Industrial rocks and minerals
of the Southwest

compiled by George S. Austin

Chronology of the Forums on Geology of Industrial Minerals

Year, place, theme	Sponsor	Source of proceedings
First, 1965, Columbus, Ohio "Industrial limestone and dolomite"	Ohio State University Dept. of Geology	Ohio Academy of Science 445 King Avenue Columbus, Ohio 43201 (Ohio Journal of Science, v. 66, no. 2, p. 97-191)
Second, 1966, Bloomington, Indiana "Cement raw materials"	Indiana Geological Survey and Indiana University	Indiana Geological Survey Dept. of Natural Resources Indiana University 611 North Walnut Grove Bloomington, Indiana 47401 (A Symposium on Geology of Cement Raw Materials) (out of print)
Third, 1967, Lawrence, Kansas "Industrial mineral exploration and development"	State Geological Survey of Kansas and University of Kansas	State Geological Survey of Kansas University of Kansas Lawrence, Kansas 66044 (Spec Distribution Pub. 34, 174 p.)
Fourth, 1968, Austin, Texas "Geology of chemical raw materials" and "Depositional models in economic stratigraphy"	Texas Bureau of Economic Geology and University of Texas (Austin)	Texas Bureau of Economic Geology Box X, University Station Austin, TX 78712 (Proceedings of the Fourth Forum on Geology of Industrial Minerals, 174 p.)
Fifth, 1969, Harrisburg, Pennsylvania "Geology of sand and gravel deposits" and "Can industrial mineral production survive urbanization and conservation?"	Pennsylvania Bureau of Topographic and Geologic Survey	Pennsylvania Bureau of Topographic and Geologic Survey Bureau of Publications 10th and Market Streets Harrisburg, Pennsylvania 17125 (Mineral Resources Rept. M 64, 278 p.)
Sixth, 1970, Ann Arbor, Michigan "Chemical and industrial rocks and minerals" and "Building and construction stones and minerals"	Michigan Geological Survey, University of Michigan, and University of Toledo	Information Services Center Michigan Dept. of Natural Resources Box 30028 Lansing, Michigan 48909 (Michigan Geol. Survey Miscellany 1, 155 p.)
Seventh, 1971, Tampa, Florida "Geology of phosphate, dolomite, limestone, and clay deposits"	Florida Bureau of Geology	Florida Bureau of Geology P.O. Box 631 Tallahassee, Florida 32302 (Spec. Pub. 17, 228 p.)
Eighth, 1972, Iowa City, Iowa "Limestone, gypsum, and shale"	Iowa Geological Survey, University of Iowa, Iowa State University, and University of Northern Iowa	Iowa Geological Survey 16 West Jefferson St. Iowa City, Iowa 52240 (Public Inf. Circ. 5, 194 p.)
Ninth, 1973, Paducah, Kentucky "Geology of fluorspar"	Kentucky Geological Survey and Illinois State Geological Survey	Kentucky Geological Survey 307 Mineral Resources Building Lexington, Kentucky 40506 (Series X, Spec. Pub. 22, 107 p.)
Tenth, 1974, Columbus, Ohio "Reclamation of pits and quarries" and "Carbonate rocks in environmental control"	Ohio State University Dept. of Geology and Ohio Geological Survey	Ohio Geological Survey Dept. of Natural Resources Fountain Square Columbus, Ohio 43224 (Misc. Rept. 1, 100 p.)
Eleventh, 1975, Kalispell, Montana "Industrial minerals of Montana and contiguous states and provinces"	Montana Bureau of Mines and Geology and Flathead Valley Community College	Montana Bureau of Mines and Geology Room 203-B, Main Hall Montana College of Mineral Science and Technology Butte, Montana 59701 (Spec. Pub. 74, 186 p.)
Twelfth, 1976, Atlanta, Georgia "Industrial minerals of Georgia and the southeastern states"	Georgia Geologic Survey and Georgia State University	Georgia Geologic Survey Dept. of Natural Resources 19 Hunter Street, SW Atlanta, Georgia 30334 (Info. Circ. 49, 78 p.)

Circular 182

New Mexico Bureau of Mines & Mineral Resources

A DIVISION OF
NEW MEXICO INSTITUTE OF MINING & TECHNOLOGY

Industrial rocks and minerals of the Southwest

compiled by George S. Austin
in cooperation with 20 other coauthors

A symposium on
industrial rocks and minerals of the Southwest,
held May 12-15, 1981, in Albuquerque, New Mexico

Cover—ADOBE YARD IN NORTHERN NEW MEXICO

SOCCORO 1982

First Printing, 1982

Circular 182 abstract

On May 13–15, 1981, in Albuquerque, New Mexico, the 17th Annual Forum on the Geology of Industrial Minerals was held. This volume contains 18 expanded papers from the Forum, along with 2 abstracts of papers given at the Forum but not presented for publication.

Preface

The southwest part of the United States holds an abundance of industrial rocks and minerals. The purpose of the 17th Annual Forum on the Geology of Industrial Minerals was to acquaint participants with industrial minerals of the Southwest. We not only heard about a number of occurrences, but also visited the areas on several field trips. Field trips included visits to the potash mines of southeast New Mexico, trips north from Albuquerque and along the Rio Grande to a gypsum mine, an adobe factory, a pegmatite gypsum operation, a pumice mine, and an Indian pueblo. Another trip to the Grants mineral belt was principally to see the perlite deposit near that city. Spouses' trips included trips to the Old Town area of Albuquerque to see an example of restored architecture and shops from 60 yrs ago.

The conference was held from May 13 through May 15, 1981, in Albuquerque and was attended by approximately 140 geologists and engineers. Most were from the United States, but we had representatives from Canada and other foreign countries as well. Circular 182 represents the final link in the entire project and is the end result of a tremendous amount of work by all the participants, chairpersons, and workers on the many committees, the authors of the papers, the editors and secretaries, and even the general chairman. Circular 182, "Industrial rocks and minerals of the Southwest," contains 18 expanded papers given at the convention plus 2 abstracts without papers. These works contain much of the latest informa-

tion on the area and include several significant papers dealing with the industrial minerals that are so important to this part of the country.

Although the staff of the New Mexico Bureau of Mines and Mineral Resources and I have done some editing, the responsibility of each paper belongs to the author. We felt that, except for obvious problems, we should let most of the papers go through with little or no change. Therefore, the approaches to the problems may vary, but the quality, we believe, remains high.

I would like to thank Frank Kottlowski, Director of the Bureau, the Executive Committee of the Forum on the Geology of Industrial Minerals, the authors, and my colleagues here at the New Mexico Bureau of Mines and Mineral Resources for their help in putting together a good conference. Those that deserve some special recognition for having worked so hard on the project include Mark Logsdon, Registration Chairman; Bob Weber, Field Trip Chairman; Steve Blodgett, Assistant Editor; Jeanette Chavez, Secretary; and Marj Austin, Women's Program Chairman.

George S. Austin
Deputy Director, New Mexico Bureau of
Socorro   Mines and Mineral Resources
January 25, 1982   General Chairman of symposium

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INDUSTRIAL MINERALS OF NEW MEXICO IN 1981

by G. S. Austin and F. E. Kottlowski, *New Mexico Bureau of Mines and Mineral Resources, Socorro, New Mexico 87801*
and
W. T. Siemers, *Phillips Petroleum, Bartlesville, Oklahoma 74003*

Abstract

New Mexico has more than 225 operations for mining, milling, preparing, and smelting. Of that number, approximately 160 are related to the production of potash, salt, perlite, mica, pumice, peat, clay, barite, fluorite, gypsum, stone, sand and gravel, lime and cement, and the preparation of abrasives, fillers, and insulating materials. In 1980 New Mexico ranked first in the United States in production of potash and perlite with values of $314,423,000 and $14,721,000, respectively (U.S. Bureau of Mines, 1981). Nonmetallic resources are distributed throughout the state, but production is concentrated in the western and southeast parts of New Mexico. The value of industrial mineral products in the state has increased substantially through the last 10 yrs, but the quantities of these materials produced has remained more or less constant. The majority of industrial rocks and minerals products is shipped out of New Mexico. However, as the population of the state continues to increase, this situation will probably change. For economic reasons, industry needs to decrease transportation costs and thus is tending to utilize more materials closer to the source. This should make New Mexico less dependent on out-of-state sources of finished products. Availability of water may be the major factor limiting population and industry.

Introduction

Bates (Lefond, 1975) describes industrial rocks and minerals as ". . . any rock, mineral or other naturally occurring substance of economic value, exclusive of metallic minerals, mineral fuels, and gem stones"; such materials are thus the same as nonmetallics. In 1980 New Mexico had more than 225 operations for mining, milling, preparing, and smelting. Of those, approximately 160 were related to the production of in-

dustrial minerals and rocks. New Mexico is the nation's leading producer of potash and perlite and is a major producer of domestic pumice. Other important commodities produced in the state include aggregate (caliche, sand and gravel, and scoria), barite, clay, dimension stone, fluorspar, gypsum, humate, limestone, mica, salt, and silica sand (Siemers and Austin, 1979a).

Deposits of these materials occur widely throughout New Mexico; nonetheless, production tends to be concentrated in the western and southeast parts of the state (fig. 1). The total value of industrial rocks and minerals produced in New Mexico during 1980 was about $359 million (U.S. Bureau of Mines, 1981; fig. 2). This value reflects about a 30% increase over the 1979 production value and a 207% growth in the value of production since 1960.

Potash

Potash is the common term for compounds of the element potassium; sylvite (KCl) and langbenite ($K_2SO_4 \cdot 2MgSO_4$) are the most important domestic potassium minerals. New Mexico's potash resources are concentrated in the Permian Basin, near Carlsbad, in southeast New Mexico (fig. 3). The Permian Basin is subdivided into a number of smaller basins, platforms, and shelves, with the Carlsbad potash mining district lying astride the boundary between the Delaware Basin and the Northeastern Shelf. Mineralization in minable quantities is concentrated in the McNutt Potash Member of the Salado Formation of Ochoan (Late Permian) age. The Salado is a sequence of marine evaporites containing thick beds of halite,
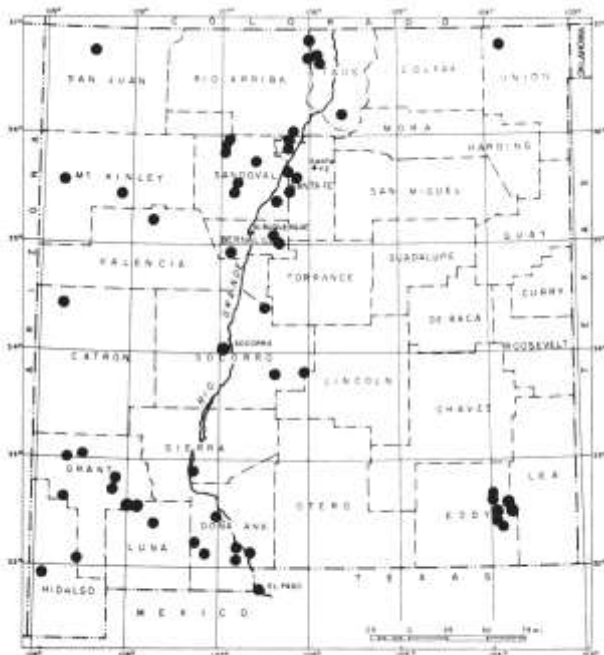
FIGURE 1—MAJOR INDUSTRIAL ROCK AND MINERAL OPERATIONS (EXCLUDING SAND AND GRAVEL) IN NEW MEXICO.
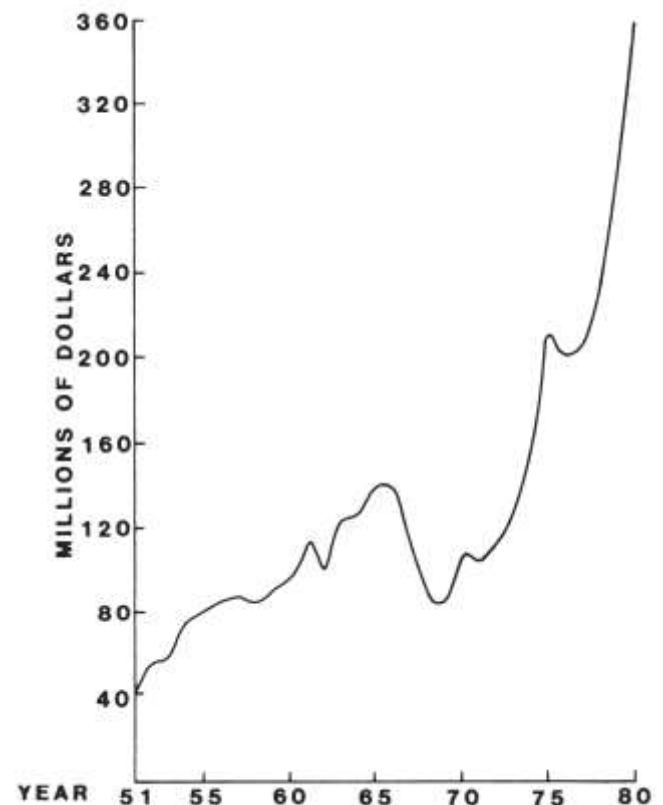
FIGURE 2—TOTAL VALUE OF INDUSTRIAL ROCK AND MINERAL PRODUCTION IN NEW MEXICO.

FIGURE 3—PERMIAN BASIN OF NEW MEXICO AND WEST TEXAS SHOWING SUBDIVISIONS AND THE CARLSBAD MINING DISTRICT.

gypsum, and anhydrite. The McNutt is about 122 m (400 ft) thick in the mining district and contains 11 ore zones of variable thickness and lateral continuity. Proven concentrations of potash occur in the 1st, 3rd, 4th, 5th, 7th, 8th, and 10th ore zones (Siemers and Austin, 1979a).

Potash was first produced in New Mexico in 1931 and total production since 1950 has amounted to 57 million metric tons (63 million tons), valued at nearly $3.0 billion. In 1980, seven mines operated by seven different companies in the Carlsbad district had a maximum daily capacity of about 48,400 metric tons (53,000 tons) of ore per day. The largest of these was the International Minerals and Chemicals Corporation mine with a daily capacity of 15,420 metric tons (17,000 tons).

Total sylvite reserves within the Carlsbad mining district held under lease, together with unleased reserves, are about l.2 billion metric tons (1.3 billion tons), having an average grade of 17% K_2O. Crude langbeinite reserves are estimated to be 180 million metric tons (198 million tons) with an average K_2O grade of approximately 9% (Alto and Fulton, 1965). At the present rate of consumption a 20-yr supply of potash remains within the enclave; however, less than one-half of the reserves have been adequately drilled and mapped. If one considers potash on unleased lands, potash bypassed because of its low grade, and the possibility of better extraction methods in the future, the sylvite and langbeinite in the Carlsbad district may last nearly 100 yrs (Austin, 1976).

Aggregate

Aggregates are hard, essentially inert materials that can be used to form a stable mass with cementing or compaction. Many igneous and sedimentary rocks are used for aggregate in New Mexico, but the largest volume comes from perlite, pumice, scoria, sand and gravel, caliche, and limestone. Aggregate resources along the Rio Grande area are very large, but on the eastern plains of New Mexico, producers must resort to using caliche, which is of lower quality than other sources of aggregate.

Perlite

Perlite originally referred only to natural, siliceous, volcanic glasses characterized by concentric fractures and a pearly luster. Industrial use of the term now includes any volcanic glass that will expand appreciably by vesiculation with a proper heat treatment.

New Mexico deposits of any importance are located west of the Rio Grande, occurring discontinuously with other volcanic rocks from the Colorado border to Mexico (fig. 4). Commercial interest in many of these deposits is currently low because most are remote, small in size, of poor quality, or involve excessive mining costs (Weber, 1965). Perlite deposits in New Mexico are associated with volcanic fields, particularly those of rhyolitic composition, and are typically Tertiary or younger in age. Some larger flow deposits are several hundred feet thick and may extend several miles laterally.

New Mexico leads the nation in the production of perlite, with approximately 85% of domestic production in 1980. Since 1953, the state has produced about 8.6 million metric tons (9.5 million tons) of perlite, valued at $104.1 million. Production in 1980 was 494,000 metric tons (544,000 tons) valued at about $14.7 million. The combined daily capacity of the state's five active operations is about 3,765 metric tons (4,150 tons) per day. More than 45% of this capacity (1,680 metric tons or 1,850 tons) is from Johns-Manville Corporation's No Agua J-M mine north of Tres Piedras, Taos County (Siemers and Austin, 1979b). Perlite resources are large in New Mexico, with developed commercial-grade reserves being sufficient to meet any forseeable demand.

Pumice

Pumice is a highly vesicular, light-colored, typically rhyolitic volcanic glass. The term is also applied to dark-colored, cellular, andesitic and basaltic rocks by industry, but here these are called scoria.

Pumice deposits occur widely throughout western New Mexico; most of the state's developed resources, however, are located in the southern and eastern Jemez Mountains in Sandoval, Los Alamos, and Santa Fe Counties (fig. 5). The deposits are associated with ash-flow and ash-fall tuffs, tuff breccias,
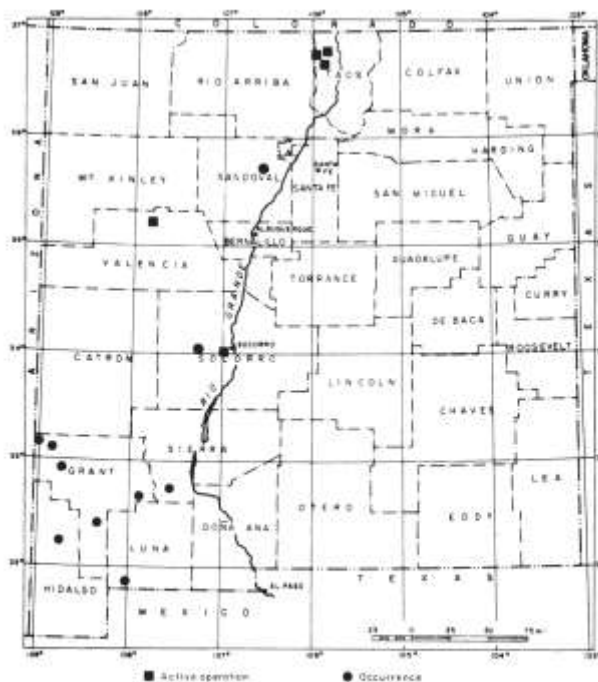
FIGURE 4—PERLITE DEPOSITS AND ACTIVE PERLITE OPERATIONS IN NEW MEXICO.
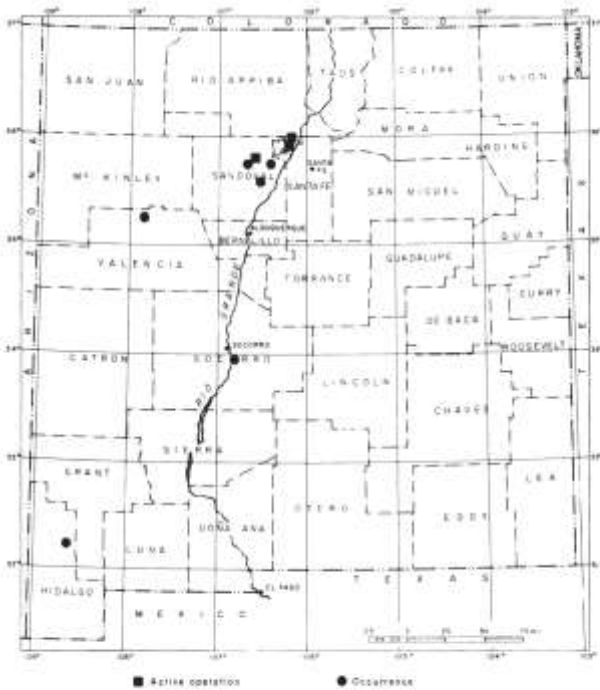
FIGURE 5—PUMICE DEPOSITS AND ACTIVE OPERATIONS IN NEW MEXICO.
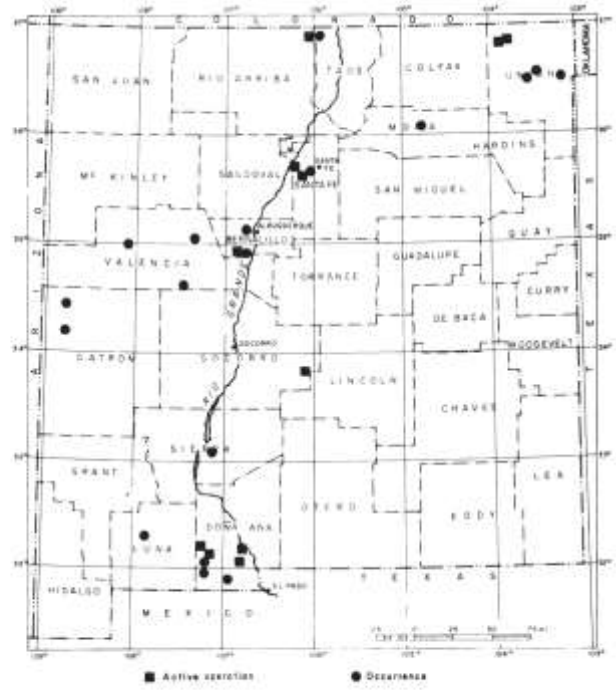
FIGURE 6—SCORIA DEPOSITS AND ACTIVE SCORIA OPERATIONS IN NEW MEXICO.

and volcaniclastic sediments of Tertiary and Quaternary age (Weber, 1965).

Pumice production in New Mexico since 1950 amounts to about 10.1 million metric tons (11.5 million tons), valued at about $26.6 million. Three pumice operations in the state produced 492,000 metric tons (542,000 tons) in 1980. The largest operation is the Copar mine in Santa Fe County, with 272 metric tons (300 tons) daily capacity (Siemers and Austin, 1979b). No estimates of pumice reserves in New Mexico are available, but reserves appear nearly inexhaustible at present consumption rates.

Scoria

Scoria is a dark-colored (commonly reddish-brown to dark-gray), cellular, usually andesitic to basaltic rock. Scoria differs from pumice, however, in its composition, darker color, higher density, coarser vesicles, more crystalline texture, and generally higher strength. Scoria deposits result from explosive eruption of basaltic lavas, with vesiculation resulting from expansion or interfragmental water vapor during eruptions. The pyroclastic fragments are usually deposited peripherally around a vent, causing the development of a cinder cone. Such deposits are widely spread in the Tertiary and Quaternary systems of western and northern New Mexico (Osburn, 1979).

Eleven operations were active in the state in 1980 (fig. 6) with a combined daily capacity of about 4,200 m³ or 5,500 yd'. The largest (Twin Mountain mine, Union County) has a maximum capacity of 1,000 metric tons or 1,000 tons per day (Siemers and Austin, 1979b). New Mexico scoria resources are extremely large and should be large enough to meet any for-seeable demand.

Sand and gravel

Sand and gravel are so widespread in New Mexico that much of a map showing geologic units containing potential deposits is covered (fig. 7). Principal deposits, nevertheless, occur as alluvial, Pleistocene to Holocene sands and gravels along the Rio Grande valley. Similar deposits are found in the Pecos and Canadian River valleys, and their tributaries. Some

sand and gravel also occurs in older (Tertiary and Cretaceous) formations in the westernmost and easternmost parts of the state.

New Mexico has produced more than 226 million metric tons (249 million tons) of sand and gravel, valued at more than $290 million. About 4.5 million metric tons (4.9 million tons) were produced in 1980 from more than 180 operations that ranged from large, permanent businesses to small, mobile, temporary operations established to meet a short-

Quaternary deposits
Early Tertiary deposits

Middle and Late Tertiary deposits
Permian Glorieta Formation
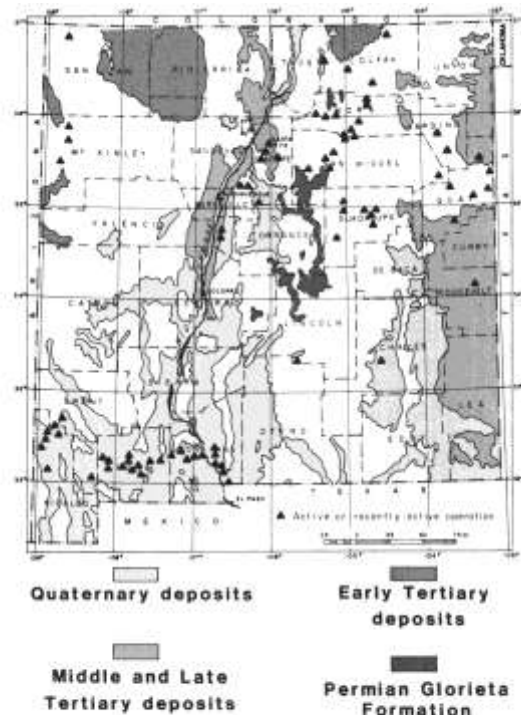
FIGURE 7—POTENTIAL SAND AND GRAVEL DEPOSITS AND SOME ACTIVE AND RECENTLY ACTIVE SAND AND GRAVEL OPERATIONS IN NEW MEXICO.

term local demand. With nearly half of the state being covered by materials that could be used for such purposes, the state's sand and gravel resources are apparently substantial.

Caliche

Caliche is a general term for a zone of secondary carbonate accumulation in surficial materials. In southeast New Mexico, where caliche horizons are well developed and widespread, such deposits are especially thick (as thick as 10-15 m) as the caprock of the Llano Estacado—the High Plains east of the Pecos River (Kottlowski, 1965). Most of the caliche in New Mexico (fig. 8) has formed by pedologic processes involving the leaching of carbonates from sandy surface sediments by downward-percolating soil water and subsequent precipitation of the carbonate as cementing material deep in the soil profile. From the early 1960's to 1978, caliche production in New Mexico is estimated to have been about 136,000 metric tons (150,000 tons) per year. In 1976, caliche produced in New Mexico was valued at about $154,000. Large caliche reserves occur throughout the eastern part of the state, but principally in the southeast.

Gypsum

Gypsum, or hydrous calcium sulfate ($CaSO_4 \cdot 2H_2O$), has the unique property of readily giving up or taking on water of crystallization; most uses for the mineral are related to this property. Many of New Mexico's gypsum deposits occur in the central and southeast parts of the state (fig. 9). The deposits are largely chemical precipitates resulting from the evaporation of Phanerozoic sea water. Such deposits occur in the Pennsylvanian (Panther Seep Formation), Permian (Abo-Hueco Formation, Epitaph Dolomite, Yeso Formation, San Andres Formation, Artesia Group, Castile Formation, and Rustler Formation), Jurassic (Todilto Formation), and lower Cretaceous strata, Tertiary redbeds, and as Quaternary lake deposits (Weber and Kottlowski, 1959). Recent gypsum dune sands are abundant in the Tularosa, Estancia, Encino, and Pinos Wells Basins.
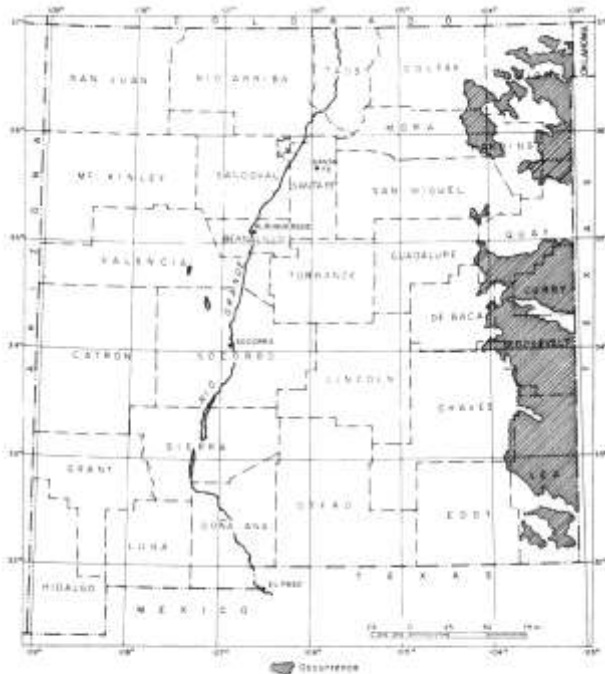
FIGURE 9—GYPSUM OCCURRENCES AND ACTIVE GYPSUM OPERATIONS IN NEW MEXICO.

There was no notable production of gypsum in New Mexico before 1959. Since that time total production has amounted to about 2.9 million metric tons (3.2 million tons) valued at more than $19.2 million. Nearly all of the production has come from the Todilto Formation in central New Mexico. Three gypsum mines are active in the state; all of these mines occur in the Todilto Formation. The largest, is the White Mesa mine in Sandoval County which boasts an annual capacity of about 165,000 metric tons (180,000 tons), all of which is utilized by the American Gypsum Company in Albuquerque (Siemers and Austin, 1979b).

New Mexico's gypsum reserves are substantial; surface reserves alone could supply the total United States demands for hundreds of years. The White Mesa deposit, for example, contains measured reserves of 89 million metric tons (98 million tons) and an estimated additional 112 million metric tons (123 million tons) available by stripping a relatively thin overburden.

Clay and shale

Clay material is defined here as any naturally occurring, fine-grained, earthy, clayey material and includes clays, shales, and soils. In the Southwest, impure clay minerals, commonly high in smectitic clay minerals, are widely used to make adobes (sun-dried bricks), which are still commonly used in New Mexico (Smith, E. W., 1981). Clay and shale deposits occur widely throughout New Mexico (fig. 10). The clay minerals that occur in the state include kaolinite, montmorillonite, illite, halloysite, sepiolite, chlorite, and mixed-layer clay minerals.

Refractory (kaolin-rich) clays occur under many types of geologic conditions in the state in hydrothermally altered tuffs. Halloysite frequently occurs in small veins and pockets, the largest of these pockets being a channel-fill deposit at the base of the Dakota Sandstone (Cretaceous) northeast of Amistad in the northeast part of the state (Patterson, 1965). Bentonite deposits are widespread in New Mexico, but have been mined on only a small scale at a few localities. They occur in sedimentary rocks ranging from Permian to Quaternary age, with the best quality Na-rich bentonites being Tertiary lake deposits.

FIGURE 8—MAJOR CALICHE DEPOSITS IN NEW MEXICO (modified from Kottlowski, 1965).
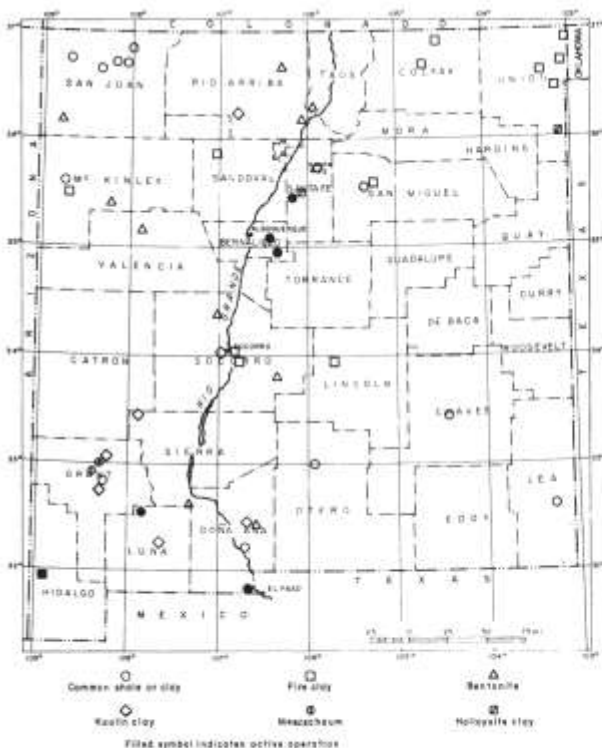
FIGURE 10—CLAY AND SHALE DEPOSITS AND ACTIVE CLAY AND SHALE OPERATIONS IN NEW MEXICO.

FIGURE 11—BARITE OCCURRENCES AND ACTIVE OPERATIONS IN NEW MEXICO.

Total production of clay and shale in New Mexico since 1950 amounts to more than 1.48 million metric tons (1.63 million tons), valued at more than $3.24 million; production amounted to about 36,000 metric tons (40,000 tons) in 1980, Currently, there are five clay producers in the state. The largest is the Kinney Brick Company with a plant in Albuquerque and an annual capacity of about 16,330 metric tons (18,000 tons; Siemers and Austin, 1979b). El Paso Brick Company operates another large clay pit just west of the Rio Grande near the Mexican border.

Clays and shales used for structural clay products occur in nearly inexhaustible quantities. This is also true of the clay resources for making adobe bricks. The prospects for further production of bentonite also appear good. The future production of kaolin in the state is a definite possibility, but inadequate information is available to presently evaluate kaolin resources or uses in New Mexico. Fire-clay resources, whose production amounted to about 10,430 metric tons (11,500 tons) in 1976 (Siemers and Austin, 1979a), appear adequate to meet only a limited local demand; coal stripping in the Gallup area, however, may uncover kaolinitic clays suitable for use as low- and moderate-duty fire clay.

Barite

Barite (BaSO$_4$) is a primary source of barium and is extremely important in the petroleum industry, which consumes approximately 80% of the world's production of barite for drilling mud. Known occurrences of barite in New Mexico are restricted to the state's central mountainous region (fig. 11). Within this region are 60 known large deposits of barite and 24 miscellaneous occurrences, many on the White Sands Missile Range (Smith, T. J., 1981).

The deposits occur in veins or veinlets and as wall rock replacement in igneous, sedimentary, and metamorphic units. Total known production of crude barite from the state amounted to about 35,510 metric tons (39,145 tons) in the period 1951-1978, with 90-95% of the production coming
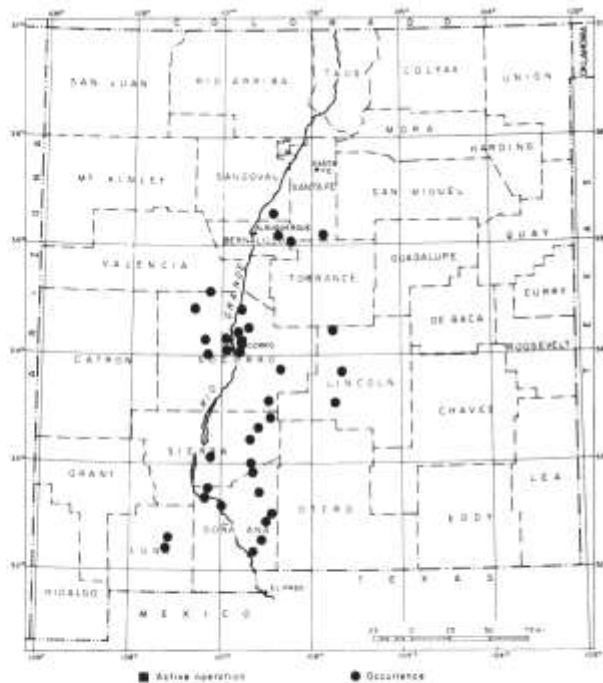
from one mining property, the Mex-Tex Group, in Socorro County. Nearly all of this production (96%) occurred in the period 1951-1965.

The one recently active barite operation in New Mexico began operation in June 1978. Ranger Industries, Inc., produced about 180 metric tons (200 tons) per day in 1979 from the Elaine Group and other barite mines in Socorro County. The barite was shipped to Midland, Texas, where it was used as a weighting agent for drilling muds.

Development of New Mexico's barite resources depends on demand and production costs. The association of New Mexico barite with fluorite has also caused slow development of the state's barite industry; nevertheless, improved methods of separating barite-fluorite complexes is raising many submarginal deposits to the status of ore reserves. If industrial demand shifts to New Mexico or if larger deposits that can be more cheaply mined are located in the state, New Mexico barite may become competitive in the Southwest barite market.

Fluorspar

Fluorspar is a mineral aggregate containing enough fluorite (CaF$_2$) to be of economical interest. New Mexico's fluorspar deposits are localized in the southwest and central areas of the state (fig. 12). Most deposits occur in Tertiary intrusives or volcanics, Precambrian granites, or Paleozoic sedimentary units that are spatially associated with Tertiary or Cretaceous extrusive or intrusive rocks. McAnulty (1978) indicated that the most common and most important deposits are fissure veins, stratiform replacement bodies (mantos) in limestones, and replacements in limestones in contact zones adjacent to silicic-alkalic intrusive igneous rocks. The deposits are typically small (18,000-30,000 metric tons or 20,000-33,000 tons) and range in grade from approximately 35 to 85% CaF2; ores commonly contain about 60% CaF2 and 20% CaCO$_3$.

Fluorspar mining began in New Mexico in the early 1880's in the southeast part of the state, reached peak production of 900,000 metric tons (990,000 tons) in 1944, and essentially ceased by 1955, because of low prices and foreign competition. Total production to 1978 is about 608,600 metric tons
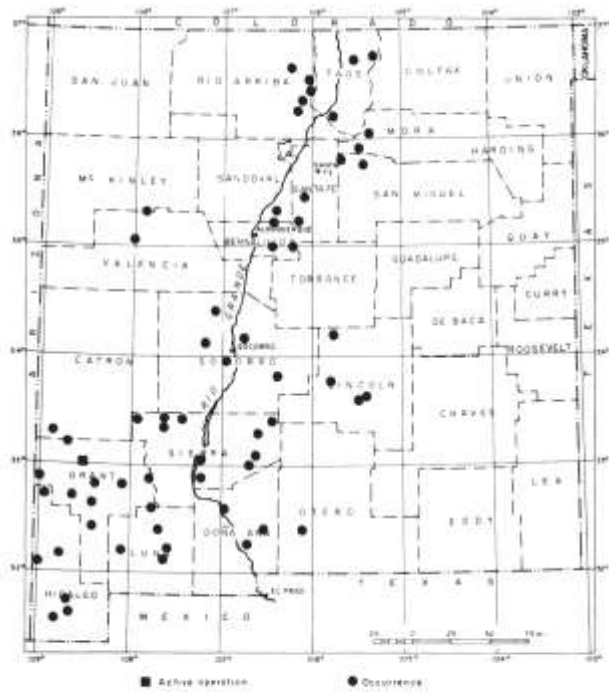
FIGURE 12—FLUORSPAR DEPOSITS AND ACTIVE FLUORSPAR OPERATIONS IN NEW MEXICO.

FIGURE 13—SALT OCCURRENCES AND ACTIVE SALT OPERATIONS IN NEW MEXICO.

(671,000 tons); 87% was from seven localities including the Zuni, Caballo, and Burro Mountains, Cooke's Peak, Fluorite Ridge, the Gila district, and the Tortuga deposit. Current production is restricted to small sporadic shipments from the Great Eagle mine near Red Rock in southwest New Mexico.

Fluorspar reserves in New Mexico are estimated to be about 900,000 metric tons (990,000 million tons) of 50% CaF_2 ore and about 3.8 million metric tons (4.2 million tons) of 27% $CaF2$ ore (Quan, 1978). Most fluorspar occurs in the seven regions that have contributed the bulk of the state's past production.

Salt

Salt is defined here as the mineral halite or sodium chloride (NaCI). New Mexico's salt resources occur in many sedimentary units of Permian age in southeast New Mexico and in the Paradox Formation of the Hermosa Group (Pennsylvanian) in extreme northwest New Mexico (fig. 13). Salt also occurs in playas and salt lakes, and considerable production is obtained from the refinery waste of the potash industry.

New Mexico produced more than 1.9 million metric tons (2.1 million tons) of salt, valued at more than $14.5 million, from 1950 to 1978. In 1979, three operations produced 108,840 metric tons (119,975 tons) of salt, valued at about $1-.8 million. The largest of these, the Salt Lake mine in Eddy County, southeast New Mexico, has a daily capacity of about 545 metric tons (600 tons). A solar operation (now defunct) at Zuni Salt Lake in west-central New Mexico produced about 180 metric tons (200 tons) of salt in 1980 (Siemers and Austin, 1979b). The state's salt supplies, including brine lakes, potash tailing piles, and subsurface rock salt, are substantial and large enough to meet any forseeable demand.

Limestone and dolostone

Limestone and dolostone are sedimentary rocks whose major constituents are calcium carbonate (calcite) and calcium-magnesium carbonate (dolomite). Limestone occurs throughout central New Mexico from Colorado to the Mexican border (fig. 14). Most occurrences are shallow marine, lake,
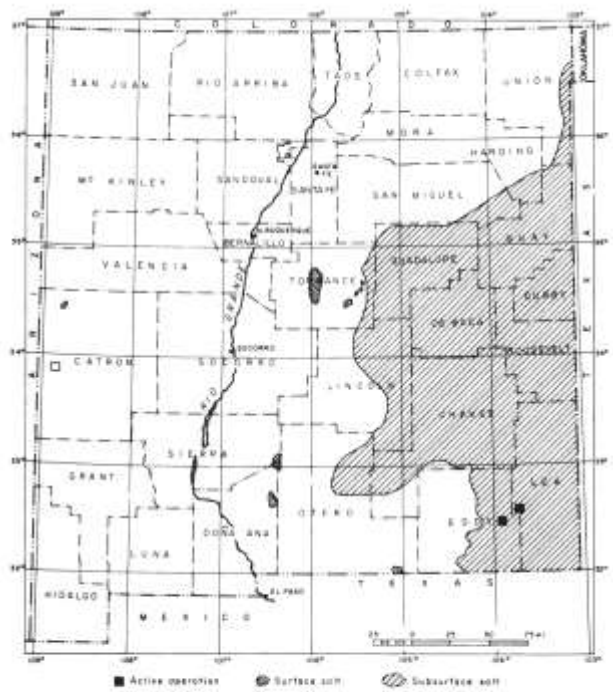
or spring deposits that range from early Paleozoic to Quaternary age. Most high-calcium limestones occur in late Paleozoic (Mississippian through Permian) formations.

Abundant limestone reserves occur in widespread Mississippian and Pennsylvanian units, as thick as 300 m. What proportion of these reserves are high-calcium, however, remains to be determined, but reserves are believed to be substantial. Dolostones are restricted to early Paleozoic (Ordovician and Silurian) units that occur in southern New Mexico.

Nearly all of the high-calcium limestone produced in New Mexico is consumed in the manufacture of lime and cement.

FIGURE 14—LIMESTONE OCCURRENCES AND ACTIVE LIMESTONE OPERATIONS IN NEW MEXICO.

The earliest record of lime production in New Mexico was in 1956. From 1956 to 1978, the state has produced more than 526,000 metric tons (580,000 tons) of lime, valued at more than $5.5 million. The only two active lime operations in the state are the Hurley lime quarry (Kennecott Copper Corporation) and the Mathis lime quarry (Mathis and Mathis Mining and Exploration) near Silver City. Their combined daily capacity is about 480 metric tons (530 tons).

Ideal Basic Industries' Tijeras plant is the only cement plant in New Mexico. The Tijeras plant presently has a designed capacity of about 475,000 metric tons (525,000 tons) per year. The company announced plans to increase its annual production by about 20% to 75,000-90,000 metric tons (85,000-100,000 tons) beginning in 1980.

Mica

Most of New Mexico's mica is associated with Precambrian schists that occur in the north-central part of the state (fig. 15). Sheet-quality mica also occurs in this area as large crystals scattered throughout unzoned pegmatites or in certain units of zoned pegmatites.

Production during the 1970's has ranged between 9,000 and 14,000 metric tons (10,000 and 15,000 tons) per year. Currently, the Tojo mine (Minerals Industrial Commodities of America, Inc., or MICA), with a 12,000 metric-ton (13,000-ton) annual capacity, is the ony active mica operation in the state (Siemers and Austin, 1979b). Resources of sheet mica in New Mexico are uncertain because of incomplete available data, but resources of scrap mica are believed to be large.

Silica sand

Potential high-silica sandstones in New Mexico are poorly studied and records of past production in the state are not available. The Brockman silica quarry, near Playas in southwest New Mexico and operated by Phelps Dodge Corporation, is the largest silica operation in New Mexico with an annual capacity of about 63,500 metric tons (70,000 tons; Siemers and Austin, 1979b). The quarry occurs in the Mojado Formation (Cretaceous).

Humate

As used here, humate is defined as a brownish, carbonaceous mudstone. Crushed humate is marketed as a soil conditioner, although the effects of its use for such purposes have not been completely determined or documented. Claimed effects (Siemers and Wadell, 1977) of humate application to a soil include: 1) increased retention of water and soluble fertilizers, 2) increased acidity, 3) improved physical character, and 4) beneficial color changes.

New Mexico humates occur as dark-brown to brownish-gray, carbonaceous, blocky to slightly fissile mudstones in the Menefee Formation (Cretaceous) in the northwest part of the state (fig. 16). The deposits reflect deposition in moderately to poorly drained swamps that received abundant clay material and organic matter.

No record of humate production in New Mexico exists prior to 1976; in that year, production amounted to about 24,000 metric tons (26,500 tons), valued at $236,000. The Clodbuster mine operated by Farm Guard Products and located between Cuba and La Ventana in northwest New Mexico, has a daily capacity of 70-90 metric tons (80-100 tons). Production, however, is typically sporadic, depending largely on intermittent sales. The state's humate resources are plentiful, with Cretaceous strata in northwest New Mexico bearing millions, if not billions, of metric tons of humate.

Dimension stone

The Placer mine, a flagstone operation in northern Luna County (southwest New Mexico), and an ornamental calcite operation in the Cedar Hills of northern Dona Ana County (south-central New Mexico) known as the Rainbow Marble mine, are the only active dimension-stone operations in the state.

Outlook

New Mexico is rich in many industrial rocks and minerals and contains a superabundance of others. One of New Mex-

FIGURE 15—MICA DEPOSITS AND ACTIVE MICA OPERATIONS IN NEW MEXICO.
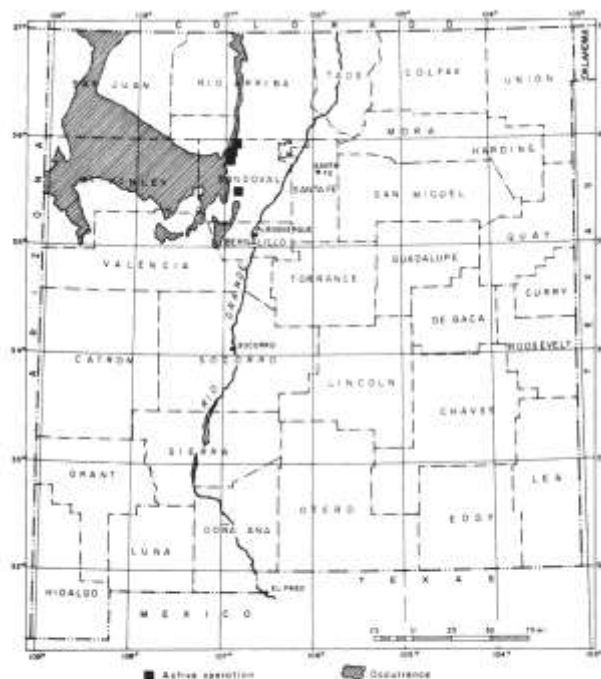
FIGURE 16—ACTIVE HUMATE MINES IN NORTHWEST NEW MEXICO AND AREAS THAT ARE LIKELY TO CONTAIN COMMERCIAL HUMATE DEPOSITS (modified from Siemers and Wadell, 1977).

ico's major problems is the scarcity of sizable, local markets; moreover, the availability of water in the state's semiarid climate will likely limit substantial increases in population and consequent industrial growth. Prospects for the development of sizable, local markets for industrial rocks and minerals are poor. Use of such resources in New Mexico is expected to continue to be limited to relatively small, scattered population centers. Large markets, with the exception of the Albuquerque and Las Cruces-El Paso areas, will have to be sought elsewhere. Major growth in New Mexico's industrial rock and mineral industry will center on local production and processing of quality materials for which a national market or markets in Albuquerque, Las Cruces, and El Paso either exists or can be created.

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GEOLOGIC OVERVIEW OF THE CARLSBAD POTASH-MINING DISTRICT

by George B. *Griswold, Tecolote Corporation, Albuquerque, New Mexico 87123*

Abstract

The Carlsbad potash district lies in southeast New Mexico, between the cities of Carlsbad and Hobbs. Mining commenced in 1931 soon after the initial discovery of commercial ore. The seven producers in the district last year produced an aggregate of 19.73 million tons of ore that after refining yielded muriate containing the equivalent of 2.77 million tons K_2O. The district accounts for 80% of domestically produced potash fertilizers. The potash deposits are contained within the McNutt Member of the Salado Formation. The Salado is part of the thick Ochoan evaporite sequence that fills much of the Permian Basin. The regional setting of these evaporites can be described in fairly simple terms; however, commercial deposits of potash are quite complex in detail. Mining methods utilize both the room and pillar and longwall systems. Ore is enriched to commercial fertilizer grades by a variety of processes. Flotation is the common method for treating simple ores, but heavy-media, base-exchange, fresh-water-leach, and recrystallization processes are also used.

Location

The Carlsbad potash district lies in southeast New Mexico (fig. 1), and mines are located between the communities of Carlsbad and Hobbs. Potash is known throughout a much larger area, extending as far to the southeast as Midland and northeast to Lubbock. However, the beds are too deep for conventional mining or contain minerals not amenable to concentration into commercial products. The western edge of the district is a well-defined salt-dissolution front where ground waters have removed the commercial ore beds.

History of discovery and development

The denial of access to German sources of potash fertilizers during World War I was sufficient incentive to cause a nation-wide search for a domestic source in known bedded-salt deposits such as in the Michigan, Paradox, Williston, and Permian Basins. Early investigations by the U.S. Geological Survey favored the Permian Basin because the mineral polyhalite had been recognized over wide areas of southeast New Mexico and west Texas from cuttings recovered from holes

drilled for oil (Hoots, 1925). The desire to find domestic sources was demonstrated by the enactment of Public Law 424 in 1926 which authorized a 5-yr exploration program, including core drilling, by the U.S. Geological Survey and U.S. Bureau of Mines. The government drilled a total of 24 widely spaced holes between 1927 and 1931; 13 were in New Mexico, 10 in Texas, and 1 in Utah. Simultaneously, private industry joined the search, with geologists carefully examining cuttings from holes drilled for oil but penetrating salt on the way down. Fortunately, most of the drilling still was done with cable tools during the late 1920s which insured that a saturated brine would build up in the hole and thus retard dissolving soluble potash minerals.

Although the U.S. Geological Survey initiated the potash-exploration program, serendipity played its usual role in the discovery. Drill cuttings from the Snowden-McSweeney No. 1 McNutt wildcat provided the first evidence of commercial amounts of sylvite. The well was spudded on July 5, 1925, and reached its eventual total depth of 4,416 ft on March 20, 1926. V. H. McNutt is credited with the identification of several significant intercepts of sylvite by examination of the cuttings. That serendipity played its role rests on these eventual revelations: 1) the hole was drilled for oil and none was found, 2) the hole location was almost at the exact "center of gravity" of the numerous orebodies now known in the district, 3) potassium mineralization was present in all 11 of the now accepted ore horizons, but 4) no potash mining has been done in the immediate vicinity of the hole because of the low grade in comparison with surrounding areas.

The American Potash Company was formed initially as a subsidiary of the Snowden-McSweeney Oil Company. A core-drilling program immediately proved up reserves of sufficient size and grade to justify mine development a short distance west of the initial discovery. Holdings of American Potash were transferred and reorganized into the U.S. Potash Company, which in turn became a subsidiary of U.S. Borax. Mining commenced in 1931. The Potash Company of America opened a second mine in 1934. The capacities of these mines were steadily increased and still other companies opened new mines until there are now seven companies working in the district (fig. 2).

Although the U.S. Geological Survey did not participate in the actual discovery of the Carlsbad district, their role should not be underestimated. The USGS laid the initial groundwork and alerted the oil industry to the possibility of potash over-
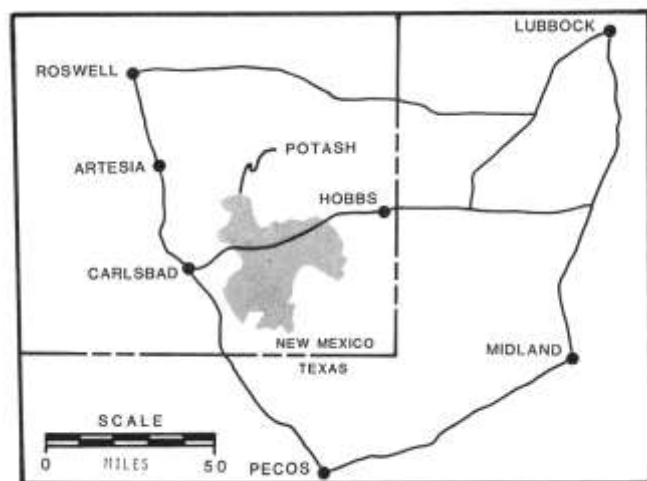
FIGURE 1—LOCATION MAP OF THE CARLSBAD POTASH DISTRICT.
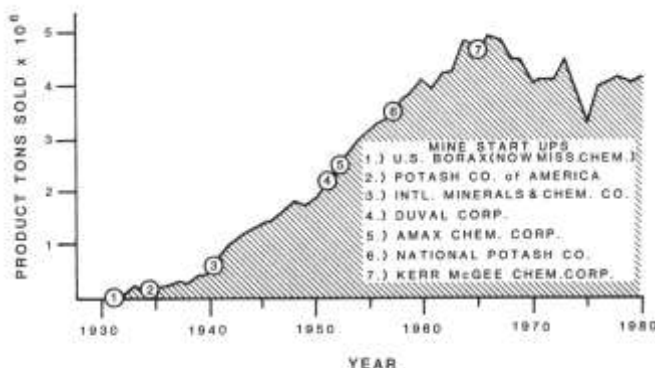
FIGURE 2—POTASH PRODUCTION IN THE CARLSBAD POTASH DISTRICT. Includes all potassium salts (sylvite, langbeinite and K_2SO_4).

lying Permian Basin oil fields. The interested reader may wish to compare the outline of the potential potash-bearing area on a map given by Mansfield in 1933 with the outline of the presently known district. The congruency is remarkable.

Published studies

All people interested in potash mineral deposits look forward to the eventual publication of a Professional Paper on the Carlsbad district by USGS geologists. Such a paper would serve a real need to synthesize in a single volume all of the simplicities and complexities concerning genesis, useful ore controls, and exploration techniques that have been developed over the past 50 yrs. In the interim, one can rely on the considerable published literature, some of which is referenced in this paper. Special note should be taken of the works by Lang, Adams, and King for discussion of the Ochoan evaporites and by Jones for the potash beds that occur within them. The published contributions by Charlie Jones have been augmented by many years of both wise and free counsel to both mining and exploration companies.

Structural setting of the Permian Basin

Perhaps the best way to describe the basic setting of the west Texas-New Mexico portion of the Permian Basin is to present first the major structural elements and then to superimpose the known mineral resources. The major elements are shown in fig. 3 and the relationships to resources are given in fig. 4. The basin can be divided into three sub-basins: Tatum, Delaware, and Midland. Marine-carbonate banks were built along the flanks of the north and central basin platforms during Guadalupian times. At its close, a pronounced barrier reef (Capitan Reef) completely surrounded the Delaware Basin. Back reef carbonates, shales, and some sandstone accumulated in the northwest shelf. Ochoan evaporite deposition commenced with the filling of the Delaware Basin by the deposition of thick salt and anhydrite beds of the Castile Formation. Ocean drying continued as evaporites formed the Salado and Rustler Formations not only in the deeper Delaware Basin but also on the back reef shelf and shallower basins. These evaporites were then buried by the Dewey Lake red beds that marked the close of the Permian. The present-day situation is shown by the north-south cross section (Jones, 1954; fig. 5). The trace of the cross section would be about midway between Carlsbad and Hobbs.

Permian oil-gas reservoirs follow the margins of the north and central platforms and continue into the northwest shelf. Smaller but significant production has occurred in sandstones
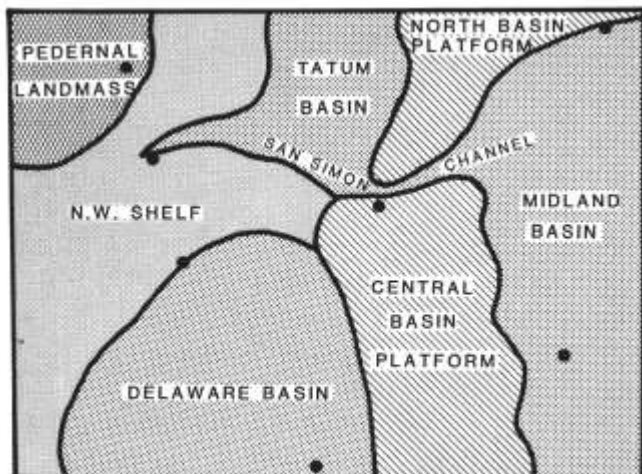
of the Delaware Mountain Group in the Delaware Basin. Pennsylvanian sediments are now being aggressively prospected, mainly for gas, west and north of the Delaware Basin. However, the trends of the Pennsylvanian gas fields are north-northeast and bear no relation to the Permian structures.

The potash-rich evaporites of the Salado Formation do not reveal any obvious structural control. I assume that most of Permian Basin was one large sabkha environment with numerous shallow basins in which bitterns accumulated. One of the largest of these basins was the area that the Carlsbad district now occupies and which is unique only in that it contains minable amounts of potash minerals.

Stratigraphy

Fig. 6 presents the overall, and then detailed character of Delaware Basin stratigraphy. The Ochoan is divided, from oldest to youngest, into three formations: 1) the Castile, principally containing salt and anhydrite, 2) the Salado, consisting of three members of which the middle or McNutt is potash rich, and 3) the Rustler, with anhydrite, salt, reddish mudstones and dolomite. The cyclical nature of deposition, commencing with mudstone at the base followed by anhydrite and finally salt, is much in evidence throughout the Salado. These rhythmic cycles, which average 4 ft thick but may be as thick as 20 ft and as thin as less than 1 ft, are thought to result from periodic climatic changes. The anhydrite beds are easily recognizable on certain geophysical logs and are amazingly consistent laterally throughout most of the mining district.

A uniform stratigraphic code has been developed by the U.S. Geological Survey to describe the Salado (Smith, 1938; Jones and others, 1960). The code designates 43 marker bed horizons, commencing with 100 at the top down to 143 which lies just above the Cowden Anhydrite. The Cowden and two other beds (Vaca Triste and Union) retain formal names to agree with prior usage in southeast New Mexico. The base of the Salado is normally located at the top of the first massive banded anhydrite in the Castile. Marker beds consist of a thin seam of mudstone at the base followed by a variable thickness of anhydrite. Frequently the anhydrite has been either partially or completely replaced by polyhalite.

Commercial quantities of potash are limited to the McNutt Member of the Salado. The McNutt is defined as the interval from marker bed 126 up to the Vaca Triste, a mudstone-sandstone bed. Eleven ore horizons normally are recognized within

FIGURE 3—MAJOR STRUCTURAL ELEMENTS OF THE TEXAS-NEW MEXICO PORTION OF THE PERMIAN BASIN.
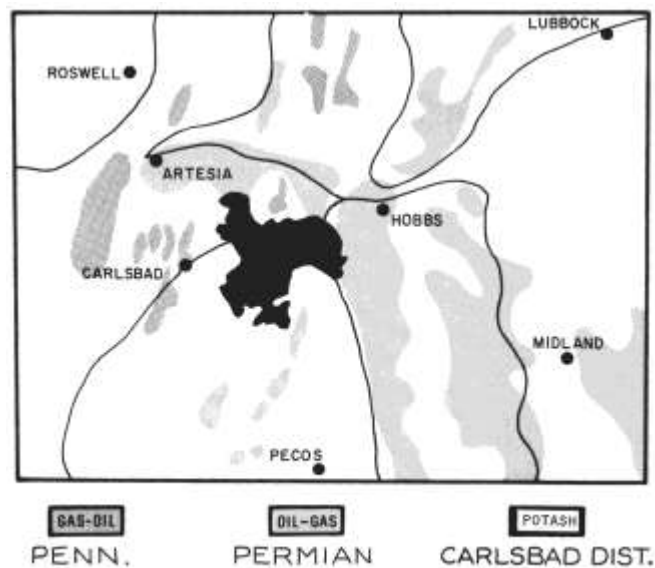
FIGURE 4—MINERAL RESOURCES OF PART OF THE PERMIAN BASIN.
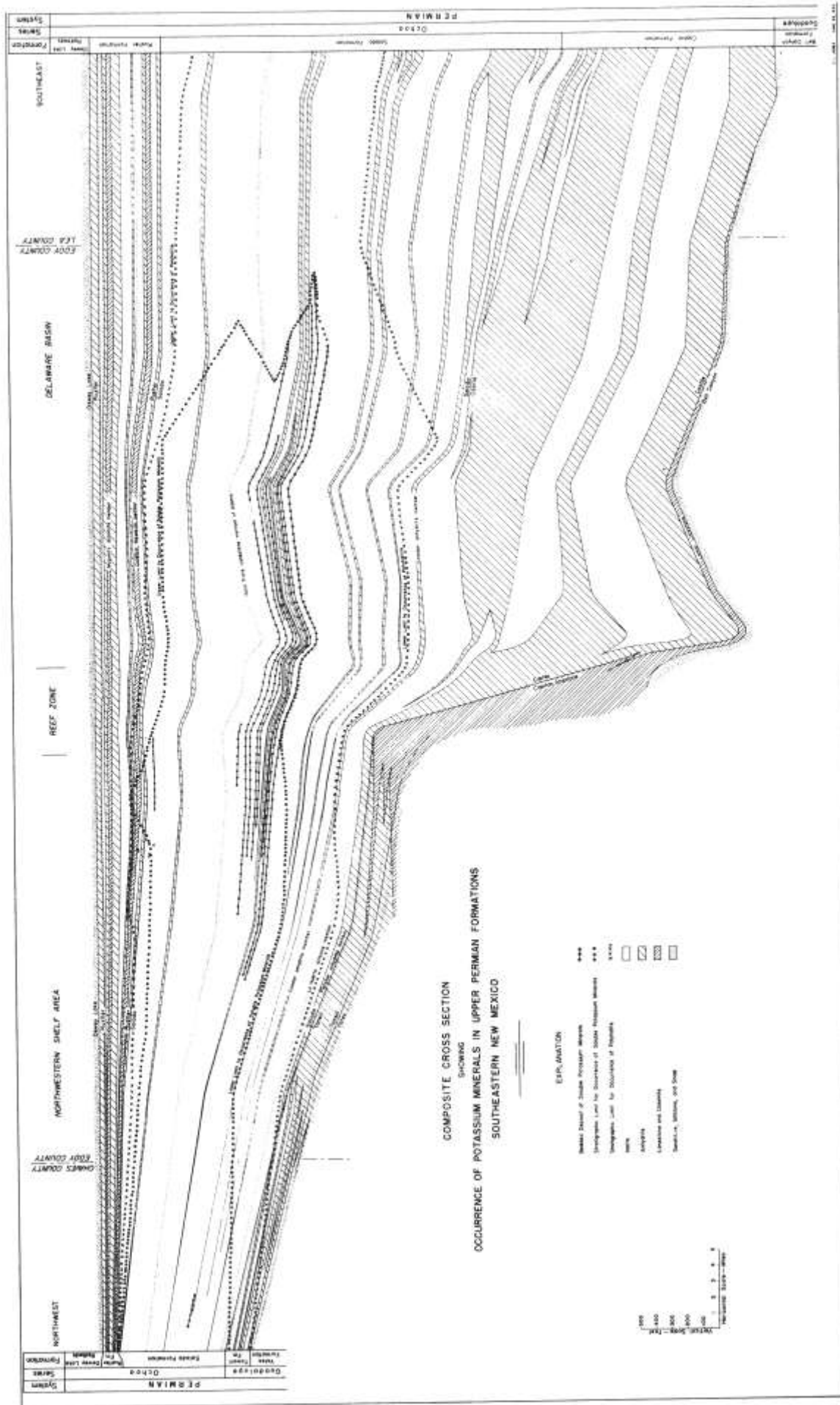
FIGURE 5—GENERALIZED NORTH-SOUTH CROSS SECTION OF THE CARLSBAD POTASH DISTRICT. The trace of the section is about midway between Carlsbad and Hobbs (from Jones, 1954).
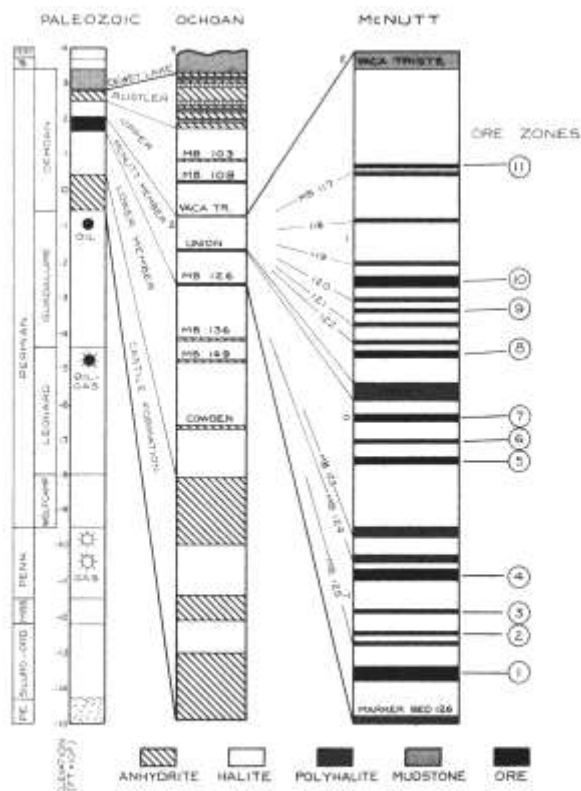
FIGURE 6—STRATIGRAPHIC COLUMN WITH EXPANDED SECTIONS OF THE OCHOAN EVAPORITE AND McNUTT POTASH MEMBER OF THE SALADO FORMATION. The pre-Salado section is typical of the central portion of the district but is south of the Capitan Reef. The Castile is limited to the Delaware Basin as shown in fig. 5.

TABLE 1—EVAPORITE MINERALS OF THE CARLSBAD DISTRICT. Only sylvite and langbeinite are ore minerals at the present time. The hydrated potassium minerals are not amenable to existing concentration methods; *common minerals in the Carlsbad district.

Mineral	Formula	Equivalent percent			
		(K)	(KCl)	(K$_2$O)	(K$_2$SO$_4$)
*Anhydrite	CaSO$_4$	—	—	—	—
Arcanite	K$_2$SO$_4$	44.88	—	54.06	100.00
Bischofite	MgCl$_2 \cdot$6H$_2$O	—	—	—	—
Bloedite	Na$_2$SO$_4 \cdot$MgSO$_4 \cdot$4H$_2$O	—	—	—	—
*Carnallite	KCl$\cdot$MgCl$_2 \cdot$6H$_2$O	14.07	26.83	16.95	—
Erytrosiderite	2KCl$\cdot$FeCl$_3 \cdot$H$_2$O	23.75	45.28	28.61	—
Glaserite	K$_3$Na(SO$_4$)$_2$	35.29	—	42.51	78.63
Glauberite	Na$_2$SO$_4 \cdot$CaSO$_4$	—	—	—	—
*Gypsum	CaSO$_4 \cdot$2H$_2$O	—	—	—	—
*Halite	NaCl	—	—	—	—
Hydrophilite	KCl$\cdot$CaCl$_2 \cdot$6H$_2$O	13.32	25.39	16.04	—
*Kainite	MgSO$_4 \cdot$KCl$\cdot$3H$_2$O	15.71	29.94	18.92	—
*Kierserite	MgSO$_4 \cdot$H$_2$O	—	—	—	—
*Langbeinite	K$_2$SO$_4 \cdot$2MgSO$_4$	18.84	—	22.70	41.99
*Leonite	K$_2$SO$_4 \cdot$MgSO$_4 \cdot$4H$_2$O	21.33	—	25.69	47.52
Mirabilite	Na$_2$SO$_4 \cdot$10H$_2$O	—	—	—	—
*Polyhalite	K$_2$SO$_4 \cdot$MgSO$_4 \cdot$2CaSO$_4 \cdot$2H$_2$O	12.97	—	15.62	28.90
Schoenite	K$_2$SO$_4 \cdot$MgSO$_4 \cdot$6H$_2$O	19.42	—	23.39	43.27
*Sylvite	KCl	52.44	100.00	63.17	—
Syngenite	K$_2$SO$_4 \cdot$CaSO$_4 \cdot$H$_2$O	23.81	—	28.68	53.06
Tachyhydrite	CaCl$_2 \cdot$2MgCl$_2 \cdot$12H$_2$O	—	—	—	—

the McNutt commencing with the first ore zone, which occurs between marker beds 126 and 125, up to the eleventh, located immediately above marker bed 117. Thus marker beds are numbered sequentially downward, and ore zones are numbered sequentially upward. Two other potash-rich beds are known to occur above the McNutt, but commerical amounts of potash mineralization are not known in them.

Characteristics of ore zones

The 11 ore zones are consistent in their stratigraphic relationship between marker beds; however, the thickness and mineralogy are quite variable. Mineral suites and ore-zone characteristics are given in tables 1 and 2.

Ore zones exhibit considerable effects of diagenetic processes. Jones (1954) divided potassic mineral occurrences into:
1) Accessory minerals
2) Stratified deposits in sulfate strata
3) Bedded deposits in mixed halite-clastic strata
4) Veins or lens deposits that have replaced or displaced the strata

The accessory-minerals category includes widespread but very low grade occurrences of soluble potassium minerals in halite beds. The stratified deposits in sulfate strata are those original anhydrite marker beds that have been at least partially replaced by polyhalite. The bedded deposits constitute the true ore zones of the Carlsbad district. The spacial relationship between the ore zones and the more widely distributed polyhalite and accessory-mineral occurrences of potash mineralization is observed in fig. 5. The polyhalite and accessory-mineral occurrences can be considered as halos. Veins or lenses of pure or mixed assemblages of potassium minerals occur, but in and adjacent to ore; their contribution to minable reserves is quite small.

TABLE 2—MINERALOGY AND MINABILITY OF ORE ZONES IN THE CARLSBAD POTASH DISTRICT.

Ore zone	Marker bed nearest base	Mineralogy	Minability
Eleventh	MB 117	Mostly carnallite, minor sylvite and leonite	Not commercial to date
Tenth	MB 120	Sylvite	Second best in the district
Ninth	MB 121	Carnallite, kierserite, minor sylvite	Not commercial to date
Eighth	Union	Sylvite	Moderate reserves, important in future
Seventh	—	Sylvite	Moderate reserves
Sixth	—	Carnallite, kierserite, etc.	Not commercial to date
Fifth	MB 123	Sylvite and langbeinite	Moderate reserves
Fourth	—	Langbeinite and sylvite	Principal source of langbeinite
Third	—	Sylvite	Ranks 3rd in production of sylvite
Second	MB 125	Carnallite, kierserite, etc.	Not commercial to date
First	MB 126	Sylvite	Was the major sylvite producing zone

Although ore zones remain in consistent stratigraphic positions, the original sedimentary features of the beds are mostly obliterated by authigenesis and recrystallization processes that followed original deposition. Though these processes caused the formation of commercial deposits in some localities, it is also apparent that late-moving fluids, unsaturated in potassium, destroyed what would have constituted ore. The apparent haphazard relationship between ore zones and barren horsts of salt is shown in fig. 7.

The north, south, and eastern edges of the Carlsbad district are based on drilling information while the western edge is the result of subsurface dissolution of the Salado by circulating ground water. The general area was tilted eastward by uplift of the Guadalupe Mountains on the west during Laramide time. Ground waters associated with the Pecos River drainage basin have attacked the salt. The effect on the Salado is shown by the isopach contours in fig. 7. Considerable amounts of ore may have been lost to dissolution on the west. For a more complete description of the dissolution process see Brokaw and others, 1972.

Mining operations

The first ore zone was by far the thickest, highest grade, and most extensive sylvite-ore mining horizon in the district. This zone probably has accounted for 80% of the past production of the district. This zone is mostly mined out and current production is being sustained from the tenth, seventh, fifth and third ore zones. Langbeinite is produced principally from the fourth zone, with important amounts of mixed sylvite and langbeinite being mined from the fifth zone. The capacities of mines and ore zones being mined are given in table 3.

The capacities given are for raw ore, not product produced after refining. Current ore grades are about 14% K_2O equivalent for sylvite and 8-10% K_2O-equivalent for langbeinite

TABLE 3—OPERATING COMPANIES AND ORE ZONES MINED IN THE CARLSBAD POTASH DISTRICT.

Company	Capacity (tons/day)	Ore zones mined (past)	Ore zones mined (present)
Mississippi Chemical	5,000	1	5, 7
Potash Company of America	10,000	1	1, 3
International Minerals & Chemical	14,000	1, 4, 5	4, 5
Duval Corporation	3,500	1, 4, 10	4
AMAX Chemical	9,200	1	3
National Potash	10,000	1, 10	10
Kerr-McGee Chemical	9,000	10	10

ores. The district produces 4 million tons of muriate (KCl) on the average with a K_2O equivalent content of 60-63%. The langbeinite production is on the order of 250,000 tons per year with an approximate K_2O content of 22%. In addition, some K_2SO_4 is produced by one company as a by-product of its sylvite- and langbeinite-refining operations. The aggregate value of these products sold in 1980 was $314 million. One company is capable of producing $MgCl_2$, but market conditions are not favorable now.

Each of the seven mining companies has completely integrated plants that include mines, processing plants, and connecting rail and power networks. Underground mining has always been efficient. In the earlier years, conventional drill, blast, load, and rail-haulage methods were used to implement room and pillar mining. Mechanized coal-mining equipment has been modified or customized to fit the harder potash-salt formations. Today all haulage is done via conveyors that are fed by diesel or electrically driven shuttle cars. Mining, particularly in the thinner beds (down to 4 ft), increasingly is being done with boring machines. In the hard langbeinite ore, mechanical undercutters are used to prepare faces for subsequent drilling and blasting. In all cases, mechanical arm-type loaders are used. Productivity is high, with an overall average of 40 tons per underground man shift. In areas mined by room and pillar, 60% of the ore is normally removed during initial mining. Subsequent removal of pillars allows overall extraction to exceed 90%. In thinner seams, modified longwall techniques are used, again with excellent recovery.

Commercial products of muriate (KCl), langbeinite ($K_2Mg_2(SO_4)_3$), and K_2SO_4 are produced by several techniques. The flotation process is preferred for the simple sylvite ores. In the eastern part of the district, the tenth ore zone contains considerable quantities of dispersed clay. These ores are treated by a leach-crystallization process. The langbeinite ores are treated by washing, flotation, heavy media, or any combination thereof. Potassium sulfate is produced by a base-exchange process between sylvite and sulfate-rich potassium-magnesium liquors. Overall recoveries exceed 90% for sylvite and 80% for langbeinite ores. On the average, 25 tons of raw ore are processed per man shift including underground, surface, plant, and administrative personnel.

FIGURE 7—KNOWN POTASH LEASING AREA (KPLA). The U.S. Geological Survey has declared the area within the boundary as a known resource area based on mining and drill-hole information. The extent of mining as of 1976 along with known barren areas is also shown. The contours are thickness in feet of the Salado Formation; note the effect of dissolution of salt from the west (adapted from Aguilar and others, 1976).

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LANGBEINITE IN POTASH DEPOSITS

by Norbert T. Rempe, *International Minerals and Chemical Corporation, Carlsbad, New Mexico 88220*

Abstract

Langbeinite ($K_2SO_4 \cdot 2MgSO_4$) occurs in evaporite deposits worldwide. Economically significant concentrations, however, are limited to New Mexico and the Ukraine. Langbeinite has been mined continuously for over 40 yrs from the Permian Delaware Basin and intermittently from the Tertiary Sub-Carpathian molasse trough since at least the 1930's.

Langbeinite deposits are found in many tectonic regimes. Examples are:
1) alpinotype: Stebnik mine, Ukraine, U.S.S.R.;
2) germanotype: Hildesia-Mathildenhall mine, Lower Saxony, Germany;
3) epeirogenic: Northern Delaware Basin, New Mexico, U.S.A.

Langbeinite cannot precipitate directly out of seawater and is always the alteration product of a primary mineral paragenesis. Either the existing stock of elements is merely rearranged by partial or complete melting and recrystallization, or the invasion of extraneous solutions, commonly from below, causes partial or complete solution, replacement, and reprecipitation. Definite metamorphic textures in langbeinite-bearing seams testify to the presence of normal or progressive geothermal metamorphism, with gradual enrichment of migrating solutions in $MgCl_2$, and/or of retrograde metamorphism, with decreasing $MgCl_2$ concentrations of alteration brines.

Introduction

Langbeinite is a rare potash mineral that nevertheless occurs in evaporite deposits worldwide. Although many salt and potash deposits contain langbeinite in only minor amounts, it is concentrated in a few deposits in such quantity that extraction is economically feasible.

Langbeinite, a double sulfate of potassium and magnesium, has the chemical formula of $K_2SO_4 \cdot 2MgSO_4$. Just like the simple sulfate of potassium, K_2SO_4, which does not exist naturally and is therefore manufactured from other potash minerals, langbeinite is used as a specialty fertilizer for crops that cannot assimilate potassium in the chloride form. Potatoes, sugarbeets, and tobacco are a few of these crops.

Principal occurrences

Langbeinite is mined and processed in only two regions of the world today. The most important one for the free world is in southeast New Mexico where it is produced by two mines in the Carlsbad potash district (fig. 1). The other region is
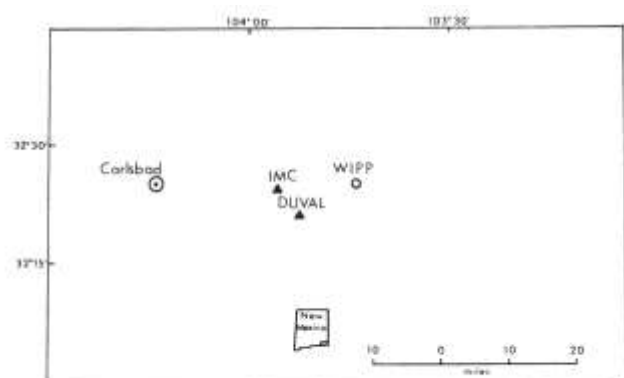
FIGURE 1—LANGBEINITE MINES IN THE CARLSBAD POTASH DISTRICT.

the southern Ukraine, U.S.S.R., formerly part of Poland, in the northern foredeep (Molasse Basin) of the Carpathian Mountains (fig. 2). The two geologic settings could hardly be more different.

Somewhere between these two extremes lies a third langbeinite occurrence. This deposit was not mined for its langbeinite, but for the other potash minerals it contained, and is situated in the salt anticline of the Hildesheimer Wald, south of the city of Hannover, in the Federal Republic of Germany (fig. 2).

This paper briefly compares these three langbeinite occurrences. Information on the Ukrainian and German locations was derived for the most part from the available literature, whereas the New Mexico deposit was accessible to direct observation and study by the author.

Geologic framework

Langbeinite has been mined since at least the 1930's (Johnson, 1944) from the Lower Miocene strata at the Stebnik mine in the Ukraine. The product was sold under the tradename "Kalimag." The original sedimentary suite underwent intensive folding and thrust-faulting (Gunzert, 1967). Utilizing some old-fashioned geotectonic terminology, the general environment can suitably be characterized as alpinotype.

The host beds of the now inactive Hildesia-Mathildenhall mine in the province of Lower Saxony, Germany, are Zechstein 2, approximately 230 m.y. old; however, careful petrographic and structural analysis (Kokorsch, 1960) revealed that the langbeinite in this deposit is of a much younger age, somewhere between Upper Jurassic and Upper Cretaceous. Radiometric age determinations support these findings with a langbeinite mineral age of only 120 m.y. (Pilot and Blank, 1967).

Langbeinite from southeast New Mexico has been produced by International Minerals and Chemical Corporation (IMC) since 1940 (sold under the trade name "Sul-Po-Mag") and by Duval Corporation since 1964, sold under the trade name "K-Mag." Both mines are situated in the southern portion of the Carlsbad district within the Delaware Basin. Studies on the genesis of the potash minerals, and especially of the langbeinite in the district, are sorely lacking, but a single radiometric langbeinite mineral-age determination (Schilling, 1973) established an age compatible with the Late Permian origin (245 ± 10 m.y.) of the overall evaporite sequence, very much in contrast to the German results.

Metamorphism in evaporites

Langbeinite forms by metamorphic processes (Braitsch, 1971, p. 13).

At this point, a digression on metamorphism in evaporites is appropriate. Borchert and Muir (1964, p. 96) state in this context:

"R. Kuehn (1955b, p. 57) finds the use of the term 'metamorphism' objectionable, because he thinks it should be applied only to mineralogical changes in silicate rocks. There is absolutely no reason to so confine it. Mineralogical and textural changes are fundamentally similar in silicates and evaporites. In both cases `authigenic liquors' play an important role. Evaporites are metamorphosed at shallow depths and low temperatures where the silicate rocks would undergo only diagenetic changes and not true metamorphism, but that does not alter the fact that the texture of most Zechstein evaporites is largely similar to that of crystalline quartz schists formed in the katazone. In each case the texture results from increase in temperature and pressure."

FIGURE 2—STEBNIK MINE, UKRAINE, U.S.S.R., AND HILDESIA-MATHILDENHALL MINE, LOWER SAXONY, GERMANY.

In a subsequent chapter Borchert and Muir (1964, p. 221) write, emphasizing langbeinite's non-primary character: "In any case it is very unlikely that important amounts of a langbeinite paragenesis could ever have precipitated from sulfate-deficient natural brines . . ."

Braitsch (1971, p. 93) contributes the following: "In salt deposits, percolating secondary solutions cause such far-reaching alterations, even if the temperature is unchanged, that they are best considered under metamorphism . . ."

Even the author of probably the most concise text on metamorphic (silicate) rocks agrees when he states (Winkler, 1974, p. 2): ". . . this book is not concerned with salt deposits (evaporites) because their metamorphism takes place at considerably lower temperatures and pressures than the metamorphism of silicate and carbonate rocks."

In conclusion, no authority on the subject any longer debates the question of diagenesis vs. metamorphism with respect to post-depositional alterations of evaporites. Petrographic evidence and physicochemical data, as well as the quotations above, amply support the conclusion that langbeinite is as much an index mineral for metamorphism in evaporites as sillimanite or garnet are index minerals for metamorphism in silicate rocks.

Metamorphic processes

Langbeinite forms by two principal processes:
1) Halite and kainite (KC I MgSO$_4$ • 2.75H$_2$O) recrystallize at 83 °C into langbeinite, sylvite (KCl), and kieserite (MgSO$_4$ • H$_2$O) (Braitsch, 1971, p. 111, table 13).
2) Sylvite and kieserite change into langbeinite and halite at the same temperature (Braitsch, 1971, p. 123, table 16, and p. 177).

Langbeinite of unquestionably primary origin has never been reported anywhere.

In the Sub-Carpathian deposit, langbeinite is thought to have developed from kainite and halite by thermal metamorphism (D'Ans and Kuehn, 1940). The necessary temperatures were probably reached during the Moldavian orogenesis (Stille, 1953) that took place at the turn from the Tortonian to the Sarmatian in the Upper Miocene.

A far more complex history is attributed to the Hildesia-Mathildenhall deposit (Kokorsch, 1960). The essence of the argument here is that brines of varying composition moved upward on joints and faults that opened up during germano-type tectonic movements. These movements took place in the Kimmeridgian (Upper Jurassic), during the Young-Kimmerian orogenesis (Stille, 1924), and again in the Senonian (Upper Cretaceous), during the Sub-Hercynian orogenesis (Stille, 1924).

To explain the origin of brines of varying composition, Kokorsch (1960) drew heavily on the concept developed primarily by Borchert (1940) and Borchert and Muir (1964), of normal or progressive and retrograde metamorphism in evaporites. A very simplified version of this concept follows:

First, magnesium-chloride-rich brines and later, sodium-chloride- and calcium-sulfate-rich brines, migrated upward and changed the chemistry and mineralogy of the overlying evaporite suite. The brines probably existed long before their migration upward, but they could not begin their ascent until after the required pathways had been opened.

Magnesium-chloride-rich brines are generated by the transformation of primary, water-bearing magnesium salts into minerals containing little or no water due to increases in temperature and/or pressure (normal or progressive metamor-

phism). Sodium-chloride- and calcium-sulfate-rich brines can be derived from the conversion of primary gypsum into secondary anhydrite (causing retrograde metamorphism). This process releases 0.486 cm' of water from each cm' of gypsum (Borchert and Muir, 1964, p. 109). NaCI is picked up during the migration through halite beds.

Analogies and discrepancies

The German and New Mexican deposits are both underlain by massive anhydrite. In the Carlsbad district the anhydrite beds within the Castile, below the potash-bearing Salado Formation, are most massive in the southern part of the basin, seaward from the Capitan reef (Jones, 1954). This area is, coincidentally, also where langbeinite occurs in minable concentration and quantity. Sulfates are known to exist north of the reef as well, but there they are scarce and not minable.

Metamorphic conditions in German evaporite deposits came about either by deep burial or by magmatic events or by combinations of both. No magmatic source of sufficient energy output is known from the Carlsbad district or its vicinity, but educated guesses about the original thickness of the almost completely removed Cretaceous strata range up to 2,000 ft. That thickness, in addition to today's remaining overburden of up to 1,000 ft, may have provided enough pressure and geothermal gradient to trigger the metamorphic process described above.

At present, no satisfactory pathways for the metamorphic solutions have been positively identified in the Carlsbad district. The Delaware Basin has been a stable, epeirogenic environment since the salt formations were originally deposited. Aside from some minor and, in this context, insignificant tilting (Powers and others, 1978), no orogenic movements are known to have occurred. The only known igneous feature in the basin is a Tertiary lamprophyric dike whose alteration zone does not extend beyond a maximum of a few feet on either side (Loehr, 1979).

Despite the conditions listed above, petrographic and mineralogical evidence in cores and mine openings clearly demonstrates the role played by non-primary processes in the formation and alteration of langbeinite as well as other evaporite minerals. For comparison with and analogies to the selected features from the Carlsbad district depicted in figs. 3-7, the reader is referred to the pertinent illustrations in Kokorsch (1960), Borchert and Muir (1964), and Lobanova (1953).

FIGURE 4—LANGBEINITE TETRAHEDRA IN 5TH ORE ZONE, HEAVILY CORRODED BY KAINITE/LEONITE (petrographic identification by T. K. Lowenstein); see also Borchert and Muir, 1964, p. 127 identical with Kokorsch, 1960, table 3 and fig. 5.
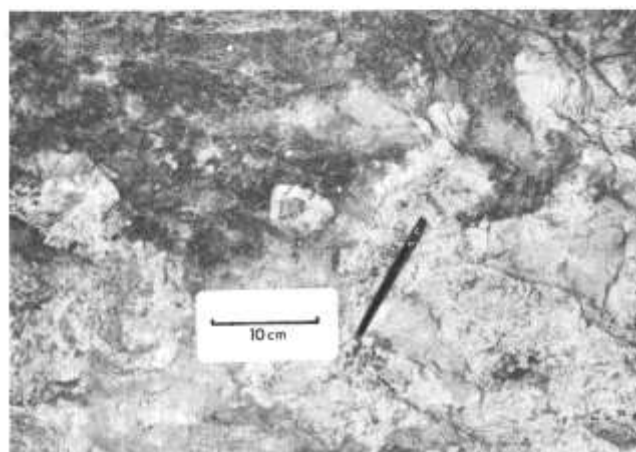
FIGURE 5—LANGBEINITE IN 1ST ORE ZONE, RIMMED BY LEONITE/KAINITE; see also Lobanova, 1953, p. 146, fig. 3.

FIGURE 3—LANGBEINITE TETRAHEDRA IN 5TH ORE ZONE, SURROUNDED BY THIN RIMS OF KAINITE/LEONITE (petrographic identification by T. K. Lowenstein); see also Borchert and Muir, 1964, p. 127 (identical with Kokorsch, 1960, table 3, fig. 5); Borchert and Muir, 1964, p. 127: "It is only very rarely that idiomorphic crystals represent original precipitation; more commonly, they are produced by metamorphism."

FIGURE 6—LANGBEINITE TETRAHEDRA IN 1ST ORE ZONE WITH ROUNDED INCLUSIONS OF HALITE, KIESERITE, AND POSSIBLY SYLVITE; crust of hydrated sulfates in upper right quadrant, below a thin clay seam; see also Borchert and Muir, 1964, p. 137 (identical with Kokorsch, 1960, table 2, fig. 5) and Kokorsch, 1960, table 6, fig. 8; Borchert and Muir, 1964, p. 135: "When the cores of idiomorphic tetrahedra of langbeinite contain rounded inclusions of kieserite, sylvite, and halite, we must conclude that a hartsalz paragenesis (a mixture of sylvinite and kieserite) has been altered by increasingly sulphatic solutions. As langbeinite therefore became the stable solid phase, porphyroblasts of this mineral developed."

FIGURE 7—ZONED LANGBEINITE IN 4TH ORE ZONE, with core of clear langbeinite, middle zone of sylvite with little kieserite, polyhalite, and langbeinite, rim of clear langbeinite, with thin outermost band of dark red sylvite, and interstitial red-brown, langbeinite-bearing clay (mineral identification by S. Tschappler); see also Lobanova, 1953, p. 147, fig. 4; Borchert and Muir, 1964, p. 103: "There can in fact be no doubt that the langbeinite (often 5–10 cm in diameter) of the Pre-Carpathian deposits is truly metamorphic, and formed by intracrustal, geothermal melting."

Summary

Langbeinite in Ukrainian and German potash deposits has formed by metamorphic processes. In both cases, sources of metamorphism have been identified. Langbeinite from southeast New Mexico closely resembles many of the petrographic and mineralogical features described and depicted from these other occurrences. Although no definite source of metamorphism has yet been identified in the Carlsbad district, there can be no doubt that all langbeinite in potash deposits is of metamorphic origin.

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POTASH IN LIBYA

by Gus H. Goudarzi, *U.S. Geological Survey, Reston, Virginia 22092*

Abstract

Potash deposits in Libya that occur in salt flats, locally known as sebcha, are formed as lagoonal deposits near the Mediterranean coast and as playas in closed depressions in the desert interior. The lagoonal deposits along the coast are natural salt flats in depressions separated from the Mediterranean Sea by a narrow sandy barrier. These ephemeral deposits of brine-impregnated salt are formed during the hot summer months by solar evaporation of seawater percolating through the sandy barrier into coastal depressions. An example of this type of deposition is in the Pisida deposits (Mellahat al Bariqah) in northwest Libya, where surface evaporites contain an average of 88% NaCl, 6.8% KCl, and 0.9% $MgCl_2$, and the brine contains 9% $MgCl_2$. In the desert interior, the salts, which are dissolved out of the rocks, find their way into depressions where, by evaporation, salt deposits such as those at Edri in south-central Libya are formed on the surface. At Edri, the salts contain an average of approximately 69.9% NaCl, 2.8% KCl, and negligible $MgCl_2$. The Marada deposits, also in the interior, are thought to be concentrations of saline residues left by receding Miocene seas. Meteoric waters dissolve the residual salts from the surrounding sediments and transport them to the low parts of the basin, supplementing the already concentrated brine of the sebcha. Salts are deposited through a cycle of percolation, capillary migration, evaporation of the brine, and accumulation of saline deposits on the surface. The brines yield a mixture of $MgSO_4$, $MgCl_2$, NaCl, and KCl, and the primary mineral is carnallite ($KMgCl_3$ $6H2O$). Grab samples of the salts contain as much as 51% NaCl and 25% KCl; the naturally concentrated brine contains about 34% $MgCl_2$. Extensive drilling and studies made recently by private companies suggest that recovery of potassium, sodium, and magnesium from the Pisida and Marada deposits is economically feasible. The deposit at Edri could be worked repeatedly if selective mining were practiced and if only part of the area were mined each year.

Introduction

Saline deposits in Libya, locally known as sebcha, are formed as lagoonal deposits along the Mediterranean coast by infiltration of seawater through or over a sandy barrier into coastal depressions. During the hot and dry season, salts are deposited by solar evaporation (fig. 1). In the desert interior, meteoric waters dissolve the residual salts from the surrounding sediments and transport them to the low parts of the basins where, by evaporation, salt deposits are formed on the surface.

These deposits occupy large areas along the Mediterranean coast at Pisida, near Tripoli, and on the west side of the Gulf of Sirte at Tauroga; deposits are found at Tamimi, near Bengasi, and along the coast of the Gulf of Sirte where they are separated from the sea by a coastal ridge. In the desert interior, deposits are found as salt flats and playas at Marada, Giarabub, Kufrah, Wau al Kabir, Gatrun, Ghat, Traghan, Sebha, Edri, and Ghadames (fig. 2).

The deposits near Tripoli and Bengasi occur in basins slightly below sea level and have been utilized as a source of common salt for many years. The salts are recovered from several manmade evaporating pans in a basin or depression fed directly from the sea by a canal. Brines are concentrated by solar heat and crystallized salts are harvested and marketed in crude form.

Italian geologists studied the saline deposits in Libya beginning in 1926; the results of their investigations were reported by Desio (1943). Studies were made by the author from 1954 to 1958 (Goudarzi, 1956, 1968, 1970) as part of a U.S. Geological Survey cooperative program with the Libyan Ministry of Industry, under the auspices of the Agency for International Development (AID) of the U.S. Department of State. During these studies, many of the saline deposits were examined and sampled as a possible source of potash. Detailed sampling and experiments were conducted at the Marada, Edri, Tauorga, and Pisida deposits to determine the economic feasibility of recovery of potash and magnesium salts. The present paper summarizes these studies and incorporates the results of the more recent exploration by private companies (J. Krason and others, unpublished data, 1978).

FIGURE 1—TYPICAL VIEW OF A SEBCHA (SALT FLAT) SHOWING A WHITE CRUST OF SALT FORMED BY EVAPORATION. Photograph taken at Marada by Harry F. Thomas.

FIGURE 2—MAP SHOWING LOCATION OF SALINE DEPOSITS IN LIBYA.

Saline deposits

PISIDA

The Pisida saline deposit is approximately 165 km west of Tripoli on the Mediterranean coast at approximately 33° N latitude, 12° E longitude. This deposit is approximately 5 km wide and more than 15 km long in Libya but extends westward and crosses the Libyan border into Tunisia. That part of the deposit in Libya has an area of about 50 km^2.

The Pisida deposit is in a coastal depression separated from the sea by a narrow, sandy barrier. Seawater percolating through this sandy barrier accumulates in the depression where, by capillary migration and evaporation during the hot and dry summer months, a layer or crust of salt is deposited on the surface. This crust is as much as 10 cm thick and contains an average of 88% NaCl, 6.8% KCI, and 0.9% MgCl$_2$. This surface evaporite (crust) is underlain by about 40 cm of partially crystallized material (undercrust) having a high brine content. This material averages 66% NaCl, 6% KCl, and 1.2% MgCl2. The brine is at a depth of about 50 cm and averages approximately 17% NaCl, 1.5% KCI, and 9% MgCl2 (fig. 3).

The order of deposition of the salts is controlled by the relative solubility and the abundance of the components. The low concentrations of potassium and magnesium salts in the surface at Pisida may be due to the relatively cool and humid temperatures of the coast which would prevent completion of the ideal cycle of evaporation. Halite (NaCl) and sylvite (KCl), the least soluble salts, crystallize at the surface (crust) and the subsurface (undercrust), respectively, resulting in a relatively high magnesium-chloride (MgCl$_2$) content in the residual brine.

By comparison with some of the well-known brines in other areas in the world, brines at Pisida are relatively rich in NaCl but low in KCl and MgCl$_2$ (table 1).

The Pisida deposit was first investigated by Italian chemists in 1929 under the direction of Dr. Niccoli. A process, called the "Niccoli process," was devised to use solar energy to concentrate the brine. This highly concentrated brine was further treated to obtain a marketable potassium and magnesium salt. During the present study, laboratory experiments indicated that extraction of potassium and magnesium salts from the Pisida brines may be possible by fractional crystallization.

During the investigation reported in this paper, samples were collected on a grid basis (fig. 4) at 19 locations, and two holes were drilled along the periphery of the deposit to determine the existence of bedded salts at depth, but none were encountered. Recent study and detailed exploration conducted on behalf of the Libyan Government by Geo-Explorers International, Inc., Denver (Jan Krason, personal communication,
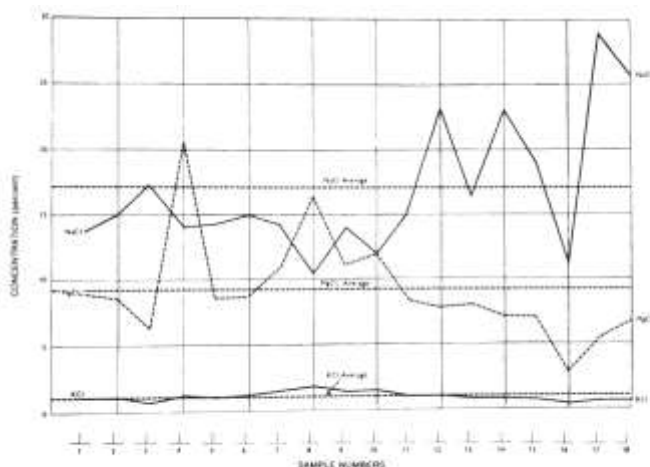
TABLE 1—COMPARISON OF SALT CONTENTS (IN GRAMS PER LITER) OF PISIDA BRINE WITH THOSE OF BRINES IN OTHER AREAS.

Specific gravity	KCl	NaCl	MgCl₂	MgSO₄	Area
1.2	11.5	85.0	138.0	—	Dead Sea brine
1.2	15.0	110.0	179.0	34.0	Tunisia-Sebcha al Melah Aioun
1.2	17.2	197.0	74.0	32.0	California bitterns (U.S.)
1.2	6.6	272.0	50.0	—	Pisida brine

1981), indicate that, in fact, bedded-salt deposits do exist at depth in the basin.

Geo-Explorers International, Inc., reports that 70 holes were drilled, ranging in depth from 15 to 45 m. As much as 11 m of salts interbedded with gypsiferous clay and pure gypsum were found in drill holes. The deposit was estimated to contain a total of 170 million metric tons of reserves, of which about 54 million tons were 98% pure NaCl and 31 million tons were potassium salts; reserves of gypsum were estimated at several million tons of 100% pure gypsum; the brine was estimated to contain 11.5 million tons of potassium salts (Jan Krason, personal communication, 1981).

MARADA

The Marada saline deposits, between 18° 57 ' and 19° 25' E. long., 29° 6' and 29° 10' N. lat., are approximately 125 km south of the Gulf of Sirte and cover an area of about 150 km^2.

The deposits occur in the lowest area of an erosional depression and now lie at the level of contact of lower Miocene and Oligocene formations; they are believed to be concentrations of saline residues left by receding Miocene seas. The presence of saline deposits at Marada can be explained by the fact that they occur in a closed depression. Rainwater, in its course of travel over and through the Eocene and Oligocene rocks, dissolves the residual salts and transports them to the lowest part of the depression; these solutions supplement the already concentrated brine of the basin. The Marada brine consists of a mixture of MgSO$_4$, MgCl$_2$, and KCI, and the principal minerals are halite (NaCl) and carnallite (KMgCl$_3$ • 6H$_2$0). The deposits are superficial, of recent origin, and are being formed by evaporation and concentration during the hot, dry summer months.

Deposition takes place through a cycle of percolation, capillary migration, and evaporation of the brine, and accumulation of saline deposits on the surface. The order of deposition is controlled by the relative solubility of the components of the brine; the least-soluble salts, gypsum and halite, are found in the western part of the basin, whereas in the lowest

FIGURE 3—SALT CONCENTRATIONS IN THE PISIDA BRINE. Sample locations are shown in fig. 4.
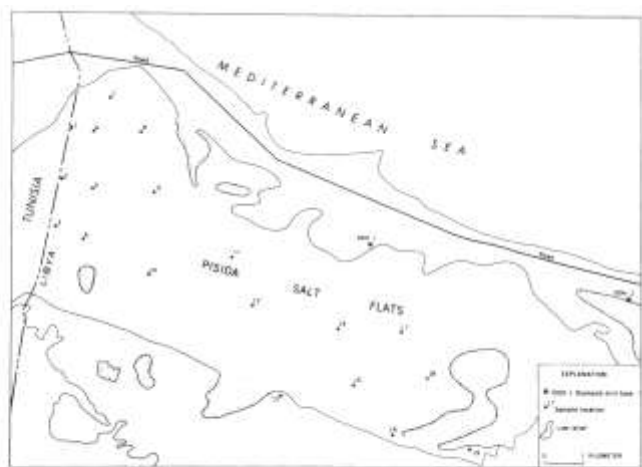
FIGURE 4—LOCATIONS OF DRILL HOLES AND SAMPLES IN THE PISIDA SALT FLATS.
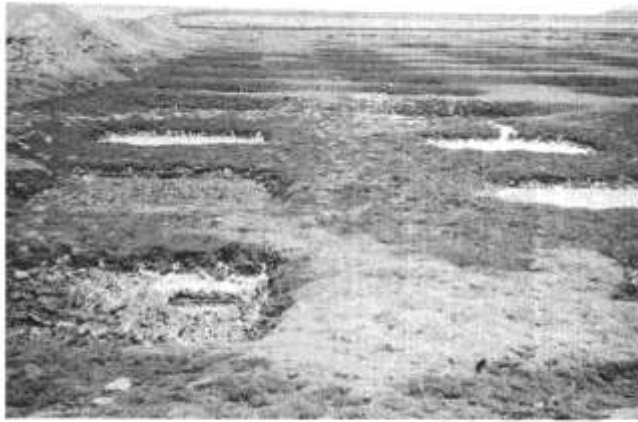
FIGURE 5—VIEW OF SALT FLATS AT MARADA IN 1956. The white material is carnallite in experimental pits.

TABLE 2—ANALYSES OF SAMPLES FROM MARADA (in percent); water-insoluble components not determined; Ledoux and Co., Teaneck, N.J., analysts.

Sample	NaCl	KCl	MgCl$_2$
Crystalline salts (grab sample)	51.0	25.0	9.0
Evaporites in pits	68.0	23.0	2.0
Natural concentrated brines in pits	30.0	25.0	34.0
Surface evaporite deposits	11.5	3.5	20.5
Harvested material	62.0	9.0	8.5

TABLE 3—AVERAGE COMPOSITION OF SALTS FROM EDRI (in percent); water-insoluble components were not determined.

	NaCl	KCl	MgCl$_2$
Crust	69.9	2.8	1.6
Undercrust	13.5	.6	.4
Brine	11.2	.5	.6

part of the area, the east-central part, the more soluble potassium and magnesium chlorides accumulate. Residual material on the bottom of the depression consists of alternate beds of clay and sandstone interbedded with salt rock.

During the years 1938-1940, an investigation was made by the Italians, at which time 15 km^2 of the area were found to contain high concentrations of potassium and magnesium salts. A number of pits 2 m square (fig. 5) and approximately 50 cm deep were dug and the brines were allowed to evaporate. Within a week, about 80 kg of crystallized salts (a mixture of MgSO4, MgCl$_2$, NaCl, and KCl) had accumulated and were recovered (harvested) from each pit. This material was further purified, and a product containing 40-50% K20 equivalent was obtained. On the basis of drill-hole information, reserves were estimated to be l.6 million metric tons of potassium salts containing 40% K$_2$0 equivalent and 7.5 million metric tons of MgCL$_2$.

The Marada deposits were examined and sampled (table 2) during this investigation, and the results suggest that the deposits offer favorable opportunities for commercial exploitation (Goudarzi, 1968). More recent studies by private industry have indicated substantial reserves of bedded salt at Marada; no new reserve estimates are available.

EDRI

The Edri salt flats, located near 27 °30 'N. latitude and 13° 10 'E. longitude, 135 km west of Brack in south-central Libya, cover an area of about 35 km^2. The deposits lie in a closed depression in a region of Devonian rocks where interbedded sandstone and sandy siltstone are capped by a conglomeratic, ferruginous sandstone.

The origin of the deposits is attributed to percolating meteoric waters that dissolve the soluble constituents from the surrounding rocks and carry them into the Edri depression. These waters are concentrated to a brine that saturates the sand; then, by percolation, capillary action, and evaporation, salt crusts accumulate on the surface (fig. 6).

The area was sampled by digging 25 pits (fig. 7, prefix P) from the crystallized salts at the surface (crust), the material below (undercrust), and the brine; the average compositions of these samples are given in table 3.

In a second phase of the study, 14 localities (fig. 7) were selected, and the surface crusts were removed in plots of about 30 m square. In each plot 25 pits, each approximately 3 m square and approximately 1 m deep, were dug and the brines were allowed to accumulate and evaporate by solar heat. Brine samples were collected periodically; although a trend towards a gradually higher concentration of salts in the brine was noted, no appreciable increase in the total dissolved solids was observed. Fig. 8 shows different stages of evaporation in the pits.

The third phase of studies was fractional crystallization of the salts by solar evaporation. Two pans, designated as A and B, were built in each of the 14 localities. Brines were poured into pan A, sampled, and allowed to evaporate. After several days the residual brine was transferred to pan B, sampled, and allowed to evaporate to dryness. The crystallized salts were then collected, weighed, and sampled. A fairly pure table salt was obtained from the brines on the first stage of crystallization, and the experiments indicate that small amounts of potash may be obtained as a byproduct during the second stage of crystallization. Laboratory experiments indicate that approximately 200 kg of a product containing 6-10% potash (K2O) equivalent may be recovered from 1 metric ton of the naturally formed salts.

The deposit contains approximately 3 million metric tons of naturally formed crude salts (crust) representing approximately 2 million tons of table salt (halite), and about 100,000 tons of potash (K20) equivalent. If the crust is removed, the salts probably will continue to accumulate on the surface and can be periodically harvested and processed. Consequently, the deposit could be worked repeatedly if selective mining were practiced and if only part of the area were mined each year.

FIGURE 6—ACCUMULATION OF SALINE RESIDUE IN PLAYA DEPRESSION AT EDRI. Note the hummocky, rough surface of the deposits.

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FIGURE 7—LOCATIONS OF SAMPLES, EXPERIMENTAL PLOTS, AND TEST PITS IN THE EDRI SALT FLATS.
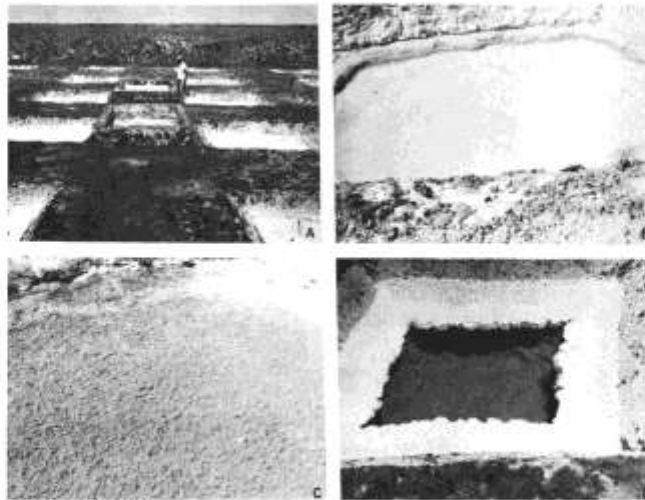
FIGURE 8—SEQUENCE OF CRYSTALLIZATION OF BRINE BY SOLAR EVAP-
ORATION AT EDRI; a) View of experimental plots; evaporation pans
shown in center; b) Supersaturated solution in experimental pit; c)
Crystal formation at the bottom of the pit (crystals were formed in
about 2 weeks); d) Salt crusts formed on top of the pit.

NEW MEXICO'S RESOURCES:
A MINERAL-DOMAIN POTPOURRI

by Robert W. Eveleth, *New Mexico Bureau of Mines and Mineral Resources, Socorro, New Mexico 87801*

Abstract

New Mexico has a rich cultural heritage dating back many centuries. This heritage, although historically interesting, has resulted in a multitude of land classifications and mineral ownerships. Lands in New Mexico are administered by the Bureau of Land Management and U.S. Forest Service, State of New Mexico, Indian tribes and trusts, grants, and other private owners. In addition, lands are withdrawn from mineral entry by several land classifications. Development and disposal of minerals and mineral materials on these various domains are similarly governed by a multitude of laws and regulations. Although mineral substances on State, Indian, and private lands are obtained exclusively by lease arrangement, those minerals on the public lands may be obtained under one or more Federal laws including the General Mining Law of 1872, Mineral Leasing Act of 1920, and Minerals Disposal Act of 1947. To further complicate the subject, some mineral substances such as limestone or sandstone, are generally classified as common rock varieties and are thus obtained under the Minerals Disposal Act and Multiple-Use Act of 1955. In certain cases, however, these substances may be locatable under the General Mining Law. Each of the various major domains and their laws and regulations are discussed.

Introduction

The purpose of this report is to introduce the reader to the State of New Mexico from the perspective of the prospector, mineral developer, and miner. New Mexico's rich cultural heritage dates back many centuries and has resulted in a variety of land ownership and administration. However, an in-depth study is beyond the scope of this presentation, and the interested person is directed to the list of references for additional information.

Historical background

New Mexico's present multitude of land classifications and associated mineral domains, perhaps more complex than any other state, originated with the signing of the Treaty of Guadalupe Hidalgo in 1848. Prior to that time, all lands within the present state boundaries were owned by either the Government of Mexico or the Republic of Texas. With the signing of the treaty came the first private lands: one stipulation of the treaty was official recognition by the United States Government of the many land grants issued by Spain prior to 1821 and by Mexico thereafter. All other lands within the territory became part of the public domain.

In 1850 and later in 1853, additional lands were added to the new territory through the Texas and Gadsden purchases and, with only a few minor changes, these lands make up the present State of New Mexico. Additional land classifications have appeared since that time, however. These include the various Federal, Indian, State, and other private lands.

The complexity of ownership and administration of these various lands is such that the modern-day prospector and mineral developer should become as familiar as possible with the different land classifications and their pertinent mineral laws and regulations before formulating any serious plans for mineral development.

Nearly 78 million acres of land lie within the present-day boundaries of New Mexico. More than 33% of this land is under Federal ownership and control while 9% is Indian land, 16.4% is State land, and over 41% is privately owned. All major land classifications and their respective owners/agencies are shown in table 1.

Federal lands

Lands under Federal ownership and administration total just under 26 million acres, most of which is included in the public domain. The mineral estate on these lands is administered by the Department of the Interior through the Bureau of Land Management. This includes the recordation and patenting of mining claims on lands within the national forests. Other Federally owned lands include National Parks, fish and wildlife refuges, and military reservations.

TABLE 1—NEW MEXICO LAND OWNERSHIP AND ADMINISTRATION; 1 Recordation and patenting of mining claims administered by BLM for USFS; 2 Previously Bureau of Reclamation; 3 Includes lands owned and administered by U.S. Air Force, NASA, DOE, and other federal agencies; 4 Includes patented mining claims, homesteads, and others; 5 Bureau of Land Management estimate; may include unconfirmed land grants; 6 Includes permanent bodies of water; 7 Fish & Wildlife lands differ in each case; thus, title to each must be examined; 8 Mineral entry, when allowed, only under leasing laws; A Hardrock minerals available by location under General Mining Law of 1872; B Fossil fuels and compounds of sodium, potassium, etc., available under Leasing Act of 1920; C Common varieties such as sand, gravel, and cinders available under Material Disposal Act of 1947 as amended by Multiple Surface Act of 1955;
D All minerals available under Acquired Lands Leasing Act of 1947 or Materials Disposal Act of 1947 as amended by Multiple-Use Act of 1955. However, reconveyed lands that may be included in this total are managed the same as public lands; E All minerals available by lease with Tribal Authority and/or BIA; F All minerals available by lease arranged with State Land Office; G All minerals obtained by private lease arrangement (in most cases; see discussion under Spanish land grants); W withdrawn from mineral entry.

Domain	Administration	Area (acres)	Total area (%)	Mineral access
Federal	BLM public lands	12,612,755.0	16.20	A, B, C
	acquired	227,680.6	.29	D
	BLM Total	12,840,455.6	16.49	-
	USFS public lands'	8,666,286.0	11.13	A, B, C
	acquired	577,327.9	.74	D
	USFS total	9,243,613.9	11.87	-
	U.S. Army	2,419,630.0	3.11	W
	Fish & Wildlife	316,818.8	.41	7
	National Park Service	244,093.3	.31	W
	Water and Power'	199,064.2	.26	A, B, C
	Corps. of Engineers	27,752.9	.03	8
	Other'	582,316.0	.75	W
	Total Federal	25,873,744.7	33.23	—
Indian	Tribal Authority & **BIA**	**6,965,203.0**	**8.95**	E
State	State Land Office	12,794,718.0	**16.43**	F
Private	Spanish land grants'	11,138,239.2	14.30	G
	Railroad grants	3,355,179.0	4.31	G
	Other°	17,739,155.9	22.78	G
	Total private	32,232,574.3	41.40	G
	Total area	77,866,240[6]	100.00	

Except for those land classifications closed to mineral entry, acquisition of mineral rights on the Federal lands falls under three basic statutes. These are the General Mining Law of 1872, the Minerals Leasing Act of 1920, and the Materials Disposal Act of 1947 as amended by the Multiple Surface Use Act of 1955. Because a significant portion of New Mexico's mineral resources is found on the public lands, a working knowledge of these laws is beneficial.

Public lands are those lands that have never left Federal ownership since original acquisition by treaty or purchase. Most mineral commodities such as barite, fluorspar, perlite, and the metallic minerals are locatable on these lands by mining claim under the 1872 law. When a deposit is found on unappropriated public domain, the discoverer obtains the right to possess and develop the deposit by staking and re-cording, in the proper manner, one or more mining claims, Prior to 1979, the task of determining which lands were unappropriated as opposed to those lands under claim could be quite difficult. Claim notices were filed only with the local county court house, and the information given on the notice was, in many cases, so ambiguous as to render exact location of the claim impossible. Since 1979, however, the mineral claimant has had to file, in addition to the notice at the court house, more specific information with the Bureau of Land Management regional office at Santa Fe. This information includes a metes and bounds description tied to either the public survey or a topographic feature and a map of the claim or claims with a scale of not less than 'A inch to the mile. Thus, the prospector may readily determine land under claim and unappropriated land by checking the filing records at Santa Fe.

Certain classes of mineral commodities are excluded from location under the General Mining Law of 1872. These minerals are fossil fuels, phosphates, compounds of sodium, potassium, and sulphur. Such minerals are acquired by lease only under the Minerals Leasing Act of 1920.

A third class of mineral substances, generally referred to as common varieties, includes stone, sand, gravel, cinders, and similar commodities that are obtainable under the Materials Disposal Act of 1947 as amended. Unlike locatable and leas-able minerals, disposal of common-variety substances may be administered by either the Bureau of Land Management or the U.S. Forest Service depending upon where the deposit is found.

Exceptions exist where otherwise common varieties may be locatable under the General Mining Law of 1872. According to regulations of the Department of the Interior (43 C.F.R. subpart 3710, sec. 3711.l, b), "Mineral materials which occur commonly shall not be deemed to be 'common varieties' if a particular deposit has distinct and special properties making it commercially valuable for use in a manufacturing, industrial, or processing operation." Examples of otherwise common varieties that have been declared locatable under the General Mining Law of 1872 are: 1) high-silica sandstone valuable as flux in a copper-smelting process and 2) high-calcium lime-stone for use in manufacture of cement.

The statutes listed above pertain to mineral acquisition on the public lands. These lands should not be confused with government-acquired lands that are often closely associated with or surrounded by public lands. By definition, acquired lands are those "in Federal ownership which are not public lands. Acquired lands were obtained by the Government through purchase, condemnation or gift, or by exchange" for same (U.S. Department of Interior, 1979). As such all minerals on acquired lands are reserved to the government and no locations are permitted under the General Mining Law of 1872. Mineral rights are obtained under the Acquired Lands Leasing Act of 1947 which is similar to the 1920 Leasing Act. In addition to public and acquired lands, other Federal land classifications may be of interest to the mineral developer.

Other Federal lands

During the early part of this century, certain tracts of federal land were classified as potential water-power sites and subsequently withdrawn from mineral entry. The Mining Claims Right Restoration Act of 1955 reopened these lands under the general mining laws. The Secretary of the Interior may set certain restrictions, however, up to and including denial of entry, specifically for placer-mining operations.

Game refuges are generally closed to mineral entry, but mineral status should be determined on a case-by-case basis as title to each refuge area differs. Entry may be allowed under the leasing laws in some cases. Approximately 2.4 million acres of military lands within the White Sands Missile Range are completely closed to mineral entry. Many Federal lands have been recently considered for inclusion into wilderness areas. The situation seems to change on a yearly basis. These areas may be closed or severely restricted to mineral entry. The Bureau of Land Management or the U.S. Forest Service should be consulted for up-to-date information.

Indian lands

Lands owned or held in trust for the various Indian tribes constitute nearly 9% of present-day New Mexico. These lands currently are owned by 24 tribes and cover approximately 7 million acres, located principally in the northwest portion of the state. The one major exception is the Mescalero Reservation in northern Otero County.

Acquisition of mineral rights on Indian lands is by lease only, regardless of commodity. Tribal lands are leased from the tribal authority, while allotted lands are leased from the individual allottee. The Bureau of Indian Affairs generally negotiates the lease and the U.S. Geological Survey acts in the capacity of technical advisor. Recently the various tribal authorities have taken a more active role in lease negotiations.

Indian lands are primarily noted for extensive deposits of coal, uranium, oil, and gas. These lands were originally thought to be non-mineral in character when granted. Only later was the extent of coal deposits known. Uranium was dis-covered relatively recently. Large tracts of land with no mineral potential whatever are, in fact, rare or non-existent; this will be further demonstrated when discussing state and private lands.

The extent and nature of other mineral deposits on these lands is not so well known. However, studies currently under-way by the U.S. Bureau of Mines, U.S. Geological Survey, Bureau of Indian Affairs, and New Mexico Bureau of Mines and Mineral Resources will investigate and assess mineral resources on the various Indian lands. This work may be con-tinued on the main reservations at a later date.

In addition to the mineral products previously mentioned, gypsum, sand, and gravel are among other commodities cur-rently produced from Indian lands. Leasing information can be obtained directly from the various tribal councils or the Bureau of Indian Affairs.

State lands

Lands owned and administered by the State of New Mexico currently total more than 13 million acres. The Ferguson Act of 1898 and, later, the Enabling Act of 1910 awarded sections 2, 16, 32, and 36 in each township to New Mexico for the pur-pose of generating revenue for the Common School Fund. The sections so granted had to be classed as non-mineral in character and be otherwise unappropriated. For the case of lands conflicting with the above, alternate or "in lieu" sections were selected. Selection of these "in lieu" lands has resulted in the formation of several large blocks of State land throughout New Mexico. Concentrations of State lands are

found, for example, in, much of Lea County, northern Luna County, and eastern Socorro County, among others.

Among many mineral deposits known to exist on State lands are gas, oil, coal, and potash; however, a much needed comprehensive study for other commodities has not been undertaken. The existence and extent of gas, oil, and other minerals, as with the Indian lands, was generally unknown at the time of selection; thus, these lands were classed as non-mineral in character. Significantly, few metalliferous and other lode deposits occur on State lands. Exceptions are a few barite and fluorspar occurrences, chiefly in the southern half of the state.

Though State lands are sometimes called public lands, no locations are permitted under the general mining laws. Acquisition of mineral rights is by lease only. Applications can be obtained from the Commissioner of Public Lands in Santa Fe.

Private lands

Taken as a whole, privately held lands comprise approximately 32 million acres (41% of total land), the largest single land classification in the state. However, as with the Federal lands, several categories exist. These include the Spanish and Mexican land grants, railroad grants, university grants, and other private lands.

The Spanish and Mexican land grants, as previously mentioned, were the first private lands to appear in New Mexico. Today, about 147 confirmed grants contain approximately 11 million acres (Office of Cadastral Engineer, personal communication, 1981) or just over 14% of the total state area. The validity of other grants may still be pending in the future. Most prominent of the confirmed land grants are the Maxwell grant in Colfax County, the Tierra Amarilla grant in Rio Arriba County, the Mora and Las Vegas grants in Mora and San Miguel Counties, and the Pablo Montoya grant in eastern San Miguel County.

Mineral rights on the various grants are generally acquired through a privately negotiated lease with the grant owner. Exceptions to this lease procedure are covered by a Congressional Act passed in 1926, in which all deposits of gold, silver, and quicksilver (mercury) are reserved to the United States Government except for those cases where such mineral rights were specifically conveyed to the grantee by the terms of the original grant. Presumably all other mineral commodities are owned in fee by the grantee, his heirs or assigns. For those grants where the above minerals were reserved to the government, the Secretary of the Interior may assign a lease, but only to the grant owner.

Railroad grants constitute another substantial group of privately owned lands. These grants also appeared quite early in New Mexico history when the United States Government granted sizeable tracts to some of the early railroads to aid in construction of their lines. Though the lands so granted were supposed to be non-mineral in character, rights to any unknown deposits were acquired when patent was issued.

These grants eventually amounted to approximately 3.3 million acres. Much of the surface estate of these grants has since been sold; however, the railroad company has retained most of the mineral rights. Here, as with other private lands, the prospector should examine the ownership plats for each tract to determine which minerals, if any, were reserved to the government. Mineral acquisition is through privately negotiated lease for those minerals retained by a railroad.

The largest group of private lands in New Mexico is that for which title has passed from Federal to individual ownership through various laws and statutes such as the early Homestead Acts and, to a lesser degree, mining claims patented under the 1872 Mining Law. Lands acquired under the Homestead Acts account for most of the private holdings, many of which are concentrated in the eastern part of the state. Only after 1909 did the United States Government begin reserving minerals in some of these patents. In 1916, the United States Government reserved all minerals in stock-raising or enlarged homesteads.

Acquisition of mineral rights on patented homesteads and mining claims is, like other private lands, through private lease with the owner. Titles should be checked carefully beforehand to determine reservations, if any, to the government. Private leases may be difficult to obtain; many individuals are inclined to demand impossible royalty rates.

Conclusions

The foregoing discussion serves to demonstrate the complexity of New Mexico's numerous land classifications. The complexities can, however, be sorted through smoothly. Toward this end the following general procedure is suggested to the would-be prospector:

1) determine areas favorable for desired mineral substances through literature search and/or field examination, if possible;
2) determine land status and mineral ownership by examination of township plats and other records at Bureau of Land Management in Santa Fe; and
3) proceed to acquire mineral rights under appropriate Federal or state statute or by private lease.

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MICA RESOURCES OF THE WESTERN UNITED STATES

by Richard J. Beckman, *U.S. Gypsum Company, Chicago, Illinois 60606*

Abstract

An examination of the sources of muscovite mica west of the Mississippi River reveals that although this ubiquitous mineral occurs in many deposits throughout the west, only one company is currently producing mica in significant amounts to serve the construction industry in the western marketplace. Muscovite occurring in hard-rock pegmatites is not a commercial source of scrap mica if mined only for that mineral because it makes up only 5-10% of the rock. Quartz-muscovite schists derived from the alteration of rhyolitic rocks are excellent sources of clean muscovite, but are not always completely converted from the parent material. Other schist and sericite deposits that have been exploited have never been clean enough to meet the specifications for ground-mica filler applications and therefore cannot compete with eastern micas. A review of the important mica districts in the west and the problems encountered in developing an economic deposit are discussed.

Introduction

Mica, as defined here, refers to the light-colored or color-less muscovite quarried directly from pegmatite or mica-schist deposits and concentrated as scrap or flake mica. Mica is used in a variety of industrial applications, generally in the ground form, and must meet fairly rigid specifications as to color, sizing, purity, and bulk specific gravity. Commercially, muscovite mica is divided into two broad classifications: sheet mica and scrap mica. The end-use applications for both types differ widely.

Sheet mica consists of coarse (>1 inch) flat plates that can be trimmed, punched, and stamped into various specified shapes for industrial use, mainly in the electrical industry. Sheet mica is found only in pegmatites, is mined and sold in pounds, not tons, and represents only about 1% of all mica production. Scrap mica, on the other hand, is the mainstay of the mica industry. In a broad sense, scrap mica is any mica that is below the specifications for sheet mica because of size, color, or quality. It is generally muscovite that is found as loose flakes, Y2 inch to 200 mesh in size and, after concentration, is ground to specified fineness. Scrap and/or ground mica is sold in tons and has widespread application as a filler and surface coating in numerous products, including roofing materials, joint cements, paints, rubber, plastics, and drilling muds.

The production and distribution of ground mica is shown in fig. 1, which also forecasts the various end uses in the year 2000. The major use of ground mica is in joint cement, a product used exclusively in the construction industry. The total demand for mica by the year 2000 is forecast to be 223,000 tons, or about 100,000 tons more than in 1981 (Zlobik, 1970). Coupled with the fact that only one company is presently mining and processing mica in significant amounts west of the Mississippi River, a reasonable assumption is that a mica-supply problem will exist in the western United States when demand increases during the expected resurgence of construction activity in the mid-1980's.

Structure of the industry

Presently, 15 companies are involved in mica production and/or grinding of muscovite mica. Only seven companies mine mica as a primary product. All other mica is mined as a by-product from feldspar and spodumene operations and is
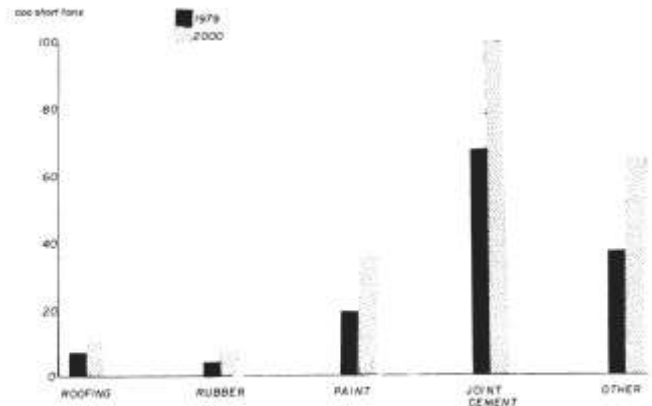
FIGURE 1—PRODUCTION AND FORECASTS FOR UNITED STATES GROUND MICA END USE.

sold to mica grinders. Therefore, much of the mica production is a direct function of feldspar and spodumene mining and is dictated by demands for those two minerals. As a result, these by-product producers cannot increase or decrease their mica production to meet fluctuations in the demand for mica.

Mica operations are concentrated in the east, mainly in North Carolina. Only two are west of the Mississippi; one is in South Dakota and the other is in New Mexico. Production levels of scrap mica have been in the range of about 125,000 tons per year over the past several years with about 15% accounted for west of the Mississippi. North Carolina produces approximately 65% of the total while the remaining 20% is produced in Connecticut, Georgia, and South Carolina (fig. 2).

Sources of supply

Deposits of muscovite mica in the United States occur in three distinct geologic environments:

1) Granitic pegmatites and alaskites (igneous) (fig. 3),
2) Quartz-muscovite schists (metamorphic), and
3) Kaolinized or weathered pegmatites and alaskites (residual).

In the western states, mining of scrap mica has been an intermittent activity and began as an outgrowth of mining com-

FIGURE 2—MAJOR MICA PRODUCTION IN THE UNITED STATES, 1979.

FIGURE 3—MAJOR PEGMATITE AREAS OF THE UNITED STATES.

mercial sheet mica from pegmatites. As demand for ground mica increased in the 1920's, there was renewed mining of pegmatites for muscovite. These hard-rock pegmatites were tunneled and mined along narrow, tortuous workings until the mica played out or water and poor ventilation prevented further mining. In modern-day practice, bulk operations aimed at working entire pegmatite structures for their mica alone would not be a profitable venture. The primary minerals of pegmatites are quartz and feldspar and these would have to be mined also to extract the mica. Mica rarely accounts for more than 5% of the rock even though, from outward appearance, pegmatites often seem to be saturated with mica flakes due to the high reflectance given off by muscovite. Unless the mica is being processed as a by-product from mining feldspar and/or other valuable minerals (spodumene, beryl, etc.), a pegmatite is only a geologic curiosity or interesting feature, not a commercial mica deposit.

In the east, particularly in North Carolina, mica is mined from large pegmatite and alaskite bodies as a by-product from feldspar, kaolin, and spodumene operations. The mica is removed by flotation methods as a step in the concentration of the feldspar and spodumene ores. This by-product mica represents about 50% of the total production of scrap mica in the United States. The other source of mica in the east comes from residual deposits resulting from the weathering of pegmatites and alaskites. Although containing only 10-12% mica, these soft, kaolinized deposits can be worked for mica alone because no blasting or crushing is required. The ore is easily quarried with scrapers and dozers and the mixture of sand, clay, and mica is treated in a mill process where the mica is removed by flotation techniques.

The western states are not so endowed with soft, weathered mica deposits nor do they have large by-product operations where mica might be obtained. Instead, the exploitation of mica has centered around the metamorphic mica schists, rocks that contain a much greater volume of muscovite than do pegmatites. The principal problem is that the mica is much finer grained and is intimately mixed with other minerals, making separation of the mica difficult. Most schist deposits also contain biotite, the black mica that is practically inseparable from muscovite in flotation practices. The consequence of mining mica-schist deposits is that they cannot compete with the higher quality eastern micas and, therefore, most such ventures have failed.

Western mines and prospects

CALIFORNIA

Although muscovite mica occurs in pegmatites in California, none has become a profitable source of any commercial grade. Even the most micaceous parts of the pegmatites generally average only a few percent muscovite and therefore are not profitable to mine. Various attempts have been made to mine mica that occurs in schist deposits, but none has ever been proved successful over the long term.

Mica schist has been quarried near Ogilby in Imperial County from the so-called "Micatalc" deposit. This inactive property at one time supplied a mica product that was used as a lubricating dust for the surfaces of rolled roofing. It was not a ground product. The mica was obtained from several discontinuous lenses as much as 100 ft thick and several hundred ft long. Rock was crushed and screened to liberate the mica, but no attempt was made to upgrade the ore by flotation or air separation.

Mica schist has also been quarried from Hodge in San Bernardino County from a series of steeply dipping lenses of quartz-muscovite schists derived from the alteration of metavolcanic rock. High costs and increasing overburden limited activity at this deposit to a few thousand tons.

In Mariposa County, a muscovite-quartz schist has been mined in the White Rock district near LeGrand by the Kelly-Moore Paint Company. In fact, this was the last major effort to mine and beneficiate mica in the state. Kelly-Moore built a modern flotation mill in 1967 and for a very short time was operating, but it soon became apparent that the ore was not suitable for the quality they desired and the plant was shut down.

Sericite-rich materials are presently mined in the Victorville-Barstow area of San Bernardino County and in the White Mountains of Mono County. Sericite is an exceedingly fine-grained mixture of nearly inseparable muscovite and quartz. These mixtures do not meet the specifications for most ground-mica uses, but are used in some filler applications.

ARIZONA

Mica is widely distributed in pegmatites that occur sporadically in a belt 250 mi long and 30-80 mi wide, extending southeastward from Lake Mead through parts of Mohave, Yavapai, Yuma, and Maricopa Counties to points south of Phoenix. Many attempts have been made at mining these pegmatites for mica, but all have been short lived. What has been said previously about mining pegmatites solely for mica applies to all these Arizona misadventures. The erratic mica distribution in a pegmatite structure has doomed many investors in ill-advised mining commitments.

The only property that has produced a significant amount of mica is the Buckeye sericite deposit where a series of narrow, highly micaceous schist lenses were quarried for several years. The discontinuity of these veins and the steep dips associated with their occurrences prohibited large-scale, open-pit mining. Instead, many small pits and some underground workings were developed and supplied the mill with a material that was screened, bagged and marketed in the Los Angeles area. The product served as a partial substitute filler in joint compounds and also as an anti-sticking agent in rolled-roofing products.

Some fine-grained sericite was mined from a copper property near Tombstone where it occurs as narrow veins associated with copper mineralization. The mica was to be concentrated by differential settling in large vats, but this project never really got started and was largely promotional.

COLORADO

Enormous unexploited reserves of mica are stated to be available in Colorado. However, past and present production, investigations, surveys, and experience of private operators have failed to indicate any such reserves. Many hundreds of pegmatite deposits and other mica-bearing rocks occur in the

Precambrian areas of the state, but most are too small, of poor quality, or too isolated to be of commercial significance.

The only known deposit considered to have economic potential is a quartz-muscovite schist located near Salida in Chaffee County. The schist is present as a nearly vertical lens approximately 1 mi long, varying between 300 and 600 ft wide. It is extremely white and contains an average of 25% mica with sizing between 30 and 100 mesh. Although numerous, thin, black diabase dikes cut the deposit, it is still considered a good prospect. The greatest drawback in the development of the deposit is its location in a National Forest which places it in a category subject to public and federal scrutiny.

NEW MEXICO

In north-central New Mexico are two major areas of mica occurrence: the pegmatites of the Petaca region in Rio Arriba County and the quartz-muscovite schists in the Picuris Range in Taos County.

The Petaca region has long been known for its concentration of pegmatites and the mining of sheet mica dates back to the early 1700's. Scrap mica has also been mined, but the total production probably does not exceed 25,000 tons (Jahns, 1946). Probably more scrap mica is found on the waste dumps of many of these old workings today than was ever mined and sold for scrap and sheet when the mines were active. The last effort to mine scrap was in the 1950's when a large processing mill was constructed near Petaca. This mill was the fabrication of several promoters and had the monetary backing of the Catholic Church, but the complexities of mining-claim ownership coupled with high manufacturing costs in the mill forced the closing of the plant within a few years.

More recently, an attempt was made to mine a sericite deposit near La Madera in Rio Arriba County. This highly promotional scheme did not succeed because the low-grade material quarried did not yield the quantity of mica originally anticipated nor did the processing system meet expectations; the project was abandoned with only a scar on the mountainside to confirm its existence.

The most successful operation to date west of the Mississippi is the M.I.C.A. (Mineral Industrial Commodities of America) mica operation located in the Picuris Range near Taos. Mica is mined from a quartz-muscovite schist that occurs with a series of metavolcanic rocks of Precambrian age. This is an area where a high degree of metamorphism is evident. The zone between the metavolcanics and the overlying quartzite shows evidence of intense deformation, and recrystallization has produced a micaceous ore zone along the contact. M.I.C.A. mines this zone, some 60-100 ft wide, and uses only a dozer and front-end loader to dig the material. The soft material, a mixture of mica (muscovite) and quartz, is screened to eliminate much of the coarse quartz, thus upgrading the mica for final grinding. This material competes with eastern micas and, although its bulk density is slightly higher, it enjoys a distinct freight advantage to west-coast markets.

Recently, Gresens and Stensrud (1974a) have completed studies of the geochemistry and petrology of the quartz-muscovite schists in north-central New Mexico. Their conclusion is that these rocks are derived from rhyolite by metasomatic recrystallization. The conversion is most pronounced along shear zones and results in a rock containing about 38% muscovite and 60% quartz. This is typically true of the M.I.C.A. deposit. Of over 200 samples taken from core tests of the deposit, the average muscovite content was found to be 38%. Alteration requires the physical and chemical breakdown of feldspar in the host rock (rhyolite) which releases alumina needed to form muscovite. No biotite is created. Quartz is stable and is preserved, thus surviving the recrystallization process and offering important evidence for the

recognition of preexisting rhyolite. A number of places in both the Picuris Range and the Petaca pegmatite area manifest this type of recrystallization. The alteration is not always complete, however, resulting in a rock type described as "slabby sugary schist" by the authors. These schists contain abundant feldspar remnants of the original rock and are not enriched with muscovite due to the incomplete conversion. The recognition of feldspar remnants is further documentation that the quartz-muscovite schists were derived from rhyolites.

Lending heed to these observations, other areas where volcanic rocks have been altered may be potential sites of enriched muscovite concentrations similar to the M.I.C.A. commercial operation.

SOUTH DAKOTA

Pegmatites are distributed over an area of about 304 mil in the Black Hills and number well over 1,500. In all, about 30,000 tons of sheet and scrap mica have been mined from this district since the early 1900's, a small amount compared to present yearly production figures. At present, the only producer is Pacer Corporation, which processes mica as a byproduct of its feldspar operation and produces less than 1,000 tons of muscovite per year. Recently, a company called Concepts West, Inc., announced plans to mine and process mica from a series of pegmatites in the Custer area, including the old Crown mine, a pegmatite that once produced sheet mica.

More pegmatites probably are concentrated in the Black Hills of South Dakota than anywhere else in the United States. The potential for establishing a scrap and ground-mica operation is not good for reasons mentioned previously regarding mica in pegmatites, and because in most cases the cost of shipment to mica markets exceeds the value of the ground mica.

IDAHO, MONTANA, AND WYOMING

Pegmatites are common in several places in these states and in times past some were mined for sheet mica. None, however, have been important scrap producers, nor will they be in the future because of their remote location from established markets and the non-uniform mica distribution typical of mica-bearing pegmatites.

Conclusions

Muscovite is an important raw material that is used in a number of filler applications, particularly in the home-building industry. The anticipated growth in the Sunbelt states of the west will create a greater demand for ground mica; at present, only one company is mining a mica deposit of significant size and quality in the west to compete with the eastern micas. The development of any new deposits will require an extensive search and should be concentrated in areas of altered volcanic rock. Recrystallization and chemical alteration associated with rocks of rhyolitic composition result in light-colored mica schists enriched in muscovite. These rocks offer much greater potential than do mica-bearing pegmatites.

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NEW MEXICO LIMESTONES-GEOLOGY, USES, AND ECONOMIC POTENTIAL OF HIGH-CALCIUM LIMESTONE

by W. T. Siemers, *Phillips Petroleum, Bartlesville, Oklahoma 74003*

Abstract

Limestone occurs over central New Mexico from Colorado to the Mexican border. The deposits range from early Paleozoic to Quaternary in age, with most high-calcium limestones occurring in formations of Mississippian, Pennsylvanian, and Permian age. Locally, Jurassic and Lower Cretaceous limestones in northwest New Mexico and Cenozoic caliche and travertine deposits in central New Mexico are also high in calcium. Most of the high-calcium limestones are shallow to deep marine, lacustrine, or spring deposits, with some exceeding 30 m (locally even 300 m) in thickness. Calcium carbonate contents as high as 99.8% are reported for some limestones, but most units show considerable lateral and vertical variability in composition.

Nearly all high-calcium limestone produced in the state is consumed in the manufacture of lime and cement. Since 1956, New Mexico has produced about 526,000 tons of lime valued at more than $5.5 million. Two currently active lime operations (Hurley Lime Quarry, Kennecott Copper Corporation, and Mathis Lime Quarry, Mathis & Mathis Mining and Exploration) have a combined daily capacity of about 500 tons. Ideal Basic Industries' Tijeras Plant, the only cement plant in New Mexico, has an estimated annual capacity of about 518,000 tons.

Reserves of high-calcium limestone in New Mexico are difficult to estimate. Widespread carboniferous units contain abundant limestone resources; thus, high-calcium limestone reserves are believed to be substantial. Nevertheless, the proportion of these limestones that are indeed high-calcium remains to be determined.

Introduction

Limestone, a sedimentary rock in which the primary component is calcium carbonate, is used for both its chemical and physical properties in industries as diverse as construction, agriculture, and manufacturing. Low-calcium limestone, which is used primarily for its physical properties, is a valuable source of aggregate and dimension stone. If such limestones contain appropriate amounts of silica, clay, and iron oxide, they may also be used directly in the manufacture of portland cement.

High-calcium limestone, defined as limestone containing 95% or more calcium carbonate, is a primary source of lime (calcined limestone—CaO). These limestones are important in manufacturing such materials as cement, paper, glass, alkalies, and calcium carbide. They are also used as a metallurgical flux.

The geology and economic aspects of major limestone-bearing units in New Mexico are reviewed here. In this review the work of many others has been drawn upon; especially important in this respect is Kottlowski's (1962) survey of high-calcium limestones in New Mexico and Jicha's (1956) observations regarding the Mesa del Oro travertine deposit.

Production

Most of the limestone produced in New Mexico is used in the production of lime and cement. Production for most of the years between 1955 and 1974 ranged between 200,000 and 300,000 tons per year. In the middle and late 1970's production increased to between 400,000 and 500,000 tons annually. Presently, three operations are active in the state. The largest of these, the Tijeras Cement Plant operated by Ideal Basic Industries, is located east of Albuquerque in Tijeras. The annual production of cement from this plant is believed to be about 518,000 tons.

Two smaller lime-producing operations are located in the southwest part of the state near Silver City. The Hurley Lime operation, operated by Kennecott Copper Corporation, and the Mathis Lime operation, operated by Mathis & Mathis Mining and Exploration, have a combined daily production capacity of about 500 tons.

Resources

Abundant limestone resources occur in limestone-bearing units that crop out in central New Mexico from Colorado to the Mexican border (fig. 1). Those most likely to be high in calcium are generally restricted to late Paleozoic (Mississippian through Permian), shallow marine limestones and Cenozoic travertines. Although the proportion of these that are high calcium remains uncertain, reserves of high-calcium limestone in New Mexico are believed to be substantial.

NORTHERN NEW MEXICO

In northwest New Mexico, limestones occur in the Pennsylvanian, Permian, Jurassic, and Cretaceous Systems. The quality of most, however, is poor. Some limestones in the San
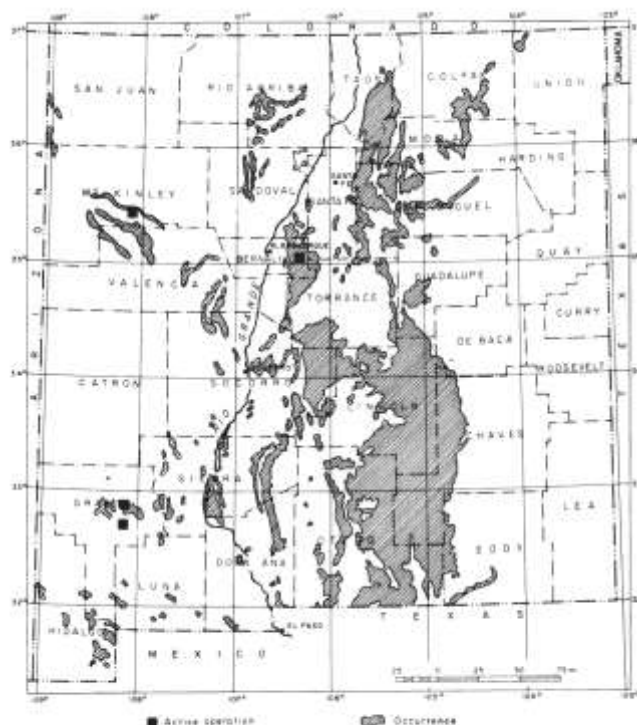
FIGURE 1—LOCATION MAP OF NEW MEXICO LIMESTONE OCCURRENCES AND ACTIVE LIMESTONE OPERATIONS.

Andres (Permian) may be locally high in calcium, but most are cherty, arenaceous, or dolomitic. Limestones occurring in the Todilto Formation are locally relatively thick (5-15 m) and have calcium-carbonate contents as high as 94.3% (Kottlowski, 1962). The upper part of the unit, however, is usually gypsiferous and the lower part of the unit is sandy. The limestones are typically laminated, carbonaceous calcarenites with silt partings.

In north-central New Mexico, limestones that occur within the Mississippian and Pennsylvanian are often silty and cherty. Locally, some crinoidal limestones in the Cowles Member of the Tererro Formation (Mississippian) are as much as 15 m thick, do not appear to be dolomitic, and contain less than 1% insoluble residues (Kottlowski, 1962). Some thick, biohermal limestones in the overlying Madera Limestone (Pennsylvanian) are relatively noncherty and yield insoluble residue contents of less than 1%. Permian limestones are commonly silty, dolomitic, and gypsiferous. Limestones of the Todilto Limestone (Jurassic) are thin and impure, and the dark-colored limestones in the Mancos Shale (Cretaceous) are very argillaceous.

In northeast New Mexico, impure limestones occur in the Todilto, Greenhorn, and Fort Hayes Limestones and as the caprock caliche of the Ogallala Formation. Caliches are used locally for crushed stone and road metal, but do not appear to have any potential for use as high-calcium limestone.

SOUTHERN NEW MEXICO

Limestones occur throughout the Paleozoic section in southwest New Mexico. Lower Paleozoic (pre-Mississippian) beds, however, are typically silty or sandy and dolomitic. Kottlowski (1962), nonetheless, suggested that a concentrated search of the El Paso Group (Ordovician) may discover some high-purity limestone in that unit.

Upper Paleozoic limestones are more promising. Limestone units in the Lake Valley, Kelly, and Escabrosa formations (Mississippian) are thick (40-175 m) and include many beds of high-calcium grade. The Pennsylvanian section is 250-750 m thick in the region and is composed chiefly of limestone. Most of the larger Pennsylvanian outcrop areas contain high-purity limestone.

Some Wolfcampian limestones in the Hueco Formation and the Horquilla and Colina Limestones are high in calcium, whereas the thin, lenticular, blocky limestones that occur in the Cretaceous (Mancos Shale) are usually argillaceous. Some high-calcium limestone also occurs in the Palm Park Formation (Tertiary) in the more remote Caballo Mountains.

In south-central New Mexico, pre-Mississippian limestones are impure and high in magnesium. All of the Mississippian rock formations contain many thick limestone sections, but only the Lake Valley Limestone is high in calcium (Kottlowski, 1962). Thick, relatively pure limestones also occur in the lower Pennsylvanian in this region. Middle and upper Pennsylvanian limestones, however, tend to be cherty and contain considerable terrigenous material. Most Permian units contain limestones that are high in calcium, with the exception of the San Andres, which is typically very dolomitic.

In southeast New Mexico, some high-calcium limestone occurs in the Permian but, for the most part, these limestones are very dolomitic, gypsiferous, or silty. Thick, high-calcium limestones occur in the core of the Capitan reef. The reef, however, has a complex dolomitization history and careful, detailed sampling is required to assess its true potential. The caliche of the Llano Estacado contains local deposits of high-calcium limestone, but most of the caliche in the region is high in insoluble residues and is suitable only for use as road metal.

CENTRAL NEW MEXICO

Limestones occur in beds of Mississippian through Permian, late Cretaceous, and Cenozoic age that crop out in cen

tral New Mexico. Little economic potential is associated with the thin, argillaceous carbonates that occur in the Mancos Shale (Cretaceous). The Kelly Formation (Mississippian) and some Cenozoic traventines hold some promise of being locally high in calcium. The most likely sources of high-calcium limestone in central New Mexico are the Madera Limestone (Pennsylvanian) and, on a more limited scale, the San Andres Formation (Permian).

KELLY FORMATION—In central New Mexico the Kelly Limestone ranges up to about 25 m in thickness. The unit thins to the north and eventually wedges out about 40 km north of Socorro. The absence of marked lithologic or faunal changes near the wedge-out suggests that this area was not close to the northern strandline. The Kelly Formation probably extended many kilometers northward, with the pinchout of the formation north of Socorro resulting from pre-Atokan erosion. The relatively large brachiopod, and endothyroid fauna yielded by the typical Kelly Limestone indicates affinities with the Keokuk Limestone of late Osage age (Armstrong, 1958).

The Kelly is bounded by the Caloso Formation below and the Sandia Formation above. The unconformable Caloso-Kelly contact is a gently undulating surface with 5-10 cm of clean, disseminated sand filling lows developed on the top of the Caloso. The unconformity between the Kelly Formation and overlying Atokan beds is characterized by as much as 10 m of relief. Broad, shallow, pre-Atokan channels developed on top of the Kelly are filled with shale and chert pebbles derived from the Kelly Formation. Dark-colored shales and cross-bedded, coarse-grained sandstones mark the base of the Sandia Formation.

The Kelly Formation is typically a medium- to thick-bedded, medium-gray, medium-crystalline crinoidal limestone in the lower part of the section. The upper 3-5 m of the unit are more thinly bedded and somewhat finer grained. The Kelly typically consists of massive crinoidal grainstones and packstones cemented by coarsely crystalline, equant calcite spar. Such limestones reflect deposition on a broad, open, well-circulated, shallow marine shelf.

The Kelly Formation crops out in the Magdalena, Lemitar, and Ladron Mountains. Because of the presence of much chert and the inaccessibility of these sections, however, the economic potential of the Kelly in these areas is almost nil. Mississippian rocks in the Sandia Mountains (Arroyo Penasco Formation) are 12-23 m thick and contain 91.0% calcium carbonate (Kottlowski, 1962).

MADERA LIMESTONE—The thickness of the Madera Limestone ranges from 64 to 716 m in central New Mexico. The formation is thicker along a north-south-trending axis through the southern Ladron, Magdalena, and southeast San Mateo Mountains. The Madera eventually wedges out toward the east against the flanks of the Pennsylvanian Pedernal uplift.

The Madera Limestone overlies the more terrigenous Sandia Formation. The contact is gradational and somewhat arbitrary, but is commonly placed at the top of a prominent sandstone unit below which terrigenous lithologies dominate and above which limestone dominates.

Although the Madera ranges from Atokan to Virgilian in age, throughout much of central New Mexico the lowermost Madera beds are no older than Desmoinesian. Apparently a hiatus exists between the Atokan and Desmoinesian; any interruption in sedimentation, however, probably involved a relatively short time interval with only minor, local diastrophism. The Madera is disconformably overlain by Permian red beds.

Diverse carbonate and terrigenous lithologies comprise the Madera Limestone in central New Mexico. The mean clastic ratio of the unit (0.45) illustrates the slight dominance of limestone. Sandstone seldom constitutes more than about 5%

of the section. Desmoinesian beds are as much as 340 m thick in the region and are characterized by clastic ratios ranging from 0 to 0.94. Upper Pennsylvanian beds are as much as 341 m thick and are more terrigenous. On this basis, the Madera has been divided into a lower gray limestone member and an upper arkosic limestone member. The Missourian-Virgilian biostratigraphic boundary coincides closely with this lithostratigraphic boundary.

Carbonate intervals in most Madera sections typically range between 5 and 15 m in thickness. In the Ladron, Lemitar, and southern Manzano Mountains, however, limestone intervals as thick as 140 m are virtually uninterrupted except for a few thin calcareous shales. The limestones are usually gray, with darker hues predominating. Throughout most of the region, micritic mudstones are the predominant carbonate lithology in the Madera. Brownish, subtranslucent, microcrystalline calcite typically constitutes from 75% to more than 90% of these limestones. Bioclastic material (mostly crinoidal, algal, and brachiopod debris), minor peloidal material, and occasional clay- to silt-sized terrigenous material comprise the remainder of the rock. Some beds may become quite carbonaceous.

Silicification of Madera limestones is variable, but tends to be concentrated in the more thickly bedded, dark-gray limestones in the Desmoinesian section. The silica occurs as disseminated blebs, as nodules or lenses, and as small pods of chalcedony. Inclusion of fossil material within many nodules suggests that they are probably secondary replacement features.

Madera limestones were developed in a rather long, narrow, north-trending extension of the Orogrande Basin. This so-called Magdelena prong of the Orogrande Basin (Siemers, 1978) was bounded on the east by the Pedernal uplift and on the west by the Zuni-Defiance highlands. Most of the dark, cherty, Desmoinesian limestones were deposited under normal outer-marine-shelf conditions. Missourian and Virgilian deposition occurred under more diverse conditions, ranging from nearshore, possibly even coastal, conditions along the flanks of the Pedernal uplift, to outer-marine-shelf conditions near the axis of the basin.

Some of the most promising potential high-calcium limestone deposits in New Mexico occur within the Madera Limestone. Calcium-carbonate contents exceeding 98% have been documented at many localities. In many areas, these limestones underlie the dip slopes of hogbacks and mesas where they could be quarried from open pits with the removal of only a thin overburden. The most attractive deposits occur on the flanks of the eastern dip slope of the Sandia-Manzano Mountains where calcium-carbonate values from upper Madera limestones range between 95 and 98.5%. Desmoinesian beds contain 4-10% insoluble residues (mostly silica); Missourian and Virgilian limestones commonly contain less than 5% insoluble residues (Kottlowski, 1962). The area is also attractive because it is accessible and close to Albuquerque, the state's population center.

Very pure Upper Pennsylvanian limestones also occur along the eastern dip slope of the northern Oscura Mountains. The Council Springs unit, for example, has shown 98.4% calcium carbonate, 0.5% magnesium carbonate, and 0.5% silica. Although the area is reasonably accessible, it is somewhat remote.

Extensive outcrops of Madera limestones also occur in the Socorro region. In the Cerros de Amado, east of Socorro, Missourian and Virgilian limestones are characterized by calcium-carbonate contents of 96-98%, silica contents of l.02.5%, and magnesium-carbonate contents of less than 1.0%. Thick sections of Pennsylvanian limestone in the Ladron Mountains yield calcium-carbonate values of about 97%, but the outcrops are extremely remote and can be reached only over long stretches of poorly maintained ranch roads.

SAN ANDRES FORMATION—The San Andres Formation is Leonardian to Guadalupian in age. The thickness of the formation within central New Mexico ranges from approximately 65 to 230 m. Typical limestones in the San Andres are light- to dark-gray, dolomitic mudstones. Massive, occasionally laminated, medium to very thick beds characterize the formation. The abundance of gas-escape features (for example, birdseye structures) suggests that deposition occurred under nearshore, possibly intertidal or supratidal, conditions.

Locally, the economic potential of San Andres limestones may be quite high. Throughout much of central New Mexico, however, the beds are dolomitic, gypsiferous, or high in siliciclastics. East of Socorro, many high mesas are capped by thick San Andres limestones. One such outcrop has been quarried north of the ghost town of Carthage. Several samples from the quarry have yielded calcium-carbonate contents near 98%. The greatest potential for high-purity limestone in the San Andres, however, probably occurs farther to the south in the Sacramento and San Andres Mountains of south-central New Mexico.

CENOZOIC TRAVERTINES—Extensive deposits of travertine occur on the western flanks of the Ladron Mountains and on the eastern and western margins of Lucero Mesa. West of the Ladron Mountains, the Riley travertine occurs along a belt about 5 km wide and 15 km long. The so-called travertine is more likely a lacustrine or playa limestone (Kottlowski, 1962). The unit has yielded calcium-carbonate values as large as 99%, but the deposit is very remote and nearly inaccessible.

The travertines in the Lucero Mesa area are much smaller in size; in most places, they cover a belt that is less than 1 km wide. Some of these spring deposits form bench-capping beds that are erratic in thickness and composition; others are nearly vertical veins along joints and faults. The travertines are typically rather impure in that they commonly yield more than 10% insoluble residues and only about 85% calcium carbonate (Kottlowski, 1962).

The Mesa del Oro travertine deposit (Jicha, 1956) is economically more promising. This spring deposit is approximately 2 km wide and nearly 10 km long. Its thickness is somewhat variable, but averages about 46 m. The body is thick in its northern and eastern parts, but thins sharply to the west and more gradually to the south. Petrographic analyses show that the travertine consists of 1-2% moderately sorted, angular grains of very fine, sand-sized, monocrystalline quartz, and small algal pisolites in a groundmass of very finely crystalline calcite. The deposit is reasonably accessible, fairly large (in excess of 100 million tons), and reasonably pure (about 97.5% calcium carbonate and less than 2% insoluble residues).

Outlook

New Mexico's high-calcium-limestone resources are substantial. Because of expansion of the state's construction and coal industries, increased development of the state's limestone resources appears likely. A recent report (Robinson & Washington Pathfinders, Inc., 1979) suggests the cement market in the Southwest would presently support another cement plant. Much of the demand, however, will come from the state's northern area where the coal industry is expanding rapidly. Such an increased demand for high-calcium limestone in northern New Mexico will be difficult to satisfy, for most of the state's high-purity limestones occur in the distant central and south-central areas.

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GYPSUM RESOURCES OF NEW MEXICO

by Mark J. Logsdon, *New Mexico Bureau of Mines & Mineral Resources, Socorro, New Mexico 87801*

Abstract

Gypsum deposits of potential commercial value occur in north-central, south-central, and southeast New Mexico. Thick, relatively pure gypsum beds are located within 120 km of the major population centers of the Rio Grande valley over a distance of more than 400 km; somewhat thinner gypsum beds crop out near the Pecos River in southeast New Mexico. Extensive gypsum dune beds occur in the Tularosa Basin and the Estancia Valley. Railroads and major state and federal highways provide access to or near the major deposits. Three active or recently active gypsum mines, all in the Todilto Formation (Jurassic), have a combined annual capacity of approximately 240,000 metric tons. Producing gypsum beds are generally 12-30 m thick and crop out as massive gypsum with little or no overburden. All of New Mexico's 1980 production of gypsum was used for construction materials. Resources of high-purity gypsum in New Mexico are in the billions of tons. As population and industry continue to move to the Sunbelt states of the Southwest, demand for high-purity New Mexico gypsum and markets for gypsum products are expected to expand.

Introduction

In the United States, gypsum is used primarily in the manufacture of construction materials. Of the 18.3 million metric tons (all figures given in metric tons; 1 metric ton = 1×10^6 g) of crude gypsum consumed in the U.S. in 1980, 67% was used to manufacture wallboard, 19% was used as retarder in portland cement, and 4% was used to produce building and industrial plaster. Almost all of the remaining 10% was used in agriculture as a soil conditioner, with only a fraction of 1% consumed in miscellaneous industrial applications. All of New Mexico's 1980 production of 189,000 metric tons was used to produce construction materials: 91% for gypsum wallboard and 9% as portland cement retarder. The economic performance of gypsum is and will continue to be tied closely to that of the construction industry.

Although gypsum and anhydrite are potential sources of sulfur for industrial chemicals such as ammonium sulfate and sulfuric acid, the calcium-sulfate minerals are not competitive sources for sulfur in North America, which has large resources of native sulfur (Appleyard, 1975). One special exception to this situation is the production by the Frasch method of naturally-occuring elemental sulfur from diagenetically altered anhydrite (now limestone) in the Ochoan rocks of the Delaware Basin. Although there is no Frasch production of sulfur in New Mexico at this time, Duval Corporation's Culberson mine, the world's largest Frasch-mining operation, in the Rustler Hills of Culberson County, Texas, is only 30 km from the Texas-New Mexico border, and there is active exploration for sulfur in the geologically equivalent Ochoan units in the New Mexico portion of the Delaware Basin.

Gypsum is a widely distributed rock-forming mineral and is produced from deposits that are large enough to sustain commercial mines or quarries in 22 states. Canada, Mexico, several Caribbean islands, and even countries in western Europe export substantial amounts of crude gypsum to the U.S., taking advantage of relatively low marine-freight rates to deliver gypsum to large U.S. seaports along the Atlantic and Gulf coasts. In 1980, over one-third of the total U.S. gypsum consumption was provided by imports of crude gypsum.

The low unit value of crude gypsum, averaging about $8.50/metric ton in 1980, and the weight and bulk of finished products, especially wallboard, dictate that in most cases gypsum plants must be near both mines and markets. Three geographic factors are particularly relevant to an examination of the gypsum resources of New Mexico. First, New Mexico is far removed from most of the major population centers and markets of the U.S. Chicago is 1,850 km from Albuquerque; Dallas-Fort Worth is 950 km; Los Angeles is 1,100 km. Even Denver is more than 700 km from Albuquerque. Second, even within New Mexico, distance is a major issue. New Mexico is the fifth largest state in area in the U.S. A trip from Farmington (northwest New Mexico) to Carlsbad (southeast New Mexico) is roughly equivalent to a trip from Dayton, Ohio to Raleigh, North Carolina. Third, although New Mexico is the fifth largest state in area, it is the 37th in population, with only 1.3 million people. Only one city, Albuquerque (near the center of the state), has a population over 100,000, and 50% of the state's population lives within 150 km of Albuquerque. The limited markets and high freight rates implicit in these geographic parameters place restrictions on the economic evaluation of New Mexico's gypsum resources and should be kept in mind in the following discussion.

Gypsum has been used as an industrial mineral in New Mexico for at least 1,000 yrs. Rock gypsum and selenite crystals were shaped into ornaments and ground for pigments by the Pueblo Indians and probably by their ancestors (Weber, 1965). In the oldest buildings still standing at Acoma Pueblo (70 km west of Albuquerque), selenite cleavage flakes were used for windows, a practice that continued well into the American Territorial period of the 19th century. Calcareous gypsite was used as a mortar material, and gypsum-based plaster was locally produced for covering buildings from the Spanish Colonial period until after World War I (Weber, 1965). Gypsum has been used in agriculture for many years, and, following World War II, agricultural gypsum from the White Mesa deposit northwest of Albuquerque was used extensively as far away as the San Luis Valley of southern Colorado (A. Pomeroy, personal communication, 1981).

However, not until 1959-1960 did a modern gypsum industry develop in New Mexico. In 1959, the Tijeras mine and mill of the Cement Division of Ideal Basic Industries, located just east of Albuquerque, began operations, requiring gypsum as a retarder for their portland cement. Two wallboard plants, the American Gypsum Company plant in Albuquerque and the Rosario plant of the Kaiser Gypsum Company, approximately 50 km north of Albuquerque, opened in 1960.

The following review of the gypsum resources of New Mexico will consider the geologic occurrence and distribution of gypsum in the state, review the recent production history, and present some projections for the development of New Mexico's gypsum resources for the future. Much of the information and even the structure of the presentation is drawn from the reviews of Weber and Kottlowski (1959) and Weber (1965).

Geologic occurrence and distribution

Gypsum is widely distributed in New Mexico (fig. 1). The major deposits of the state are located close to the Rio Grande and Pecos River valleys, where more than 80% of New Mexico's population is concentrated. Gypsum deposits are found in rocks ranging in age from Pennsylvanian to Pleistocene, with local dune sands and gypsite of Holocene age (tables 1-3). Most of the bedded gypsum in central, southern, and southeast New Mexico is middle Permian in age and represents gypsum precipitated during evaporation of sea water. In north-central New Mexico the gypsum is Upper Jurassic in age and was deposited as a chemical sediment during the
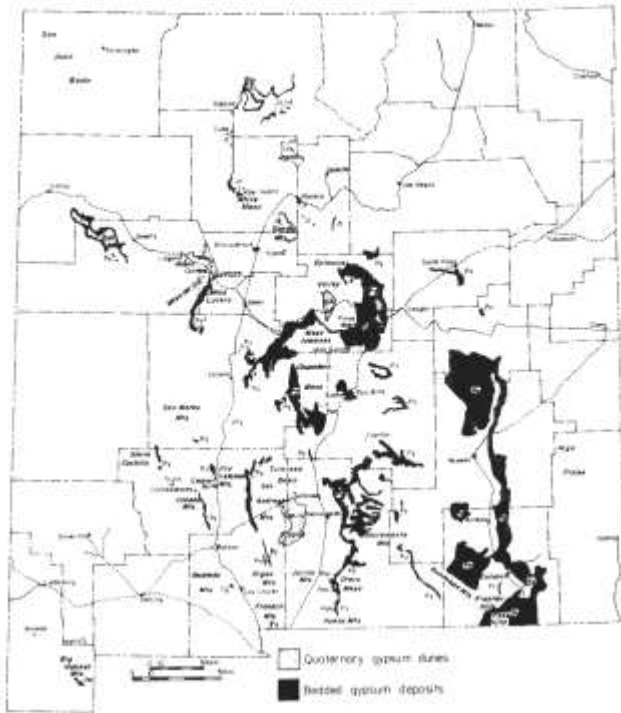
FIGURE 1—INDEX MAP OF GYPSUM OCCURRENCES OF NEW MEXICO. *Pg:* Pennsylvanian gypsum; *Pah:* Permian Abo Formation–Hueco Limestone; *Py:* Yeso Formation (Permian); *Pe:* Epitaph Dolomite (Permian); *Psa:* San Andres Limestone (Permian); *Pa:* Artesia Group (Permian); *Pc:* Castile Formation (Permian); *Pr:* Rustler Formation (Permian); *Jt:* Todilto Formation (Jurassic); *Tg:* Tertiary gypsum. (Modified from Weber and Kottlowski, 1959 and Weber, 1965).

TABLE 2—GENERALIZED STRATIGRAPHIC NOMENCLATURE IN SOUTH-EAST NEW MEXICO; Rustler Formation, Castile Formation, Tansill Formation, Seven Rivers Formation, Grayburg Formation, San Andres Limestone, Yeso Formation, Abo Formation, Hueco Limestone (all Permian) and Panther Seep Formation (Pennsylvanian) are gypsiferous (from U.S. Geological Survey, 1965, Mineral and water resources of New Mexico: New Mexico Bureau of Mines and Mineral Resources, Bull. 87, 437 p.).

TABLE 1—GENERALIZED STRATIGRAPHIC NOMENCLATURE IN SOUTH-WEST NEW MEXICO; Yeso Formation and Epitaph Dolomite of the Colina Limestone (both Permian) are gypsiferous (from U.S. Geological Survey, 1965, Mineral and water resources of New Mexico: New Mexico Bureau of Mines and Mineral Resources, Bull. 87, 437 p.).

TABLE 3—GENERALIZED STRATIGRAPHIC NOMENCLATURE IN NORTH-CENTRAL AND NORTHWEST NEW MEXICO; Yeso Formation (Permian) and Todilto Formation (Jurassic) are gypsiferous (from U.S. Geological Survey, 1965, Mineral and water resources of New Mexico: New Mexico Bureau of Mines and Mineral Resources, Bull. 87, 437 p.).

[New Mexico Bureau Mines & Mineral Resources, Circular 182, 1982]

evaporation of a very large, hypersaline lake (Anderson and Kirkland, 1960; Rawson, 1980) or paralic basin (M. B. Goldhaber, personal communication, 1981). The third major gypsum occurrence is as extensive Quaternary dune sands in the Tularosa Basin and in and near the Estancia Valley. Exposures of Pennsylvanian, Cretaceous, and Tertiary bedded gypsum, Quaternary lake deposits, and hydrothermal gypsum-anhydrite veins and stringers are less extensive than the Permian and Jurassic bedded deposits and the Quaternary dune sands but could be of local economic importance.

In New Mexico, calcium-sulfate minerals at and near the surface are virtually always gypsum. Significant nodular anhydrite begins to appear in the Todilto Formation at the White Mesa deposit at a depth of about 20 m; in the Permian beds of south-central and southeast New Mexico, gypsum occurs to depths of approximately 150 m, below which depth the calcium-sulfate mineral is anhydrite. Chemical analyses of 20 gypsum samples from New Mexico are listed in Weber and Kottlowski (1959) and Weber (1965).

In southwest New Mexico, there are two occurrences of gypsum in the vicinity of the Big Hatchet Mountains (fig. 1, table 1). Gypsum in the Epitaph Dolomite (Permian) is up to 100 m thick; grab samples indicate that much of the gypsum is greater than 95% pure (Weber and Kottlowski, 1959). The gypsum is substantially deformed, and the deformation and extraordinary thickness of the gypsum are probably due to imbricate thrusting in the New Mexico portion of the Laramide overthrust belt. The deposit is reasonably accessible, and much of it could be quarried in open pits. Although this deposit is probably too distant from major markets to justify a wallboard plant, it could supply agricultural gypsum to the irrigated farming areas of southwest New Mexico, as suggested by Weber and Kottlowski (1959). The second occurrence is gypsum interbedded with clastic beds of the Hell-to-Finish Formation (Lower Cretaceous) in the foothills south of the Big Hatchet Mountains. These beds are more remote, thinner, and less favorable for mining than are the Permian beds.

Both areally and volumetrically, the largest gypsum reserves are in southeast and central to south-central New Mexico, where gypsum occurs in units of Pennsylvanian, Permian, Tertiary, and Quaternary ages (fig. 1, table 2). Areally small but locally thick gypsum occurs in the Panther Seep Formation (Pennsylvanian) of the southern San Andres and northern Franklin Mountains. Weber and Kottlowski (1959) deduced that the gypsum in the Panther Seep Formation was deposited originally in a basin that was approximately 80 km north-south by 40 km east-west. Present thickness of the gypsum in the Panther Seep Formation decreases from 20 to 30 m in the San Andres Mountains to 3 to 10 m in the Franklin Mountains. Gypsum from this unit has been mined by the El Paso Portland Cement Company (Weber, 1965).

Most of the bedded gypsum in central, southern, and southeast New Mexico is Permian in age. Permian gypsum occurs in rocks of the Abo Formation-Hueco Limestone, Yeso Formation, San Andres Limestone, Artesia Group, Castile Formation, and Rustler Formation (fig. 1, table 2). The Abo-Hueco gypsum occurrences in southern Otero County are up to 8 m thick, but are discontinuous and often overlain by thick clastic units (Weber and Kottlowski, 1959).

"Yeso" is the Spanish word for gypsum, and the most widely distributed gypsiferous unit in the state, forming a major part of the total gypsum resource, is the Yeso Formation. The Yeso Formation is particularly widely distributed over central and south-central New Mexico (fig. 1). Twelve percent of the 215-m-thick type section of the Yeso Formation, located 25 km northeast of Socorro, and 40% of the 500-m-thick Yeso section in the San Andres Mountains is gypsum (Weber, 1965). The Canas Gypsum Member of the Yeso Formation is 15-50 m thick over much of the area from Mesa del Yeso and Chupadera Mesa southeast to the Sacramento Mountains, and the Torres Member also contains gypsum

beds (Weber and Kottlowski, 1959). However, gypsum in the Yeso Formation is typically remote from major markets and is often unfavorable for mining because of thick overburden, interbedding with clastic and carbonate units, and tectonic and geomorphic complications.

Significant gypsum deposits occur in the San Andres Limestone in central New Mexico (fig. 1). A plaster mill that used the local San Andres gypsum and its associated gypsite was active at Ancho in western Lincoln County in the early part of this century. Solution collapse has seriously affected the continuity and distribution of gypsum in the San Andres Limestone, especially in the Ancho-Carrizozo region (Weber, 1965).

Gypsum in the Grayburg, Seven Rivers, and Tansill Formations of the Artesia Group crops out along the Pecos River valley from Santa Rosa in east-central New Mexico to Carlsbad in the southeast corner of the state (fig. 1, table 2). Although the gypsum beds are locally thin, discontinuous, or grade laterally into dolomitic limestone, there is a substantial amount of gypsum in the Artesia Group. Extensive irrigated farming along the Pecos River and on the Llano Estacado of eastern New Mexico and west Texas could provide a sizable market for agricultural gypsum from the Artesia Group.

Very large volumes of gypsum are present in the Upper Permian (Ochoan) Castile and Rustler Formations in extreme southeast New Mexico (fig. 1, table 2). Castile Formation gypsum beds, which range from 92 to 97% gypsum, lie at or near the surface over an area of 250 km^2 southeast of the Guadalupe Mountains; in much of this area, the gypsum beds are as thick as 30 m. Relatively poorly exposed gypsum is also present in the Rustler Formation between Carlsbad and the New Mexico-Texas border. As with the Artesia Group gypsum, the most likely market for the Castile and Rustler gypsum would be as agricultural gypsum.

Gypsum of Tertiary age occurs in the Robledo Mountains, approximately 8 km northwest of Las Cruces (fig. 1). Beds of gypsum ranging in thickness from a fraction of a meter to 3 m are interbedded with redbeds near the base of a Tertiary volcanic sequence. The highly productive agriculture of the Mesilla Valley provides a potential market for this gypsum, but problems of access and of minability probably render the occurrence uneconomic (Weber, 1965).

No discussion of gypsum resources or of the geology of southeast New Mexico would be complete without reference to the White Sands and other Quaternary dune fields. The White Sands are one of New Mexico's most prominent geologic features, even from space (Short and others, 1976, p. 114). The White Sands dune field in the Tularosa Basin of southeast New Mexico is a pure white, active dune field covering 450 km^2 with 3- to 15-m-high dunes, most of which is at least 99% gypsum. To the north and south of the main dune fields are hundreds of square kilometers of more dunes of lower purity. The following notes on the hydrogeologic origin of the White Sands summarize the study of Allmendinger and Titus (1973). Precipitation falling on the San Andres, Oscura, and Sacramento Mountains during the Late Wisconsin Stage (22,000-11,000 yrs B.P.) of the Pleistocene Epoch dissolved gypsum from the Permian units in the mountains. Runoff and ground water rich in dissolved solids fed ancient Lake Otero, a large, pluvial lake in the southwest part of the Tularosa Basin. As the climate changed toward today's aridity, Lake Otero shrank and eventually evaporated to dryness, with concommitant increases in the concentration of dissolved solids in solution. Saline lacustrine deposits with thick beds of gypsum, including large selenite crystals, were precipitated from the hypersaline waters. Deflation of the playa surface by the predominant southwest winds exposed the gypsum and blew gypsum flakes to the east-northeast across the Tularosa Basin. Similar hydrologic processes are active today: gypsum derived from ground water dissolution of Permian evaporites and Quaternary lacustrine deposits precipitates in the capillary

fringe of the playa bed of modern Lake Lucero, and continued deflation of the dry lake bed and the alkali flats surrounding it contributes small amounts of modern gypsum to the active dune fields (Allmendinger and Titus, 1973). Although the gypsum dunes of the Tularosa Basin represent an easily exploitable gypsum resource totaling billions of tons, the gypsum may never be developed because most of it lies within the White Sands National Monument or the White Sands Missile Range.

Similar but less pure and much less extensive Quaternary gypsum dunes related to saline playa lakes occur in basins north of the Tularosa Basin (fig. 1). Partially to fully stabilized dunes are exposed in the southern Estancia Valley, east of Pinos Wells, and west of Vaughn in east-central New Mexico. The dunes are composed of selenite cleavage flakes, clay, and silt. Some of the purer dunes, containing more than 75% gypsum, are probably suitable for agricultural use if local markets arise (Weber, 1965).

The economically most important gypsum resources of New Mexico are in the Todilto Formation (Jurassic) of north-central New Mexico, within 80 km of Albuquerque. The Todilto Formation is a nonmarine or shallow, paralic, laminated sequence of limestone and gypsum. The formation consists of a thin, finely laminated, basal limestone member of wide extent and thick, upper gypsum member that is restricted in distribution to the center of the basin in which the Todilto was deposited (fig. 2). In most Todilto Formation exposures, the basal limestone is 2-3 m thick; the gypsum member, generally 12-30 m thick, is thickest toward the center of the depositional basin. Petrographic examination shows that limestone laminae, although often deformed, persist throughout the massive gypsum (Anderson and Kirkland, 1960).

The contact between the basal limestone of the Todilto Formation and the underlying Entrada Sandstone (Jurassic), the upper part of which is largely eolian in origin, is transitional. Similarly, a transitional contact occurs between the massive, upper gypsum member and the overlying Morrison Formation (Jurassic), which is fluvial in origin. Thus, it appears that the Todilto Formation was deposited during the transition from an arid climate, which produced an eolian sandstone environment, to a more humid climate, which produced the mudstones of the Morrison Formation (Anderson and Kirkland, 1960).

By the beginning of Todilto time, a large basin had formed in northwest New Mexico and southwest Colorado. Streams draining the Permian and Triassic carbonate, evaporite, and shale units that surrounded the basin carried relatively heavy dissolved-solid loads and suspended sediment into the basin. Some seawater probably entered the basin from the northwest (M. B. Goldhaber, personal communication, 1981). On the basis of detailed petrographic studies, Anderson and Kirkland (1960) established that there was cyclical sedimentation of the Todilto Formation, yielding regular combinations of clastic, organic, and evaporite components. Further work showed that the cycles could be closely correlated with sunspot cycles, leading Anderson and Kirkland to the interpretation that the cycles were annual and climatically based. Furthermore, they recognized an overall clastic-organic-evaporitic cycle which they believed represents a full, 20,000-yr climatic cycle from predominantly humid conditions, represented by the transition between the Entrada Sandstone and the basal limestone of the Todilto, through progressively more arid conditions, represented by the limestone and the massive gypsum, and back into more humid conditions, represented by the transition into the Morrison Formation (Anderson and Kirkland, 1960).

The two active and one recently active gypsum mines in New Mexico all quarry Todilto gypsum from flat-lying to moderately-dipping benches with no overburden except gypsite. Gypsum, typically 93-97% pure, is mined to depths of 15-30 m below the gypsite cap, below which nodular anhydrite or limestone typically prevent further mining of industrial-quality gypsum.

The White Mesa mine, located approximately 65 km northwest of Albuquerque, produces high-purity gypsum from a shallowly dipping mesa, the top of which is a gently rolling surface covered with grayish gypsite. The structure of the deposit is such that an open-pit quarry can efficiently mine the gypsum to a depth of 20 m, below which anhydrite appears in significant quantities (fig. 3; A. Pomeroy, personal communication, 1981). The gypsum is trucked to the north side of Albuquerque for processing into wallboard at the American Gypsum Company plant.

The small San Felipe pit in the Todilto Formation along Tongue Arroyo, about 30 km north of Albuquerque, produces gypsum exclusively for use as retarder in portland cement at the Tijeras plant of the Ideal Cement Company, just east of Albuquerque. Gypsum forms the dip slope of a moderately diping questa along Tongue Arroyo (fig. 4). Twenty to 30 m of gypsum overlies Entrada Sandstone in the cliff face, and the gypsum appears to thicken to the east, probably due to faulting that has repeated part of the Todilto section. The

FIGURE 2—WEATHERED SURFACE OF THE MASSIVE GYPSUM MEMBER (Jtg) OVERLYING THE LAMINATED, DARK-GRAY LIMESTONE MEMBER OF THE TODILTO FORMATION (Jtl). The sandstone unit underlying the Todilto limestone member is the Entrada Sandstone (Je).

FIGURE 3—WHITE MESA MINE OF POMEROY, INC., LOCATED NORTHWEST OF ALBUQUERQUE. Gypsum is hauled to the American Gypsum Company wallboard plant in Albuquerque.

FIGURE 4—TODILTO FORMATION *(Jt)* OVERLYING ENTRADA SAND-
STONE *(Je)* ALONG TONQUE ARROYO.

San Felipe operation is small—two men in the pit and one truck driver delivering the gypsum to the cement plant—but quite efficient, providing all the gypsum needed for a half-million-tons-per-year cement plant.

The third gypsum operation in the Todilto Formation is the Rosario mine of Western Gypsum Company, located along Galisteo Creek, approximately 60 km north of Albuquerque. Twelve- to 30-m-thick gypsum crops out as an almost flat- lying bench above the cliff-forming Entrada Sandstone (fig. 5). The main line of the Atchison, Topeka, and Santa Fe Railroad follows part of Galisteo Creek, essentially alongside the gypsum deposit. Furthermore, an 8-inch natural-gas pipeline lies within 10 km of the deposit. In 1960, Kaiser Gypsum Company built a wallboard plant adjacent to the Rosario deposit and a 500-m spur line connecting the plant to the main line of the Santa Fe railroad. The mining operation consists of loading a haulage truck with a front-end loader, backing the truck up at the top of the cliff, and dumping the gypsum over the side directly into the primary crusher of the plant. The combination of ease of mining and relatively high purity of Todilto gypsum allows New Mexico gypsum products to be competitive regionally, in spite of high freight costs.

Significant gypsum deposits are widely distributed in the state. However, commercially viable deposits are limited un-

FIGURE 5—ROSARIO MINE AND WALLBOARD PLANT OF WESTERN GYP-
SUM COMPANY, LOCATED ALONG GALISTEO CREEK; Todilto Forma-
tion *(Jt)* overlying Entrada Sandstone *(Je)*.

der present market conditions to the high-purity, well-exposed deposits of Jurassic Todilto gypsum within 80 km of Albuquerque.

Production history and projections

Production figures for gypsum in New Mexico from 1973 to 1980 are given in table 4. There are two active and one recently active gypsum operations: the San Felipe Mine produces crude gypsum used for retarder in portland cement; the active White Mesa mine and the recently-active Rosario mine produce gypsum for wallboard plants. When all three mines are operating and the economy is reasonably healthy, their combined production should be in excess of 225,000 metric tons per year, as in 1973, 1978, and 1979. Markets for the wallboard produced in New Mexico include Arizona, California, Colorado, and Texas, as well as in-state construction sites. But when construction starts are down, production of gypsum is down, as in 1980. In May, 1981, the Denver-based parent company of Western Gypsum, the current operators of the Rosario mine and wallboard plant, ordered the shutdown of the Rosario operation until market conditions improve. The last time Rosario closed, in 1974, the quantity and value of gypsum produced in New Mexico dropped dramatically (table 4). The combined loss of the Rosario production and decreased demand by Ideal Cement for San Felipe gypsum will probably lower the 1981 production to a figure close to the 1974 quantity, although the dollar value will be substantially higher than in 1974.

Although the production figures for 1980 and 1981 are disappointing, the future of New Mexico gypsum remains bright. Reserves at the three Todilto Formation gypsum mines are in the hundreds of millions of tons. The well-exposed beds of Todilto gypsum in the Ojo del Espiritu Santo Grant and at Gallina are as large or larger; the Mesita and Suwanee deposits are also a significant resource of easily exploited Todilto gypsum (Weber and Kottlowski, 1959). Large resources in the Permian rocks of central, southern, and southeast New Mexico and the huge resource of the gypsum dunes of the Tularosa Basin outside the White Sands National Monument increase the total gypsum resource of New Mexico well into the billions of tons. Much of the gypsum is well exposed and close to Santa Fe and Southern Pacific railroad lines and to modern, heavy-duty highways.

Most importantly, basic demographics are on the side of New Mexico's gypsum industry. New Mexico's population grew by almost 28% from 1970 to 1980. As population and industry continue to move to the Sunbelt states, particularly the energy-rich states of the Southwest, demand for the high-purity, easily exploited gypsum resources of New Mexico is almost certain to grow.

TABLE 4—GYPSUM PRODUCTION IN NEW MEXICO. W: data withheld to avoid disclosing proprietary information; Source: U.S. Bureau of Mines; 1 ton = 10³ kg = 2204.6 lbs = 1.1023 short ton.

Year	Quantity (10³ tons)	Value (10³ dollars)
1973	231	1220
1974	142	532
1975	W	W
1976	W	W
1977	165	1227
1978	239	2649
1979	228	3244
1980	189	3005

mission to visit, sample, and photograph the Rosario, White Mesa, and San Felipe mines. Financial and technical support for this study were provided by the New Mexico Bureau of Mines and Mineral Resources.

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LARGE-SCALE ADOBE-BRICK MANUFACTURING
IN NEW MEXICO

by Edward W. Smith, *Consulting Geologist, Tesuque, New Mexico 87574*

Abstract

The use of adobe by the Pueblo Indians, Spanish, and, later, Anglo-Americans has resulted in a native-adobe architecture that is a unique aspect of New Mexico. Historically and up to the present day, New Mexico has remained the largest manufacturer and user of adobe bricks in the United States. Results of a field evaluation and sampling program have identified 48 active producers whose production in 1980 totaled 4,133,000 adobe bricks valued at $1,174,598. Engineering tests have shown that the majority of adobe bricks meet the specifications of the Uniform Building Code and the New Mexico Building Code adopted in 1977. Ten large-scale adobe-brick manufacturers produced 81% of New Mexico's total production, with the majority of these producers located in the Albuquerque and Espanola Basin areas of the Rio Grande valley. The terminology, general characteristics, and techniques for the large-scale production of adobe bricks are detailed in this report, and examples of buildings representing the four architectural periods in New Mexico covering a span of 700 yrs are included.

Introduction

In 1980, the New Mexico Bureau of Mines and Mineral Resources undertook a study of New Mexico's adobe-brick industry. The study included the distribution and location of commercial adobe-brick producers, soil types, geology and mineralogy, methods of production, market trends, and economic factors. Several types of adobe samples were collected from 10 large-scale adobe-brick producers (fig. 1) and tested at the Rock Mechanics Laboratory of the New Mexico Institute of Mining and Technology in Socorro.
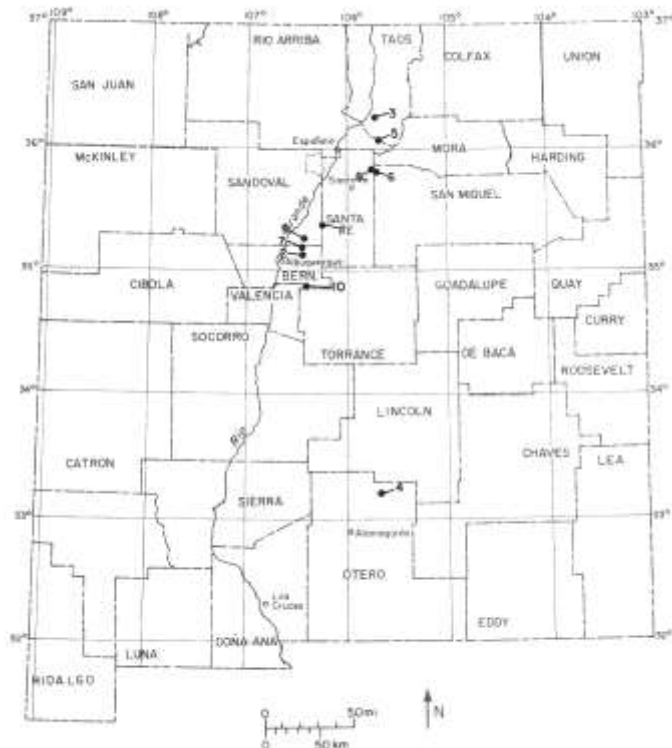
During this study, a total of 48 active adobe producers were identified in 1980 and classified as small-, medium-, or large-scale manufacturers. Estimated dollar value of production for the three major types of adobe bricks (traditional, semistabilized, and stabilized) totaled $1,174,598.00. This figure accounts for the commercial production of 4,133,000 adobe bricks (table 1), 81% of which were produced by the 10 largest adobe-brick manufacturers in the state. Not included in the production totals are the thousands of adobe bricks produced by individual homeowners for their own construction projects, which may represent an additional 3-4 million adobes per year.

History

The word "adobe," which has its roots in an Egyptian hieroglyph denoting brick, evolved through the Arab and Spanish to its present form. Today, adobe describes various earthen-building materials and techniques, referring usually to the sun-dried adobe brick now most widely used in the U.S. The word also applies to puddled structures, adobe-plastered logs or branches, and even to rammed-earth construction (pisé).

Among the oldest of technologies mastered by prehistoric man, the use of mud as a building material parallels the birth and spread of the great ancient civilizations of the world. To this day, soil remains the primary building material of at least 50% of the world's population.

Remains of adobe structures have been found in Neolithic farming villages throughout the Near and Mid-East, dating as far back as 7,000 B.C. (Steen, 1972). Hand- and form-molded bricks from this era are seen in the ruins of structures throughout Mesopotamia, Crete, Egypt, and India. The use of mud construction rapidly spread eastward through Asia and westward through North Africa and the Mediterranean Basin; as the Ice Age receded in northern Europe, these peoples also began to use adobe in construction. Meanwhile, the New World was undergoing a similar development. Younger civilizations in the western hemisphere, dating from 3,000 B.C. at Chicama Valley in Peru, used adobe (Steen, 1972). The first adobe construction in what is now the United States probably does not pre-date the 10th or 11th century A.D.

Since that time, the region that now incorporates the state of New Mexico has undergone much development in adobe architectural styles by incorporating contributions from the Indian, Spanish, and Anglo residents of the area to produce a distinct, native New Mexican architecture. The evolution of architectural styles in New Mexico can be roughly broken down into four periods: Indian, Spanish Colonial, Territorial,

FIGURE 1—LOCATION OF LARGE-SCALE ADOBE BRICK MANUFACTURERS ACTIVE IN NEW MEXICO IN 1980.

TABLE 1—ESTIMATED 1980 ADOBE-BRICK PRODUCTION; *summer adobe-brick price averages for 38 traditional adobe-brick producers, 8 semistabilized and 6 stabilized adobe-brick producers.

Type of adobes	Average* price/adobe	Estimated production total	Estimated dollar value	% of total production
Stabilized	36.9ᶜ	460,000	$ 169,740	11
Semistabilized	27.8ᶜ	2,320,000	$ 644,960	56
Traditional	26.6ᶜ	1,353,000	$ 359,898	33
Totals	30.4ᶜ	4,133,000	$1,174,598	100%

and Later American (Bunting, 1976). What is known as the Indian Period actually includes all architecture built before the arrival of Spanish settlers in 1598 and can be organized from archaeological information into several distinct periods. Puddled-adobe structures, made by patting the mud into wall shape without forms, along with stone masonry and early hand-formed bricks, are seen at Chaco Canyon, old Picuris Pueblo (fig. 2), Sapawe and the Pajarito Plateau (Steen, 1977), and in later, multi-storied housing complexes such as the one at Taos Pueblo. Structures of this era were marked by a substantial defense system that allowed entry only from above and provided for few windows or doors. The resulting solid geometric lines and modular design are distinctive of the Pueblo style.

Upon the arrival of Spanish colonists in 1598, new building techniques and forms of architecture were introduced to New Mexico. Due to the isolation of the region and severe survival conditions imposed on the new settlers, the colonial era is characterized by little technical or cultural innovation. In fact, most buildings of this era were reduced to the barest essentials, and techniques and materials remained virtually unchanged from what the Indians had used before. The major contribution of the Spanish was the introduction of the formed, standard adobe brick to the Indian population of the area. In addition, the classic Spanish linear floor plan also began to appear, sometimes forming around a central plaza area for defense. Other notable architecture of the Spanish and Mexican Colonial period includes "fortress churches"—solid, rectangular, adobe churches that resemble fortresses in their imposing size and lack of fenestration.

With the opening of the Santa Fe Trail in 1821, influences from the eastern and midwestern U.S. began to trickle slowly westward and, upon annexation of the territory of New Mexico in 1848, the flow of new materials and ideas was increased. The Territorial Period was one of rapid economic and cultural development which contrasted markedly with the centuries of cultural isolation and stagnation that directly preceded it. This development is reflected clearly in the architecture of the period, which went through several phases, but was essentially nothing more than a transported version of the Greek Revival Style that had been popular in the East during the 1820's. The Territorial Style is probably best known for its neo-Classic, elaborate wood trim on windows and doors, symmetrical floor plan based around a center hall, two-story construction, and columned verandas. The influx of new materials such as milled lumber, window glass, burned brick (used to trim the tops of adobe walls), and corrugated iron, all contributed to the new modern American look of New Mexican architecture. There was even some attempt to disguise the adobe building material altogether with cement coating scored to resemble ashlar masonry (hewn or squared stone), or even some curious examples of elaborate Gothic-style wood trim.

The Later American Period began with the arrival of the railroad in New Mexico in 1880, and did not reach some of the more remote mountain villages until after the middle 1940's. During this period, many diverse architectural movements from California and the East were represented simultaneously as New Mexico rushed to catch up with the rest of the country. The availability of iron, improved tools and fittings, and other manufactured items hastened New Mexico's arrival to the 20th century. In addition, Pueblo-like architecture was experiencing a rebirth in what became known as the Santa Fe style; all of these factors contributed to a unique and eclectic blend of architecture that continues to visually distinguish New Mexico from the rest of the nation.

Adobe bricks are still made in the traditional way by hand methods at the site of construction. However, today, even on smaller jobs, the trend has been to purchase the bricks ready-made. Bricks are manufactured commercially on a large scale at centrally located adobe yards where the right soil is available. The controlled conditions of manufacture assure a uniformly good product and a continuous supply of adobe bricks; this increasingly important technique of manufacturing adobes will be discussed next.

Terminology and general characteristics

Some of the common terms used in the adobe trade are:

1) Adobe soils—Adobe bricks can be made from a variety of local soils, but the most suitable soil found in the Rio Grande basin area is a sandy loam composed of approximately 55-75$^{\%}$ sand and 25-45% finer material (generally equal parts of silt and clay). A proper balance of clay and sand is critical to ensure a quality adobe. Clay gives strength to the brick but, in excessive amounts, it will cause shrinkage; sand is added to decrease the shrinkage and prevent cracking (California Research Corp., 1963).

2) Stabilizers—Moisture re-entering an adobe brick causes the clay to swell and release its bonding so that the entire mass disintegrates. To prevent disintegration, various soil stabilizers are added to the basic soil mix to waterproof or increase the weathering resistance of the adobe brick. The most commonly used stabilizers are sand, straw, Portland cement, lime, and bituminous and asphalt emulsions (Clifton, 1977), although as many as 20 different materials have been found in use (Wolfskill, 1970). Asphalt emulsion is used exclusively in commercial operations in New Mexico because of its superior mixing and waterproofing properties.

3) Traditional adobe brick—Often referred to as an untreated, unstabilized, or standard sun-dried adobe brick, the traditional adobe is made with soil composed of a mixture of clay, sand, and silt. Usually straw is added to the adobe brick to prevent the brick from cracking while being cured. Taos Pueblo (fig. 3), a multi-storied Indian adobe structure over. 900 yrs old, is a good example of the lasting qualities of properly maintained traditional adobes.

4) Semistabilized adobe brick—Essentially the same as a traditional adobe brick, the semistabilized adobe is classified as a water-resistant brick because of the addition of a small amount of asphalt emulsion (4% or less by weight). The adobe contains a sufficient amount of stabilizer to protect stockpiled adobes from local rains and the production cost is significantly lower than that of a fully stabilized brick. The mosque now under construction at Abiquiu is a good example of the use of semistabilized adobe (fig. 4).

5) Stabilized adobe brick—The fully stabilized adobe, referred to by the New Mexico Building Code as a treated adobe, is defined to contain an amount of stabilizer sufficient

FIGURE 2—JESUS MERMEJO OF PICURIS PUEBLO STANDING NEXT TO THE OLD PUEBLO STRUCTURE BUILT OF PUDDLED ADOBE BETWEEN 1250 AND 1300 A.D. Site visited by the Spanish conquistador Juan de Oñate on July 13, 1598.

FIGURE 3—Taos Pueblo is an example of Pueblo architecture of five stories that has been continuously occupied for over 900 yrs. Photograph taken in August, 1980 during the annual re-plastering of the Pueblo.
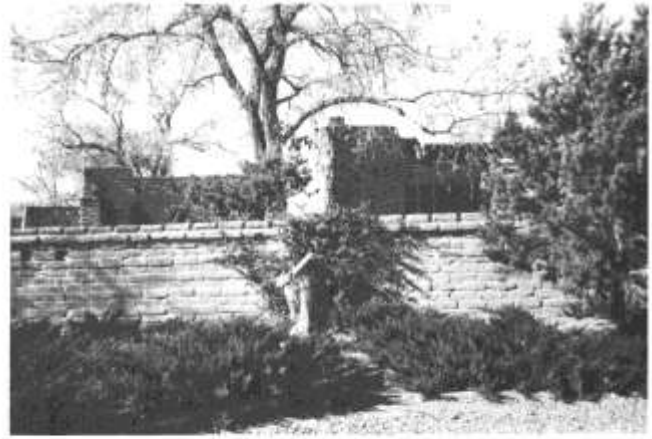
FIGURE 5—An example of recent Pueblo style architecture constructed of stabilized adobe bricks that have been left un-plastered and exposed to weathering for over 15 yrs. Located in Guadalupe woods area of Albuquerque.

to limit the brick's water absorption to less than 2.5% by weight. The fully stabilized adobe, usually manufactured with 5-12% asphalt emulsion, is a completely waterproof building material that resists water even without a protective coating. An example is a Pueblo-style house built in Albuquerque of stabilized adobe brick (fig. 5) that has been left uncovered and exposed for over 15 yrs without any appreciable weathering.

Geology

Usable adobe materials are found in large areas of New Mexico and constitute a virtually inexhaustible reserve. Adobe materials are obtained principally from the Quaternary alluvial deposits in stream channels, floodplains, terraces, and alluvial fans. The majority of deposits consist of a mix of clay, silt, sand, and gravel.

Most large-scale manufacturers of adobe bricks are located in the Albuquerque and Espanola Basin areas of the Rio Grande valley where significant quantities of good adobe materials are obtained from the floodplains of the Rio Grande and adjacent terraces and plains. Several adobe manufacturers also use crusher fines or fill dirt purchased from local sand and gravel operations. The crusher fines used by the Adobe Patch in Alamogordo usually contain between 1518% clay which produces a superior stabilized adobe brick

FIGURE 4—The Dar Al-Islam mosque project in Abiquiu was under construction in 1980 using semistabilized adobe bricks. Note arch and dome structures that are common in North Africa.

that meets both the specifications of federal agencies (such as the U.S. Department of Housing and Urban Development) and the State Building Code.

Mineralogy of adobe clays

Clay-size material is the chief binding agent in adobe bricks and most clay-size material consists of clay minerals. Four clay-mineral groups are commonly represented in New Mexico's adobe clays. The high-aluminum kaolinite group and the high-potassium illite group generally make up about one-half of the clay present in the 39 samples tested. Both groups are regarded as nonexpanding or only slightly expandable in the presence of water. The smectite group and the mixed-layer illite-smectite group contain varying expanding clay minerals and also make up about 50% of the clay minerals present. Expanding clay minerals take up both water and other materials into their structure. A member of the chlorite group was found in only one New Mexico adobe clay and the vermiculite group was not represented. Calcite ($CaCO_3$), found in the majority of adobe clay soils tested in New Mexico, is typical in adobe soils located throughout the arid Southwest. Calcite is somewhat soluble in water and, in the clay-size fraction, may act as an additional binder in the adobe brick.

The presence or absence of a particular mineral and total clay minerals present affects the physical properties of adobe bricks. A clay material with a high percentage of nonexpandable clay minerals, particularly the kaolinite group, would tend to be more brittle than one with more expandable clay minerals. Similarly, a clay material with a large percentage of expandable minerals will produce a brick with greater compressive strength than one with the same amount of clay, but a larger percentage of nonexpandable clay minerals. Thus, an adobe material with more expandable clay minerals, but less total clay, will have a strength equal to adobe with more clay but fewer expandable clay minerals. However, a clay material with an abundance of expandable clay minerals, particularly the smectite group which is the most expandable, may make a poor adobe brick. In such clays expansion and contraction of the clay minerals may cause excessive cracks in the adobe brick. More silt- or sand-like material, or even straw, can be added to the mix to minimize this effect.

Physical properties

In any building project, certain minimum requirements or specifications have been developed regarding the durability, strength, thermal characteristics, and fire safety of various

building materials used in construction (Long, 1946). The building of adobe structures in New Mexico is regulated by specifications adopted from the Uniform Building Code as outlined in Section 2405 of the 1977 revision of the New Mexico State Code. The sections of the code on unburned-clay masonry state basic requirements concerning the adobe brick, including regulations on soil composition, standards of water absorption, compressive strength, and modulus of rupture, as well as construction specifications. The New Mexico Code requires that an adobe brick have a compressive strength averaging 300 psi. The importance of this standard for a relatively heavy material such as adobe brick is apparent when considering the great amount of weight a typical wall unit must support. In addition to the weight of the roof structure supported by a load-bearing wall, each layer of bricks must support all others above it; this support depends on the compressive strength of the lower layers.

The other test required on all types of adobe bricks is the rupture test. This test helps indicate the relative cohesion of the materials that make up the adobe and the ability of the material to resist tension or shear forces that might result from settling of foundations or wind action. In testing for modulus of rupture, the adobe unit is placed on two supports and a force is applied to the center of the unit. The strength of the brick can be calculated from the amount of applied force that is taken to break the adobe brick.

Other tests are used on semi-stabilized and stabilized adobes to determine the water resistance of the bricks. The water-absorption test is only important if the bricks are left unplastered or if the test is a building requirement of the architect or federal housing agency.

In general, values of physical properties vary for the adobe bricks tested from major adobe producers located throughout New Mexico (table 2). This variability results from differences in the composition of adobe soils, methods of producing adobe bricks, and the quality of workmanship. Test results on the three major types of adobe bricks produced by the large-scale manufacturers are as follows:

1) Traditional adobe bricks—The range of compressive-strength values for the traditional sun-dried adobe brick averaged 376 psi and varied from a low of 266 to a high of 489 psi. The modulus of rupture ranged from a low of 42 to 66 psi.

2) Semistabilized adobe bricks—The compressive-strength values for the semistabilized-adobe bricks averaged 388 psi, ranging from a low of 259 to 586 psi. The modulus of rupture ranged from a low of 36 to 99 psi. Seven-day water-absorption tests averaged 1.93% and moisture content averaged 1.4%.

3) Stabilized adobe bricks—Compressive-strength values for stabilized-adobe bricks averaged 426 psi, ranging from a low of 249 to 578 psi. The modulus of rupture ranged from a low of 51 to 157 psi. The 7-day water-absorption tests on stabilized adobes ranged from 1.3 to 5.0% for an average of 2.35%, and moisture content ranged from a low of 0.4 to 1.5% for an average of 0.79%. The State Building code requires that for a treated (stabilized) adobe brick, moisture content must not exceed 4% by weight and water absorption must not exceed 2.5% by weight.

Techniques of adobe-brick production

The production of adobe bricks is a labor-intensive, fuel-efficient industry employing easily extracted indigenous materials. The production of adobe bricks is a seasonal operation and is usually limited by the number of frost-free days for a particular adobe yard. In general, the production season varies from 5 to 9 months, depending on climate and weather conditions. Bricks are made in various sizes according to their intended use. The principal standard-size adobe brick produced and used in New Mexico measures 4" x 10" x 14" and weighs approximately 30 lbs. An average of one yd^3 of adobe soil is used to produce approximately 80 bricks. Other sizes and types of adobes are usually made upon special order. The three major techniques of adobe production used in New Mexico (University of New Mexico, 1970) have been classified as follows:

TABLE 2—SUMMARY OF PHYSICAL-PROPERTY TESTS OF THE LARGE-SCALE ADOBE PRODUCERS; specification requirements for Uniform Building Code and New Mexico State Building Code: *Compressive strength:* average of 5 bricks—300 psi minimum, 1 out of 5 bricks—250 psi minimum; *Modulus of rupture:* average of 5 bricks—50 psi minimum; *7-day water absorption:* 2.5% maximum by weight; *Moisture content:* 4.0% maximum by weight; **test results are from limited sampling of adobe bricks from each adobe yard and may not be representative of the total annual-production averages. The New Mexico State Building Code recommends testing of samples selected at random from each 25,000 bricks produced.

Adobe producers**	Type of adobe brick	Size (inches)	Average adobe weight (lbs)	Compressive strength (psi)	Modulus of rupture (psi)	Water absorption (% by weight)	Moisture content (% by weight)
New Mexico Earth Alameda	Traditional	10 × 4 × 14	30	489	66	11.6	0.9
	Semistabilized	10 × 4 × 14	30	N/A	N/A	N/A	N/A
	Stabilized	10 × 4 × 10	32.4	499	89	1.3	0.5
Adobe Enterprises, Inc. Albuquerque	Semistabilized	10 × 4 × 14	30	N/A	N/A	N/A	N/A
	Stabilized	10 × 4 × 14	30	N/A	51	1.7	0.75
Eight Northern Indian Pueblos San Juan Pueblo	Semistabilized	10 × 4 × 14	30.3	317	99	4.3	0.9
	Stabilized	10 × 4 × 14	31.6	382	71	5.0	1.0
The Adobe Patch Alamogordo	Stabilized	10 × 4 × 14	33.8	578	157	2.2	0.0
The Adobe Farm Española	Semistabilized	10 × 3½ × 14	33	322	82	2.1	1.1
Victor Montano Santa Fe	Traditional	10 × 3½ × 14	30	438	46	N/A	N/A
Western Adobe Albuquerque	Semistabilized	10 × 4 × 14	32	456	96	N/A	1.0
Eloy Montano Santa Fe	Traditional	10 × 3¼ × 14	30*	320	42	N/A	N/A
Manuel Ruiz Corrales	Traditional	10 × 4 × 14	30	311	55	N/A	N/A
Rio Abajo Adobe Works Belen	Semistabilized	10 × 4 × 14	30	N/A	87	2.8	0.6
	Stabilized	10 × 4 × 14	27	486	101	1.8	0.76

FIGURE 6—THE SYSTEM OF MIXING ADOBE MUD AND STRAW BY FOOT TREADING IN A MUDPIT IS THOUSANDS OF YEARS OLD. Photograph taken in Nambé, summer of 1980.

FIGURE 8—A MICHIGAN MODEL 75-3 FRONT-END LOADER WITH A 2-YD³ BUCKET WORKING A CEMENT-LINED MUDPIT AT THE RIO ABAJO ADOBE WORKS IN BELEN.

1) Traditional handcrafted method—This relatively simple process involves mixing soil, water, and, sometimes, straw in a shallow mud pit and using a hoe or foot treading (fig. 6) to blend the mixture. Wooden forms that produce a single or multiple number of bricks are laid out on smooth and level ground. Mud is placed in them, tamped into the corners (fig. 7), the top is smoothed off, and the form removed. Forms are then washed clean, and the process is repeated. As the bricks dry (2-3 days in summer), they are turned on edge, trimmed of excess material or rough edges, and sometime later are stacked for delivery. This system, utilizing 2-3 adobe makers, can produce 300-500 adobe bricks per day.

2) Semimechanized method—This method of production, similar to the traditional handcrafted method except for the addition of mechanical equipment, most frequently a front-end loader, also represents the lowest capital investment (equipment) of the operations surveyed that are capable of large-scale adobe-brick manufacturing. Three outstanding examples of the semimechanized method, all producing a 10" x 4" x 14" adobe brick, are described as follows:

a) *The Rio Abajo Adobe Works,* Belen, located on approximately 3 acres, uses a 2-yard³-capacity front-end loader to move the stockpiled adobe soil into a cement-lined mud pit measuring approximately 30 x 65 ft and sloping to a maximum 3-ft depth at one end. Adobe material is stockpiled in

the mud pit and measured amounts of water and asphalt emulsion are added. The material is then allowed to soak overnight to assist in breaking down the clay material. The next day, mud is mixed in the mud pit by the front-end loader until a uniform mixture of adobe soil is achieved (fig. 8). Before mixing the adobe soil, over 200 oil-sprayed wooden forms, yielding 8 bricks per frame, have been placed adjacent to each other in the level laying yard of the property. The front-end loader picks up adobe materials from the mud pit and delivers mud to the wooden forms. Mud is poured into the forms and the excess material is raked and leveled (fig. 9).

The adobes are allowed to dry in the forms for several hours, depending upon the weather and, when the bricks have shrunk from the form sides, they are lifted and moved to an adjacent location for the next pouring of adobe mud. Bricks are allowed to dry for 2 or 3 days at which time they are turned on edge, trimmed and later stacked ready for delivery. This method of production, with a crew of 2-3 employees, is capable of producing 1,500-3,000 adobe bricks per day.

b) The *New Mexico Earth Company,* Alameda, is the largest producer of adobe bricks in New Mexico. More than l,000,000 adobe bricks were produced in 1979 and 700,000 bricks were produced in 1980. Soil is purchased from local sand and gravel operations which extract the material from the overburden (upper 15-30 ft) of the alluvial-plain and ter-

FIGURE 7—WHEELBARROWS DELIVER ADOBE MUD FROM THE MUDPIT WHERE IT IS PACKED INTO WOODEN FORMS THAT MOLD FOUR BRICKS AT A TIME. Photograph taken in Santa Fe, summer of 1980.

FIGURE 9—ADOBES DRYING IN WOODEN FORMS AT THE RIO ABAJO WORKS AFTER THE MUD HAS BEEN POURED AND RAKED SMOOTH. Forms will be removed the same day after the mud has partially dried and shrunk from the mold sides.

FIGURE 10—AFTER THE ADOBE MUD HAS PARTIALLY DRIED, WOODEN FORMS ARE LIFTED AND RELOCATED FOR THE NEXT MUD POURING. Photograph taken at the Adobe Enterprise yard in Albuquerque.

FIGURE 13—EXAMPLE OF A TYPICAL MIXING PUGMILL AT THE NEW MEXICO EARTH YARD IN ALAMEDA. After mixing soil, asphalt emulsion, and water, mud is dumped into the mudpit where it is removed by a Michigan Model 55 front-end loader for delivery to the wooden forms.

FIGURE 11—A 7.5-YD³ READY-MIX TRUCK MIXES THE ADOBE SOIL, WATER, AND ASPHALT EMULSION AND DELIVERS THE MIXTURE TO WOODEN FORMS FOR POURING. Photograph taken at the Adobe Farms yard in Española.

race deposits of the Albuquerque Basin. Stockpiled material is moved by a 1.5-yard³ front-end loader and placed into a 7 x 7-ft material hopper. Here, soil is screened to remove any large gravel and is then transferred into a 7-ft pugmill trough. As adobe soil is fed into the pugmill trough, water and asphalt emulsion are added to produce a uniform mud mixture. The mud is then dumped into a large mud pit, removed by the front-end loader, and carried to the adobe laying yard. Located in the leveled yard are several hundred (500-600) adobe forms of 10 molds per frame into which adobe mud is dumped and then raked and leveled. The adobe bricks are allowed to dry for several hours or until they have started to shrink from the frame sides. The forms are then lifted (fig. 10) and relocated to a new area. Adobe bricks are allowed to sun dry for 2 or 3 days at which time they are turned on edge, trimmed, and finally stacked for delivery. This operation can produce an average of 5,000-6,000 adobe bricks per day.

c) The *Adobe Farms Operation,* located at Arroyo Seco near Espanola, uses a front-end loader to move adobe material to a screening plant. Soil is then conveyed to a ready-mix truck (fig. 11). Measured amounts of water and asphalt emulsion are added and the ready-mix truck delivers the mixed adobe mud to a large number (100-200) of wooden forms laid out in the leveled adobe yard (fig. 12). The adobe laborers then rake and level the mud in the forms to dry for several hours before lifting and moving the forms to a new location. The adobe bricks are allowed to sun dry for 2-3 days before

FIGURE 12—POURING, RAKING AND LEVELING OF ADOBE MUD IN WOODEN FORMS AT THE ADOBE FARMS YARD IN ESPAÑOLA.

FIGURE 14—WATER AND OIL TANKS, PUGMILL, MUDPIT, AND CONVEYOR-BELT SYSTEM AT THE WESTERN ADOBE YARD IN ALBUQUERQUE.

FIGURE 15—Self-propelled adobe layer awaiting a load of mud delivered from the mudpit by the front-end loader. Note adoberos turning and cleaning the adobes at the Western Adobe yard in Albuquerque.
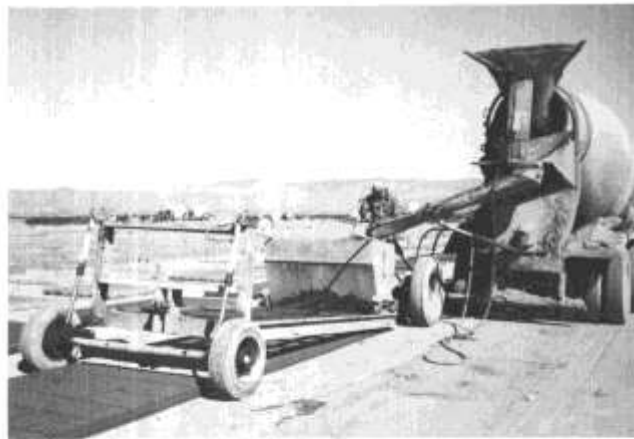
FIGURE 16—A 7.5-yd³ ready-mix truck and a mechanical adobe layer attached to the truck's power equipment are used at the Adobe Patch in Alamogordo.

they are turned on edge, trimmed, and finally stacked. This method of production, with a minimum of mechanical breakdowns, is capable of producing l,500-3,000 adobes per day with a crew of 2-3 employees.

3) *Mechanized method*—Two large-scale, mechanized techniques of adobe production are the use of the pugmill mixer with a self-propelled mechanical adobe layer, and the use of a ready-mix truck with an attached mechanical adobe layer. Examples of the first type of operation are the Western Adobe Plant in Albuquerque (fig. 13) and the Eight Northern Indian Pueblos, Native Products Division Plant at San Juan Pueblo. In both operations, sandy loam from the area is either removed by grading and leveling of the production site or delivered from nearby alluvial-terrace or flood-plain deposits. The soil is usually screened and conveyed to one end of the trough of the pugmill. Two shafts, studded with paddles, rotate in the trough of the pugmill, and an operator controls the feed of soil, water, and asphalt emulsion (fig. 14). The materials, thoroughly mixed in the trough and dumped into a mud pit, are then removed by a front-end loader that carries the mud mix to a nearby molding machine. This self-propelled mechanized adobe layer or molding machine is mounted on wheels and molds 25 4" x 10" x **14"** adobe bricks at one time (fig. 15). Different molds of various sizes and shapes can be mounted on the machine. With good weather and no mechanical breakdowns, an average of 5,000-7,000 adobes per day can be produced.

An example of the second type of mechanized, large-scale production is the *Adobe Patch, Inc.,* Alamogordo. This production system uses a 7.5-yd³ ready-mix truck into which adobe soil and measured amounts of water and asphalt emulsion are added using a front-end loader. The mud is mixed for an average of 45 minutes and is then delivered to a mechanical adobe layer (fig. 16) that is attached to the rear of the ready-mix truck and operates hydraulically using the truck's power equipment. The mixer empties adobe mud into the hopper of the adobe layer which feeds it into a steel mold to produce 42 adobes at a time. The scrapers attached to the mud hopper level the top of the mud mix and the mold is raised, leaving the bricks on the ground. The ready-mix truck and the attached adobe layer are moved forward. The mold is washed with a water spray and lowered for the next filling. After 2-3 days of drying, the bricks are turned up on edge, cleaned and scraped, and left in this position to dry thoroughly (30 days) or until shipment. This operation at La Luz (just north of Alamogordo) produces 2,000-3,000 adobe bricks per day with a crew of 3 employees.

Summary and conclusions

The growth of the adobe-brick industry in New Mexico depends upon three basic factors: 1) the ability to locate and secure a supply of low-cost adobe-soil materials from federal, state, or private lands; 2) acceptance by federal agencies (HUD and others) of the good physical qualities and advantages of adobe bricks; and 3) economic considerations related to production, transportation and the marketing of adobe bricks.

1) Adobe soil-resources

Results of this study show that adobe-soil materials are abundant in New Mexico, with extensive deposits located in the floodplains and alluvial-terrace deposits in the Albuquerque and Espanola Basin areas of the Rio Grande valley. Tests indicate that New Mexico adobe bricks have physical properties adequate to meet federal and state building codes. Adobe bricks can be produced in large quantities (150,0001,000,000) at reasonable costs with limited equipment by using the semimechanized and mechanized methods of production. In 1980, 67% of adobe-brick production consisted of semistabilized and stabilized adobe bricks using asphalt emulsion as the water-resistant, stabilizing material.

2) Acceptance of adobe by federal agencies

The use of adobes by the Spanish and Indian population of the Southwest has long been of historical and cultural importance. Recently, considerable discussion has centered around the thermal performance of adobe, the wages paid adobe layers because of the Davis-Bacon Act, and the overall costs of adobe construction. To answer one of the major questions affecting the use of adobes, the U.S. Department of Energy, U.S. Department of Housing and Urban Development, University of New Mexico's Energy Institute, Tesuque Pueblo, and the Eight Northern Indian Pueblos Council funded the Southwest Thermal Mass Study, Adobe Research Facility Phase in 1980. This project seeks to identify the characteristics of thermal mass with emphasis on adobe. Objectives of the project are: 1) to measure thermal performance of mass walls as components in buildings; 2) to measure effects of energy consumption of mass-wall coupling with thermostat setpoints and ventilation; 3) to measure U-values of stabilized, semistabilized, and traditional adobe including changes due to curing effects; and 4) to compare experimental results with predictions from techniques such as effective U-

TABLE 3—List of adobe-brick producers active in 1980; for locations see fig. 1.

Map no.	County	Operator	Telephone (A.C. 505)	Approximate annual production	Production equipment
1	Bernalillo	New Mexico Earth P.O. Box 10506 Alameda Richard Levine, owner	898-1271	700,000	Semimechanized, front-end loader, pugmill, wooden forms, delivery trucks
2	Bernalillo	Adobe Enterprises, Inc. 6000 Powerway, S.W. Albuquerque Ernest Sanchez, owner	877-4315	500,000	Semimechanized, front-end loader, wooden forms, delivery trucks
3	Rio Arriba	Eight Northern Indian Pueblos Council Box 969 San Juan Pueblo Dennis Duran, manager	753-4846	436,000	Mechanized, front-end loader, pugmill, adobe layer, delivery trucks
4	Otero	The Adobe Patch Rt. 1, Box 240 Alamogordo Robert Godby– Howard Scoggins, owners	437-9932	350,000	Mechanized, front-end loader, ready mixer, adobe layer, delivery truck
5	Santa Fe	The Adobe Farm Rt. 1, Box 193B Española Ralph Rivera, owner	753-4628	300,000	Semimechanized, front-end loader, ready mixer, wooden forms, delivery truck
6	Santa Fe	Victor Montano Rt. 6, Box 79A Santa Fe	471-2038	265,000	Semimechanized, front-end loader, wooden forms, delivery truck
7	Bernalillo	Western Adobe 7800 Tower Road, S.W. Albuquerque Dean Leach, owner	836-1839	250,000	Mechanized, front-end loader, adobe layer, delivery truck
8	Santa Fe	Eloy Montano 523 Barela Ln. Santa Fe	983-2172	235,000	Semimechanized, front-end loader, wooden forms, delivery trucks
9	Bernalillo	Manual Ruiz P.O. Box 104 Corrales	898-1913	175,000	Semimechanized, front-end loader, wooden forms, delivery trucks
10	Valencia	Rio Abajo Adobes 105 W. Aragon Belen Jerry Sanchez, owner	864-6191	150,000	Semimechanized, front-end loader, wooden forms, delivery trucks

values, M-factors, and steady-state U-values (a U-value is the coefficient of heat transmission expressed in units of BTU per hr per ft' per °F).

3) Transportation and marketing

Information obtained from adobe producers, architects, builders, and others indicates wide acceptance of adobe bricks in New Mexico, with the total production of the manufacturers usually sold by the end of November of each year. The 1980 summer market value for the three major types of adobes averaged $266.00 per thousand for the traditional adobe bricks, $278.00 per thousand for the semistabilized adobes, and $369.00 per thousand for the stabilized adobes. Transportation costs averaged $100-$200 per thousand for delivery within a 100-mi radius of the adobe-production yard.

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SCORIA EXPLORATION AND UTILIZATION IN NEW MEXICO

by JoAnne C. Osburn, *New Mexico Bureau of Mines & Mineral Resources, Socorro, New Mexico 87801*

Abstract

The large number of scoria or cinder cones in New Mexico and the relative ease of scoria mining make scoria an attractive source of lightweight aggregate. New Mexico presently ranks fifth in United States scoria production behind Arizona, Oregon, California, and Nevada. Currently, 11 scoria deposits are being commercially mined in New Mexico. The majority of New Mexico's scoria output is used to make concrete to produce strong, lightweight cinder blocks. The second most important market for scoria is as decorative stone used in desert landscaping and roofing material. Both of these uses require more control of color and size than the aggregate used for block manufacture; these controls create a higher price for landscaping and roofing scoria. Dense lava-flow materials that are waste products in the lightweight-aggregate industries are used for ballast and erosion control. Geological methods can be effectivly used in scoria-cone mining during both cone exploration and mine planning to ensure more consistent and better grade control and, ultimately, a more profitable mining operation. The ratio of height to basal diameter (aspect ratio) of the cone falls within discrete limits of 0.1-0.2 for all potentially commercial scoria cones in New Mexico. Mining only the deposits within this range would lower the number of lava flows and amount of ash encountered, thus upgrading the quality of material and lowering costs. Color and texture of the aggregate greatly influence marketability and can be largely controlled by mining the appropriate portion of the cone. Scoria cones are relatively simple structures, with most physical and chemical properties directly related to the position of the sample with respect to the vent. Scoria cones exhibit a vent-centered, concentric color pattern. Tephra are dark red brown in the vent area, dark gray with iridescent surface coatings at intermediate distances, and very dark gray at the perimeter of the cone.

Introduction

Scoria or volcanic cinder is a dark-colored, vesicular rock formed when gases, especially water vapor, expand during a volcanic eruption. Scoria piles up around the vent, producing the scoria cone. The classic deposit is cone shaped with beds farther from the vent dipping outward and beds near the vent dipping inward. The actual shape of the deposit is determined by a number of factors, including wind direction and intensity, vent shape, and lava viscosity. Noteworthy variations in form occur in many of New Mexico's scoria cones.

Strong winds during the eruption cause reduced deposition on the windward side of the cone; the resultant lopsided form is common in the cones of the Potrillo volcanic field in south-central New Mexico (fig. 1). Twin Mountain in northeast New Mexico erupted from a fissure vent to form an elongate cone.

Production

All U.S. scoria production comes from the western states (fig. 2). New Mexico ranks fifth behind Arizona, Oregon, California, and Nevada; these five states produce about 85% of all scoria mined each year. New Mexico, as the easternmost, major producing state, has cheaper transportation costs to eastern markets than other major producing states. New Mexico operators currently ship by rail to Illinois, Texas, and Wyoming; most of the material shipped out of state is used as decorative stone in landscaping.

FIGURE 1—LOCATION OF NEW MEXICO'S MINABLE SCORIA DEPOSITS.

New Mexico's scoria resources are estimated to be 320 million yds^3, near major roads and railroads (Osburn, 1980). Scoria deposits, found in widely separated parts of New Mexico (fig. 1), are of Tertiary or younger age. Currently, 11 deposits are being mined by commercial operators in the state; at least four other deposits are sporadically used for maintenance of roads and railroad beds on public, state, or railroad lands. Active mines are located either near major cities like Albuquerque, Santa Fe, and Las Cruces, or near railroads.

New Mexico currently contributes approximately 14% of U.S. production or about 448,000 tons annually (fig. 3). Scoria production in New Mexico has been closely linked to fluctuations in the economy through the years. Railroad ballast was the main use in the early 1950's but, in the late 1950's, roadbuilding and block manufacture became more important. From about 1961 to the present, the principal use (approx-

FIGURE 2—MAJOR AND MINOR SCORIA-PRODUCING STATES.
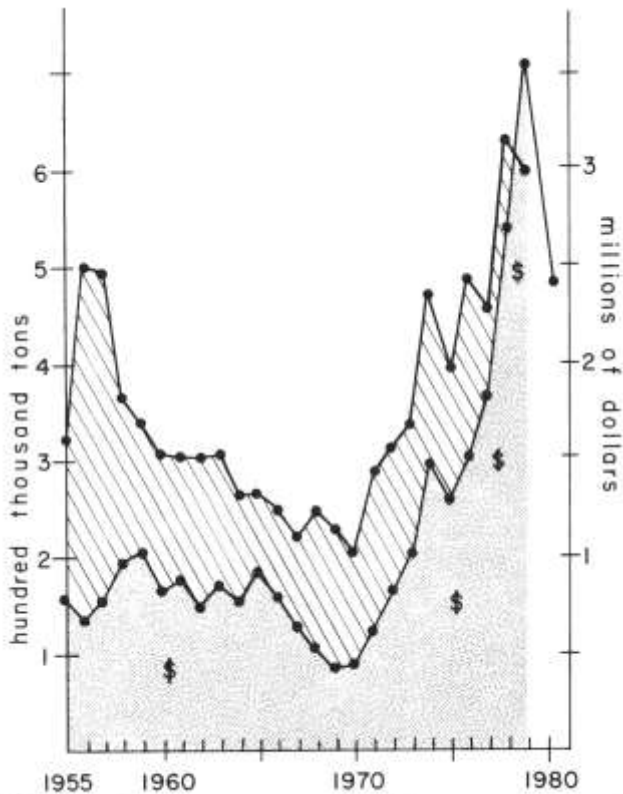
FIGURE 3—SCORIA PRODUCTION IN NEW MEXICO BY WEIGHT AND DOLLAR VALUE.

imately 35%) of scoria mined in New Mexico has been in concrete admixtures, especially in cinder-block manufacture. The commercial use of scoria as decorative aggregate was first recorded in 1961 (U.S. Bureau of Mines, Mineral Yearbook). Overall production slowed in the 1960's, reflecting a reduced demand for road metal and ballast, and slow construction periods. Both the construction industry and the lightweight-block industry were growing rapidly in the 1970's as people migrated to the southwest. Desert landscaping became very popular during this period. However, in 1980 and 1981 the construction industry was depressed due to high interest rates and a subsequent drop in housing starts.

Utilization

The principal use of scoria in New Mexico is in concrete admixtures used in the production of lightweight cinder blocks. Cinder-block-manufacturing plants operate in Santa Fe, Las Cruces, and Albuquerque; a number of scoria deposits are used exclusively for feedstock for these plants. At the mines, the aggregate is dug with tractors and rippers and then stockpiled. Scoria is crushed to size and screened at either the mine or the blockplant and then mixed with cement and water at the blockplant. The blocks are pressed and stacked using a largely automated system of presses, lifts, and conveyor belts. Blocks are cured in a moist environment for 24 hrs and then cured in the yard for three weeks prior to shipping. As with any other type of cement block, a large variety of colors and shapes are produced. Exposed aggregate and geometric designs in blocks contribute pleasing qualities to a building and are becoming more common at the expense of plain, standard blocks that contribute only structural strength.

Scoria blocks are an excellent building material, especially for institutions. Blocks made from scoria have greater structural strength than those made from pumice, mainly because of thicker cell walls in the rock. In addition, scoria blocks are sawable, nailable, vermin-proof, fireproof, have good insu

lating properties and are difficult to vandalize (Schmidt, 1957).

The popularity of scoria as decorative stone in the desert landscaping industry is largely due to low maintenance and water requirements. In 1980, 28% of scoria produced in New Mexico was used in landscaping, compared to approximately 5% used nationwide. This landscaping style is gaining acceptance in building projects ranging from college campuses to fast-food restaurants.

The most common sizes of scoria used in landscaping are ³/4-inch- and 1 3/4-inch clasts and large decorative blocks and bombs. Landscaping aggregate requires careful size and color control and thus commands much higher prices than scoria used for cinder block. Prices in 1981 for decorative stone in New Mexico ranged from $12 to $14 per yd^3; the reddish-brown cinder is generally more marketable for landscaping than the darker gray material.

Block production and landscaping material use approximately 70% of the total scoria sold in New Mexico. Other uses include roofing, railroad ballast, erosion control, and road metal. The use of road metal, much more important during the massive roadbuilding era of the 1960s, is now reduced to road maintenance.

Geological-exploration techniques

Inspection of about 75 scoria cones and further study of four partially dissected deposits have revealed a predictability in the physical properties of potentially minable deposits. The physical properties studied included the aspect ratio of the deposit and color, sorting, and grain size of the tephra. Knowledge of each of these parameters could significantly lower both exploration and mining costs.

Aspect ratio (Green, 1975) is defined as the ratio of height to average basal diameter of a volcano. A plot of all the scoria cones in New Mexico with exposed, good-quality scoria shows that all have aspect ratios between 0.1 and 0.2 (fig. 4). In general, cones with aspect ratios lower than 0.1 have thick lava flows included in them and approach the general configuration of shield volcanoes; those cones with aspect ratios greater than 0.2 tend to contain large amounts of agglutinate and approach the form of spatter cones. Agglutinate consists of scoria blocks stuck together with flow material. Large amounts of flow material and agglutinate must be blasted, but blasting increases costs and mining time and decreases mine safety. Calculation of the aspect ratio requires only a topographic map and should be determined before digging the first test pit.

All measurable physical properties of volcanic ejecta vary with respect to the position of the vent. One of the most strik-
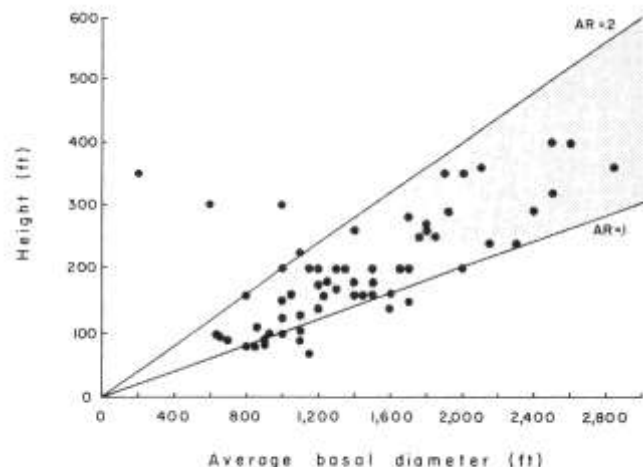
FIGURE 4—ASPECT RATIOS FOR POTENTIALLY MINABLE NEW MEXICO SCORIA CONES.

ing patterns is the concentric color zonation around the vent. Scoria tends to be dark reddish brown in the vent zone, brown to dark gray with iridescent surface coatings at intermediate distances, and very dark gray to black around the perimeter of the cone.

Chemical analyses of samples collected at regular intervals from the vent zone to the perimeter zone show a progressive decrease in the ferric-to-total-iron ratio moving away from the vent. Ferric iron constitutes about 95% of the total iron in the vent zone and decreases to only $5^{\%}$ at the perimeter of the cone. No significant changes in other major element chemistries is observed.

Heat at the vent is postulated as the cause for oxidation. When black scoria samples were heated to 400°C in a muffle furnace, sample color began to change progressively after only 1 hr. Samples develop iridescent blue and green surface coatings after about 24 hrs of heating; these samples resemble scoria found at intermediate distances from the vent. After about four days of heating, the color of the samples stabilizes at weak red or dark reddish brown (Munsell Color Comparison Chart). Tests run at 700 °C showed the same color progression, but the colors stabilize at dark reddish brown after 8 hrs rather than the 84 hrs required at 400°C. No color changes are noted in samples heated at temperatures lower than 400°C for a period of two weeks. In short, by heating scoria for variable amounts of time, the vent-centered color pattern of a scoria cone can be duplicated. The typical pattern observed is dark-reddish-brown scoria in the vent zone, where maximum heat exposure occurs, brownish to dark-gray scoria at intermediate distances from the vent, and very dark gray to black scoria in perimeter zones, where volcanic ejecta are well insulated from the vent. As the demand for dark-reddish-brown scoria for landscaping aggregate expands, the value of being able to predict color variation in a particular deposit becomes more important.

Both sorting and grain size of the tephra reflect the sedimentary nature of the deposit. All volcanic ejecta emanate from a point source, the vent. Tephra in the vent are the largest and most poorly sorted in the deposit. Agglutinate also occurs in the vent area. In contrast, at the perimeter of the cone, scoria from each eruptive pulse is well sorted and graded. The average grain size is small, with the exception of an occasional block or bomb.

Conclusions

Scoria cones have been successfully quarried for many years without using formal geologic tools. However, understanding the predictable spatial relationships among the different materials in the deposits could make mining operations more cost efficient. The potential market of scoria is dependent on both size distribution of the clasts and color of the material. Screening and crushing processes largely control the size of the aggregate shipped. However, by predicting the degree of sorting and average grain size, one can significantly lower the amount of flow material and ash that lower the value of the product. The color of volcanic aggregate is becoming important, especially in the landscaping and roofing industries. Knowledge of the characteristic color zoning in scoria cones could help predict locations of various colors within a given cone, and lower the number of test pits necessary. Selecting a cone with an aspect ratio of 0.1 to 0.2 is very important. This simple test alone could lower the number of cones in the state that have been eliminated from further testing due to the anticipated large amounts of waste materials.

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GEOLOGY OF BARITE IN NEW MEXICO

by Thomas J. Smith, *Statex Petroleum, Inc., Oklahoma City, Oklahoma 73102*

Abstract

With few exceptions, barite deposits in New Mexico occur in a north-south-trending belt that stretches for more than 200 mi along the Rio Grande rift zone. The most significant ore bodies tend to be located near the linear extremities of fault-block mountain ranges, areas characterized by numerous normal faults. All of the deposits that have been of economic interest are vein and cavity-filling deposits, generally found in Paleozoic carbonate rocks. Galena and fluorite are the minerals typically associated with the barite, but quartz, jasperoid, and oxidized copper minerals are not uncommon.

Introduction

In recent years a dramatic increase in domestic drilling activity by the petroleum industry has occurred. As a consequence, the demand for barite has also risen sharply. Nevada is currently the leading state in barite production, accounting for more than 80o of the total tonnage produced in the United States. In 1979, New Mexico recorded barite production for the first time since 1964. Actual tonnage, although quite small, illustrates the renewed interest in New Mexico barite. At present, at least three operations are actively exploring for and exploiting barite deposits within the state.

All of the known barite deposits in New Mexico may be categorized as hydrothermal vein and cavity-filling type deposits according to Brobst's (1975) classification. Although many of the deposits exhibit minor amounts of replacement, significant replacement occurs at only a few sites. Barite-bearing veins in New Mexico are found in all three major rock types but are most common in sedimentary rocks. In fact, the larger deposits appear to be associated exclusively with carbonate rocks of Paleozoic age.

Table 1 lists the minerals most commonly associated with barite deposits in New Mexico. Minerals are listed in order of decreasing frequency of occurrence, rather than decreasing abundance. Generally, though, fluorite is the most abundant accessory mineral, especially in veins in crystalline Precambrian rocks.

Almost all of the major New Mexico barite deposits occur in a 250-mi-long, north-south-trending belt that generally coincides with the Rio Grande rift zone (fig. 1). In general the veins are associated with fault or shear zones located at or near the linear extremities of fault-block mountain ranges and in areas characterized by numerous normal faults. Although most of the veins seem to be within faults that crosscut bedding, the larger deposits tend to be relatively concordant bodies, often occurring along bedding planes or formational contacts.

Access to barite deposits in New Mexico and proximity to milling facilities are almost invariably poor; consequently, the

size of the deposits is often a controlling factor for economic interest. New Mexico is not blessed with large bedded deposits like those of Nevada; on the contrary, the large orebodies in New Mexico commonly are only a few tens of thousands of tons. Three or four exceptional deposits may contain up to several hundreds of thousands of tons each, but the prospects for successful exploration in New Mexico appear to be limited to smaller operators. In the following deposit descriptions, estimated tonnages have been omitted to maintain propriety.

Barite deposits

The following descriptions of the 13 most significant barite deposits or groups of deposits known in New Mexico are based upon past production, potential production, or both. Table 2 lists properties along with the locations and types of the deposits. Deposits are listed as either vein or replacement depending upon the apparent mechanism responsible for ore deposition.

The northernmost barite deposit is the Landsend deposit (fig. 1, locality **1),** located at an elevation of over 9,000 ft near the north end of the Sandia Mountains in what is now a wil-

TABLE 1—MINERALS ASSOCIATED WITH NEW MEXICO BARITE DEPOSITS.

Minerals
1. Fluorite
2. Quartz
3. Galena
4. Chert
5. Oxidized-copper minerals
6. Jasperoid
7. Iron oxides
8. Calcite
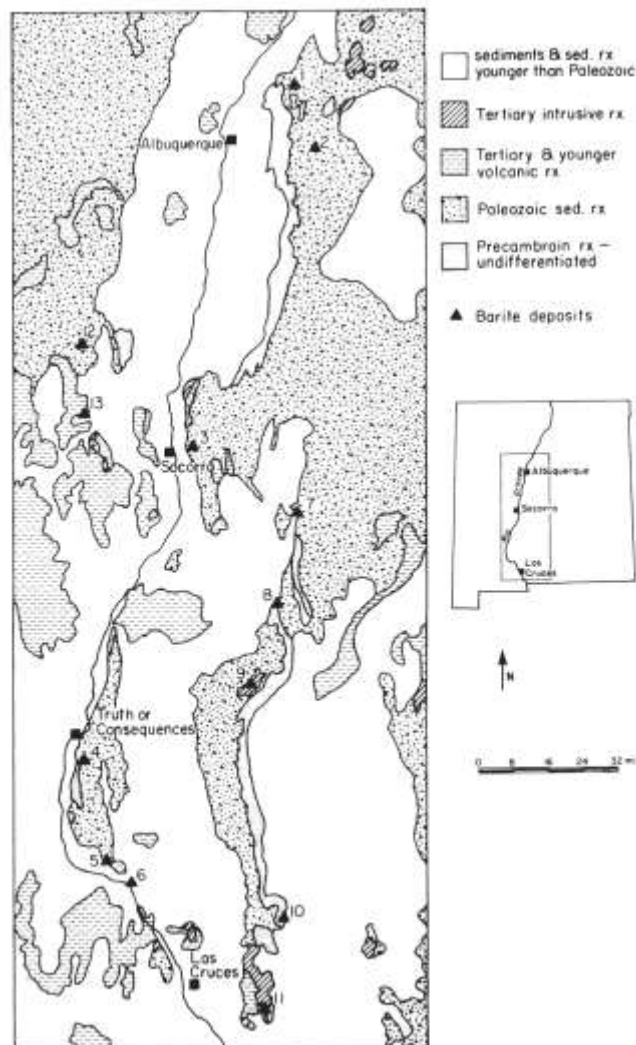
FIGURE 1—GENERALIZED GEOLOGIC MAP OF THE RIO GRANDE REGION WITH LOCATIONS OF BARITE DEPOSITS LISTED IN TABLE 2.

TABLE 2—NEW MEXICO BARITE DEPOSITS CITED IN TEXT. Locality numbers refer to site locations in fig. 1.

Locality	Deposit(s)	Location	Description
1	Landsend	T. 12 N., R. 5 E.	vein; Madera Limestone (shear zone)
2	Tina & Shockley	T. 9 N., R. 7 E.	veins; Madera(?) Limestone (fault zones)
3	Gonzales	T. 3 S., R. 1 E.	vein; Sandia Formation and Precambrian granite (fault zone)
4	Carolyn	T. 15 S., R. 3 & 4 W.	vein; Abo Formation
5	Palm Park	T. 18 S., R. 3 W.	replacement; Fusselman Dolomite (formational contact shear zone)
6	Tonuco	T. 19 & 20 S., R. 1 W.	veins; Precambrian granite
7	Hansonburg	T. 5 S., R. 5 & 6 E.	vein and replacement; Council Springs Limestone (fault and shear zones)
8	Independence	T. 9 S., R. 5 E.	vein; El Paso(?) Limestone (fault zone)
9	Salinas Peak	T. 11 & 12 S., R. 4 E.	vein; Tertiary rhyolite sill and Panther Seep Formation (fault and shear zones)
10	Stevens	T. 20 S., R. 5 E.	replacement; El Paso Limestone (fault zone)
11	Devil's Canyon	T. 23 S., R. 4 E.	replacement; brucite-serpentine marble
12	Katherine	T. 2 N., R. 4 W.	vein; Yeso Formation (fault zone)
13	North Magdalena	T. 2 S., R. 4 W.	veins; Tertiary volcanics (fault zones)

derness area. Barite mineralization occurs as a breccia cement and cavity-filling with minor replacement in a bedding-plane fracture zone in Madera Limestone (Pennsylvanian). Lesser amounts of honey-yellow, clear fluorite and quartz are also present, along with minor galena. The barite is white and very coarsely crystalline and occurs in massive, tabular aggregates varying in thickness from 1 to 4 ft. These layers are commonly diluted with large inclusions of jasperoid, chert, or limestone host rock. Numerous pits and trenches expose the ore zone for a length of more than 400 ft.

The Tina and Shockley deposits (fig. 1, locality 2) are situated less than a mile from each other in a hilly, wooded area approximately 25 mi east of Albuquerque. Each deposit consists of an east-west-trending vein in limestones that are probably of the Madera Formation. Both veins are in fracture zones and contain breccia fragments of the wall rock. Barite is associated with fluorite, quartz, and minor galena at both sites. The veins appear to vary considerably in width along their strikes, but the width of barite mineralization averages less than a foot.

Five miles east of Socorro, across the Rio Grande, is the Gonzales deposit (fig. 1, locality 3). According to Rothrock and others (1946), the deposit is located along a normal fault. A small elongated block of pink Precambrian granite is on the eastern upthrown side and limestones and sandstones of the Sandia Formation are to the west. Mineralization predominantly occurs within the brecciated sedimentary rocks. Two major ore shoots, one 240 ft long and the other 280 ft long, constitute the bulk of the deposit. Ore-zone widths range from 3 to 22 ft. Barite occurs as large, bladed crystals intimately intergrown with abundant fluorite. Quartz is also abundant in the vein, especially near the footwall.

The Carolyn group of claims (fig. 1, locality 4) is located approximately 7 mi southeast of Truth or Consequences, on the east side of the Caballo Mountains. The deposit consists of two vertically dipping veins that strike N. 50° E. through red sandstones and limestones of the Abo Formation (Permian). Barite is the dominant mineral in the veins, along with minor amounts of fluorite and quartz. According to Williams and others (1964), one of the veins, exposed 60 ft below the surface in a vertical shaft, is as much as 4 ft wide; one sample from that location contained 88.5% barite.

The Palm Park mine (fig. 1, locality 5; fig. 2) is in the southern foothills of the Caballo Mountains; at the time of the field work for this report this mine was being developed by Barite of America, Inc. Barite mineralization occurs as numerous, discontinuous, replacement lenses and cavity fillings within highly silicified Fusselman Dolomite (Silurian) at and near the unconformity of the dolomite with the overlying Percha Shale (Devonian). The ore zone is in the eastern, uplifted block of the Palm fault (Kelley and Silver, 1952). Although the ore is diluted with varying amounts of jasperoid, fluorite, and quartz, and minor amounts of calcite, galena, and iron

oxides (fig. 3), this deposit represents one of the larger barite deposits in the state.

The Tonuco mine (fig. 1, locality 6) consists of workings in two fluorite-barite veins that occur as fissure fillings in fractures within Precambrian granite and schist near the base of San Diego Mountain. Recorded production has been limited to a few thousand tons of fluorspar ore. Aggregates of clear, tabular crystals of barite are commonly intergrown with the fluorite, and often account for as much as 50% of the ore material. According to Williams and others (1964), the two minerals were easily separated at a nearby mill, but the disposition of the barite is unknown.

The Hansonburg mining district (fig. 1, locality 7), located at the northern end of the Oscura Mountains, is by far the most important producer of barite in New Mexico, having accounted for more than 90% of all recorded production in the state. Substantial reserves remain in the area; however, Minopco, Inc. of Socorro currently has the entire district under lease, and is developing selected orebodies.

Nearly all of the Hansonburg deposits occur as veins in fault breccias or as replacement bodies in sheeted fracture zones in the Council Springs Limestone. Fluorite, barite, quartz, calcite, and galena are the most common minerals, although oxidized coppper minerals are also present. At the Mex-Tex and Barite workings, the most productive barite deposits in the district, barite occurs as concordant veins along bedding-plane fractures. Both veins consist of central zones of barite lined on either side with fluorite, galena, and minor amounts of quartz.

The Independence mine (fig. 1, locality 8) is located within the White Sands Missile Range at the extreme northern end of the San Andres Mountains. Barite occurs in the El Paso Limestone as minor gangue of lead-zinc ores in a silicified and

FIGURE 2—BARITE OUTCROP 4 FT THICK AT SALINAS MINE.

FIGURE 3—Typical box-work barite crystals in Fusselman Dolomite, Palm Park deposit (Barite of America); outcrop shown is 0.75 m wide.

brecciated fault zone that strikes nearly due north. Although the workings do not suggest significant mineralization, unconfirmed reports imply that the vein extends northward under alluvium and contains considerably more barite than surface indications suggest.

Approximately 20 mi south of the Independence mine is Salinas Peak (fig. 1, locality 9). At nearly 9,000 ft, it is the most prominent feature of the northern San Andres Mountains. An extensive Tertiary rhyolite sill forms the uppermost 1,000 ft of the peak. Associated with the contact between the sill and limestones of the Panther Seep Formation are three relatively significant barite deposits, referred to as the Unnamed mine, Salinas mine, and Section 9 prospect respectively by Williams and others (1964). Both the Unnamed and Salinas mines orebodies are concordant veins within shear zones at the limestone-rhyolite contact. Barite occurs with abundant fluorite and locally abundant galena at the two deposits. In contrast, the vein at the Section 9 prospect occupies a north-south-trending fault that has dropped the limestone-rhyolite contact to the east. Fluorite and galena are present only in trace amounts, and barite constitutes nearly all of the vein material.

The Stevens mine (fig. 1, locality 10) is located at the southern end of the San Andres Mountains and, like the Independence mine and the deposits at Salinas Peak, lies within the White Sands Missile Range. Barite occurs in a group of replacement deposits associated with a low-angle fault contact between carbonate rocks of the El Paso Formation (Ordovician) and Precambrian granite. Minor amounts of fluorite and galena occur with the barite, along with traces of cerrusite, anglesite, wulfenite, and vanadinite. The Stevens mine is noteworthy because it has recorded barite production (Dunham, 1935).

The Devil's Canyon mine (fig. 1, locality 11), at the head of Target Range Canyon near the southern end of the Organ Mountains, could not be visited during the course of this study because it lies within the artillery range target area of Fort Bliss Military Reservation. This deposit is relatively important, however, because of recorded barite production (Dunham, 1935). From Dunham's description, barite occurs with minor fluorite as replacement deposits in brucite-serpentine marble.

Two other New Mexico localities deserve mention. The Katherine group of claims (fig. 1, locality 12) is in the northern Bear Mountains, approximately 20 mi north of Magdalena. Barite occurs in a brecciated fault zone in silicified limestones of the Yeso Formation (Permian). Because the general geology of this area is similar to other areas that have produced barite, the Katherine deposit may be favorable to further prospecting on a small scale.

Less than 2 mi north of Magdalena are numerous small workings on veins in fault breccias of Tertiary volcanic rocks (fig. 1, locality 13). Interest in the area has been renewed recently because many of the veins are known to contain a high percentage of barite. Most of the workings consist of inaccessible vertical shafts, so the actual extent of the deposits is unknown; however, unconfirmed reports suggest a possible widening to minable widths at depth in at least one of the veins.

Summary

The larger barite deposits in New Mexico appear to be characterized by two major factors: 1) structural setting and 2) host rock. Favorable structural settings are areas of numerous normal faults that occur near the extremities of fault-block mountain ranges. All of the ranges lie along the eastern boundaries of basins associated with the Rio Grande rift.

Host rocks for barite-bearing veins are almost always Paleozoic carbonates. That the age of the host rocks contributes to the favorability for barite mineralization is unlikely, because deposits are known in carbonate rocks ranging in age from Cambrian-Ordovician to Permian. Instead, the petrology probably contributes to the favorability of the host rock. Whether a chemical or structural feature in the carbonates favors barite mineralization is difficult to determine.

Aside from the foregoing characteristics for the larger New Mexico barite deposits, with few exceptions the deposits tend to be no greater than a few tens of thousands of tons in size. Although the possibility always exists that very large barite orebodies will be discovered in the future within the state, a more likely expectation is that exploration efforts will reward those operators who work on a relatively small scale.

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ZEOLITE DEPOSITS IN THE GILA AND SAN SIMON VALLEYS OF ARIZONA AND NEW MEXICO

by Ted H. Eyde, *GSA Resources, Inc., Cortaro, Arizona 85230*

Abstract

Several deposits of zeolite minerals occur in Cenozoic rocks in the Gila and San Simon Valleys both in New Mexico and Arizona. Zeolite deposits occur not only in the lacustrine beds within the valleys but also in vitroclastic tuff beds intercalated with the volcanic rocks that form many of the mountain ranges bounding the valleys. Though the geological processes that formed these two types of deposits are diverse, the actual geochemical reactions involved in zeolitic alteration are nearly identical. Zeolite minerals in both types are alteration products resulting from the reaction between volcanic glass and saline-alkaline waters. The saline-alkaline lake deposits that are the subject of this paper are the most important of the two types. These deposits consist of one or more stratiform beds of high-purity zeolite minerals interbedded in lacustrine beds. The deposits are the alteration product of volcanic ash initially ejected into the atmosphere by distant volcanism during Pliocene-Pleistocene time. The vitric phase of the airborne volcanic ash was carried large distances by the prevailing winds and eventually deposited in shallow lakes in the Gila and San Simon Valleys. Following deposition, the ash reacted with the saline-alkaline lake waters to form zeolite minerals. Sedimentary processes in the shallow, saline-alkaline lakes were active before, during, and following the deposition of the vitric ash. Consequently, the degree of zeolitic alteration as well as the thickness and areal extent of the zeolite beds are directly related to the depositional history of the basins. Two zeolite deposits in the San Simon and Gila Valleys are in production. The zeolite deposit near Bowie, Arizona has produced chabazite used to purify sour natural gases and also to remove cesium 137 and strontium 90 from liquid nuclear reactor wastes. The zeolite deposit near Buckhorn, New Mexico, produces clinoptilolite used to denitrify sewage effluent and to produce organic fertilizer from poultry manure. The chab

azite deposit in the Dripping Springs Valley has been explored by drilling. Several unexplored chabazite deposits occur in the vicinity of Bear Springs, near Pima, Arizona.

Introduction

Zeolites are a group of 34 separate mineral species which are crystalline, hydrated aluminosilicates that contain exchangeable alkali or alkaline-earth cations. Extensive deposits of high-purity zeolite minerals occur in both Cenozoic age volcanics and sedimentary rocks of volcanic origin. Both natural zeolite minerals and their synthetic counterparts, which are known as molecular sieves, are used in pollution control, radioactive-waste management, petroleum refining, and gas-purification processes. Natural zeolite minerals are also used in construction materials, in carriers, fillers, and coatings, aquaculture, and agriculture. Major future uses will include waste-water treatment, solar energy, and synthetic-fuel production.

Zeolite deposits of the Gila and San Simon Valleys

The zeolite mineral chabazite was tentatively identified by Lowe (1875) in an outcrop of volcanic tuff about 4 mi south of Whitlock Cienega. This is believed to be the first published account describing a bedded zeolite deposit anywhere in the world. Not until 1959 was this same chabazite deposit, now known as the Bowie deposit, rediscovered by geologists of the Linde Division of Union Carbide Corporation. The Dripping Springs, Bear Springs, and Buckhorn deposits were discovered during the same exploration program during 1960 and 1961. All of these deposits occur in the basins filled with late Tertiary lacustrine sediments in the Gila and San Simon River drainages in Arizona and New Mexico (fig. l). A more de-
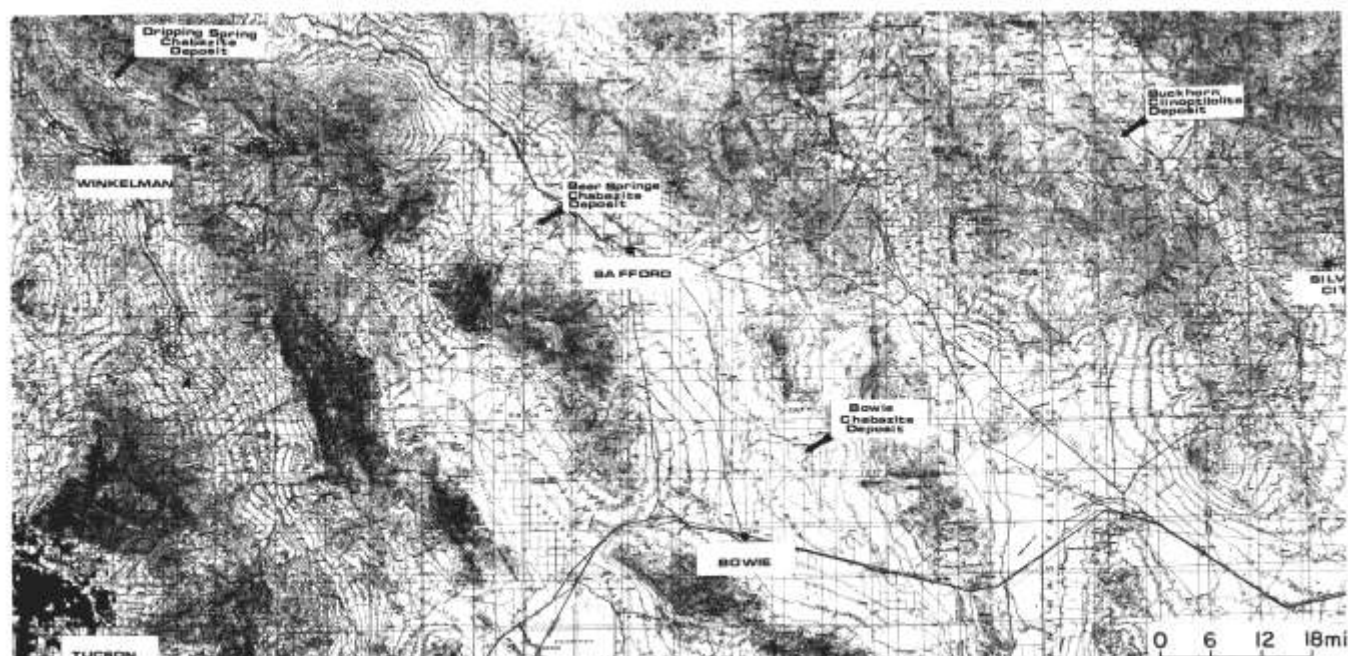
FIGURE 1—LOCATION MAP OF THE ZEOLITE DEPOSITS IN THE SAN SIMON AND GILA RIVER VALLEYS.

tailed exploration program in these basins would certainly result in the discovery of additional bedded zeolite deposits.

Extensive deposits of the zeolite minerals clinoptilolite and mordenite occur in the Cretaceous and early Teritary volcanics in the Whitlock Mountains, Blue Range, Mogollon Mountains, and Coronado Mountains. The deposits were formed from the alteration of thick vitrophyre, vitric tuff, and vitroclastic tuff members in the volcanics. Though the deposits are large, most of them would require extensive beneficiation to produce a marketable product. Even though the geochemical reactions involved in the zeolitic alteration are nearly identical, the origin of these deposits differs significantly from that of the deposits formed in a saline- or alkaline-lake environment. Discussion of these deposits is beyond the scope of this paper.

Several zeolite deposits occur in Tertiary lacustrine beds exposed in the San Simon and Gila River drainages of southeast Arizona and southwest New Mexico. The most important deposits are the Dripping Springs Valley chabazite deposit near Winkleman, Arizona, the Bear Springs chabazite deposits near Pima, Arizona, the Bowie chabazite deposit near Bowie, Arizona, and the Buckhorn clinoptilolite deposit near Buckhorn, New Mexico. These deposits formed during Pliocene-Pleistocene time when vitric ash ejected by volcanic activity in the far western United States was carried eastward by the prevailing winds and deposited in a series of saline-alkaline lakes that occupied the present San Simon, Gila River, and adjoining valleys.

The vitric ash reacted with the saline-alkaline waters contained in the lake to form zeolite minerals. The size of the vitric ash particles apparently is inversely proportional to the distance from the source. Thus, the more distant the source, the smaller the individual particles. In addition, the smaller the particle size, the larger the surface area of vitric ash available for reaction with the saline-alkaline water. In general, the zeolite deposits in the San Simon and Gila River valleys contain more chabazite and are significantly thinner than similar deposits in California, Oregon, and Nevada. This may be because the deposits in the San Simon and Gila River valleys were farther from the source of the volcanic ash and, consequently, the basins received a thinner fall of finer-grained ash. It may also explain why the zeolite deposits are generally thinner, the size of the individual crystals smaller, and the deposits of higher purity. However, it does not necessarily explain why chabazite is often the most common zeolite mineral in these deposits.

DRIPPING SPRINGS VALLEY, ARIZONA CHABAZITE DEPOSIT

A chabazite-bearing tuff horizon is exposed in a road cut on the east side of Arizona State Route 77 about 23 mi south of Globe, Arizona. The Anaconda Minerals Company acquired the deposit in 1978 and since then has drilled over 100 holes to explore it.

The chabazite horizon is interbedded in a section composed of brown mud, sandstones, and limestones of probable Pliocene-Pleistocene age. The lacustrine beds are nearly flat-lying but dip gently toward the center of the Dripping Springs Valley. In the roadcut the chabazite horizon consists of three lithologies: a lower thin-bedded to platy bed, a middle massive bed, and an upper thin-bedded to platy bed. Total thickness of the chabazite is nearly 2 ft. Sampling done in 1960 indicated that the middle massive bed averages approximately 67% chabazite. Drilling by the Anaconda Minerals Company revealed that the deposit consists of two chabazitebearing horizons separated by what Anaconda geologists refer to as the red claystone bed which is about 5 ft thick.

The Anaconda drilling program showed that the lower zeolite horizon ranges from 0.50 to 1.74 ft thick and consists of three distinct lithologies that include a lower bed 0.15-0.87 ft

thick consisting of zeolitically altered vitric ash and about 10% detrital matter. This bed is overlain by and in sharp contact with a bed 0.14-0.20 ft thick, consisting of a laminated, thin-bedded, zeolitically altered vitric ash containing a trace of detrital matter in addition to a trace of clay and calcite. Overlying this bed is a massive bed 0.30-0.67 ft thick of zeolitically altered vitric ash. The bed grades upward into the overlying red claystone.

The lower zeolite horizon grades from a clean unaltered vitric ash at the north end of the deposit to 90% chabazite at the south end of the deposit. The only other zeolite mineral present in the deposit is a trace of clinoptilolite. A system of paleo channels that flowed across the deposit has removed large portions of the deposit.

BEAR SPRINGS, ARIZONA CHABAZITE DEPOSITS

Three chabazite-bearing tuff horizons crop out on the east side of the Gila River valley near Pima, Arizona. Two horizons crop out near Bear Springs in sec. 1, T. 7 S., R. 23 E. The other crops out in sec. 27, T. 6 S., R. 24 E. Heindl (1958) locates all three zeolite horizons in the Solomonsville beds (Pliocene-Pleistocene) which are lacustrine beds composed principally of brown mudstones with a few beds of limestone, sandstone, and conglomerate. Many of the brown mudstone beds contain unaltered, vitric-ash shards. The Solomonsville beds dip gently toward the center of the Gila River valley, the apparent center of the depositional basin.

The chabazite-bearing tuff horizons were vitric-tuff beds zeolitically altered to the minerals chabazite and erionite. As is the case at the Bowie chabazite deposit, the individual chabazite crystals are usually less than 0.20 microns in size. Because of this extremely small size, it is difficult to determine accurately the percentage of chabazite and erionite in the zeolite horizon. The most abundant zeolite mineral is chabazite followed by erionite, and occasionally small amounts of phillipsite.

Two chabazite-bearing tuff horizons separated by about 30 ft of brown mudstone with a few thin interbedded sandstone beds crop out at Bear Springs. The lower chabazite horizon averages about 1.5 ft in thickness, contains approximately 45% chabazite, and consists of two distinct beds. The lower is a massive, yellowish-white bed 0.50 ft thick that contains about 60% chabazite. Overlying this bed, but separated by a thin clay parting, is a massive yellowish-white bed 1.0 ft thick containing 40% chabazite. The best exposure of the lower chabazite horizon is in banks of the ditch that drains Bear Springs. The lower horizon can be traced by float and intermittent outcrops for nearly a mile along strike.

The upper chabazite horizon is a yellowish to grayish-white, platy to massive, zeolitically altered tuff bed that has an average thickness of about 1 ft and contains approximately 50% chabazite. In unweathered exposures along the road to Bear Springs, the chabazite horizon is composed of a lower massive bed 0.50 ft thick containing approximately 50% chabazite. The upper bed, which is separated from the lower by a thin clay parting, is also about 0.50 ft thick and contains approximately 60% chabazite. The upper chabazite horizon can be traced for about 1/2 mi along strike.

Another chabazite-bearing tuff horizon crops out about 4 mi northeast of Bear Springs and an estimated 100 ft downsection in the Solomonsville beds. The flat-lying horizon consists of a single yellowish to grayish-white tuff bed 1.0 to 1.5 ft thick that contains approximately 80% chabazite. The bed crops out continuously for about 500 ft along the strike.

BUCKHORN, NEW MEXICO CLINOPTILOLITE DEPOSIT

Two clinoptilolite-bearing tuff horizons crop out along the west side of Duck Creek valley in secs. 3, 4 and 10, T. 15 S.,

R. 18 W. The outcrops are visible approximately 1/4 mi west of U.S. 180 from a point about 1 mi south of Buckhorn, New Mexico. Double Eagle Petroleum and Mining Company acquired part of the deposit in 1974 and installed crushing and screening equipment. A small tonnage of sized clinoptilolite produced from the deposit has been sold for use in advanced waste-water treatment (AWT) plants to denitrify sewage effluent. The clinoptilolite has also been used in the production of an organic fertilizer from chicken manure. Leonard Resources acquired a portion of the deposit in 1977 but has not produced any clinoptilolite. No mining has been done at the deposit since 1977.

The clinoptilolite beds are in a nearly flat-lying sequence of lacustrine beds composed of green and brown mud and claystones. Heindl (1958) placed these sediments in the Cactus Flat beds of Pliocene-Pleistocene age. Olander and Surdam (1979), who investigated the mineralogy and depositional environment of the Buckhorn deposit, correlate the deposit with diatomite deposits about 3 mi north of Buckhorn that have been dated at the boundary between early and middle Pliocene or approximately 3 m.y. ago.

These investigators recognized four distinct depositional environments at the deposit: 1) a soil horizon possibly developed along the edge of an alluvial fan, 2) an intermittent stream-channel and adjacent overbank deposits, 3) interchannel, flood-plain deposits, and 4) lacustrine or ponded deposits. This confirms what has been recognized at other zeolite deposits—that sedimentary processes were active in the depositional basins before, during, and following the deposition of the parent vitric ash and also following zeolitic alteration.

The clinoptilolite-bearing tuff horizons are separated by approximately 20 ft of green and brown mudstone. The lower horizon is a yellowish-white massive bed of zeolitically altered vitric tuff ranging from 3 to 5 ft thick (fig. 2). Sampling done in 1961 indicated that the bed contained 70-90% clinoptilolite. More detailed sampling of the zeolite horizon revealed that a lower 0.50 ft bed appears to have been an exceptionally clean vitric, tuff prior to zeolitic alteration and contains approximately 48% chabazite and an equal amount of clinoptilolite. Several samples of the bed contained from a trace to more than 23% erionite. Olander and Surdam (in preparation) also indentified the zeolite minerals heulandite and analcime in addition to smectite/illite, chert, calcite, quartz, and minor amounts of detrital hornblende and biotite.

The upper zeolite horizon is a single 1 ft thick bed of massive, yellowish to grayish-white, zeolitically altered vitric tuff that contains over 60% clinoptilolite and a trace of erionite. Both the lower and upper zeolitic-tuff horizons can be traced for over 1 mi along strike.

BOWIE, ARIZONA CHABAZITE DEPOSIT

The Bowie chabazite deposit is approximately 13 mi north of Bowie, Arizona. The marker-tuff horizon that contains the high-grade chabazite bed at its base crops out discontinuously along the 3,440 ft contour line for approximately 7 mi along both sides of the San Simon Valley. Most of the outcrops and areas being mined are on the southwest side of the valley. Only a few scattered outcrops occur on the northeast side of the valley.

Lowe (1875), in possibly the first reference to a bedded zeolite deposit, described an outcrop of what appears to be the chabazite-bearing tuff horizon approximately 4 mi south of Whitlock Cienega. He described the zeolitically altered tuff as a yellowish, soft, porous material that cropped out for about
mi and concluded that the tuff was a mixture of quartz and a hydrous silica related to chabazite or stilbite. The deposit was literally rediscovered 84 yrs later in 1959 by geologists working for Union Carbide Corporation who were attempting

FIGURE 2—LOWER CLINOPTILOLITE HORIZON AT THE BUCKHORN, NEW MEXICO DEPOSIT. The basal 0.50-ft-thick bed clearly visible to the left of the hand contains approximately 48% chabazite.

to locate the source of a chabazite-bearing ornamental stone being quarried north of Bowie (Eyde, 1959).

Ernest Baugher, a retired railroad engineer from Buffalo, New York, found the outcrops of the chabazite-bearing tuff while rock-hounding in the area. He interested Frank Meadows and Paul Sanger, who both worked for the Southern Pacific Railroad, in the deposit, and the partnership staked the BMS claims that covered the outcrops in the central and northwest end of the deposit. Frank Clark, a butcher in Bowie and a part-time prospector, staked claims covering the southeast end of the deposit. Baugher and Clark quarried the lower massive bed of the marker-tuff horizon and fabricated bookends, lamps, paper weights, and pencil holders from it. These souvenirs made of what was known as "Arizona Tuffa-Tex" could be purchased during the 1950's and 1960's at stores and shops along the Tucson-El Paso highway.

The initial X-ray diffraction analyses of samples from the marker-tuff horizon indicated the presence of only small amounts of chabazite. However, oxygen adsorption analyses of several suites of samples collected from all the lithologies within the marker-tuff horizon revealed that the lower massive bed or the so-called high grade or ore bed was composed of high-purity chabazite (figs. 3 and 4). A subsequent field examination revealed that the lower massive bed could be traced intermittently for nearly 7 mi along the southwest side of San Simon Wash (fig. 5). A re-evaluation of the X-ray diffraction analyses revealed that the microcrystallinity of the chabazite in the marker-tuff horizon produced low, broad peaks that had been incorrectly interpreted in the initial

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FIGURE 3—CHABAZITE (CHAIN-LIKE CLUSTERS OF SMALL CRYSTALS) AND ERIONITE (LARGE CRYSTALS) FROM THE HIGH-GRADE BED AT BOWIE.

analyses as indicative of a low chabazite content. This microcrystallinity is apparent in scanning electron micrographs of the ore bed.

In April 1961, the Linde Division of Union Carbide Corporation staked claims covering most of the deposit behind the outcrops that were still covered by claims held by Baugher and Clark. In 1965, the Davison Chemical Division of W. R. Grace acquired the southeast end of the deposit. In subsequent years, the Norton Company, Letcher and Associates, NRG Incorporated, Filtrol Corporation, and the Anaconda Minerals Company purchased claim groups either not acquired or dropped by Union Carbide following the completion of the exploration drilling program in 1963.

FIGURE 4—CHABAZITE (CHAIN-LIKE CLUSTERS OF SMALL CRYSTALS) AND CLINOPTILOLITE (LARGE CRYSTALS) FROM THE HIGH-GRADE BED AT BOWIE.

FIGURE 5—DIS
BOWIE CHABAZI

Over 3,000 holes have been drilled to explore the deposit. Production records are unavailable. However, based on an estimate of the area mined since 1962, it appears that the deposit has produced about 12,000 tons of crude chabazite. Activated natural chabazite sells for $1.50 to $1.60 per pound. Therefore, after deducting moisture, mining, and processing losses, the estimated market value of the activated chabazite produced would be nearly $30,000,000. Reserves are adequate to sustain the current production rate of approximately 1,000 tons per yr indefinitely.

The Bowie deposit is mined by strip-mining techniques. In order to produce the highest purity chabazite possible, it is necessary to carefully remove all the overlying low-grade bed and the underlying green clay from the high-grade bed. This operation requires a great deal of hand labor. After mining, the chabazite is stored in warehouses in Bowie for drying and is then shipped out of state by rail for processing. The chabazite is crushed and ground to minus 200 mesh, mixed with a binder, extruded into rods or pellets and activated by calcining at a temperature of 800°F. Chabazite from Bowie, Arizona is superior to synthetic zeolites for certain specialized applications. Chabazite from the Bowie deposit is stable at a pH as low as 2.5 which makes it suitable for removing hydrogen chlorine from hydrogen streams, water from chlorine, and sulfur dioxide from stack gases (Mumpton, 1975). Chabazite from Bowie is used in a gas treatment plant at the Salt Lake oil field near Los Angeles, California to remove water, carbon dioxide, and hydrogen sulfide from casing-head gas. It is also used to purify methane at the Palos Verde, California land fill. Natural chabazite from Bowie has been used successfully to recover cesium 137 from nuclear-reactor effluents. Most recently, chabazite from the Bowie deposit has been used in a blend of natural and synthetic zeolites to remove the radioactive elements from the water at the damaged nuclear reactor at Three Mile Island, Pennsylvania.

The aerial view of the deposit taken in 1960 shows that the marker-tuff bed formed prominent outcrops that extended toward the center of the San Simon Valley (fig. 6). Most of these outcrops were removed by mining operations. These outcrops were along the southwest side of the valley. Only a few scattered outcrops occurred on the northwest side of the valley and these were nearly obscured by sand dunes.

An examination of the marker-tuff horizon (fig. 7) reveals two lithologies. A lower massive bed is known as the high-

FIGURE 6—AERIAL VIEW OF THE BOWIE CHABAZITE DEPOSIT LOOKING NORTHWEST. This photograph was taken in 1960 prior to the start of mining operations.

grade or ore bed (Sheppard and others, 1976) and the overlying bed is known as the thin-bedded or low-grade bed. The low-grade bed is a grayish or yellowish-white, zeolitically altered, vitric-tuff bed that contains numerous clay partings near the base and mud and sand partings toward the top. The upper bed grades into the overlying green and brown mudstones. Ripple marks are common particularily near the contact with the underlying high-grade bed. The upper bed, which is usually very thin or absent in the northwest part of the deposit, is from 3 to more than 5 ft thick near the center of the deposit and thins again to less than a foot thick at the southeast end of the deposit.

The mineralogy of the upper bed is not well known. Even though several hundred drill holes have penetrated the upper bed in the central part of the deposit, only a few intersections were collected and still fewer of these were analyzed. The mine operators regard the upper bed as a mining problem rather than a mineral resource because the bed is difficult to strip without disturbing the underlying high-grade bed. It is known that the upper bed contains several facies with distinct zeolite-mineral assemblages that can be correlated between drill holes across the central part of the deposit. The upper bed contains approximately 40% chabazite and significant amounts of erionite and clinoptilolite. Preliminary tests indicate that the upper bed can be beneficiated to a final product containing 85% zeolite minerals in about 50% of the original weight (Mondale and others, 1976). The upper bed is a potentially valuable mineral resource that is presently being wasted.

The lower massive bed is a yellowish- to brownish-white, zeolitically altered vitric tuff that ranges from less than 0.10

to 0.70 ft thick. It contains from 60 to 90% chabazite, from a trace to more than 15% erionite, and small amounts of clinoptilolite. The microcrystallinity of the zeolite minerals makes accurate estimates of the exact percentage of each mineral present difficult. Unquestionably, the principal mineral constituent of the high-grade bed is chabazite. The mine operators attempt to produce a consistently high-grade product. Consequently, those portions of the deposit that have a relatively higher erionite content and, therefore, a lower chabazite content exhibit a significantly lower adsorption capacity. Depending on the adsorption capacity, the operators either do not mine the bed or blend it with higher purity material to produce chabazite with a consistently uniform high-adsorption capacity.

The marker-tuff bed was deposited in a closed lacustrine basin that was confined to the present San Simon Valley. The bed is enclosed in green lake beds that consist of a section of green and brown clays having an unknown thickness. Edson (1977) has identified the clay minerals as a non-interlayered mixture of smectite and illite. He also reports that one of the holes drilled by the U.S. Geological Survey in 1972 intersected another zeolitic tuff horizon within the green lake beds about 77 ft below the marker-tuff horizon. This is not surprising, because more than one zeolitic-tuff horizon occurs at all of the other zeolite deposits in the San Simon and Gila River valleys.

The airborne vitric ash was deposited in a shallow saline-alkaline lake subject to large fluctuations in depth (fig. 8). During dry cycles, the marker-tuff bed was exposed at the surface when parts of the lake dried up. During wet cycles, this bed was again submerged and covered by green and brown clays when sedimentation resumed. Several areas occur within the deposit where the marker tuff bed was exposed at the surface, eroded, and then covered by more green clay. Some of the changes in mineralogy that have affected the adsorption and cation-exchange capacities of the high-grade bed may be correlated with these weathered areas.

The northwest end of the lake was shallow. This interpretation is supported by the discovery of tracks of an extinct llama and camel in the clay beds beneath the chabazite bed (fig. 9). A network of small streams flowed into the shallow northwest end of the lake. These fresh-water streams distributed and reworked the vitric ash prior to zeolitic alteration. Zeolitic

FIGURE 7—MARKER-TUFF HORIZON AT THE BOWIE CHABAZITE DEPOSIT showing the lower massive high-grade bed and the overlying, thin-bedded, low-grade bed.

FIGURE 8—OUTLINE OF THE ORIGINAL BASIN OF DEPOSITION AT THE BOWIE CHABAZITE DEPOSIT.
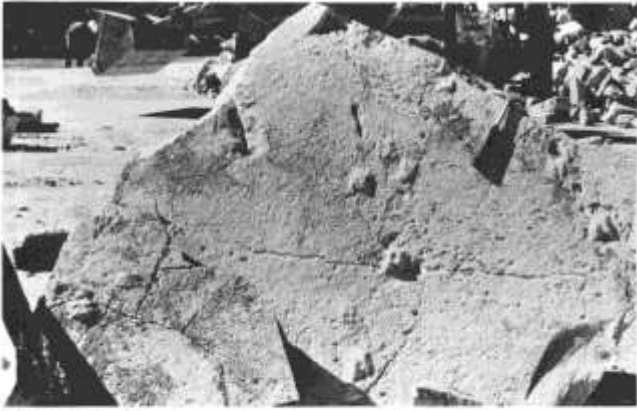
FIGURE 9—CASTS OF LLAMA TRACKS IN THE HIGH-GRADE BED OF THE BOWIE CHABAZITE DEPOSIT.

FIGURE 11—DEPRESSION IN LAKE BOTTOM AT BOWIE CHABAZITE DEPOSIT. The high-grade bed that has been cleaned but not mined was much thicker in the depression.

alteration apparently did not occur either in or adjacent to these stream beds. These so-called contemporaneous channels are filled with fine sand and unaltered vitric ash. In the northwest end of the deposit, closely spaced drilling is required to locate these channels prior to stripping and mining operations.

The initial ash fall after compaction appears to be about 0.1 ft thick and can usually be seen in the drill core as a thin white layer that approaches 100% chabazite at the base of the high-grade bed. Streams transported the ash deposited on the adjacent watershed into the lake where currents and the topography of the lake bottom controlled its distribution (fig. 10). The high-grade bed is thickest along depressions in the lake bottom (fig. 11).

The mineralogy of the high-grade bed in the marker-tuff horizon changes from the shallow northwest end to the deeper southeast end. Near Ryan Dam at the northwest end of the depositional basin, the marker-tuff horizon is a clean, white, unaltered vitric tuff that grades into a partially altered vitric tuff and then into the typical high-grade bed to the southeast. The northwest end of the deposit was a calcium-rich environment characterized by the occurrence of calcite concretions within the high-grade bed. The chabazite is a high-calcium variety that has a light-yellow color. Gypsum does not occur in the northwest end of the deposit. The southeast end of the deposit was a sodium-rich environment characterized by the occurrence of thenardite within the high-grade bed. The chabazite is a high-sodium variety approaching the composition of herschelite (Regis and Sand, 1967) which has a dark-orange-brown color. Both gypsum and analcime are common in the southeast end of the deposit.

The change from a calcium-rich to a sodium-rich depositional environment is gradational over a distance of 7 mi. Interestingly, the increase in sodium content is also reflected by the color of the high-grade bed. High-calcium chabazite is a light-yellowish-white color and high-sodium chabazite is an orange-brown color. Previous workers believed that the orange-brown chabazite contained more iron than the yellowish-white chabazite. However, chemical analyses (Sheppard and others, 1976) show that the yellowish-white, calcium-rich chabazite has a higher iron content than the orange-brown, sodium-rich chabazite.

A system of younger paleochannels has cut through the green lake beds and removed most of the marker-tuff horizon. The deposit in fact consists of several erosional remnants separated by alluvium-filled channels. Subsurface information indicates that the paleochannels on the southwest side of the San Simon Valley flowed to the northeast. Less complete subsurface information from the northeast side of the valley suggests that the channels flowed toward the southwest. This indicates that the paleochannels drained into a major northwest-flowing drainage system near the center of the present San Simon Valley, which also drains to the northwest (fig. 12).

FIGURE 10—A CONTEMPORARY CHANNEL FILLED WITH GRAY SAND AND UNALTERED ASH CUTTING THROUGH ORE BED AT BOWIE. Direction of flow was from the lower left toward the top of the photograph.
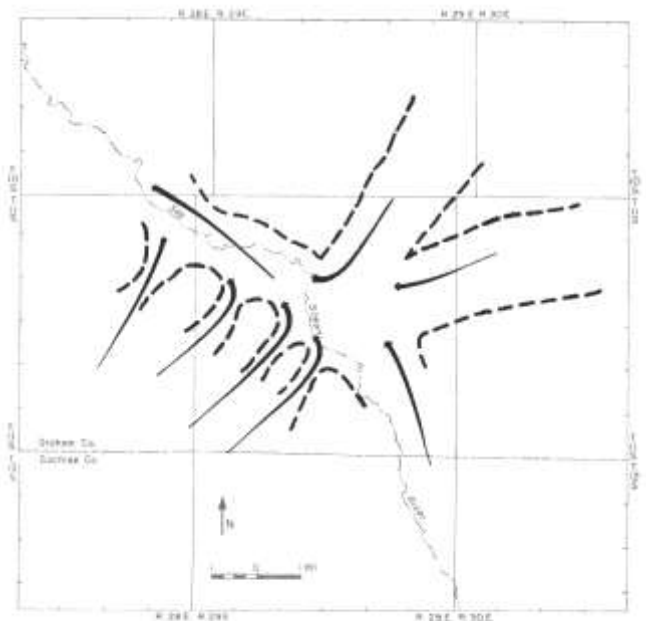
FIGURE 12—MAJOR PALEOCHANNELS AND DIRECTIONS OF FLOW AT THE BOWIE CHABAZITE DEPOSIT.

FIGURE 13—A PALEOCHANNEL (DARK-COLORED AREA NEXT TO BACK-HOE) THAT CUTS THROUGH UNMINED, HIGH-GRADE BED AT THE BOWIE CHABAZITE DEPOSIT. Direction of flow is from the bottom of the photograph toward the top.

FIGURE 14—DISTRIBUTION OF THE BROWN LAKE BEDS AT THE BOWIE CHABAZITE DEPOSIT.

The channels are filled with sand, coarse gravel, and brown mudstone (fig. 13). Large angular fragments of chabazite from the high-grade bed are found in the channel gravels in areas where the channels intersect the marker-tuff horizon. Small, well-rounded chabazite fragments from the high-grade bed can be found in the channel gravels several miles downstream from the area where the channel cut the marker-tuff horizon. The paleochannels developed after the deposition of the green lake beds and of course after the zeolitic alteration of the marker-tuff bed was complete.

The brown lake beds overlie both the paleochannels and the green lake beds. Subsurface information from the northeast side of the San Simon Valley confirms that the brown lake beds overlie paleochannels that have cut through and removed the older green lake beds and the zeolitically altered marker-tuff horizon. On the southwest side of the valley, the brown lake beds overlie the green lake beds. The basin of deposition for the brown lake beds (fig. 14) appears to have been much smaller than that of the green lake beds. Possibly, the recent erosion associated with the formation of the present San Simon Valley may have removed the brown lake beds from the southwest side of the valley. The brown lake beds are characterized by clay beds that contain disseminated halite crystals. No zeolitic tuff beds are known to occur in the brown lake beds.

Conclusions

The zeolite deposits in the Gila and San Simon Valleys are the result of the alteration of vitric ash deposited in a series of saline-alkaline lakes. Geological evidence suggests that the vitric ash from distant volcanic eruptions did not simply fall into large quiescent lakes with stable levels of water and alkalinity. Instead, the ash appears to have been deposited in shallow lakes in which the water level and alkalinity were variable and sedimentation was active. The successful exploration and development of these zeolite deposits requires an understanding of how these geological processes control their formation.
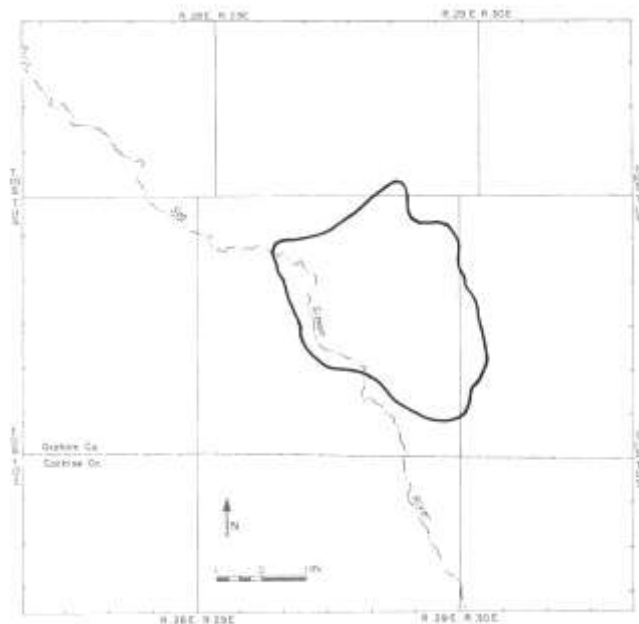
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LITHIUM-RICH TUFFS IN THE POPOTOSA FORMATION, NEW MEXICO

by Sigrid Asher-Bolinder, *U.S. Geological Survey, Denver, Colorado 80225*

Abstract

Throughout the western United States are a variety of lithium-rich di- and trioctahedral smectites. These rocks are characteristically found in closed-basin lacustrine or playa sequences of Miocene to Pleistocene age and represent lithium enrichment of sediments beyond that of ordinary solar evaporation and concentration. In most cases, caldera eruptions of calc-alkaline and basaltic to andesitic composition and related hot-spring activity have been contemporaneous with the deposition of sediments now enriched in lithium. Indeed, such volcanic systems may provide both the source and the enrichment mechanism for lithium.

In Socorro County, south-central New Mexico, there are remnants of a now-disrupted Miocene closed basin that developed during the initiation of the Rio Grande rift. The basin was characterized by syndepositional faulting, high sedimentation rate, additions of air-fall ashes from nearby calderas, and the presence of a caldera within the basin itself. Rocks deposited in that paleobasin are called the Popotosa Formation.

Today, several of the ashes within the formation are characterized by lithium values as high as 3,800 ppm. The pattern of lithium enrichment within the basin appears to be one of logarithmic increase of lithium values with distance away from the intrabasin caldera until a maximum is reached; farther away from the caldera than that, lithium values drop rapidly to near-background levels.

This halo-like effect of 50-100 times background-level enrichment occurs in a system that apparently has suffered otherwise no more than burial diagenesis and the alteration of playa sediments in an alkaline ground-water system. The halo of anomalous lithium values suggests that enrichment may have occurred shortly after deposition of the ashes, before intrabasin faulting ruptured the blankets of airfall ash, and before alteration of the ashes to clay and zeolites destroyed the initial high porosity.

Introduction

Today, the lithium industry in the United States is a small but stable one that supplies the United States and also produces three-quarters of the free world's supplies. Lithium and its compounds are used for ceramic shock resistance, lubricative properties, flux properties in aluminum smelting, and pharmacological properties. Lithium is being used in increasing amounts in the exotic and everyday fields of primary batteries.

Approximately half of United States production of lithium is derived from the roasting and leaching of spodumene from the North Carolina pegmatite belt. Lithium is also extracted from a brine of unusual composition by evaporation and chemical precipitation. This brine is ideally situated for evaporative concentration in a playa of Clayton Valley, southwest Nevada. Other lithium-rich brines and pegmatites occur worldwide but are not in production at this time, or are supplying nations such as the U.S.S.R. or the People's Republic of China, where little or no production data is given (Singleton, 1979).

Because petroleum is a finite, expensive source of energy, one alternative form of transportation being developed in the United States is the electric vehicle. The rechargeable storage-battery-powered vehicle provides clean, efficient, about-town transportation for personnel and loads, and is ultimately powered by the most efficient energy source available to the power company that recharges the battery. Although the total number and kinds of electric vehicles that will penetrate the market are unknown, one type of battery, the lithium/iron-sulfide battery, is being developed by Gould Incorporated, Eagle-Picher Industries, Argonne National Laboratories, and others. This battery and its variants utilize lithium's inherent properties of high specific energy, light weight, and safety.

Another large consumer of lithium in the next 20-40 yrs may be the nuclear-fusion industry, where lithium may serve as the insulating blanket for the reactor and as the source for the tritium required by the fusion reaction. The enormous amounts of lithium required and the relative insignificance of cost of the metal to the overall fusion operation are problems for lithium producers to consider.

For these speculative uses, and for the increasing worldwide growth of lithium use in more traditional markets, domestic lithium producers and potential producers have turned their interest to foreign lithium resources. In the last 10 yrs, rich lithium brines have been located in high Andean playas of South America such as Salar de Atacama, Chile, and Salar de Uyuni, Bolivia. Although such brines represent huge resources waiting to be developed, their ultimate control is foreign, and subject to disruption. Yet, at present, these resources are more economic to develop than are pegmatites in Canada.

The U.S. Geological Survey has been studying the occurrences and mechanisms of lithium accumulation within the United States as part of a larger study of sedimentary minerals of economic and strategic importance. The lithium-rich altered volcanic ashes (tuffs) in south-central New Mexico discussed in this preliminary report do not constitute a present resource, but our understanding of their mode of origin may allow further discovery and development of domestic lithium resources.

Sedimentary and tectonic history of the Popotosa Formation

The Popotosa Formation of south-central New Mexico is a diverse assemblage of late Oligocene to late Miocene terrestrial sediments and tuff beds that reflect the tectonic history of part of the developing Rio Grande rift. The Popotosa represents mudflow, fluvial, alluvial-fan and alluvial-flat, and playa environments produced in a climate botanically similar to today's. The rocks were deposited in the Socorro paleobasin, which is ringed by late Oligocene calderas and the Colorado Plateau, and was floored, in part, by Oligocene basaltic andesites characteristic of early rifting. The paleobasin was later reintruded by rhyolitic to dacitic volcanics before breaching. Syndepositional faulting and widening of the paleobasin provided a catchment for more than 1,500 m of sediments that reflect individual areas of basin isolation, renewed movement, and local sediment sources.

Today these sediments are exposed on the west side of the Rio Grande rift along a series of intrabasin horsts either exhumed or created in early Pliocene time. Exposures are best along two major north-south trends, and provide little chance for east-west correlation (fig. 1). Exposures on the east side of the rift are unknown. Two excellent areas of continuous expo-

FIGURE 1—EXPOSURES OF POPOTOSA FORMATION (PATTERNED) IN THE SOCORRO PALEOBASIN. Sections 1 and 2 are measured sections described in text. Stippled area represents present-day mountain ranges.

Geochemical considerations

Lithium deposits require a source for the metal, a concentrating mechanism, and a trap. Evidence in the Popotosa Formation provides some indications of what these three mechanisms were, and suggests that lithium accumulation in the tuffs was contemporaneous with basin-sediment deposition, or occurred shortly thereafter.

Although sections 1 and 2 are uncorrelatable on a bed-bybed scale (fig. 3), and although portions of the basin were apparently active or isolated at various times in basin history, a general pattern of lithium enrichment within the basin away from the Oligocene Socorro caldera emerges (fig. 4). Such a pattern may be fortuitous, but the caldera focus was chosen based on the geochemical behavior of lithium. The high-lithium clays and tuffs of the Horse Spring Formation (Miocene) of southern Nevada are associated with spring pots (Brenner-Tourtelot and Glanzman, 1978). Evidence also exists for hydrothermal alteration of the intracaldera clays that contain lithium in McDermitt caldera (Glanzman and others, 1978). The commercial lithium brine at Clayton Valley, Nevada, occurs in sediments that were derived from a caldera that still shows hot-spring activity at its margins (Davis and Vine, 1979).

In all these cases, spring activity related to volcanism is a common denominator, but the lithium-rich sediments are not those closest to the springs. This fact is more understandable if one considers the solubility of the lithium ion. In ground water, the lithium ion travels tightly bound to a water molecule, and is too small, if separated from its water, to substitute for sodium or potassium. However, limited substitution of lithium for magnesium occurs in some clays. Lithium is also the last element to precipitate from a brine. Thus, I chose a known volcanic center as a point of origin for measuring lithium concentration of tuffs of the Popotosa Formation (fig. 4). Today at the base of Socorro Peak (a remnant of the caldera), hot-spring activity continues.

If lithium had been extracted from the caldera-related volcanics by hot water, the lithium ions may have been transported by ground water to the deeper parts of the basin in very permeable and porous airfall ashes before the ashes were altered to diagenetic minerals and before the basin was disrupted.

The same airfall ashes that may have served as conduits for lithium leached by thermal waters from caldera-related volcanics may have been the source of the lithium. Experiments

sure of the basin sediments are marked sections 1 and 2 in fig. 1. These sections are the thickest and most complete in the basin, and the rocks were apparently deposited near the center of the paleobasin. Oldest and youngest ages for the more than 1,500 m of rocks exposed are 26.3 m.y. for an andesite flow near the base of one section and 4.5 m.y. for establishment of Pliocene through-flowing drainage (Machette, 1978).

Seven tuffs are present in section 1, and 12 tuffs in section 2 (fig. 2). The exact ages and sources of the tuffs remain unknown, but local intrabasin volcanic events are suspected as the sources (G. R. Osburn, personal communication, 1981). Preliminary heavy-mineral separations for fission-track age-dating did not yield sufficient material for dating. However, a tuff in the paleobasin south of section 1 yielded an age of 11.8 m.y. (K. Manley, personal communication, 1981). This tuff is local in extent and is uncorrelatable with measured sections 1 and 2.

However, the importance of the tuffs lies not so much in their ages as in their lithium content. The seven tuffs in section 1 range from 30 to 680 ppm lithium and average 355 ppm Li; their enclosing rocks average 75 ppm Li. The 12 tuffs in section 2 vary from 730 to 3,850 ppm Li and average 2,600 ppm Li; their enclosing rocks average 120 ppm Li. Lithium values greater than 300 ppm are assumed to be anomalous, and some rocks having lithium values greater than 1,000 ppm have been studied by industry and the U.S. Bureau of Mines as possible future domestic lithium sources.

FIGURE 2—AIRFALL TUFF GRADING INTO WATERLAIN TUFF, SECTION 2, POPOTOSA FORMATION, SOCORRO COUNTY, NEW MEXICO. This 2-m thick white tuff contains as much as 3,275 ppm Li.

FIGURE 3—GENERALIZED STRATIGRAPHY OF MEASURED SECTIONS 1 AND 2, POPOTOSA FORMATION, SOCORRO COUNTY, NEW MEXICO.
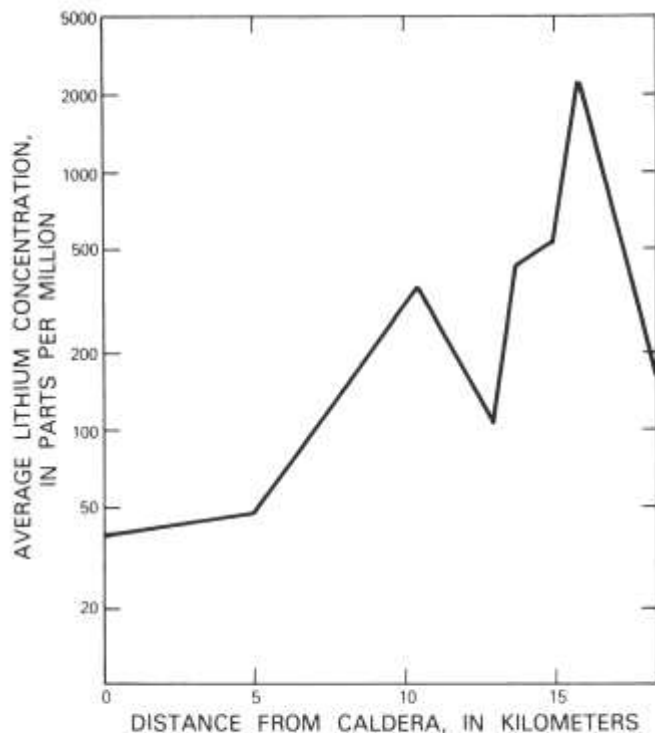
FIGURE 4—LOGARITHMIC PLOT OF LITHIUM CONCENTRATION IN TUFFS VERSUS DISTANCE FROM THE INTRABASINAL SOCORRO CALDERA, POPOTOSA FORMATION.

by Zielinski (1979) showed that as much as half of the available lithium in volcanic ashes can be leached at 120°C by alkaline ground waters. The passage of hot-spring waters through the ashes may have leached and deposited lithium in the ashes near the basin center as the waters cooled. Section 2, with its high lithium concentration, may simply represent the presence of hot-spring activity whose record in the Popotosa is now removed or remains unexposed. Thus, the hydrothermal activity that produced the deposit may not be related to the Socorro caldera farther north in the basin.

In section 1, some evidence exists for hydrothermal activity that may have supplied the driving force for lithium accumulation in section 1, and possibly in section 2. A thin black chert that bifurcates into two limestones contains stemmed and oncolitic algae. These cherts occur just below a tuff (very low in lithium) and may represent the dissolution of silica in the ash by slightly acid (organic-rich) spring ponds.

The mechanism for concentrating lithium up to 95 times its background concentration should be reflected in the sedimentation and mineralogy of the Popotosa. Evaporation appears to be a likely mechanism for concentration, as it has been in other lithium-rich basins. However, there is little evidence that the playa facies held standing water for long or that evaporative concentration reached higher levels than the dolomitic to gypsiferous levels. No bedded salines or stromatolites, and only discontinuous stringers of gypsum are found in the Popotosa; fluvial sediments are ubiquitous in the playa facies, where oxidizing conditions prevailed.

If no evidence of a pool of highly saline and alkaline brine indicates evaporative concentration as a mechanism for lithium enrichment, perhaps other evidence of high cation concentration can be found in the tuffs. Table 1 shows the mean-elemental concentrations of tuffs from the Popotosa versus early eruptive sequence of the Bishop Tuff (Pleistocene), from which the Searles Lake lithium deposit may have derived its lithium. Assuming similar original compositions of the rocks, and allowing for ground-water alteration of the much older Popotosa, the enrichments in the Popotosa of lithium and

magnesium are large. Magnesium is a major component of both di- and trioctahedral smectites formed from tuffs; lithium can be contained in both. Actual concentrations of lithium in the individual ashes may be as great as 100 times that in the Bishop Tuff. Barium and strontium values of the Popotosa are comparable to average granite.

The only zeolite recognized on the X-ray diffractograms of the tuffs, regardless of their lithium content, is clinoptilolite. Clinoptilolite is not diagnostic of alkalinity and salinity; it forms today under a variety of environments from above the water table to marine conditions but is not indicative of high alkalinities. Thus, evaporative concentration of brines or hydrothermally-enriched fluids does not adequately explain the concentration of lithium in the basin.

Glanzman and others (1979) postulated that anomalous concentrations of boron, uranium, and lithium may be used as guide elements for one another in closed basins containing volcanics such as McDermitt caldera, Oregon and Nevada, and the Date Creek Basin, Arizona. In the Popotosa, uranium values in the sedimentary rocks are generally no more than two times background level except in pockets of reduced sandstone containing plant debris. Boron values are also at background levels or less, except in gypsum beds. Thus, it is not clear that the enrichment mechanism of the lithium in the Popotosa was the same as that for boron or uranium.

Although source and concentration mechanisms cannot be completely defined in the Popotosa, the trap for the lithium seems evident. Lithium content of the whole tuff correlates fairly well with the percentage of clay-sized material separated from the tuff. X-ray diffractograms show that the clay fraction is mainly the dioctahedral smectite montmorillonite, with some clinoptilolite and calcite. Up to 50% of the weight of the whole tuff can be montmorillonite, the most logical trap for lithium. The trioctahedral smectites such as hectorite, saponite, and stevensite are not visible on X-ray diffractograms. Trioctahedral smectites can contain lithium as structural components of their octahedral layers, but montmorillonite can hold lithium only as interlayer cations that are susceptible to cation exchange in nature and, perhaps, in the laboratory. The exchangeability of lithium in the interlayer position and lithium's extreme solubility suggest that the Socorro paleobasin waters must have been nearly depleted in the tuffs; the lithium was incorporated in the clay structure before Pliocene disruption of the basin allowed through-flowing drainage to carry away lithium-enriched ground waters.

Conclusions

In the Miocene Popotosa tuffs of the Socorro paleobasin, the source of lithium seems to have been from hydrothermal fluids leaching nearby rhyolitic volcanics, or from groundwater or hydrothermal transport of lithium from one site to another within the porous ashes. The concentrating mechanism may have been evaporative concentration within the playa, but the exact mechanism is not adequately reflected in associated elements or the geology. Montmorillonite, produced by alteration of the ashes, can trap considerable amounts of lithium in its interlayers in a closed-basin environment.

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THE ORIGIN OF TALC IN THE ALLAMOORE DISTRICT, TEXAS

by W. B. Bourbon, *Consulting Geologist, Alpine, Texas 79830*

Abstract

The Allamore Formation (Precambrian) is host to numerous deposits of ceramic-grade and paint-grade talc in the Allamore Talc district, Culberson and Hudspeth Counties, Texas. Common accessories to talc are dolomite and/or quartz in virtually all orebodies with the exception of two paint-grade deposits that contain only minor accessory chlorite (penninite), and two ceramic-grade deposits that contain minor, accessory, asbestiform richterite. Calcite is notable by its absence. The talc probably formed in Precambrian time by replacement of primary sedimentary magnesite. Water of undetermined origin permeated cherty magnesite beds of the Allamoore Formation and, with heat, produced talc according to the reaction:

$$3MgCO_3 + 4SiO_2 + H_2O = Mg_3Si4O_3(OH)_2 + 3CO_2$$
$$\text{(magnesite)} + \text{(quartz)} + H_2O = \text{talc}$$
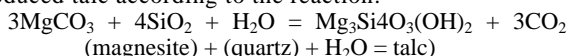
Primary dolomite and excess quartz contained in the magnesite deposits were essentially unaffected by the low-temperature (approximately 300 °C) reaction. Subsequent tectonic and intrusive igneous activity of undetermined age has significantly altered the mineralogy of several talc bodies, resulting in paint-grade talc deposits.

The magnesite deposits were formed by chemical precipitation in isolated hypersaline lagoons or in associated ephemeral lakes. Stromatolitic dolostone units which both overlie and underlie talc deposits in the Allamoore Formation were formed in the intertidal zone of restricted lagoons. Limestones in the Allamoore Formation were formed in a subtidal environment.

Introduction

The Allamoore Talc district encompasses approximately 400 km^2 of west-central Culberson and east-central Hudspeth Counties, Texas. Talc-bearing rocks crop out discontinuously for approximately 48 km westerly from Tumbledown Moun 11 km north-northwest of Van Horn, Culberson County, to the Steeruwitz Hills, 8 km east of Sierra Blanca, Hudspeth County (fig 1). Production statistics are not released in the district, but 1980 production is estimated at 300,000 tons of talc.

Previous work

Von Steeruwitz (1890, 1891, 1892, 1893) was the first geologist to study the Precambrian rocks of Trans-Pecos Texas. Richardson (1904, 1914) provided additional information on the Precambrian north of the (then) Texas and Pacific Railroad in the Van Horn area. The geology of the Sierra Diablo foothills area, encompassing the Allamoore Talc district is discussed by King (1940, 1980), King and Flawn (1953), and Davidson (1980). Reid (1974) described the Hazel Formation. Information on talc deposits in the Allamoore Formation is provided in Flawn (1958), King (1965), Rohrbacher (1973), Hawkins and Girard (1977), and Edwards (1980a). Metallic mineral occurrences within the district are described in Evans (1943), Sample and Gould (1945), and King and Flawn (1953). Gravity and magnetic data for the district are presented by Wiley (1972). Nyberg and Schopf (1980) describe fossil blue-green algae from the Allamoore Formation. Rb/Sr age determinations are discussed by Denison and Hetherington (1969), and Denison (1980).

Stratigraphy

Sedimentary rocks of the Allamoore district range in age from Precambrian to Cretaceous. Metamorphic rocks and extrusive igneous rocks of the district are of Precambrian age. Precambrian rocks include the Carrizo Mountain Group, the Allamoore Formation, and the Hazel Formation. The age of the younger Van Horn Sandstone is uncertain, but has been considered to be late Precambrian by King (1940) and McGowen and Groat (1971). King and Flawn (1953) and McGowen and Groat (1971) suggest a possible Ordovician age.
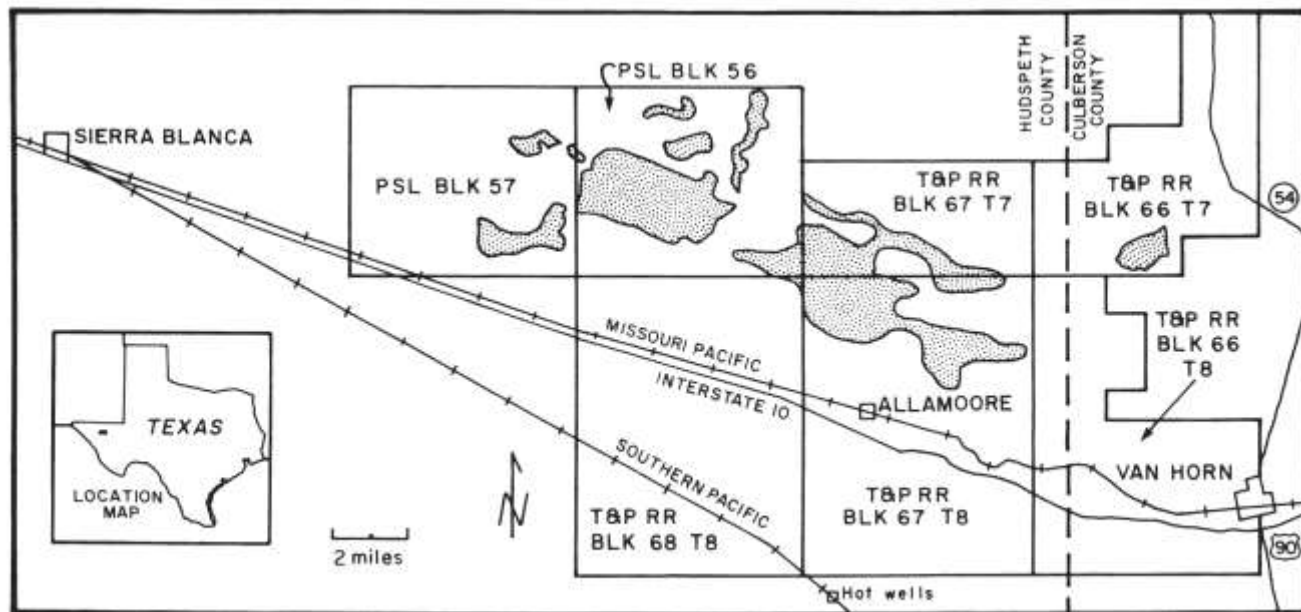
FIGURE 1—INDEX MAP OF ALLAMOORE TALC DISTRICT (STIPPLED) AND ADJOINING LANDS, SHOWING TEXAS LAND NET SURVEY BLOCKS.

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Ordovician rocks, exposed only in the extreme eastern end of the district, include the Bliss Sandstone and El Paso Limestone (both Early Ordovician), and the Montoya Limestone (Late Ordovician). The Permian Hueco Limestone (Wolf-camp Series) crops out only in the extreme western end of the district. The Campo Grande Limestone (Cretaceous) of the Comanche Series similarly is exposed only in the western portion of the district. Extensive deposits of Quaternary alluvium blanket the broad valleys.

Because economic talc deposits are found only in the Allamoore Formation, this paper will emphasize that formation. For detailed descriptions and distribution of all other units mentioned, the reader is referred to King and Flawn (1953); for information on the Hazel Formation see Reid (1974).

Allamoore Formation

The Allamoore Formation consists of massive dolostones, limestone, talc-bearing rocks, thin-bedded to massive clastic sediments, and volcanics. Total thickness of the Allamoore Formation is probably in excess of 2,600 m, but no complete section exists.

The basal Allamoore is primarily composed of massive, stromatolitic, cherty, and, in part, silicified dolostone and cherty limestone approximately 970 m thick. Contained within this basal unit are sparse volcanic beds less than 15 m thick. Low-angle crossbedding is noted in slightly arenaceous dolostone beds.

Overlying the basal carbonate sequence locally are talc-bearing rocks as much as 160 m thick. Elsewhere, the succession of carbonates continues upward without interruption by talc horizons.

Talcose rocks are overlain in turn by carbonates identical to those of the basal unit, or, locally, by concordant chloritic igneous rocks less than 15 m thick and of probable volcanic origin. Thickness of dolostone varies from a few meters to over 200 m. This carbonate sequence is truncated upsection by wedgelike, coarse clastic rocks composed of locally derived conglomerate that grades upward to medium-grained sandstone. These clastic units are similar lithologically to the clastic unit upon which the basal Allamoore rests and to the presumably younger Hazel Formation. In the west-central portion of the district, sandstone is overlain by massive limestone-boulder conglomerate. Stromatolitic and nodular dolostones less than 15 m thick are abundant within the dominantly clastic sequence. Total thickness of this clastic sequence exceeds 680 m.

Overlying the clastic wedges is the uppermost exposed Allamoore unit, consisting predominantly of dolostone with sparse talc horizons, thin-bedded clastics, and volcanics. Maximum thickness is over 600 m, with talc horizons rarely exceeding 100 m. The Allamoore Formation is unconformably overlain by the Hazel Formation.

DOLOSTONES—The more than 1,600 m (combined thickness) of dolostones in the Allamoore Formation exhibit great diversity in both composition and appearance. Pervasive silicification that has locally obliterated all sedimentary structures, elsewhere has preserved cryptalgal laminae. Most dolostone is not silicified but contains abundant chert as a primary precipitate or as a product of selective early diagenetic mineralization. Recrystallized dolomite composes the bulk of those dolostones that are not silicified.

Stratiform stromatolites occur in great abundance in sets of planar laminae (fig. 2) and convolute wavy laminae (fig. 3). Columnar stromatolites (fig. 4) and oncolites (fig. 5) are also noted. Branching stromatolites are not known to occur in the Allamoore Formation. Parallel to subparallel laminae from< 0.5 to > 5 mm thick are characteristic of Allamoore stromatolites. Laminae are consistently cherty, whereas interspaces are generally dolomitic. Interspaces are frequently as thick as, or

FIGURE 2—COLUMNAR STROMATOLITES (AT BASE OF HAMMER) OVERLAIN BY PLANAR STROMATOLITES IN CHERTY DOLOSTONE, CROSS-SECTIONAL VIEW, ALLAMOORE FORMATION.

slightly thicker than, bounding laminae. Minute (< 0.5 mm thick) vertical seams often connect cherty laminae and represent infilling of dessication cracks in the dolomitic interlayers (fig. 6).

Planar stromatolites probably are of the genus *Stratifera* (Awramik, 1977). Convolute stromatolites closely resemble *Tungussia* (Donaldson, 1976; Preiss, 1976), but may represent a morphology intermediate between stratiform and columnar. *Conophyton-like* columnar stromatolites are also observed. Oncolites were identified only in the intermediate carbonate sequence. Coccoidal and filamentous blue-green algae, both microbial mat builders, have been identified in laminated cherty Allamoore carbonates (Nyberg and Schopf, 1980; Ellison, 1981, personal communication).

Nodular dolostone (fig. 7), noted at several localities, constitutes excellent marker beds over strike distances as great as 3,000 m. The nodules are nudose (grapestone-like), range from 2 to 15 mm in diameter, and are composed of anhydrite
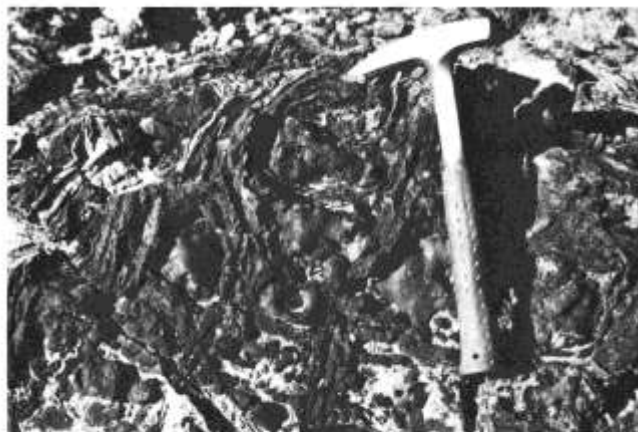
FIGURE 3—PLAN VIEW OF DOMAL ALGAL HEADS IN CHERTY DOLOSTONE, ALLAMOORE FORMATION.

[New Mexico Bureau Mines & Mineral Resources, Circular 182, 1982]
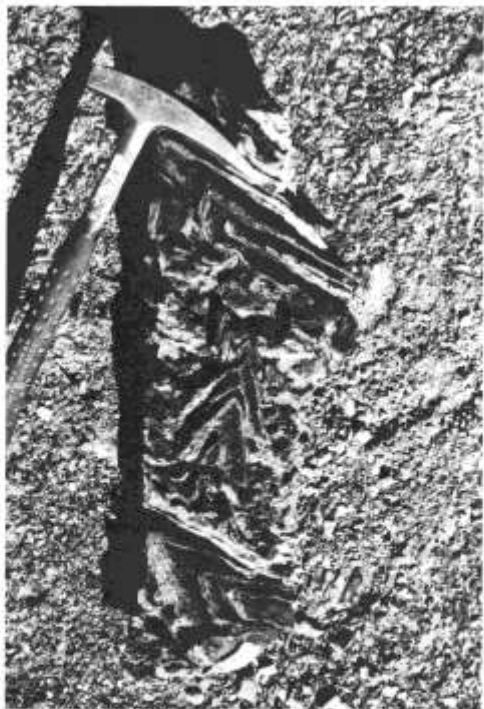
FIGURE 4—COLUMNAR STROMATOLITES IN CHERTY DOLOSTONE, CROSS-SECTIONAL VIEW, ALLAMOORE FORMATION.

FIGURE 7—CHERT NODULES AFTER ANHYDRITE IN CHERTY DOLOSTONE ALLAMOORE FORMATION.

or chert after anhydrite. Crossbedded, arenaceous dolostone (fig. 8) is uncommon in the Allamoore but, where present, may be used as a local marker bed. Dessication cracks in dolostone are prominent along the conformable contact between the basal dolostone and the overlying lower talc unit (fig. 9).

LIMESTONES—Cherty limestone beds up to 10 m thick are noted in the massive dolostone sequences. Limestones in the Allamoore are remarkably lacking in sedimentary structures and are, without exception, non-stromatolitic.

CLASTICS—The clastic sequences in the upper Allamoore are composed predominantly of massive, poorly sorted dolostone boulder, cobble, and pebble conglomerates and unsorted to poorly sorted, medium- to coarse-grained, red sandstones. Sparse, thin-bedded stromatolitic dolostones are also noted. Reid (1974) provides detailed descriptions of the Hazel Formation, which is apparently interbedded with the Allamoore Formation.

VOLCANICS—Dark-red-brown to purple, aphanitic, extrusive igneous rocks and buff to orange-brown pyroclastics are observed as thin interbeds throughout the Allamoore Formation.

Talc deposits

The terms "ceramic-grade talc" and "paint-grade talc" as used herein refer to commercial applications and are comparable, respectively, to the mixed talc ores and soft platy talc of

FIGURE 5—ONCOLITES IN DOLOSTONE, CROSS-SECTIONAL VIEW, ALLAMOORE FORMATION.

FIGURE 6—FINELY LAMINATED STROMATOLITIC CHERTY DOLOSTONE, ALLAMOORE FORMATION. Note infilling of dessication cracks (arrows).

FIGURE 8—CROSSBEDDED ARENACEOUS DOLOSTONE, CROSS-SECTIONAL VIEW, ALLAMORE FORMATION.

FIGURE 9—Dessication cracks in cherty dolostone, at interface between dolostone and overlying talc, Allamoore Formation.

TABLE 2—X-ray diffraction data; Sample A = pink talc, sample B = gray/black laminated talc, sample C = gray/black laminated talc, sample D = caliche in soil horizon overlying gray/black talc deposit; CuK α radiation, Ni filter (λ = 1.5418Å), 2°2θ/min, 8° to 52°2θ; dÅ = diameter in Angstroms; *tc = talc, pe = penninite, qt = quartz, do = dolomite, mg = magnesite; ** relative intensity determined from area under curve.

		I** Sample no.			
dÅ	Mineral *	A	B	C	D
9.34	tc	100	100	60	10
7.20	pe	20	—	—	—
4.67	tc	20	20	20	—
4.56	tc	12	20	15	—
4.26	qt	—	5	—	—
3.69	do	—	—	5	—
3.61	pe	20	—	—	—
3.34	qt	—	15	5	5
3.12	tc	100	100	60	10
2.89	do	—	12	100	40
2.74	mg	—	—	10	100
2.67	do	—	—	6	—
2.60	tc	30	25	15	—
2.54	do	—	—	6	—
2.50	mg	—	—	—	30
2.48	tc	65	40	10	—
2.21	tc	20	5	—	—
2.19	do	—	5	40	10
2.13	qt	—	—	—	—
2.10	tc/mg	20	5	5	60
2.01	do	—	5	10	5
1.87	tc	40	10	—	—
1.82	qt	—	—	—	—
1.79	do	—	5	—	—
1.70	mg	—	—	5	60

Roe (1975). Local standards for ceramic-grade talc generally require less than 15% (by volume) disseminated quartz and dolomite to ensure that loss on ignition during calcining does not exceed 11%. Paint-grade talc is of much higher purity, contains very little or no disseminated quartz or dolomite, and yields loss on ignition less than 7%. Sub-economic, talc-bearing rock is herein referred to as talcose rock.

CERAMIC-GRADE TALC

Ceramic-grade talc occurs in tabular bodies up to 2,000 m long and as much as 160 m thick. Mining and drilling to date have failed to yield conclusive information regarding the third (downdip) dimension of the talc bodies.

All ceramic-grade deposits contain accessory quartz and dolomite. Field observations and petrographic and x-ray-diffraction analyses show broadly varying concentrations of these contaminants from < 5 to > 30% by volume (table 2). Quartz and dolomite occur in talc as disseminated irregular grains (<0.05-15 mm in diameter) and as interbeds 1 cm-m thick, parallel to relict bedding. Concentrations of bedded dolostone tend to be greatest near both the upper and lower boundaries of a talc deposit.

Asbestiform richterite, an alkali-rich amphibole, is present only at two inactive prospects. Magnesite in minute quantities was detected in talc by x-ray-diffraction procedures and was also found in caliche overlying talc deposits at two locations. Calcite is notable by its absence.

Chemical analyses show Allamoore talc to be extremely low in potassium, sodium, iron, and aluminum (table 1). Fluorine is present in quantities ranging from 0.1 to 0.5% (Ross and others, 1968; additional proprietary data).

Ceramic-grade Allamoore talc is characterized by fine laminae (figs. 10 and 11) 0.5 mm-1 cm thick, varying in color from gray/white sets to black/dark-gray sets which represent at least in part relict cryptalgal laminae. Other primary sedimentary structures observed in ceramic-grade talc deposits in-

TABLE 1—Talc analyses; *LOI = loss on ignition, ** total H₂O only; Sample data: 1-4, Allamoore District, Texas (dark-gray talc); 5, Montana (talc); 6, No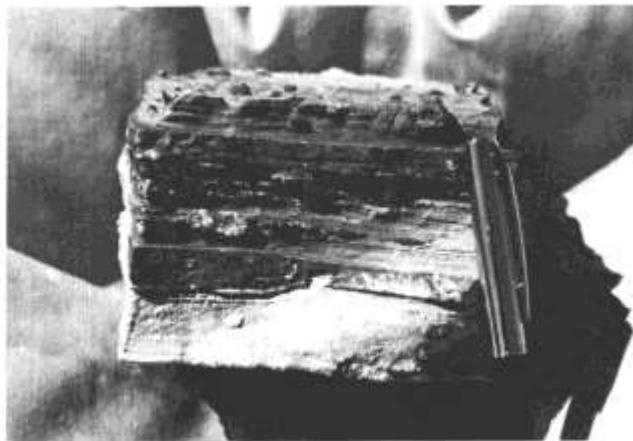rth Carolina (talc); 7, Death Valley, California (talc); 8-9, Allamoore District, Texas (pink talc); 10, theoretical composition; Data source: 1-4, 6-9, analyses provided by U.S. Dept. of Interior, National Park Service; 5, Okuma (1971).

Sample number	Oxide							
	SiO₂	Fe₂O₃	Al₂O₃	CaO	MgO	Na₂O	K₂O	LOI*
1	56.74	0.11	0.04	3.66	28.87	0.06	0.06	10.11
2	56.14	0.93	0.52	4.20	27.00	0.19	0.17	10.56
3	55.88	0.08	0.19	3.50	30.16	0.09	0.07	9.64
4	51.28	0.26	0.67	7.36	25.07	0.12	0.12	14.41
5	62.25	0.71	0.27	0.05	31.13	0.06	0.02	5.03
6	55.80	1.65	4.34	0.04	31.39	0.06	0.03	6.56
7	57.40	0.36	0.51	5.32	27.13	1.98	1.16	5.82
8	59.88	0.05	0.10	1.36	31.48	0.06	0.04	6.67
9	60.94	0.07	0.10	0.66	31.98	0.06	0.06	5.72
10	63.36	—	—	—	31.89	—	—	4.75 **

FIGURE 10—Finely laminated ceramic-grade talc, cross-sectional view, Allamoore Formation. Note imbricate rip-up clasts (A), and edge-wise flat pebble conglomerate (B), exposed on fracture cleavage surface (parallel to plane of photograph).

FIGURE 11—FINELY LAMINATED CERAMIC-GRADE TALC, ALLAMOORE FORMATION. Note fracture cleavage perpendicular to laminae.

FIGURE 12—MINUTE DESSICATION CRACKS IN DOLOSTONE (PLAN VIEW) FROM CERAMIC-GRADE TALC, ALLAMOORE FORMATION.

FIGURE 13—DESSICATION CRACKS IN CERAMIC-GRADE TALC (PLAN VIEW), ALLAMOORE FORMATION.

FIGURE 14—QUARTZ/DOLOMITE PSEUDOMORPHS AFTER DISPLACIVE HOPPER HALITE CRYSTALS, FROM CERAMIC-GRADE TALC, ALLAMOORE FORMATION.

FIGURE 15—QUARTZ/DOLOMITE AFTER ELLIPTICAL ANHYDRITE NODULES IN CERAMIC-GRADE TALC, ALLAMOORE FORMATION.

dude rip-up clasts, edge-wise flat pebble conglomerate (fig. 10), and dessication cracks (figs. 12 and 13).

Abundant quartz/dolomite pseudomorphs after displacive hopper halite crystals (Gornitz and Schreiber, 1981) occur in ceramic-grade talc throughout the district. The maximum dimension of these distorted pseudomorphs ranges from 1 to 10 cm (fig. 14). One occurrence of quartz and dolomite after anhydrite nodules was observed (fig. 15). Well-developed fracture cleavage is common, but not pervasive, in Allamoore ceramic-grade talc (figs. 10 and 11).

PAINT-GRADE TALC

Very high purity paint-grade talc occurs in only a few localities in the Allamoore district. Without exception these deposits show intensely deformed structure and are intruded by basic igneous rocks (fig. 16). The geometry of paint-grade deposits is accordingly quite complex. Paint-grade talc in the Allamoore Formation is essentially pure talc (tables 1 and 2). None of the disseminated quartz or dolomite so common in ceramic-grade deposits is present; however, quartz and dolomite do occur as remnants of original beds and as veins. The only microscopic contaminant detected was minor penninite at two localities. Calcite and quartz were observed in close proximity to paint-grade deposits as fracture fillings and in tension gashes (fig. 17).

Paint-grade talc is readily distinguished from ceramic grade talc by color and by the absence of laminae and other relict sedimentary structures. Pale- to moderate-pink talc that fades to grayish pink or white on prolonged exposure to sunlight is most prominent. Very pale green and white talc is commonly found in contact with pink talc. Paint-grade talc grades laterally into ceramic-grade talc in at least two localities. Elsewhere, lateral relationships are not exposed. Well-developed foliation is pervasive in all paint-grade-talc deposits. In thin foliae, the talc is translucent and slightly transparent.

Talc origin

Previous writers have attributed the origin of talc in the Allamore Formation to regional metamorphism of siliceous

FIGURE 16—PAINT-GRADE TALC DEPOSIT, ALLAMOORE FORMATION WITH IGNEOUS INTRUSIVE MASS (ARROW) ADJACENT TO TALC (T).

FIGURE 17—CALCITE FILLING TENSION GASHES IN CARBONACEOUS DOLOSTONE. ALLAMOORE FORMATION.

dolostone (Flawn, 1958; Chidester and others, 1964; King, 1965; Ross and others, 1968; Rohrbacher, 1973; Denison, 1980) or magnesium-rich tuff (King and Flawn, 1953; Flawn, 1958; Chidester and others, 1964; King, 1965; Davidson, 1980; Edwards, 1980b).

Chemical and x-ray-diffraction analyses, field observations, mineralogical and chemical considerations fail to support either of these hypotheses for the following reasons:

1) Talc formation from siliceous dolostone generates three parts calcite for each part talc; however, Allamoore talc contains no calcite (table 2), nor is there evidence of a calcite depositional front in surrounding rocks.
2) Well-preserved primary sedimentary structures in talc (figs. 10, 11, and 13) attest to the fact that no volume loss due to dissolution of calcite has occurred.
3) That only isolated siliceous dolostones could be altered to talc while other siliceous dolostones, which are in immediate contact, could be totally unaffected is unlikely (fig. 9).
4) Talc formed from magnesian tuff would, of necessity, contain large quantities of aluminum, but Allamoore talc contains negligible quantities of aluminum (table 1).
5) No minerals commonly associated with low-grade metamorphism of magnesium-rich (mafic or ultramafic) igneous rocks are present (for example, hornblende or biotite).
6) Fluorine content in Allamoore talc exceeds the maximum determined by Ross and others (1968) for talc formed from mafic igneous rocks.

SEDIMENTARY-MAGNESITE HYPOTHESIS

Allamoore talc was formed, probably in Precambrian time, by hydrothermal alteration of sedimentary magnesite by the reaction:

$$3MgCO_3 + 4SiO_2 + H_2O = Mg_3Si_4O_{10}(OH)_2 + 3CO_2$$
$$(magnesite) + (quartz) + H_2O = (talc)$$
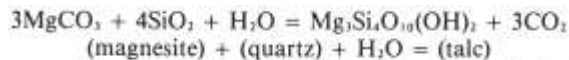
Quartz and magnesite essential to the process were derived in situ, thus requiring that only water, either connate or of meteoric origin, and heat be added to the system to promote the reaction. The process occurred at fluid pressures between 330 and 1,000 bars and at temperatures between 300° and
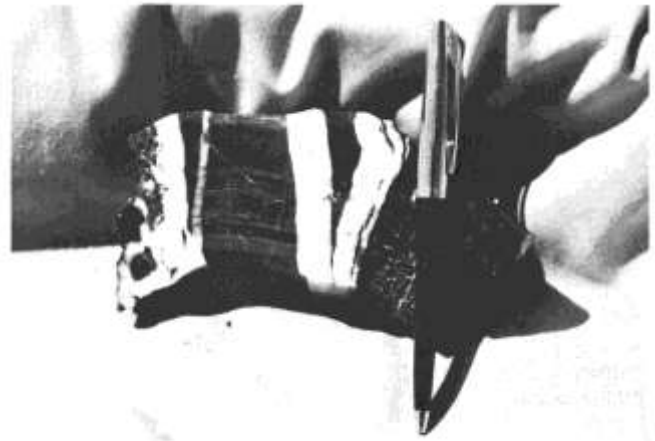
360°C. During talc formation, quartz in excess of that required to form talc from magnesite failed to react with contained dolomite due to low temperature and pressure. For the same reason, underlying and overlying siliceous dolostones were totally unaffected by the process.

Johannes (1969) formed talc from quartz and magnesite at 300°C and 330 bars. Skippen (1974) demonstrated equilibrium in the reaction:

$$3MgCO_3 + 4SiO_2 + H_2O = Mg_3Si_4O_{10}(OH)_2 + 3CO_2$$
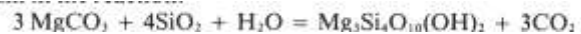
at 300°C and 1 kb. Greenwood (1967) established the upper limit of this reaction at 450°C and 2 kb, where the talc-magnesite-quartz assemblage gives way to talc-magnesite-quartz-f orsterite.

By the reaction:

$$3Ca,Mg(CO_3) + 4SiO_2 + H_2O = Mg_3Si_4O_{10}(OH)_2 + 3CaCO_3 + 3CO_2$$
$$(dolomite) + (quartz) = (talc) + (calcite)$$
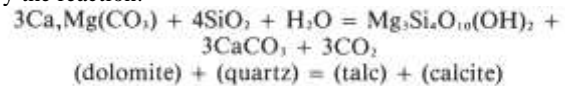
Gordon and Greenwood (1970) formed talc at 2 kb only when T > 460°C. Skippen (1971) determined that the above reaction proceeds to the right at 2 kb only when T > 417 °C. Similarly, Winkler (1974) showed formation of talc by this reaction at 1 kb only when T > 400 °C.

Asbestiform richterite in the Allamoore Formation has been identified by Rohrbacher (1973) at the Diablo prospect and the Buck Mine. He believes that these occurrences represent concentrations of sodium and potassium, and that the richterite formed at low temperatures.

Paint-grade talc in the Allamoore Formation was derived from previously formed ceramic-grade talc as a result of subsequent intrusive igneous activity. The intrusive bodies locally elevated temperatures to the point where dolomite and quartz within the previously formed ceramic-grade talc reacted to form additional talc, thus resulting in talc bodies of high purity.

CARBONATE-DEPOSITIONAL MODEL

Carbonates of the Allamoore Formation were precipitated in restricted marine ragoons and associated ephemeral lakes. Dolomite precipitation was favored in the intertidal and supratidal zones of restricted lagoons. Magnesite, dolomite, and amorphous silica were precipitated in isolated hypersaline lagoons or in associated ephemeral saline lakes.

Elliptical stromatolites, nodular and oncolitic dolostones, crossbedded, arenaceous dolostone, and dessication cracks, all encountered in intertidal or supratidal environments (Heckel, 1972; Lucia, 1972; Wray, 1977), are common in the massive dolostone sequences of the Allamoore Formation. Planar microbial laminae, rip-up clasts, displacive hopper halite crystals (as pseudomorphs), edge-wise flat pebble con-

glomerate, anhydrite nodules (as casts), dessication cracks, and interlayered amorphous chert, all characteristic of restricted supratidal or ephemeral lake environments (Heckel, 1972; Lucia, 1972; von der Borch, 1976; Gornitz and Schrieber, 1981) are common in ceramic-grade-talc deposits of the Allamoore Formation. As discussed elsewhere in this paper, ceramic-grade talc was formed from primary magnesite deposits that are characteristic of saline ephemeral lakes associated with restricted lagoons (von der Borch, 1965, 1976).

Massive Precambrian magnesite deposits in dolostone in Manchuria are believed to have been precipitated in partially enclosed lagoons in a warm, arid, or semi-arid climate (Nishihara, 1956). A similar chain of lentiform deposits of magnesite in Southern Pithoragarh, Kumaun Himalya, India, is described by Valdiya (1968). Similarities between the Indian magnesites and talc of the Allamoore Formation are numerous.

The time of formation of the Indian deposits is not known, but Valdiya (1968) feels that the stromatolites fix the age between Algonkian and early Ordovician. Denison and Hetherington (1969) have fixed the minimum age of the Allamoore Formation at 950 m.y. The Indian magnesite beds are lentiform and occur along a linear trend approximately 50 km long. Allamoore ceramic-grade-talc beds are lenticular to tabular and occur along a linear trend 48 km long. Individual Indian magnesite beds have strike-wise continuity up to 10 km. Allamoore magnesite beds are believed to have lateral (strike) extent approaching 5 km. Similarly, strike length is always much greater than thickness in both the Indian and the Allamoore deposits. Concordant alternation of dolostone and magnesite (talc in the Allamoore) is also observed in both regions. Mineralogy of the Indian deposits consists of "predominantly magnesite with small quantities of dolomite, siderite, breunerite, chert and talc." (Valdiya, 1968, p. 926). Furthermore, "chemical composition of the magnesite is uniform throughout . . ." (Valdiya, 1968, p. 928). Similar composition prior to talc formation is evident in the Allamoore Formation.

The relationship between stromatolitic dolostone and primary magnesite in the Allamoore Formation parallels the association described by Valdiya (p. 932):

> The almost invariable association of stromatolites and the magnesite deposits is certainly not fortuitous. This association suggests that algae played an active and significant part in the origin of these deposits. Marine algae, now represented by abundant stromatolites, must have formed biohermal barriers in the shallower coastal belt of the Gongolihat sea, and given rise to coastal lagoons, barred embayments and ponds in the back-reef shelf. The algae must have also created biochemical conditions favorable for the formation of the carbonates of higher Mg-content. These conditions were: Increased pH, possibly 10 or above, by utilization of CO_2; liberation of ammonia resulting from death and decay. . . . the death and accumulation of algal debris might have added to the concentration of magnesium of the basin waters. Chilinger's studies (1960) have demonstrated that the Mg/Ca ratio of carbonate rocks increases towards the shore, which is attributed to the abundance of Mg-rich algal sediments in the near-shore waters.

This commentary, written to describe deposits in India, could have been written as well for the magnesite deposits from which talc was derived in the Allamoore Formation.

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FLUORSPAR DEPOSITS OF CHRISTMAS MOUNTAINS DISTRICT, BREWSTER COUNTY, TEXAS

by Franklin W. Daugherty, *D and F Minerals, Inc., Alpine, Texas 79830*

Abstract

The newest fluorspar district in the United States is located in the Christmas Mountains adjacent to Big Bend National Park. Production from September 1971 to the end of 1980 was approximately 70,000 tons of metallurgical-grade fluorspar.

Practically all fluorspar mined thus far consists of relatively small isolated bodies replacing Cretaceous limestone adjacent to Paisano Peak, a prominent landmark near the northwestern edge of the Christmas Mountains. The peak is a deroofed Tertiary laccolithic intrusion of sodic rhyolite about 1 mi in diameter. A broad sill of rhyolite to the north may be an apophysis of the laccolith.

The bulk of the production has been obtained from the Buda Limestone overlying the northern edge of the laccolith and the adjacent sill and along the laccolithic contact on the east. Recently an exploratory adit disclosed fluorspar mineralization in the Santa Elena Limestone beneath the floor of the laccolith. Noneconomic bodies of fluorspar also are found in the Boquillas Formation along the rhyolite contact, in addition to minor veinlets and replacements in the rhyolite. In the higher elevations of the Christmas Mountains, fluorspar occurs in the Santa Elena Limestone near contacts of rhyolitic intrusions.

The principal gangue minerals are calcite and quartz, with minor pyrite and siderite. Some of the fluorite is anomalously high in uranium, thorium, beryllium, and molybdenum.

The fluorspar is of hydrothermal rather than contact metasomatic origin, in spite of its close proximity to rhyolitic intrusions. Ground preparation, mostly fractures and faults of small displacement, controlled localization of mineralization.

A tentative hypothesis offered to explain the source of the ore-forming solutions involves: a) generation of sodium-rich magma in the mantle along a subduction zone, b) continental rifting resulting from tension developed as subduction ceased, c) upward migration of magma accompanied by an increase in silica content by partial assimilation of crustal rocks to produce sodic rhyolite, and d) separation of a hydrothermal phase enriched in fluorine, uranium, thorium, beryllium, and molybdenum.

Introduction

The only fluorspar deposits of economic importance presently known in Texas are located in the Christmas Mountains fluorspar district in Brewster County, adjacent to Big Bend National Park (fig. l). Approximately 70,000 tons of direct-shipping, metallurgical-grade fluorspar has been produced from the district from the inception of mining activities in late 1971 through the end of 1980.

Although the occurrence of fluorite was recognized in the nearby Terlingua mercury district (fig. 1) at least as early as 1945 (Yates and Thompson, 1959), no exploration for fluorspar deposits was conducted prior to 1959. The discovery and exploitation of very large fluorspar deposits in northern Coahuila, 50 mi to the east, led the Dow Chemical Company to conduct a reconnaissance exploration program in Trans-Pecos Texas in 1952. This program, under the direction of W. N. McAnulty, Sr. and Stephen E. Clabaugh, resulted in the discovery of several fluorspar occurrences in the Christmas Mountains. Additional discoveries were made the following year. In 1971, D and F Minerals, Inc. commenced mining ac-
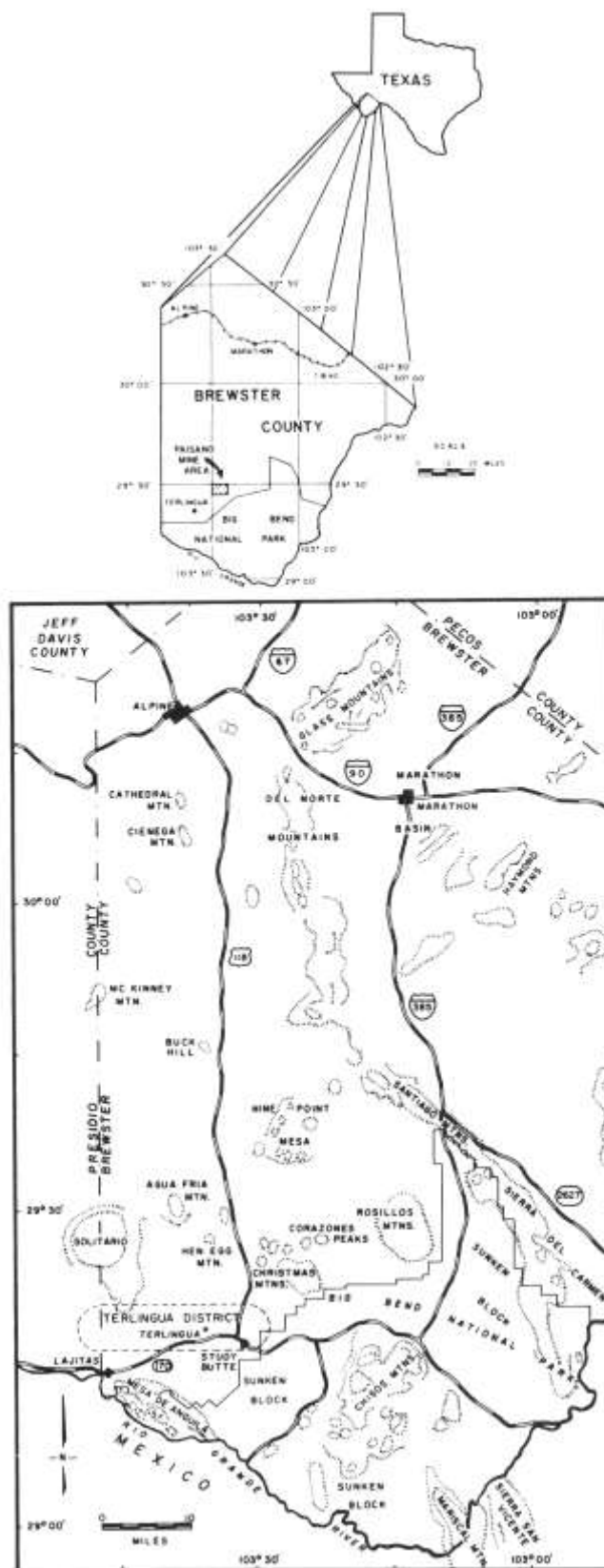
FIGURE 1—INDEX MAPS OF CHRISTMAS MOUNTAINS AREA.

tivities on the northern flank of Paisano Peak, a prominent landmark on the northwest edge of the Christmas Mountains.

The first detailed geological study of the Christmas Mountains was done by Bloomer (1949). Maxwell and others (1967) included the Christmas Mountains in a much larger study of Big Bend National Park. Fluorspar deposits of the Christmas Mountains have been described by W. N. McAnulty, Sr. (1967, 1974), Stevens (1974), and Daugherty and Fandrich (1978).

Geologic setting

The Christmas Mountains are part of an arid mountainous region within the Mexican Highlands section of the Basin and Range physiographic province. Widespread Late Tertiary igneous activity, basin-and-range-type block faulting, and ensu

ing differential erosion have carved out an area of spectacular topographic relief. The highest peaks are composed of highly resistant rhyolite in the form of stocks and laccoliths. Steep escarpments, hogbacks, and cuestas are commonly formed where resistant massive limestone of Early Cretaceous age has been uplifted by an igneous intrusion or along a normal fault.

The Christmas Mountains lie within a structurally depressed area known as the "sunken block" (fig. 1). The sunken block is characterized by large northwest-trending normal folds and faults (parallel to the boundary faults), and by abundant igneous intrusions. The Christmas Mountains are a northwest-trending elliptical dome of Cretaceous sedimentary rock, but the domal structure has been complicated by block faulting and the intrusion of many diverse igneous bodies (fig. 2). The origin of the dome is uncertain, but it may have been initiated by Laramide folding and accentuated by a
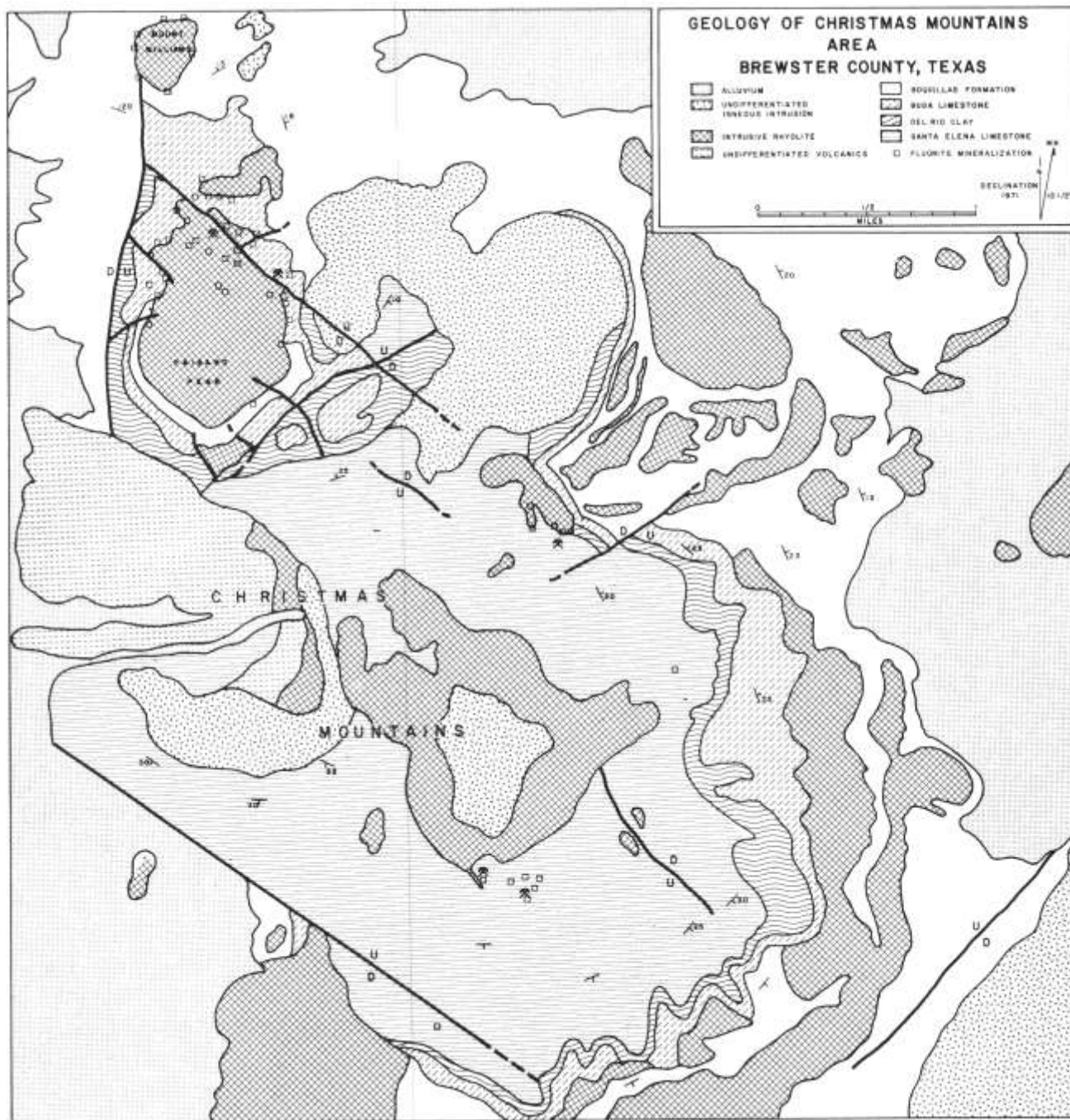
FIGURE 2—GEOLOGY OF CHRISTMAS MOUNTAINS AREA, BREWSTER COUNTY, TEXAS.

later intrusion of a large gabbroic laccolith in the core of the range.

Intrusive bodies include dikes, sills, laccoliths, and stocks. The igneous rocks are part of the Early Tertiary, alkalic igneous province of west Texas and can be classified as a) rhyolite and trachyte, b) peralkaline rhyolite and trachyte, c) analcime-bearing basic rocks, and d) analcime-free basic rocks. Except for the gabbroic body that forms the core of the range, the basic rocks occur mostly as a few small dikes and sills.

The sedimentary rocks exposed in the Christmas Mountains are entirely of Cretaceous age with the exception of some Quaternary alluvium and colluvium. A summary of the Cretaceous stratigraphy is presented in fig. 3. Rocks of the Comanche Series (Cretaceous) include the Santa Elena Limestone, Del Rio Clay, and Buda Limestone. The overlying Gulf Series is represented by the Boquillas, Pen, and Aguja Formations. Excellent descriptions of the stratigraphy and paleontology are given by Maxwell and others (1967) to whom the reader is referred for additional information.

Fluorspar deposits

Practically all the fluorspar mined thus far consists of a number of relatively small replacement deposits in Cretaceous limestone adjacent to Paisano Peak (fig. 4), a deroofed laccolith of peralkaline rhyolite approximately 1 mi in diameter. Most of the fluorspar occurs as a fine-grained replacement of Buda Limestone in mantos above sill-like apophyses of the laccolith on the north and northeast sides of Paisano Peak (fig. 5) and near-vertical discontinuous bodies along or above the edge of the laccolith (fig. 6).

Some fluorspar has been produced from replacement bodies of Santa Elena Limestone beneath the floor of the Paisano Peak laccolith. Siliceous fluorspar is known to exist as

FIGURE 4—SOUTHWESTWARD VIEW OF PAISANO PEAK.

replacement of flaggy limestone of the Boquillas Formation along the southeast edge of the laccolith. Small quantities of fluorspar also have been found as veinlets in and replacement of rhyolite of the Paisano Peak laccolith and related bodies. In several localities in the higher parts of the Christmas Mountains, open-space filling of fluorspar occurs in Santa Elena Limestone in veins and collapse structures (sinks) and as replacement bodies near contacts with rhyolitic intrusions.

The principal gangue minerals are calcite and quartz, sometimes accompanied by minor siderite and pyrite. Some of the fluorite is anomalously high in beryllium, molybdenum, uranium, and thorium.

The average composition of the fluorspar produced is approximately 80% CaF_2, 5% SiO_2, and 10% $CaCO_3$ for a grade of 67.5% effective CaF_2. The effective CaF_2 grade is calculated by deducting a penalty of 2 1/2 times the percentage of silica in an ore from the percentage of calcium fluoride. Thus, an ore with 90% CaF_2 and 8% SiO_2 has a grade of 70% effective CaF_2.

The fluorspar occurs in a variety of colors: white, light to dark gray, green, purple, and black. Replacement fluorspar is very fine grained, almost aphanitic. In some cases, replacement has been so perfect as to preserve sedimentary structures and fossils. Well-crystallized fluorite is rare, but some cubes up to 6-8 mm in size are found filling open spaces.

The effect of local controls on the deposition of fluorspar is obvious in most of the deposits. Fluorspar along the edges of rhyolitic intrusions is localized by brecciation of the limestone, created by emplacement of the igneous body. Where fluorspar occurs in limestone above concordant igneous intrusions, mineralization is confined to discontinuous linear zones overlying fractures or faults that cut the rhyolite and overlying limestone. Where fractures or faults intersect, mineraliza-

FIGURE 3—CRETACEOUS STRATIGRAPHY OF CHRISTMAS MOUNTAINS.

FIGURE 5—FLUORSPAR MANTO (CaF_2) IN BUDA LIMESTONE.
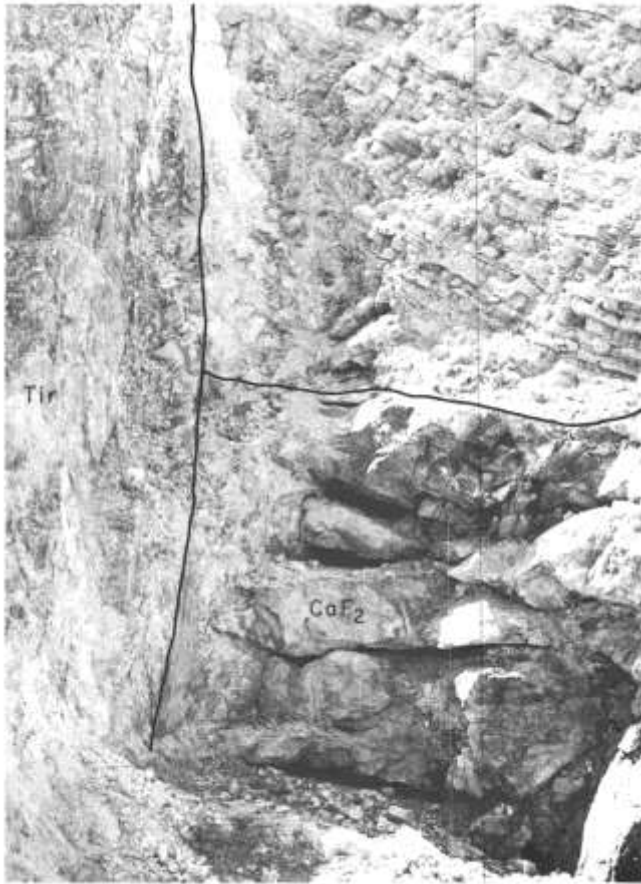
FIGURE 6—BUDA LIMESTONE REPLACED BY FLUORSPAR (CaF_2) ALONG VERTICAL CONTACT WITH RHYOLITE (Tir).

tion is thicker and more extensive. Where limestone and the underlying rhyolite is dense and relatively unbroken, little or no fluorspar is found along the contact.

Origin of fluorspar deposits

The persistent association of fluorspar deposits with peralkaline rhyolites in the Christmas Mountains and in the nearby Pico Etereo fluorspar district of Coahuila is striking. This association has led more than a few geologists to hypothesize that these deposits are the result of igneous metamorphism or contact metasomatism; that is, fluids emanating from the rhyolite provided fluorine that reacted with the limestone to form fluorite.

Critical evaluation of all available evidence, however, indicates that the rhyolite was not the source of the mineralizing fluids. For example, fluorspar deposits are not invariably associated with peralkaline rhyolites in contact with carbonate rocks. In fact, few of the peralkaline rhyolites in the Christmas Mountains and Pico Etereo districts have any associated fluorite. In addition, the fluorspar deposits associated with any given rhyolitic body are localized only in settings where the permeability of the rhyolite and limestone has been enhanced by fracturing, faulting, or brecciation. Furthermore, veins of fluorite in rhyolite are not at all uncommon and clearly show that the fluorite mineralization postdated intrusion of the rhyolitic magma. The evidence points to a hydrothermal origin for the fluorspar deposits.

A strong case can be made, however, that the fluorspar deposits are comagmatic or consanguineous with the peralkaline rhyolites. Most of the fluorite associated with peralkaline rhyolite of both the Christmas Mountains and the Pico Etereo districts is irregularly enriched in beryllium, molybdenum, uranium, and thorium. Fluorspar associated with more nor-

mal rhyolites, or not associated with rhyolites at all, have negligible concentrations of these elements. These observations lead to the conclusion that the F-Be-Mo-U-Th association is related to a high concentration of alkalies (Na and K) in the parental magma. Furthermore, this association also suggests that these elements were transported as fluoride-complex compounds.

Geologists have noted that many fluorspar deposits of the western United States are localized along or near deep-seated rifts. This observation is often cited as evidence that large volumes of fluorine-bearing solutions migrated upward from the lower crust or mantle. Van Alstine (1976) described the association of continental rifts and fluorspar deposits in the western United States with the observation that these deposits are often associated with barite, zinc, molybdenum, uranium, thorium, niobium, tin, beryllium, and rare-earth mineralization. Van Alstine (1976) also defined a sinuous belt of fluorspar deposits along a continental rift zone extending from the Big Bend region of Texas to Montana. The belt extends from the Christmas Mountains northwest to southern New Mexico, thence north through New Mexico and Colorado, and finally northwest through Wyoming, Idaho, and Montana.

A recent paper by Lamarre and Hodder (1978) explains Van Alstine's belt of fluorspar deposits, and a less well-defined western belt, in terms of plate tectonics. Their model involves subduction of oceanic crust beneath continental crust in two imbricate subduction zones of Early to Middle Tertiary age. The alkalic igneous rocks associated with the fluorspar deposits are believed to have originated by melting of oceanic lithosphere beneath the continental crust at depths of 300 km or more. Fluorine is derived by melting of phlogopite and concentrated in the magma by bonding with lithophile elements such as K, Na, Ca, and Mg. The silica content of rising alkalic magma increases by assimilation of $SiO2$ from crustal rocks, allowing much of the fluorine to be partitioned to the vapor phase as SiF4. Continental rifting resulting from tension following cessation of subduction provides avenues for the ascent of SiF4 to the upper crust. Reaction of SiF4 with connate and meteoric waters provides HF and SiO_2 which are precipitated as fluorite and quartz.

Following in part the model suggested by Lamarre and Hodder, a tentative hypothesis offered to explain the origin of the ore-forming solutions of the Christmas Mountains fluorspar district involves: a) generation of sodium-rich magma in the mantle along a subduction zone, b) continental rifting resulting from tension developed as subduction ceased, c) upward migration of magma accompanied by an increase in silica content by partial assimilation of crustal rocks to produce sodic rhyolite, and d) separation of a hydrothermal phase enriched in fluorine, uranium, thorium, beryllium, and molybdenum.

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GEOLOGY OF THE PERLITE DEPOSIT AT NO AGUA PEAKS, NEW MEXICO

by David Whitson, *Johns-Manville Sales Corp., Denver, Colorado 80225*

Abstract

Two partially dissected endogenous rhyolitic domes and several associated coulees form the No Agua Peaks. Hydration of the glass margins of the extrusion structures has produced a perlite deposit considered to be the world's largest. These domal structures and coulees are the result of at least two separate episodes of volcanism during the late Tertiary. Some of the outstanding features of the No Agua domes include similar internal structures and lithologic characteristics that are also partially reflected within the coulees. Field examinations and petrographic studies of the rocks forming the domes have identified three distinct, concentric lithologic zones within the structures: 1) a devitrified felsic core surrounded by 2) an interior, partially devitrified glass envelope, which in turn is surrounded by 3) the exterior glass unit. Both glass envelopes show a systematic change in physical characteristics toward the flow margins. Almost complete hydration of the enveloping glass units of the domes has resulted in the development of perlite structures in the glasses coupled with an increase in the average total combined water content from approximately 0.20 to 2.9%. Marked differences in the combined or hydroxyl water content of the glasses is useful in identifying perlites of differing relative ages at No Agua.

Introduction

The No Agua Peaks which rise conspicuously above the Taos Plateau volcanic field of north-central New Mexico, are composed of late Tertiary extrusive rhyolitic rocks dated at 4.2 m.y. (I. I. Friedman, 1980, personal communication), in contrast to the compositionally diverse basaltic lava flows of the surrounding plateau (Lipman and Mehnert, 1979). The peaks cover an area of 4 mil and consist of a group of four prominent peaks that represent the erosional remnants of two rhyolitic dome structures and a less conspicuous group of hills and ridges that make up the low hills area. The low hills area consists of several rhyolitic flows that are believed to have been derived from the lower western flanks of the westernmost prominent peak.

Secondary hydration processes (Friedman and others, 1966; Laursen and Lanford, 1978) acting upon the glassy rocks of the exterior chill margins of the rhyolitic extrusive bodies have resulted in nearly complete perlitization of the obsidians. Over geologic time, this process has produced the world's largest single perlite deposit. Mining operations here currently account for more than two-thirds of American production.

This paper interprets data collected from surface geologic mapping and extensive diamond-core drilling. Any discussion of the type occurrence of economically important perlite at No Agua has been excluded from the following interpretation because it has little bearing on this subject or on the resulting conclusions. In the following discussion the term "perlite" is used in the context of a combination of the original petrographic definition and the commercial definition. Perlite was originally defined as a glassy rhyolite possessing a pearl-like luster and exhibiting numerous concentric cracks resembling an onionskin (Johannsen, 1939). Commercially, the term applies today to any acid volcanic rock containing 2-5% water that upon rapid heating to the softening point of the glass, coincident with the volatilization of the combined water, expands or "pops" (Kadey, 1963). The use of a combination of these definitions is necessary, because the No Agua Peaks contain perlitic glasses that display the above commercial characteristics but do not show the classical perlitic characteristics described by Johannsen (1939).

The following interpretation of the geology of the No Agua Peaks is based on field investigations conducted since 1976 by field geologists with the Johns-Manville Sales Corporation's Exploration department. Included are descriptions of the stratigraphic relationships within the flow structures and an examination of preliminary geochemical data that has been collected from selected obsidian and felsite samples. These data, combined with surface geologic information, form the basis for interpretation of the structure of the extrusion features, which are manifested both as endogenous domes and as blocky rhyolite flows.

Previous studies of the No Agua rhyolites have included limited geochemical studies conducted by members of the U.S. Geological Survey (Zielinski and others, 1977; Friedman, 1980-81, personal communications), and a doctoral thesis on the geology and extrusion history of the perlites of the No Agua Peaks area (Naert, 1974). Because of the commercial nature of the deposit, a major need developed for further investigation and re-interpretation of the geology and extrusion mechanisms that were active during formation of the No Agua flow structures. The results of this investigation follow.

Stratigraphy

Rhyolitic structures that result from the extrusion of highly viscous magmas characteristically form steep-sided, areally restricted, block-type lava structures or domes. Previous field investigations have revealed silicic flow structures that, typically, are quenched at the exterior margins to form a glassy rhyolite (obsidian), while more slowly cooled interiors undergo primary crystallization to dense felsites (Ross and Smith, 1955; Friedman and Smith, 1958; Noble, 1968). Felsites may also result from complete devitrification of the parent glass. The complete development of such a cooling unit is dependent upon the temperature, thickness, and thermal gradient prevalent within the flow unit upon emplacement. A parallel process is the development of cooling units within welded ash-flow sheets (Smith, 1960).

A generalized graphical representation of the stratigraphy of the No Agua extrusive structures is shown in fig. l, and is representative of section A-A ' of fig. 2. The stratigraphic sequence changes laterally with regard to the glass envelopes relative to the source of extrusion. This change is manifested in relative vertical extension and compression of the exterior chill margin in relation to the proximity to the vent. Conversely, the interior felsite unit thickens toward the source of extrusion. A detailed description of the characteristics of the major identifiable lithologic units seen in the extrusive structures at No Agua follows. No attempt has been made to describe the small-scale flow structures within the glasses.

EXTERIOR CHILL MARGIN

The exterior chill margin consists of an obsidian developed at the margin of a rhyolitic flow structure where the cooling regime of the lava is dominated by a steep thermal gradient.
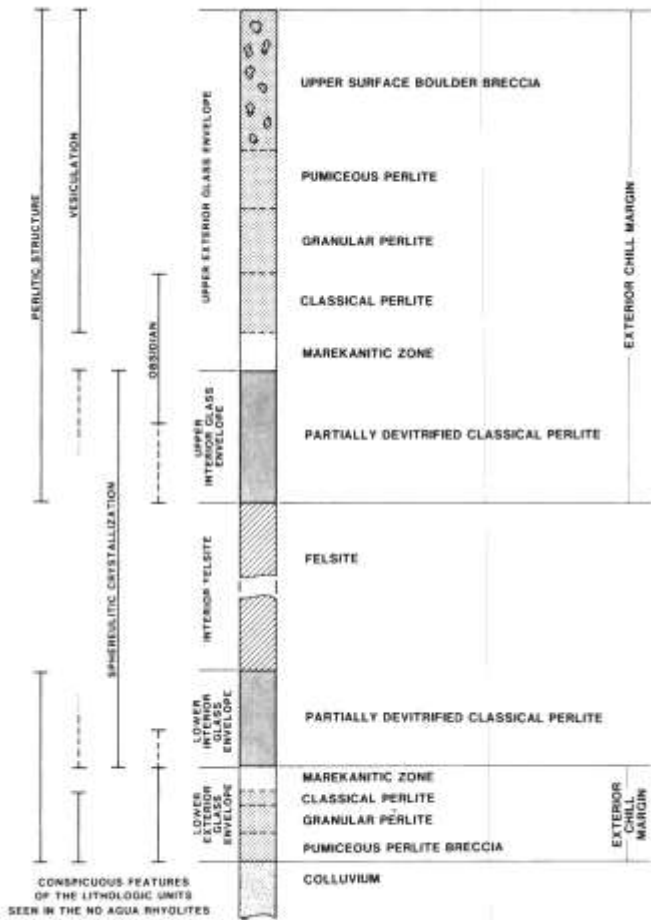
FIGURE 1—GENERALIZED STRATIGRAPHIC COLUMN OF THE NO AGUA PEAKS DOMES AND ASSOCIATED COULEES.

The result is a glass carapace, or envelope, surrounding a dense felsitic core.

The exterior chill margin, as observed in the structures of the No Agua Peaks, is divided into two distinct lithologic units, the exterior and interior glass envelopes. As the terms imply, these lithologic units are found in conjunction with each other in a concentric layered relationship (fig. 2).

The exterior glass envelope consists of a typically holohyaline, vesicular perlite that varies in physical characteristics as a function of its primary structure and texture. This envelope also contains the great majority of nonhydrated obsidian that remains in the rocks of the No Agua Peaks.

The interior glass envelope is composed of hypohyaline, perlitic glass, characteristically showing well developed classical perlitic structure and devitrification in the form of coarse, open spherulites distributed throughout the glass. Vesiculation within this unit is restricted to the outer extremities, where the vesicles are localized into flow laminae and are commonly lined with vapor-phase or hydrothermal-alteration products. The interior portion of this lithologic unit is commonly massive.
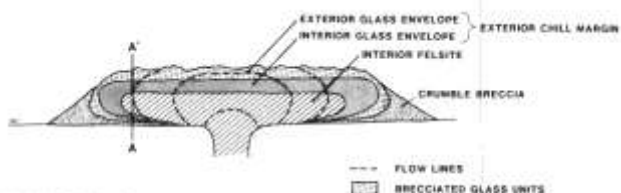
FIGURE 2—SCHEMATIC CROSS SECTION OF AN ENDOGENOUS DOME SHOWING IDEALIZED LITHOLOGIC ZONATION AS SEEN AT NO AGUA PEAKS, NEW MEXICO (adapted from MacDonald, 1972).

EXTERIOR GLASS ENVELOPE

The exterior glass unit consists of four distinctive petrologic subunits as identified by their physical and textural characteristics (fig. 1), which are manifested both megascopically and microscopically. These subunits result from primary vesiculation and autobrecciation of the glass during emplacement, and secondary hydration. The descriptions of the following rock types represent end-member textural types that dominate each subunit. The subunits should be considered as textural transition zones, because rocks displaying one unique texture are uncommon at the No Agua Peaks. This variability reflects the effects of differential shear and distortion of vesicles from differential confining lithostatic and hydrostatic pressures within the complex flow material.

The outer marginal breccia subunit is produced by autobrecciation (MacDonald, 1976) of the brittle expanding exterior margins of the rhyolitic extrusive bodies and is restricted to the outer pumiceous subunit that forms the upper surface of the dome and flow structures at No Agua. This subunit is not identifiable as a distinct type in the lower exterior glass unit, where autobrecciation is common within all of the subunits.

The outer marginal breccia is a moderately well-consolidated boulder and cobble breccia. Consolidation of the breccia results from the interlocking of the block boundaries, where the blocks comprise predominately pumiceous perlitic glass. However, welding at block boundaries has not been observed within this subunit at No Agua. This feature differentiates the subunit from the autobrecciated portions of the lower pumiceous subunit of the exterior glass envelope, which shows incipient to complete welding at the breccia-block boundaries.

Progressing toward the interior of the cooling unit, the pumiceous-perlite subunit is encountered. The glass that composes this subunit has a typically frothy to pumiceous appearance, with the vesicles showing a systematic and progressive distortion toward the lower portion of the subunit. Fig. 3 is a thin-section photomicrograph of a typical pumiceous perlite. The development of perlitic structure in the glass (Friedman and others, 1966) occurs only within the glassy junctions formed by intersecting vesicle walls. The lower manifestation of this subunit is generally quite thin. This thinness reflects the destruction of most primary vesiculation in the glass, due to compression caused by hydrostatic pressure and differential shear during emplacement.

FIGURE 3—THIN-SECTION PHOTOMICROGRAPH OF PUMICEOUS PERLITE OF THE EXTERIOR GLASS ENVELOPE SHOWING PARTIAL DISTORTION OF THE VESICLES (32.5 ×, transmitted light).

Inside of the pumiceous subunit is the dense perlite glass with a characteristic saccharoidal texture and blocky fracture that makes up the granular subunit. Fig. 4 is a thin-section photomicrograph of a typical granular perlite that is characterized by compressed vesicles throughout the glass, and a crude fracture system developed normal to the plane of flattening. Perlitic fracture structures within the glass are rare, but crude fracture and foliation patterns within the glass combine to impart the saccharoidal texture.

The innermost portion of the exterior glass envelope is the classical perlite subunit. This subunit, which consists of glassy rhyolite with a pearl-like luster and numerous concentric cracks resembling an onionskin, displays a brittle, spherical fracture habit and well-developed perlitic structure visible to the naked eye. This perlite is commonly light gray to brown and, in some cases, color banded, and contains obsidian nodules isolated in the central cores of perlitic fracture systems.

Obsidian, which occurs throughout the subunits of the exterior glass envelope in minor amounts, is most common in the classical perlite glass, where it shows a progressive increase toward the interior of the cooling unit. This results in a marekanitic zone at the base of the classical perlite subunit in which the obsidian makes up the bulk of the rock.

Fig. 5 is a photomicrograph of a classical perlite, showing the characteristic perlitic fractures in the glass. Vesicles are commonly localized in the axial regions of the folds during flow but are uncommon elsewhere within the glass.
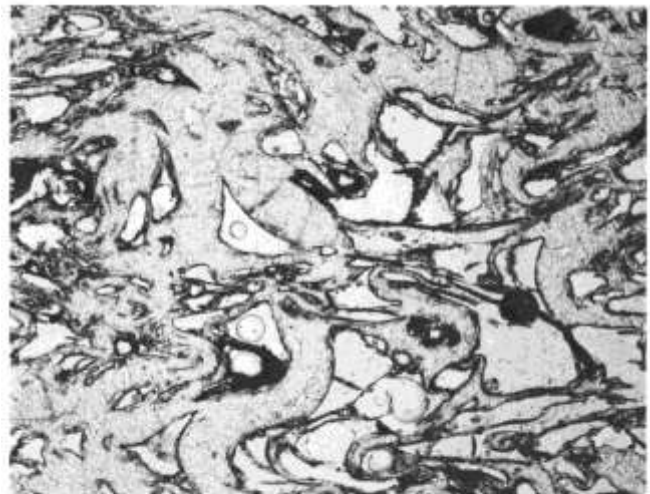
The exterior glass envelope, with its four subunits, is characterized by its vesicular nature and progressive distortion and localization of vesicles toward the interior. These characteristics reflect the lithostatic and hydrostatic pressure and flowage conditions present in the glass during emplacement. The presence of obsidian within the perlite, although not exclusively restricted to the exterior glass envelope of the outer chill margin, represents remnant primary obsidian that has not been subjected to secondary hydration. This condition is thought to be a function of isolation of the primary glass from the presence of meteoric water in unfractured glass zones within the flow structures. The time- and temperature-dependent process of hydration, coupled with devitrification and erosion of the metastable glass unit, would be expected to act in conjunction to destroy the definitive character of the exterior glass unit at No Agua Peaks. In much older rhyolitic extrusion features, such as those seen in the commercial per
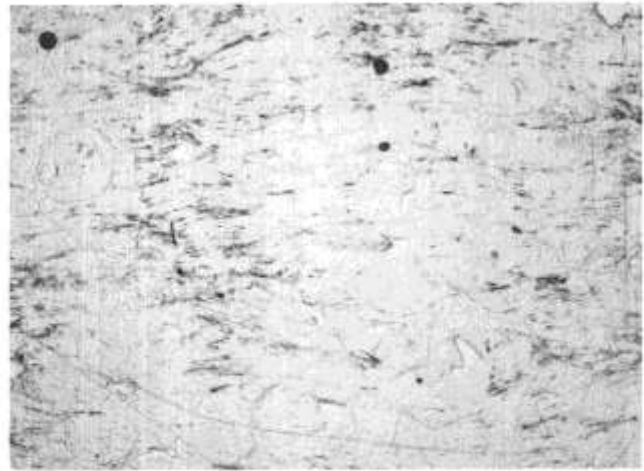
FIGURE 5—PHOTOMICROGRAPH OF A TYPICAL CLASSICAL PERLITE FROM THE EXTERIOR GLASS ENVELOPE. Distortion and localization of vesicles and the characteristic perlitic fracture are seen in upper left corner of photograph (32.5×, transmitted light).

lite deposits of Grants, New Mexico, and the Brushy Mountain perlite occurrence near No Agua, the units described above are notably absent.

INTERIOR GLASS ENVELOPE

The interior glass envelope forms the major portion of the exterior glass carapace of the extrusive bodies at No Agua. This lithologic unit is composed of partially devitrified classical perlite that shows many microscopic features seen in the classical perlite subunit described previously. The definitive features unique to this glass reflect the cooling history and secondary alteration of the parent glass and are evident from megascopic and microscopic studies.

In hand specimen, this partially devitrified classical perlite is green to greenish gray, with poorly developed perlitic structure, and commonly has a greasy or oily luster and spheroidal fracture that is typical of the classical perlites. Outcrop exposures are commonly flow laminated in the outer portion of the unit, becoming massive toward the interior of the cooling unit. Flow lamination in the outer extremities is primarily due to localization of vesicles into foliate layers that result from differential shear within the glass during emplacement. The vesicles are commonly lined with hydrothermal and vapor-phase minerals. Figs. 6 and 7 are photomicrographs of a partially devitrified classical perlite with well developed perlitic structure present throughout the hypohyaline glass. Devitrification of the metastable glass is apparent from the presence of coarse, open-structured spherulitic crystal groups of feldspar crystals arranged in divergent radial clusters. Strain birefringent halos are associated with these spherulites. Vesiculation within the massive glass is not present.

The spherulitic groups are both randomly dispersed throughout the glass and localized into axiolitic structures along flow bands composed of cryptocrystalline minerals. This type of spherulitic texture was produced experimentally in a rhyolitic glass under conditions of rapid cooling and water undersaturation. Subsequent secondary hydration in the presence of low-alkaline solutions takes place within the glass at temperatures below 200° C. This type of devitrification is described as "glassy stage" spherulitic crystallization (Lofgren, 1971). The number of spherulitic crystal groups within the devitrified classical perlite glass increases toward the interior margin of this lithologic unit and ranges from <1% to >25%. However, the forms and relative dimensions do not appear to vary greatly throughout the unit.
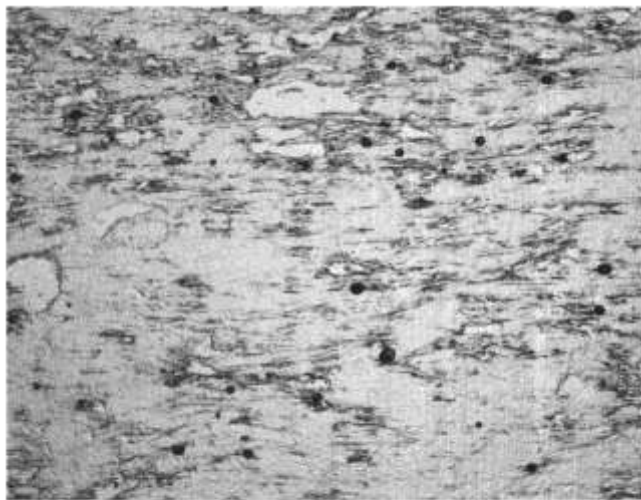

FIGURE 4—GRANULAR PERLITE OF THE EXTERIOR GLASS ENVELOPE SHOWN IN THIN SECTION. The glass shows marked distortion of the vesicles and notably lacks perlite structure (32.5×, transmitted light).
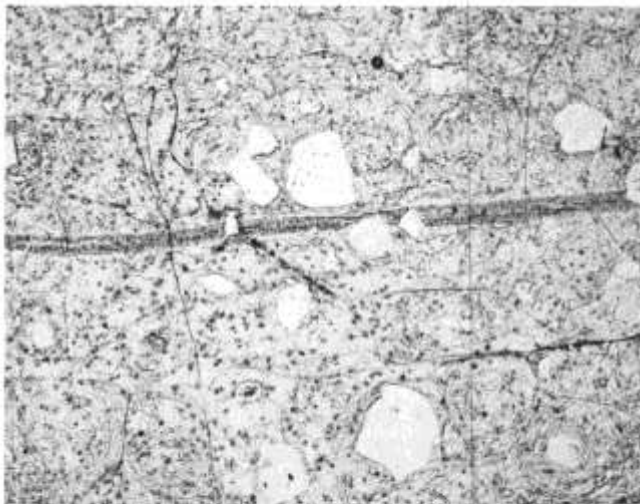
FIGURE 6—PHOTOMICROGRAPH OF A PARTIALLY DEVITRIFIED CLASSICAL PERLITE OF THE INTERIOR GLASS ENVELOPE showing well-developed perlitic structure and random dispersion of open, coarse, radial spherulite-crystal groups. Void spaces result from plucking of the centers of perlitic structures during thin-section preparation (32.5×, transmitted light).

FIGURE 8—PHOTOMICROGRAPH OF A SPHERULITIC BANDED FELSITE OF THE INTERIOR FELSITE UNIT. Only the continuous matrix of spherulites within the massive rock is shown. The fine, compact spherulite forms developed along flow banding are not shown. Note the mottled areas of the sheet matrix contain a micro-poikilitic texture (32.5×, transmitted light).

The abundance of nonhydrated glass within the interior glass envelope decreases markedly away from the marekanitic zone of the exterior glass envelope and toward the central portion of the extrusive body. The marekanitic zone transcends the upper boundary of the interior glass envelope in some instances, but rarely by more than several tens of feet. This zoning is most commonly seen in the glasses of the westernmost peak, where it is found within the massive glass, interior to the foliation planes.

INTERIOR FELSITE UNIT

The interior felsite unit, encountered in the central core of the cooling unit, consists of dense, hypocrystalline, massive-banded and spherulitic felsite and is found in sharp contact, sometimes interfingering, with the interior glass unit. The limited exposure of the felsite has thus far restricted study of the rock to the outer portions of the unit and to intrusive felsites—both dikes and sills.

A petrographic study of the felsites shows that they are derived from complete devitrification of a glassy rhyolitic material, rather than primary crystallization within the massive, slowly cooling interior of a flow structure. Figs. 8 and 9 are photomicrographs of a typical felsite exposed on the inner flank of the northernmost peak. Two basic types of spherulites are present in the rocks; l) compact, fine-structured spherulites (Keith and Padden, 1963) that are localized along flow bands and 2) coarse, open spherulites (Keith and Padden, 1963) that form a continuous-sheet network in the massive portions of the rock. The sheet network is formed by the coalescing of dispersion halos associated with the radial crystal groups. Superimposed locally upon the sheet are areas displaying a micropoikilitic texture (Lofgren, 1971). Similar spherulitic textures have been produced experimentally in natural rhyolitic glasses under conditions of slower cooling rates at higher temperatures, with greater availability of water than those conditions conducive to glassy-stage devitrification. This process has been described as the spherulitic stage of devitrification (Lofgren, 1971).

From the above description of the generalized stratigraphy, the development of the lithologic units is apparently depen-

FIGURE 7—PARTIALLY DEVITRIFIED CLASSICAL PERLITE SHOWN IN THIN SECTION. Notable strain birefringent halos are developed about the spherulitic crystal groups in the glass. Note the development of axiolitc structures along cryptocrystalline flow bands (32.5×, crossed nicols).

FIGURE 9—PHOTOMICROGRAPH OF FELSITE SHOWN IN FIG. 8 IN DOUBLY POLARIZED LIGHT.

dent upon the cooling history and secondary processes active within the extruded rock. Primary crystallization is manifested within the rocks of No Agua by the presence of flow-segregated bands of crypto- to microcrystalline minerals and by euhedral feldspar phenocrysts. Nowhere at No Agua, in the material thus far examined, do primary minerals constitute more than 1% of the rock mass.

The cooling unit consists of an initial rhyolitic glass which degasses in the outer margins and chills quickly to produce a holohyaline, ve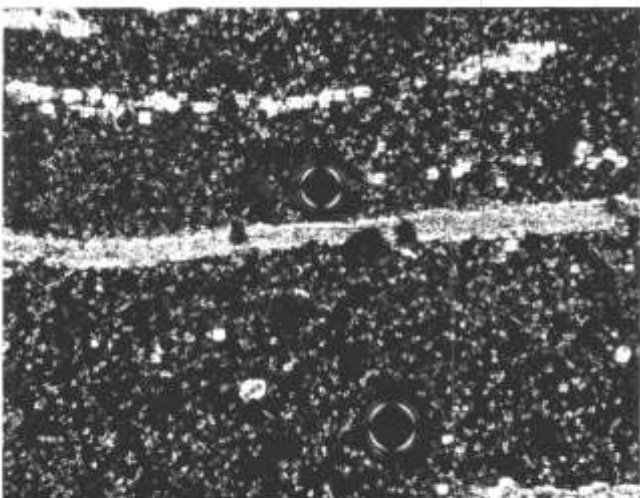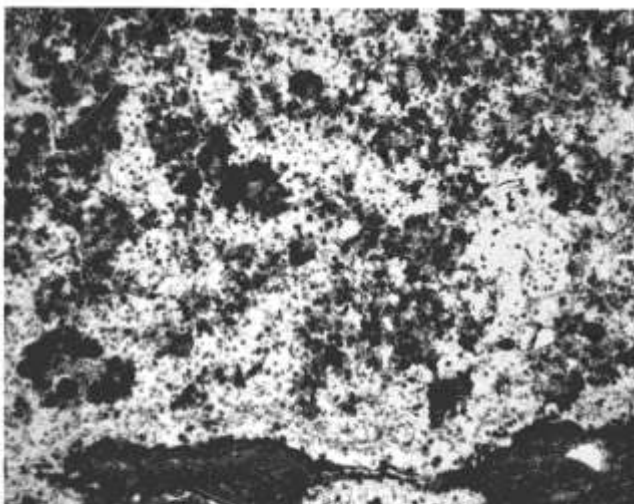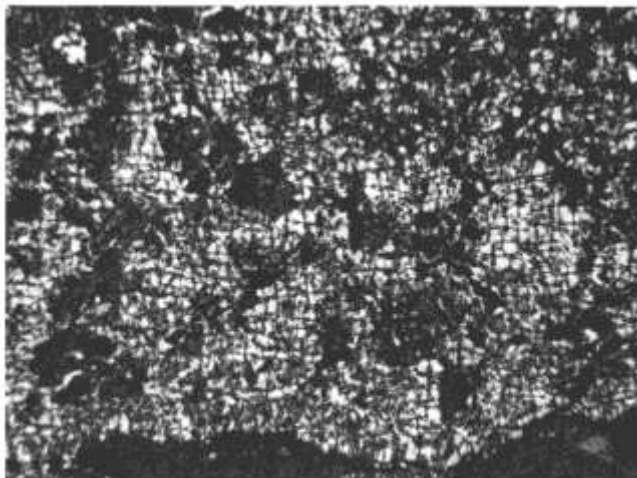sicular exterior margin during emplacement. The development of this margin creates an insulation layer that protects the flow interior and thus decreases the thermal gradient within the flow structure. The effects of the reduced thermal gradient are shown by the trends in abundance and development of spherulites in the interior lithologic units.

Surface geology and the geologic model

Recent workers have noted that extrusion of highly viscous magma, such as rhyolite or dacite, has been from central-vent or fissure-type eruptions. These extrusions typically are blocky sheet flows or domal structures that are formed by endogenous or exogenous processes (MacDonald, 1976). The resultant flow and dome structures are usually areally restricted and form steep-sided structures that typically have a high thickness-to-width ratio. Rhyolitic flows show a characteristic lithologic zonation from interior to exterior as a function of cooling (Christiansen and Lipman, 1966). Similar lithologic relationships exist within the interior of many rhyolite flows such as those seen at Beaver Creek in southern Colorado, the Devil's Punch Bowl at Mono Craters, California, and at the No Agua Peaks, New Mexico.

Glass chemistry

Obsidian samples were collected from the marekanitic portions of the exterior glass envelopes of all of the No Agua Peaks and, together with felsites, were analyzed for selected major, minor, and trace elements. These data were combined with existing data (Zielinski, 1977; Asaro, 1981) for interpretation.

Table 1 contains the results of major- and minor-element analyses from four selected obsidian samples, one derived from each of the peaks. Data indicate that the glasses are of typical silica-rich rhyolitic composition. No marked compositional differences between the glasses are seen. However, data collected from trace-element analyses of these obsidians show quite different results with respect to rubidium, strontium and barium.

Trace-element analyses of the glass and felsites from each of the four hills, including the obsidians and felsites of the low hills area are shown in fig. 10. Marked differences in trace-element abundances indicate the presence of rhyolitic material

FIGURE 10—SELECTED TRACE-ELEMENT ANALYSES OF THE NO AGUA PEAKS OBSIDIANS AND RELATED ROCKS.

of contrasting trace-element compositions. The rhyolitic material associated with the westernmost peak and low hills area is notably enriched in rubidium but depleted in strontium and barium relative to that of the north, east, and south hills of the deposit. These results indicate that at least two groups of rhyolitic glass exist within the No Agua Peaks area and are considered to be derived from two compositionally different magma sources. The rhyolites were extruded during separate episodes of volcanism at No Agua. The rocks of the No Agua Peaks are also anomalously depleted in barium relative to the surrounding flow rocks of the plateau. Fig. 11 shows the areal extent of the different extrusive rocks that form two overlapping, subround areas in plan view.

Domes of the No Agua Peaks

Two types of volcanic domes are recognized in the literature today. Exogenous domes are built by successive extrusion through a summit vent to form overlapping flows, and endogenous domes are formed by internal distension of a lava mass above a vent area. The recognized domal structures at No Agua are of the endogenous type.

TABLE 1—MAJOR- AND MINOR-ELEMENTAL ANALYSIS OF OBSIDIANS FROM THE LOWER MAREKANITIC ZONES OF THE NO AGUA PEAKS.

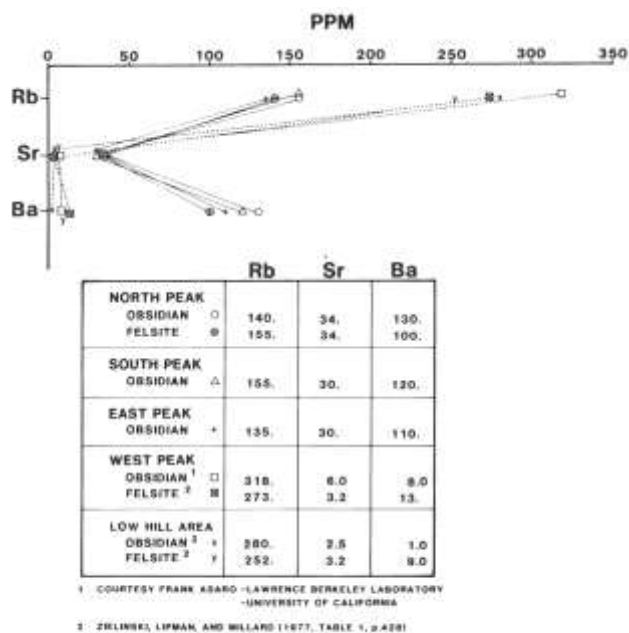| %	North Hill	South Hill	East Hill	West Hill
SiO_2	74.5	75.0	75.5	75.0
Al_2O_3	13.3	13.7	14.6	13.0
Fe_2O_3	0.7	0.9	0.85	1.2
CaO	1.3	1.1	0.55	1.3
MgO	0.3	0.35	0.1	0.25
Na_2O	2.3	3.5	3.1	4.1
K_2O	3.3	4.3	4.2	4.0
MnO	0.04	0.06	0.06	0.13
TiO_2	0.05	0.05	0.05	0.05
P_2O_5	0.04	0.05	0.04	0.05
H_2O+	0.03	0.03	0.04	0.03
H_2O-	0.07	0.2	0.21	0.43

FIGURE 11—NO AGUA PEAKS TOPOGRAPHIC MAP SHOWING TWO AREAS CONTAINING CHEMICALLY SIMILAR EXTRUSIVE RHYOLITIC ROCKS AS DETERMINED BY SELECTIVE TRACE-ELEMENT ANALYSIS.

MacDonald (1976) gives the following description of the extrusion and endogenous domal growth ". . . lava being squeezed up through a vent simply distends the mass above it so growth is somewhat like that of an expanding balloon. But the amount of stretching the cooling outer part of the growing dome can sustain is limited. The shell of the dome is literally torn open and lava oozes out through the fractures forming flows. . . ." A simplified schematic cross section of such an idealized dome has been adapted from MacDonald (1976) by superimposing a model of litholgic units as seen at No Agua over the dome (fig. 2). Rhyolitic rocks exposed at the No Agua Peaks show a continuous variation from felsite to pumiceous perlite. Only when the compositionally different rocks are separated do their relationships, both lithologic and structural, become discernible. Extrusive rocks associated with the northern, eastern, and southern peaks, as identified by their chemical composition, represent the erosional remnants of a large domal structure that resulted from the first extrusive rhyolitic extrusive episode at the No Agua Peaks. The outcrop pattern as seen on the geologic map of the No Agua Peaks area (fig. 12) is one of a bullseye pattern with the felsite forming the central mass.

This bullseye pattern has exposed a massive, banded, spherulitic felsite unit on the interior of the flanks of the north, east, and south peaks relative to the central depression area at No Agua that is formed by the arrangement of the four prominent peaks. Outside of the interior felsite core are the interior and exterior glass envelopes, respectively. Islands of the exterior glass envelope appear isolated within the interior glass envelope on the summits of the north and east peaks. These islands are erosional remnants of the upper portion of the exterior glass envelope (a manifestation of the third dimension).

This type of outcrop pattern can be produced by dissecting a model dome such as the one shown in cross section in fig. 2. The removal of a cone-shaped mass (the apex penetrating the interior felsite unit) from the interior portion of the dome (plan view) and normal erosion of the steep-sided exterior margins of the flow, would produce outcrop patterns similar to those seen at No Agua).

The structure of the original dome is reflected by flow laminae of the rock units. The internal structure of the dome is one of a massive, crudely bedded core changing to a well-foliated glassy structure in the distal portions of the dome. The outer foliation forms an upturned fan-shaped structure as is commonly seen in the distal portions of domes and blocky lava flows (Fenner, 1923; MacDonald, 1976). From the available structural evidence, the extrusion event forming the large dome appears to have been a central vent-type eruption from a vent located in the area occupied by the central depression of the No Agua Peaks.

Evidence from the major, identifiable flow directions indicates that the history of the dome is quite complicated compared with the distended mass of magma expanding as a balloon (MacDonald, 1976). Indeed, growth appears to have occurred by internal distension of a lava body and by the formation of lobes from the initial flow front of the dome. This is not the same mechanism as breaching. Rather, this mechanism is a function of irregular rates of flow in response to gravity, existing topographic barriers, or viscosity differences in the magma along an advancing flow front.

The younger structures of the westernmost peak and adjacent low hills, as determined by stratigraphic and chemical composition and relationships, comprise the rhyolitic extrusive rocks in the western portion of the No Agua Peaks area. The structures include a small endogenous dome, forming the westernmost peak, and several coulee flow structures.

The endogenous dome that forms the western peak, although similar in lithologic character to the dome previously described, differs in surface expression by the presence of a skirt of laharic debris and an exterior glass envelope along the interior flank of the hill, relative to the central valley depression. Interlayered, partially devitrified, classical perlite and felsite within the foliated portions of the interior glass envelope is common. This interlayering apparently is due to concordant intrusion of felsitic magma along contraction failure planes within the cooling interior glass envelope. A large discordant intrusion, observed in the northern outcrop exposures of the peak, penetrates both the interior and exterior glass envelopes and may have reached the surface and produced a minor flow atop the structure after emplacement of the dome.

The internal structure of the younger western peak dome differs markedly from the older dome. Although not entirely exposed by erosion, the internal structure is shown by concentric foliation planes. Layering within the interior glass envelope that is due to felsite produces a concentric structure much like the shells of an onion. A similar type of internal structure was described within a dacitic dome of Raker Peak, Lassen Volcanic National Park, California (Williams, 1932).

Rhyolite flows

Rhyolite flows are associated with domal structures of the No Agua Peaks and are found to have structural characteristics similar to the block lava flows described by MacDonald (1976) and Christiansen and Lipman (1966).

Rhyolite flows are most common within the low hills area where they form low hummocks and ridges. These coulee structures emanate from the lower exterior flanks of the dome structures, apparently through breaching of the expanding flow front. The flows show limited development of the stratigraphy characteristic of a cooling unit only in the thicker coulees of the low hills area. Commonly, the characteristics of only the exterior glass envelope are reflected in flow stratigraphy, and the flows generally have a thickly developed, upper, exterior, marginal-breccia subunit on top (fig. 13).

The major coulee structures at No Agua are located within the low hills area and were not accessible for this report. Because of this, a detailed examination of the well-developed rhyolitic structures was not made. The flow direction of these structures, as noted by general observations and discussions

Explanation

EXTERIOR GLASS ENVELOPE		ALLUVIUM	
INTERIOR GLASS ENVELOPE		BASALT	
FELSITE		MAREKANITIC ZONE	

SCALE 1:24000

FIGURE 12—GEOLOGY OF THE NO AGUA PEAKS

W E
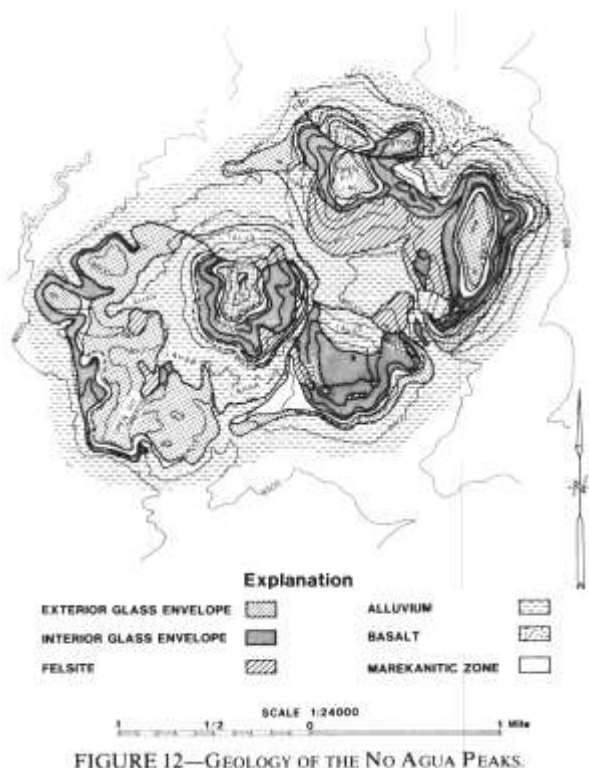
FIGURE 13—IDEALIZED CROSS SECTION OF THE NO AGUA RHYOLITE DOMES.

with Lipman, Friedman, and Zielinski of the U.S.G.S. in 1980, is to the west and southwest from the vicinity of the western flank of the westernmost peak. This orientation would suggest that flow directions were controlled by existing topographic barriers, such as a highland to the east.

Water content of the glasses

A preliminary study of the way in which water is bound within the perlitic glasses of the No Agua Peaks was undertaken recently and is still in progress. Preliminary data from this study indicate that the information gathered from such studies may be useful in identifying chemically different perlitic glasses of close relative ages and spatial arrangement. The obsidians of the No Agua Peaks have been found to contain 0.01-0.46% water although this varies considerably throughout equivalent glass samples. Water content of the perlitic glasses studied varies from 2.8 to 3.7%.

The percentage of the water in perlites that is found as molecular water or hydroxyl ions bonded in the glass has proven to be useful in identifying perlitic glasses derived from different parent glasses, resulting from the extrusion events at No Agua. These glasses were derived from the obsidians associated with the different extrusion events at No Agua.

Summary

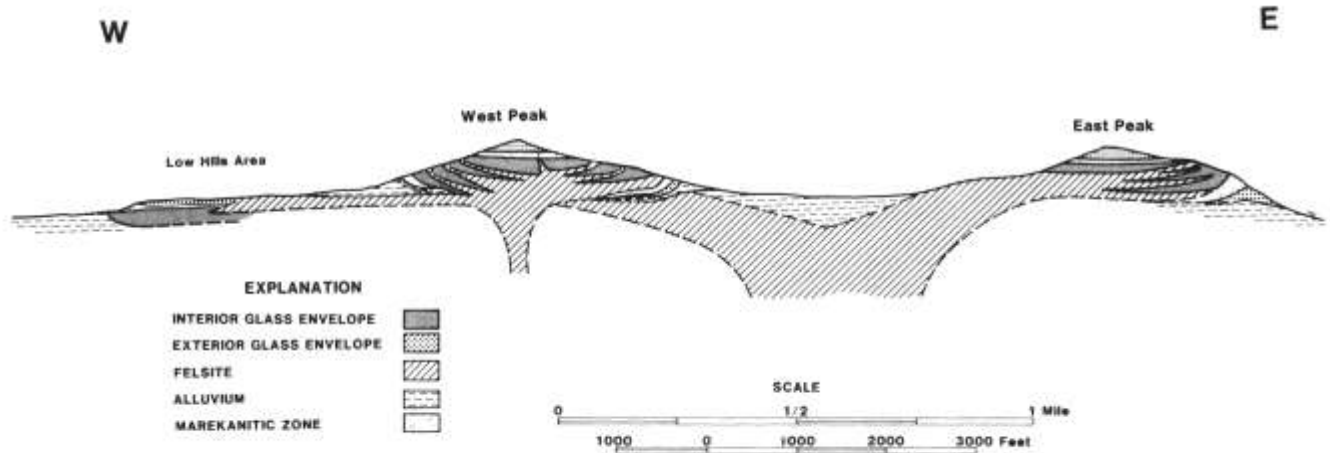The commercial investigation of the No Agua Peaks perlite deposits has provided a detailed examination of the major rhyolitic structures of this composite volcanic feature. Geologic surface mapping coupled with petrographic study of the rocks has identified three distinct lithologic units within the rhyolitic structures. These include the holohyaline, vesicular rocks of the exterior glass envelope, the hypohyaline, partially devitrified, classical perlites of the interior glass envelope, and the hypocrystalline rocks of the interior felsite unit. The first two units are portions of the chilled exterior carapace common to most rhyolitic flow structures.

The lithologic units are a manifestation of the cooling history of the rhyolitic glasses of the extrusive bodies as seen in the textural changes in vesiculation and in the degree of devitrification. Therefore, they are defined as a cooling unit. Application of this cooling-unit concept to the surface geology and analytical data from the obsidians occurring in the rock units has resulted in the identification of two endogenous domal structures in the No Agua Peaks area. Continuing investigations have revealed the existence of several block lava flows associated with the domes. The two domes and their associated coulee structures were formed by at least two eruptive events, neither of which are separated by a time interval detectable by potassium-argon dating methods.

A study of the nature of the water in perlitic glasses has corroborated the existence of two types of source rock. Further research is necessary, and such studies may be useful in future investigations of composite rhyolitic structures.

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PERLITE IN NEW MEXICO

by Robert H. Weber and George S. Austin, *New Mexico Bureau of Mines and Mineral Resources, Socorro, New Mexico 87801*

Abstract

539,000 short tons of domestic perlite valued at $14.4 million dollars were mined in New Mexico in 1980. Although production of crude perlite is reported by mines located in six states, New Mexico is the leading U.S. producer with 85-90% of total domestic production in recent years. The chief value of perlite is its ability to expand or "pop" when heated to temperatures above 1,400° F, producing a lightweight, nearly inert product. New Mexico has five active perlite mines. Two are located in the No Agua area and one is located at nearby Brushy Mountain in northwestern Taos County, north-central New Mexico. Another is in Valencia County about 13 km northeast of Grants. The fifth mine is located in central New Mexico, southwest of the town of Socorro. Industrial-grade perlites in the five operating open-pit mines consist largely of texturally nonperlitic, flow-banded, microvesicular glass in volcanic domes of rhyolitic composition with combined water contents ranging approximately from 2.5 to 3.5%. Age determinations indicate that perlite is largely restricted to Oligocene or younger volcanic units. Numerous other deposits, most of which are texturally classic "onionskin" perlites, are known in the western half of the state. However, these deposits remain largely undeveloped because of factors that include relatively small size, abrupt variations in physical properties, and relatively high costs of mining and transportation.

Introduction

The term perlite was originally applied to natural siliceous glasses of volcanic origin characterized by abundant concentric fractures and a pearly luster. This is classic or "onionskin" perlite (fig. 1). Today, the term perlite has been extended by industrial usage to include any naturally occurring siliceous volcanic glass that will expand appreciably when heated to a suitable point in its softening range. Crude perlite expands or "pops" because it contains between 2 and 6% combined water that flashes into steam upon heating. Expanded perlite is a cellular aggregate of low density (fig. 2); hence it is bulky, has large surface area, low heat and sound transmission, and an inert chemical nature for most environments.

Expansion of perlite is accomplished by injection of the properly sized perlite aggregate into gas- or oil-fired furnaces at temperatures of 1600 °-2000 °F. The specific gravity of

crude perlite ranges from 2.2 to 2.4 with a bulk weight of 139- 150 lbs/ft^3. Bulk weight of the expanded products ranges from about 2 to 20 lbs/ft^3.

Approximately one-half to two-thirds of perlite consumption is related to construction, commonly as aggregate for plaster, concrete, and insulating board, and for thermal insulation. The second major use is as filtration media. Lesser quantities are used as soil conditioners and soil substitutes, for packing material, and as extenders, fillers, and carriers.

The United States is by far the leading producer of crude perlite and consumer of expanded products. A total of 539,000 tons of crude perlite, valued at $14.4 million, was mined in New Mexico in 1980, according to the U.S. Bureau of Mines. New Mexico continues to be the principal producing state, contributing about 85-90% of the total crude, domestic perlite mined in recent years.

General description of deposits

Perlite occurs as mushroom-shaped and bulbous masses in volcanic domes and as short tongues in and near the vents of siliceous volcanoes; it also occurs as intrusive necks, dikes, and sills and in the basal zones of welded ash-flow tuffs. Glassy rocks, including obsidian and pitchstone, commonly are part of an eruptive sequence that includes pyroclastic and lithoidal phases. Individual bodies of perlite range from a few feet to several hundred feet thick and extend over areas of up to several square miles. Perlite domes are particularly favorable for commercial development because of their large size and shallow overburden, which permits low-cost mining by open-pit methods.

Most perlites have a rhyolitic chemical composition, but compositions extending into the dacitic range have been reported. Analyses of five volcanic glasses from New Mexico show them to be high-silica rhyolites (Weber, 1965, table 44). Crystalline inclusions are common, consisting of microlites, crystallites, and spherulites and phenocrysts of quartz, feldspar, biotite, hornblende, magnetite, hematite, and sphene.

Perlites are largely restricted to Oligocene or younger volcanic assemblages because they tend to devitrify with age into microcrystalline aggregates of quartz and feldspar. They also are subject to hydrothermal alteration to clay minerals and zeolites, commonly smectite and clinoptilolite.
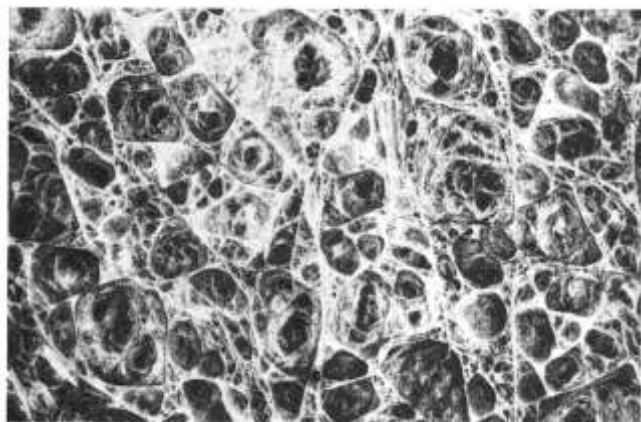

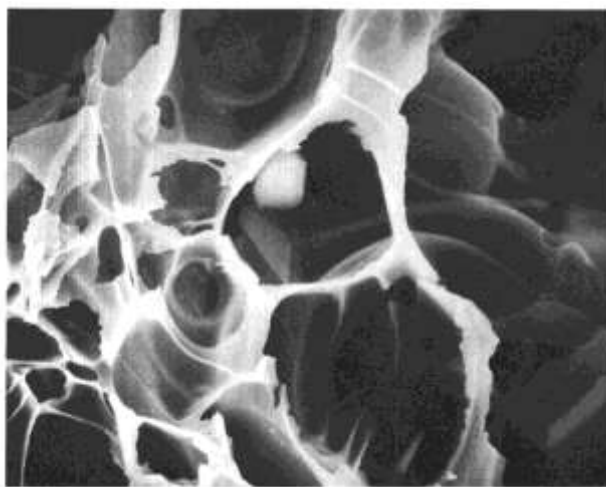FIGURE 1—CLASSIC ONIONSKIN PERLITE, MORROW RANCH AREA.

FIGURE 2—ELECTRON MICROGRAPH OF EXPANDED PERLITE (×2,000).

Origin

The origin of volcanic glasses has been the subject of considerable speculation and controversy, particularly with regard to the source of the water in perlite and pitchstone. Investigations beginning in the 1930's have shown that the combined water in obsidian differs significantly from that in perlite. The few tenths of a percent of water and other volatiles in obsidian were confirmed to be pristine magmatic components, whereas the several percent of water in coexisting perlite appeared to be of post-emplacement origin. Deuterium-hydrogen compositions of water in obsidians differ from that of coexisting perlites, the perlitic water more nearly approximating the composition of local ground waters. These relationships favor origin of perlite by hydration of obsidian with meteoric water and flushing of accessory volatiles. Further support for an obsidian-hydration origin of perlite is provided by perlites surrounding nodular obsidian nuclei (fig. 3), differences in water content and index of refraction between obsidian and perlite, lab tests of hydration rates at moderate pressures and temperatures, and the rates of development under atmospheric conditions of hydration rinds on obsidian artifacts of known age (Shepherd, 1938; Ross and Smith, 1955; Friedman and Smith, 1960; Friedman and others, 1966).

These physical and chemical relationships pertain primarily to classic or "onionskin" perlites intimately associated with coexisting obsidian. However, many texturally non-perlitic glasses of similar composition, which extend into pitchstones of even higher water contents, show field relationships contradictory to this type of supposed origin. Hydration of these glasses, if caused by water from outside the magma, probably occurred in shallow chambers prior to emplacement, or during emplacement in the upper to middle cooling temperatures. Introduction of water during either of these earlier stages would result in a marked decrease in viscosity and extension of the plastic stage of the melt to lower temperature levels.

Currently active perlite operations

Perlite deposits are widely distributed in the rhyolitic volcanic fields of the north-central and southwestern sections

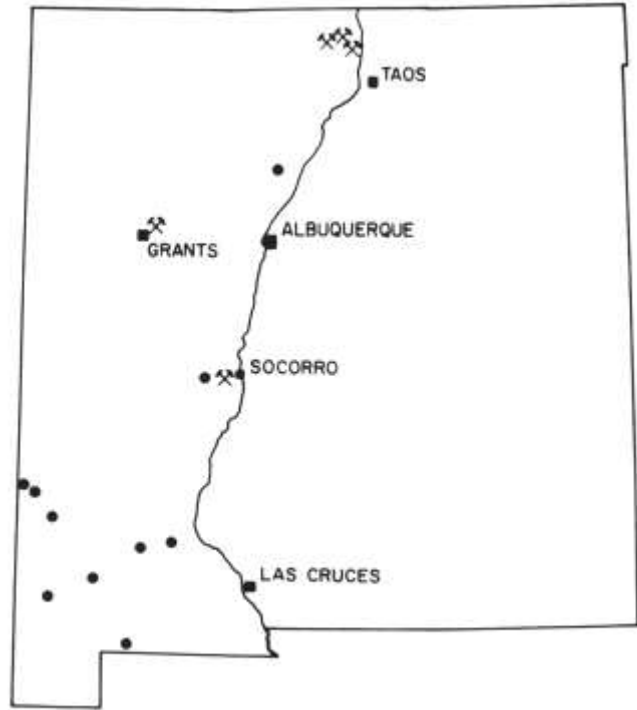FIGURE 3—MULE CREEK PERLITE WITH OBSIDIAN NODULES.

FIGURE 4—INDEX MAP OF NEW MEXICO PERLITE DEPOSITS.

of New Mexico (fig. 4). This paper will focus on five deposits that have yielded significant production of commercial perlite, with only brief mention of others that illustrate additional features of interest.

NO AGUA AREA

The perlite deposits of the No Agua area in extreme north-central New Mexico are the largest of any in New Mexico and were first recognized in 1948. The two largest mines are located on a group of four hills known collectively as No Agua Mountains or No Agua Peaks, which rise to a maximum height of 270 m above the surrounding plain at the western edge of the Taos Plateau. The No Agua area has been described in detail in an unpublished doctoral dissertation by Naert (1974) and also is the subject of a paper by David Whitson of Johns-Manville Corporation, elsewhere in this volume.

The mine on the north hill currently is operated by the Johns-Manville Corporation. The El Grande mine of Grefco (General Refractories Company) is located on the western flank of the No Agua Mountains. Volcanic rocks composing the No Agua Mountains have been included by Butler (1946) in the Cordito Member of Los Pinos Formation of middle Tertiary age, but Schilling (1960) inferred a later Pliocene-Pleistocene age. Lipman and others (1970) reported a fission-track age of 4.8 m.y. on obsidian from the Grefco mine. A potassium-argon age of 3.9 m.y. on obsidian was obtained later by Lipman and Mehnert (1979).

The four rhyolitic hills of No Agua overlie basalt flows and interflow sands and gravels that have been identified as the Jarita Basalt Member. Naert (1974) reported that a drill hole in the vicinity of No Auga penetrated about 80 ft of gravel, 380 ft of perlite, 180 ft of basalt, and 460 ft of the Esquibel Member of the Los Pinos Formation overlying highly crystalline, quartzitic basement rock.

GREFCO DEPOSIT—In Grefco's El Grande mine, extensive open-pit workings provide excellent exposures of a relatively uniform, pale-to-medium gray and brownish-gray, planar flow-banded, microvesicular and microperlitic volcanic glass. The rock typifies the so-called pumiceous perlites. Vesicular-

ity varies from slight to medium in various parts of the pit and marked differences in this property commonly are noteworthy between adjacent, thin flow bands. In general, the degree of vesicularity decreases with increasing depth, a probable result of increasing confining pressures downward.

The perlite is sufficiently fractured to be loosened with a ripper mounted on a tractor and then picked up by scrapers for transport to the mill (fig. 5). Naert (1974) estimates that the perlite flow is 25-100 m thick in the vicinity of the Grefco pit, and that the western flank of No Agua Mountain in the vicinity of the Grefco mine contains 100 million tons of commercial-grade perlite.

JOHNS-MANVILLE DEPOSIT—In the Johns-Manville mine (now the Manville Products Corp. mine) on the north hill, the upper compact perlite unit that underlies the crests of the hills is composed of flow-banded, brownish-gray, vesicular glass. Vesicles constitute 5-15% (by volume) of the rock and are accompanied by variable proportions of phenocrysts of sodic plagioclase, quartz, biotite, and opaque minerals. The thickness of this unit ranges here from 15 to 55 m. Gray perlitic glass containing abundant subangular obsidian nodules is exposed in road cuts below the working level of the pit and the southeast side of the hill. Earlier open-pit operations encountered numerous zones of obsidian-rich material in the perlitic-textured unit.

BRUSHY MOUNTAIN AREA

The Uniperl mine of Silbrico Corporation is located at Brushy Mountain, about 19 km east of No Agua Mountains. Brushy Mountain actually includes a cluster of several small, steep-sided, east-trending ridges and small knobs. The open-pit mine is situated on the southwestern flank of the southern-most ridge.

The Brushy Mountain perlite consists of vitrophyric perlite breccia ranging from 10 to 15 m in thickness in the mined area. Blocks ranging from very pale gray, highly vesicular perlite to medium-brownish-gray, dense perlitic glass reach diameters of up to several meters. All are highly porphyritic and some are distinctly flow banded. Phenocrysts consist of blocky sanidine and plagioclase crystals up to 4 mm in diameter, quartz, and sparse biotite. The matrix of the blocks is finely pulverized glass and clayey alteration products. The perlite breccia rests on a rhyolitic air-fall tuff, containing basalt and andesite clasts, that is overlain, in part, by up to 7 m of light-brown mudstone. The potassium-argon age of 22.3 m.y. obtained on sanidine from this deposit by Lipman and Mehnert (1979) is significantly older than the indicated age of the No Agua perlite. Mudstone in the western end of the pit is overlapped by the distal end of an andesite flow that probably issued from one of the nearby cones. Above the perlite, on the upper south slope of the ridge, is a sequence of devitrified and highly spherulitic rhyolite flows.

A separate segment of the perlite was exposed recently in exploratory cuts across a small valley 500 m south of the present pit. This location is at the northern foot of Cerro Montoso, one of a number of conspicuous basaltic cones that dot the surface of the Taos Plateau. The distal end of a vesicular basalt flow from Cerro Montoso directly overlies the weathered surface of the perlite.

EAST GRANTS RIDGE AREA

Perlite is intimately associated with obsidian and lithoidal rhyolite in a domal complex that forms much of East Grants Ridge near the southwestern foot of the composite cone of Mount Taylor. The rocks of the ridge, as described by Kerr and Wilcox (1963), consist of two distinct groups, an early rhyolitic assemblage and a later basaltic cone and flows. These volcanics are intrusive into and overlie exposed sedimentary rocks ranging from Triassic to Cretaceous in age.

The rhyolitic assemblage consists of a basal air-fall tuff and a cross-cutting domal complex of perlite, obsidian, and rhyolite. Within the dome, a core of lithophysal rhyolite is surrounded by a collar of obsidian and perlite that also caps much of the core. Nearly vertical flow banding in the center of the dome, and inward-dipping bands around the exposed base, indicate a bulbous domal form. Potassium-argon dating of obsidian and perlite by Bassett and others (1963a, b) yielded an apparent age of 3.3 m.y.

Perlite also occurs in satellitic masses that form a partial ring separated from the central dome by tuff. One of these masses on the eastern flank of East Grants Ridge forms a conical hill from which commercial perlite has been mined by U.S. Gypsum Company since 1953 (fig. 6). Evidently this body was extruded from a parasitic vent that probably is connected at depth with the central dome vent. The regularity of the flow banding in the central vent led Kerr and Wilcox (1963) to conclude that the perlite originated by direct intrusion rather than by hydration of obsidian.

Several intergradational varieties of perlite are exposed in the mine pit and adjacent road cuts and slopes. All varieties, however, are slightly to moderately porphyritic, with phenocrysts of albite, quartz, and orthoclase, plus traces of biotite, totalling about 12% by volume in a specimen of the flow-banded pumiceous phase in which vesicles constitute about 7%.

Internal structural features are particularly illustrative of the narrow limits between deformation by plastic flow and

FIGURE 5—GREFCO WORKING PIT, NO AGUA.

FIGURE 6—U.S. GYPSUM COMPANY PIT, GRANTS; columnar rhyolite forms cliffs in background.

brittle fracture during emplacement that characterizes siliceous volcanic glasses. This is particularly evident in the flow-banded, pumiceous type (fig. 7) in which transverse lenticular tension fractures, squeeze-outs, and breccia lenses formed simultaneously with elongation by laminar flow.

Perlite in the pit is ripped loose with power shovels and loaded onto trucks for transport to the mill at Grants where it is crushed, sized, and shipped out principally by rail. Bassett and others (1963a, b) obtained potassium-argon ages averaging 3.3 m.y. from several samples of obsidian collected west of the perlite mine. This age is believed to reasonably represent the age of the perlite.

SOCORRO-MAGDALENA AREA

Several perlite deposits are known near the northeast edge of the Datil-Mogollon volcanic field in the vicinity of Socorro and Magdalena (Weber, 1965). One of these, near Socorro, has the distinction of having been the principal domestic source of commercial perlite during the infant years of development of the perlite industry (Weber, 1963).

The Socorro perlite deposit is located on the southeastern flank of the Socorro Mountains, 5 km southwest of the town of Socorro. The deposit was worked from 1949 to 1961 by Great Lakes Carbon Corporation but then was inactive until 1975 when Grefco began construction of a new mill that is now in operation.

The deposit has the form and internal structure of a volcanic dome with an exposed vertical extent of more than 136 m and horizontal dimensions of 600 by 800 m. Vitric tuff and breccia of the same lithology overlap the north and south margins, whereas east and west margins are bounded by high-angle faults between which the dome is a horst block.

Most of the perlite body consists of gray to yellowish-gray, microvesicular, pumiceous perlite (fig. 8) with prominent, inward-dipping, concentric flow banding. Transverse tension fractures and lenses of auto brecciation are prevalent. Lithologically, the bulk of the deposit is very similar to the pumiceous phases of the No Agua and Grants deposits, consisting of about 93% glassy groundmass, 6% vesicles, and a little more than 1% quartz and oligoclase phenocrysts. Vesicles are flat and lenticular in shape, largely less than 1 mm in maximum length, and oriented parallel to the flow banding.
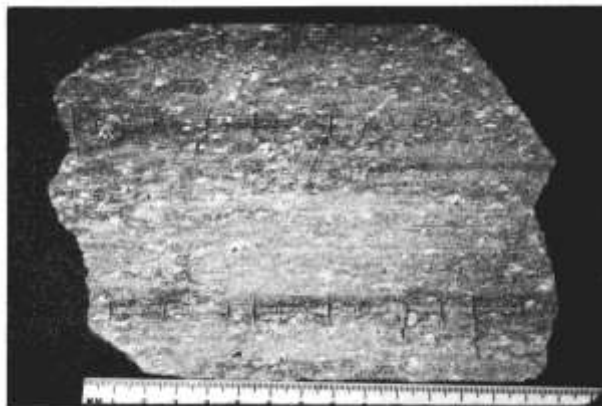

FIGURE 8—PUMICEOUS PERLITE, GREFCO DEPOSIT, SOCORRO.

The Socorro perlite is a late element in a series of domes of the Socorro Peak Rhyolite that were emplaced near the northeastern margin of the Socorro cauldron. A whole-rock potassium-argon age of 7.4 m.y. reported by Chamberlin (1980) probably more nearly approximates the true age of the deposit than the contradictory ages reported previously by Weber and Bassett (1963).

Mining is by a tractor-mounted ripper that loosens the highly fractured rock. The perlite is then pushed into an adjacent chute with grizzly and jaw crusher to reduce the oversized blocks. Broken rock then is fed onto a covered conveyer belt that transports it to a surge pile at the plant from which it is drawn as feed for the crushing-sizing circuit.

Perlite in southwest New Mexico

A number of perlite occurrences are widely distributed in the extensive volcanic region of southwest New Mexico (Weber, 1965). Several of these deposits have been intermittently exploited to a limited extent, but no sustained production has resulted. In contrast to the major, developed deposits in the northern half of the state, these perlites largely consist of the classical onionskin variety, only a few of which contain residual obsidian nodules. Water-rich glasses (> 6% H_2O) that have been designated as pitchstones and perlitic pitch-stones are noteworthy in some areas.

One occurrence of classic onionskin perlite is located near the Arizona border 9 km west of Mule Creek, immediately north of NM 78 in Grant County. Here, a rhyolitic dome complex contains local masses of perlite with abundant residual obsidian nodules (fig. 3). The perlite is interlain with flow-banded, spherulitic rhyolite capping several small peaks. A potassium-argon age of 18.6 m.y. was obtained for obsidian nodules from this deposit (Weber and Bassett, 1963; Weber, 1971).

A probably correlative rhyolitic sequence forms the main mass of the Mule Mountains, 18 km east of the previously mentioned deposit. On the southeast side of this small, isolated range, a 15-m tabular body of dark-gray, onionskin perlite crops out. Neither of these deposits offers much encouragement for commercial development.

A linearly extensive belt of deposits crops out in the vicinity of the Morrow and McDonald Ranches about 40 km south of Silver City, Grant County. The perlite there occurs in elements of a dome-flow complex that crop out at a number of places along a 10-km section of Burro Cienega. Hedlund (1978) inferred an Oligocene age for this complex.

Several small-scale efforts at commercial development have been made at various locations in the area, but all have been of short duration because of the diverse character of the perlite, much of which is of the classic onionskin type (fig. 1). Although locally of good quality, the perlite occurs in lenticular bodies of variable character and limited size. Previous mining

FIGURE 7—U.S. GYPSUM COMPANY PUMICEOUS PERLITE SHOWING TENSION FRACTURES, SQUEEZEOUT, AND BRECCIATION.

operations have not given adequate consideration to these factors, nor to the relatively high costs of mining and transportation. In the last operation, crude perlite was crushed at the pit, transported by truck to a railroad siding some 40 km south, and shipped to expansion plants in other states.

In the Leitendorf Hills, 12.5 km south-southwest of Lordsburg, Hidalgo County, perlitic pitchstone forms massive outcrops as much as 150 m high and extending over an area approximately 3.2 km long and up to 1.2 km wide. The overall form is suggestive of emplacement as a volcanic dome. Flege (1959) found evidence of intrusive emplacement as an endogenous dome or laccolith. The rock consists of reddish-brown to dark-green glass of resinous luster and fine-grained weak perlitic structure. Flow banding is locally prominent, and lithoidal clots and zones of alteration and devitrification are irregularly distributed. Chemical composition indicates a high-silica sodic rhyolite with a total water content (8.95%) in the pitchstone range (Weber, 1965). Although Flege estimated a volume of 30 million yds^3, the deposit has supported only short-lived, intermittent commercial operations during the past 30 yrs.

Conclusions

Several important conclusions can be drawn from a study of the geologic features of the productive perlite deposits of New Mexico:

1) The currently productive New Mexico perlites occur prevailingly as mushroom-shaped and bulbous domes and short flows overlying and adjoining rhyolitic volcanic vents.

2) Classic onionskin perlite is less characteristic of these deposits than the pumiceous or microvesicular variety.

3) Obsidian is relatively uncommon in these deposits, occurring abundantly only in parts of the Manville Products Corp. deposit.

4) Hydration prior to or during emplacement is indicated for at least major segments of these commerical deposits.

5) Zones of alteration and devitrification and the prevalence of lithoidal inclusions and flow bands seriously limit the commercial potentialities of many deposits.

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PERLITE IN EL SALVADOR, **CENTRAL AMERICA**

by W. Lorenz and **P.** Muller, *Federal Institute for Geosciences and Natural Resources,*
P.O. Box 510153, D-3000 Hannover, Federal Republic of Germany

Abstract

Perlites have been mapped in El Salvador. The most promising ones economically have been explored by drilling programs. The El Rosario deposit, Dept. Cuscatlán, is the best known and is described here in more detail as an example of perlite deposits in El Salvador. Drill holes and geological vertical sections through the deposits show the following zonation: Perlite on top (up to 30 m), spherulitic perlite beneath it (a few meters), grading into felsitic rhyodacite.

Expanded perlites exhibit bulk densities of 110 g/L to 240 g/L with an average of 157 g/L. They have been successfully tested for use in lightweight concrete. Chemical-analysis data compare well with those of other perlites worldwide. Thin sections of perlites contain phenocrysts of plagioclase, amphibole, hypersthene, and ore (altogether 15-20 volume percent) imbedded in a glassy (perlitic) matrix. Towards the base of the perlites, feldspathic macrospherulites (diameter 1-2 mm) can be observed, becoming more frequent with depth; at the same time, their diameters become greater. A parallel increase in the content of cristobalite, revealed by x-ray analysis, indicates that it is an integral part of the macrospherulites. Cristobalite is considered to be the high-temperature poly-morph of quartz, which therefore must have formed during an early stage of the cooling of the rhyodacite dome. On the other hand, there are smaller (younger) microspherulites (diameter about 0.1 mm) in the bottom part of the perlite deposit consisting of opal-CT (lussatite), the low-temperature polymorph of quartz.

Based on these observations, the following theory for the genesis of perlites in El Salvador is profered:
1) A rhyodacite dome rises to the surface of the earth, crystals forming deep inside its core;
2) Chilling forms an outer hyaline zone that alters to perlite by influx of meteoric (and juvenile?) water;
3) More or less simultaneous formation of macrospherulites occurred during an early stage of devitrification;
4) With time, glassy perlite devitrified to felsitic rhyodacite at depth;
5) During a later stage of devitrification, microspherulites were formed.

The stages mentioned overlap to a greater or lesser degree.

Because of the continuous devitrification of perlites, it must be assumed that most of the (felsitic) rhyodacite was formed from perlite.

Introduction

Within the framework of Salvadorian-German technical cooperation, numerous occurrences of nonmetallic mineral raw materials including perlite were geologically investigated in El Salvador from 1975 to 1977 (table 1). In petrography, "perlite" is defined as an acidic, glassy, volcanic rock with a perlitic texture, that is, small, semi-spherical cracks along which the rock disintegrates into small (mm-size) spherules. The technical term "perlite," however, includes all glassy volcanic rocks that contain water and expand to many times their original volume when heated. Both raw and expanded perlite are traded simply as "perlite." Perlite occurs all over the world in young, acidic volcanic rocks. In El Salvador, the occurrence of acidic volcanics is restricted chiefly to the Cuscatlan Formation (c2). All known perlite occurrences in El Salvador are shown in fig. 1. The El Rosario perlite deposit (no. 12, fig. l), Dept. Cuscatlán, was investigated in detail and is representative of other perlite occurrences in El Salvador.

El Rosario perlite deposit

The El Rosario perlite deposit was mapped at a scale of 1:10,000 (fig. 2). Detailed mapping revealed the complex structure of this deposit with its complicated distribution of perlite, spherulitic (that is, partly devitrified) perlite and felsitic rhyodacite.

On the basis of the detailed geological map, a core-drilling program was carried out to investigate the underground conditions. Because the deposit has a very complicated structure, a total of 35 core boreholes (fig. 2) with a total length of 954 m (3,130 ft) were sunk. The data obtained from the drillholes made it possible to construct an isopach map (fig. 3) that shows the spatial distribution of perlite thicknesses within the deposit. Areas of great thickness suitable for economic recovery are located near the western and eastern edges of the deposit. In general, the greatest thicknesses occur within the center of a perlite area.

A geological section (fig. 4) through the deposit showed a definite zoning from top to bottom. The perlite (the outer skin of a lava dome) is underlain by spherulitic perlite grading indistinctly into felsitic rhyodacite. This pattern was a valuable aid in the exploration of perlite occurrences in El Salvador. Moreover, the easily erodable perlite is preserved only as a comparatively thin cap (up to 30 m or 98 ft thick) on thick, felsitic rhyodacite that has not yet been drilled.

Laboratory tests

A number of laboratory tests were carried out in the Federal Institute for Geosciences and Natural Resources. According to chemical analysis, the composition of the Salvadorian perlite is comparable to perlites from other parts of the world (fig. 5). Minor variations occur in SiO_2, Fe_2O_3, and K20 content. The mappable vertical zoning that can be recognized macroscopically was studied for its mineralogical structure using X-ray-diffraction methods (fig. 6). These studies, together with those of thin sections, demonstrated that a phenocryst generation of plagioclase, amphibole, hypersthene, and ore (approximately 15-20% by volume) is embedded in a glass matrix. With increasing depth, the glass matrix is interspersed with an increasing number of brownish, feldspathic macrospherulites (fig. 7), whose diameter (approximately 1-2 mm

TABLE 1—LIST OF KNOWN PERLITE DEPOSITS IN EL SALVADOR.

No.	Name of deposit	Map sheet (year of printing)	Coordinates
1	Cerro Afate	Sta. Ana (1954)	438 500/302 700
2	Cerro Pacho	Sta. Ana (1954)	437 500/302 500
3	Cerro El Cobano	Nueva Concepcion (1973)	449 150/331 600
4	San Ignacio	San Ignacio (1967)	477 300/356 800
5	Ciudad Delgado	San Salvador (1975)	481 350/288 850
6	Venecia	San Salvador (1975)	483 250/288 850
7	Cerro El Paraiso	San Salvador (1975)	498 300/294 100
8	Cerro Quemado	San Salvador (1975)	499 300/294 300
9	Cerro El Carmen	San Salvador (1975)	500 000/290 000
10	Cerro Buena Vista	San Salvador (1975)	499 400/288 500
11	Cerro Chachacaste	Cojutepeque (1959)	507 500/292 800
12	El Rosario	Cojutepeque (1959)	510 000/294 300
13	El Carmen	Cojutepeque (1959)	510 650/289 700
14	El Cimarron	Rio Titihuapa (1962)	531 650/296 850
15	Agua Zarca	Jocoro (1956)	587 200/268 750
16	Yamabal	Jocoro (1956)	588 750/282 750
17	San Francisco Gotera	San Francisco Gotera (1962)	597 900/283 500
18	El Triunfo	San Francisco Gotera (1962)	598 350/283 600
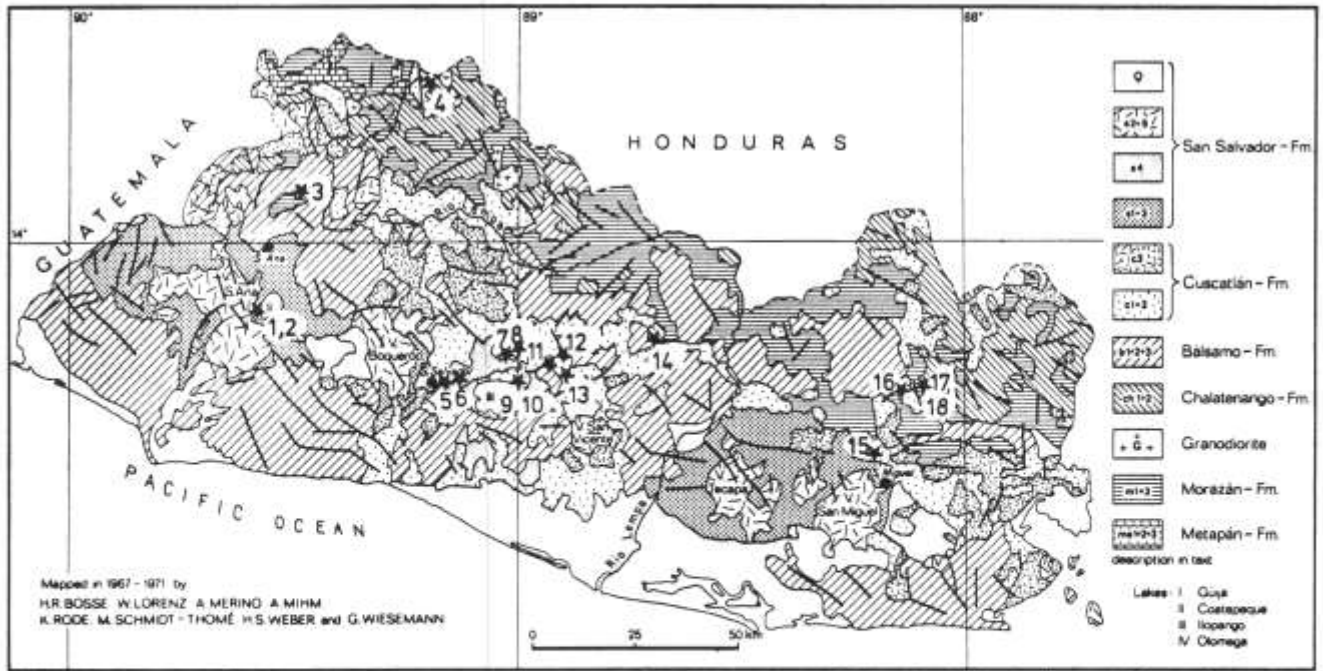
FIGURE 1—PERLITE DEPOSITS IN EL SALVADOR (black dots). Base map: Mapa geologico general de la República de El Salvador, scale 1:500,000, published in 1974 by the Federal Institute for Geosciences and Natural Resources, Germany, in collaboration with the Centro de Estudios e Investigaciones Geotécnicas of El Salvador.
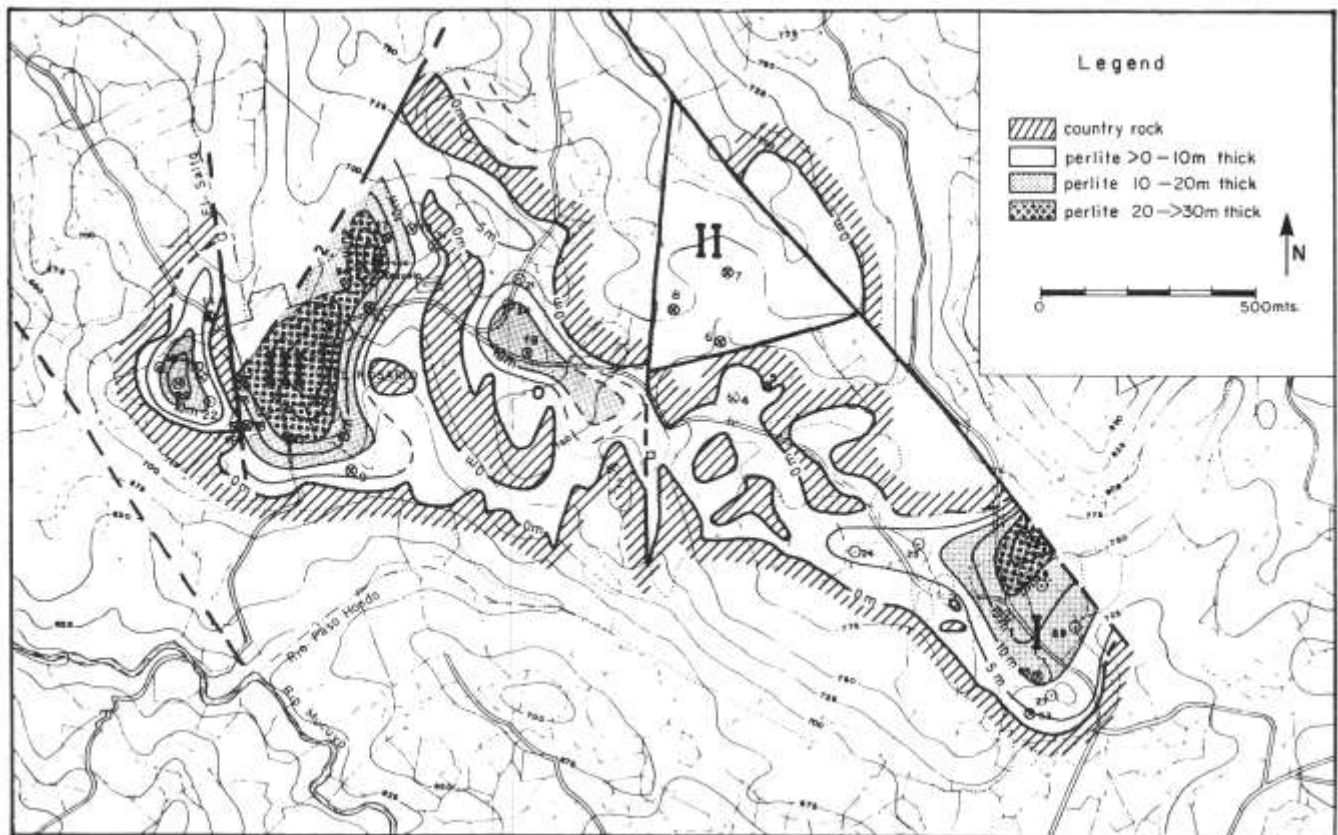

FIGURE 2—GEOLOGICAL MAP (SCALE 1:10,000) OF EL ROSARIO PERLITE DEPOSIT, DEPT. CUSCATLÁN, EL SALVADOR.
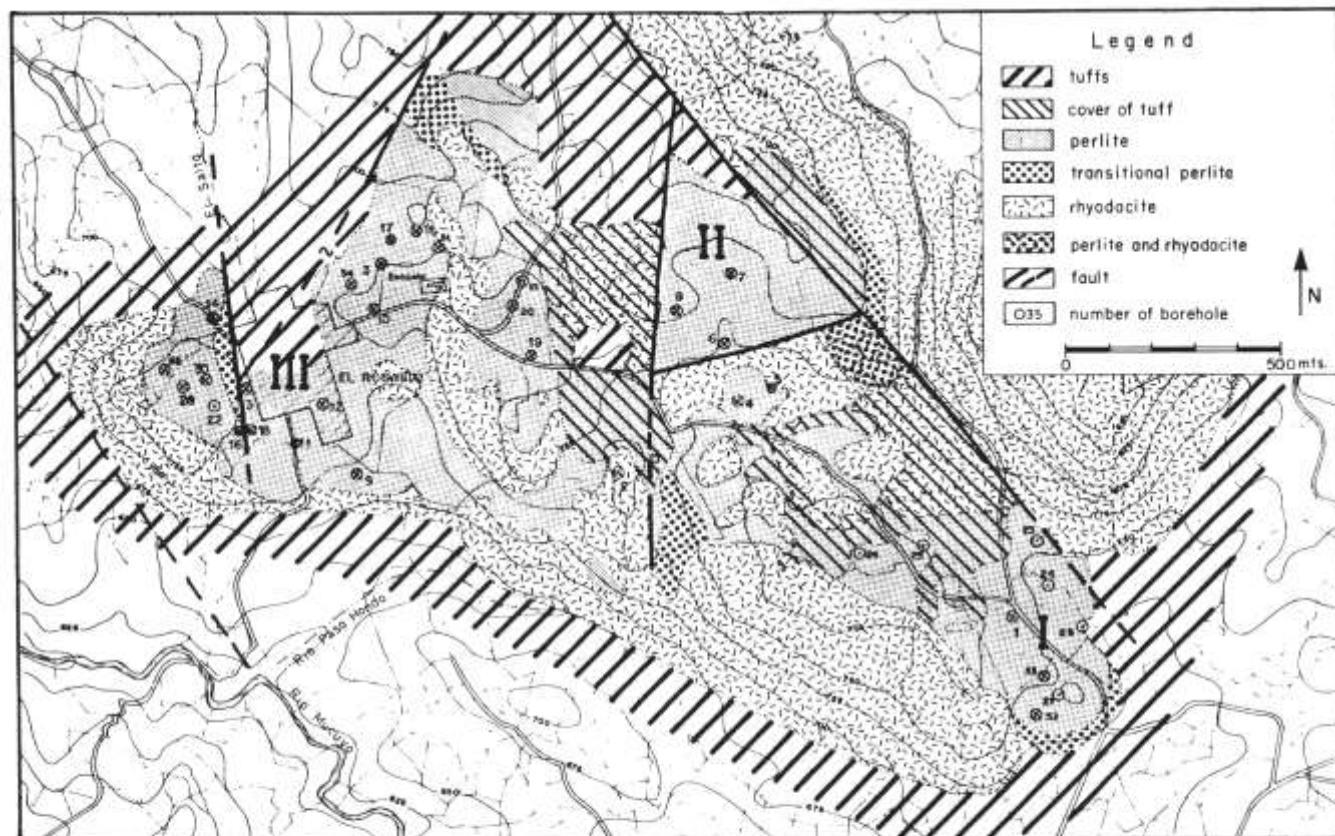
FIGURE 3—ISOPACH MAP OF PERLITE BED IN EL ROSARIO DEPOSIT. Greatest thicknesses (more than 25 m) occur near the western and eastern edges of deposit.
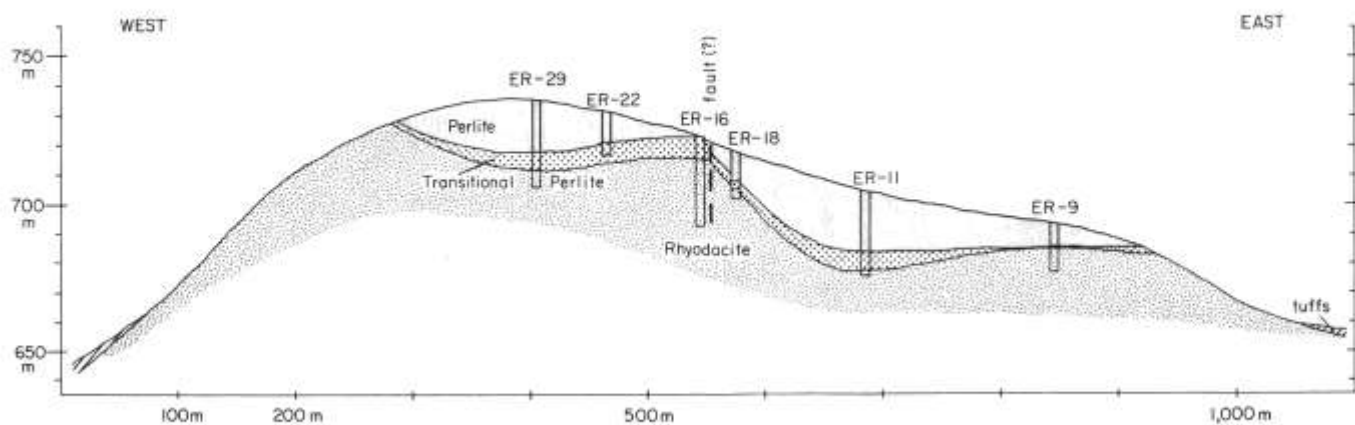

FIGURE 4—GEOLOGIC CROSS SECTION (EAST-WEST) THROUGH EL ROSARIO PERLITE DEPOSIT. The following zonation from top to bottom is clearly visible: perlite, spherulitic perlite, thick felsitic rhyodacite (vertical exaggeration × 2.5); ER-1 = drill hole.

[New Mexico Bureau Mines & Mineral Resources, Circular 182, 1982]

Comparison of chemical analyses		
	perlites, worldwide[x]	perlite, El Rosario
SiO_2	71.0 – 75.0	70.9
Al_2O_3	12.5 – 18.0	13.7
Fe_2O_3	0.5 – 1.5	1.9
FeO	0.0 – 0.1	–
CaO	0.5 – 2.0	1.9
MgO	0.1 – 0.5	0.5
Na_2O	2.9 – 4.0	3.6
K_2O	4.0 – 5.0	3.0
H_2O,comb.	3.0 – 5.0	3–4

[x] source: Perlite Institute, Inc., New York

FIGURE 5—COMPARISON OF THE CHEMICAL COMPOSITION OF 15 PERLITES FROM EL ROSARIO WITH PERLITES FROM OTHER PARTS OF THE WORLD; source: Perlite Institute Inc., New York.
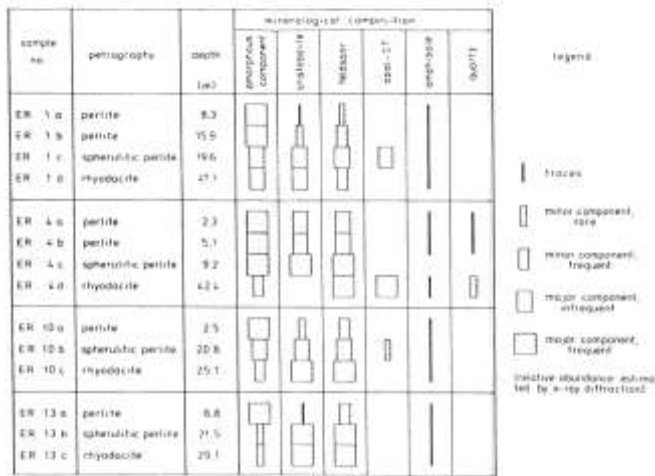

FIGURE 6—MINERALOGICAL COMPOSITION OF PERLITE, SPERULITIC PERLITE, AND RHYODACITE IN 4 BOREHOLES IN EL ROSARIO DEPOSIT. Amorphous components (that is, perlitic glass) decrease with increasing depth, whereas feldspar and cristobalite contents increase with increasing depth. The occurrence of cristobalite is confined to the bottom of the perlite, especially to the spherulitic perlite.
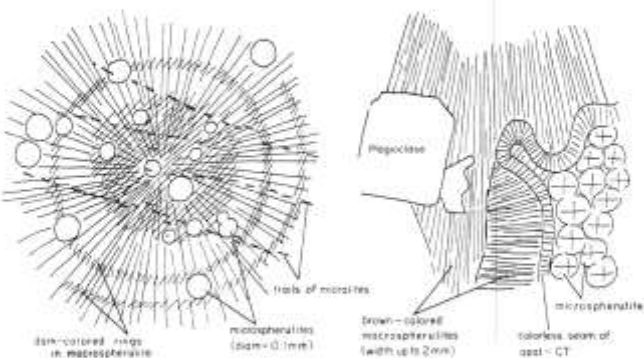

FIGURE 7—LEFT: ALKALI FELDSPAR-QUARTZ (CRISTOBALITE) MACROSPHERULITE (DIAMETER APPROXIMATELY 2 MM) WITH DARK CONCENTRIC RINGS AND COLORLESS, TRANSPARENT MICROSPHERULITES (DIAMETER APPROXIMATELY 0.1 MM) OF OPAL-CT (LUSSATITE), BOTH CROSSED BY SUBPARALLEL MICROLITE TRAILS. RIGHT: MACROSPHERULITES WITH A PLAGIOCLASE CORE AND INTERNAL RADIAL FIBROUS STRUCTURE, SURROUNDED BY A RIM OF WATER-WHITE OPAL-CT, direction of fibres in the rim is perpendicular to the surface of the macrospherulite; aggregations of microspherulites consist of opal-CT (sample ER 1d); drawings and evaluation of thin sections by P. Müller.

or 0.04-0.08 inches) increases with depth. Approaching the spherulitic perlite, the percentage of cristobalite also increases with depth. This increase implies that the macrospherulites consist of feldspar and cristobalite, the high-temperature polymorph of quartz (SiO_2); accordingly, the macrospherulites must have formed during a rather early stage of the cooling of the rhyodacite dome. On the other hand, very small, younger, and colorless microspherulites (diameter approximately 0.1 mm or 0.004 inches) consisting of opal-CT (fig. 6) are also encountered in the deeper parts of the perlite deposit; opal-CT is regarded as a low-temperature polymorph of quartz (SiO_2).

Formation of Salvadorian perlites

On the basis of the above-mentioned field and laboratory studies, the following ideas have been developed on the formation of the Salvadorian perlites (fig. 8). Acid magma rose to the earth's surface and crystallized in its innermost parts, whereas the outer chilling zone turned into a glassy mass that was then transformed into perlite by the uptake of meteoric and juvenile (?) water. More or less simultaneously, while the temperatures were still high, macrospherulites formed. With the passage of time, the perlite body was increasingly devitrified towards depth and was transformed into felsitic rhyodacite. During the late stage of devitrification, microspherulites of opal-CT finally formed at low temperatures. Because the devitrification of the perlitic rock bodies progressed steadily, it must be assumed that most of the (felsitic) rhyodacites were formed from perlite.

Ideas on the formation of perlite as outlined in this paper are valid, for the time being, only for the El Rosario type and for most of the perlites occurring in El Salvador. Indications are that comparable deposits may also occur in other countries.

Potential uses

Numerous perlite samples were tested for their expansion properties (fig. 9). The bulk density of the expanded perlite is between 120 and 240 g/L (7.5-15 lb/ft^3), whereas the bulk density of the raw perlite is approximately 1,300 g/L (81 lb/ft^3). The average bulk density of 65 samples of expanded perlite from El Rosario was 157 g/L (kg/m' or 2.8 lb/ft^3). Apparently, expansion properties do not depend on depth. The possibilities for using Salvadorian perlite depend primarily on the bulk density. Approximately 85-90% of the perlite traded in the world is used for construction purposes (for insulation, as fire-resistant plaster, ultralight concrete, etc.). Salvadorian perlite fully meets the requirements of the ASTM (American Standard for Testing and Materials). Perlite can also be used as a filter aid, although the quality of Salvadorian perlite is possibly not sufficient for this purpose; additional uses are found for perlite in horticulture and as carrier material for insecticides.

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Development of present erosion level. Near surface alteration produced by weathering

Temperature, time

Growth of microspherulites

50°C

Crystallization of felsitic rhyodacite

Formation of perlite by influx of meteoric and (?)juvenile water

Devitrification — growth of spherulites

300–500°C

Rise of rhyodacitic lava dome. Chilling produces hyaline outer zone. Crystallisation of inner core

900°C

Legend

microspherulites (opal-CT)

felsitic rhyodacite

perlite with spherulites (of cristobalite and feldspar)

perlite

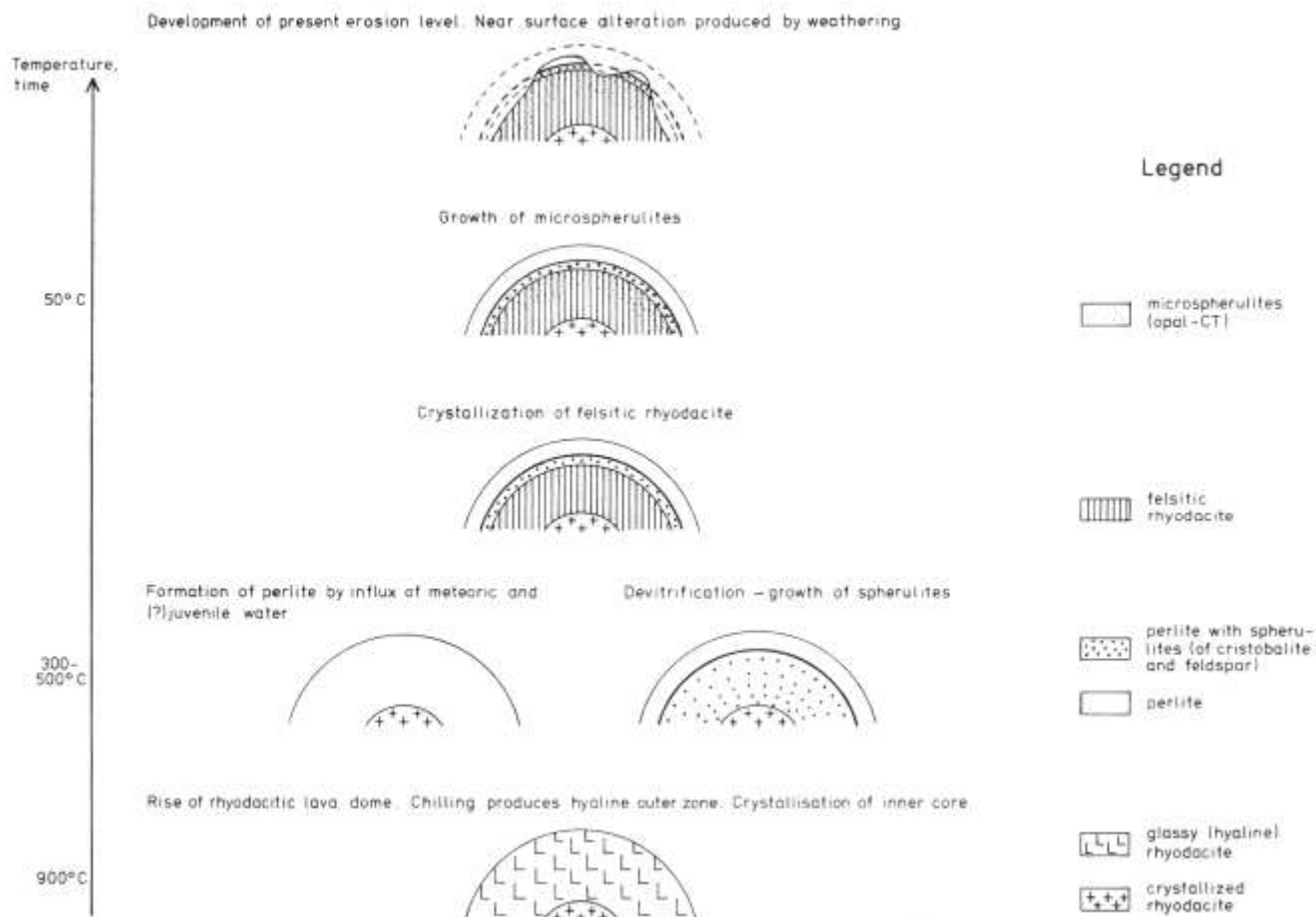glassy (hyaline) rhyodacite

crystallized rhyodacite

FIGURE 8—THEORY OF FORMATION OF PERLITES IN EL SALVADOR. Temperatures given on the vertical axis must be regarded as a basis for discussion. The various phases of formation of the perlite, shown here as separate units, overlap in reality to greater or lesser degrees.

g/l (=kg/m³)

40   perlite for cryogenic use

80   perlite for filteraids

120

160   perlite for construction

200

240

perlite of inferior quality

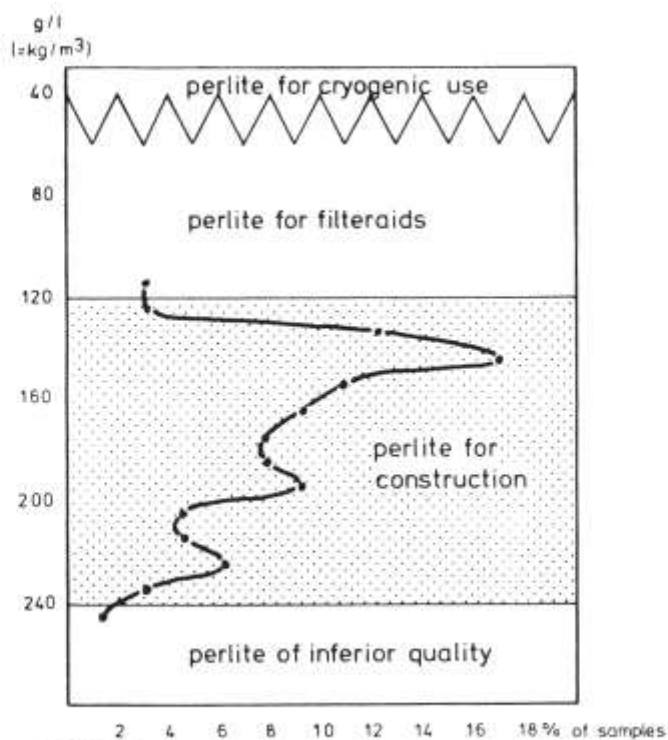2 4 6 8 10 12 14 16 18 % of samples

FIGURE 9—BULK DENSITY AND POSSIBILITIES FOR USING EXPANDED PERLITES FROM EL ROSARIO DEPOSIT. The perlites can be generally regarded as very good for use in the construction industry; data from 65 samples from the El Rosario deposit.

GEOLOGY OF THE ASH MEADOWS CLINOPTILOLITE DEPOSIT, INYO COUNTY, CALIFORNIA, AND NYE COUNTY, NEVADA

by K. N. Santini and A. R. Shapiro, *Anaconda Copper Company, Denver, Colorado 80225*

Abstract

The Ash Meadows clinoptilolite deposit crosses the California-Nevada border in the southern part of the Basin and Range province. It is located approximately 70 mi northwest of Las Vegas, Nevada, and about 5 mi east of Death Valley Junction, California. The deposit is a Tertiary vitric ash-flow tuff that has been diagenetically altered by alkaline ground water to clinoptilolite over a strike length of approximately 9.6 km (6 mi). The northerly striking and easterly dipping horizon has a stratigraphic thickness up to 46 m (150 ft) in California, and 122 m (400 ft) in Nevada. Sample analyses indicate that the deposit contains large tonnages of 90% clinoptilolite with excellent ammonia adsorption and attrition characteristics.

BARITE BENEFICIATION   MODERN MILL DESIGN

by Thomas J. Plouf, *Continental Oil Company, Denver, Colorado 80225*

Abstract

Barite use is increasing in proportions greater than that of any other industrial mineral. The availability of high-grade deposits, from which its extraction is a simple gravity upgrading, is decreasing at an even more rapid rate. Four general methods of barite beneficiation are reviewed with specific emphasis given to barite flotation. The general design parameters for each section of a barite flotation mill are reviewed and discussed. A method of capital investment costing and operating costing to obtain the optimum mill size for a given ore reserve is discussed. A case study of the application of this method to a mill design project is presented to help geologists understand the economics of the beneficiation of a barite orebody.

Appendix

REGISTRANTS AT 1981 FORUM ON GEOLOGY OF INDUSTRIAL MINERALS.

Allen, Hank
Martin Marietta Cement
6801 Rockledge Drive
Bethesda, MD 20034

Anderson, Cindy
Duval Corporation
4715 E. Ft. Lowell
Tucson, AZ 85712

Arndt, Robert H.
Oklahoma Geological Survey
830 Van Vleet Oval
Norman, OK 73019

Asher-Bolinder, Sigrid
U.S. Geological Survey
M.S. 950, Federal Center
Denver, CO 80225

Ault, Curtis H.
Indiana Geological Survey
611 N. Walnut Grove
Bloomington, IN 47405

Austin, George S.
New Mexico Bureau of Mines and Mineral
   Resources
Campus Station
Socorro, NM 87801

Bates, Robert L.
Ohio State University
125 South Oval Mall
Columbus, OH 43210

Beckman, Richard J.
U.S. Gypsum
422-1 W. Russell
Barrington, IL 60010

Bell, Rex R.
Foote Mineral Company
Route 100
Exton, PA 19341

Bentzen, Edwin H.
C.S.M. Research Institute
P.O. Box 112
Golden, CO 80401

Berendsen, Pieter
Kansas Geological Survey
University of Kansas
1930 Avenue "A"
Lawrence, KS 66050

Boardman, Ray C.
Reynolds Metals Company
6601 W. Board Street
Richmond, VA 23261

Blazoric, P.
International Minerals & Chemical Corp.
666 Garland Place
Des Plaines, IL 60016

Bourbon, W. B.
Consulting Geologist
Box 368
Alpine, TX 79830

Bourne, H. Lyn
Consulting Geologist
P.O. Box 5322
Northville, MI 48167

Breese, Richard
Johns-Manville Sales Corporation
Mailstop 2-13, J-M Deer Creek Canyon Rd.
Littleton, CO 80217

Brown, Howard
Pluess-Staufer Calif. Inc.
P.O. Box 825
Lucerne Valley, CA 92356

Burchett, Raymond R.
Nebraska Geological Survey
901 N. 17th, Nebraska Hall
Lincoln, NB 68588

Burgin, Lorraine B. U.S. Bureau
of Mines Intermountain Field
Operations Center State
Services Section, Building 20
Denver Federal Center
Denver, CO 80215

Carmichael, Virgil W.
North American Coal Corporation
1013 Anderson Street
Bismarck, ND 58501

Cathcart, Robert S.
Medusa Corporation
P.O. Box 5668
Cleveland, OH 44101

Ciesco, Richard
Marblehead Lime Co.
300 W. Washington
Chicago, IL 60606

Clifton, Robert A. U.S.
Bureau of Mines 2401 E
Street, N.W.
Washington, DC 20241

Cohen, Bernard
FMC Corporation
P.O. Box 8
Princeton, NJ 08540

Conner, John M.
Kings Mountain Mica Company
Grover Road
Kings Mountain, NC 28086

Cutcliffe, William E.
Dunn Geoscience Corporation
5 Northway Lane North
Latham, NY 12110

Daugherty, Frank W.
D & F Minerals, Inc.
P.O. Box 329
Alpine, TX 79830

Day, Dennis L.
NERCO, Inc.
111 Columbia S.W., Station 800
Portland, OR 97201

Deal, Edmond
Duval Corporation
4715 E. Ft. Lowell
Tucson, AZ 85712

Dickson, Ted
Metal Bulletin Inc.
708 3rd Ave., 23rd Floor
New York, NY 10017

Donegan, Ben
Leonard Minerals Company
3202 Candelaria N.E.
Albuquerque, NM 87107

Elston, Wolf
University of New Mexico
Department of Geology
Albuquerque, NM 87131

Eveleth, Robert
New Mexico Bureau of Mines and Mineral
   Resources
Campus Station
Socorro, NM 87801

Eyde, Ted H.
GSA Resources, Inc.
Box 1127
Cortaro, AZ 85230

Feldman, Sandra C.
Consulting Geologist 5934
Pauline N.W.
Albuquerque, NM 87107

Fullton, Jim
Radcliff Materials, Inc.
P.O. Box 2008
Mobile, AL 36602

Geitgey, Ron
Duval Corporation
4715 E. Ft. Lowell
Tucson, AZ 85712

Golson, Glenn S.
A. P. Green Refractories
Green Blvd.
Mexico, MO 65265

Goudarzi, Gus H.
U.S. Geological Survey
658 Pemberton Ct.
Herndon, VA 22070

Grisafe, David A.
Kansas Geological Survey
University of Kansas
Lawrence, KS 66044

Griswold, George B.
Tecolote Corporation
531 Wagontrain Drive S.E.
Albuquerque, NM 87123

Gunning, Donald F.
International Marble & Stone Co.
10691 Shellbridge Way, #190
Richmond, B.C. V6X 2W8

Harben, Peter
Metal Bulletin Inc.
708 3rd Ave., 23rd Floor
New York, NY 10017

Haumesser, Andrew F.
American Aggregates Corporation
P.O. Box 160
Greenville, OH 45331

Hoagberg, Rudolph K.
E. K. Lehmann & Associates, Inc.
1409 Willow Street
Minneapolis, MN 55103

Holmes, L. A.
International Minerals & Chemical Corp.
666 Garland Place
Des Plaines, IL 60016

Hopkins, David A.
J. E. Baker Company
Box 1189
York, PA 17405

Hora, Z. D.
B.C. Ministry of Energy, Mines, &
    Petroleum Resources
Parliament Buildings
Victoria, BC V8V 1X4

Huber, Gary C.
Canyon Resources Corporation
2207 Jackson Street, Suite B-101
Golden, CO 80401

Ives, Robert E.
Michigan Geological Survey
Box 30028
Lansing, MI 48909

Jenkins, David
Grefco Inc.
621 N. 5th Street
Lompoc, CA 93436

Jensen, Robert E.
FMC Corporation
P.O. Box 8
Princeton, NJ 08540

Johnson, Kenneth S.
Oklahoma Geological Survey
830 Van Vleet Oval
Norman, OK 73019

Jones, Charles L.
Kailum Chemicals
600 S. Cherry Street, Suite 1120
Denver, CO 80222

Jorgensen, Duane B.
U.S. Gypsum Company
RR 3 Box 434
Abingdon, VA 24210

Keith, Barry L.
NERCO, Inc.
111 S.W. Columbia, Suite 600
Portland, OR 97201

Kelley, Robert W.
New Mexico Bureau of Mines and Mineral
    Resources
Campus Station
Socorro, NM 87801

Kottlowski, Frank E.
New Mexico Bureau of Mines and Mineral
    Resources
Campus Station
Socorro, NM 87801

Knostman, Richard W.
Anaconda Copper Company
555 17th Street
Denver, CO 80217

Krier, Ned
Duval Corporation
4715 E. Ft. Lowell
Tucson, AZ 85712

Krueger, Al
Alpha Aggregates
P.O. Box 217
Martins Creek, PA 18063

Kuchenbuch, Pam
Duval Corporation
4715 E. Ft. Lowell
Tucson, AZ 85712

Kuff, Karen R.
Maryland Geological Survey
The Johns Hopkins University
Merryman Hall
Baltimore, MD 21218

Larsen, Veryl
Colorado School of Mines Research
    Institute
Box 112
Golden, CO 80401

Lewis, William L.
United States Gypsum Company
19 Green Acres, Site 1, Box 21
Kingston, Nova Scotia BOP 1R0

Lobdell, David G. Taiga
Consultants 1300 8th Street,
S.W., Station 100 Calgary,
Alberta T2R 1B2

Logan, William J.
Ontario Ministry of Natural Resources
P.O. Box 9000
Huntsville, Ontario POA 1KO
Canada

Logsdon, Mark J.
New Mexico Bureau of Mines and Mineral
    Resources
Campus Station
Socorro, NM 87801

McKinley, James
University of New Mexico
Albuquerque, NM 87131

McLemore, Virginia
New Mexico Bureau of Mines and Mineral
    Resources
Campus Station
Socorro, NM 87801

Mallory, William M.
Anglo-American Clays Corporation
P.O. Box 471
Sandersville, GA 31082

Malloy, Robert W.
1MCO Services, M & E Department
P.O. Box 861
Battle Mountain, NV 89820

Malone, Virgil
Medusa Corporation
P.O. Box 5668
Cleveland, OH 44101

Manson, G. Robert
Mountain Minerals Co.
P.O. Box 700, 529 6th Street
Lethbridge, Alberta T1J 1L2
Canada

Martin, James A.
Missouri Geological Survey
Box 250
Rolla, MO 65401

Massa, Philip
Johns-Manville
1826 Grand Ave.
Billings, MT 59102

Matthews, Sharon E.
U.S. Geological Survey
3001 Portland Ridge
Norcross, GA 30071

Meadows, John R.
P.O. Box 25626
Albuquerque, NM 87125

Menzie, David
New Mexico Bureau of Mines and Mineral
    Resources
Campus Station
Socorro, NM 87801

Minnes, D. G.
Ministry of Natural Resources
Room 4648, Whitney Block
99 Wesley Street West
Toronto, Ontario M7A 1W3
Canada

Montgomery, Edward
Duval Corporation
4715 E. Ft. Lowell
Tucson, AZ 85712

Oak, Leo C.
Martin Marietta Aggregates
Central Division
P.O. Box 50815
Indianapolis, IN 46250

Olson, Richard H.
Industrial Minerals Evaluations, Inc.
14618 W. 6th Avenue, Suite #202
Golden, CO 80401

Orbock, Edward
Billiton Metals & Ores, Inc.
711 Hazelpine S.E.
Albuquerque, NM 87102

Osburn, JoAnne C.
New Mexico Bureau of Mines and Mineral
    Resources
Campus Station
Socorro, NM 87801

Oxford, Edward F.
J. M. Huber Corporation
Route 4-Huber
Macon, GA 31298

Papke, Keith G.
Nevada Bureau of Mines and
Geology University of Nevada Reno,
NV 89557

Parduhn, Philip
VePed Traffic Controls, Inc.
11313 N. Bdwy
Oklahoma City, OK 73114

Peirce, H. Wesley
Bureau of Geology and Min. Tech.
845 N. Park Ave.
Tucson, AZ 85719

Peterson, Stephen
Leonard Minerals Co.
3202 Candelaria N.E.
Albuquerque, NM 87107

Pickering, Sam M.
Georgia Kaolin Co.
Box 28
Sandersville, GA 31082

Plouf, Thomas M.
Continental Oil Company
555 17th Street
Denver, CO 80202

Power, W. Robert
Department of Geology
Georgia State University
Atlanta, GA 30034

Rempe, Norbert T.
IMC Corporation
P.O. Box 71
Carlsbad, NM 88220

Richards, Donald J.
Michigan DNR-Ref.
5145 Ballantine Rd.
Bath, MI 48808

Root, Robert L.
Consultant
3379 Aztec Road
Doraville, GA 30340

Rowbottom, T. E.
Domtar, Inc.
1136 Matheson Blvd.
Mississauga, Ontario
Canada

Runvik, R. C.
U.S. Gypsum Company
101 South Wacker Drive
Chicago, 1L 60606

Santini, Kenneth N.
Anaconda Copper Company
555 17th Street
Denver, CO 80217

Sarles, John A.
Johns-Manville Corporation
P.O. Box 338
Antonito, CO 81120

Schwab, Geraldine
W. K. Summers & Associates, Inc.
P.O. Box 684
Socorro, NM 87801

Schwochow, Stephen
Colorado Geological Survey
1313 Sherman Street, Room 715
Denver, CO 80203

Shapiro, Alan
Anaconda Copper Company
555 17th Street
Denver, CO 80217

Sharpe, Roger D.
U.S. Gypsum Company
101 S. Wacker Drive
Chicago, 1L 60606

Seibert, Walter E., Jr.
907 Country Club Drive
Teaneck, NJ 07666

Siemers, William Terry
Phillips Petroleum Company
116 TRW
Bartlesville, OK 74004

Smith, Edward W.
Eight Northern Indian Pueblos Council
Box 969
San Juan Pueblo, NM 87566

Sorensen, Harry
Michigan Geological Survey
Dept. of Natural Resources
Box 30028
Lansing, MI 48909

Speczik, Stanislow
Kansas Geological Survey
University of Kansas
1950 Avenue "A"
Lawrence, KS 66050

Stewart, Charles T.
Malvern Minerals Co.
P.O. Box 1246
Hot Springs, AR 71901

Stonehouse, D. H.
Energy, Mines, & Resources
580 Booth Street
Ottawa, Ontario K1A
OET Canada

Strong, Bob
Sunbelt Mining Company, Inc.
P.O. Box 2106
Albuquerque, NM 87103

Sweet, Palmer
Virginia Division of Mineral Resources
P.O. Box 3667
Charlottesville, VA 22901

Timmons, Bobby J.
Florida Rock Industries, Inc.
P.O. Box 2319
Jacksonville, FL 32202

Turner, James A.
U.S. Bureau of Land Management
7009 Pioneer Place, N.W.
Albuquerque, NM 87120

Via, Edwin King
Lone Star Ind. Inc.
P.O. Box 2880
Greenwich, CT 06830

Walish, Robert C.
Chromalloy Mining & Milling
P.O. Box 429
Elko, NV 89802

Weber, Robert H.
New Mexico Bureau of Mines and Mineral
  Resources
Campus Station
Socorro, NM 87801

Weeks, Donald R.
Romaine Construction Co.
8819 Fairfield Greens Drive
Midwest City, OK 73110

Wells, Robert L.
California Portland Cement Company
P.O. Box 947
Colton, CA 92324

Westcott, James F.
A. P. Green Refractories Co.
Green Blvd.
Mexico, MO 65265

Whitson, David
Johns-Manville
Box 338
Antonito, CO 81120

Williams, Barrett
Southern Railway System
125 Spring Street, S.W.
Atlanta, GA 30303

Wilson, Henry
Bureau of Land Management
Box 1449
Santa Fe, NM 87501

Wise, Sanford
Mountain Minerals Co., Ltd.
P.O. Box 700, 529 6th Street
Lethbridge, Alberta TIJ 1L2
Canada

Wolfe, Steven W.
Cyprus Industrial Minerals
P.O. Box 3299
Englewood, CO 80155

Zupan, Alan-Jon
South Carolina Geological
Survey Harbison Forest Road
Columbia, SC 29210

Chronology of the Forums on Geology of Industrial Minerals

(continued from inside front cover)

Year, place, theme	Sponsor	Source of proceedings
Thirteenth, 1977, Norman, Oklahoma "Gypsum, silica-rich sediments, natural brines, and energy in the industrial-minerals industry"	Oklahoma Geological Survey and University of Oklahoma	Oklahoma Geological Survey 830 Van Vleet Oval, Rm. 163 Norman, Oklahoma 73069 (Circular 79, 107 p.)
Fourteenth, 1978, Albany, New York "Geology of industrial minerals in the 1970's"	New York State Geological Survey, Empire State Concrete & Aggregate Producers Assoc., State Univ. of New York at Albany, New York Dept. of Transportation, New York Dept . of Environmental Conservation, Renssalaer Polytechnic Inst., and Dunn Geoscience Corp.	New York State Geological Survey State Education Building Albany, New York 12234 (Bulletin 436, 120 p.)
Fifteenth, 1979, Golden, Colorado "Industrial minerals in Colorado and the Rocky Mountain region"	Colorado Geological Survey, Colorado School of Mines Research Inst., and Colorado Sand & Gravel Producers Assoc.	Colorado Geological Survey 1313 Sherman St., Rm. 715 Denver, Colorado 80203 (Resource Series 8, 140 p.)
Sixteenth, 1980, St. Louis, Missouri "Geology of barite, fireclay, and stone deposits" and "Planning for industrial-mineral development"	Missouri Geological Survey, Dept. of Natural Resources, and University of Missouri (Rolla)	Missouri Geological Survey Dept. of Natural Resources Division of Geology and Land Survey P.O. Box 25 Rolla, Missouri 65401 (Information Circular 28, in press)
Seventeenth, 1981, Albuquerque, New Mexico "Industrial rocks and minerals of the Southwest"	New Mexico Bureau of Mines and Mineral Resources	New Mexico Bureau of Mines and Mineral Resources Campus Station Socorro, New Mexico 87801 (Circular 182)
Eighteenth, 1982, Bloomington, Indiana "Construction materials"	Indiana Geological Survey and Dept. of Geology, Indiana University	Indiana Geological Survey 611 N. Walnut Grove Bloomington, Indiana 47401

Nineteenth, 1983, Toronto, Ontario

Twentieth, 1984, Maryland Twenty-

first, 1985, Arizona

INDUSTRIAL MINERALS OF THE SOUTHWEST

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