



## ***Origin of barite-fluorite-galena deposits in the southern Rio Grande rift, New Mexico***

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## ORIGIN OF BARITE-FLUORITE-GALENA DEPOSITS IN THE SOUTHERN RIO GRANDE RIFT, NEW MEXICO

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**Abstract**—Rio Grande rift (RGR) deposits, formerly called sedimentary-hydrothermal barite-fluorite-galena deposits or Mississippi Valley-type (MVT) deposits, are found within the RGR in central New Mexico. RGR deposits are characterized by a unique regional geologic setting, simple mineralogy dominated by fluorite and/or barite, low temperature of formation (<200–250°C), deposition as open-space fillings with little or no replacement of host rocks, common paragenetic sequence, and no obvious direct association with any magmatic or volcanic activity. These deposits occur along faults, fractures, contact zones, unconformities, shear zones, bedding planes, and solution cavities in diverse Precambrian, Paleozoic, and Tertiary rocks. Age determinations and field relationships suggest that RGR deposits formed during the last 12 Ma coincident with the later stages of rifting in southern New Mexico. Low-temperature formation waters or basinal fluids that accumulated in sedimentary rift basins and were heated by high heat-flow or magmatic activity formed RGR deposits in New Mexico. The warm convecting waters leached barium, sulfate, lead, silver, and other ions from source rocks. Mixing of magmatic and meteoric fluids may have occurred locally, thereby providing fluorine and other constituents. The mineralized fluids were injected along open-spaces as a result of gravity-driven flow from topographic highs and/or compaction of basin sediments. Although density-driven flow produced by thermal contrasts along the basin flanks could also mobilize fluids, this process is not likely to occur in the RGR. Precipitation occurred as a result of cooling of the fluids, decrease in pressure, change in water chemistry, and/or mixing of the mineralized fluids with subsurface brines or meteoric water. Numerous factors, such as surface topography, development of rift structural architecture, erosion and tectonic elimination of aquitards, unroofing of carbonate aquifers by karsting, climate, rates of extension, basin sedimentation, subsidence, compaction, overpressuring, and crustal magmatism, may have played roles in the location, types, and intensity of RGR deposit evolution at any particular site. As a result, deposits of different ages, size, temperature of formation, and accessory mineralogy are a common feature of the RGR deposits.

### INTRODUCTION

Numerous barite-fluorite-galena deposits are found within the Rio Grande rift (RGR) in central New Mexico (Fig. 1; Table 1; Van Alstine, 1976; McLemore and Barker, 1985; North and McLemore, 1985; Hill, 1994). North and McLemore (1986, 1988) previously called these deposits sedimentary-hydrothermal barite-fluorite-galena ( $\pm$  Ag) deposits and Roedder et al. (1968) and Putnam et al. (1983) called them Mississippi Valley-type (MVT) deposits. However, because of the uniqueness of these deposits, they have more recently been called Rio Grande rift (RGR) barite-fluorite-galena deposits (Sutphin, 1997; McLemore and Lueth, 1996; McLemore, in press). The purpose of this paper is to describe these deposits, compare them with classic MVT deposits, and present some theories on their origin.

### DESCRIPTION

RGR deposits are characterized by a unique regional geologic setting, simple mineralogy dominated by fluorite and/or barite, low temperature of formation (<200–250°C), deposition as open-space fillings with little or no replacement of host rocks, common paragenetic sequence, and no obvious direct association with any magmatic or volcanic activity. The deposits are typically hosted by Paleozoic carbonate rocks as vein-fillings, breccia cement, and cavity-fillings adjacent to, or within, major fault zones (Kottlowski and Steensma, 1979; Smith, 1981; McLemore, 1994). RGR deposits are also locally concentrated along other structural discontinuities and features such as minor faults, fractures, stratigraphic and intrusive contact zones, shear zones, bedding planes, and solution cavities in diverse Precambrian, Paleozoic, and Tertiary rocks (McLemore and Barker, 1985; North and McLemore, 1986; McLemore, in press).

These deposits consist predominantly of barite, fluorite, and argentiferous galena. Specific deposits may also contain minor amounts of copper, zinc, molybdenum, uranium, vanadium, and locally, other minerals. Malachite and azurite are the most common secondary copper minerals. Manganese oxides are common in many RGR deposits (Sarkar, 1985), but have been overlooked by many workers. The Hansonburg district is known for 84 mineral species (Northrop, 1959, 1996; Taggart et al., 1989). Dolomitization occurs in many districts, while silicification and jasperoid formation are common features in most districts. In some districts, jasperoids form in several stratigraphic horizons (such as Rincon, Hot Springs, Potrillo Mountains, and Salado Mountains), but only a few are mineralized. The amount of barite and fluorite far exceeds that of sulfides in nearly all RGR districts (Table 1). Galena, typically argentiferous, is the most abundant sulfide mineral, whereas pyrite, sphalerite, and chalcopyrite are found in lesser amounts. Chemical analyses of samples from 11 RGR deposits suggest that trace metal concentrations vary over a wide range (Table 2). Most RGR deposits appear to contain little or no gold, moderate to low concentrations of lead, silver, copper, vanadium, molybdenum, and arsenic (Table 2). The concentration ranges presented in Table 2 suggest that these metal concentrations are highly variable within individual deposits as well as among deposits from different districts.

### PRODUCTION AND ECONOMIC POTENTIAL

Most RGR deposits in New Mexico are too small to be economic; nevertheless, some deposits contained high-grade pockets or zones of barite, fluorite, or galena (lead-silver ore) that were selectively mined in the past (Table 1). RGR deposits typically contain less than a few thousand metric tons (mt) of ore (Table 3), but some deposits can be traced along strike for several thousand meters and

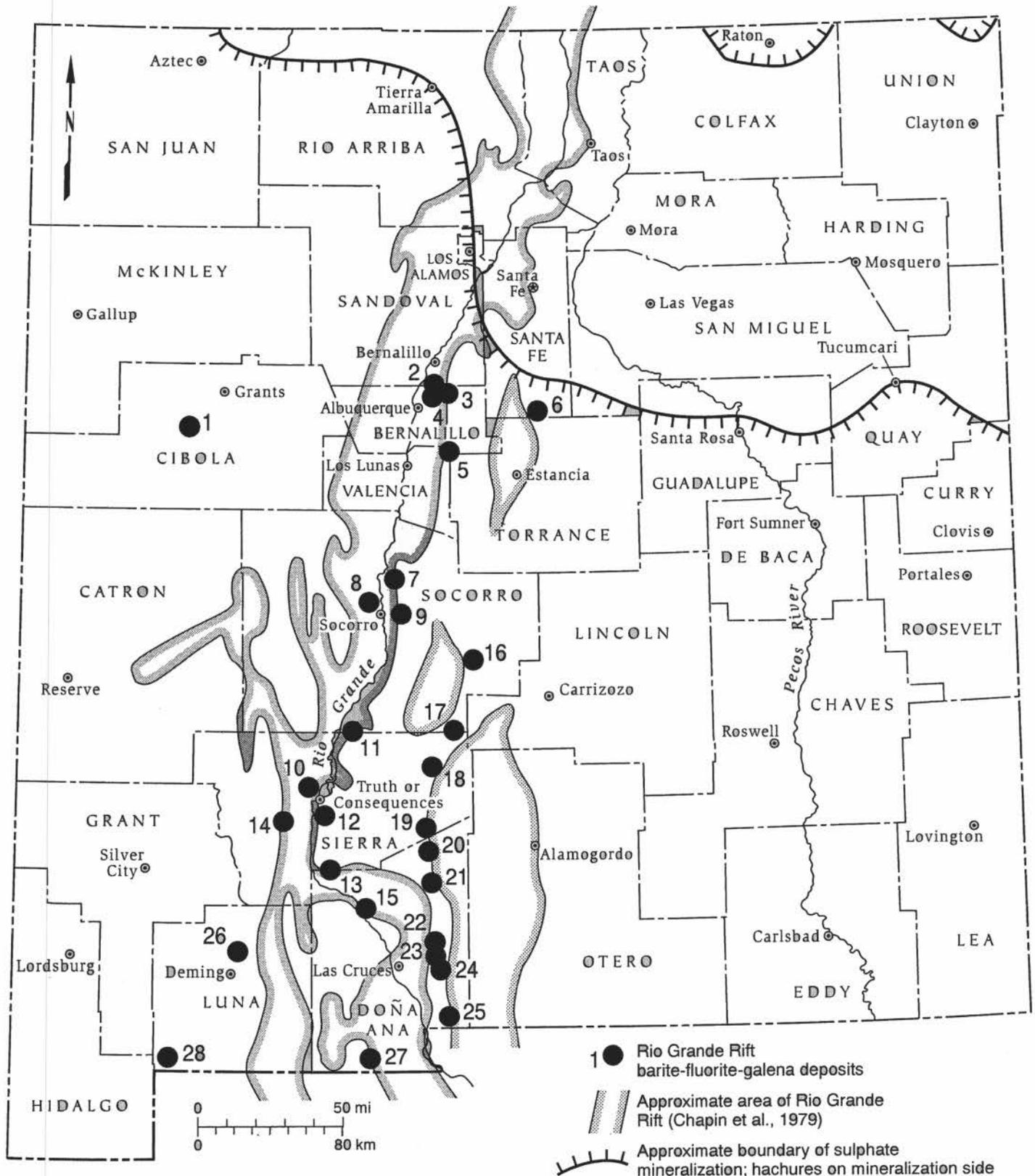


FIGURE 1. Rio Grande rift (RGR) barite-fluorite-galena deposits in New Mexico. Numbers are keyed to Table 1. Boundaries of the RGR are modified from Chapin et al. (1978). The approximate boundary of sulfates in Permian evaporates is shown with hachures on the mineralization side (McLemore and Barker, 1985). Note that RGR deposits occur south of this boundary.

TABLE 1. Rio Grande rift (RGR) barite-fluorite-galena deposits in New Mexico (modified from McLemore and Lueth, 1996). Numbers refer to Figure 1. Production statistics are compiled from U. S. Bureau of Mines (1927-1990), McNulty (1978), and unpublished ore production records (NMBMMR file data) by V. T. McLemore.

DISTRICT	LITHOLOGY	PRODUCTION OF FLUORITE (short tons)	PRODUCTION OF BARITE (short tons)	PRODUCTION OF LEAD (lbs)	PRODUCTION OF ZINC (lbs)	METAL ABUNDANCE	REFERENCE
1. Zuni Mountains	Precambrian granite, Permian sandstone	182,244	—	—	—	F>Ba>>Pb	Godard (1966)
2. Placitas	Precambrian granite, Paleozoic limestones	—	—	10,357	580	Ba>F>Pb>Cu>Zn	McLemore and Barker (1985)
3. Coyote Canyon	Precambrian granite, Paleozoic limestones	200	—	38,756	—	Ba=F>Pb>Zn>Cu	North and McLemore (1986)
4. Tijeras Canyon	Precambrian granite, Paleozoic limestones	400	—	3,982	—	F>Ba>Pb>Cu>Zn	McLemore and Barker (1985)
5. Edgewood	Paleozoic limestones	—	—	—	—	Ba>F>Cu>Pb>Zn	McLemore and Barker (1985)
6. El Cuervo Butte	Yeso Formation, Glorieta Sandstone	—	—	—	—	Ba>F>Pb>Cu>Zn	North and McLemore (1985)
7. Joyita Hills	Precambrian granite, Paleozoic limestones, Tertiary andesite	—	—	small	—	F>Ba>Cu>Pb>Zn	North and McLemore (1986)
8. Lemitar Mountains	Precambrian granite, Paleozoic limestone	—	some	small	—	Ba>F>Pb>Cu>Zn	North and Tuff (1986)
9. Chupadero (Gonzales)	Precambrian granite, Paleozoic limestone	50	some	—	—	F>Ba	—
10. Hot Springs	Paleozoic limestones	—	—	small	—	F>Cu>Pb>Zn	—
11. Fra Cristobal Mountains	Precambrian granite, Paleozoic limestones	—	—	—	—	F>Ba>Pb>Zn>Cu	DeLillo (1991), Allen et al (1984)
12. Caballo Mountains	Precambrian granite, Paleozoic limestone	24,405	—	102,220	—	F>Ba>Pb>Cu>Zn	North and Tuff (1986), Korzeb et al. (1995)
13. Rincon	Paleozoic limestones, Tertiary rhyolite, andesite	8,816	10,250	—	—	Ba>F>Pb=Zn=Cu	Sarkar (1985), Filsinger (1988), McLemore et al. (1996)
14. Salado	Penn limestones	—	—	—	—	F>Ba>Pb=Zn>Cu	—
15. Tonuco (San Diego Mountain)	Precambrian granite	7,720	200	—	—	F>Ba	McLemore et al. (1996)
16. Hansonburg	Paleozoic limestone	—	37,500	11,047,400	33,000	Ba=F>Pb>Zn=Cu	Putnam (1980), Norman et al. (1985)
17. Mockingbird Gap	Precambrian granite, Paleozoic limestones	—	—	71,200	—	F>Ba>Pb>Zn>Cu	McLemore (1994)
18. Salinas Peak	Paleozoic limestones, Tertiary rhyolite	—	—	72,700	10,000	F>Ba>Pb=Zn>Cu	McLemore (1994)

TABLE 1. (continued).

DISTRICT	LITHOLOGY	PRODUCTION OF FLUORITE (short tons)	PRODUCTION OF BARITE (short tons)	PRODUCTION OF LEAD (lbs)	PRODUCTION OF ZINC (lbs)	METAL ABUNDANCE	REFERENCE
19. Granview-Sulfur Canyons	Paleozoic limestones	—	—	small	—	F>Ba>Cu>Pb>Zn	McLemore (1994)
20. San Andrecito-Hembrillo	Paleozoic limestones	—	50+	small	—	Ba>F>Cu>Pb>Zn	McLemore (1994)
21. San Andres Canyon	Paleozoic limestones	—	—	small	—	Ba>F>Cu>Pb>Zn	McLemore (1994)
22. Bear Canyon	Precambrian granite, Paleozoic limestones	—	50	small	—	Ba>F>Pb>Cu>Zn	McLemore (1994), McLemore et al. (1996)
23. Black Mountain	Precambrian granite, Paleozoic limestones	1,100	—	small	—	Ba>F>Cu>Pb>Zn	McLemore (1994), McLemore et al. (1996)
24. Organ Mountains (Bishops Cap)	Paleozoic limestones	1,650	600	—	—	F>Ba>Pb>Zn>Cu	McLemore (1994), Macer (1978), Glover (1975), Kramer (1970)
25. Northern Franklin Mountains	Paleozoic limestones	—	—	small	—	Ba>F>Pb>Zn>Cu	McLemore et al. (1996)
26. Fluorite Ridge	Tertiary andesite, granodiorite	87,533	—	—	—	F>Ba	McLemore et al. (1996)
27. Potrillo Mountains	Paleozoic limestones	—	—	small	—	F>Ba>Cu>Pb>Zn	Jenkins (1977), McLemore et al. (1996)
28. Carrizalillo	limestones	—	—	—	—	F>Ba>Pb=Zn=Cu	Gates (1985), McLemore et al. (1996)
29. Tortugas	Paleozoic limestones	20,751	100	—	—	F>Ba>Pb	McAnulty (1978)

TABLE 2. Range of chemical analyses of samples collected from RGR deposits. Data are from Korzeb et al. (1995), McLemore et al. (1996), and unpublished data (V. T. McLemore). Samples consist of chip sampling across the vein or replacement bodies as well as select and grab sampling of dumps. Analytical methods consisted of induced-coupled plasma spectrometry, flame atomic absorption spectrometry, X-ray fluorescence, and fire assay.

DISTRICT	Au (ppb)	Ag (ppm)	Cu (ppm)	Pb (ppm)	Zn (ppm)	As (ppm)	Mo (ppm)	V (ppm)	NUMBER OF SAMPLES
Bear Canyon	<2-150	<0.7-5.4	2-5000	<15-51,000	42-210	—	<2-140	—	14
Caballo Mountains	<5-360	<5-22	7- >20,000	<2- >10,000	6-4699	<1-749	2-3012	<1- >2000	201
Fra Cristobal	—	0.04-2	0.96-61	0.93-1188	5-530	<0.9-85	0.2-288	—	60
Grandview-Sulfur Canyons	<2-496	<2-238	66-119,711	7-16,147	5-31,654	<10-11	<2-30	2-107	19
Mockingbird Gap	<2-47	<0.4-22	19-10,000	41-47,185	43-39,746	<5-14	2-20	—	23
Northern Franklin Mts.	<1	0.7-11	3.4-15.3	28-18,000	15-12,000	33-5200	32-370	<10-30	8
Potrillo Mountains	<680	<0.7-269	3-63,000	14-5100	14-41,000	—	—	—	21
Salado	<5-200	<0.2-2	12-16	20-41	5-79	<5-165	4-34	—	6
Salinas Peak	<2-152	0.5-52	11-14,649	548-46,313	21- >99,999	—	<2-162	3-124	24
San Andrecito-Hembrillo	<680	<0.7-115	<50-178,000	14-42,000	18-1700	<10-210	—	—	14
San Andres	—	<2-22	21-270,000	14-152,000	7-330	11-260	—	<4-280	14

TABLE 3. Grade and tonnage data for selected Rio Grande rift deposits in New Mexico (from NMBMMR unpublished file data). Data presented do not include total production.

DISTRICT	MINE	TONNAGE				
		(short tons)	Pb%	Zn%	Cu%	Ag oz/ton
Hansonburg	Mex-Tex	95,000	1.3	0.01	0.01	0.08
Salinas Peak	all producing mines	530	6.9	0.94	0.01	0.58
Mockingbird Gap	all producing mines	833	4.3	—	—	0.14

contain 100,000 mt or more of ore. One of the largest RGR deposits is found in the Hansonburg district and consists of equal amounts of barite, fluorite, and galena. However, the lead and zinc grades at Hansonburg (1.3 and 1.0%, respectively) are much lower than skarn and carbonate-hosted Pb-Zn replacement deposits elsewhere in the state (Table 2; McLemore and Lueth, 1996; McLemore, in press). Some RGR deposits were selectively mined for zinc, copper, vanadium, uranium, molybdenum, and jarosite. Zinc was produced at Hansonburg and Palomas Gap in the Caballo Mountains. From 1909 to 1911, 1500 lbs of  $V_2O_5$  as vanadinite were also produced from Palomas Gap (Eveleth, 1986) and during 1914 a small amount of molybdenum as wulfenite was produced from the Gladys claim at Palomas Gap (Horton, 1916; Schilling, 1965). Several hundred metric tons of jarosite were mined from the Copiapo mine in the northern Franklin Mountains at Webb Gap for use as pigment in paints (Dunham, 1935). Jarosite is also found at Hansonburg, Bishops Cap, and in the Potrillos Mountains. In 1955, 12 short tons of ore grading 0.06%  $U_3O_8$  and 0.04%  $V_2O_5$  were produced from the Blue Star mine at Bishops Cap in the Organ Mountains district (McLemore, 1983). Production of mineral specimens by collectors is common in many RGR deposits.

## REGIONAL GEOLOGIC SETTING

RGR deposits in New Mexico are restricted to geologic settings associated with the Rio Grande rift (Fig. 1). No significant barite-fluorite-galena deposits occur north of the Placitas district in north-central New Mexico. This absence of significant RGR mineralization has been attributed to a lack of Paleozoic evaporite deposits in the stratigraphic section north of the Placitas district that could provide sulfur needed to form these barite-rich deposits (Fig. 1; McLemore and North, 1984; McLemore and Barker, 1985). Nearly all RGR deposits are found in structurally high geologic terrains. Most of the major RGR deposits occur in late Tertiary uplifts that are superimposed over Laramide (Late Cretaceous-Eocene) structural highs (Fig. 2). Examples of RGR deposits over Laramide uplifts are found in the Zuni Mountains, Rincon, Fra Cristobal, Salado, Tonuco (San Diego Mountain), Bishops Cap (Organ Mountains), and Tortugas districts. Modern-day convective geothermal systems of the RGR have an identical regional setting and are found on structural highs at locally low elevations (Witcher, 1988).

## CENOZOIC THERMAL REGIME

Shallow, typically less than 1 km depth, heat-flow measurements indicate that much of the RGR is characterized by a very high heat flow, 85–100 mW/m<sup>2</sup> (Reiter et al., 1975, 1978, 1979, 1986; Decker and Smithson, 1975; Decker et al., 1988; Reiter and Barroll, 1990). Reduced heat-flow models that account for crustal radiogenic heat sources, and thermal extension models suggest that additional heat from the underlying mantle may be required (Roy et al., 1968; Decker and Smithson, 1975; Lachenbruch and Sass, 1978; Morgan et al., 1986; Decker et al., 1988; Keller et al., 1990; Reiter and Barroll, 1990). Convective addition of mantle-derived magma to the crust is a common process that enhances the thermal regime of

rifts (Morgan, 1982). Cenozoic basaltic andesite volcanism, indicative of magma generation in the upper mantle, accompanied initial rift extension between 36 to 24 Ma and is also associated with intense periods of volumetrically important silicic ignimbrite volcanism (Ratté et al., 1984; Elston, 1984, 1994; McIntosh et al., 1992; Mack et al., 1994; Hawkesworth et al., 1995). With convective magmatic heat from the mantle, anomalous upper crustal heat has probably characterized the RGR since shortly after extension began about 36 Ma ago (Morgan et al., 1986; Keller et al., 1990; Barroll and Reiter, 1990; Bussod and Williams, 1991).

Ground-water flow in the upper 5–7 km of the crust dramatically redistributes heat by suppressing heat flow in areally large recharge regions (<80 mW/m<sup>2</sup> heat flow), and increasing heat flow in relatively restricted discharge areas to values that can greatly exceed 105 mW/m<sup>2</sup> (Barroll and Reiter, 1990). Heat flow above 400 mW/m<sup>2</sup> is estimated over the discharge area of some currently active hydrothermal systems in the RGR (Reiter et al., 1986; Barroll and Reiter, 1990; Witcher, 1991). Measured subsurface temperatures and chemical geothermometry indicate that temperatures between 50 and 160°C are common at depths less than 1 km in present day RGR geothermal systems outside the Valles caldera.

RGR barite-fluorite-galena deposits are associated with a region that has had high heat flow for at least the last 36 Ma. Three components of thermal regime characterize the rift during this time. The first is a diffuse, but high, conductive heat flow from deep crustal regions and the mantle. Upper crustal radiogenic heat augments the mantle convective flux. Second, transient and local point source thermal impulses from cooling silicic and intermediate composition magma bodies in the shallow crust are ephemerally superimposed on the diffuse regional regime. Hydrothermal systems existed over many of these intrusions resulting in volcanic-epithermal gold, silver, and base metal mineralization (i.e., Steeple Rock, Mogollon, and other districts in New Mexico; McLemore, 1993; Kamilli, 1993, 1994; Kamilli, and Ratté, 1995; McLemore, 1996). Finally, topographic-driven (gravity) advective flows of ground

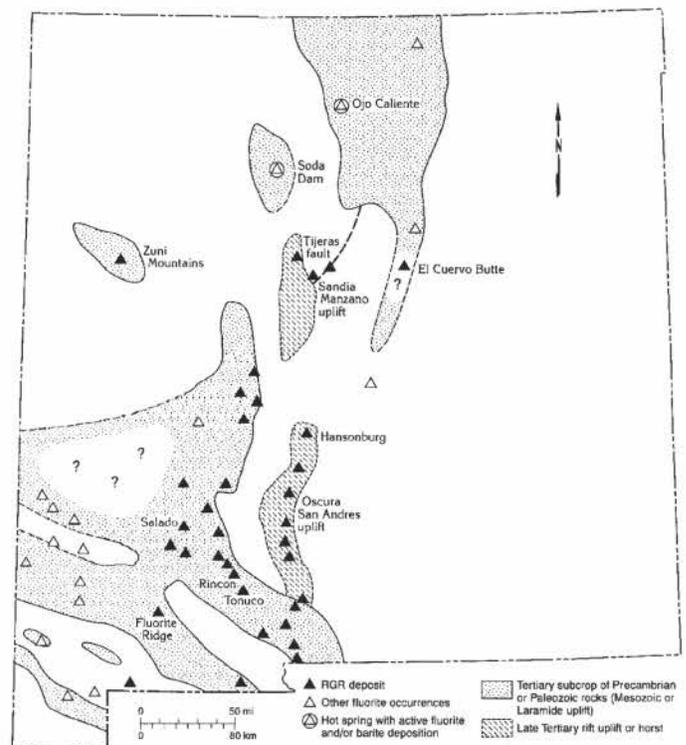


FIGURE 2. Relationship of RGR deposits and Laramide structural highs and selected late Tertiary rift uplifts.

water in the upper-most crust concentrate heat associated with the first two thermal regimes. As the rift has evolved, the loci for recharge and discharge of regional ground-water flow have changed dramatically. Former low-lying discharge areas may now be recharge areas at relatively high elevations.

### TEMPERATURE AND CHEMISTRY OF RGR MINERALIZING FLUIDS

Fluid inclusion and geochemical studies of RGR deposits at Hansonburg (#16, Fig. 1, Table 1) and other smaller districts are the principal source of direct data on the chemistry and temperature of RGR mineralizing fluids (Table 4; Roedder et al., 1968; Macer, 1978; Putnam, 1980; Norman et al., 1985; North and McLemore, 1985; North and Tuff, 1986; Bohlke and Irwin, 1991; Hill, 1994). These data suggest that the mineralizing fluids responsible for forming RGR deposits were low temperature (95–220°C) and low-to-moderate salinity (0.3–20 eq. wt. % NaCl). Thus, mineralizing fluids ranged in character from brackish water to true brines of moderate salinity.

The most intensely studied RGR district is Hansonburg (Kottlowski and Steensma, 1979; Roedder et al., 1968; Putnam, 1980; Putnam et al., 1983; Norman et al., 1985; Taggart et al., 1989; Bohlke and Irwin, 1991; Hill, 1994). Geochemical models based on information from these studies indicate that mineralizing fluids at Hansonburg ranged in temperature from 125 to 210°C, were of moderate salinity (10–18 eq. wt. % NaCl), and acidic (pH of 4.3–5.1). Data also suggests that deposition took place at low pressures of 150–200 bars, which is consistent with deposition at shallow depths of 1.5–2 km (Table 4). Conodont geothermometry indicates that the sediments were heated to 110–140°C (Cook, 1986). Similar low-temperature, low-to-moderate-salinity fluids formed the deposits at El Cuervo Butte (North and McLemore, 1985), Potrillo Mountains (Jenkins, 1977), Rincon (Filsinger, 1988), and elsewhere along the RGR in central New Mexico (Table 4; Macer, 1978; North and Tuff, 1986; Hill, 1994).

Fluid inclusions from Hansonburg have a high aqueous sulfate to sulfide ratio and contain organic gases and locally organic liquids (Roedder et al., 1968; Norman et al., 1985). The predominant volatile is H<sub>2</sub>O, whereas CO<sub>2</sub>, C<sub>n</sub>H<sub>n</sub>, and N<sub>2</sub> are the predominant gases (Norman et al., 1985). Organic material and CO<sub>2</sub> also have been found in other RGR deposits (North and Tuff, 1986). Norman et al. (1985) interpreted the organic constituents at Hansonburg as being possibly derived from biological and sedimentary sources in the rift basin and indicating a basinal-brine origin for the mineralizing fluids. Bohlke and Irwin (1991) analyzed noble gases and halide ratios in fluid inclusions from the Hansonburg district and found that Br/Cl and I/Cl ratios are similar to those of modern brines formed by dissolution of Permian NaCl-bearing marine evaporates in southern New Mexico. They also found that concentrations of <sup>39</sup>Ar and <sup>84</sup>Kr in these fluid inclusions are similar to concentrations found in fresh water in equilibrium with air at approximately 20°C. <sup>40</sup>Ar/<sup>39</sup>Ar ratios indicate excess <sup>40</sup>Ar in the fluid. Together, these observations suggest that Hansonburg mineralizing fluids originated as relatively dilute meteoric waters and that after recharge this dilute fluid acquired the bulk of its chloride and sulfate in the subsurface by dissolving halite-bearing Permian marine evaporates. The brine then mobilized lead and excess radiogenic <sup>40</sup>Ar through interactions with non-evaporate aquifer rocks, probably clastic sedimentary or Precambrian basement rocks (Bohlke and Irwin, 1991).

This fluid evolution model is consistent with available sulfur- and lead-isotopic data. Stable isotopes and geochemical analyses of fluid inclusions in a regional study by Hill (1994) indicate that

meteoric waters formed many RGR deposits in New Mexico. From a sulfur-isotopic study of the deposits in the Hansonburg district, Allmendinger (1974, 1975) concluded that the mineralizing fluid acquired its sulfur from Permian evaporates. Sulfate also could be derived from lacustrine deposits of evaporitic gypsum, Permian gypsum deposits, or sulfides in black shales. It appears that sulfate was not the result of dissolution of halite-bearing evaporates for RGR deposits with low fluid salinities between 0.3 and 8 eq. wt. % NaCl, but was the result of dissolving constituents from carbonate, siliciclastic, and crystalline source rocks within the rift basin. The source for fluorine in RGR deposits is problematic and discussed below. Lead-isotopic data indicate that the galena from Hansonburg, the Lemitar Mountains, Salinas Peak, Caballo Mountains, and Joyita Hills was derived from rocks derived from upper crustal sediments and/or Precambrian rocks (Slawson and Austin, 1962; Ewing, 1979; Stacey and Hedlund, 1983). Lead-isotopic data also suggest a source for barium compatible with derivation from basement rocks (Slawson and Austin, 1960; Ewing, 1979) or Permian arkoses derived from basement rocks (Beane, 1974).

### PARAGENESIS

The depositional sequence of mineralization, or mineral paragenesis, appears to be surprisingly uniform for most RGR deposits. Four distinct stages of mineralization are recognized in detailed studies of the Hansonburg district (Roedder et al., 1968; Putnam, 1980; Putnam et al., 1983; Norman et al., 1985; Taggart et al., 1989), although minor differences exist between interpretations. Stage 1 is characterized by extensive silicification and dolomitization followed by deposition of minor amounts of pyrite and sphalerite. Stage 2 is characterized by deposition of galena in open spaces accompanied by minor deposition of chalcopyrite. Stage 3 is characterized by barite-fluorite mineralization in open spaces accompanied by deposition of quartz. During this stage, early fluorite preceded barite deposition and is followed by simultaneous deposition of both minerals. The culminating stage, stage 4, is characterized by deposition of calcite and quartz. Roedder et al. (1968) reports five recognizable substages of fluorite deposition with intermediate quartz and barite deposition. However, Putnam et al. (1983) found fluorite paragenesis to be ambiguous. Both studies found a general decrease in fluid temperatures from early sulfide deposition at 150–210°C to later barite-fluorite mineralization at 125–180°C. Putnam et al. (1983) suggested that salinities decreased from sulfide deposition to barite-fluorite deposition, while Roedder et al. (1968) reported evidence for an increase in salinities of mineralizing fluids during stage 3 fluorite deposition. Two or three stages of fracturing and faulting occurred during the depositional sequence. Taggart et al. (1989) proposed an oxidation stage of mineralization that occurred at the end of stage 3, after deposition of barite and fluorite, when the fluids became cooler and more oxidizing. This gradual change in the hypogene chemical environment from the more reduced mineralizing fluids to cooler, more oxidizing fluids resulted in the formation of a significant amount of sulfate, carbonate, silicate, and oxide mineral species found in the district. Some of these minerals are primary, while other minerals were formed at the expense of pre-existing mineral species during both the hypogene and supergene stages (Taggart et al., 1989).

McMahon (1989) described the paragenesis of four RGR districts (Caballo Mountains, San Diego Mountain, Tortugas, and Bishops Cap) and constructed a complete paragenetic sequence based on petrographic data and field relationships from individual deposits. He found three well-defined stages. Stage 1 involved the formation of jasperoid with minor deposition of vein quartz, cal-

TABLE 4. Summary of fluid inclusion data from selected Rio Grande rift deposits in New Mexico.

DISTRICT	TEMPERATURES OF HOMOGENIZATION °C	SALINITIES eq. wt.% NaCl	MINERALS	REFERENCE
Zuni Mountains	108-220	7-20.4	fluorite	Hill (1994)
El Cuervo Butte	95-190	6-16	fluorite	North and McLemore (1985)
Caballo Mountains	120-346	0-19.8	fluorite	North and Tuff (1986), Hill (1994)
Fra Cristobal	176-252	0.5-5.9	fluorite	Sarkar (1985)
Hansonburg	125-210	10-18	fluorite, quartz	Putnam (1980), Putnam et al. (1983)
Rincon (Doña Ana County)	163-341	4.8-17.0	quartz, barite, fluorite	Filsinger (1988)
Lemitar Mountains	138-220	2.8-5	fluorite	North and Tuff (1986)
Gonzales, Chupadero	143-189	1.1-8.1	fluorite	Hill (1994)
Fluorite Ridge	145-224	0.8-8.1	fluorite	North and Tuff (1986), Hill (1994)
Potrillo Mountains	185-220	—	quartz	Jenkins (1977)
Bishops Cap, Organ Mountains	134-187	0.3-0.6	fluorite	Macer (1978), North and Tuff (1986)
Tortugas Mountains	138-186	0.3	fluorite	Macer (1978)

cite, galena, and pyrite. At most deposits, brecciation of jasperoid, probably due to fault movement and/or hydrofracturing, separates it from the second stage. Stage 2 is characterized by barite and fluorite deposition, typically simultaneously. Fluorite was found in all of the deposits McMahon (1989) studied, but barite was absent in a few. The final stage 3, followed the onset of fracturing at the end of stage 2 and is characterized by local deposition of minor calcite and significant quartz, followed by major calcite deposition and finally hematite deposition. Filsinger (1988) found a similar paragenesis at the Palm Park deposits in the Rincon district.

These and other separate studies suggest that RGR deposits generally developed in four stages. Stage 1 consists of early dolomitization and silicification, typically resulting in the formation of jasperoids. Stage 2 is characterized by minor sulfide deposition. Stage 3 is characterized by fluorite-barite deposition. The final stage 4 is characterized by deposition of calcite and quartz. Fracturing appears to be prevalent during the early part of stages 2 and 3. Manganese oxides are typically deposited during the later stage 3 or early stage 4, when the fluids became more oxidizing. The similar paragenesis for RGR deposits within a region 150-km long suggests that structural, thermal, and geochemical conditions responsible for RGR mineralization were, to some extent, controlled regionally. Local conditions, such as fluid-host rock interactions and localized fault systems, are primarily responsible for the mineralogical and chemical differences between individual RGR deposits.

### AGE OF RGR DEPOSITS

Most RGR deposits are found in structurally high bedrock (pre-Tertiary) adjacent to clastic-filled rift basins (Table 1, Fig. 1). For many of these deposits, geohydrologic evidence and host lithology would suggest that RGR deposits are either pre-Tertiary or concurrent with rifting and basin development, but before host terrains were uplifted above a regional water table. However, previous workers have assigned these deposits to the mid-Tertiary (Allmendinger, 1974, 1975; Beane, 1974; Ewing, 1979; Putnam et al., 1983). More recent studies (McMahon, 1989; Lueth and Goodell, 1996; Goodell et al., 1997; Kelley et al., 1997; Lueth and Heizler, 1997; Lueth et al., this guidebook) suggest that mineralization began as early as middle Miocene and extended to late Miocene-early Pliocene (Table 5). From early Pliocene to the present, hydrothermal systems in the RGR have deposited minor amounts of barite, fluorite, sulfide, and manganese. Detrital man-

ganese (psilomelane) in the late Miocene Rincon Valley Formation provides stratigraphic evidence in the Rincon and Caballo districts of a pre-late-Miocene age. The tectonically-juxtaposed and underlying early-Miocene Hayner Ranch Formation exhibits manganese and barite mineralization in faults and fracture zones (Seager et al., 1971; Seager and Hawley, 1973). In a nearby drill hole in an active geothermal system, euhedral barite is observed in fractures in the Rincon Valley and Camp Rice Formations, indicating a late-Miocene to Holocene age for mineralization.

At San Diego Mountain (Tonuco district), mineralization can be bracketed between early-middle Miocene and late Miocene based on structural and stratigraphic evidence (McMahon, 1989). In this district, the early-middle Miocene Hayner Ranch Formation has been silicified and mineralized. Also, at San Diego Mountain, the West Tonuco fault truncates some of the fluorite-barite veins that cut Precambrian granite. The West Tonuco fault is a normal fault that is related to the latest Miocene and Pliocene rifting event that produced the modern uplifts in the RGR (Chapin et al., 1978; Seager et al., 1984). It is possible that there was no movement on the fault until the Pliocene (Seager et al., 1971). Recent apatite fission dating corroborate these conclusions regarding the age of mineralization at San Diego Mountain, but suggest that minor barite mineralization has continued through the Holocene. Kelley and Chapin (1997) report an apatite fission track (AFT) cooling age of  $6.5 \pm 2.2$  Ma on granite from San Diego Mountain that is mineralized with fluorite and barite, and they interpret this age as cooling after uplift through the 120°C fluorapatite isotherm. Therefore, AFT dating provides a minimum age for the mineralization at this site. High heat flow (Reiter et al., 1979, 1986) and outcropping siliceous sinter on the west slopes of San Diego Mountain indicate a mid-Pleistocene to Holocene intermediate-temperature (100–150°C) geothermal system. On the west side of San Diego Mountain, a remnant outcrop of Plio-Pleistocene Camp Rice fluvial sand on the hanging wall of the West Tonuco fault is silicified and contains abundant barite mineralization.

Methods of directly determining the ages of these deposits are currently being used (Table 5). The Gonzales deposit in the Chupadera district, Socorro County, may be as young as 8 Ma according to AFT cooling dates of the granitic host rock (Kelley et al., 1992, 1997).  $^{40}\text{Ar}/^{39}\text{Ar}$  age determinations of jarosite from Hansonburg indicate an age of  $6.36 \pm 0.1$  to  $5.98 \pm 0.06$  Ma (Lueth and Heizler, 1997). Jarosite from Bishops Cap, Organ Mountains district (#24, Fig. 1, Table 1) is 5.4–5.1 Ma (Lueth et al., this guidebook). Jarosite from

the Copiapo deposit, Northern Franklin Mountains district (#25, Fig. 1, Table 1) indicates an age of  $5.0 \pm 0.3$ – $4.6 \pm 0.06$  Ma; jarosite from the Schneider claim (Northern Franklin Mountains district) is  $3.25 \pm 0.29$  Ma (Lueth and Goodell, 1996; Goodell et al., 1997; Lueth et al., this guidebook). These limited age determinations may suggest that RGR deposits are younger in the southern RGR, but more data are needed to confirm this observation.

### COMPARISONS BETWEEN RGR AND MISSISSIPPI VALLEY-TYPE DEPOSITS

Because of similarities between the Hansonburg RGR deposits and classic MVT deposits located in the North American mid-continent, it has been suggested that RGR deposits represent a subclass of MVT deposits (Roedder et al., 1968; Putnam, 1980; Norman et al., 1985; Morgan et al., 1986). The features most favorably compared include ore textures and mineralogy, mineralizing-fluid chemistry, host-rock lithology, and genetic relationship to sedimentary basins. However, two problems arise in trying to classify RGR deposits as MVT deposits. First, there is disagreement regarding the assignment of a large number of deposits as MVT deposits (Bjorlykke and Sangster, 1981; Gustafson and Williams, 1981; Sangster, 1983, 1990; Sangster et al., 1994). Second, and perhaps even more important, many of the features of Hansonburg and other RGR deposits are quite different from those of the MVT "type-section," the Upper Mississippi Valley zinc-lead district in southwest Wisconsin and similar MVT deposits in the North American mid-continent and around the world. A far better analogy would be between RGR deposits and other rift-related fluor spar (RRF) deposits (Van Alstine, 1976; Plumlee et al., 1995).

#### MVT deposits

The general characteristics of classic MVT deposits (the "type section" and other Zn-Pb MVT deposits in the mid-continent of the United States) have been critically documented by Sangster (1983, 1990), Sverjensky (1986), and Sicree and Barnes (1996). These characteristics include the following:

MVT deposits occur principally in limestone or dolomite that overlie crystalline basement. The mineral deposits are never found in crystalline basement rocks, but in some districts (most notably, the Viburnum Trend of southeast Missouri) small amounts of ore minerals are found in basal siliclastic sediments between crystalline basement and carbonate host rocks.

Deposits are stratabound and consist of bedded replacements, open-space fillings in vugs between breccia fragments, and in veins; all of which are strongly controlled by individual strata. There is always evidence of dissolution of the carbonate host rocks expressed by thinning, slumping, collapse, and brecciation.

Predominant mineral phases include galena, sphalerite, pyrite, and marcasite, although the relative abundances vary markedly between individual deposits and districts. Barite and/or fluorite are accessory minerals in some districts, but are entirely lacking in a majority of districts. Other minor minerals include chalcocopyrite, dolomite, calcite, and quartz.

The most obvious examples of wall-rock alteration associated with mineralization are dissolution, recrystallization, and brecciation of the host carbonate rocks, along with various degrees of silicification and dolomitization.

MVT deposits are constrained in regional geologic setting to platform sequences peripheral to intracratonic basins and, in some cases, are spatially related to basement highs within the basin.

Deposits were formed in areas of mild deformation expressed by fractures, broad domes and basins, and gentle folds.

TABLE 5. Age of Rio Grande rift deposits in New Mexico.

AREA	AGE Ma	METHOD	REFERENCE
Gonzales, Chupadera	$8.2 \pm 4.8$	Fission track, apatite from granite	Kelley et al. (1992)
Copiapo, Northern Franklin Mountains	$5.0 \pm 0.3$ – $4.6 \pm 0.06$	$^{40}\text{Ar}/^{39}\text{Ar}$ , jarosite	Lueth and Goodell (1996), Lueth et al. (1998), Goodell et al. (1997)
Scheider, Northern Franklin Mountains	$3.25 \pm 0.29$	$^{40}\text{Ar}/^{39}\text{Ar}$ , jarosite	Lueth and Goodell (1996), Lueth et al. (1998), Goodell et al. (1997)
Bishops Cap Hansonburg	$5.4$ – $5.10$ $6.36 \pm 0.1$ – $5.98 \pm 0.06$	$^{40}\text{Ar}/^{39}\text{Ar}$ , jarosite $^{40}\text{Ar}/^{39}\text{Ar}$ , jarosite	Lueth et al. (1998) Lueth and Heizler (1997)
San Diego, Tonoco Mountain	$6.5 \pm 2.2$	Fission track, apatite from granite, stratigraphic	Kelley and Chapin (1997), McMahon (1989)
Rincon	Miocene	stratigraphic	Seager and Hawley (1973), McMahon (1989)

Deposits are not associated with igneous rocks.

The ore minerals were deposited at shallow depths and were probably never deposited at depths greater than about 1500 m.

Fluid inclusions in sphalerite, fluorite, barite, and calcite always consist of dense, saline aqueous fluids and locally contain oil and natural gas. Total dissolved salts range from 10 to 30 eq. wt.% NaCl and consist predominantly of sodium/calcium chloride. Homogenization temperatures are generally in the range of 50–220°C.

Hydrogen and oxygen isotopic compositions of the fluid-inclusion fluid are similar to those of pore water in sedimentary basins. Sulfur isotopic systematics demonstrate that the source of sulfur was not magmatic and must have been sedimentary.

Zinc and lead are the most common metals that determine economic viability. In some districts silver, cadmium, germanium, barite, and fluorite have been recovered, but only as byproducts.

Individual MVT deposits generally contain between 300,000 and 3,000,000 mt of ore with zinc grades ranging from 2 to 6% (maximum 16%) and lead grades between 1 and 3% (seldom exceeding 4%). For most districts, Zn/(Zn + Pb) ratios fall between 0.6 and 0.8.

During the past several decades, the formation of MVT deposits has been the subject of intense study. Important questions remain regarding transport and deposition mechanisms of ore constituents. However, there is general consensus that these deposits were formed when hot fluids, similar to oil-field brines, migrated out of the deeper regions of sedimentary basins through aquifers, eventually forming mineral deposits in platform carbonates at or beyond the basin margin (Sverjensky, 1986; Sangster, 1990). Lead and strontium isotopic compositions and trace element geochemistry suggest that crystalline basement and overlying clastic sandstones are the predominant sources for the lead, zinc, barium, and other metals in these deposits (Sangster, 1990; Viets et al., 1992; Christensen et al., 1995; Goldhaber et al., 1995). Sulfur isotopic compositions of fluid inclusions, sulfide minerals, and barite demonstrate that sulfur in the mineralizing fluids was derived from sea water as trapped pore water in sediment or through the dissolution of evaporites (Sangster, 1990). Evidence suggests that MVT mineralizing fluids were driven from the hot, deeper parts of the basin to the basin periphery by topography-driven (gravity) flow that was initiated by foreland basin uplift during an associated orogenic event (Sangster, 1990; Brannon et al., 1992; Sverjensky and Garven, 1992; Garven and Raffensperger, 1997). These regional fluid-flow systems appear to be very large scale, on the order of 100s of km and reaching distances of 700 km in the case of the MVT ore formation associated with the Late Pennsylvanian/Early Permian Alleghenian/

Ouachita orogeny. This event is responsible for uplift of the Appalachian foreland basin and flushing of ore fluids from the Arkoma, Illinois, and Michigan basins to form the classic MVT deposits of the North American mid-continent (Brannon et al., 1992; Sverjensky and Garven, 1992;). Similar plumbing systems can be documented for most MVT districts around the world (Sangster et al., 1994; Garven and Raffensperger, 1997).

### Illinois-Kentucky fluor spar district

The Illinois-Kentucky fluor spar district has been classified by many as a MVT deposit, albeit significantly anomalous (Sangster, 1990; Spry and Fuhrman, 1994; Symons, 1994). Deposits in this district have some similarities with MVT deposits of the mid-continent including a genetic relationship. However, there are significant differences. Mineralization is dominated by fluorite, with only minor amounts of barite, sulfides, calcite, and dolomite. The district is located on the southeastern flank of the Illinois basin in an area that overlies the early Paleozoic Reelfoot rift system. Mineralization is associated with a faulted and collapsed domal anticline that has as its apex a cryptovolcanic feature named Hicks Dome. The Early Permian igneous system associated with Hicks Dome is alkaline and consists of ultramafic dikes and sills and intrusion breccias, containing fragments of silicic syenite and peridotite. These rock types are typical of mantle-derived alkaline magmatism. The horst and graben structure of the rift post-dates these igneous rocks and many of the faults in the rift are mineralized where they cut Mississippian host rocks. The two predominant styles of ore include: (1) fluorite fissure-filling veins in Mississippian-age rocks along graben-bounding faults, which also penetrate Precambrian basement, and (2) stratabound, flat-lying replacement and open-space filling deposits in limestone of Middle to Late Mississippian age adjacent to major graben-bounding faults. A minor third style of mineralization occurs as fluorite and barite cementing explosion (intrusion) breccias in Ordovician rocks overlying Hicks Dome. Fluid inclusion studies demonstrate that deposition temperatures ranged from 60 to 160°C and that mineralizing fluids had salinities between 17 and 23 eq. wt. % NaCl (Spry and Fuhrman, 1994). Geologic and geochemical data suggests that these mineralizing fluids were derived from the flushing of the Arkoma and related basins during the same Early Permian orogenic event responsible for MVT mineralization in the mid-continent (Viets et al., 1992; Chesley et al., 1994; Plumlee et al., 1995). However, Symons (1994) suggests a late Jurassic age for the Illinois-Kentucky fluor spar deposits.

The available data from the Illinois-Kentucky fluor spar district along with recent geochemical modeling of fluorite deposition in this district (Plumlee et al., 1995) build a convincing case for the superposition of continental rift and intracratonic basinal processes in the formation of fluorite mineralization in this area. Plumlee et al. (1995) tested a variety of deposition models and concluded that the fluorine in the district is most likely derived as HF expelled from crystallizing alkaline magma in the rift. This HF then mixes at depth with basinal brines being expelled from the Arkoma basin and deeper parts of the rift. These fluorine-enriched basinal brines then move upward through horst and graben-bounded faults mineralizing these faults within Mississippian limestones as well as forming stratabound deposits within the limestones. Plumlee et al. (1995) point out the similarity of rift environments (structural style and alkaline magmatism) and style of mineralization associated with the Illinois-Kentucky fluor spar district and other RRF districts around the world and suggest a similar source of fluorine for such deposits, including those in the RGR.

### RGR deposits

From the description of RGR and MVT deposits above, it is clear that these two deposit types are similar only in a few superficial aspects. Furthermore, local processes within the RGR typically determine many of these similarities, including nature of host rock, relative amounts of minerals, and deposition textures. Of more importance are the differences between these two classes of deposits and the similarity of RGR deposits with other RRF districts. Most fundamentally, MVT deposits were developed in mildly deformed platform carbonate rocks at intracratonic basin margins in response to compressional orogenic events hundreds of kilometers away. RGR and RRF deposits were developed in a wide variety of lithologies, that were highly deformed by normal faulting (horst and graben structures) associated with basin extension, and are regionally related to alkaline magmatism developed within continental-rift systems.

The plumbing systems are significantly different. For the most part, the plumbing system for MVT mineralizing fluids is controlled by the stratigraphy of intracratonic basins. However, the plumbing systems for RGR and other RRF deposits are controlled, in large part, not only by basin stratigraphy, but also by the normal faults developed within these strata and extending into the basement rocks. As mentioned above, fluid-inclusion salinities of fluorite are highly variable in different RGR deposits compared to fluid inclusions from MVT. Together, this evidence suggests that RGR mineralizing fluids were originally dilute meteoric waters that were buried or migrated to the deepest parts of the rift basin and into the basement. These fluids then leached varying amounts of brine constituents (Na<sup>+</sup>, Cl<sup>-</sup>, SO<sub>4</sub><sup>2-</sup>, Ca<sup>2+</sup>, etc.) by reacting with local host rocks (including evaporites, siliciclastics, carbonates, and crystalline rocks). On the other hand, MVT mineralizing fluids were derived from formation waters that evolved ultimately from seawater and seawater components. Another significant difference between RGR and MVT deposits is the dominance and ubiquitous presence of fluorite in RGR and RRF deposits compared to the minor abundance or lack of fluorite in MVT deposits. We believe these major differences between RGR and MVT deposits are sufficient to preclude RGR deposits from being classified as MVT deposits. Furthermore, we believe that it is reasonable to assign RGR deposits as a subclass of RRF deposits, based upon similar geology, plumbing system, mineralogy, and geochemistry.

## ORIGIN OF RGR DEPOSITS

### Physical model

The wide range of geologic, geochemical, and heat flow data, along with circumstantial evidence as summarized in this paper have given rise to several genetic models for RGR deposits (Roedder et al., 1968; Macer, 1978; Norman et al., 1985; McLemore and Barker, 1985; North and McLemore, 1986, 1988; McMahan and Giordano, 1988; McMahan, 1989; Hill, 1994; McLemore and Lueth, 1996; McLemore, in press). We propose the following generalized model for RGR mineralization and then discuss some of the more controversial aspects of this model. The geologic data along with trace element and isotopic data summarized in this paper provide strong evidence that RGR mineralization began approximately 12 Ma (middle Miocene), but that the predominant period of deposition occurred between 8 and 5 Ma. During this time, meteoric waters within the Tertiary and older sediments of the rift evolved chemically, as they became warmer with depth due to burial and/or basin compaction, by dissolving rock constituents to produce waters of varying salinities and calcium concentrations.

Reactions with Permian evaporite deposits to produce NaCl brines rich in sulfate formed the most concentrated fluids. Sulfate in less concentrated fluids may have been derived from lacustrine evaporite deposits, pyrite-rich black shales, or other sulfur-bearing units. As these warm brines interacted with Precambrian basement rocks, Tertiary volcanic rocks, and arkoses derived from these sources, they mobilized lead, barium, and other metals. Some of these brines mixed with fluorine-bearing fluids derived from the alkaline and silicic magmatism within the rift to produce fluorine-enriched brines (see discussion below). As the rift developed, favorable discharge zones allowed the mineralizing fluids to migrate up along fractures, faults, and unconformities, thus giving rise to deposition within a wide range of lithologies comprising the margins of rift basins as well as horst blocks along the rift.

### Controversial aspects

Somewhat controversial is the source of fluorine in RGR deposits. Macer (1978) proposed a meteoric water source for fluorine, but provided no geochemical model to show that significant fluorite deposition could take place from waters containing several ppm fluorine. Worl (1974) suggested a deep, rift-related source of fluorine, perhaps as volatiles released from crystallizing magma. Plumlee et al. (1995) conducted numerous geochemical models of fluorite deposition in the Illinois-Kentucky fluorospar district and concluded that significant fluorite deposition, both open-space filling and replacement of limestone, can only be accounted for by addition of HF to basinal brines. This addition gives rise to fluorine-rich acidic fluids capable of depositing fluorite by cooling and interaction with carbonate host-rocks. The source of HF in their model is from volatiles expelled from crystallizing alkaline magmas in the rift. They suggest a similar source of fluorine in other RRF deposits, including those in the RGR. We believe this hypothesis is credible for the source of fluorine in RGR deposits. Volcanism during early rift development (36–20 Ma) was dominated by silicic and then basaltic andesite volcanism, while during the last 10 Ma, magmatism has been characterized by alkali basalts along with minor silicic igneous activity (Seager et al., 1984; Anthony et al., 1992). Thus, alkaline magmatism in the RGR coincided with the most intense stage of fluorite deposition in the rift.

Another possible source of fluorine could be from fluorite and apatite in Precambrian basement rocks that could be leached by warm, acidic fluids. Apatite is a common constituent in most Precambrian granites, whereas fluorite is a rare accessory mineral in some areas (Condie and Budding, 1979). Fluorite also occurs as disseminations and in veins within Cambrian–Ordovician syenites in the Caballo Mountains (McLemore, 1986) and Cambrian–Ordovician carbonatites in the Lemitar and Chupadera Mountains (McLemore, 1987).

Probably the most controversial aspect regarding the origin of RGR deposits is the nature of the plumbing system responsible for mobilization, transport, and deposition of ore constituents. As mentioned above, it is generally agreed that mineralizing fluids began as meteoric water. However, beyond this, there is little agreement. Three general hypotheses have been proposed: (a) basin dewatering by compaction (McLemore and Barker, 1985; Norman et al., 1985; Lueth and Goodell, 1996), (b) topography-driven (gravity) flow (Morgan et al., 1985; Witcher, 1988; McMahan, 1989; McLemore, 1997), and (c) density-driven flow produced by thermal contrasts along the basin flanks (McLemore, 1997). In the basin-dewatering model, formation waters or basinal brines accumulated in rift and earlier sediments and were subsequently heated in part as a result of burial and compaction within the high heat-flow environment of the rift. These warm fluids after leaching mineral con-

stituents from sediments and crystalline rocks were ejected along faults and fractures, and unconformities during early diagenesis, later compaction, or uplift of sedimentary basins during extension (Hanor, 1979; McLemore and Barker, 1985; Norman et al., 1985).

An alternative model uses present-day rift geothermal systems as an analog model for the RGR deposits. Erosion and tectonic stripping of Mesozoic and early Cenozoic aquitards over structural highs created geohydrologic windows for discharge of deeply-circulating, forced-convective (gravity-driven) regional hydrothermal systems (Witcher, 1988). The circulation framework for these systems is largely independent of the predominantly lateral shallow hydrodynamics of clastic rift basins and utilizes fracture permeability with greater potential for vertical flow in pre-Tertiary rocks, especially in the upflow or discharge zones. Flow in these systems is topographically controlled with recharge in highlands and discharge in structurally high, pre-Tertiary bedrock or highly fractured Tertiary volcanic rocks in the lowlands.

The third model assumes that during the later thermal cooling or flexural stage of rifting, fluids of contrasting densities and temperatures are produced in aquifers of different permeabilities. Pearson and Garvin (1994) modeled fluid flow during this stage and found that density driven fluid-flow produced by temperature differences dominates in permeable aquifers, whereas in lesser permeable units compaction-driven flow continues. Pearson and Garvin (1994) suggested that the fluids could move long distances to the edge of the rift basins and mineralization could result. However, according to their model, density-driven flow occurs only in the thermal cooling stage of rifting when thermal contrasts develop as a result of cooling. The RGR has not reached this thermal cooling stage (Seager and Morgan, 1979; Chapin and Cather, 1984), and therefore this model is not appropriate for formation of RGR deposits.

In addition, the close association of a few RGR deposits with early rift silicic volcanic centers, such as the Organ cauldron, suggests that some RGR deposits may fit a volcanic-epithermal mineralization model (Buchanan, 1981; McLemore, 1993, 1996). However, the different ages between the mineralization associated with the Organ cauldron (32.2–30.8 Ma; McLemore et al., 1995; Lueth and McLemore, this guidebook) and the mineralization at Bishops Cap (5–6 Ma; Lueth et al., this guidebook), implies that different processes resulted in the deposits found in these areas.

With all of these models, the convecting waters leached barium, sulfate, lead, silver, and other ions from various source rocks such as arkosic sediments, evaporites, Precambrian rocks, and Precambrian mineral deposits. Mixing of magmatic and meteoric fluids may have occurred locally, thereby providing fluorine and other constituents. The mineralized fluids were advected into open-spaces such as bedding planes, faults, fractures, and solution collapse breccias. Putnam et al. (1983), Norman et al. (1985), and North and McLemore (1985) evaluated deposition mechanisms for the Hansonburg deposit. The results of these studies suggest that precipitation of RGR mineralization could have occurred as a result of cooling of the fluids, decrease in pressure, change in water chemistry, and/or mixing of the mineralized fluids with other subsurface brines or meteoric water. Any of these processes could have operated at any time during development of the basin and resulted in deposits of varying age.

Although concentrated brines are common in upper crustal sediments (e.g., oil/gas field brines), a question may arise as to the presence of such saline waters migrating in crystalline basement rock. Furthermore, if present-day geothermal systems in the RGR are analogs of systems responsible for RGR mineralization, then why are their fluids typically less than 2 eq. wt.% NaCl? Salinity of fluids from many RGR deposits and current day geothermal systems adjacent to mid-Tertiary silicic volcanic centers show less than

1–2 eq. wt.% NaCl. Several explanations may apply. In both cases formation waters may have been flushed out of basin rocks. Given enough time, other topographically driven systems and magmatically driven hydrothermal systems probably sweep the adjacent country rock with low salinity meteoric water as recharge inflows. Seismogenically generated fluids from deep sources in the Precambrian basement may help explain higher, but highly variable salinities of mineralizing fluids (Sibson, 1987). These flows would be transient and repeated many times over the life span of a mineralizing hydrothermal system. In periods of seismic quiescence when deep fluid sources are sealed or constricted by a silica or pressure, cooler, but still hot, lower salinity fluids would dominate the near surface thermal regime.

### CONCLUDING REMARKS

Just as the RGR has evolved through the Cenozoic, the nature of heat redistribution by ground-water advection has also been dynamic. Numerous factors, such as surface topography, development of rift structural architecture, erosion and tectonic elimination of aquitards, unroofing of carbonate aquifers by karsting, climate, rates of extension, basin sedimentation, subsidence, compaction, overpressuring, and crustal magmatism, may have played roles in the location, types, and intensity of Cenozoic RGR deposit evolution at any particular site. As a result, deposits of different ages, size, temperature of formation, and accessory mineralogy are a common feature of the RGR deposits.

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View to the northwest of the Sierra de las Uvas, composed of the Oligocene Bell Top Formation and Uvas Basaltic Andesite in the background and the late Miocene Rincon Valley Formation and Plio-Pleistocene Camp Rice Formation in the foreground. Photograph by Greg Mack.