A Strontium Isotope Evolution Model for Cenozoic Magma Genesis, Estern Great Basin, U.S.A. *

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Abstract

Cenozoic volcanism in the Great Basin is characterized by an outward migration of volcanic centers with time from a centrally located core region, a gradational decrease in the initial Sr^{87}/Sr^{86} ratio with decreasing age and increasing distance from the core, and a progressive change from calc-alkalic core rocks to more alkalic basin margin rocks. Generally each volcanic center erupted copious silicic ignimbrites followed by small amounts of basalt and andesite. The Sr^{87}/Sr^{86} ratio for old core rocks is about 0.709 and the ratio for young basin margin rocks is about 0.705. Spatially and temporally related silicic and mafic suites have essentially the same Sr^{87}/Sr^{86} ratios.

The locus of older volcanism of the core region was the intersection of a north-south trending axis of crustal extension and high heat flow with the northeast trending relic thermal ridge of the Mesozoic metamorphic hinterland of the Sevier Orogenic Belt. Derivation of the Great Basin magmas directly from mantle with modification by crustal contamination seems unlikely. Initial melting of lower crustal rocks probably occurred as a response to decrease in confining pressure related to crustal extension. Volcanism was probably also a consequence of the regional increase in the geothermal gradient that is now responsible for the high heat flow of the Basin and Range Province.

High Sr isotopic ratios of the older core volcanic rocks suggests that conditions suitable for the production of silicic magmas by partial fusion of the crust reached higher levels within the crust during initial volcanism than during production of later magmas with lower isotopic ratios and more alkaline chemistry. As the Great Basin became increasingly attenuated, progressively lower portions of the crust along basin margins were exposed to conditions suitable for magma genesis. The core region became exhausted in low tem-

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perature melting components, and volcanism ceased in the core before nearby areas had completed the silicic-mafic eruption cycle leading to their own exhaustion of crustal magma sources.

Introduction

Initial Sr^{87}/Sr^{86} ratios of igneous rocks are limited indicators of chemical characteristics of magma source rocks. An initial isotope ratio represents the integrated accumulation of Sr^{87} from Rb^{87} in a rock system and its predecessors throughout geologic time; because of the concentrations of Sr and Rb depend upon the highly variable chemical evolution or rock systems, the initial isotopic ratio thus is a consequence of an unknown, possible complex chemical evolution. In rare circumstances Sr isotopes can provide an unambiguous indication of a mantle or crustal origin of magmas, but in most cases supporting information for magma genesis models is required from other isotopic, chemical, petrographic, and geologic observations of the rock system and its geologic environment. We have examined the temporal and spatial relationships of initial Sr^{87}/Sr^{86} ratios and bulk chemistry of eastern Great Basin volcanic rocks and will discuss a possible model for the magmatic evolution of the region.

Regional Geologic Framework

The Great Basin is a region of internal drainage that covers most of Nevada and parts of Utah and California (Fig. 1). Precambrian basement with an age of 2000-2500 m.y. underlies the northern half of the eastern Great Basin and 1500-17000 m.y. old basement underlies the southern half (STACEY *et al.*, 1968), but these rocks are only rarely exposed at the surface. Throughout the Paleozoic, geosynclinal deposits accumulated in the Cordilleran mio-geosynoline along the western margin of the stable continental interior. The change from shelf to geosyncline is roughly coincident with the eastern edge of the Sevier Orogenic Belt of Cretaceous age (Fig. 2). The Missisippian Antler Orogenic belt is a zone where western eugeosynclinal Paleozoic facies are tectonically intermingled with miogeosynclinal rocks (GILLULY, 1963). It forms a convenient western boundary for the eastern Great Basin, the area with which we are concerned. During Mesozoic time miogeosynclinal rocks were deformed, metamorphosed, and thrust eastward over thin stable shelf sediments to form the Sevier Orogenic Belt and its hinterland (ARMSTRONG, 1968*a*). West of the folded and overthrust Sevier belt a northeastward trending series of exposures of a metamorphic infrastructure developed during the

FIG. 1 - Cenozoic normal fault pattern in the western United Statdcs. The heavy dashed line delineates the extent of internal drainage that defines the Great Basin. The Colorado Plateau is bordered by the lighter dashed line. Approximately 50 to 90 kilometers of crustal extension is recorded in displacements in surficial Cenozoic normal faults across the Great Basin. Modified after GILLULY (1963).

Mesozoic in Paleozoic and Precambrian rocks extends across the Great Basin from southeastern California to Idaho (ARMSTRONG, 1968*b*; ARMSTRONG and HANSEN, 1966). The predominantly Cenozoic K-Ar dates from rocks of this Mesozoic metamorphic belt (ARMSTRONG and

HILLS, 1967) probably represent the time of uplift, erosion, and cooling associated with Cenozoic block faulting; the metamorphic rocks are proof of a broad and irregular thermal ridge which developed within the crust during the Mesozoic; the young K-Ar dates indicate that the thermal anomaly persisted into the Cenozoic, and may well

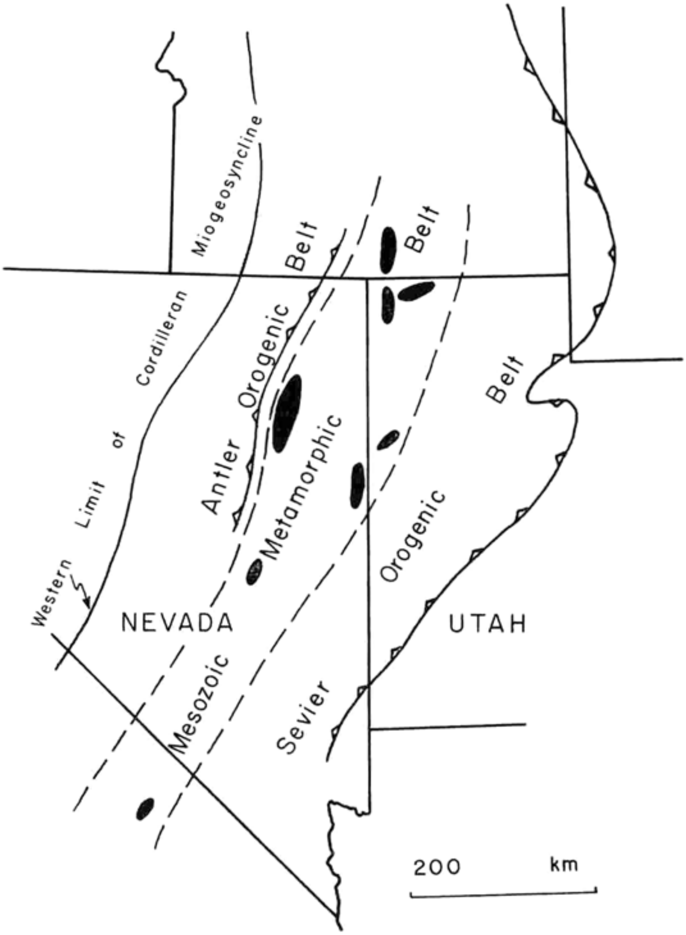

FIG. 2 - Pre-Cenozoic tectonic framework. The Paleozoic Cordilleran geosyncline is represented by a miogeosynclinal carbonate sequence that is much thicker than the sediments on the craton east of the Sevier Orogenic Belt. Eugeosynclinal Paleozoic rocks are tectonically intermingled with mio-geosynclinal rocks in the Mississippian Antler Orogenic Belt. The Cretaceous Sevier Orogenic Belt and its hinterland consist of folded, thrust, and metamorphosed Paleozoic strata. Ranges that contain Paleozoic and Precambrian rocks metamorphosed during the Mesozoic are indicated by the solid pattern. After ARMSTRONG (1968*a*).

have been of significance for the Cenozoic magma genesis that will be discussed later.

The Basin and Range Province, which includes the Great Basin, is a Cenozoic rift system of north trending grabens and horsts formed by regional east-west crustal extension (COOK, K. L., 1965; THOMPSON, 1965); between 50 and 90 kilometers of crustal extension along normal faults (THOMPSON, 1959; J. GILLULY, oral communication, 1969) has been estimated for the widest portion of the Great Basin (Fig. 1). Normal faulting began in early Oligocene time (NOLAN, 1943), reached a maximum in Mio-Pliocene time (MOORES, SCOTT, and LUMSDEN, 1968) and is presently active (THOMPSON, 1965).

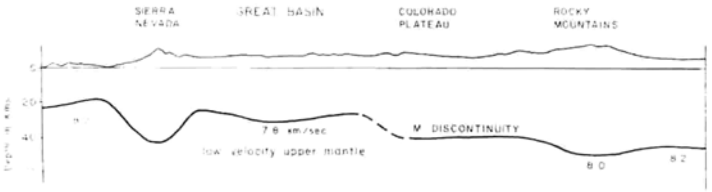

FIG. 3 - Crustal and mantle structure in an east-west Great Basin cross section. The thinner crust and low velocity upper mantle are the distinctive characteristics beneath the Great Basin. Compiled from EATON (1965), ROLLER (1965) and JACKSON and PAKISER (1965) by BLACKWELL (1969).

Underlying this surficial rift system is a relatively thin crust (25-30 km thick) and a low-velocity upper mantle (7.6-7.8 km/sec P-wave velocity) (COOK, K. L., 1965; EATON, *et al.*, 1965; ROLLER, 1965; JACKSON and PAKISER, 1965) (Fig. 3). The Basin and Range Province lies on a zone of relatively high heat flow (Fig. 4) (1.5 to 3.0 μ cal/cm^2 sec) with boundaries roughly coincident with the boundaries of the Basin and Range Province (BLACKWELL, 1969).

Extensive Cenozoic volcanism covered most of the Great Basin with wide-spread sheets of voluminous dacitic to rhyolitic ignimbrites and smaller andesite and basalt flows (MACKIN, 1960; COOK, E. F., 1965; ARMSTRONG *et al.*, 1969). Individual ignimbrite cooling units average about 50 meters thick but attain thicknesses as great as 800 meters in some localities (COOK, E. F., 1965; SCOTT, 1965) and cover as much as 50,000 square kilometers. In several localities the total volcanic accumulations are over 1.5 km thick. A conservative estimate of the amount of volcanism suggests that 200,000 square kilometers of the Great Basin were covered with an average of 500 meters of volcanic material representing about 100,000 cubic kilometers of eruptive rock. These volcanic rocks include calcic to peralkalic suites (NOBLE *et al.*, 1965; NOBLE *et al.*, 1968).

The distribution of volcanic centers for silicic Cenozoic volcanism followed a distinct trend in time and space within the Great Basin (ARMSTRONG *et al.*, 1969). Most older (30-40 m.y.) volcanic centers are grouped in a core area centrally located in the Great Basin (Fig. 5). Subsequent centers are found progressively farther from the core with decreasing age; a marked absence of younger volcanic rocks characterizes the inactive core. The simple radial pattern of time transgressive volcanism is best documented for regions south of the core zone, including the area of concern in this paper; north and

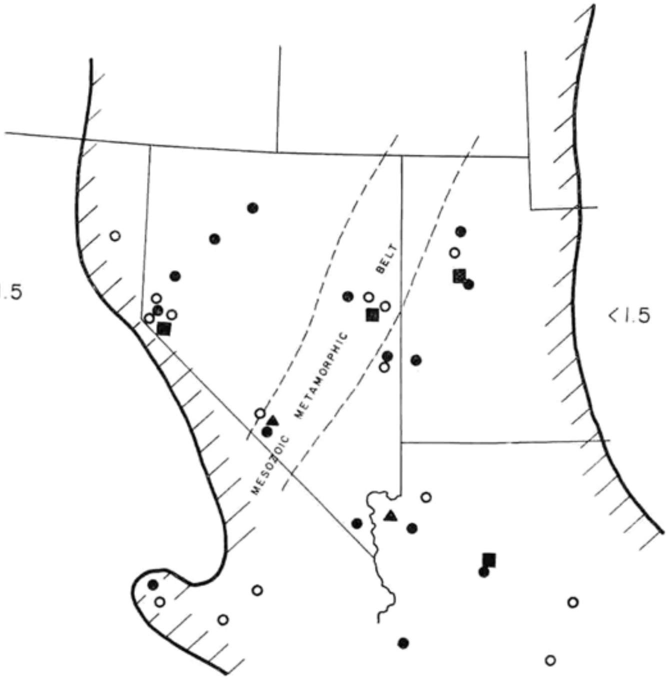

FIG. 4 - Heat flow in the Great Basin and vicinity. The hatched boundaries enclose a region of heat flow greater than 1.5 μ cal/cm² sec. The open circles indicate heat flow measurements of 1.5-2.0 μ cal/cm² sec.; filled circles, 2.0-2.5; triangles, 2.5-3.0; and squares > 3.0. The dashed borders outline the region where the thermal ridge formed during Mesozoic metamorphism persisted into the Tertiary; most of the K-Ar dates for these rocks with Mesozoic metamorphic fabrics are Tertiary. After BLACKWELL (1969).

east of the core zone the volcanic history is more complex. Most centers initially extruded voluminous silicic ignimbrites and lavas followed by less voluminous mafic lavas - the « capping lavas » of VITALIANO (1969).

The significance of the superimposed rift system, volcanism, and

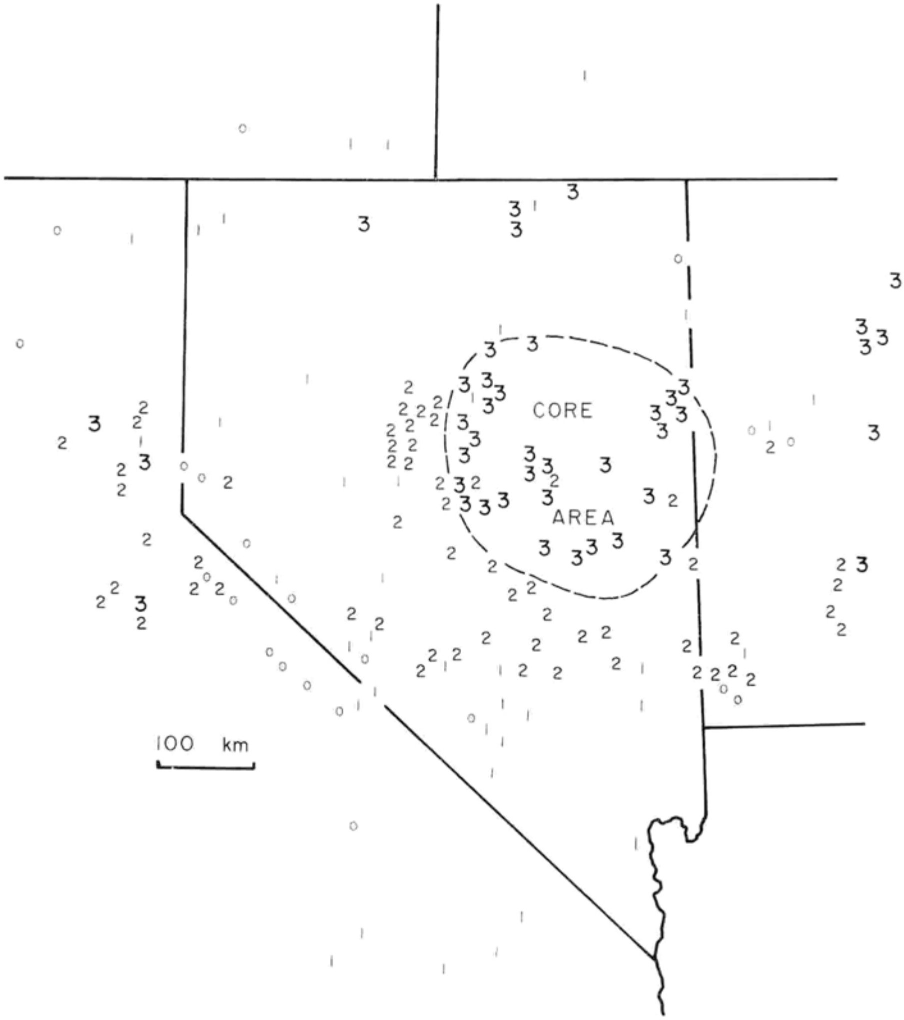

FIG. 5 - Age distribution of silicic volcanic rocks in the Great Basin. The 30 to 40 m.y. old volcanic centers are indicated by « 3 », the 20 to 30 m.y. centers by « 2 », the 10 to 20 m.y. centers by « 1 », and the 0-10 m.y. centers by « 0 ». The core area is delineated by the dashed line. After ARMSTRONG *et al.* (1969).

TABLE I

UNIT	LOCATION	SAMPLE	Sr^{87}/Sr^{86} [1] (observed ratio)	Rb + (ppm) (XRF)	Sr + (ppm) (XRF)	Rb/Sr	K-Ar [2] date, m.y.	Sr^{87}/Sr^{86} (initial ratio)
SILICIC UNITS								
Stone Cabin Fm. [4]	Lat. 38°26′20″N.	vitric rock	0.7145	212	104	2.03	—	0.7121
Rhyolitic ignimbrite	Long. 115°16′55″W.	vitric rock	0.7153*	—	—	—	32.4±0.6	0.7121
	Grant Range, Nye		0.7098	6	39	0.15	—	0.7098
	Co., Nevada		0.7093	—	—	—	—	0.7093
Windous Butte Fm. [4]	Lat. 38°36′20″N.	vitric rock	0.7100	252	220	1.15	—	0.7097
Rhyolitic ignimbrite	Long. 115°16′55″W.	plagioclase	0.7078	3	983	0.003	32.0±0.5	0.7078
	Grant Range, Nye	plagioclase	0.7071	—	—	—	—	0.7071
	Co., Nevada							
Needles Range Fm. [4]	Lat. 38°35′54″N.	vitric rock	0.7096	151	441	0.342	—	0.7092
Basal dacitic	Long. 115°17′0″W.	devitrified rock	0.7084	156	444	0.351	—	0.7080
ignimbrite	Grant Range, Nye	glass	0.7099	194	305	0.636	—	0.7092
	Co., Nevada	plagioclase	0.7088	2	1579	0.001	29.7±0.3	0.7088
		plagioclase	0.7099	—	—	—	—	0.7099
		biotite	0.7264*	213	20	10.7	—	0.7141
		hornoblende	0.7118*	20	137	0.15	—	0.7116
Leach Canyon Fm. [5]	Lat. 37°51′0″N.	vitric rock	0.7102	212	229	0.925	24.0±0.5	0.7091
Rhyolitic ignimbrite	Long. 115°1′20″W.	vitric rock	0.7105	—	—	—	—	0.7094
	White River Narrows							
	Lincoln Co., Nev.							
Swett Ibnimbrite [5]	Lat. 37°50′20″N.	vitric rock	0.7071	169	266	0.622	23.3±0.2	0.7064
Rhyolitic ignimbrite	Long. 115°1′55″W.	plagioclase	0.7049	2	1940	0.001	—	0.7049
	White River Narrows							
	Lincoln Co., Nev.							
Bauers Ignimbrite [5]	Lat. 37°50′20″N.	vitric rock	0.7069	202	280	0.721	—	0.7060
	Long. 115°1′55″W.	devitrified rock	0.7082	175	351	0.498	—	0.7077
	White River Narrows	devitrified rock	0.7077	—	—	—	21.5±0.5	0.7071
	Lincoln Co., Nev.	glass	0.7086	235	107	2.20	—	0.7062
		plagioclase	0.7062	13	1743	0.007	—	0.7062
Harmony Hills Ignim. [5]	Lat. 37°51′35″N.	devitrified rock	0.7094	101	739	0.136	—	0.7092
Dacitic ignimbrite	Long. 114°19′45″W.	devitrified rock	0.7092*	—	—	—	—	0.7091
	Condor Canyon	devitrified rock	0.7093*	—	—	—	21.0±0.2	0.7092
	Lincoln Co., Nev.	plagioclase	0.7083	0	2052	0	—	0.7083
		plagioclase	0.7081*	—	—	—	—	0.7081
		biotite	0.7166*	320	30	10.7	—	0.7084
Racer Canyon Tuff [3]	Racer Canyon,	plagioclase	—	—	—	—	20.3±0.5	0.7053
	Washington Co.,							
	Utah							

	Range, Lincoln Co. Nevada		—	—				
Ox Valley Tuff [4]	Lat. 37°35′N. Long. 114°20′W. Caliente depression, Nevada	sanidine	—	—	—	—	15.1±0.4	0.7054
MAFIC UNITS								
Trachytic basalt [6]	Lat. 39°30′55″N. Long. 115°57′20″W. N.E. of Eureka Eureka Co., Nev.	whole rock	0.7089	56	854	0.065	34.8±0.7	0.7089
Basaltic andesite [8]	Lat. 39°11′50″N. Long. 115° 23′45″W. Illipah Area, White Pine Co., Nevada	whole rock	0.7094	39	610	0.064	33.9±0.7	0.7094
Trachytic basalt	Lat. 38°3′10″N. Long. 115°24′30″W. Golden Gate Range Nye Co., Nevada	whole rock	0.7076	67	450	0.15	20.1±0.5	0.7076
Basaltic andesite [5]	Lat. 37°15′50″N. Long. 115°27′40″W. Meadow Valley Wash Lincoln Co., Nev.	whole rock	0.7061	159	961	0.165	8.5±0.3	0.7061
Basaltic andesite [7]	Lat. 35°50′50″N. Long. 115°35′20″W. 15 miles SE of Hoover Dam, highway 93-466, Arizona	whole rock	0.7048	12	510	0.024	5.8>1.0	0.7048
Camptonite dike [7]	Lat. 35°55′25″N. Long. 114°37′55″W. 9 miles SE of Hoover Dam highway 93-466, Arizona	hornblende (Kaersutite) whole rock	0.7024 0.7024 0.7035*	— — 36	— — 133	— — 0.27	4.7±0.3 — —	0.7024 0.7024 0.7035

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* - Determined at Australia National University, Canberra.

· - Determined at University of Adelaide, Adelaide, Australia.

1 - normalized $Sr^{88}/Sr^{86} = 0.1194$.

2 - ARMSTRONG (1969).

3 - Sr and K-Ar data from NOBLE *et al.* (1968).

4 - Stratigraphic information; SCOTT (1966) and COOK, E. F. (1965).

5 - Stratigraphic information; COOK, E. F.

6 - Stratigraphic information; NOLAN (1962).

7 - Stratigraphic information; LONGWELL (1963).

8 - Stratigraphic information; HUMPHREY (1960).

thin crust was emphasized by MENARD (1961) when he suggested that an extension of the East Pacific Rise under the continental crust of the western United States produced these features. Since that time additional information concerning the nature of the crust, upper mantle, and heat flow in the western United States has strengthened convictions that all these phenomena, including magma generation, are integrally related as part of a complex major tectonic system (COOK, K. L., 1968; MACKENZIE and MORGAN, 1969).

For the Sr isotopic studies a sequence of Tertiary silicic and mafic volcanic rocks were selected that represents a time and space profile from the Grant Range in the core region, southeastward toward the margin of the Great Basin near the Utah-Arizona-Nevada border junction. The petrology and chemical characteristics of the silicic rocks studies have been investigated in considerable detail by SCOTT (1965 and 1966) and NOBLE (1968, and *et al.*, 1967) but only the generalized results of those studies having specific bearing on problems of magma genesis will be dealt with in this paper.

Analytical Methods

Rock, glass, and mineral separates were analyzed for rubidium/strontium ratios by X-ray fluorescence at Univ. of Adelaide using the NBS feldspar 70 ($Sr = 65$ ppm and $Rb = 530$ ppm) as a standard. Strontium-isotope compositions were determined with a 6 inch radius, Nier-type mass spectrometer equipped with expanded scale recorder. The Eimer and Amend standard $SrCO_3$ (lot No. 492327) was analyzed several times during the study and gave an average Sr^{87}/Sr^{86} ratio of 0.7073 ± 0.0001. To correct for isotopic fractionation all results reported are normalized to $Sr^{88}/Sr^{86} = 0.1194$. Replicate analyses indicate that the reproducitility for the unspiked Sr^{87}/Sr^{86} ratio measurements is approximately $\pm$ 0.0006 (σ).

The normalized Sr^{87}/Sr^{86} ratio, initial Sr^{87}/Sr^{86} ratio, Rb/Sr ratio, K-Ar date and location of the samples are listed in Table 1.

Internal Isotopic Variation within Individual Units

Before we can discuss the regional trends of initial Sr^{87}/Sr^{86} ratios we must consider the question of the exact significance of the values measured for whole rock and separated phases of individual ash flows. HEDGE and NOBLE (1967) have observed variation in the isotopic composition of compositionally zoned ignimbrite cooling units of the southern Great Basin: a 0.006 to 0.008 decrease in isotopic ratio was observed from the base to the top of vertical sections of compositionally zoned, voluminous, compound cooling units. The more calcic upper portions of the units gave lower, presumably less contaminated ratios. In the present study this problem or internal variation related to magma chamber zoning and reaction of low Sr magmas with radiogenic wall rocks is proably less serious. All of the ignimbrites studied are either simple cooling units with no apparent compositional zoning or compound cooling units with very slight compositional zoning evident in vertical profiles. Moreover, the Sr contents are relatively high, even in the slightly zones units, and this should diminish the effects of high level crustal contamination or contamination with ground water Sr.

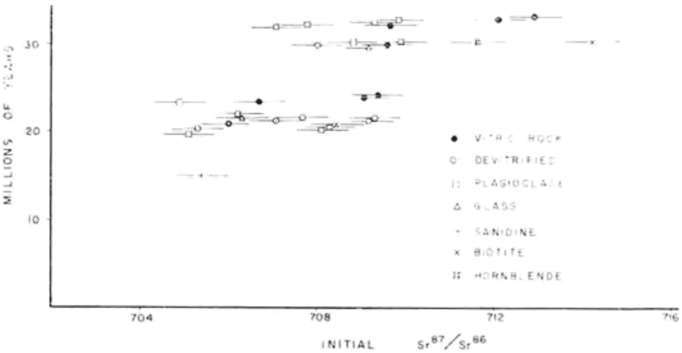

FIG. 6 - Evolution of Sr isotopes of silicic volcanic rocks. The horizontal lines indicate the range of Sr isotope error. The youngest three points are from NOBLE *et al.* (1968). Clusters of overlapping points are separated on the vertical scale enough to distinguish between them.

The analysis of separate mineral phases as well as whole rock samples provides a worthwhile check on processes that might disturb the isotopic composition of pyroclastic volcanic rocks. These processes include late mixing of magmas from distinct sources, high-level crustal contamination with radiogenic Sr from magma chamber or vent wall rock reactions, and alteration of groundmass isotopic composition by ground water equilibrated with Sr from the Paleozoic limestones that underlie the region studied. The data of Table 1 demonstrate that variation of Sr isotopic composition between phases of individual ignimbrites is quite common. The generalization is evident that the feldspars are usually, but not invariably, slightly lower (0.001-0.002) than whole rocks in Sr^{87}/Sr^{86} ratio (Fig. 6). Therefore, a modest amount of high-level crustal contamination has occurred so that the exact ratios observed for specific samples cannot have too much significance attached to them. Similar disequilibrium between feldspars and the parent magma has been noted by DASCH (1969). The isotopic composition trends for rock and feldspar, however, are identical and thus the trends are real and worthy of further discussion. Rather large excesses of radiogenic Sr observed in mafic mineral separates from the Needles Range Formation are further evidence of lack of isotopic equilibration between the phases deposited by ash flows as a consequence of complex contamination processes. A corollary of this observation of lack of interphase equilibrium is the doubtful validity of the comparison of phase chemical compositions with experimental systems when discussing petrogenesis.

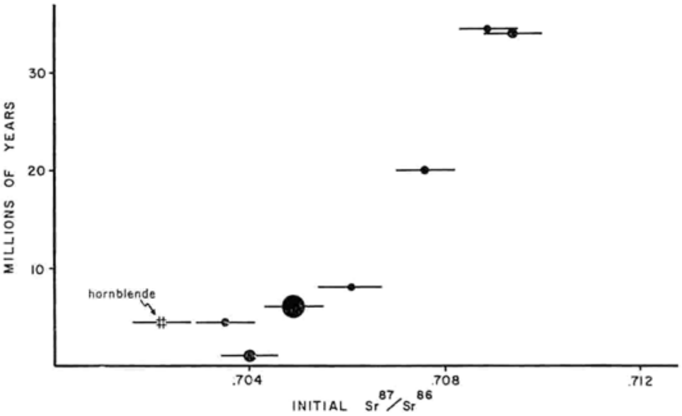

FIG. 7 - Evolution of Sr isotopes in mafic volcanic rocks. The horizontal lines indicate the range of Sr isotope error. The size of the black spot indicates the precision of the K-Ar age measurement. All isotopic analyses are of whole rocks except the indicated hornblende. The youngest point represents the average of analyses of late Cenozoic alkali-olivine basalts by MANTON and LEEMAN (1969).

Variation of Initial Sr^{87}/Sr^{86} Ratio in Time and Space

Pronounced trends are found in the initial Sr^{87}/Sr^{86} ratios for rocks and phase separates of both silicic and mafic rocks. Sr^{87}/Sr^{86} ratios of silicic rocks decrease from about 0.709 to 0.705 with decreasing age during a 20 million year period (Fig. 6). Associated mafic rocks have ratios that decrease from about 0.709 to 0.704 during a 35 million year period (Fig. 7). The similarity in these trends may be seen by comparing Figures 6 and 7. Lower initial Sr^{87}/Sr^{86} ratios are characteristic of the younger rocks which are found nearer the margins of the Great Basin (Fig. 8). The chemistry of volcanic rocks from the Great Basin changes from the predominantly calc-alkalic rocks erupted in the core area to alkalic rocks (even comendites and pantellerites) closer to the margins (Fig. 9). This relationship between location of volcanic centers and chemistry characterizes the suite of units chosen for this study; for example, the Ox Valley Tuff and the camptonite dike are considerably more alkalic than the Stone Cabin and Windows Butte Formations of the Core region (Table 1). Thus, Great Basin calc-alkalic volcanic suites have initial Sr^{87}/Sr^{86} ratios near 0.709 and alkalic suites have 0.705 ratios. This general correlation between alkalic magmas and low initial Sr^{87}/Sr^{86} ratios conforms with most observations of alkalic lavas (Hamilton, 1965, p. 119-120); Hamilton (1968) reports a range of 0.703 to 0.705 and a mode of 0.704 for Sr^{87}/Sr^{86} for most alkalic basalts, their differentiates, and carbonitites.

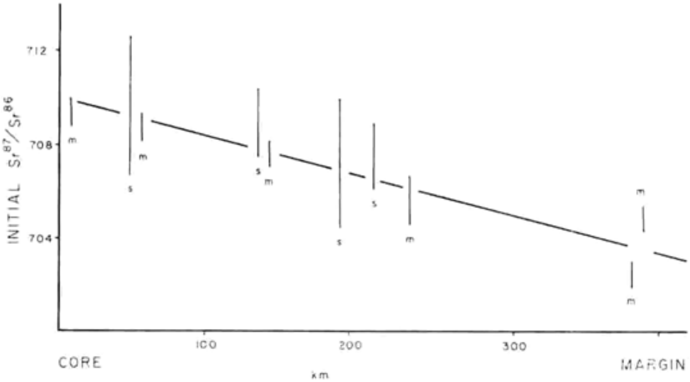

Fig. 8 - Lateral change in Sr isotopic composition from the core southeastward to the margin of the Great Basin. Silicic volcanic centers are represented by « s », and « m » represents mafic volcanic centers. The vertical lines represent the precision of Sr isotopic measurements.

Great Basin volcanic rocks generally have higher initial Sr^{87}/Sr^{86} ratios (0.707 ± 0.003) than typical continental volcanic rocks (0.7045 ± + 0.0015) (Hedge, 1966). This is in agreement with the observations of Eastwood (collaborator in Damon, 1968, p. 54-56) and Bikerman

(1967) that the ratios of volcanic rocks in the Basin and Range Province of Arizona are higher (0.707 ± 0.003) than typical for continental rocks. On the Colorado Plateau and in eastern Arizona (east

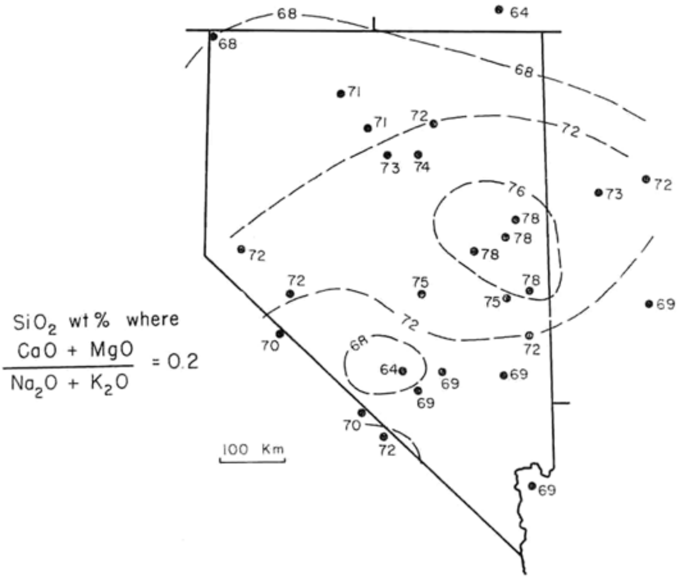

FIG. 9 - Average chemical trends of volcanic rocks in the Great Basin. Chemical trends taken from literature (listed in abbreviated from below) were plotted on a graph of SiO_2 weight % versus $CaO + MgO / Na_2O + K_2O$. The SiO_2 % at the intercept of the average trend with $CaO + MgO / NaO_2 + K_2O = 0.2$ is indicated on the map. The index of 76-78 is typical of the calc-alkalic suites. The 68-64 index characterizes alkalic rocks. Literature references: Callahan, 1937, Univ. Nevada Bull., VXXXI, No. 5, p. 1; Callaghan, 1939, Am. Geophys. Union Trans., Vol. 20, p. 438; Carmichael, 1967, Contr. Mineral. Petrol., Vol. 14, p. 36; Christiansen and Noble, 1964, Geol. Soc. America Spec. Paper 82, p. 246; Cook, 1969, Geol. Soc. America Memoir 116, p. 107; Drewes, 1963, U.S. Geol. Survey Prof. Paper 413; Drewes, 1967, U.S. Geol. Survey Prof. Paper 557; Gilbert, 1941, Geol. Soc. America Bull., Vol. 52, p. 781; Gilluly and Masursky, 1965, U.S. Geol. Survey Bull. 1175; Gilluly and Gates, 1965, U.S. Geol. Survey Prof. Paper 465; Hotz and Willden, 1964, U.S. Geol. Survey Prof. Paper 431; Lindgren and Loughlin, 1919, U.S. Geol. Survey Prof. Paper 107; Lipman, 1966, Am. Jour. Sci., Vol. 264, p. 810; Lipman and Christiansen, 1964, U.S. Geol. Survey Prof. Paper 501-B, p. 74; Longwell, 1963, U.S. Geol. Survey Prof. Paper 374; Moore, Lamphere, and Obradovich, 1968, Econ. Geology, Vol. 63, p. 612; Muffler, 1964, U.S. Geol. Survey Bull. 1179; Noble *et al.*, 1964, Geol. Soc. America Spec. Paper 82, p. 143; Noble, 1965, U.S. Geol. Survey Prof. Paper 525B, p. 85; Noble, 1966, Geol. Soc. America Spec. Paper 87, p. 117-118; Noble *et al.*, 1968, Science, Vol. 160, p. 1337; Nolan, 1935, U.S. Geol. Survey Prof. Paper 177; Roberts, 1964, U.S. Geol. Survey Prof. Paper 459-A, p. 1; Ross, 1961, Nevada Bureau Mines Bull. 59; Scott, 1965, Ph. D. thesis, Rice Univ., Houston, Texas; Stoatz and Carr, 1964, U.S. Geol. Survey Prof. Paper 415; Thompson and White, 1964, U.S. Geol. Survey Prof. Paper 458A, p. 1; Walker, 1961, U.S. Geol. Survey Prof. Paper 424C, p. 142C; Willden, 1964, Nevada Bureau Mines Bull. 59.

and north of the Basin and Range Province) PUSHKAR and DAMON (collaborators in DAMON, 1969, p. 25-26) observe lower values (0.7033 to 0.7054, average of 0.704). In the regions adjacent to Arizona's rift system (EASTWOOD, collaborator in DAMON, 1968, p. 54-56) and in the rift system (BICKERMAN, 1967) Sr isotope initial ratios show a consistent change with time: the rocks have a linear trend from a 27 m.y. porphyry with a 0.709 initial ratio to a 10 m.y. basalt with a 0.704 initial ratio. The extra-rift rocks define a linear trend from a 27 m.y. basaltic andesite with a 0.7054 initial ratio to a 2 m.y. basalt with a 0.7033 initial ratio. The Arizona rift trend is indistinguishable from that of the Great Basin rocks; the extra-rift Arizona trend is comparable to Great Basin rocks except for the lower Sr^{87}/Sr^{86} ratios of the Arizona rocks. These isotopic data suggest that similar evolutionary processes have been active under these regions. The crustal differences between the Colorado Plateau and the Great Basin (and Basin and Range Province) may explain the different range of isotopic ratios as outlined below.

HEDGE (1966) considers that most continental felsic volcanic rocks are significantly enriched in radiogenic Sr and therefore, have been derived within the crust or appreciably contaminated by crustal material. EASTWOOD (collaborator in Damon 1968, p. 54-56) concludes that the higher isotopic ratios of Basin and Range samples are indicative of a lower crustal magma source or mantle-derived magmas diluted by large amounts of crustal material. PUSHKAR (1966) postulates from the isotopic ratios of Central American ignimbrites that most of their magmas might have formed by melting of crustal rocks, or by large scale assimilation of crustal rocks by mantle-derived magmas. The consensus of these authors is that specific magma sources cannot be identified from the isotopic data alone.

Eugeosynclinal sedimentary rocks have Sr and Rb concentrations and Sr^{87}/Sr^{86} ratios compatible with the hypothesis that these rocks are the source of anatectic granitic magmas in orogenic belts (PETERMAN, *et al.*, 1967; EWART and STIPP, 1968). But Paleozoic miogeosynclinal strata overlie Precambrian crystalline rocks in the eastern Great Basin and, therefore, Phanerozoic eugeosynclinal sediments cannot be a source for western Great Basin silicic volcanic rocks (Fig. 2).

MANTON and LEEMAN (1969) and MANTON (1967; and oral communication, 1969) report that late Cenozoic alkali-olivine basalts and trachyandesites from the Great Basin and Basin and Range Province have an average Sr^{87}/Sr^{86} ratio of 0.704 with 25 ppm Rb and 780 ppm Sr. These volcanic rocks are younger than virtually all those shown in figure 5 and occur as small, local cones and flows throughout the region. They commonly contain olivine nodules; this has not been observed in any of the rocks studied by us. Manton and Leeman suggest that the low ratios of the alkalic basalts indicate derivation from the mantle. If this conclusion is correct and the mantle isotopic composition has remained relatively constant in the Cenozoic, then it follows that older Great Basin extrusive suites with higher initial ratios must be derived either from crustal rocks or mantle rocks contaminated by crustal rocks.

Contamination schemes can be devised that explain any individual Sr isotope ratio, but consistent trends of the spatial and temporal evolution of Sr isotopes cannot be the result of random contamination processes. The similarity of isotopic compositions of associated silicic and mafic extrusive rocks requires that mantle derived silicic parent magmas must consistently assimilate about half as much radiogenic Sr-rich crust per unit volume as the associated less voluminous mafic magmas. Using the contamination scheme of Pushkar (1966) (also used by Ewart and Stipp, 1968, *p.* 730) the fraction of contaminating crust can be calculated by the equation:

$$\left(\frac{Sr_m}{Sr_m + X\,Sr_c}\right)(Sr^{87}/Sr^{86})_m + \left(\frac{X\,Sr_c}{Sr_m + X\,Sr_c}\right)(Sr^{87}/Sr^{86})_c = (Sr^{87}/Sr^{86})_{cm}$$

where the following values and definitions are used:

	contamination models	
	mafic magma	*silicic magma*
Sr_m = Sr in parent magma	800 ppm (7)	300 ppm (7)
Sr_c = Sr in contaminant	440 ppm (5)	440 ppm (5)
$(Sr^{87}/^{86})_m$ = ratio of parent magma . . .	0.704 (6)	0.704 (6)
$(Sr^{87}/Sr^{86})_c$ = ratio of contaminant . . .	0.725 (6)	0.725 (6)
$(Sr^{87}/Sr^{86})_{cm}$ = ratio of contaminated magma .	0.709 (7)	0.709 (7)

X = parts of contaminant for one part parent magma.

Parent magma = continental basalt (5, 6) silicic differentiates (?) of continental basalt or possibly mantle derived silicic magma.

Contaminant = average silicic crust (5, 6).

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(5) Turekian and Kulp (1956).

(6) Faure and Hurley (1963).

(7) Measured by us for Oligocene silicic and mafic volcanic rocks.

For the mafic contamination model, X = 0.5; this requires that 33 % of the Oligocene mafic extrusive rocks of the Great Basin consists of average silicic crust contaminant approximately andesitic in composition. Obviously a basalt altered enough to have the required isotopic ratio would no longer resemble an ideal basalt and might be rather similar to the alkali and alumina rich mafic volcanic rocks with which we are concerned. The silicic contamination model yields X = 0.2 or about 17 % average silicic crust, an amount that might easily be present yet not distinguishable from the parent magma in terms of bulk chemistry. To explain the observed trends by a contamination model requires that the contamination process be systematically linked with geography and time in such a way that contamination decreases progressively towards the margins of the Great Basin as time passes. Because of the rather arbitrary assumptions involved in discussing such contamination models on a quantitative basis it is pointless to argue whether the isotopic and chemical data themselves are proof or disproof of a given hypothesis.

POWELL (1969) recognizes several criteria by which possible contamination of magmas may be recognized; these include a positive relationship between Sr^{87}/Sr^{86} and Rb/Sr ratios and between Sr^{87}/Sr^{86} ratios and SiO_2 wt. % and an inverse hyperbolic relationship between Sr^{87}/Sr^{86} ratios and Sr ppm. The scattered plot of initial Sr^{87}/Sr^{86} ratios versus Sr ppm (Fig. 10A) shows a weak inverse trend suggestive of limited contamination. However, when the initial rock ratios are replaced by the initial ratios of their respective coexisting plagioclase (shown by arrowhead), this weak trend is no longer present. As discussed above, a modest amount of high-level crustal contamination has occurred and is characterized by isotopic disequilibrium between plagioclase and rock. The absence of any trend between plagioclase initial ratios and Sr ppm suggests that most of the contamination occurred after formation of plagioclases. A weak positive slope exists between initial Sr^{87}/Sr^{86} ratios and Rb/Sr ratios (Fig. 10B) even if plagioclase initial ratios are used. We disagree with Powell's conclusion that a positive relationship in this graph is indicative of contamination. A graph of Rb/Sr ratios versus Sr^{87}/Sr^{86} ratios of lavas derived from different levels in a differentiated crust (lower Rb/Sr ratios with greater depths) would produce a positive slope. Therefore, application of Powell's criteria to isotopic data presented in this paper supports evidence of limited, high-level crustal contamination and does not support a contamination hypothesis for the origin of Sr isotopic trends of Great Basin volcanic rocks.

Obviously, the nature of the low velocity upper mantle under the Great Basin (Fig. 3) is very critical to hypotheses of magma generation. ARCHAMBEAU and DAVIES (1969) consider the low velocity zone to consist of a region of partial melt, extending from 35 to 160 km in depth that was produced by depression of the melting curve associated with dehydration of upper continental mantle. They suggest that decoupling between 100 and 150 km and sliding of the continental upper mantle over a region of high heat flux caused the melting. COOK, K. L. (1965) postulates that the low velocities represent a mantle-crust mix, but the nature of this mix is not clear. The low velocity upper mantle of the Basin and Range region is similar in geophysical properties to the upper mantle underlying oceanic rises.

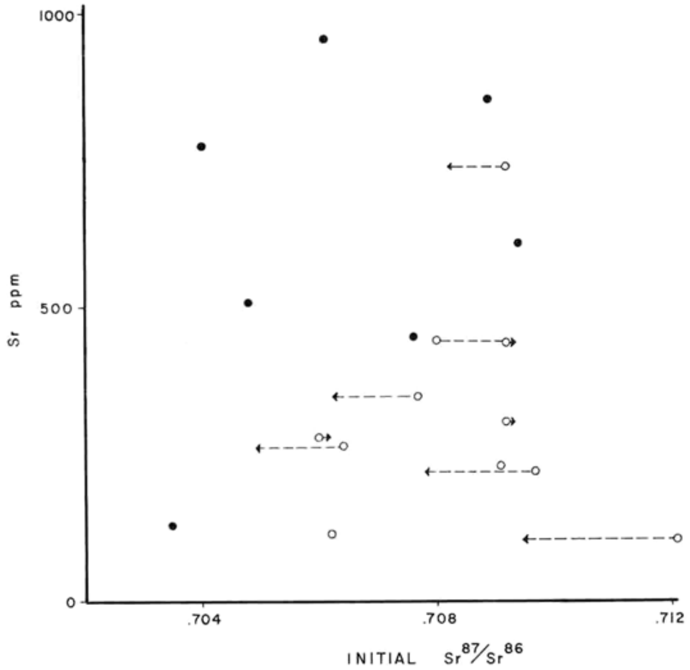

FIG. 10 A - Relationship between initial Sr^{87}/Sr^{86} ratios and Sr ppm of Great Basin volcanic rocks. The circles are silicic rocks and the solid circles are mafic rocks. The dashed lines connect rocks to coexisting plagioclases at arrowheads. The plagioclase initial ratios probably are more indicative of uncontaminated magmatic initial ratios than are the whole rock initial ratios.

In terms of the concept of continental drift and the new global tectonics, it is difficult to accept that the low velocity mantle in both regions is profoundly different in chemical composition and physical character. In regions where continental crust is lacking, the low velocity upper mantle seems capable of yielding only tholeiitic to alkalic basaltic magmas and small amounts of their differentiation products. Thus it does not seem acceptable to propose that the same upper mantle beneath continental regions should yield large volumes of silicic calc-alkalic to alkalic primary magmas. We feel a more satisfactory explanation for the silicic magmas must be an origin largely within the continental crust itself. Only the very young, nodule bearing alkaline basalts with low Sr isotopic ratios that occur

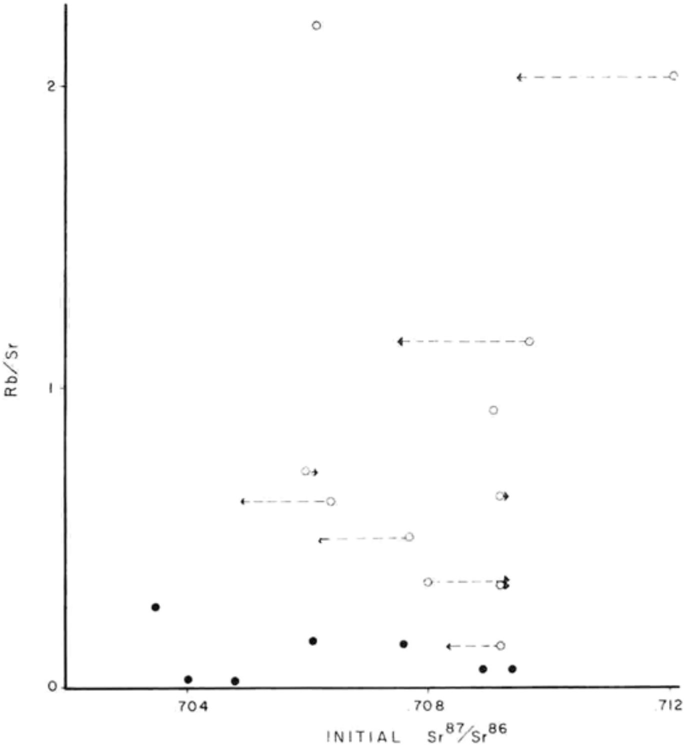

FIG. 10 B - Relationship between initial Sr^{87}/Sr^{86} ratios and Rb/Sr ratios of Great Basin volcanic rocks. Symbols are the same as in figure 10 A. Analyses of magmas derived from different levels in a differentiated crust (lower Rb/Sr ratios with depth) would have a positive relationship between Rb/Sr ratios and initial Sr^{87}/Sr^{86} ratios.

scattered through the Basin and Range Province seem likely to be mantle-derived magmas, with negligible crustal contamination.

Proposed Model for Magma Genesis and Evolution in the Eastern Great Basin

We propose that the volcanic core area of the Great Basin originated at the intersection of an axis of Cenozoic crustal extension and high heat flow with a Mesozoic metamorphic belt that still retained its steep thermal gradients, although plastic deformation at depth had largely ceased by early Cenozoic time (Fig. 4). Thus the thermal condition of the crust inherited from its previous tectonic history exerted a pronounced influence on the Cenozoic tectonic evolution under a grossly different tectonic regime. The initial condition for Basin and Range rifting was definitely not one of an isotropic crustal plate subject to tensile stress; rather it was one of extension of a crustal plate containing a warm, relatively weak zone that was first to respond to the new tectonic regime and first to complete its magma generation cycle. It is even permissible to speculate that the thick warm crust itself nucleated the extension process with consequent upwelling of mantle material independent of the East Pacific Rise. Only later in Miocene and Pliocene time would the two zones of extension have merged more or less in the manner discussed by MACKENZIE and MORGAN (1969) in their paper on triple junctions. In any case, the Basin and Range Province is today closely linked with the East Pacific Rise and San Andreas transform fault system (MENARD, 1961; COOK, K. L., 1968) and is clearly a zone of continuing crustal extension and transcurrent fault movements.

Initial crustal extension would result in partial fractional melting of warm crustal rocks due to decrease in tectonic confining pressure, upward flow of material to lower pressure zones under approximately adiabatic conditions, and to temperature increases caused by convective transfer of heat from deep zones by magmas generated within crust and upper mantle (Fig. 11A). The first Cenozoic normal faults in the core area proceded the oldest volcanism in the core area by only a few million years (KELLOGG, 1964; MOORES *et al.*, 1968). Faulting seems to have progressed toward the Great Basin margins along with the volcanic transgression. An intimate interrelationship between crustal extension and volcanism is thus inescapable.

The silicic volcanic rocks and capping mafic lavas were derived by successive stages of fractional melting of the lower crust with relatively low Rb/Sr ratios (Heir, 1964) and therefore, with Sr^{87}/Sr^{86} ratios lower than average silicic crust, but higher than mantle rocks. Fractional melting appears to have yielded first the volatile rich (although not necessarily water saturated) intermediate to silicic

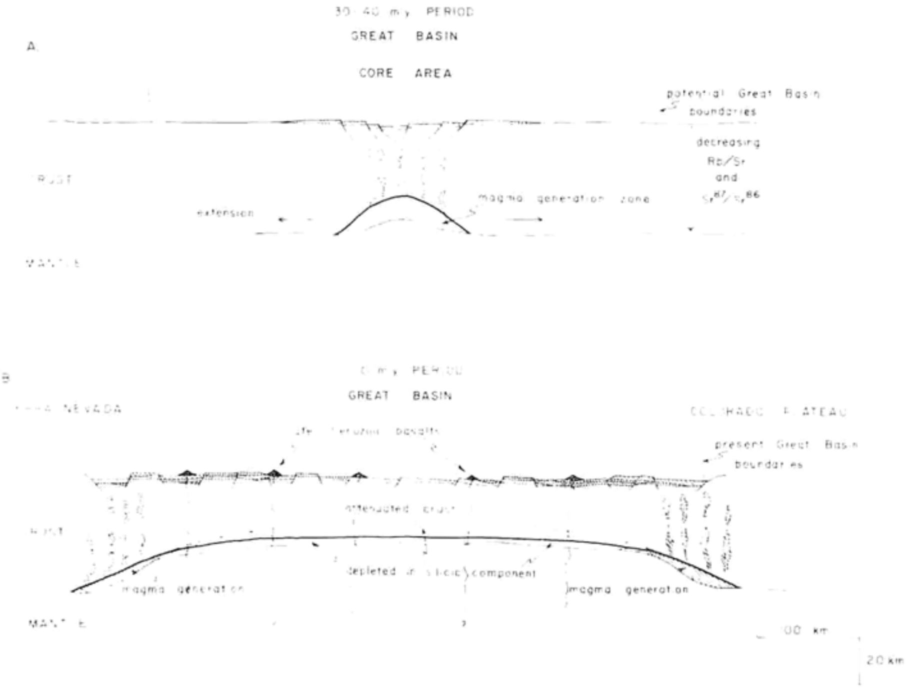

Fig. 11 - East-west section across the Great Basin showing proposed magma genesis and crustal evolution in the Cenozoic Era. Initial volcanism during the 30-40 m.y. period is shown in Fig. 11 A. Late Cenozoic volcanism (10 m.y.) period is shown in Fig. 11 B. The stippled pattern represents probable regions of calc-alkalic magma genesis in the 30-40 m.y. period and the resulting extrusive units in the core area. The left-sloping and vertical line patterns represent extrusive rocks of an intermediate age. The right-sloping line pattern distinguishes regions of genesis of more alkalic magma and resulting extrusive units near the Great Basin margins. Late Cenozoic alkali-olivine basalts are indicated in black. The dashed line pattern covers regions of the crust presumably depleted in low temperature melting fractions. The thickness of extrusive units is greatly exaggerated. The heavy black line is the schematic upper limit of conditions capable of producing melting of the silicic component of the crust in the presence of water. Because of the effect of confining pressure on the liquidus, it is not exactly an isotherm.

magmas that formed the ash flow sheets and soon thereafter a second, higher temperature, drier melt of mafic, aluminous character formed the capping lavas. The refractory residue from the fractional melting process constitutes the lower crust of the volcanic depleted areas.

In the core region, the temperature and pressure conditions capable of producing volcanic magmas included shallower crustal zones than elsewhere (Fig. 11A). Relatively shallow magma genesis produced the calc-alkalic melts with highest Sr^{87}/Sr^{86} ratios. The similar isotopic ratios of the mafic melts derived at high temperatures indicates that they were from approximately the same crustal zones.

Continued extension of the crust and exhaustion of lower temperature melting fractions of the crust in the core region brought volcanism there to an end. This must have been accompanied by a relatively dramatic decrease in the ductility of the core region relative to surrounding crustal areas so that the locus of extension shifted continually to the thicker crustal areas surrounding the previous zone of crustal extension. If this were not so then we would have a Red Sea type of rift rather than the broad zone of distributed rifting of the Basin and Range Province. The location of volcanic centers throughout Cenozoic time (Fig. 8) indicates that the spreading of magma generating conditions was at a rate of slightly more than 1 cm per year. As spreading of volcanic activity toward the margins of the Great Basin occurred, partial melting was restricted to lower portions of the crust because the thermal anomaly inherited from the Mesozoic metamorphism became less pronounced away from the core zone and, as time passed, the metamorphic rocks continued to cool. More alkalic magmas with lower Sr^{87}/Sr^{86} ratios were derived from deeper crustal zones close to present day Great Basin margins (Fig. 11B). At the scattered localities throughout the region where young basalts with low isotopic ratios are found we accept that tapping of deeper magma sources could occur after exhaustion of the crustal zones of magma genesis.

Because silicic and mafic volcanic rocks from the Great Basin, the Basin and Range Province of Arizona and the adjacent Colorado Plateau all show a trend toward lower Sr^{87}/Sr^{86} ratios with decreasing age (EASTWOOD, PUSHKAR, and DAMON, collaborators in DAMON, 1968 and 1969; BIRKERMAN, 1967) similar processes of evolution of magma sources must affect these regions. The Great Basin has the spatial relations that requires the somewhat unique explanation above. But the time-isotopic trends in Arizona are apparently not related to lateral transgressions of volcanic activity. Production of magmas from different crustal levels must be invoked explain these trends. Extraction of the relatively low-temperature melting fraction from higher crustal zones with higher Rb/Sr and Sr^{87}/Sr^{86} ratios gave rise to the older lavas; then, as isotherms rose, partial fusion of progressively less hydrous, deeper zones in the crust and upper mantle with lower Rb-Sr and Sr^{87}/Sr^{86} ratios produced the younger lavas.

Conclusions

Our generalized observations and interpretations concerning magma genesis for Great Basin Cenozoic volcanic rocks can be summarized as follows:

1) The Sr^{87}/Sr^{86} ratios of silicic volcanic rocks decrease from 0.709 for the 32 m.y. old, calc-alkalic rocks of the core area to 0.705 for the 15 m.y. old, more alkalic rocks nearer the margins of the Great Basin.

2) In a like manner, the Sr^{87}/Sr^{86} ratios of mafic volcanic rocks decrease from 0.709 for the 35 m.y. old calc-alkalic rocks of the core area to 0.705 for the 5 m.y. old more alkalic rocks nearer the margins of the Great Basin.

3) The volcanic core area marks the intersection of the relic thermal structure of a Mesozoic metamorphic belt with a zone of crustal extension and high heat flow that originated in the late Eocene-early Oligocene time and is still active. The partial fusion of the deeper crust was due to depression of the melting curve in lower pressure environments and may have been additionally enhanced by regional heat flow increases related to mantle convection that itself may have arisen as a consequence of crustal extension.

4) The source of the abundant silicic and mafic magmas was probably the lower crust with relatively low Rb/Sr and Sr^{87}/Sr^{86} ratios; the silicic partial melt fraction was extracted first and was followed by a mafic fraction as the temperature rose in the same region of the crust.

5) Progressively lower Sr^{87}/Sr^{86} ratios and more alkalic magma chemistry were produced away from the core zone because the upper boundary of conditions that could produce a silicic partial melt in

the crust progressively descended in the crust so that lower Rb/Sr and Sr^{87}/Sr^{86} ratios and higher pressure conditions characterized the zone of partial fusion. Attenuation of the Great Basin crust occurred as the process of magma generation continued.

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