intrusive phases, and the east-northeast-trending composite body is approximately 65 km long and 11 km wide. The most extensive part of the pluton is composed of the biotite granite and quartz monzonite of Roxy Ann Lake, but compositions range from diorite to leucogranite.

Fluorite and elevated fluorine concentrations are common in the Mount Ethel pluton, but their distributions are not coincident. Non-fluorite fluorine occurs primarily in biotite, with lesser amounts in hornblende, apatite, and sphene (Snyder and Hedge, 1978). Thus, the amount of fluorine is greatest in the most mafic phases of the pluton (1,500 to 4,000 ppm) and least in the most felsic phases (100 ppm) (Snyder and Hedge, 1978), regardless of their location within the complex. Fluorite, in contrast, is most abundant in the outer margins of the plutonic complex, regardless of the rock composition. It forms interstitial crystals within the plutonic units, late-magmatic veins with magmatic silicates, and deuteric or post-magmatic grains. Snyder and Hedge (1978) concluded that all of the fluorite was magmatic in origin, and that fluorine and the various fluorite stages reflect progressive magmatic differentiation and concentration, with fluorite deposited along the margins of the complex near the end of magmatism. As discussed in the section on "Fluorine Related to the Rio Grande Rift," Snyder and others (1987) suggested that the fluorine in Neogene fluorspar districts in the North Park area had a provenance in the Mount Ethel fluorine and fluorite.

The St. Kevin Granite in the Sawatch Range west of Leadville contains elevated amounts of fluorine and trace amounts of interstitial, magmatic fluorite (Tweto and Pearson, 1964). In addition, fluorite veinlets related to the Oligocene intrusive complex at Turquoise Lake (see fig. 10) cut the St. Kevin Granite, as described in the section on "Fluorine Related to Middle Tertiary Calc-alkaline Igneous Centers." Like the Mount Ethel pluton, the St. Kevin Granite has several intrusive phases ranging from granite, the most-abundant felsic composition, to lesser amounts of granodiorite. Fluorine contents range from 400 to 3,400 ppm (Tweto and Pearson, 1964; Alan Wallace, unpubl. data, 1984). The granitic normal facies is the most extensive intrusive phase, and two samples contained an average of 900 ppm fluorine. The highest reported fluorine value came from the fine-grained facies, which is composed of quartz monzonite and occurs along the margin of the composite intrusive. One sample of this phase contained 3,400 ppm fluorine (Tweto and Pearson, 1964).

Other Mesoproterozoic plutons in central Colorado contain somewhat elevated fluorine but little or no fluorite. Analyses of the Silver Plume Granite in the central Front Range range from 200 to 6,000 (average 1,100) ppm fluorine, and those of the Sherman Granite in the northern Front Range show 400 to 2,000 ppm fluorine (average 700) (**Appendix**; click on the link to access the Excel file). No data are available regarding the specific intrusive phases or compositions related to these analyses. Snyder and Hedge (1978) reported magmatic fluorite in the granite of the Log Cabin pluton, a Silver Plume-equivalent unit northwest of Fort Collins.

The 285 Mesoproterozoic pluton samples in the USGS rock database contain an average of 1,750 ppm fluorine, with a range of 100 to 170,000 ppm (table 1; figs. 3, 7). All but one of the samples average 1,116 ppm and have a median of 500 ppm fluorine; the one sample contains 17 percent (170,000 ppm) fluorine, which must indicate secondary enrichment. These values are similar to those of the Paleoproterozoic plutonic and metamorphic rocks. This is consistent with the isotopic evidence cited earlier for derivation of the Mesoproterozoic magmas from the metamorphic units. This was the second time that these rocks had been partially melted, and the restite from the 1.7-Ga melting event may have been relatively enriched in fluorine.

Despite the presence of primary magmatic fluorite in some intrusive phases, the Mesoproterozoic rocks are not, on average, significantly more enriched in fluorine than the Paleoproterozoic plutons and metamorphic rocks. The formation of magmatic fluorite may have been related to differentiation of the Mesoproterozoic magmas during emplacement, as proposed by Snyder and Hedge (1978) for the Mount Ethel plutonic complex, a process that apparently did not take place or affect fluorine in the Paleoproterozoic magmas during their emplacement. However, because the Mesoproterozoic plutons are not as deeply eroded as the older plutons, any fluorite that formed in the upper, more volatile-rich parts of the plutons in this younger suite had a better chance of being preserved.

Fluorine and Fluorite in 1.1-Giga-annum Plutonic Rocks (Pikes Peak Batholith)

Fluorite and fluorine are abundant in the Pikes Peak batholith west of Colorado Springs. Fluorite is present as a primary magmatic mineral in the granitic intrusive rocks and related pegmatites, as well as in younger, possibly Tertiary veins at Saint Peters Dome, which are discussed in the section on “Fluorine Related to Middle Tertiary Alkalic Complexes.” In addition, the intrusive rocks have high concentrations of fluorine.

The 273 samples of the Pikes Peak Granite and related rocks in the USGS rock database average 2,140 ppm fluorine, with a range of 60 to 13,300 ppm and a median of 1,700 ppm (table 1; figs. 3, 7). In addition, one sample of “rare-earth-bearing” granite contains more than 100,000 ppm fluorine. Data reported by Barker and others (1975) show that the batholith as a whole averages 3,500 ppm, and that the fluorine contents of the sodic and potassic series plutons range from less than 1,000 to about 7,000 ppm. The average and median fluorine contents of the batholith are significantly higher than those of Mesoproterozoic and Paleoproterozoic plutonic and metamorphic rocks in the region, which average about 1,100 ppm fluorine and have medians of 500 to 645 ppm.

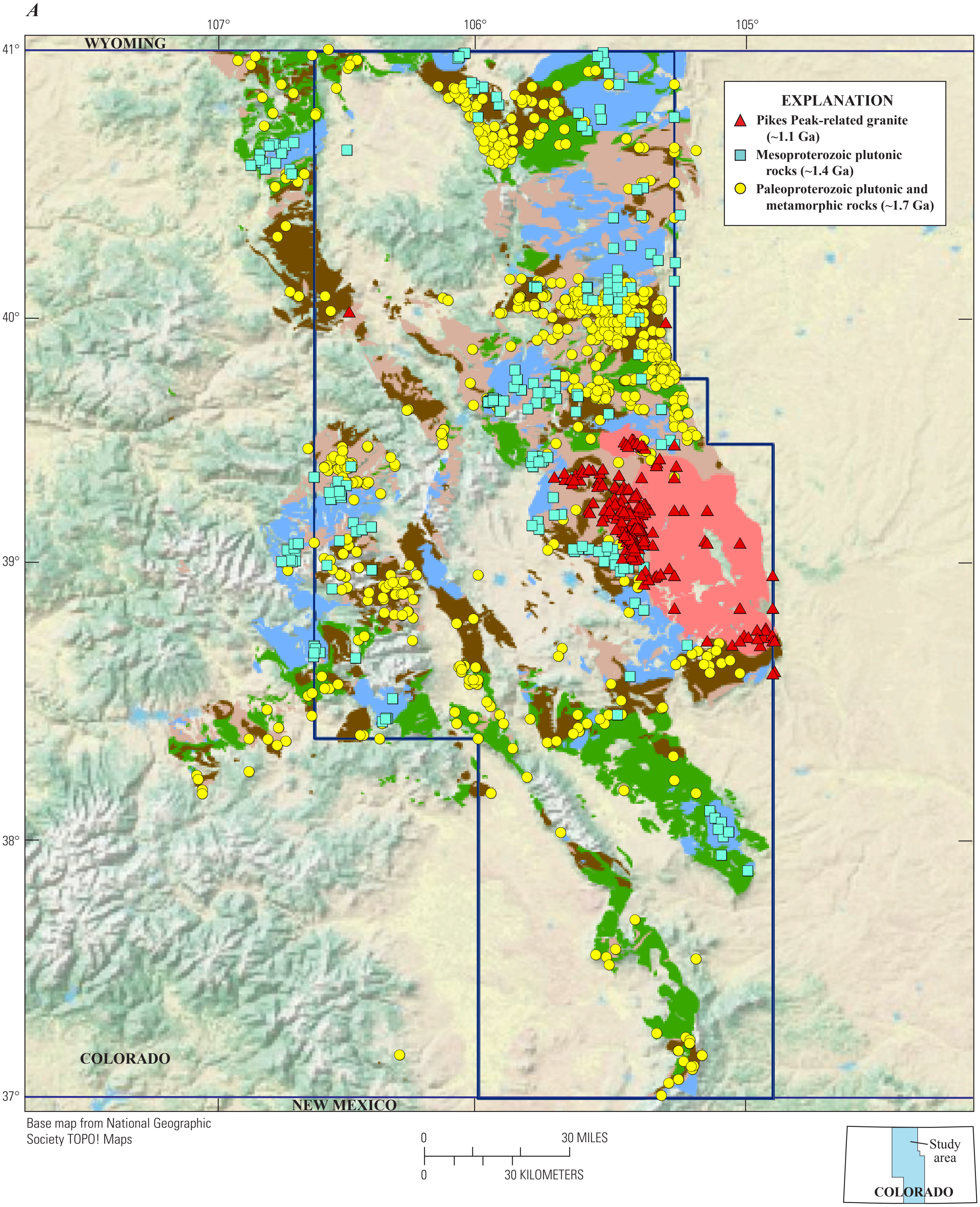
Fluorite is a common primary accessory mineral in the Pikes Peak batholith, including the main phase, the sodic and potassic plutons, and pegmatites. Fluorite in the Pikes Peak batholith is intimately intergrown with silicate minerals in the granitic rocks. Some fluorite from the Pikes Peak area is of gem quality and, as a result, has been the focus of numerous gem-related and mineralogical studies (see summaries in Eckel, 1997). The magmas likely had high fluorine contents, leading to the precipitation of fluorite in the late stages of crystallization, the formation of fluorine-rich biotite and amphibole, and alteration of mafic silicate minerals to form fluorite and fayalite (Barker and others, 1975). The presence of cryolite at some of the Proterozoic-age pegmatites at Saint Peters Dome clearly demonstrates that the Pikes Peak Granite evolved to form late-stage, fluorine- and volatile-rich hydromagmatic fluids (Dolejš and Baker, 2007).

In the Buffalo Park center, fluorite-bearing pegmatites are spatially related to the monzogranite intrusive, which may have had a mafic mantle component. In the Pikes Peak center, fluorite-bearing pegmatites are spatially related to small intrusive bodies of the Mount Rosa sodic intrusion. Fluorite-bearing pegmatites are within and near the Lake George sodic + potassic center, but it is not clear to which series they are related. Other fluorite-bearing pegmatites are scattered throughout the batholith, and at least two are spatially related to plutons of the potassic series. On the basis of fluid inclusion data from fluorite and quartz in the South Platte pegmatite district, fluorite was deposited from relatively low temperature, late-stage magmatic-hydrothermal fluids that were exsolved from the Pikes Peak Granite magma (Gagnon and others, 2004).

Fluorine in Other Proterozoic Pegmatites

Most of the pegmatites that formed during the 1.7- and 1.4-Ga igneous events are not known for high fluorine contents or fluorite, unlike the pegmatites associated with the Pikes Peak Granite. Although one sample of a 1.7-Ga pegmatite north of Berthoud Pass contained 20,200 ppm fluorine, 16 other samples averaged 400 ppm fluorine, with a median of 200 ppm. Fluorite and fluorspar have been reported and locally mined from a few 1.7-Ga pegmatites, notably in the Trout Creek Pass and Phantom Canyon pegmatite areas. Fluorite also has been reported from a pegmatite east of Salida (fig. 4).

Fluorite has been reported from only two pegmatites associated with the 1.4-Ga plutonic suite. These are located near Evergreen, possibly associated with the Silver Plume Granite, and in the northern Front Range, associated with the Sherman Granite. The USGS fluorine database contains only four analyses of 1.4-Ga pegmatites. The fluorine contents range from 80 ppm to 11,900 ppm, with intermediate values of 1,300 and 3,000 ppm. These very limited data suggest higher fluorine contents in the 1.4-Ga pegmatites than in the older ones. However, the scarcity of fluorite in the younger pegmatites is unusual given the apparent propensity of fluorine in this plutonic suite to fractionate into residual, volatile-rich phases, as described previously herein.



Base map from National Geographic Society TOPO! Maps

Figure 7. (*A*) Simplified geologic map showing the distributions of the principal age groups of Proterozoic rock units and the locations of samples of those rocks that were analyzed for fluorine. Some age-classified samples do not correspond to the same-age lithology on the base map because of the generalized nature of the map. Refer to figure 1 for general geographic locations. (*B*) Locations of high-fluorine Proterozoic samples.

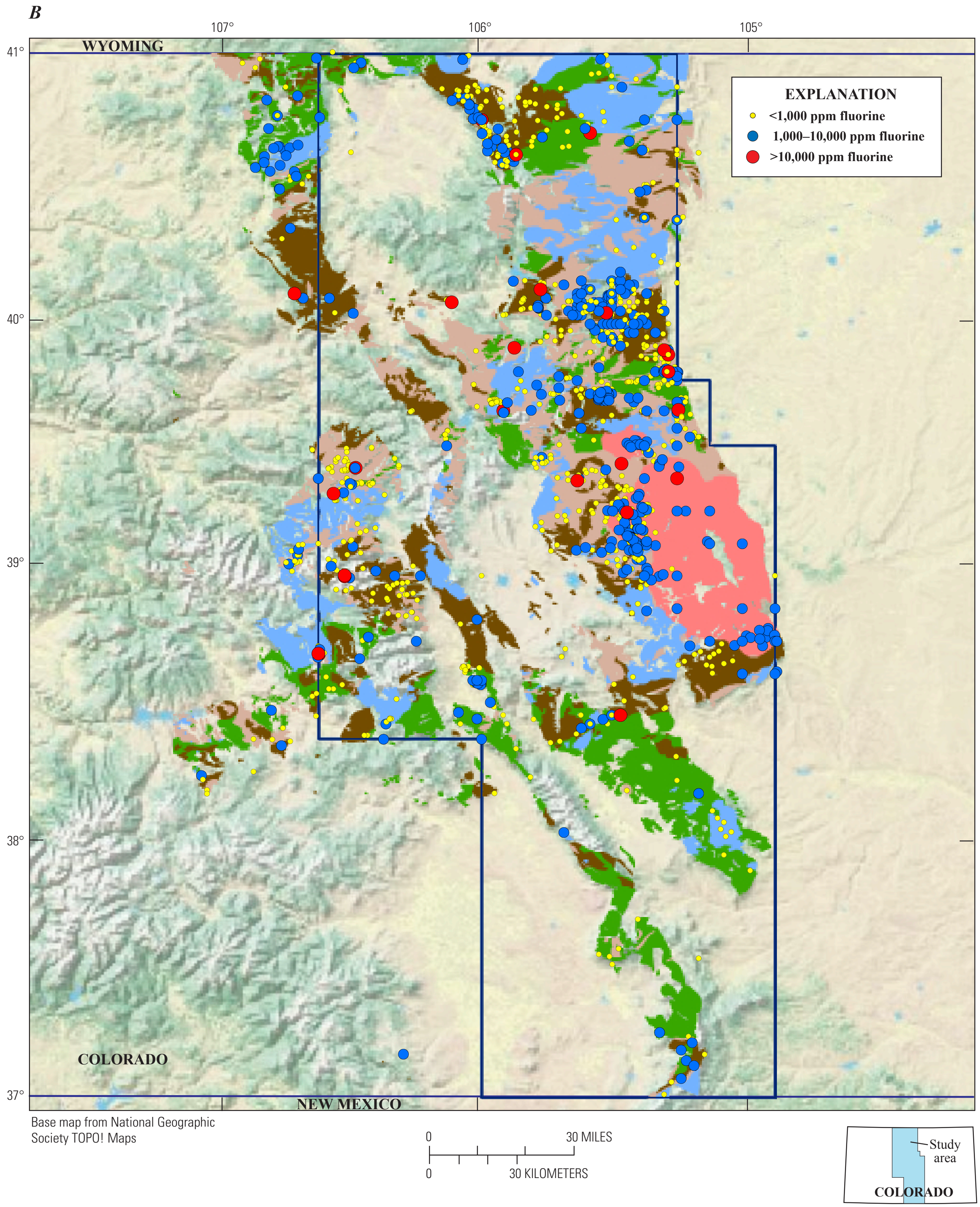


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Figure 7—Continued. (*A*) Simplified geologic map showing the distributions of the principal age groups of Proterozoic rock units and the locations of samples of those rocks that were analyzed for fluorine. Some age-classified samples do not correspond to the same-age lithology on the base map because of the generalized nature of the map. Refer to figure 1 for general geographic locations. (*B*) Locations of high-fluorine Proterozoic samples. The base is the same as in figure 7*A*.

Neoproterozoic to Early Paleozoic Alkalic Complexes

Geologic History

Igneous activity was notably sparse to absent in Colorado throughout much of the late Neoproterozoic, Paleozoic, and Mesozoic. The exception was in southern Colorado, where alkalic intrusive complexes were emplaced in the Powderhorn district south of Gunnison in the late Neoproterozoic (~570 Ma; Olson and others, 1977) and in the northern Wet Mountains north of Silver Cliff in the Cambrian (511 and 535 Ma) (fig. 8; Olson and others, 1977; Olson and Hedlund, 1981; Armbrustmacher, 1988). The complexes include carbonatites, lamprophyres, alkali syenites, pegmatites, and fenites, and they contain locally abundant thorite veins. Other Late Cambrian to Early Ordovician igneous rocks in the region include west-northwest-striking mafic dikes in southern Colorado (Hansen, 1971; Larson and others, 1985; Wallace, 2004) and alkalic complexes in New Mexico (McLemore and others, 1999).

The alkalic complexes were related to regional Cambrian extension that produced a west-northwest-trending zone of rifting that extended through southern Colorado (Larson and others, 1985; McMillan and McLemore, 1999). Alkalic igneous magmas were emplaced during rifting, with evidence for shallow emplacement and local venting to produce volcanic units (Bickford and others, 1989; McMillan and McLemore, 1999). Differential erosion has exposed these complexes at different levels of emplacement.

The principal alkalic intrusive centers in the central Colorado study area are the intrusive complexes at McClure Mountain, Gem Park, and Democrat Creek in the Wet Mountains area (Armbrustmacher, 1984). The three igneous complexes include stocks and dikes that were emplaced into Paleo- and Mesoproterozoic rocks. Compositions include mafic-ultramafic units, nepheline and quartz syenites, lamprophyres, and primary carbonatites. The McClure Mountain and Democrat Creek complexes contain the full spectrum of rock types, whereas the Gem Park complex is composed only of mafic-ultramafic rocks and carbonatite. Red syenite dikes were emplaced along synintrusive, west-northwest-striking faults, generally southeast of the three complexes. Secondary magmatic processes produced fenite, replacement carbonatite, and quartz-barite-thorite veins (Armbrustmacher, 1984). Almost all of the igneous and related rocks and veins are located between the long-lived Ilse fault to the east and the Texas Creek and Westcliffe faults to the west.

The rocks in the Wet Mountains are bimodal and have low initial strontium values (0.7028–0.7046), pointing to a mantle origin for the parent magmas. Trace-element and isotopic data suggest that the diverse intrusive rocks in the three complexes formed from three separate magma types, including alkali basalt, tholeiitic basalt, and nepheline basalt (Armbrustmacher and Hedge, 1982). This complexity is typical of carbonatite and related magmas; derivation from mantle sources during extension as well as fractional crystallization and volatile exsolution during ascent appear to be common attributes of these magmas (see summaries in Bell and others [1999] and Bell [2001]). Except for metasomatism during emplacement, interaction with the crust generally was minimal.

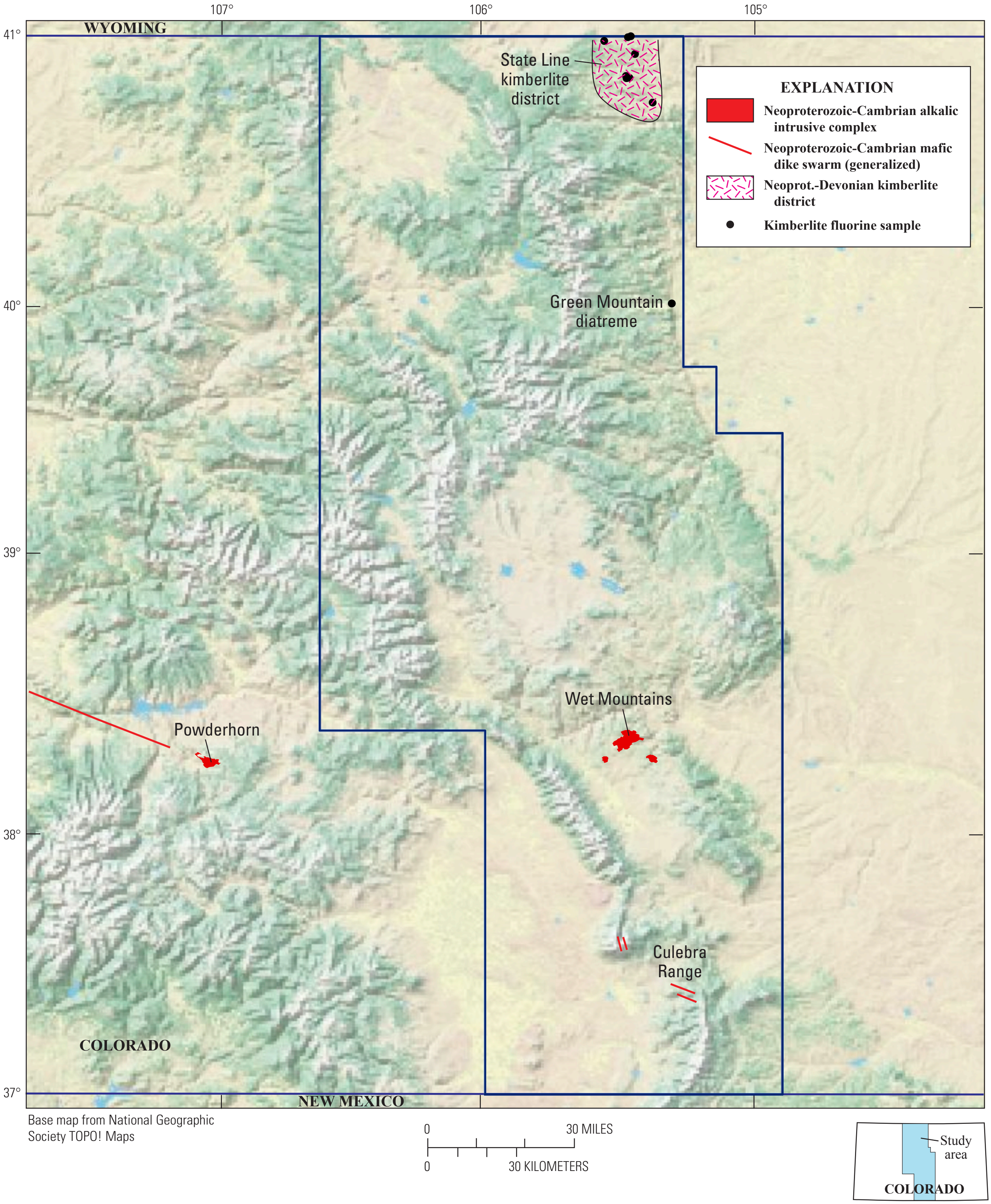
Fluorine and Fluorine-Rich Minerals

Cambrian alkalic rocks in the USGS rock database (80 samples) averaged 970 ppm fluorine, with a median of 400 ppm (table 1; fig. 3). All but 14 of these samples came from the Wet Mountains area, and those 66 rocks averaged 1,070 ppm (median 400 ppm). Twelve samples collected from the Powderhorn area averaged 420 ppm fluorine, with a median of 280 ppm. Two samples from Cambrian mafic dikes in the Culebra Range and near Blanca Peak contained 340 and 680 ppm fluorine, respectively. In the Wet Mountains area, other analyses reported by Armbrustmacher (1984) showed that the mafic intrusive rocks contained between 100 and 1,600 ppm fluorine (average of 433 ppm for 12 samples), and syenites contained 600 to 3,700 ppm fluorine (average 1,600 ppm, 14 samples).

Although fluorspar has not been mined from the alkalic complexes in the Wet Mountains, fluorine and fluorine-rich minerals are associated with the igneous rocks. Accessory fluorite is rare in most of the alkalic intrusive rocks. However, the quartz syenite of the Democrat Creek complex has accessory fluorite, and the red syenite dikes contain fluorapatite, fluorite, and epigenetic bastnaesite. Although fluorine analyses of the carbonatites are not available, these rocks contain various fluorine-bearing minerals, including fluorite, fluorapatite, cryolite, aluminofluoride minerals, bastnaesite, pyrochlore, and synchysite ($([Ce, La]Ca[Co_3]_2F)$) (Armbrustmacher, 1988). Heinrich (1977) also identified a complex suite of fluorine-bearing minerals from the Goldie carbonatite dike in the McClure Mountain complex.

The quartz-barite-thorite veins are the youngest products of the intrusive complexes. They formed, along with red syenite dikes, along west-northwest-striking faults and fractures in all

Figure 8 (facing page). Locations of Neoproterozoic to Cambrian alkalic complexes (red areas) and related mafic dike swarms (red lines), and of Neoproterozoic to Devonian kimberlite diatremes. Sampling sites for the southern Colorado alkalic complexes are not shown because the locations provided were too generalized (see the section on fluorine data from central Colorado for an explanation). The locations of samples of the kimberlites that were analyzed for fluorine are shown with black dots. The kimberlite diatremes are too small to show at this scale. The State Line District is generalized from McCallum and others (1975). Geology simplified from Tweto (1979a), with mafic dike locations from Larson and others (1985) and Wallace (2004). Refer to figure 1 for general geographic locations.



Base map from National Geographic Society TOPO! Maps

older rocks, including red syenite dikes along the same fractures. The veins are distributed over a large area, from north of the Arkansas River to southeast of Rosita. Fluorine minerals in the veins include fluorapatite and synchysite. Fluid inclusions in smoky quartz in one vein indicated two populations of inclusions. The water-rich inclusions had homogenization temperatures of about 125°C and salinities of about 12 weight percent sodium chloride (NaCl) equivalent, and coexisting carbon dioxide (CO_2)-rich inclusions may have formed above 200°C. The combined fluid inclusion information indicated probable CO_2 effervescence that led to precipitation of the vein minerals (Armbrustmacher, 1988).

The work by Armbrustmacher (1979, 1984, 1988) and Armbrustmacher and Hedge (1982) showed that all fluorine and fluorine-bearing minerals were products of the igneous systems. On the basis of the likely mantle derivation of the parent magmas and the lack of significant crustal contamination, the fluorine was derived originally from the mantle and then fractionated as a volatile phase into the more felsic, later magmas and the late-stage veins. Although a large percentage of the igneous rocks are concentrated in the three intrusive centers, the wide distribution of the fluorine-rich red syenite dikes and quartz-barite-thorite veins indicates that fluorine is also widely distributed through the northern Wet Mountains between the Ilse and Westcliffe–Texas Creek fault zones.

Neoproterozoic to Early Paleozoic Diatremes

About 20 small kimberlite diatremes intruded Proterozoic rocks of various ages in the northern Front Range and the southern Laramie Mountains near the Colorado-Wyoming border (fig. 8). Two additional diatremes—the Green Mountain diatreme just west of Boulder, and the Estes Park diatreme near Estes Park—are south of the main cluster of diatremes. On the basis of clasts of Cambrian, Silurian, and Devonian sedimentary rocks in some of the diatremes, as well as a few fission-track and rubidium-strontium (Rb-Sr) dates, the age of emplacement for all of the diatremes was assumed to be late Paleozoic. More recent $^{40}Ar/^{39}Ar$ and samarium-neodymium (Sm-Nd) dates show that at least two of the diatremes were emplaced in the late Neoproterozoic and that the diatremes undoubtedly were emplaced during more than one intrusive event (Lester and others, 2001).

The kimberlites contain primary olivine, enstatite, perovskite, and a variety of other less-common minerals, including phlogopite (McCallum and others, 1975). Xenoliths and xenocrysts derived from the lower crust and mantle are common in some of the diatremes. The diatremes near the border between Colorado and Wyoming (the Stateline District) contain abundant small diamonds, which led to some production in 1980s and 1990s. The kimberlites have mantle-derived affinities and thus, in part, may be related to the igneous events that produced the alkalic complexes in southern Colorado (McCallum and others, 1975; Lester and others, 2001). Uranium-lead (U-Pb) dating of zircons from the xenoliths produced 1.7- to 1.4-Ga ages, similar to the ages of the crust in that region (Abbasi and others, 1998). The tops of the diatremes were removed during the formation of the Neogene erosion surface (see later discussion in section on “Weathering of Fluorine-Bearing Rocks”).

The primary igneous rocks were weakly to strongly altered, probably by late-magmatic fluids, producing serpentine minerals, carbonates, talc, montmorillonite, and chlorite. Fluorite has not been reported in the published literature. However, two samples in the USGS fluorine database were described as having “hi [high] fluorite,” and one of these also was described as “greisen,” implying that the fluorite formed during late-stage alteration. Thirty-five samples in the USGS rock database that were collected from the Stateline District (table 1; fig. 8) averaged 2,070 ppm fluorine, with a median of 1,500 ppm and a range of 20 to 8,600 ppm. While these values, even the median, are high compared to average crustal and regional abundances, the diatremes and their contained fluorine are very local.

Paleozoic and Mesozoic Sedimentary Rocks

Geologic History

Sedimentary rocks were deposited across all or parts of central Colorado from the Cambrian to the present (fig. 9). Erosion has removed these units from most, but not all, of the mountain ranges, and extensive Tertiary and Quaternary sedimentary and volcanic units conceal older sedimentary units in many areas. From the Cambrian through the Mississippian, central Colorado was covered by a series of transgressing and regressing shelf and shelf-margin environments above the craton. The resulting sedimentary rocks included limestone, dolomite, sandstone, and quartzite. Orogenic uplift in the Pennsylvanian produced the northwest-trending Uncompahgre uplift in southwestern Colorado, the Ancestral Front Range in central to northern Colorado, and the smaller Sawatch uplift in central Colorado. Alluvial fans adjacent to these uplifts carried debris into nearby basins, and the fans interfingered with marine environments that fringed some of the uplifted areas.

Mesozoic sedimentary rocks reflect nonmarine to marine sedimentation as orogenic activity waned and inland seas episodically invaded the area. Late orogenic clastic sedimentation continued through the Triassic as erosion beveled the highlands and sediments were deposited in various marginal marine, braided stream, and lacustrine environments. Jurassic and Cretaceous sediments were deposited during several nonmarine to marine sedimentary cycles separated by periods of epeirogenic uplift and subsequent erosion. Incipient Late Cretaceous uplift related to the Sevier and Laramide orogenies created a northeastward retreat of the sea and attendant shift from marine to near-shore and continental sedimentation.

Fluorine in Paleozoic and Mesozoic Sedimentary Rocks

Fluorite unrelated to hydrothermal mineralization has not been reported from Phanerozoic sedimentary rocks in the study area. Except for a few locally intense studies, very few sedimentary rocks in central Colorado have been analyzed for fluorine. However, many of the original sediments—and their contained fluorine—were derived from bedrock sources in Colorado that have been analyzed for fluorine. As such, weathering and erosion of those source rocks released fluorine, either as a dissolved element or as clastic grains, which ultimately were transported toward and possibly into sedimentary basins.

The many steps between the source rocks and the site of sedimentation, such as weathering, transport, water chemistry, and deposition, affect the ultimate distribution and occurrence of fluorine in the sedimentary units. The mode of occurrence of fluorine in the source rock also has an important effect on how the fluorine is transported. For example, fluorine that occurs as a trace element in hornblende or biotite can be released by source-area weathering and transported in surface water, or the entire mineral grain can be transported along with its constituent fluorine. As a result, sedimentary units that contain abundant fluorine-bearing minerals, such as biotite, can contain greater amounts of fluorine than do sediments composed largely of fluorine-poor minerals, such as quartz and feldspar. Therefore, a micaceous shale or mudstone may be more fluorine rich than a quartz arenite or quartzite.

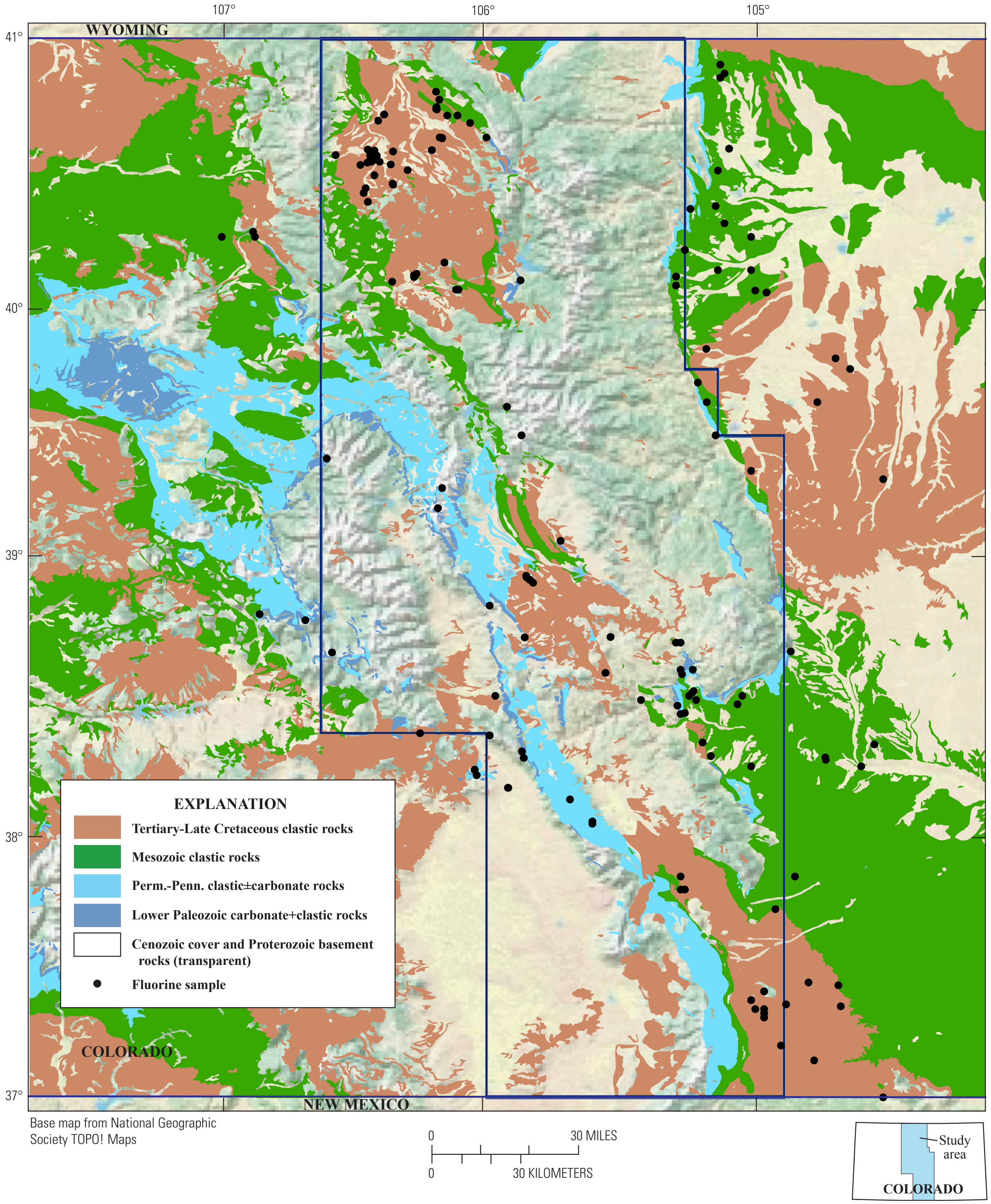
The average fluorine content of all 587 Phanerozoic sedimentary rocks in the USGS rock database was 1,990 ppm. One drill core in the Cretaceous Niobrara Formation produced 138 of these samples, which had an average fluorine content of 730 ppm fluorine. In addition, two samples contained 110,000 and 240,000 ppm fluorine, which could indicate contamination or unreported secondary fluorite. By removing those two samples and using the average value for the drill core to remove the bias from that single site, the remaining 448 samples had an average of 1,600 ppm fluorine and median of 560 ppm (table 1; fig. 9).

Limestone and dolomite samples contained an average of 1,670 ppm fluorine and a median of 900 ppm, which is significantly higher than the global average for carbonates of 330 ppm (Turekian and Wedepohl, 1961). The carbonate samples were collected from lower Paleozoic sedimentary units and Cretaceous limestones. Fluorine in limestones typically occurs in fluorapatite and detrital clay minerals. Most of the fluorine in marine carbonates is precipitated directly from seawater and thus does not reflect the fluorine contents of rocks exposed in nearby areas. However, the relatively high fluorine content of the carbonates in Colorado may indicate that the samples contain more detrital clay than was noted in the abbreviated sample descriptions, thereby elevating the fluorine content.

Most of the sandstone and quartzite samples were collected from lower Paleozoic formations in west-central, central, and east-central parts of the study area. With five very anomalous exceptions, 89 samples contained an average of 670 ppm fluorine, with a median of 400 ppm. This is higher than the global average for sandstones of 270 ppm fluorine (Turekian and Wedepohl, 1961). Two samples of Ordovician quartzite contained 2,100 ppm fluorine. These samples were collected from southwest of Pikes Peak and southeast of Salida in areas where quartzite samples with more normal fluorine contents were collected, suggesting local epigenetic introduction of fluorine. The detrital grains in the clastic sedimentary rocks were derived in part from nearby highlands, and some originated from much more distant sources and were deposited in regional shoreline sand facies. Thus, more detailed information on the actual unit sampled and its depositional setting is required before a correlation between the fluorine contents in source areas and the nearby sedimentary units can be made. Most of the fluorine in the sandstones is likely contained in detrital biotite, amphiboles, and apatite.

Most of the shale (including mudstone and claystone) samples were collected from Cretaceous and Paleocene sedimentary units, including the Pierre Shale, in North Park, near Boulder, and east of the Wet Mountains and Culebra Range. Other samples were collected from various Lower Paleozoic formations in the central and east-central parts of the study area. With the exception of two very anomalous samples, these shale samples contained an average of 1,130 ppm fluorine, with a median of 500 ppm. Five of the Cretaceous and Paleogene shale samples contained coal, based on the brief sample-submittal information, and those five samples contained an average of 420 ppm fluorine, with a range of 350 to 500 ppm. The fine-grained detrital sediments in shale units, especially the Pierre Shale, were derived from sources well outside of the study area, thus precluding a correlation with bedrock units in Colorado. Coal-bearing units, such as in the Upper Cretaceous and Paleocene Raton and Upper Cretaceous Vermejo Formations, are more locally derived, and the contained fluorine may have been derived from nearby uplifts or possibly from residual brackish ocean water.

The Upper Cretaceous Mancos Shale in western Colorado is roughly correlative with the Upper Cretaceous Pierre Shale, which is exposed in various parts of the study area. Analyses of drill core of the Mancos Shale collected from just west of the central Colorado study area produced fluorine values ranging from 500 to 1,540 ppm, with a median of 867 ppm (Richard Grauch, U.S. Geological Survey, written commun., 2006). Soil samples collected from above the Mancos Shale in the same area contained between 75 and 7,500 ppm fluorine, with an average of 780 ppm (no median given; Tuttle and others, 2007). Given the similarities in depositional environments and ages, these fluorine values in the Mancos could be used to infer the values in the Pierre Shale, analyses for which are not present in the USGS rock database.



Base map from National Geographic Society TOPO! Maps

Figure 9. Distribution of pre-Quaternary Phanerozoic sedimentary rocks in central Colorado and the locations of samples of these rocks that were analyzed for fluorine. Geology simplified from Tweto (1979a). Refer to figure 1 for general geographic locations.

Cenozoic Igneous and Sedimentary Rocks and Mineral Deposits

Cenozoic Geologic History

The magmatic and tectonic setting of central Colorado is described in somewhat more detail here due to the fact that most of the fluorspar deposits formed during this period of time, and most are related in some fashion to the magmatic history. The distributions of Laramide and Tertiary plutonic and volcanic rocks are shown in figure 10*A*.

Tectonism in central Colorado related to the regional Laramide orogeny began in the Late Cretaceous. The orogeny was a response to subduction of the Farallon plate beneath North America, and it produced generally east-northeast-directed compression that, in Colorado, involved basement rocks. This compression reactivated older faults, such as those along the northeast-trending Homestake and Idaho Springs–Ralston shear zones (Tweto and Sims, 1963; Caine and others, 2006), and created new structures. East- and west-vergent thrust faulting generated a number of uplifts, such as the Front, Park, and Sawatch Ranges and the Sangre de Cristo Mountains (fig. 1), with some ranges forming through more vertical uplift, such as the Gore Range. Middle Eocene orogenic sedimentary rocks in the Denver and Raton basins indicate that Laramide tectonism continued to that time, although its effects were waning (Dickinson and others, 1988).

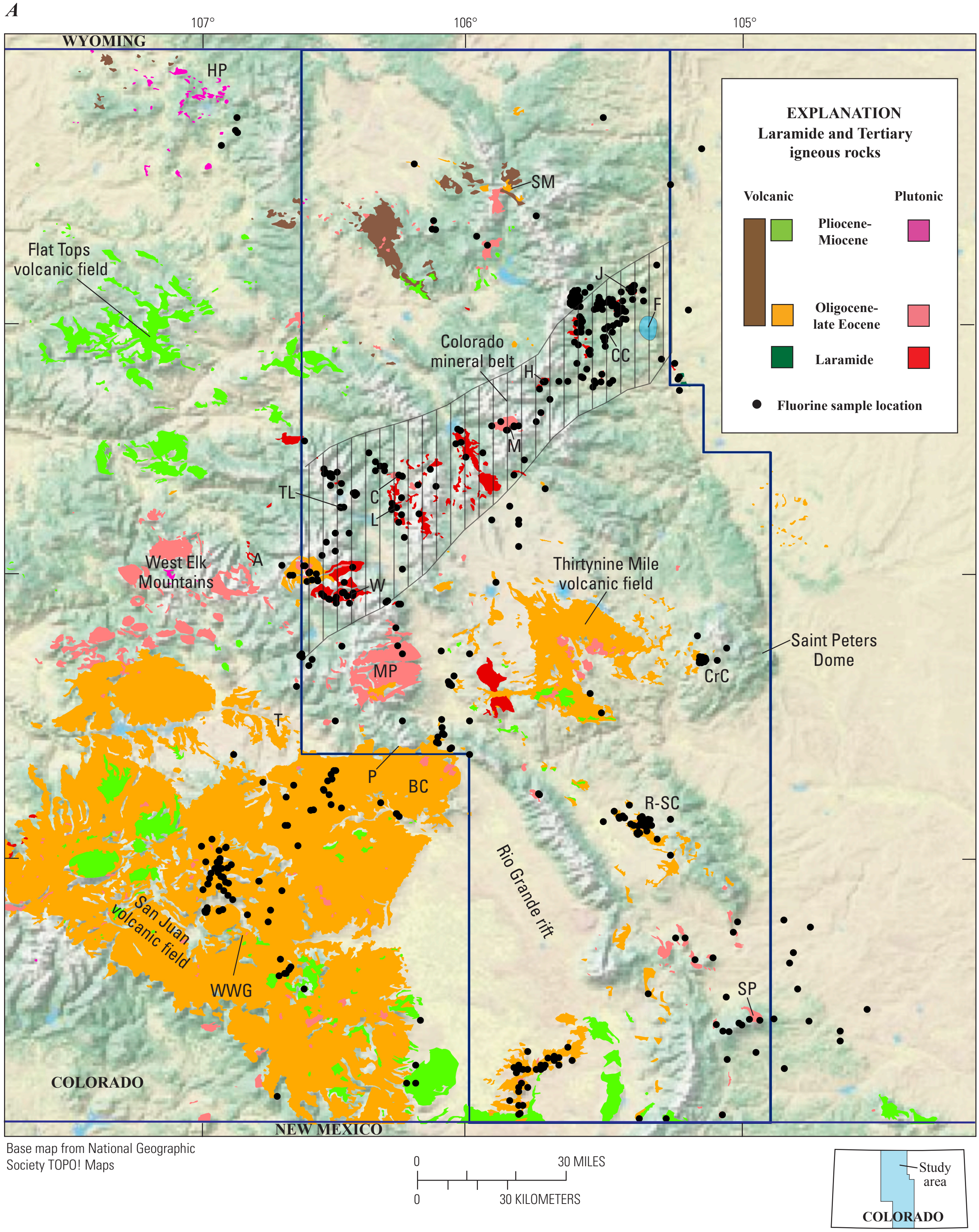
The shallow-dipping subducted slab that extended beneath the region induced magmatism in central Colorado starting at 72–70 Ma (Lipman, 1981). In general, magmatism progressed rapidly in a northeast direction and then, as the slab began to fail and subside, began to retreat to the southwest and south starting at about 45 to 40 Ma (Bookstrom, 1990; Humphreys, 1995). Laramide intrusive complexes commonly were calc-alkalic with local, more alkali-rich phases, such as in the Central City and Jamestown districts (fig. 10*A*; Rice and others, 1985; Bookstrom, 1990). The magmas formed batholiths, stocks, sills, and dikes in host rocks of Proterozoic to Cretaceous age. Some of these magmas vented to the surface, but Eocene erosion removed any evidence of volcanic units except for a few 65.5-Ma shoshonite lavas west of Denver and some volcaniclastic materials in adjacent coeval basins (North Park: Hail, 1968; Denver Basin: Wilson, 2002).

At a regional scale, Laramide magmatic activity in central Colorado was restricted to a broad, northeast-trending zone that was roughly coincident with the Homestake and Idaho Springs–Ralston shear zones (Tweto and Sims, 1963). Those Proterozoic fractures in part served to localize magmas as they ascended through the thickened continental crust, permitting their emplacement at subvolcanic depths. The localization of magmatism along those fractures may have been due to their weakly oblique orientation to the compression direction, which would have caused dilation, as compared to Proterozoic fractures that were strongly oblique to or perpendicular to the compression direction and thus did not permit magma ascent at a crustal scale. Caine and others (2006) documented this process on a district scale in the Front Range.

Starting in the middle Eocene, tectonic and magmatic activity waned as compression decreased and the subducted plate began to collapse (Lipman, 1981; Humphreys, 1995). With the cessation of uplift, the ensuing erosion of the mountainous terrain produced a widespread, relatively gentle late Eocene erosion surface (Epis and Chapin, 1975; Kelley and Chapin, 2004). The decreased compression, coupled with asthenospheric heating of the lower crust and possible weak, local extension in the wake of the collapsing slab (Lipman, 1981; Mutschler and others, 1988; Humphreys, 1995), triggered renewed calc-alkaline magmatism and the formation of andesitic stratovolcanoes and more silicic calderas. The largest concentration of these volcanoes was in the San Juan Volcanic Field of southwestern Colorado, but several volcanic centers were centered in the Bonanza, Marshall Creek, Mount Aetna, Mount Princeton, Grizzly Peak, and Thirtynine Mile areas in central Colorado (fig. 10*A*; the Central Colorado volcanic field of McIntosh and Chapin, 2004). The widespread, low-relief Eocene erosion surface allowed outflow tuff sheets to travel great distances from their source calderas. In addition, other magmas were emplaced at shallow levels, commonly in Paleozoic sedimentary rocks, to form plutons, sills, and dikes throughout central Colorado, including those in the Leadville, Gilman, and Breckenridge mining districts. Erosion removed many volcanic edifices, although Oligocene intracaldera units, outflow sheets, and flows are variably preserved in the Sawatch Range (Varga and Smith, 1984; Shannon, 1988; Fridrich and others, 1991), the Thirtynine Mile volcanic field (Epis and Chapin, 1975), the Wet Mountains and Silver Cliff area, and in the West Elk and San Juan Mountains to the west (fig. 10*A*).

Figure 10 (pages 26–27). (*A*) Distribution of Laramide and Tertiary plutonic and volcanic rocks in central Colorado. The black dots indicate the locations of samples of these rocks that were analyzed for fluorine; in some areas, more than one sample was collected at the same site. The Colorado mineral belt is shown with the vertical lined pattern; the Rio Grande rift and areas with related late Cenozoic extensional faults are shown with the stippled pattern. Major labeled volcanic fields include the Flat Tops, San Juan, and Thirtynine-Mile volcanic fields. Abbreviations for mines, districts, and other features: A, Aspen; BC, Bonanza caldera; C, Climax; CC, Central City; CrC, Cripple Creek; fluorine, fluorine-quartz cement in breccia reefs (blue oval area); H, Henderson; HP, Hahn's Peak; J, Jamestown; L, Leadville; M, Montezuma; MP, Mount Princeton (and general location of Mount Antero); P, Pitch uranium mine; R–SC, Rosita–Silver Cliff; SM, Specimen Mountain: SP, Spanish Peaks; T, Tomichi Dome; TL, Turquoise Lake; W, Winfield; WWG, Wagon Wheel Gap. Geology simplified from Tweto (1979a), with additional information from Tweto and Sims (1963) and Tweto (1979b). Refer to figure 1 for general geographic locations. (*B*) Locations of Laramide and Tertiary ore samples with greater than 1,000 parts per million fluorine. The sampling sites and the geologic base are the same as in figure 10*A*.

A



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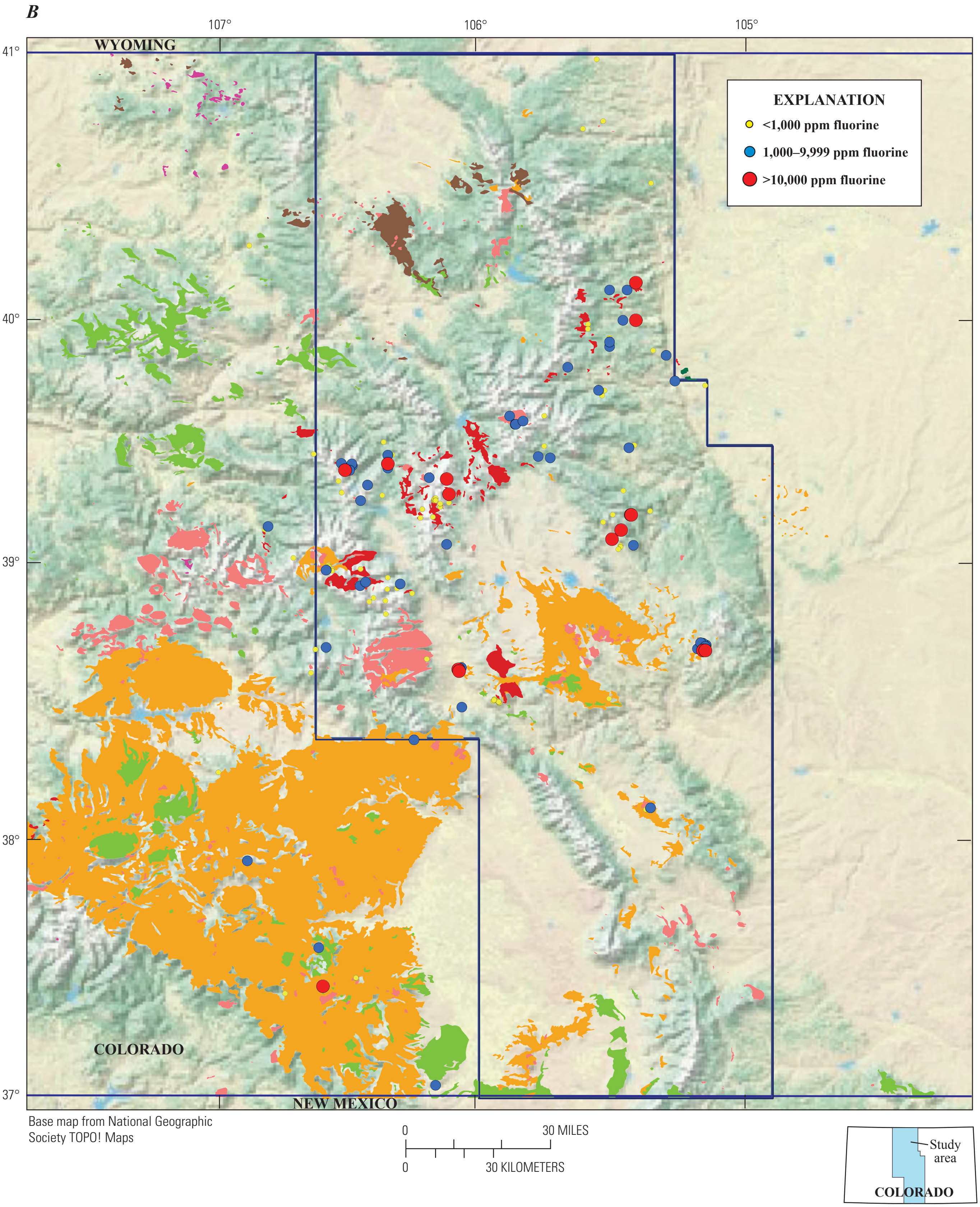

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At 35 to 30 million years ago (mya), crustal relaxation gave way to incipient extension, and the north-trending Rio Grande rift began to form. The rift evolved during two stages: an early period of modest extension, propagating northward from New Mexico and forming broad sag basins, and subsequent increased extension and the creation of a series of en echelon grabens and related horsts that extends from southern New Mexico into north-central Colorado (fig. 10*A*; Tweto, 1979b). Many of the steep mountains near the rift, including the Sangre de Cristo Mountains, Sawatch Range, Gore Range, and parts of the Park Range, formed as a result of rift-margin uplift during the younger stage of rifting. Rift-related, high-angle normal faulting also cut the 33 Ma Ceresco molybdenite orebody at Climax and displaced it downward to the west by more than 2,000 m (Wallace and others, 1968). Fault activity along the rift has continued to the present day (Colman, 1985), and modern hot springs are common along some frontal fault zones (see section on Modern Hot Springs).

Farther from the rift itself, coeval faulting was asymmetric. Areas west of the rift axis experienced little to no Neogene faulting. Much of the geomorphic development in that area has been erosional, with streams draining westward from the topographic high created by rift-margin uplift. However, limited Climax-type plutonism and mineralization did migrate to the west, from the 17-Ma Mt. Emmons system north of Gunnison (see fig. 2) to the 4.5-Ma Rico system in the western San Juan Mountains. These systems typically formed in extensional environments, which suggests that the crust to the west of the rift did undergo some extension that did not produce notable coeval faulting. In contrast, Miocene high-angle faulting was relatively common east of the rift (Taylor, 1975; Colman, 1985; Matthews, 2003). Postmineralization, down-to-the-west normal faults formed near the Henderson molybdenum deposit (White and others, 1981), and the Proterozoic Ilse fault in the Wet Mountains was reactivated during the Miocene (Taylor, 1975). However, this faulting did not significantly disrupt eastward drainage. Streams continued to flow to the east and produced a broad Miocene to Pliocene erosion level that was incised into the Eocene erosion surface; incision progressively deepened to form steep canyons (Scott, 1975a). These erosional processes were related to a combination of rift-related uplift in the headwaters, climate, isostatic uplift related to erosion, and far-field lowering of the base level (Leonard, 2002; Kelley and Chapin, 2004).

With the Laramide slab collapsing to the southwest during a period of crustal relaxation and rift-related extension propagating northward, the middle Tertiary magmatic transition between these two crustal events varied and, in some locations, overlapped in both time and space. Two significant magmatic events took place during this slightly pre- to early-extensional period: alkalic igneous activity and rhyolite-dominated Climax-type plutonism. The alkalic magmatism was focused in a belt along the eastern Front Range, the western Wet Mountains, and the eastern side of the southern Sangre de Cristo Mountains. The magmas were moderately to strongly alkalic and produced the shallow intrusive and eruptive centers at Cripple Creek, Rosita, Silver Cliff, and the Spanish Peaks (Sharp, 1978; McLemore, 1995; Kelley and Ludington, 2002). These magmas were generated when post-subduction upper mantle materials, which may have been relatively enriched in fluorine relative to chlorine during earlier slab devolatilization (Straub and Layne, 2003), ascended into and partially melted the lower crust, producing oxidized, alkaline magmas (Kelley and others, 1998; Kelley and Ludington, 2002; Richards, 2009).

The Climax-type magmatism produced a bimodal suite of high-silica, alkali-rich granites with minor amounts of lamprophyres (Bookstrom and others, 1988) that range in age from about 29 to 12 Ma. Like the alkalic plutons, they formed slightly before or at the onset of extension. Extension in the lower crust, perhaps enhanced by mafic mantle underplating of the crust, can induce decompression melting and the formation of low-viscosity fluorine-rich magmas (Orozco-Esquivel and others, 2002). Plutons that formed from these magmas typically are associated with Climax-type molybdenum deposits or prospects, such as at Climax, Henderson, and other locations in central Colorado. Climax-type magmatism took place over a broader area, however, extending south into northern New Mexico (Questa) and west into the San Juan Mountains (Chicago Basin, Rico). Trace-element and isotopic data indicate a lower crustal, possibly granulitic, source for the high-silica granites and the molybdenum and tungsten that are present in related mineral deposits (Westra and Keith, 1981; White and others, 1981; Bookstrom and others, 1988). Volatiles such as fluorine may have been derived from either or both the mantle or the crust. The presence of magmatic fluorite in some lamprophyre dikes, magmas for which originated in the lithospheric upper mantle (Bookstrom and others, 1988), argues for at least some mantle contributions of fluorine. In the Silver Cliff–Rosita area, the emplacement of highly evolved, Climax-like rhyolites alternated with more intermediate-composition to weakly alkalic magmatism, indicating complex mechanisms of magma generation, ascent, and emplacement of these two magma series without mutual contamination. Similarly, in the Latir volcanic field, which includes the Questa molybdenum deposit in northernmost New Mexico, high-silica rhyolite magmas derived from the crust slightly predated alkalic magmatism that was the product of combined mantle input and crustal assimilation (Johnson and Lipman, 1988).

The third and latest stage of magmatism began in the late Oligocene (about 26 mya) and produced a mafic-dominated bimodal suite that is more typical of a rift environment. Most of the rocks that were erupted during this period were basaltic, with much smaller amounts of rhyolite. They change in composition with distance from the rift axis, from tholeiitic basalts along the rift to more alkalic rocks farther from the rift axis (Lipman and Mehnert, 1975). With time, most of the volcanic activity was mafic, producing extensive basalt flows in the Flat Tops region (fig. 10*A*) and more local flows in many other places. The basalts had a predominantly asthenospheric to lithospheric mantle source (Stein, 1985; Leat and others, 1988; Johnson and Thompson, 1991).

The processes that formed the Cenozoic magmas, and thus the mineral deposits associated with them, were extremely complex and reflected the tectonic regime, the effects of slab advance and retreat, the degree of partial melting of the lower crust, the composition of the crust itself, the amount of mantle contributions to the magmas, and undoubtedly other processes that have yet to be recognized. These variables ultimately affected the compositions of the magmas and therefore the amount of fluorine that was carried along with them to more shallow levels. Some trace-element and isotopic data indicate that Laramide magmas were derived from a highly metasomatized mafic lower crust, whereas the Oligocene magmas were derived largely from a granulitic, intermediate-composition to felsic crust (Stein, 1985; Bookstrom and others, 1988). However, significantly more geochemical and geochronological work is required to arrive at a full understanding of what magmas formed, where, and why.

Quaternary climatic cooling induced glaciation from about 500,000 years ago into the Holocene. During three glacial maximums, ice blanketed the higher parts of the ranges and filled the valleys. The modern alpine topography with extensive morainal deposits and deep U-shaped valleys is largely a product of glacial erosion. As discussed in the section on "Weathering of Fluorine-Bearing Rocks," this has produced many areas with a juvenile weathering profile, resulting in different degrees of fluorine mobility in the surficial environment.

Cenozoic Mineral Deposits

With few exceptions, the major ore deposits of central Colorado are the products of the Laramide through middle Oligocene magmatic activity related to crustal compression and noncompression, or of the Oligocene and younger magmatism and high heat flow related to extension. As a result, the northeast-trending Colorado mineral belt (fig. 10*A*), as defined by Tweto and Sims (1963), has a pronounced bulge in its central region where these northeast- and north-trending magmatic and tectonic systems intersect. Shawe (1976) noted that a number of mineral deposits and Laramide intrusive rocks have northwest-oriented trends as well, perhaps reflecting structural controls by Laramide and older faults. A good example of this is the northwest-trending belt of plutons and related mineral deposits in the northern Sawatch Range (Wallace and others, 1989). The mineral deposits in central Colorado are widely varied in their metal contents, source rocks, ages, host rocks, and styles of mineralization, but for the most part, they and their associated igneous rocks share a genetic affinity to major structural flaws in the continental crust. A number of deposits, notably Climax-type molybdenum deposits and fluorspar deposits, have high concentrations of fluorine (fig. 10*B*). The fluorine contents of most deposits, however, are not anomalous (see table 1 and the discussion in the "Fluorspar Deposit Database section").

Most of the Laramide mineral deposits are vein-dominated base- and precious-metal deposits that were emplaced in the Proterozoic basement, and they are associated with Laramide stocks and dikes. The major deposits are concentrated along the Colorado mineral belt in the Front Range. Major districts include Central City, Idaho Springs, Georgetown, Empire, Jamestown, and various districts and deposits in Boulder County and the northern Sawatch Range.

The late Eocene and Oligocene mineral deposits are widespread and, in the cases of Leadville and Gilman, some of the largest deposits of their type in the world. The deposits include polymetallic veins associated with epizonal plutons (Turquoise Lake, Montezuma/Geneva Basin area), pluton-related manto and vein deposits in Paleozoic carbonate units (Leadville, Gilman), caldera-associated veins (Bonanza), and veins and breccias related to more alkalic or mixed-composition plutonic suites (Cripple Creek, Silver Cliff/Rosita) (figs. 10, 11).

Mineral deposits associated with the late Oligocene and younger early-extension to rift-related systems extend beyond the mineral belt to the north (the Hahn's Peak molybdenum deposits and Northgate fluorspar district) and south (the Browns Canyon fluorspar district, and the Questa molybdenum deposit in northernmost New Mexico). Within the central part of the mineral belt, the major deposits are the enormous Climax-type molybdenum orebodies at Climax and Henderson, which are associated with highly evolved granites.

Some mineral deposits, including the carbonate-hosted ores at Aspen and those at the Sherman mine in the Leadville district, permissively may have formed during the late Paleozoic, Laramide, or middle Tertiary. The Aspen ores have characteristics similar to both Paleozoic and Oligocene deposits, but, due to the absence of clear crosscutting relationships with dated igneous rocks, the age of mineralization is unknown (Stegen and others, 1990).

Fluorine in Cenozoic Sedimentary Rocks

Most of the materials in the Cenozoic sedimentary rocks were derived from erosion of surrounding highlands during various stages of tectonism. Thus, the fluorine contents are a mixture of whatever rocks were in the source areas. None of these rocks contain fluorite or fluorspar that is not related to postsedimentation hydrothermal processes.

Samples in the USGS fluorine database came from the Paleocene and Eocene Coalmont Formation, the Eocene Huerfano and Cuchara Formations, the Miocene Troublesome Formation, and Miocene tephras (volcanic ash) (**Appendix** [click on the link to access the Excel file]; fig. 9). Most of the samples of the Paleocene and Eocene units and the Troublesome are size fractions and thus may not represent the fluorine contents of the original rocks. The 42 samples of Paleocene and Eocene sedimentary materials averaged 710 ppm fluorine, with a range of 100 to 2,500 ppm and a median of 400 ppm.

Fluorine contents of 21 samples of the Troublesome Formation averaged 22,050 ppm, with a range of 200 to 240,000 and a median of about 250 ppm. Six of these samples contained more than 10,000 ppm fluorine, and two of these were carbonaceous, suggesting chelation of fluorine by carbon.

Some of the Miocene basins contain minor to major amounts of air-fall tephra that were derived from major eruptions hundreds to more than a thousand miles away. These typically formed thin beds within otherwise epiclastic fluvial to alluvial strata in the basins, and they comprise a small percentage of the sediments in each basin. Although these beds are volumetrically small, their ages can be used to determine the age and rate of sedimentation, in the basins (Izett, 1975; Hubbard and others, 2001; Perkins and Nash, 2002). The 77 samples of tephra in the USGS rock database averaged 830 ppm fluorine, with a range of 300 to 3,400 ppm and a median of 600 ppm (table 1). Most of these were collected from the Troublesome Formation in the North Park region. The composition of the erupted magma determined the fluorine content of the resultant tephra and thus was not a function of local source rocks or processes.

Fluorine Related to Laramide Igneous Centers

Early Laramide igneous activity began at about 74 Ma and continued to about 45 Ma. Most of the Laramide plutons were calc-alkalic, although some of the plutons emplaced in the Central City, Ward, Eldora, and Jamestown districts were at least mildly alkalic (figs. 10, 11; Bookstrom, 1990; Kelley and Ludington, 2002). Despite the abundance of early Laramide (74–60 Ma) plutons in central Colorado, fluorite of this age is uncommon with the exception of the large fluorspar deposits in the Jamestown district, described herein. Only two other MRDS records describe fluorite of Laramide age, and both are in the central Front Range. Fluorite was deposited in a gold-telluride vein during intrusion of the late-stage, more alkaline parts of the Central City intrusive complex (Sims and others, 1963; Rice and others, 1985; Wallace, 1989a). This stage of igneous activity was dated at about 60 Ma (Rice and others, 1982, 1985). Fluorite also was deposited in an isolated vein and an igneous breccia west of the Eldora district, possibly related to emplacement of the 58-Ma Eldora stock.

In addition, two small veins just south of Evergreen in the Front Range contain fluorite. Their ages are unknown and they are not associated with any Laramide or Tertiary intrusive unit. One of the veins was mined for copper.

A large number of Laramide intrusive rocks were analyzed for fluorine. As described herein, sample descriptions for samples collected in the Jamestown district were insufficient to differentiate mineralized from unmineralized samples. As a result, those were considered separately. The remaining 286 samples of Laramide intrusive rocks contained an average fluorine content of 850 ppm, with a median of 510 ppm and range of 200 to 33,900 ppm. Bostonite dikes in the central Front Range had an average of 2,290 ppm fluorine, the highest of any single lithology of this age.

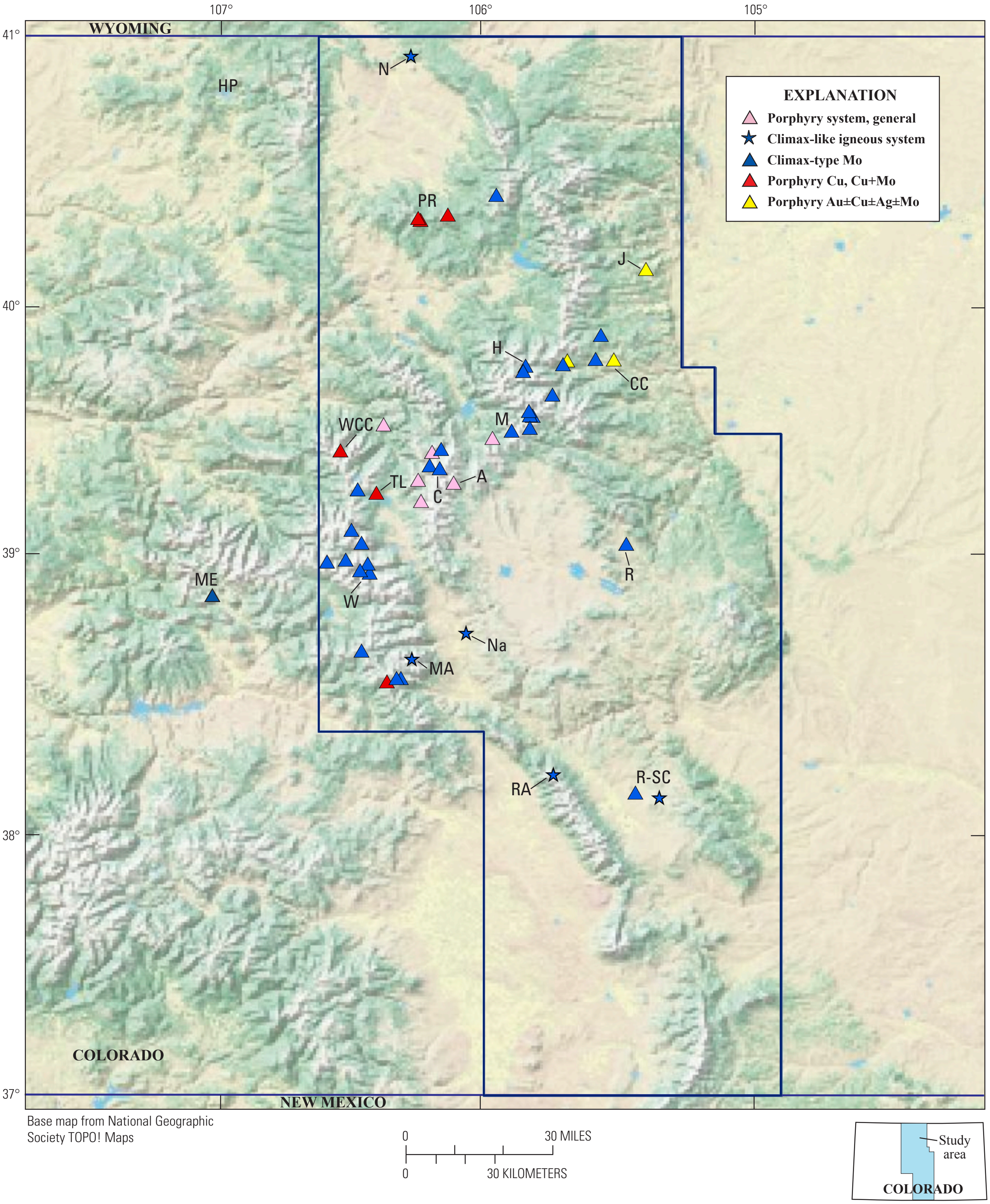
All but one of the 21 analyses of Laramide volcanic rocks were from the 65.5 Ma shoshonitic lava flows at North and South Table Mountains west of Denver. Those flows averaged 1,030 ppm fluorine, with a median of 700 ppm and range of 40 to 2,700 ppm.

Jamestown District

The fluorspar and gold deposits of the Jamestown district in the northwestern Front Range (figs. 10, 11) are related to the 54 to 56 Ma Porphyry Mountain sodic granite pluton (Nash and Cunningham, 1973; Cunningham and others, 1994; Terry Klein, U.S. Geological Survey, written commun., 2008). The district was one of the leading producers of fluorspar in the State. Fluorspar occurs as a primary component of the pluton and as veins, breccia pipes, and dikelike fluorite and vanadium mica veins that cut the pluton and its nearby country rocks. Fluid inclusion studies showed fluorite formation temperatures of 250° to 375°C and salinities of 26 to more than 50 percent NaCl equivalent (Nash and Cunningham, 1973). The presence of fluorite in both the pluton and the related breccia pipes indicates a primary magmatic source, likely the sodic granite, for the fluorine. The regional late Eocene Rocky Mountain erosion surface, which formed after mineralization, cut down to within 700 to 1,200 m of the current tops of the deposits (Nash and Cunningham, 1973; Kelley and Chapin, 2004), so the deposits formed at even greater depths. Postmineralization regional downcutting, much of it in the Pliocene, has exposed the tops of the deposits.

Samples were collected from drill core and mine and surface exposures in the Jamestown district in the early 1970s, and a few were collected in the early 2000s. Only two sets of

Figure 11 (facing page). Locations of porphyry-related deposits of molybdenum, copper, and other metals, including vein, replacement, and greisen deposits related to these prophyry systems. Several locations, shown with a blue star, are either volcanic systems or plutons that lack true porphyry characteristics but which have chemical and mineralogical affinities to Climax-type igneous systems (“Climax-like”). As described in the text, many of these porphyry systems contain large to moderate amounts of fluorine and fluorite, and the mineralizing effects of each system extends far beyond the actual site of the mineral deposit. All of the igneous systems are latest Cretaceous and Tertiary, with the exception of the greisen of the Mesoproterozoic Redskin Granite. Abbreviations for select igneous systems: A, Alma; C, Climax; CC, Central City; H, Henderson; HP, Hahn’s Peak; J, Jamestown; M, Montezuma–Geneva Basin; MA, Mount Antero; ME, Mount Emmons; N, Northgate (porphyry origin uncertain); Na, Nathrop volcanic field; PR, Poison Ridge; R, Redskin; RA, Rito Alto; R–SC, Rosita–Silver Cliff; TL, Turquoise Lake; W, Winfield; WCC, West Cross Creek. Locations from Bookstrom (1990), with additional information from Art Bookstrom and Steve Ludington (U.S. Geological Survey, written commun., 2009). Refer to figure 1 for general geographic locations.



Base map from National Geographic Society TOPO! Maps

latitude and longitude coordinates were used for the vast majority of the samples (one set each for submittal batches in 1970 and 1971), so specific sample locations are largely unknown. All of the samples, though, were collected from near the general center of the district. The 255 samples all were described as felsic intrusive, felsic, granodiorite, or quartz monzonite, with no mention of mineralization products. Analytical data for the 255 samples ranged from 200 ppm to 36.8 percent (368,000 ppm) fluorine, with an average of 27,800 ppm fluorine and a median of 2,900 ppm. Of these samples, 96 contained 1 percent (10,000 ppm) or more fluorine. This high median, coupled with descriptions of the Porphyry Mountain pluton as being, in places, as fluorite-rich as some of the veins and breccias (Nash and Cunningham, 1973), makes a distinction between mineralized and unmineralized a somewhat moot point.

Fluorine in Non-Magmatic Laramide Mineral Deposits and Settings

Laramide Shear Zones in the Front Range

At least eight regionally extensive, northwest-striking fault zones (“breccia reefs”; Lovering and Goddard, 1950) cut Paleoproterozoic granites, metavolcanic rocks, and metasedimentary rocks in the east-central Front Range. The faults originally formed in the Proterozoic, were active during Laramide uplift of the Front Range, and were reactivated modestly in the Neogene. At a regional scale, the fault zones are linear; at more local scales, they are linear to sinuous and are composed of brecciated zones several tens of to more than a hundred meters wide.

The brecciated fault zones contain three types of cement: carbonate-adularia, quartz-hematite, and fluorite-quartz. The carbonate-adularia cement is restricted to areas where the faults cut the metamorphic rocks. The quartz-hematite cement is most common in fault zones that cut granitic rocks but is present in the metamorphic rocks as well (Wallace, 1982). In the transition zone between the two types of cements, the carbonate-adularia cement is younger. The components of these two cements were derived from their respective lithologic hosts (Wallace, 1982; Wallace and Whelan, 1986), and the hydrothermal fluids moved along the fracture zones during Laramide faulting at about 69 Ma (Ludwig and others, 1985).

The carbonate-adularia-cemented fault zones host hydrothermal uranium deposits, such as the large Schwartzwalder deposit and the smaller Ladwig, Mena, and Bonzo deposits, all in Jefferson County near the range front. The principal ore mineral is pitchblende, with adularia, carbonates, and various iron, molybdenum, arsenic, and copper sulfide minerals (Wallace and Whelan, 1986). Fluorite is present in only small amounts, and it was deposited very late in the paragenetic sequence from fluids that had a predominantly meteoric composition (Wallace and Whelan, 1986). The Schwartzwalder vein materials contain an average of 277 ppm fluorine, as compared to the fresh hornblende- and biotite-rich host rocks that have an average fluorine content of 1,083 ppm (Wallace, 1983).

Segments of several of the fault zones contain abundant fluorite and quartz within a north-northeast-trending belt that roughly overlaps the carbonate-quartz transition zone and is independent of host lithologies (“F” in fig. 10; Wallace, 1982). The fluorite and quartz form veinlets and stockworks that cut both the carbonate-adularia and quartz-hematite cements. The fluorite-quartz cement is not brecciated, indicating formation during the waning stages of, or after, faulting. Amethyst, barite, galena, and secondary copper minerals are locally present in the veins and cement, and fluid-inclusion data from fluorite indicate formation temperatures of about 150°C (Wallace, 1982). No fluorine analyses are available for the fluorite-rich breccias, although the abundance of fluorite indicates fluorine contents on the order of tens of thousands or more parts per million. The origin of the fluorite is uncertain, but the oblique orientation of the fluorite-rich zone relative to the fault zones, the lack of correlation with host lithology, the association with base metals, and the relatively late tectonic age suggest a possible magmatic fluorine contribution. However, the relatively low temperatures and absence of daughter minerals in the fluid inclusions distinguishes these deposits from the higher-temperature, saline fluids that formed the pluton-related fluorite deposits at Jamestown, as described previously (Nash and Cunningham, 1973).

Laramide Uranium Deposits in Thrust Zones

Uranium deposits at the Pitch mine occur in breccia zones along thrust or upthrust faults south of Monarch Pass and just west of Marshall Pass (fig. 10; Nash, 1988). The deposits are most common where the faults cut the Mississippian Leadville Dolomite, but they also are present in the Pennsylvanian Belden Shale and underlying Paleoproterozoic igneous and metamorphic rocks. Pitchblende and coffinite are the primary ore minerals, and pyrite, marcasite, and minor quartz are the major gangue minerals; fluorite has not been reported. Supergene weathering has produced a variety of secondary iron oxide and hexavalent uranium minerals to depths of more than 100 m, with reduced, commonly carbonaceous ores in deeper parts of the deposit. The Pitch mine exploited the uranium ores in the 1980s.

Nash (1988) analyzed the fluorine contents of core samples of fresh and altered host rocks (principally carbonates and shale) and ore. Those analyses indicate fluorine contents ranging from less than 100 ppm to as much as 2,100 ppm, with a geometric median of 300 ppm, lower than the range of fluorine in carbonate and shale units elsewhere in central Colorado (see section on “Fluorine in Paleozoic and Mesozoic Sedimentary Rocks”). There is no correlation between fluorine and uranium (Nash, 1988), and the fluorine and any variations in concentration are not related to the hydrothermal system that introduced the uranium. More likely, the original compositions of the host rocks or the relative abundances of those lithologies in the mineralized fault breccias controlled the fluorine contents of the samples analyzed. Thus, although other uranium deposits may be present in similar geologic settings in central Colorado, their fluorine contents probably would not differ from those in the host rocks.

Fluorine Related to Middle Tertiary Calc-Alkaline Igneous Centers

The middle Tertiary (roughly 45 to 25 Ma) calc-alkaline igneous activity in central Colorado produced numerous shallow plutons and large, caldera-forming eruptions (figs. 10, 11). Mineralization was associated with many of these igneous centers, and some of them deposited fluorite or fluorine-rich minerals, either in fluorspar deposits or as accessory minerals in other types of mineral deposits. The fluorspar deposits at Wagon Wheel Gap were among the largest sources of fluorspar in Colorado. Other areas that contain small, noneconomic amounts of fluorite include the intrusive complexes near Mount Antero and Turquoise Lake and, to a lesser degree, the Tincup and Bonanza mining districts.

The 29.6-Ma Mount Antero Granite in the southern Sawatch Range is an unusual evolved, rare-metal granite. It contains primary magmatic fluorite, and abundant miariolitic cavities and late-stage pegmatites contain fluorite, topaz, and a variety of other minerals (Sharp, 1976). Unfortunately, no fluorine analyses from the granite have been published. Electron-microprobe analyses of biotite from the granite showed a median fluorine content of 0.88 weight percent, pointing to a relatively high amount of fluorine in the parent magma (Toulmin and Hammarstrom, 1990). Progressive crystallization of the magma concentrated volatiles, including fluorine, into the late-stage phases to form the fluorine-bearing minerals near the upper carapace (Thompson and Pulfrey, 1973).

The 36-Ma Turquoise Lake porphyry system west of Leadville and the Bonanza caldera southwest of Salida contain high-silica granites and gangue fluorite (and topaz at Turquoise Lake) in related veins. Fluorine and fluorite are dispersed around the Turquoise Lake porphyry copper-molybdenum system, which is intermediate between a Climax system and a typical calc-alkaline porphyry copper deposit (Craig, 1980). Fluorite at the 36-Ma, calc-alkaline Bonanza caldera complex is a common gangue mineral in veins at the south end of the district. The veins and abundant hydrothermal alteration formed during the slightly postcaldera emplacement of rhyolite intrusive bodies and exogenous domes (Burbank, 1932; Varga and Smith, 1984).

Eruption of high-silica magmas near Nathrop at 30–29 Ma produced chemically evolved rhyolite flow units. The rhyolite contains topaz and spessartine garnet in vesicles and lithophysae (Van Alstine, 1969; Scott, 1975b; Christiansen and others, 1986). The presence of topaz suggests a genetic link to Climax-type systems (Seedorff and Einaudi, 2004b), but the K_2O/Na_2O ratios and trace-element compositions are more similar to calc-alkaline magmas (White and others, 1981; Christiansen and others, 1986). Fluorine in 13 USGS samples from the Nathrop rhyolites averaged 1,080 ppm, with a median of 600 ppm. In addition, three samples of volcanic glass contained 1,800 (obsidian), 2,300 (perlite), and 5,500 (vitrophyre) ppm fluorine (Van Alstine, 1969; Christiansen and others, 1986).

Analyses of 117 middle Tertiary calc-alkaline intrusive rocks averaged 1,040 ppm fluorine, with a median of 770 ppm and range of 100 to 5,800 ppm. Similar-aged calc-alkaline volcanic rocks, many of which are exposed in the eastern San Juan Mountains area, averaged 910 ppm fluorine (n=246). Many of the volcanic rocks are ash-flow tuffs that were erupted from a caldera and emplaced over extremely large areas. The behavior and migration of volatiles such as fluorine during the eruption, emplacement, and cooling of ash-flow tuffs can produce widely variable fluorine contents in the resulting rocks. For example, two samples of vitrophyre from the Wall Mountain Tuff contained 1,700 ppm fluorine, but the tuff as a whole otherwise averaged only 360 ppm. More locally derived lava flows, such as basalt and andesite flows in the San Luis Hills south of Alamosa (fig. 2), also showed some variation from the average: samples from that area average 330 ppm fluorine, and two additional samples contain 1,080 and 2,400 ppm with no evidence of secondary mineralization.

Wagon Wheel Gap

High-grade fluorite veins cut Oligocene (~27 Ma; Hon and Mehnert, 1983) volcanic rocks in the Wagon Wheel Gap district, 14 km southeast of Creede (figs. 4, 10). This district is outside the Central Colorado study area, but its setting is similar to that found in some of the western parts of the study area, and runoff from the district drains into the nearby Rio Grande, which flows east into the study area. The veins were mined in the first half of the 20th century, and the district has been inactive since then. Several hot springs have formed along the downhill, western extensions of the faults that host the fluorspar veins (Emmons and Larsen, 1913). As described herein, these springs contain elevated amounts of fluorine.

The geology and ore deposits in the district were described by Aurand (1920) and Vanderwilt and others (1947), and the district was visited briefly during the current study. The wide, steeply south dipping veins generally strike to the east (fig. 12). The wall rocks were intensely brecciated and silicified near the veins, and these features can extend for more than 200 m into the hanging wall. The veins themselves were not brecciated after mineralization, indicating that mineralization took place after the formation of the faults. The veins contain coarse- to fine-grained fluorite and 5- to 10-cm barite crystals in the upper parts of the veins (fig. 12). The fluorite commonly is banded, indicating multiple phases of deposition.

Figure 12. Photographs of the Wagon Wheel Gap fluorspar district. (*A*) Mill for fluorspar ores. The east-striking veins extend up the hill in the background, and one dump is visible behind the mill. Photograph taken looking southeast. (*B*) Steeply dipping fluorite-barite vein in Tertiary volcanic rocks. Red pen about 15 centimeters long. (*C*) Photograph looking northwest from the hill and fluorspar veins shown in figure 12*A*. The mill is in the valley behind the trees to the lower left. The margin of the Creede caldera is shown: rocks underlying the hills in the left background are within the caldera, and the fluorspar veins are just outside of the caldera margin. The Rio Grande is in the valley near the right edge of the photograph.

Fine-grained pyrite is common in adjacent wall rocks in the lower parts of the mine, and orange clay is more common in the upper parts of the exposed veins.

Trace-element analyses of the veins show locally large amounts of barium due to the coarse barite that is intergrown with the fluorite (table 3). Other anomalous elements in some of the samples include arsenic, mercury, and thallium, which are common elements found in many near-surface epithermal mineral deposits (Berger and Silberman, 1985).

More recent work indicates that the deposits formed just outside of the margin of the 27-Ma Creede caldera (fig. 12; Ratté and Steven, 1967). A lake formed in the caldera moat and began to drain to the east along a paleovalley in the vicinity of the fluorite district by about 23 Ma (Steven and others, 1995). Basalt flows were erupted into the paleovalley at about 22 Ma, and the fluorite veins had formed by about 21 Ma (minimum age on fluorite; Steven and others, 1995; Campbell and Barton, 2005). The analytical uncertainties in these two sets of ages suggest that the veins may have formed at the same time as the basaltic volcanism. Pliocene and younger (Steven and others, 1995; Rye and others, 2000) erosion has incised the early Miocene paleosurface and eroded the upper parts of the veins. As noted by Barton and others (2000), the currently exposed top of the vein system is about 400 m beneath the early Miocene paleosurface, and the veins clearly extended to a shallower depth before the tops were eroded. The combination of active faults, an east-draining groundwater system, and high heat flow (and possibly fluorine) from the basaltic volcanism provided the necessary combination of plumbing, water, and heat to produce the fluorite veins.

The age of this deposit is slightly young for the middle Tertiary suite of magmatism, and the deposits instead may have formed during the early stages of rifting. Like Questa and the related Latir volcanic field, the Creede caldera is slightly young and silicic compared to other San Juan calderas and may be related to the early stages of extension. Due to the at least spatial association with the Creede caldera, Wagon Wheel Gap is included in this section on middle Tertiary igneous centers. Although other basalt-related fluorspar deposits have not been reported in the San Juan Mountains volcanic field, similar lower Miocene basalts are present in the region, including the western part of the central Colorado study area. As such, the basaltic magmas have contributed to the formation of fluorite veins elsewhere if the hydrology and faulting provided the other key elements.

Fluorine Related to Middle Tertiary Alkalic Complexes

Mid-Oligocene magmatism produced various alkaline intrusive and volcanic complexes east of the future Rio Grande rift and generally but not exclusively south of the latitude of Salida. Major centers that generated alkaline or strongly alkalic rocks include Cripple Creek, Rosita Hills, Silver Cliff, and the Spanish Peaks (figs. 10, 11). This stage of magmatism ranged in age from about 32 Ma to about 23 Ma.

Cripple Creek is best known as one of the world’s largest gold deposits produced by alkalic magmatic systems. The Cripple Creek deposits were related to alkaline diatremes, phonolites, and various alkaline mafic units that were emplaced into Proterozoic rocks (Loughlin and Koschmann, 1935; Thompson and others, 1985; Kelley and others, 1998). The system formed between about 32 and 28 Ma in a shallow, near-surface environment (Thompson and others, 1985; Kelley and others, 1998). Fluorite is a common gangue mineral in most of the hydrothermal stages, especially in those that produced large hydrothermal breccias; it also is a minor, relatively uncommon alteration mineral, especially immediately adjacent to veins (Thompson and others, 1985). Hydrothermal alteration is pervasive in the district except in abundant, postmineralization phonolite dikes. Fluorine analyses of 13 relatively unaltered alkaline igneous rocks indicated an average of 790 ppm fluorine and a median of 750 ppm (Kelley and others, 1998). The 25 Cripple Creek samples in the USGS rock database, which included fresh to altered rocks and fluorite-bearing phonolitic breccia, averaged 1,700 ppm fluorine and a median of 800 ppm. Omitting the one sample of fluorite-bearing breccia that contained 24,000 ppm fluorine, the average and median were 770 and 800 ppm, respectively, and the highest fluorine content was 1,300 ppm. These data suggest that fresh igneous rocks did not contain abnormally large amounts of fluorine, and that the fluorine was introduced into veins and breccias, but not to any great extent into the broad alteration zone, by the hydrothermal fluids. Kelley and others (1998) used isotopic data to conclude that those fluids were derived from the phonolitic magmas as they were emplaced and differentiated. The fluorine, like other volatiles, was concentrated into the fluid phase during this process, leading to the fluorine-rich fluids.

The Silver Cliff and Rosita volcanic centers near Westcliffe produced a wide variety of volcanic flows and intrusive rocks with compositions that include high-silica, Climax-type rhyolites, alkali-calcic rhyolites and andesites, and trachytes and trachyandesites. Magmatic activity took place between about 35 and 27 Ma (Sharp, 1978; McEwan and Rice, 1996; Terry Klein, U.S. Geological Survey, unpubl. data, 2009). Gold-silver, base-metal, and minor molybdenite mineralization accompanied various stages of igneous activity, producing the mineral deposits in the Silver Cliff and Rosita districts. As such, elements of several types of mineralizing systems can be found in the two districts. No fluorite deposits are present in the district, although fluorite has been reported in a few veins. Analyses of 36 samples of fresh and altered rocks in the USGS rock database from the Silver Cliff and Rosita areas showed an average fluorine content of 840 ppm and a median of 800 ppm, with a range of 100 to 2,300 ppm fluorine. The freshness of these rocks was not noted, although alteration is widespread in both areas.

Saint Peters Dome

Fluorite and quartz form veins in the Saint Peters Dome area in the southeastern part of the Pikes Peak batholith (figs. 6, 10), and some of the veins were mined for fluorspar. The veins fill en echelon shears, with brittle deformation textures, within the Pikes Peak Granite near the Mount Rosa intrusive body. The shear zones cut all Proterozoic fabrics, dikes, and structures (Steven, 1949), including lamprophyre and aplite dikes and cryolite-bearing pegmatites. A second population of lamprophyre dikes was emplaced along the shear zones.

Steven (1949) concluded that the second stage of dikes, along with the fluorite veins, formed during a Tertiary epithermal event that was distinctly different from the deeper magmatic conditions that formed the fluorine-rich pegmatites related to the Pikes Peak Granite (see discussion in the section on “Fluorine and Fluorite in 1.1-Giga-annum Plutonic Rocks”). Mineralization began with deposition of chalcedonic silica and chlorite, followed by a period of brecciation and subsequent deposition of fluorite and quartz, with minor base-metal sulfides and barite.

The fluorine-rich Cripple Creek district is 23 km to the west and formed at about 29 Ma (fig. 6; Kelley and others, 1998). Small, satellitic igneous bodies related to that igneous center are only 10 km west of Saint Peters Dome. Lamprophyre dikes are common at Cripple Creek and were part of the larger igneous system. Although the relation

Table 3. Trace-element geochemistry for central Colorado fluorspar districts.

[Some elements analyzed by more than one method. Analyzed for but not detected in SQS analyses (detection limit in parts per million in parentheses): Ag (8), Bi (40), Cd (8), Eu (8), Ho (20), Ta (200). <, less than; >, greater than. Methods: SQS, semiquantitative spectrographic analysis; ED, energy dispersive X-ray analysis; ICP, induced coupled plasma analysis; DN, delayed neutron analysis. Units: ppm, parts per million; %, weight percent. Analysts (all U.S. Geological Survey, Denver, CO): B. Roushey, L. Bradley, L. Hageman, G. Gourtin, R. McGregor, J. Kent, P. Briggs. All samples collected from fluorite-rich surface veins and mineralized outcrops by A.R. Wallace; fluorine content likely tens of a percent for each sample. One latitude and longitude listed for each district]

Method unit	Ca SQS %	As SQS ppm	Au ICP ppm	Ba SQS ppm	Be SQS ppm	Ce SQS ppm	Co SQS ppm	Cr SQS ppm	Cu SQS ppm	Ga SQS ppm	Hg ICP ppm	La SQS ppm	Li SQS ppm	Mn SQS ppm	Mo SQS ppm
Northgate District (40°56.00, 106°17.00)															
NG1	0.55	1,000	<0.002	71	<4	30	<4	<4	20	<20	0.92	10	40	30	1,300
NG10	0.1	50	<0.002	390	6	30	<4	7	<4	<20	0.03	8	140	470	<8
NG11	1.0	<40	<0.002	470	4	43	<4	10	<4	<20	0.05	20	88	270	<8
NG12	32	<40	<0.002	140	<4	43	<4	<4	8	<20	0.02	20	20	120	<8
NG13	2.6	60	<0.002	63	4	30	<4	<4	200	<20	0.08	20	95	700	<8
NG2A	42	<40	<0.002	46	<4	<20	<4	<4	<4	<20	<0.02	8	<8	<20	<8
NG2B	49	<40	<0.002	61	<4	30	4	<4	<4	<20	0.05	10	<8	<20	<8
NG2C	46	<40	<0.002	10	<4	30	4	<4	<4	<20	<0.02	20	20	<20	8
NG5	45	<40	<0.002	10	<4	98	5	<4	<4	<20	0.02	20	20	<20	<8
NG6	35	50	<0.002	350	7	30	5	<4	41	<20	0.22	10	20	1,000	10
NG7	27	80	<0.002	34,000	30	30	48	10	400	190	0.14	10	10	120,000	44
NG8	38	<40	<0.002	160	6	50	<4	4	6	<20	0.18	10	<8	450	<8
NG9A	40	<40	0.008	100	4	47	5	<4	20	<20	0.06	20	10	130	<8
NG9B	35	50	<0.002	78	<4	30	<4	<4	30	<20	0.07	10	10	270	10
NG9D	0.51	<40	0.004	1,400	<4	42	<4	6	65	<20	0.07	20	70	58	<8
Crystal District (40°40.50, 106°35.25)															
CR1	53	<40	<0.002	97	5	<20	4	<4	<4	<20	0.03	<8	10	<20	<8
CR2	47	<40	<0.002	90	4	20	<4	<4	<4	<20	<0.02	<8	10	<20	<8
CR3	22	50	<0.002	57	<4	<20	5	<4	<4	<20	0.07	<8	78	680	20
Delaney Butte (40°41.83, 106°29.083)															
DB1	39	<40	<0.002	94	<4	40	<4	<4	4	<20	0.04	20	20	<20	10
DB2	15	<40	<0.002	58	<4	<20	<4	<4	<4	<20	0.03	<8	89	40	<8
DB3	0.2	<40	<0.002	1,400	<4	130	<4	10	<4	<20	0.10	57	20	52	<8
DB5B	0.1	<40	<0.002	120	<4	<20	<4	<4	5	<20	0.10	9	160	20	10
DB6	1.0	<40	<0.002	200	<4	75	30	66	51	30	0.03	40	40	1,700	<8
Brown's Canyon District (38°38.50, 106°04.50)															
BC1	43	<40	<0.002	130	<4	49	<4	<4	5	<20	0.15	10	10	20	<8
BC2	0.3	<40	<0.002	210	<4	30	<4	<4	<4	<20	0.08	8	10	20	<8
BC3	48	<40	<0.002	130	<4	30	<4	<4	<4	<20	<0.02	<8	<8	<20	<8
PS1	46	<40	0.14	980	<4	65	5	<4	<4	<20	0.07	<8	<8	<20	<8
Poncha Springs District (38°29.75, 106°0455)															
PS1	46	<40	0.14	980	<4	65	5	<4	<4	<20	0.07	<8	<8	<20	<8
PS2	1.7	<40	<0.002	2,400	<4	91	<4	<4	<4	<20	<0.02	51	<8	150	<8
Wagon Wheel Gap District (37°44.83, 106°49.58)															
WW2A	48	<40	<0.002	31,000	<4	79	30	<4	5	<20	<0.02	<8	<8	30	<8
WW2B	27	<40	<0.002	10,000	<4	44	10	<4	<4	<20	<0.02	<8	<8	<20	<8
WW2C	0.3	110	<0.002	1,200	<4	80	<4	<4	5	<20	0.33	42	45	20	20
WW3B	0.09	350	<0.002	690	<4	58	<4	<4	9	<20	1.6	30	64	<20	10
WW4	48	<40	<0.002	1,200	6	20	6	<4	<4	<20	<0.02	<8	<8	<20	<8
WW5	54	<40	<0.002	540	<4	53	<4	<4	<4	<20	0.02	<8	<8	<20	<8

between Cripple Creek and the Saint Peters Dome area is unknown, Steven (1949) noted that the sequence of mineralization at Saint Peters Dome, including the emplacement of lamprophyre dikes, was similar to that at Cripple Creek. As such, the Saint Peters Dome fluorite deposits appear to have been formed by Oligocene alkalic magmatism. The abundance of fluorine in the Pikes Peak Granite host rocks permits derivation of at least some of the fluorine from those rocks, with hydrothermal circulation driven by the Tertiary magmatism, but the fluorine at Saint Peters Dome also could have been derived from the Tertiary magmas as at Cripple Creek.

Fluorine Related to Climax-Type Molybdenum Deposits

Central Colorado contains some of the world's largest molybdenum deposits ("Climax-type" deposits), including those that have been mined at Henderson and Climax. Several other molybdenum prospects were discovered but not mined in the Sawatch Range (Winfield and the lower fluorine West Cross Creek system), the Montezuma district, the north-central Front Range, possibly at Northgate in North Park, and other locations (figs. 10, 11).

Table 3. Trace-element geochemistry for central Colorado fluorspar districts.—Continued

Method unit	Nb SQS ppm	Nd SQS ppm	Ni SQS ppm	Pb SQS ppm	Rb ED ppm	Sc SQS ppm	Sr ED ppm	Te ICP ppm	Th DN ppm	Tl ICP ppm	U DN ppm	V SQS ppm	Y SQS ppm	Yb SQS ppm	Zn ED ppm	Zr ED ppm
Northgate District (40°56.00, 106°17.00)																
NG1	30	20	<8	30	150	<8	37	<0.05	<36	63.0	197	40	280	30	27	77
NG10	<20	<20	<8	<20	58	<8	73	<0.05	3.2	0.85	2.01	80	<8	<4	<10	86
NG11	<20	<20	<8	<20	194	<8	66	<0.05	6.98	1.30	2.92	20	10	<4	18	144
NG12	<20	<20	<8	20	200	<8	79	<0.05	60.7	1.40	3.75	<8	100	7	23	55
NG13	30	<20	<8	30	174	<8	34	0.05	<14	1.15	66.7	74	180	20	16	34
NG2A	<20	<20	<8	<20	<10	<8	73	<0.05	93.3	0.90	<0.14	<8	240	8	<10	<10
NG2B	<20	<20	<8	<20	<10	<8	61	<0.05	84.8	0.05	<0.14	<8	150	5	<10	<10
NG2C	<20	<20	<8	<20	<10	<8	61	<0.05	78.0	0.70	<0.14	<8	75	<4	<10	12
NG5	<20	<20	<8	<20	<10	<8	78	<0.05	80.3	<0.05	<0.14	<8	80	<4	<10	12
NG6	<20	<20	<8	<20	74	<8	210	<0.05	65.8	1.15	3.10	10	110	6	17	27
NG7	20	<20	10	100	60	<8	720	<0.05	52.1	76.0	11.1	40	64	7	<10	32
NG8	<20	<20	<8	<20	25	<8	340	<0.05	67.7	0.25	0.905	<8	30	<4	<10	21
NG9A	<20	<20	<8	<20	77	<8	176	<0.05	72.8	1.10	8.32	<8	62	5	12	28
NG9B	<20	<20	<8	<20	76	<8	180	<0.05	67.9	1.00	6.92	20	140	10	<10	24
NG9D	<20	<20	<8	30	134	<8	148	0.05	10.4	1.15	4.12	30	30	<4	16	102
Crystal District (40°40.50, 106°35.25)																
CR1	<20	<20	<8	<20	<10	<8	500	<0.05	84.5	<0.05	2.28	<8	30	4	<10	11
CR2	<20	<20	<8	<20	22	<8	470	<0.05	80.5	0.20	2.75	<8	30	<4	<10	19
CR3	<20	<20	<8	<20	<10	<8	100	<0.05	48.1	0.75	40.4	<8	68	8	27	<10
Delaney Butte (40°41.83, 106°29.083)																
DB1	<20	<20	<8	<20	37	<8	85	<0.05	72.7	0.40	2.26	<8	50	<4	< 10	29
DB2	<20	<20	<8	<20	87	<8	42	<0.05	45.7	0.80	6.02	<8	40	<4	< 10	50
DB3	30	46	<8	<20	315	10	100	<0.05	<6.9	3.40	21.3	100	20	<4	13	335
DB5B	<20	<20	<8	<20	52	<8	47	<0.05	25.3	1.90	8.66	10	<8	<4	<10	61
DB6	20	20	40	<20	39	20	67	<0.05	21.6	0.20	14.8	170	20	<4	205	395
Brown's Canyon District (38°38.50, 106°04.50)																
BC1	<20	<20	<8	<20	16	<8	102	<0.05	72.6	0.10	1.43	20	63	5	10	31
BC2	<20	<20	<8	<20	28	<8	20	<0.05	<1.6	0.10	1.02	<8	<8	<4	<10	23
BC3	<20	<20	<8	<20	<10	<8	186	<0.05	94.9	<0.05	<0.13	<8	64	<4	<10	23
PS1	<20	<20	<8	<20	<10	<8	270	<0.05	88.7	<0.05	<0.14	<8	110	10	<10	11
Poncha Springs District (38°29.75, 106°0455)																
PS1	<20	<20	<8	<20	<10	<8	270	<0.05	88.7	<0.05	<0.14	<8	110	10	<10	11
PS2	<20	40	<8	<20	30	<8	76	<0.05	10.3	0.10	4.02	<8	83	9	12	315
Wagon Wheel Gap District (37°44.83, 106°49.58)																
WW2A	<20	<20	<8	<20	<10	<8	920	<0.05	85.6	<0.05	0.810	<8	250	10	<10	23
WW2B	<20	<20	<8	<20	34	<8	>5,000	<0.05	55.2	<0.05	<0.13	<8	200	10	13	12
WW2C	<20	20	<8	20	180	<8	52	<0.05	12.6	1.50	5.94	10	20	<4	10	245
WW3B	<20	<20	<8	20	73	<8	86	<0.05	10.2	5.10	4.05	10	10	<4	18	192
WW4	<20	<20	<8	<20	<10	9	198	<0.05	88.6	<0.05	<0.14	<8	140	7	<10	<10
WW5	<20	<20	<8	<20	<10	<8	188	<0.05	90.1	<0.05	<0.14	<8	150	8	<10	<10

Climax-type deposits formed during the shallow, clustered emplacement of one or more middle Tertiary granitic magmas (White and others, 1981). At Henderson, as many as 12 stocks were emplaced into three intrusive centers between 29.9 and 27 Ma (Carten and others, 1988; Geissman and others, 1992; Seedorff and Einaudi, 2004b), each contributing to the formation of the overall molybdenum orebody. The magma that formed the Seriate stock at Henderson may have contained as much as 13,000 ppm molybdenum (Carten and others, 1988). The hydrothermal fluids associated with each of these magmas were high in fluorine, producing fluorine-rich minerals such as fluorite, topaz, biotite, sericite, and spessartine garnet. These minerals occur as primary magmatic minerals, as well as in associated veins, disseminations, and alteration zones (Gunow and others, 1980; White and others, 1981; Seedorff and Einaudi, 2004a, b). Interaction between the hydrothermal fluids and primary hydrous mafic minerals, such as biotite, in the host rocks can increase the fluorine level in the host-rock minerals (Gunow and others, 1980). One sample from Climax in the USGS rock database contained 43,000 ppm fluorine, and four others averaged 4,002 ppm fluorine. Analyzed biotites from Henderson contained up to 75,000 ppm fluorine, and sericite contained up to 44,000 ppm fluorine (Gunow and others, 1980).

The distribution and extent of introduced fluorine around Climax-type systems can be enormous. Fluorite and topaz (see section on “Topaz Deposits”) in high-temperature veins generally formed in the core of each magmatic center and ore deposit. In contrast, relatively lower temperature, fluorine-rich veins and alteration haloes tended to form a broad halo around the entire intrusive center rather than around each intrusive body (Seedorff and Einaudi, 2004a, b), grading out from sericite-rich to chlorite-rich alteration assemblages. As such, the intrusive systems as a whole introduced a very large amount of fluorine to a large volume of host rocks.

To show the broad area affected by fluorine-rich hydrothermal fluids, Eppinger and Theobald (1985) mapped the alteration zones in the Vasquez Peak Wilderness Area, a few kilometers north and west of the Henderson molybdenum deposit (figs. 10, 11). There, they identified many areas of hydrothermal alteration and identified abundant fracture-filling fluorite over a broad area (Eppinger and Theobald, 1985). Most commonly, the fluorite-bearing veins there occur in chlorite-altered Proterozoic rocks, including the 1.4-Ga Silver Plume Granite. Similar broad halos of fluorite-filled veinlets are present around a number of other Climax-type intrusive systems, such as at Climax and near Winfield.

At the Sweet Home mine in the Alma district, between Climax and Fairplay (figs. 10, 11), 26-Ma fluorine-rich fluids derived from either Climax or a related but concealed Climax-type system invaded fractures in Proterozoic and lower Paleozoic rocks and minor Oligocene dikes. Structurally, the Alma area is domal, and veins fill fractures that radiate outward from the core of the dome (Bookstrom, 1989). The veins are related to a structurally lower part of the overall, east-tilted Climax-Alma intrusive complex. The veins contain fluorite, topaz, and fluorapatite, along with molybdenite and other Climax-type ore minerals (Bookstrom, 1989; Bartos and others, 2007).

Fluorine Related to the Rio Grande Rift

High heat flow, perhaps locally coupled with young, near-surface igneous activity, produced fluorspar deposits and fluorite veins along the Rio Grande rift from near the Colorado-Wyoming border south to the San Luis Valley in southern Colorado (table 2; figs. 4, 10). Although the deposits have not been dated, geologic relations and indirect dates indicate that these deposits formed during rifting in the Miocene and Pliocene. As described later, hydrothermal activity in some locations, such as at Poncha Springs and Chalk Mountain west of Nathrop, currently produces fluorine-rich geothermal systems and hot springs.

Fluorite deposits and prospects are present at the Crystal district and Delaney Butte area along the west side of North Park and at the Northgate district at the north end of North Park. Fluorspar was mined at Northgate and Crystal, and the fluorite veins at Delaney Butte have small prospects but no recorded production. Poncha Springs and Browns Canyon near Salida were two of the principal rift-related fluorspar-producing districts in central Colorado. The two small deposits in the Sangre de Cristo Mountains had little to no recorded production.

The following descriptions of the districts are based on published information and field work, petrography, and chemical analyses done as part of this study.

Delaney Butte

Delaney Butte is a short, elongate knob near the western side of North Park (figs. 4, 13). Mesoproterozoic granite of the Mount Ethel pluton is exposed at the top of the butte. The granite was carried westward along an east-dipping Laramide thrust fault, and it forms an isolated, fault-bounded mass surrounded on all sides and at depth by Mesozoic and Paleogene sedimentary rocks (Hail, 1965).

Fluorite-quartz veins at Delaney Butte fill narrow, east-striking fractures in the granite (fig. 13), and replacement of early bladed calcite by quartz indicates boiling during mineralization. The fluorite veins contain higher vanadium, zinc, and lithium than any of the other Colorado fluorspar deposits sampled (table 3). In samples collected for this study, fluid inclusions in the fluorite contain hydrocarbons, indicating that the hydrothermal fluids passed through petroliferous Mesozoic reservoir rocks in the basin, such as the Dakota Sandstone and Pierre Shale. Burial history studies in North Park indicate hydrocarbon maturation and migration to reservoirs during the Eocene (Wandrey and others, 2000). The combined data indicate that heated fluids (greater than 100°C) migrated through the hydrocarbon-bearing sediments in the mid- to late Tertiary, scavenged hydrocarbons and some trace elements, and continued up along the thrust to fractures in the brittle granite at Delaney Butte. There, they boiled and precipitated fluorite. The source of the fluorine is unknown, but it could have been derived from the fluorine-rich Mount Ethel pluton at depth or from the Mesozoic pelitic sedimentary rocks along the flow path. This mode of formation, with fluid migration through sedimentary rocks, is similar to that of some fluorite deposits in New Mexico (Hansonburg; Putnam, 1980) and Mexico (Purisima; Gonzalez-Partida and others, 2002), where basinal brines were important components of the ore fluids.

Crystal District

The Crystal district, on the east flank of the Park Range, includes north-northwest-striking veins along faults in the Mesoproterozoic granite of Mount Ethel (figs. 1, 4, 13). A steep, northeast-striking normal fault truncates the south end of these faults and forms the structural border between the Park Range and North Park to the east (Hail, 1965; Snyder and others, 1987). The vein contacts with the granite host are sharp, and the wall rocks have strong hematite and limonite staining but are not silicified or otherwise altered. Most of the fluorite is crystalline and radiating with growth bands; the youngest stage has a mamillary, colloform texture. At the open-pit mines and at small prospects along the north rim of Red Canyon, about 2 km to the northwest, banded to massive chalcedony occurs either alone or as a pre-fluorite stage of mineralization. Local bladed calcite in these zones indicates boiling during mineralization. Chemical analyses of fluorite-rich vein materials do not indicate any anomalous trace-element concentrations (table 3). Overall, the alteration and vein textures and mineralogies point to a shallow, low-temperature but boiling hydrothermal system that was focused along the faults. The Mount Ethel pluton has abundant primary fluorite, and hydrothermal circulation could have leached fluorine along its flow path (Snyder and Hedge, 1978). The absence of arsenic, molybdenum, and base metals argues against a magmatic source for the fluorine.

Northgate District

The Northgate district is located on the flanks of Pinkham Mountain at the north end of North Park (figs. 4, 13). It was one of the largest fluorspar producers in Colorado, and most of the mining was done with large, linear open cuts (fig. 13). The northwest-striking fluorspar veins cut Mesoproterozoic quartz monzonite, Paleoproterozoic gneissic rocks, and Eocene to Oligocene sedimentary rocks of the White River Formation that overlie the crystalline rocks. The quartz monzonite is related to the Mount Ethel pluton, which is exposed in the Park Range to the west and contains abundant primary fluorite (Snyder and Hedge, 1978). Details on the geology of the district were amply described by Steven (1960).

Two west-striking faults form the south edge of the district: the very shallow dipping Independence Mountain thrust fault (Steven, 1960) and the steeper North Boundary reverse fault of Blackstone (1977). Steven mapped both faults as part of a single thrust fault system, but Blackstone argued that the North Boundary fault was younger and offset the Independence Mountain thrust fault. The fluorspar veins are in the hanging wall of the North Boundary fault. Steven (1960) described brecciated fluorite-bearing quartz veins along the Independence Mountain fault, but it is unclear to which of these two faults he was referring. On the basis of textures, he felt that these veins could be Laramide or at least older than the major fluorspar veins.

The two large, northwest-striking fault systems filled by the fluorspar veins clearly are younger than the late Eocene to early Oligocene-age White River Formation. Steven (1960) determined that the faults, and thus the fluorspar deposits, formed after deposition of the North Park Formation, which in the North Park area was deposited between about 20 and 12 Ma (recalculated from fig. 3 in Izett, 1975). The late Miocene (Barstovian) Medicine Bow erosion surface beveled the faults and veins (Steven, 1960; Montagne, 1991). Thus, the major fluorspar veins appear to have formed in the mid- to late Miocene.

The western of the two veins was mined at the Penber, Garo, and Springer mines. Reconnaissance work for this study found large amounts of early pyrite and quartz along this vein, which was not reported by Steven (1960). At the Garo, Penbar, and Fluorspar mines, pyrite and quartz permeate the host granite, in places obscuring the original texture, and they form abundant veinlets in the granite. Where two strands of the fault system converge near the Fluorspar mine, the intervening granite is strongly pyritized. Pyrite decreases in abundance north along the vein system to the Springer mine. The northward extension of the vein, as exposed on the surface and in exploration cuts, has increasingly abundant hematite and no pyrite. Jarosite that might indicate weathering of pyrite is not evident. The fluorite along the western vein system is coarsely crystalline to colloform. A stage of brecciation followed quartz-pyrite mineralization, and the fluorite coats fragments of quartz-pyrite-altered rock as well as fragments that were not altered (fig. 13).

The eastern of the two veins (Fluorite–Camp Creek vein) cuts both Proterozoic rocks and the White River Formation. Pyrite and quartz are scarce along this vein system except for minor occurrences at the southeastern end. The sedimentary beds in the White River Formation contain abundant hematite alteration as much as 200 m away from the veins (fig. 13), and they locally were intensely and pseudomorphically replaced by fine-grained silica. Manganese oxides are much more common along this vein than along the vein to the west. Although some of these oxides may be secondary, the association between hypogene manganese oxides and fluorite has been documented in many veins in the Southwestern United States (Hewitt, 1964). For the most part, the Proterozoic rocks are fresh adjacent to the veins, and the fluorite fills the veins and envelopes unaltered fragments of wall rock. As noted by Steven (1960), north-northeast-striking tension fractures within the vein system also contain fluorite.

Trace-element data from all veins show anomalous amounts of arsenic and molybdenum (table 3). Two samples from the Fluorite–Camp Creek vein contain weakly detectable gold (4–8 ppb). The pyrite-rich material from the Gero-Penber area contains anomalous arsenic, molybdenum, thallium, mercury, and uranium. The manganese-altered White River Formation adjacent to the Fluorspar–Camp Creek vein contains several percent each of manganese and barium, anomalous thallium, and weakly anomalous beryllium.

The arsenic and molybdenum suggest an association with a moybdenum-rich intrusive system (White and others, 1981), and the district was explored briefly in the 1970s as a molybdenum prospect. The highest arsenic and molybdenum values came from the pyrite-quartz parts of the veins, which formed early during mineralization but along the same Miocene faults that controlled the fluorite stage. The anomalous arsenic, mercury, and thallium are consistent with a shallow epithermal environment (Berger and Silberman, 1985). The banded, open-space textures along the veins, as well as the fluid inclusion homogenization temperatures of 100 to 168°C (Steven, 1960), also support an epithermal origin for the fluorite. The fluorine could have been derived from leaching of the Mesoproterozoic granite, as proposed for the Crystal district by Snyder and Hedge (1978), but the association with molybdenum is similar to that seen in Climax-type and copper-molybdenum porphyry systems in central Colorado (Craig, 1980; White and others, 1981).

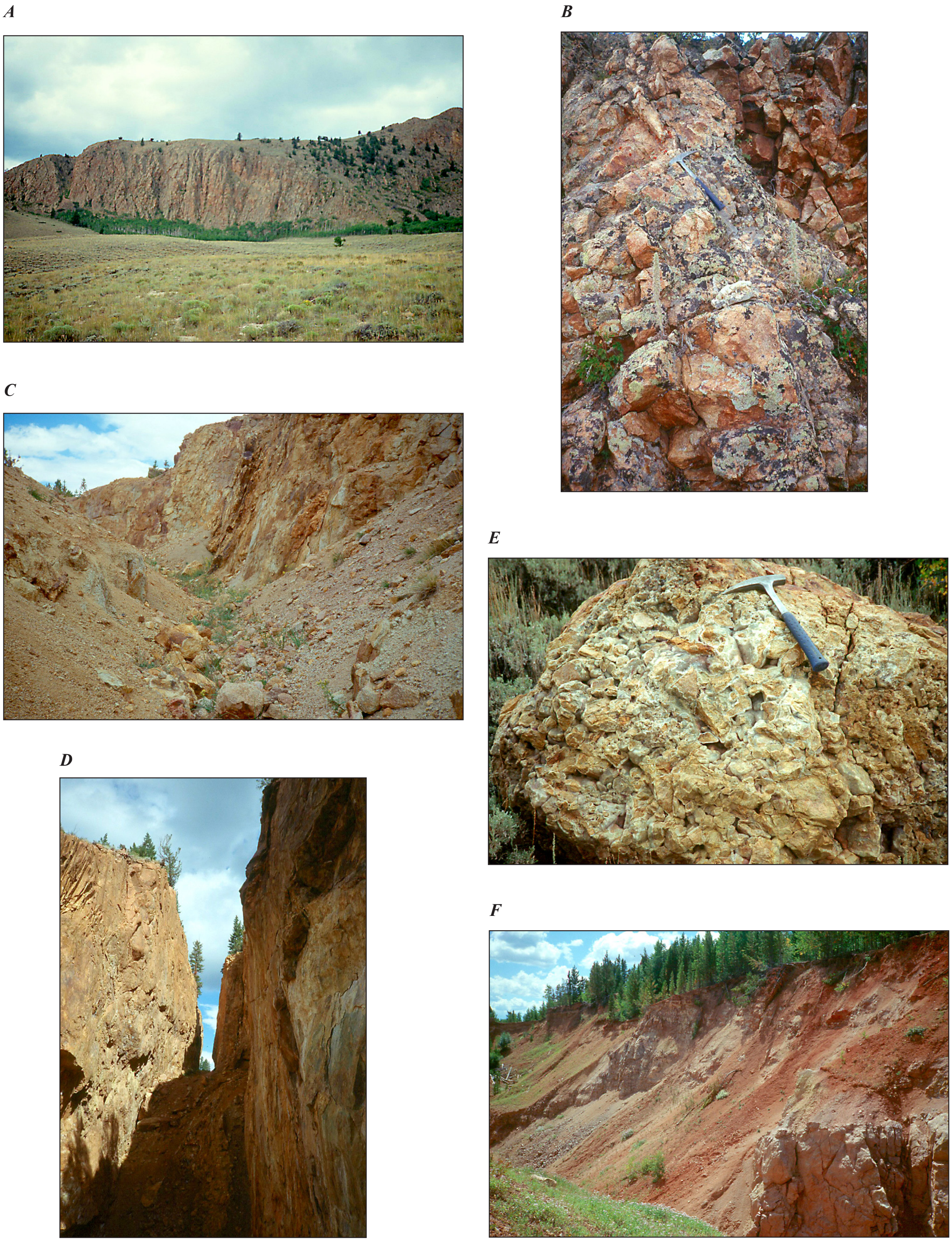

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Figure 13 (above and facing page). Photographs of fluorspar deposits at Northgate, Crystal, Delaney Butte, Browns Canyon, and Poncha Springs. (*A*) Close-spaced vertical fractures in Proterozoic granite at Delaney Butte. (*B*) Fluorite-filling breccia along fracture at Delaney Butte. (*C*) Steeply dipping, fluorite-filled fractures in Proterozoic granite, Crystal district. (*D*) Open cut along fluorspar vein at the Fluorspar mine, Northgate district. (*E*) Brecciated vein material incompletely cemented by fluorspar, Fluorspar mine, Northgate district. (*F*) Hematite alteration in middle Tertiary White River Formation adjacent to the fluorspar vein (lower right), Northgate district. (*G*) Colorado-American fluorspar mine, Browns Canyon district, looking north-northeast. The fluorspar veins extend toward the ridgeline behind the brown headframe in the left center. (*H*) Fluorspar vein (white, steeply dipping) along fault zone that separates the Oligocene Wall Mountain Tuff (left) and Proterozoic granite (right). (*I*) View looking west-northwest from the Poncha Springs fluorspar district. The fluorspar veins are along the hillside in the near foreground, and the modern Poncha hot springs are near the curving road in the lower left. A major normal fault that is part of the Poncha accomodation zone extends from the lower left side of the photograph along the topographic break in the middle left of the photograph. The fault placed Proterozoic rocks on the left (south) against semi-consolidated sediments of the Miocene and Pliocene Dry Union Formation on the right (north). The southern Sawatch Range is in the far distance. (*J*) Fluorite coating slickensides in faults in the Poncha Springs district.

Browns Canyon District

The Browns Canyon district is located on the west side of the Arkansas River 14 km north of Salida (figs. 4, 13). Fluorspar veins in the district were mined until about 1949 (Brady, 1975). Van Alstine (1969) thoroughly described the geology and mineral deposits in the district, and that report can be consulted for details. The following summary provides the general context of the deposits, augmented by data collected for this and other recent studies.

The steeply dipping fluorspar veins fill faults that strike northwest and juxtapose Proterozoic granitic rocks to the east and the 36.7-Ma Wall Mountain Tuff to the west (Van Alstine, 1969; McIntosh and Chapin, 2004). The veins, in places as wide as 10 m (del Rio, 1960), cut epiclastic sedimentary units of the middle Miocene Browns Canyon Formation that overlie the older rocks. On the basis of Van Alstine's (1969) descriptions, the Browns Canyon Formation likely is the basal part of the middle Miocene Dry Union Formation. Lower Dry Union strata southwest of Salida were dated at about 14 Ma (Hubbard and others, 2001). The main part of the Dry Union Formation, which was dated near Salida to the south at 10.5 to 8.5 Ma (Hubbard and others, 2001), overlies the veins and all of the older rock units and was not mineralized. Thus, faulting and mineralization at Browns Canyon likely took place between about 14 and 11 Ma.

Massive botryoidal to finely crystalline fluorite fills fractures and coats breccia fragments in the veins. Quartz, chalcedony, and opal, along with minor pyrite, calcite, iron and manganese oxides, and barite, are also present in the veins, with locally abundant limonite at the surface. Late-stage fluorspar has horizontally banded textures in open spaces. Fault movement produced multiple stages of brecciation during mineralization. Alteration of the wall rocks was minor and confined to within a few tens of centimeters of the vein walls. On the basis of the vein textures, Van Alstine (1969) concluded that the veins formed in a shallow epithermal environment. The horizontally banded and colloform textures in the upper parts of some veins support a very shallow origin. If the top of the Browns Canyon Formation represents the approximate middle Miocene paleosurface at the time of mineralization, then hydrothermal fluids likely extended to the surface in the form of hot springs.

Trace-element analyses of three vein samples from the district did not indicate any anomalous elements (table 3). Mercury, however, was measurable in two samples (0.08 and 0.15 ppm, respectively), consistent with the near-surface setting during mineralization.

Poncha Springs District

The Poncha Springs fluorspar veins are in the uplands south of the Arkansas River Valley, several kilometers south of the town of Poncha Springs (figs. 4, 13). The mine workings include elongate open pits along the veins, bulldozer roads and cuts, and numerous prospects. Most of the mining took place in the 1940s (Brady, 1975). Modern Poncha Hot Springs are located along the western edge of the mined area.

The fluorspar veins occur in a west-northwest-trending, Proterozoic-cored horst that formed during late rift-related faulting. In the middle Miocene, epiclastic sediments of the Dry Union Formation blanketed the areas of the future horst and adjacent Arkansas River graben. The basal Dry Union sediments are as old as 14.4 Ma (Hubbard and others, 2001). After about 8.5 Ma, the youngest age of the Dry Union in this area, the west-northwest-striking faults that formed the horst juxtaposed Proterozoic rocks on the south and the Dry Union Formation on the north (Van Alstine, 1968; Hubbard and others, 2001). The resulting erosion removed much of the Miocene strata from the horst, including the area of the fluorspar veins.

The steeply dipping, northwest-striking faults along which the veins formed cut the Proterozoic igneous and metamorphic rocks in the horst. They extend northwest to, and apparently terminate at, the west-northwest-striking, horst-bounding fault. To the west, similar north-striking faults controlled deposition of the Dry Union (there dated at about 14 Ma; Hubbard and others, 2001), and those faults also terminate at the horst-bounding faults. Brief reconnaissance for this study did not find any evidence of fluorite mineralization in the Dry Union north of the horst-bounding fault. Several kilometers west of the fluorspar district and south of the horst-bounding fault, Van Alstine (1974) reported calcite and chalcedony along faults that cut the basal Dry Union, and minor fluorite is present in the underlying Proterozoic gneiss in that area. Therefore, fluorite mineralization postdated deposition of at least the oldest part of the Dry Union and predated the formation of the horst.

The fluorite in the Poncha Springs district is white to light purple and forms 1-mm- to 5-cm-thick veinlets along joints, fractures, and dilational vugs in the Proterozoic host rocks and along faults. The fluorite is banded and forms radiating, small crystals, and it commonly coats fragments of gneiss along the faults. At higher elevations, the vein fluorite and thin fracture coatings are more colloform than at lower elevations. Vein minerals other than fluorite were not observed. Overall, the fluorspar deposit appears to have formed when fluorine-rich hydrothermal fluids ascended along and permeated a broad fracture zone in the Proterozoic rocks. Fluorite coats steeply dipping slickensides along one fault, and none of the fluorite in the veins was brecciated after deposition, indicating that mineralization took place near the end of, or after, faulting.

The host faults continue to the southeast, and the amount of fluorite in the faults decreases in the same direction. A planar erosional surface extends to about 1 km southeast of the mine workings. This surface may be the middle Miocene paleosurface and thus represents the paleosurface at or near the time of mineralization. The tops of the exposed veins are 200 to 400 m beneath the projection of that surface over the mineralized area, and it is possible that the hydrothermal system extended to the surface.

Fluorspar also occurs at a small prospect (Poncha Pass or Divide claims) 7 km south of the main fluorspar district. It was inaccessible during this study, but del Rio (1960) and Brady

(1975) reported that the small veins formed along a steep, north-striking fault that may have had an en echelon relation to those to the north. The fine-grained to colloform fluorite coats breccia fragments and forms small veinlets. Quartz, chalcedony, and manganese oxides also are present in the veins.

One sample of a 10-cm-thick fluorite vein from the Poncha Springs district contained 0.14 ppm gold (table 3), which is notably anomalous. Trace-element analyses of that and another sample did not indicate any other anomalous elements, including those typically related to hot springs, such as mercury, thallium, and arsenic (table 3).

The overall geology of the district indicates that the fluorite veins formed at a very shallow depth along faults related to the early formation of the Rio Grande rift. The age of mineralization is less than 14 Ma but likely middle Miocene overall; the Browns Canyon district to the north, and in a similar structural setting, formed between about 14 and 11 Ma (see previous discussion of the Browns Canyon District). The vein textures indicate that mineralization took place after major fault movement but before the <8.5-Ma formation of the horst. The modern hot springs indicate that hydrothermal activity has centered on this area, either continuously or episodically, for several million years. The fluorine in those springs may be recycled from the fluorspar deposits.

Sangre de Cristo Mountains Deposits

Two small deposits along the west side of the Sangre de Cristo Mountains contain fluorite veins. These include the Liberty mine just north of Great Sand Dunes National Park and the Beryl mine along Rito Alto Creek 12 km north of Crestone (fig. 4). The Liberty mine has two northwest-striking fluorite veins that cut quartzite. The fluorite coats quartzite fragments in the veins, which also contain hydrothermal quartz and lesser calcite (Aurand, 1920). The USGS MRDS record indicates sporadic production, possibly in the early 1900s. At the Beryl mine, manganese was the primary commodity, but the north-northwest-striking vein locally contained up to 60 percent fluorite (Brady, 1975), similar to other manganese-fluorite veins described by Hewitt (1964). Production, if any, from this deposit was small. Fluorite also was a minor gangue mineral at the San Luis gold mine, 12 km northeast of San Luis in the Culebra Range east of the southeastern part of the San Luis Valley (figs. 1, 4; Benson, 1997).

The fluorite-bearing mineral deposits in the Sangre de Cristo Mountains and Culebra Range have three main origins. The first was related to the middle Tertiary calc-alkalic magmatism. This event produced the ~35-Ma Rito Alto stock, 12 km north-northeast of the Beryl mine in the northern Sangre de Cristo Mountains (Lindsey and others, 1985; Lindsey and Soulliere, 1987), and the ~35-Ma Grayback stock in the northern Culebra Range (Wallace, 1997; Dan Miggins, U.S. Geological Survey, unpubl. data, 2006), which produced iron and gold skarns but no reported fluorite. The second event produced ~26–22-Ma, generally north-striking, high-silica rhyolite dike swarms in both ranges that have chemical affinities to Climax-type rhyolites (Alan Wallace, Ed DeWitt, U.S. Geological Survey, unpubl. data, 1994) and locally contain sparse quartz-molybdenite veinlets and patches of disseminated molybdenite (Art Bookstrom, U.S. Geological Survey, written commun., 2009). The 22-Ma San Luis gold deposit is situated in one of these swarms, and small high-grade gold veins are centered on another swarm on the southern flank of Blanca Peak at the extreme south end of the Sangre de Cristo Mountains (Alan Wallace, unpubl. mapping, 1995). The third event produced the Rio Grande rift and related faults and high heat flow. The early stages of this event overlapped in time with the intrusion of the rhyolite dikes (Benson, 1997), and both the San Luis and Liberty mines may be related to a combination of rhyolite magmatism and incipient faulting related to rifting.

Modern Hot Springs

Thermal springs of various temperatures are scattered throughout central Colorado (fig. 14). Many are located along the Rio Grande rift or in areas with late Cenozoic extensional faults and(or) related volcanic rocks. Most are located in areas of relatively high heat flow, although several are not. These springs include 10 unnamed systems that generally follow the axis of the San Luis Valley graben: Cottonwood, Mount Princeton, Browns Canyon, and Poncha Springs in the upper Arkansas River graben; Hot Sulphur Springs in northern Middle Park; and several hot springs on both sides of the Park Range, including Strawberry and Steamboat hot springs on the west side. Additional thermal springs include Waunita hot springs east of Gunnison, scattered thermal springs near Cañon City, and a few other isolated hot springs. Hot springs also are located at the Wagon Wheel Gap fluorspar district.

As described earlier, modern thermal springs are located in or near several of the rift-related fluorspar districts. The fluorspar deposits are relatively young, and rift-related faulting has continued into the Quaternary. Hypothetically, then, some of the associated thermal areas may represent long-lived thermal centers, or their modern locations may indicate optimal structural, heat-flow, and hydrogeologic conditions that have persisted for millions of years.

Cappa and Hemborg (1995) published chemical analyses, including fluorine, for many low-temperature geothermal areas in Colorado. Some thermal areas had multiple analyses from different springs in that system, whereas other areas had only one analysis. For the central Colorado study area and vicinity, fluorine values ranged from below detection limits (0.1 ppm) to 20 ppm (table 4). The average fluorine content from 112 analyses was a maximum of 4.2 ppm (maximum because 0.05 ppm was used as a proxy for the 19 samples with fluorine below detection limits), with a median of less than 2.9 ppm. Where multiple analyses came from the same thermal area, the fluorine values were consistent (Browns Canyon, with three samples with fluorine below detection limits) to widely varied (Cottonwood and Mount Princeton each have samples with 0.1 ppm, although the rest of the samples from each area contain more than 8 ppm).

Table 4. Fluorine in central Colorado thermal springs.

[Condensed from Cappa and Hemborg (1995). Type: HS, hot spring; W, warm spring. Sites with no pH measurement are shown with --. Fluorine (F) reported in mg/L (ppm). Locations are shown in figure 14]

Name	Latitude	Longitude	County	Type	pH	F
Antelope Warm Spring	37.743330	–107.037220	Mineral	HS	--	2.00
Birdsie Warm Spring	37.728330	–107.053610	Mineral	HS	9.2	0.05
Brands Ranch	40.704440	–106.534440	Jackson	W	6.4	0.05
Browns Canyon (Chimney Hill)	38.644440	–106.078060	Chaffee	W	--	0.05
Browns Canyon Warm Spring	38.653610	–106.053060	Chaffee	HS	8.0	0.05
Browns Grotto Warm Spring	38.636940	–106.073890	Chaffee	HS	7.0	0.05
Canon City Hot Springs	38.432220	–105.261390	Fremont	HS	6.2	1.50
Cebolla A, (Powderhorn)	38.273890	–107.098330	Gunnison	HS	6.7	4.28
Cebolla B, (Powderhorn)	38.273890	–107.098330	Gunnison	HS	--	5.80
Cebolla C, (Powderhorn)	38.273890	–107.098330	Gunnison	HS	--	4.60
Cement Creek Warm Spring	38.835000	–106.826110	Gunnison	HS	7.1	1.68
Chinaman Canyon	37.233330	–104.716670	Las Animas	HS	8.5	0.30
Clark Spring	38.258060	–104.609720	Pueblo	W	6.8	1.40
Cokedale	37.139720	–104.612500	Las Animas	HS	8.4	0.50
Conundrum Hot Springs	39.011940	–106.890830	Pitkin	HS	--	2.30
Cottonwood (Jump Steady)	38.812780	–106.222500	Chaffee	HS	7.6	10.68
Cottonwood (Merrifield Well)	38.809720	–106.222500	Chaffee	W	8.8	12.00
Cottonwood Hot Springs	38.812780	–106.225830	Chaffee	HS	7.6	14.33
Crowley Ranch Reserve	37.019440	–106.802780	Archuleta	W	6.8	0.05
Deganahl (Yampa)	40.135830	–106.961940	Routt	W	7.2	0.10
Deganahl-Watson Creek(Yampa)	40.145830	–106.955560	Routt	W	8.2	0.05
Desert Reef (Florence)	38.369170	–105.048610	Fremont	W	6.5	1.09
Dexter Spring	37.294720	–105.785000	Conejos	HS	7.9	0.05
Don K Ranch	38.170280	–105.013330	Pueblo	W	6.5	1.90
Dotsero	39.630560	–107.101390	Eagle	HS	7.1	0.55
Dotsero South	39.625280	–107.099440	Eagle	HS	6.9	0.27
Dutch Crowley	36.998890	–106.771940	Archuleta	W	6.9	0.50
Eldorado Springs A	39.932220	–105.279720	Boulder	W	6.9	0.20
Eldorado Springs B	39.932220	–105.279720	Boulder	HS	6.6	0.30
Eoff	37.191110	–106.995280	Archuleta	W	7.0	0.05
Florence	38.414720	–105.045280	Fremont	W	6.3	1.10
Fremont Natatorium	38.460280	–105.195830	Fremont	W	6.8	0.50
Glenwood Springs (Big Spring)	39.549440	–107.321670	Garfield	HS	6.3	2.25
Glenwood Springs (Drinking Spring)	39.549440	–107.321670	Garfield	HS	6.4	2.30
Glenwood Springs (Graves Springs)	39.555000	–107.335280	Garfield	HS	7.0	2.90
Glenwood Springs (Railroad Spring)	39.554440	–107.314170	Garfield	HS	6.8	2.25
Glenwood Springs (Spring A)	39.549440	–107.319440	Garfield	HS	6.3	2.20
Glenwood Springs (Spring B)	39.550560	–107.317780	Garfield	HS	6.8	2.15
Glenwood Springs (Vapor Cave)	39.550830	–107.319720	Garfield	HS	6.8	1.71
Grassy Creek (Hayden)	40.442500	–107.133610	Routt	W	7.5	0.30
Hartsel (Sping A)	39.018060	–105.794440	Park	HS	--	2.10
Hartsel (Spring B)	39.018060	–105.794440	Park	HS	6.7	1.95
Haystack Butte	40.100280	–105.238890	Boulder	W	8.0	4.40
Hooper Aquaculture Well	37.706110	–105.872220	Alamosa	W	8.8	0.05
Hot Sulphur Springs (Spring A)	40.074720	–106.111390	Grand	HS	6.9	11.60
Hot Sulphur Springs (Spring B)	40.074720	–106.111390	Grand	HS	6.7	12.00
Hot Sulphur Springs (Spring C)	40.074720	–106.111110	Grand	HS	7.0	10.75
Hot Sulphur Springs (Spring D)	40.074440	–106.111110	Grand	HS	7.1	9.10
Idaho Springs (Lodge Well)	39.739440	–105.511940	Clear Creek	W	6.9	3.50
Idaho Springs (Spring A)	39.738890	–105.511940	Clear Creek	HS	6.8	4.18
Idaho Springs (Spring B)	39.739170	–105.511940	Clear Creek	HS	--	4.80
Idaho Springs (Spring C)	39.738610	–105.512220	Clear Creek	HS	--	2.90
Jacks Mine	37.194440	–104.714440	Las Animas	HS	9.2	1.40
Lake City Airstrip	38.076390	–107.297220	Hinsdale	W	7.6	0.05
Lake San Cristobal	37.983330	–107.294440	Hinsdale	W	7.7	0.05
Lower Waunita Hot Springs (Spring A)	38.515830	–106.515560	Gunnison	HS	7.8	0.05

Table 4. Fluorine in central Colorado thermal springs.—Continued

[Condensed from Cappa and Hemborg (1995). Type: HS, hot spring; W, warm spring. Sites with no pH measurement are shown with --. Fluorine (F) reported in mg/L (ppm). Locations are shown in figure 14]

Name	Latitude	Longitude	County	Type	pH	F
Lower Waunita Hot Springs (Spring B)	38.515830	–106.515560	Gunnison	HS	7.9	16.00
Lower Waunita Hot Springs (Spring C)	38.515830	–106.516110	Gunnison	HS	7.8	0.05
Lower Waunita Hot Springs (Spring D)	38.516390	–106.516670	Gunnison	HS	7.8	20.00
Marigold	38.663060	–105.218610	Teller	W	7.0	0.05
McIntire Warm Spring	37.280560	–105.818330	Conejos	HS	8.2	0.05
MGP Well	37.133610	–104.808610	Las Animas	W	7.6	4.30
Mineral Hot Springs (Spring A)	38.168890	–105.918060	Saguache	W	6.8	3.95
Mineral Hot Springs (Spring C)	38.168330	–105.919720	Saguache	HS	--	4.20
Mineral Hot Springs (Spring D)	38.167780	–105.922220	Saguache	HS	6.8	3.93
Moffat	37.983890	–105.901390	Saguache	W	8.8	2.50
Mosca West	37.661390	–105.927220	Alamosa	W	8.2	3.00
Mt. Princeton Hot Spr. (Hortense Well)	38.732780	–106.174170	Chaffee	W	--	14.00
Mt. Princeton Hot Spr. (Hortense)	38.732500	–106.175000	Chaffee	HS	7.8	17.30
Mt. Princeton Hot Spr. (Spring A)	38.732780	–106.161390	Chaffee	HS	7.5	9.35
Mt. Princeton Hot Spr. (Spring F)	38.732780	–106.161390	Chaffee	HS	--	8.30
Mt. Princeton Hot Spr. (Woolmington)	38.723330	–106.177220	Chaffee	W	--	0.10
Mt. Princeton Hot Spr. (Wright Well E)	38.732220	–106.166110	Chaffee	W	--	10.00
Mt. Princeton Hot Spr. (Young Life Well)	38.732220	–106.175560	Chaffee	W	--	9.20
Mullenville (Rhodes) Warm Spings	39.163610	–106.066110	Park	HS	7.6	0.20
Pagosa Springs (Big Spring)	37.264440	–107.010280	Archuleta	HS	6.6	4.62
Pagosa Springs (Courthouse well)	37.266110	–107.010560	Archuleta	W	6.5	4.50
Pagosa Springs (Spa Motel well)	37.265280	–107.009720	Archuleta	W	6.5	4.40
Papeton well (Colorado Springs)	38.878890	–104.796110	El Paso	W	6.7	0.30
Penny Hot Springs (Granges Spring)	39.230560	–107.226110	Pitkin	HS	7.2	3.33
Poncha Springs (Spring A)	38.496940	–106.076940	Chaffee	HS	7.7	12.00
Poncha Springs (Spring B)	38.496940	–106.076940	Chaffee	HS	7.5	12.00
Poncha Springs (Spring C)	38.498060	–106.076110	Chaffee	HS	7.7	11.23
Rainbow	37.509170	–106.874440	Mineral	HS	--	2.20
Ranger	38.813060	–106.874440	Gunnison	HS	7.0	1.35
Routt [aka Strawberry] (Spring A)	40.559440	–106.850000	Routt	HS	7.8	16.30
Routt [aka Strawberry] (Spring B)	40.559440	–106.850000	Routt	HS	7.1	17.00
Sand Dune Pool	37.778330	–105.855560	Saguache	W	8.3	5.90
Sarcillo Canyon Well	37.161110	–104.782220	Las Animas	W	8.0	0.50
Shaws	37.750280	–106.316940	Saguache	HS	9.0	3.00
Smith Canyon Spring	37.172220	–104.751940	Las Animas	HS	7.5	0.30
South Canyon Hot Springs (Spring A)	39.553060	–107.411110	Garfield	HS	7.3	3.68
South Canyon Hot Springs (Spring B)	39.553060	–107.411110	Garfield	HS	7.1	4.00
Splashland	37.488610	–105.858610	Alamosa	W	8.3	4.20
Steamboat Springs (Heart Spring)	40.482780	–106.826940	Routt	HS	8.1	5.30
Steamboat Springs (Steamboat Spring)	40.488890	–106.840560	Routt	HS	6.7	2.90
Steamboat Springs (Sulphur Cave Spring)	40.484170	–106.839440	Routt	HS	6.5	3.00
Stinking Springs	37.035000	–106.807780	Archuleta	HS	7.0	0.60
Swissvale (Spring A)	38.480560	–105.890560	Fremont	HS	7.0	0.05
Swissvale (Spring F)	38.480560	–105.890560	Fremont	HS	7.0	0.05
Sylvester Gulch Warm Spring	38.913330	–107.439440	Gunnison	HS	7.1	0.20
Two Mile Road	37.546390	–105.848060	Alamosa	W	7.5	0.05
Valley View (Orient) Hot Spr. (Spring A)	38.192220	–105.813610	Saguache	HS	7.2	0.45
Valley View (Orient) Hot Spr. (Spring B)	38.191940	–105.813890	Saguache	HS	7.1	0.40
Valley View (Orient) Hot Spr. (Spring D)	38.191110	–105.809170	Saguache	HS	7.3	0.30
Wagon Wheel Gap (4UR Ranch Spring)	37.748610	–106.831110	Mineral	HS	6.9	6.97
Wagon Wheel Gap (CFI Spring)	37.748330	–106.830560	Mineral	HS	6.5	7.15
Waunita Hot Springs (Spring A)	38.513890	–106.507500	Gunnison	HS	8.3	17.00
Waunita Hot Springs (Spring C)	38.513890	–106.507500	Gunnison	HS	8.3	18.50
Waunita Hot Springs (Spring D)	38.513890	–106.507500	Gunnison	HS	8.3	18.00
Wellsville	38.486110	–105.912500	Fremont	HS	7.2	0.67
Wet Canyon	37.234720	–104.891940	Las Animas	HS	8.2	0.50

For the three locations that have both a modern spring and a fluorspar deposit, the correlation between fluorine content in the springs and the presence of a fluorspar deposit is variable. Modern Poncha Springs, located along the same set of faults that host the fluorspar veins (figs. 4, 14), contains 11 to 12 ppm fluorine, some of the highest values in the dataset. Some, and possibly a lot, of the fluorine may have been leached from the fluorspar veins. In contrast, analyses of springs at Browns Canyon, as noted previously, measured virtually no fluorine, even though those springs, like Poncha Springs, are immediately adjacent to fluorspar mines and formed along the faults that host the veins. The springs at Wagon Wheel Gap are along the western extensions of the faults that host the fluorspar veins. Two samples of the springs at Wagon Wheel Gap had about 7 ppm fluorine, which was above the average and median for the dataset. Vanderwilt and others (1947) reported 2,200 ppm (“0.22 percent”) fluorine from these springs, which is unreasonably high given the values in the more recent database. Emmons and Larsen (1913) suggested that the meteoric waters in the springs may have flowed through the rocks that host the fluorspar veins, implying fluorine recycling without addressing the fluorine specifically.

Besides Poncha Springs, four thermal centers contain all of the hot springs with more than 10 ppm fluorine. These include Waunita, Cottonwood, Mount Princeton, and Strawberry hot springs (fig. 14). Waunita is anomalous as a thermal center because it is not along the rift, although, like much of the rift, it is an area of elevated heat flow (Berkman and Carroll, 2007). Of the seven samples collected there, five contained more than 16 ppm fluorine and two had fluorine below detection limits. This area is adjacent to the middle Tertiary Tomichi Dome, which has been considered to be a very high-level Climax system (Steve Ludington, U.S. Geological Survey, written commun., 2009), has associated fluorite-bearing veins and topaz, and may have contributed fluorine to circulating meteoric waters. Despite several studies, the origin of these hot springs is enigmatic (Zacharakis, 1981).

Cottonwood and Mount Princeton hot springs are situated along the eastern base of the Sawatch Range, and each contained high amounts of fluorine (table 4). These two centers are associated with an area of extremely high heat flow (Berkman and Carroll, 2007). They also are located within and near the fluorine-rich Mount Antero Granite, which may have provided fluorine to the geothermal system.

The two samples from Strawberry hot springs, north of Steamboat Springs, contained 16 to 17 ppm fluorine. This area underlies the Mesoproterozoic, fluorine-rich Mount Ethel pluton. As was proposed for the Crystal fluorspar district (Snyder and Hedge, 1978), meteoric water circulation through the pluton also may have contributed fluorine to the modern hot springs. Steamboat Springs had about 3 to 5 ppm fluorine and does not overlie the pluton. Only two other hot springs—Ranger and Cement Creek north of Gunnison—are within or near Mesoproterozoic granites, and those springs contained less than 2 ppm fluorine.

The thermal areas in the middle of the San Luis Valley contained as much as 5.9 ppm fluorine, but most of the samples contained much less fluorine. The San Luis Valley has very low heat flow (Berkman and Carroll, 2007), and the meteoric waters in the springs pass through considerable amounts of upper Tertiary sedimentary units before reaching the surface. The east side of the valley along the base of the Sangre de Cristo Mountains has above-average heat flow, but Valley View hot springs, the only measured springs along that zone, contained less than 1 ppm fluorine.

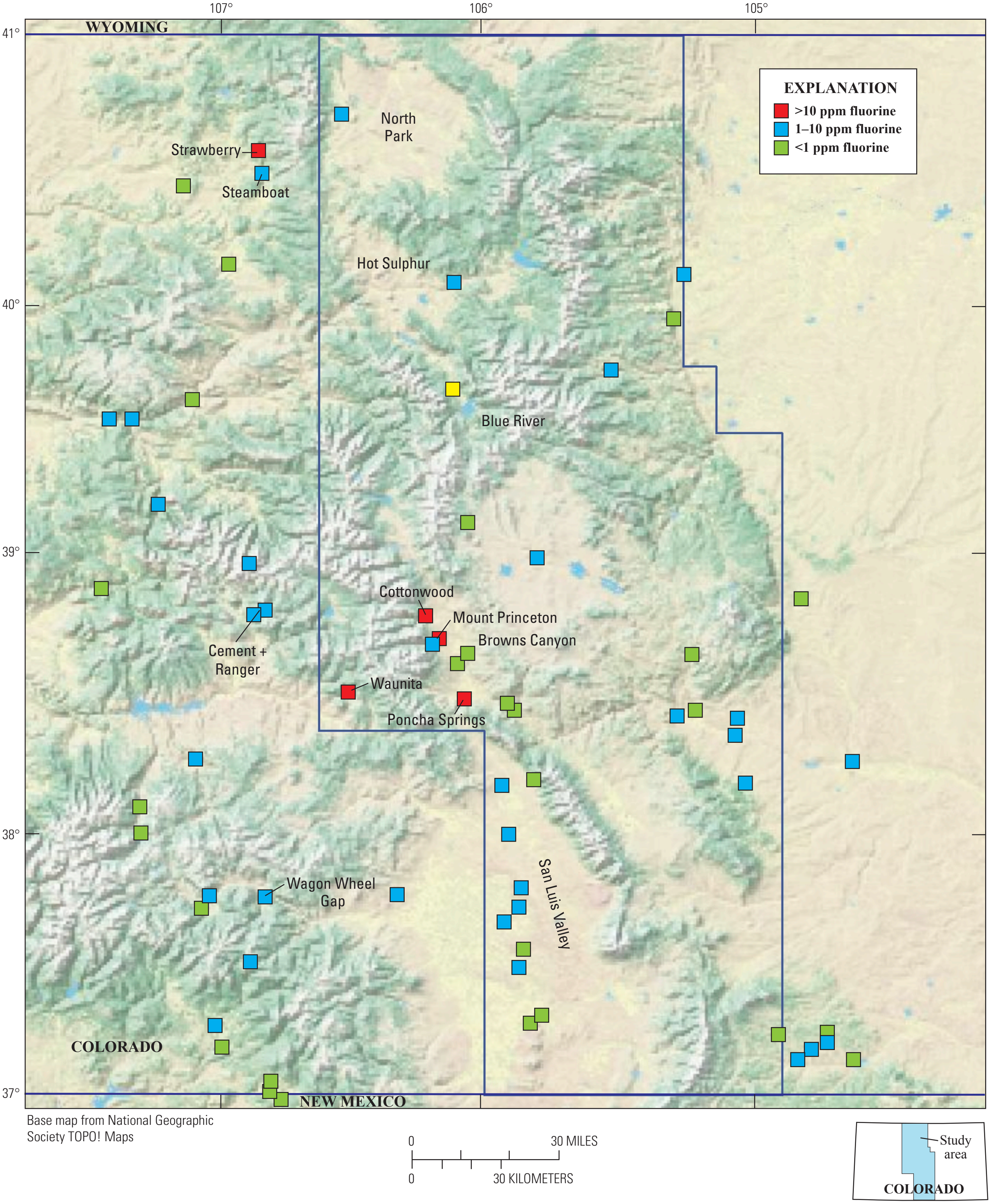
The USGS MRDS database contains one location along the Blue River near Silverthorne (site DC05604 [Frisco] in table 2; fig. 14). There, colloform fluorite reportedly coats cobbles in Quaternary glacial outwash, suggesting young, local geothermal activity.

Limited isotopic and chemical data indicate that most of the water in central Colorado’s thermal springs has a meteoric origin. The water circulates to various depths and through diverse rocks, deriving heat and elements along the circulation path before ascending and emerging as hot springs. High heat flow associated with extension could provide the necessary heat but not necessarily the fluorine. Most of the fluorine in these thermal waters may have been derived from rocks along the flow paths. However, as noted in several sections of this report, fluorine was a common constituent in some late Tertiary magmatic systems. Isotopic studies of fluorspar deposits along the rift in New Mexico point to an asthenospheric source for the fluorine (Partey and others, 2005), even though the hydrothermal fluids were largely connate to meteoric. By analogy, a similar process could have introduced magmatic fluorine to some of the meteoric water-dominated thermal centers in Colorado.

Topaz Deposits

Topaz ($Al_2SiO_4[F, OH]_2$) is a fluorine-bearing mineral that formed in a variety of magmatic and hydrothermal environments in central Colorado. In most places, topaz is a minor, non-economic component of a rock or mineral deposit. In some places, the quality of the topaz has attracted mineral collectors, who have mined it on a selective basis. Topaz is resistant to near-surface chemical weathering, and topaz crystals and grains commonly remain as a surficial residue when the host rock weathers and disaggregates. Sediment in streams draining these areas also can contain modest amounts of topaz grains.

A tabular summary of known topaz localities is provided in table 5, and locations are shown in figure 15. Much of the data for that table was derived from descriptions in Eckel (1997) and other publications, from information in the MRDS database, and from various descriptions that were available on the Internet, including the Web sites of magazines and clubs devoted to mineral collecting. In addition to the name and location of the deposit, the table lists the type of deposit in which the topaz is present and the age of the deposit or occurrence. The locations of the deposits vary in accuracy. Some references provided very accurate latitude and longitude data, whereas others gave the location only as a few miles in some direction from a known landmark, such as a named mountain or town. In most cases, the online Web site Google



Base map from National Geographic Society TOPO! Maps

Figure 14. Locations of thermal springs in central Colorado where the water has been analyzed for fluorine. Yellow square is a thermal spring with no chemical analyses but which has deposited fluorite. Many locations shown on the map have multiple springs and analyses; see table 4 for the list of all springs and analyses. Compare this figure with the map of the fluorspar deposits in the study area (fig. 4), especially those related to the Rio Grande rift. Refer to figure 1 for general geographic locations.

Table 5. Locations of topaz-bearing rocks in the central Colorado study area.

No.	Site	County	Latitude	Longitude	General deposit type	General deposit age
TAW32	Alma district	Park	39.310300	–106.116800	vein/replacement	Oligocene (27 Ma)
TAW25	Chalk Mountain	Lake/Summit/Eagle	39.381600	–106.206300	Climax Mo	Oligocene (27 Ma)
TAW23	Climax mine	Lake	39.370400	–106.166500	Climax Mo	Oligocene (27 Ma)
TAW06	Henderson mine	Clear Creek	39.757700	–105.841700	Climax Mo	Oligocene (28 Ma)
TAW01	Mt. Antero	Chaffee	38.649200	–106.274000	pegmatite	Oligocene (28 Ma)
TAW26	Specimen Mountain	Larimer	40.444900	–105.807800	volcanic	Oligocene (28 Ma)
TAW02	Ruby Mountain	Chaffee	38.750000	–106.070000	volcanic	Oligocene (30 Ma)
TAW08	Rosita Hills	Custer	38.127300	–105.335000	volcanic	Eocene (35 Ma)
TAW24	St. Kevin district	Lake	39.277500	–106.406600	porphyry Cu-Mo	Eocene (36 Ma)
TAW19	Tomichi Dome	Gunnison	38.490000	–106.530000	porphyry Cu-Mo	Eocene (38 Ma)
TAW03	Winfield	Chaffee	38.976500	–106.433400	Climax Mo	Eocene (39 Ma)
TAW31	Badger Flats	Park	38.970800	–105.369400	pegmatite	Mesoproterozoic (1.1 Ga suite)
TAW14	Bear Creek Canyon	El Paso	38.805300	–104.904800	pegmatite	Mesoproterozoic (1.1 Ga suite)
TAW35	Catherine No. 1 & Susy Belle claims	Jefferson	39.381400	–105.213100	pegmatite	Mesoproterozoic (1.1 Ga suite)
TAW40	Crystal Park	El Paso	38.831667	–104.931389	pegmatite	Mesoproterozoic (1.1 Ga suite)
TAW33	Crystal Peak (Florissant)	Teller	38.995556	–105.288611	pegmatite	Mesoproterozoic (1.1 Ga suite)
TAW09	Devils Head	Douglas	39.251400	–105.119980	pegmatite	Mesoproterozoic (1.1 Ga suite)
TAW34	Glen Cove area	Teller	38.868200	–105.074400	pegmatite	Mesoproterozoic (1.1 Ga suite)
TAW30	Harris Park	Park	39.530100	–105.469900	pegmatite	Mesoproterozoic (1.1 Ga suite)
TAW10	Long Hollow	Douglas	39.212300	–105.064700	pegmatite	Mesoproterozoic (1.1 Ga suite)
TAW11	Pine Nook	Douglas	39.383600	–105.072200	pegmatite	Mesoproterozoic (1.1 Ga suite)
TAW21	South Platte pegmatite district	Jefferson	39.395600	–105.214400	pegmatite	Mesoproterozoic (1.1 Ga suite)
TAW15	Stove Mountain (N of St. Peters Dome)	El Paso	38.776100	–104.900800	pegmatite	Mesoproterozoic (1.1 Ga suite)
TAW29	Tarryall Mountains, Aspen Lode claims	Park	39.339722	–105.394444	pegmatite	Mesoproterozoic (1.1 Ga suite)
TAW22	Wigwam Creek area	Jefferson	39.250000	–105.350000	pegmatite	Mesoproterozoic (1.1 Ga suite)
TAW27	Red Head	Larimer	40.883300	–105.466700	pegmatite	Mesoproterozoic (1.4 Ga suite)
TAW39	Comet Group	Gunnison	38.992222	–106.733056	pegmatite	Meso- or Paleoproterozoic
TAW04	Beaver Brook	Clear Creek	39.709444	–105.399167	pegmatite	Paleoproterozoic
TAW38	Brown Derby/White Spar/ Opportunity	Gunnison	38.551111	–106.579444	pegmatite	Paleoproterozoic
TAW28	Crystal Mountain district	Larimer	40.538600	–105.432000	pegmatite	Paleoproterozoic
TAW17	Devils Hole	Clear Creek	39.690000	–105.410000	metamorphic	Paleoproterozoic
TAW12	Holy Cross Mountain	Eagle	39.470000	–106.480000	pegmatite	Paleoproterozoic
TAW36	Jacobson Ranch claim	Clear Creek	39.702222	–105.398056	pegmatite	Paleoproterozoic
TAW20	Medicine Bow Mountains	Jackson/Larimer	40.750000	–106.000000	pegmatite?	Paleoproterozoic
TAW16	Mitchell Gulch	Fremont	38.447400	–105.792300	metamorphic	Paleoproterozoic
TAW18	Ohio City district	Gunnison	38.511360	–106.668300	pegmatite	Paleoproterozoic
TAW05	Upper Bear Creek	Clear Creek	39.625200	–105.456900	pegmatite	Paleoproterozoic

Earth was used to visually find disturbed ground in the general vicinity of those vague locations that might indicate a prospect. This proved successful in most, but not all, cases.

Topaz in central Colorado is associated primarily with Proterozoic igneous rocks, and lesser amounts are present in or near Tertiary plutonic and volcanic rocks. Brief descriptions of these settings are provided here.

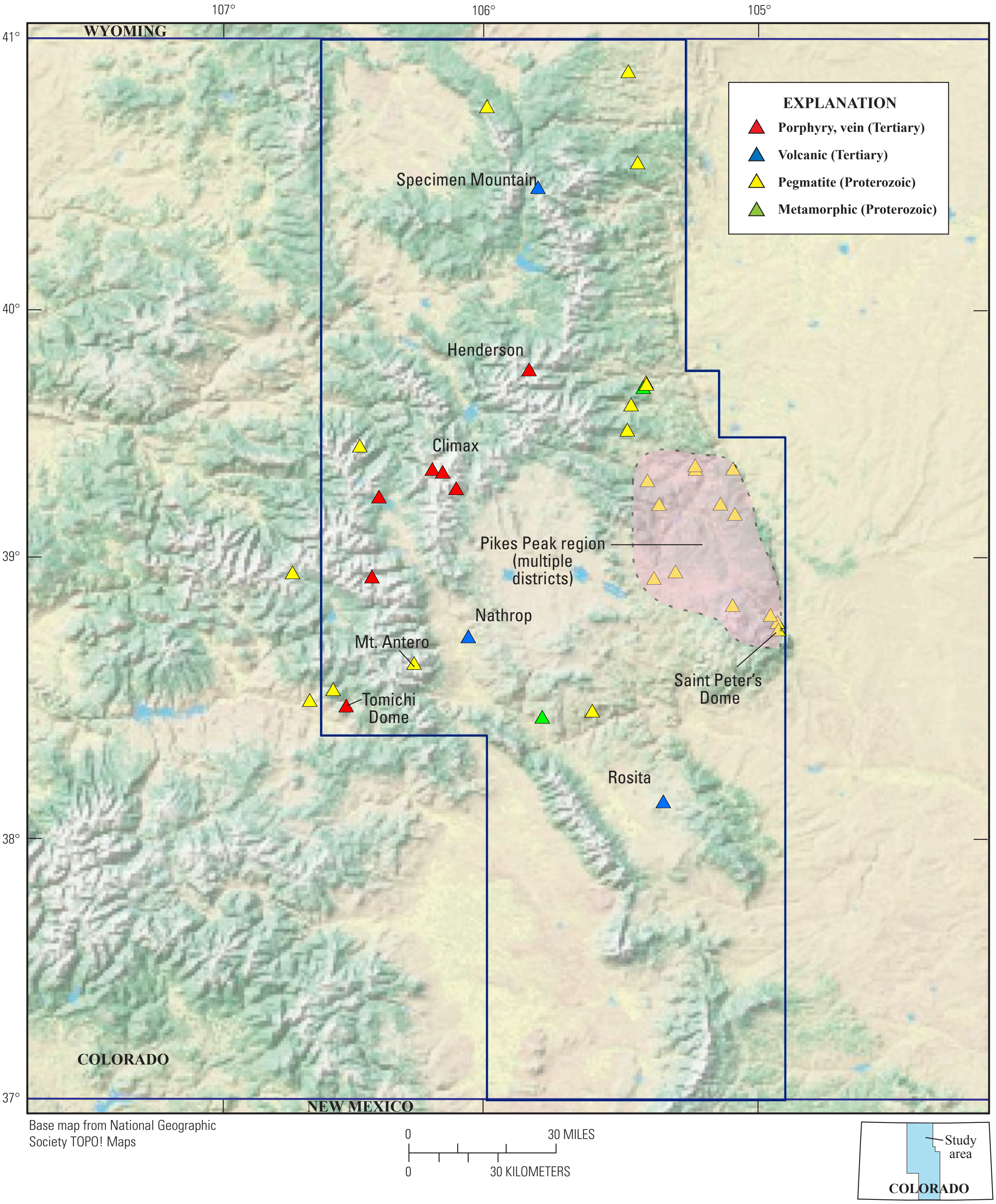
Topaz in Proterozoic Pegmatites and Skarns

Most of the Proterozoic-age, topaz-bearing rocks are pegmatites associated with the Pikes Peak batholith intrusive complex (figs. 6, 15), as described earlier for fluorite-bearing pegmatites. Some gem-quality topaz occurs in fluorite-bearing pegmatites, and some is present in pegmatites that do not contain fluorite, although the fluorine contents clearly must be elevated. The genesis of these pegmatites is described previously in this report.

Topaz also has been reported from small, scattered pegmatites associated with the 1.4- and 1.7-Ga plutonic suites. In addition, a few Paleoproterozoic metamorphic rocks that have fluorine-rich protoliths (high-fluorine rhyolites or pelitic sedimentary rocks) contain topaz along with other metamorphic minerals.

Topaz in Tertiary Igneous Systems

Within the central Colorado study area, topaz is an accessory mineral in high-silica granite intrusives (at Climax, Henderson, Winfield, Mount Antero), rhyolite volcanic flow units (at Nathrop, Silver Cliff), and other shallow, rhyolitic intrusive rocks (at Specimen Mountain, Tomichi Dome) (fig. 15). Some of the topaz-producing magmas were high-silica Climax-type systems, and others were alkali- and silica-rich end members of calc-alkaline systems. The topaz is present as part of the igneous rock and in late-stage vugs and cavities in some of the plutons and rhyolite



Base map from National Geographic Society TOPO! Maps

Figure 15. Locations of notable topaz occurrences in the central Colorado study area. See table 5 for a list of the topaz occurrences. Refer to figure 1 for general geographic locations.

flow units. At Henderson, Seedorff and Einaudi (2004b) showed that topaz formed when the fluorine activity of the magmatic-hydrothermal system increased with a decrease in temperature and pressure. At the Sweet Home mine in the Alma district, which was related to a 27-Ma Climax-related felsic intrusive body in Proterozoic rocks, the topaz occurs in replacement vugs and veins along with other hydrothermal minerals (Bartos and others, 2007). As summarized by Christiansen and others (1986, 2007), topaz-forming igneous systems formed at the onset of extensional faulting and during and after the related transition between calc-alkaline and bimodal volcanism described previously in this report. Fluorine levels in the central Colorado host rocks range from 1,400 to 5,500 ppm (Christiansen and others, 1986). The average fluorine content of 927 Cenozoic igneous rocks of all compositions in the USGS rock database is 1,080 ppm. Where the annotations in the database indicate the general rock composition, the felsic rocks in that database tend to contain somewhat higher fluorine levels than the more mafic rocks. Therefore, the fluorine contents of the topaz-bearing rocks are not atypical of the rest of the felsic part of the database.

Weathering of Fluorine-Bearing Rocks

Weathering and erosion of rocks that contain fluorine can redistribute the fluorine into soil, stream sediment, and water. The amount of redistribution depends on the original site of the fluorine in the source minerals and rocks and the processes, which can range from large-scale to microscopic, that can remove the fluorine and transport it elsewhere. Fluorine has been used for decades to prevent tooth decay, but it also may have adverse health effects at higher concentrations, leading to recent debates about the merits of adding fluorine to drinking water. Therefore, this final step between fluorine-bearing rocks and the hydrosphere may have implications for the levels of naturally occurring fluorine in central Colorado water.

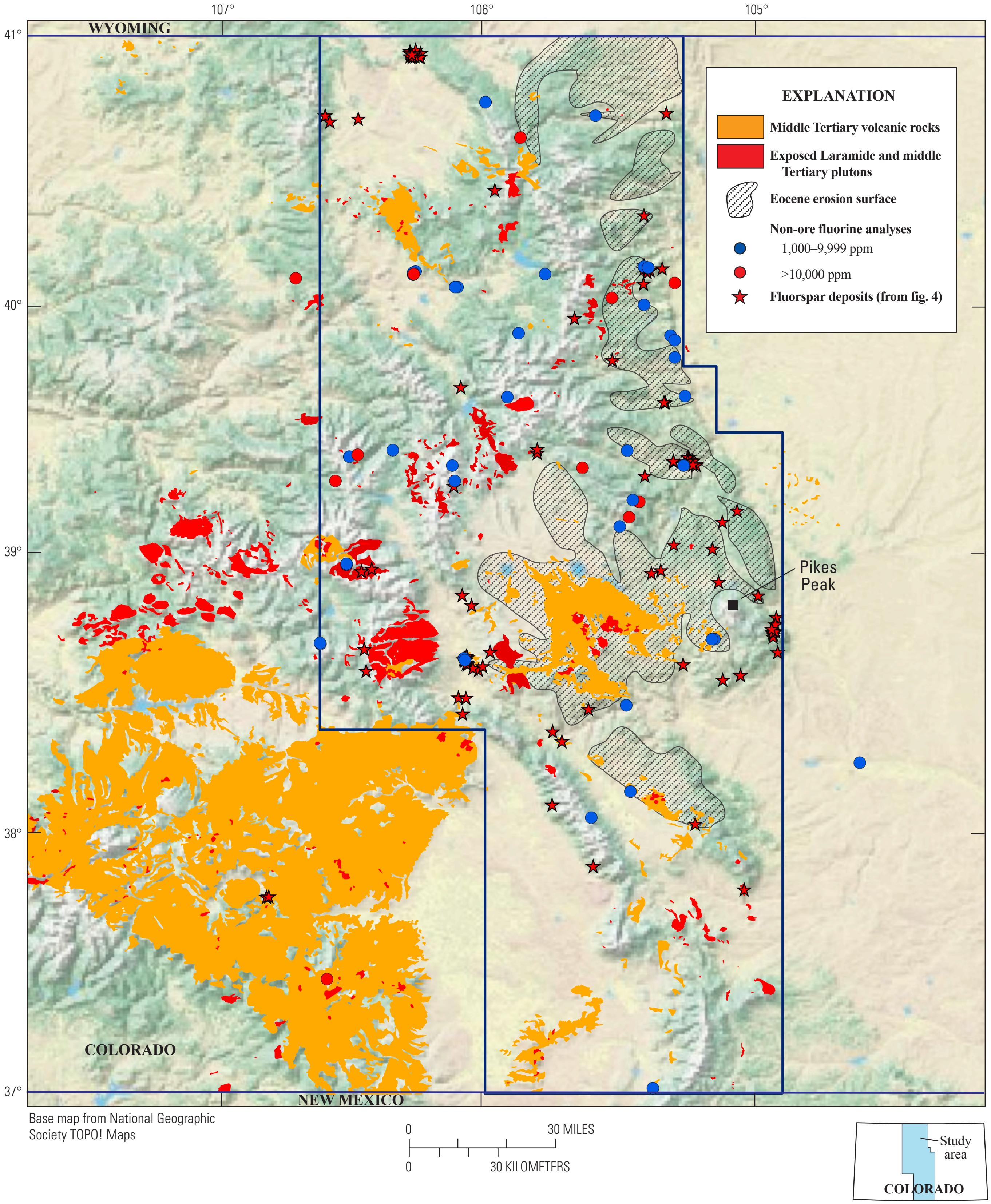
For the most part, fluorine in central Colorado occurs as a trace element in minerals. Less commonly, fluorine is a major component of fluorite. Minerals with trace amounts of fluorine include those that are relatively more susceptible to weathering, such as biotite and hornblende, and those that are relatively more resistant to weathering, such as apatite and topaz. Weathering-resistant minerals can accumulate in soils and be incorporated into stream sediment. Minerals that are susceptible to weathering can break down, either partly or completely, by complex chemical and mechanical processes; their elemental components either remain in place as new mineral phases (such as iron oxides or hydroxides) or are removed from the original site. Independent variables, such as climate, microbial and other organic activity, pH of the soil and water, and time are important contributors to the weathering process. Thus, depending on the combination of processes and the source rocks, fluorine-bearing minerals such as biotite can break down and release fluorine in situ, or the host mineral, even if it is partly weathered, can be transported along with its contained fluorine into other settings.

Much of the central Colorado landscape is the product of multiple Tertiary and Quaternary tectonic and geomorphic events. Broad erosion surfaces developed over much of the region during the Cenozoic, shedding sediments into basins and exposing uplands to long-term weathering. Two of the major erosion surfaces formed in the late Eocene and in the Neogene. The Eocene erosion surface is well defined in south-central and southwestern Colorado and in the Front Range. In the San Juan Mountains, the erosion surface largely remains buried by middle Tertiary volcanic rocks, and other exposures of middle Tertiary volcanic rocks throughout central Colorado, such as the Thirtynine Mile volcanic field, also help delineate the erosion surface (fig. 16). In those covered areas, the formation of the erosion surface exposed fluorine-bearing rocks to weathering only for the period (about 40 to 30 Ma) between erosion and volcanism, and those weathered rocks have been protected by the volcanic cover since that time.

In contrast, many areas of the Front Range have remained exposed since the formation of the Eocene erosion surface (fig. 16). As a result, weathering of rocks beneath the Eocene surface can be very deep, such as in the Pikes Peak region and other areas; in some parts of the Front Range, the full weathering profile can reach a depth of 8 m or more (Campbell and Dethier, 2005). The Neogene erosion surface was incised into the Eocene surface in the Front Range and Wet Mountains. With a lower base level, this incision has induced erosion of the higher, older surface as streams migrated and incised headward with time (Scott, 1975a; Taylor, 1975; Leonard and Sak, 2005). This process continued into the Quaternary, with added erosion and weathering during periods of glaciation.

As a result, regions such as the Front Range have broad, low-relief areas that have been exposed to weathering for tens of millions of years and narrow, steep-sided valleys that are relatively juvenile with minimal weathering. The Front Range contains many areas with high background levels of fluorine in the country rock, as well as several large fluorspar or fluorite deposits. Combined with the dual geomorphic history, the weathering history of these fluorine-bearing rocks and the redistribution of fluorine into soils, streams, and water in the Front Range likely have been very complex. For example, fluorine-bearing rocks of the Pikes Peak Granite that underlie the Eocene erosion surface have abundant microjoints, are deeply weathered, and have a locally thick cover of grus. Groundwater in those areas contains high levels of fluorine (Wanty and others, 1992). Leaching experiments on fresh equivalent rocks released considerably less fluorine, a situation that may proxy for less-weathered units found along the more juvenile valleys.

Local weathering differences also affected fluorine-rich rocks. For example, the quartz-fluorite-cemented breccia reefs in the east-central Front Range (fig. 10) have weathered differently depending on the host rocks: those in metamorphic rocks have been more resistant to weathering and form narrow spines (thus the term “reefs”), and those in the plutonic rocks are somewhat less resistant to erosion than the granitic rocks and thus occur in swales (Wallace, 1982). As with the Pikes Peak area, Tertiary erosion produced older erosion



Base map from National Geographic Society TOPO! Maps

Figure 16. Features related to Cenozoic erosion surfaces and the locations of fluorspar deposits and samples with greater than 1,000 parts per million fluorine. See the text for a discussion of the various features. The Front Range erosion surface was generalized from Kelley and Chapin (2004); the extent of the Cenozoic igneous rocks was generalized from Tweto (1979a).

surfaces and younger incised valleys. Thus, erosion of fluorite and transport of fluorine from the breccia reefs has differed depending on the host lithology and the location relative to the pediments and streams.

Elsewhere, in the northern Wet Mountains, the late Eocene erosion beveled the landscape (Taylor, 1975; Epis and others, 1976). The fluorine-bearing Cambrian intrusive rocks were exposed to weathering until Oligocene volcanic rocks, including those of the fluorine-rich Rosita–Silver Cliff volcanic centers, buried the paleosurface. Renewed Neogene erosion removed part of the volcanic cover and reexposed the Cambrian rocks, inducing a second period of weathering along the new paleosurface. Progressive upstream cutting of the upper Arkansas River drainage system has progressively incised into the Neogene erosion surface, and the materials formed during both periods of weathering are locally being transported toward the Arkansas River.

As described previously, some Laramide and younger Tertiary intrusive rocks contain significant amounts of fluorine. The plutons were intruded to various depths, but combinations of uplift and erosion have exposed them at the surface (fig. 16). Depending on the age of intrusion and the timing of uplift and erosion, the plutons have been subjected to long to relatively short periods of weathering. In the Sawatch Range, the large, Laramide Twin Lakes stock was already exposed when the 34-Ma Grizzly Peak caldera erupted through its western margin (Fridrich and others, 1991), and much of the stock has remained exposed since that time. Just to the south, however, the Mount Aetna cauldron and related plutonic rocks formed at about the same time as the Grizzly Peak caldera system, but the current deep exposure of the cauldron and plutons largely is a product of Neogene uplift of the Sawatch Range (Shannon and others, 1987; Shannon, 1988). Thus, fluorine-bearing rocks in two nearby areas have very different erosional and weathering histories.

Regional Implications of the Data

Is Central Colorado a Fluorine-Rich Province?

With abundant fluorspar deposits, fluorite in many rock units of various ages, and fluorine in modern thermal systems, it is tempting to call central Colorado a fluorine-rich province. In reality, the region has an extremely skewed distribution of fluorine, with 64 percent of the samples containing less than 1,000 ppm fluorine and less than 4 percent of the samples containing more than 1 percent (10,000 ppm) fluorine. Without a doubt, various geologic processes have concentrated fluorine in specific locations or rock units, such as at the Jamestown and other fluorspar districts or in the Pikes Peak Granite, but those locations and areas spatially comprise a small part of the region. In contrast, the fluorine contents of the vast majority of the rocks in the area are very similar to the average crustal abundance, both for the crust as a whole and for the various general lithologic categories. By that measure, the region as a whole should be considered average in terms of its overall fluorine content.

Fluorine Has Been Recycled and Added Through Magmatic Processes

Almost all of the fluorine-rich rocks are related in some way or another to magmatic processes, and those magmatic processes are a 1.8-Ga-long combination of crustal recycling and additions of new mantle materials. Isotopic studies have shown that many central Colorado magmas, ranging in age from 1.7-Ga batholiths to Tertiary plutons, were derived to a large extent from partial melting and assimilation of older lower crust (Peterman and others, 1968; Barker and others, 1975; White and others, 1981; Armbrustmacher and Hedge, 1982; Hannah and Stein, 1986; Bookstrom, 1990; Stein and Crock, 1990; Smith and others, 1999; Kelley and Ludington, 2002). This lower crust started out as a mixture of mantle-derived arc volcanic rocks and volcaniclastic and epiclastic sedimentary rocks. With each successive tapping of the lower crust for melts, the crust evolved and thus was able to contribute different components to new melts through time.

The ascent of mantle- and subduction-derived mafic magmas into the lower crust commonly induced partial melting of the crust. Mantle materials were incorporated into those magmas to varying degrees or, in the cases of the Cambrian alkalic magmas and certain phases of the Pikes Peak batholith, were the principal magma source. Mantle materials, including fluorine, that did not become involved in the magmas remained at the base of or within the crust to be available for later melting. Slab devolatilization retained fluorine in the slab, but later melting of the slab (or its residue in the asthenosphere) could have produced fluorine-rich melts that, when they ascended into the lower crust, could mix with crustal melts (Straub and Layne, 2003). Christiansen and others (1986, 2007) argued that significant amounts of fluorine in topaz rhyolites, such as those exposed at Rosita and Nathrop, could be the product of fractional crystallization of similar mantle-derived basaltic magmas. For example, fluorine is less susceptible to mobilization during initial partial melting than other volatiles and ligands (Christiansen and others, 2007), and the residual starting crust may have become relatively enriched in fluorine through time.

The evolution of the magmas as they ascend influences the behavior of fluorine, especially near the end stages of crystallization. In granitic melts, the fluorine content during magma evolution depends on the precipitation or not of aluminum phases and the total amount of aluminum in the melt (Dolejš and Baker, 2007). At lower aluminum concentrations, precipitation of aluminosilicate minerals, such as micas, can consume all available fluorine. This, in part, may account for the paucity of magmatic fluorite or topaz in the 1.7-Ga plutons. If aluminum remains in excess, either due to a high starting content or the lack of early precipitation of aluminosilicate minerals relative to phases like quartz and alkali feldspar, then enough fluorine remains to become concentrated in the residual fluids, theoretically reaching more than 30 weight percent (Dolejš and Baker, 2007). These residual fluids can form late-stage fluorite, topaz, or cryolite (rare in Colorado), depending on the compositions and concentrations of alkalis and the site-specific hydromagmatic processes (Seedorff and Einaudi, 2004b; Dolejš and Baker, 2006). This scenario might pertain to the 1.4- and 1.1-Ga plutons, as described by Snyder and Hedge (1978) and Barker and others (1975). The late-stage segregation of volatile phases such as fluorine also may have been a contributing factor in the formation of fluorite- and topaz-rich granites and fluorite in related hydrothermal and hydromagmatic systems, such as in the middle Tertiary Mount Antero and Climax-type granites.

The fluorine in the rift-related fluorspar deposits has several possible sources: recycled from Proterozoic igneous rocks and Phanerozoic sedimentary rocks, and directly from rift-related magmas. Some of the fluorspar districts in central Colorado are spatially associated with fluorine-bearing 1.4-Ga granites, such as the Mount Ethel pluton near the North Park fluorspar deposits. Not all fluorspar districts, however, have this association, and no isotopic studies have been done to indicate if the plutons contributed the fluorine. The presence of hydrocarbons in fluid inclusions in fluorite in the Delaney Butte district indicate, at minimum, the migration of the hydrothermal fluids through petroliferous Mesozoic rocks in the area. Isotopic (Pb, S, Cl, Nd, Sr) studies at a variety of fluorspar deposit types in New Mexico indicate both fluid interaction with both Proterozoic granites and Phanerozoic sedimentary rocks and derivation of fluorine from asthenospheric, presumably rift-related, magmas at depth (Hill and others, 2000; Partey and others, 2005). On basis of those studies, it is very possible that at least some of the fluorine in central Colorado fluorspar deposits was recycled from older rocks in the area.

Postmineralization Processes Affected Fluorite-Rich Areas

Most fluorspar deposits and major fluorite occurrences formed in epithermal environments or the upper parts of plutonic systems, generally within a kilometer of the surface. That near-surface environment is very susceptible to erosion during uplift (Wallace and others, 2004). Given the complex and protracted uplift and erosional history of central Colorado, almost all of the preserved epithermal fluorspar deposits are Miocene or younger in age largely because young deposits have had less of a chance to be eroded. Conversely, the nearly complete erosion of the Laramide paleosurface has removed any near-surface mineral deposits of that age. Exploration for and assessments of epithermal mineral deposits should focus initially on areas where the coeval paleosurface has been preserved, either through lack of erosion or because it was protected by younger cover. For example, the extents of the late Eocene paleosurface and the volcanic rocks that cover that paleosurface (fig. 16) could be used to define areas where epithermal mineral deposits might have been preserved. Similarly, the abundance of fluorspar deposits along the Rio Grande rift empirically show that tectonically young areas with little erosion are better places to find a fluorspar deposit.

Preservation of the upper parts of magmatic systems depends on the depth of formation of that environment and the amount of subsequent uplift and erosion. As previously described, a period of major crustal erosion took place between the emplacement of the 1.7- and 1.4-Ga plutons. No fluorite is associated with the 1.7-Ga suite, but most of the 1.4-Ga granites have fluorite in their upper or peripheral margins. It is possible that some of the 1.7-Ga plutons had primary fluorite in their upper margins that was removed during the postintrusion period of erosion. Similarly, in the Pikes Peak batholith, the three main phases are now exposed at different levels of emplacement.

Most of the Laramide and Tertiary plutons were emplaced at relatively shallow depths, and the areas where those plutons were emplaced underwent multiple episodes of uplift and erosion. Laramide plutons in the Sawatch Range and elsewhere had been uplifted and eroded by the late Eocene. If, as with other magmatic systems, any fluorine in the Laramide magmas had been concentrated in the upper parts of the plutons, then postemplacement erosion could in part explain the relative paucity of magmatic fluorite seen today in that suite of plutons. Postuplift, middle Tertiary plutons were emplaced at similarly shallow levels, relative to the middle Tertiary paleosurface, and they were emplaced next to the deeper levels of adjacent Laramide plutons. Although uplift and erosion have affected those younger intrusive bodies, the upper, fluorite-bearing parts of many plutons are still preserved. As a result, fluorite-bearing plutons appear to have intruded fluorite-poor older plutons, where in reality the level of exposure may have had a greater influence on the amount of fluorite seen today. In some cases, such as at Henderson and several other Climax-type molybdenum deposits, many fluorite-bearing phases of the intrusive system remain buried even today.

Using High-Fluorine Areas as Exploration Guides

As shown earlier, fluorite is associated with a wide variety of mineral-deposit types. Because fluorite commonly forms in the peripheral zones of mineralizing systems, the presence of abundant fluorite and(or) high fluorine levels possibly could be used as an exploration guide for concealed deposits of various metals. Several of the fluorspar deposits in central Colorado contain weakly anomalous amounts of gold, molybdenum, and other elements that often are associated with precious-metal deposits (table 3). Globally, both fluorspar and some types of precious-metal deposits commonly formed in areas of crustal extension, and a few deposits were mined initially for fluorspar and then for gold when the latter was discovered to be present in economic amounts (Wallace, 1989b; Greyback and Wallace, 1991).

Climax-type porphyry systems produced broad haloes of fluorine and fluorite, and the presence of such a halo can be used as an exploration guide for concealed molybdenum deposits. Eppinger and Theobald (1985) observed extensive fluorine and fluorite west of the Henderson deposit and then predicted that concealed molybdenum deposits might be present in that area. Similarly, the large fluorine and fluorite halo around the Turquoise Lake copper-molybdenum system (Craig, 1980) could serve as a model for exploring for similar deposits elsewhere. At the Northgate fluorspar district, the fluorspar and related molybdenum anomalies (including molybdenite in one drill hole) led to the hypothesis that the district might represent the top of a concealed Climax-type deposit (Art Bookstrom, U.S. Geological Survey, and Roger Steininger, Minerals Consultant, oral commun., 2006).

Trace elements in fluorite have been used to predict the possible presence of different types of mineral deposits. Eppinger and Closs (1990) used the trace-element and rare-earth-element compositions of fluorite from known mineral-deposit types to show that the geochemical signatures could be used as exploration guides for tungsten and precious-metal deposits. Similarly, Hill and others (2000) used trace elements and stable isotopes from fluorite to predict the possible presence of concealed precious-metal deposits along the Rio Grande rift.

The fluorine-rich thermal springs in central Colorado are modern analogs of the processes that formed the rift-related fluorspar deposits. At Poncha Springs, both modern and fossil systems used the same fault systems. Several fluorine-rich thermal springs, notably in the Mount Princeton and Cottonwood areas, have no associated fluorspar, but the high concentrations of fluorine in the waters suggest the possibility that fluorspar deposits are being formed at a shallow depth.

This study of fluorine, fluorite, and fluorspar in central Colorado has led to the following conclusions:

1. Although some parts of central Colorado have high levels of fluorine, it is not a fluorine-rich province overall.
2. Fluorine has been recycled and new fluorine has been added through geologic time.
3. Magmatic processes were the most important fluorine-adding and fluorine-concentrating events.
4. The past and current levels of exposure have been important for preserving or destroying areas with high concentrations of fluorine.
5. Areas with high fluorine (and fluorite and fluorspar) are indicative of processes that may have produced other types of mineral deposits.

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